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The codes and methods of plotting given in Section III have been superseded.

Details of current codes are given in the *Admiralty List of Radio Signals*, Vol III.

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Preface and Acknowledgements

The Admiralty Weather Manual consists of three sections :—

Section I containing detailed instructions for the taking of meteorological observations at sea, the use and care of meteorological instruments carried in H.M. Ships, and the forwarding of weather reports to the appropriate shore meteorological services.

Section II deals with the physical aspects of meteorology and forms the theoretical basis of

Section III Synoptic Meteorology. This section contains instructions for the plotting of weather charts, followed by discussions of pressure systems and the polar front. It concludes with a chapter on general forecasting, illustrated by a number of specimen charts and forecasts.

Thanks are due to the following societies, publishers and authors for permission to include brief extracts from the books and papers named :—

To the Oxford University Press, " Meteorology," by D. Brunt, " Climate," by W. G. Kendrew, and " The Climates of the Continents," by W. G. Kendrew.

To the Cambridge University Press, " Physical and Dynamical Meteorology," by D. Brunt.

To Sir Napier Shaw and Messrs. Constable & Co., Ltd., " Forecasting Weather," by Sir Napier Shaw.

To Messrs. Blackie & Sons, Ltd., " Meteorology," by A. E. M. Geddes.

To Messrs. Kegan Paul, Trench, Trubner & Co., " Weather," by Abercromby and Goldie.

To Lt.-Col. E. Gold and the Royal Meteorological Society, " Fronts and Occlusions," by E. Gold.

To the Royal Aeronautical Society, " Line Squalls," by M. A. Giblett.

To the Royal Society, Sir George Simpson and F. J. Scrase, " The Distribution of Electricity in Thunderclouds," by Simpson and Scrase.

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SECTION I

METEOROLOGICAL OBSERVATIONS AND INSTRUMENTS

CHAPTER I.

Instrumental Observations

Observations are made and records kept of weather experienced by H.M. Ships for two purposes, firstly as a basis on which forecasts of future weather can be prepared and secondly for extraction for climatological purposes when the ships' meteorological logs are received in the Admiralty.

Observations may be divided into two classes, surface and upper air. The former are concerned with the weather experienced on board ships, the latter with wind, temperature and humidity observed by soundings of the upper atmosphere by aircraft or pilot balloons. Some elements of weather such as pressure and temperature are of a quantitative nature and require for their observation scientific instruments specially adapted for the purpose. Some of these instruments are self-recording, that is, they furnish continuous records of the meteorological elements they measure and thus enable the forecaster to see at a glance what changes have been taking place during the last few hours. Other elements of weather such as visibility, cloud-form and weather itself do not require instrumental aid for their observation. Such observations are termed non-instrumental observations.

It is difficult in H.M. Ships to arrange a satisfactory exposure for the wind-vane of the wind-speed and -direction recorders and the bulbs of the distant reading thermograph so that they will be free from the effects of eddies, funnel smoke and the heating of the metal parts of the ship. The recording parts of these instruments are normally placed in the ship's meteorological office, which is usually situated in the fore superstructure, but even in this position the effects of vibration and movement of the ship may be appreciable. It must be accepted therefore, that records will sometimes be unreliable owing to the above-mentioned causes.

H.M. Ships' List of Observations and Instruments.

Observations.	**Instruments.**
Pressure	Mercury Barometer with Gold Slide.
	Aneroid Barometer.
	Barograph.
	Open Scale Barograph.
Temperature and Humidity of the Air	Dry- and Wet-bulb Thermometers in Portable Screen.
	Electric Psychrometer.
	Whirling Psychrometer.
	Distant Reading Thermograph with dry and wet bulbs.

H.M. SHIPS' LIST OF OBSERVATIONS AND INSTRUMENTS.—*contd.*

Observations.	**Instruments.**
Sea Temperature	Sea Thermometer with canvas bucket.
	Distant Reading Sea Thermograph.
Wind	Wind-Speed Recorder (Munro).
	Portable Wind-Speed Indicator.

ATMOSPHERIC PRESSURE.

The Mercury Barometer (*Pattern No.* 512).—In its simplest form the mercury barometer consists of a glass tube about 33 inches long, closed at one end and filled with mercury. The tube is inverted with its lower open end dipping into a metal cistern, also containing mercury, which is open to the atmosphere. The tube being held vertically, the mercury column in it falls to a height of about 30 inches leaving an air-free space at the upper closed end. The weight of the mercury column is balanced by the pressure of the atmosphere on the mercury in the cistern thus affording a measure of the pressure in terms of the length of the column. The mercury barometer as supplied to H.M. Ships is designed to facilitate the measurement of length of the mercury column and the conversion of the measured length into pressure units; it is shown in Fig. 1.

The cistern is made of stainless steel which ensures the mercury being kept as clean as possible. It is closed, apart from a small hole in its upper end which permits communication with the external air. The escape of mercury through the hole is prevented by a washer of leather which is permeable to air but not to mercury. The barometer tube is enclosed in an outer tube of brass, portions of which are cut away near the upper end in order to render the top of the mercury column visible. This portion of the brass tube on which is engraved the scale for measuring the length of the mercury column is enclosed, with the vernier, in a larger glass tube or shade. The vernier slide consists of two portions accurately aligned, one in front of the mercury column and the other behind it. It is actuated by a milled screw-head carried on the outer brass tube. The temperature of the instrument, on which depend the corrections to be applied to readings is measured by a small thermometer fixed to the front of the outer tube and known as the attached thermometer. In barometers now issued to H.M. Ships the attached thermometer is now incorporated with the "Gold slide" which enables the correction to be read directly.

For the greater portion of its length the bore of the tube is constricted to ·063 inch diameter and a short length consists of fine capillary tube. The object of these constrictions is to reduce the amount of pumping, that is the oscillation of the top of the mercury column due to the movement of the ship in a seaway and

To face page 2.

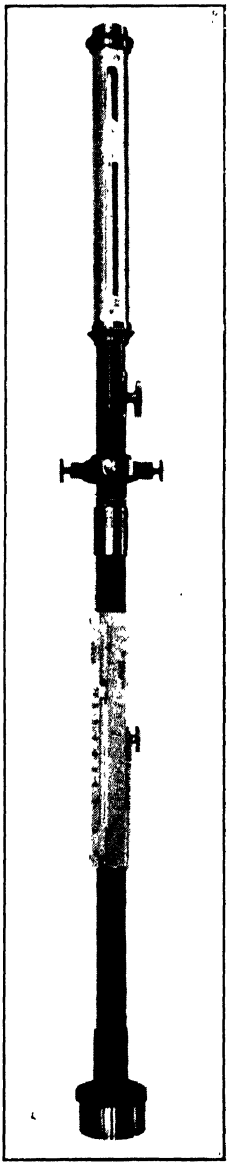

FIG. 1. MERCURY BAROMETER WITH GOLD SLIDE.

the gustiness of the wind. The constrictions render the barometer sluggish in responding to changes of pressure and it may be assumed that the pressure obtained from reading the barometer is equal to the actual pressure about five minutes previously. Since normally the change of atmospheric pressure in a period of five minutes is too small to be detected the effect of lag is negligible. At the top of the mercury column the bore of the tube is of ·315 inch diameter, which gives a sufficiently pronounced curvature to the upper surface of the mercury to make accurate reading possible.

In order to prevent air and moisture finding their way into the tube there is sufficient mercury in the cistern to cover the lower end of the tube at whatever angle the instrument may be inclined to the vertical. As a further precaution against air which may accidentally find its way into the vacuum space at the top of the mercury column, an air trap is provided at the lower end of the constricted portion of the tube. The principle of the air trap is shown in section in Fig. 2.

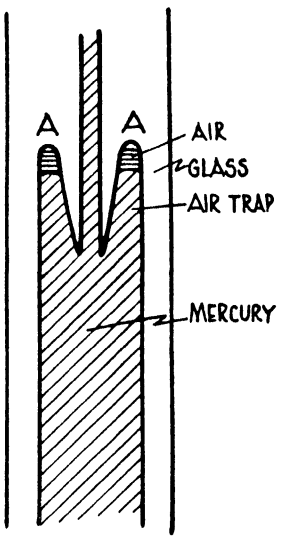

FIG. 2—BAROMETER AIR-TRAP

Air which may bubble up through the mercury in the wide portion of the tube is diverted by the funnel-shaped trap and imprisoned in the space *A* where it cannot affect the readings appreciably.

Barometer Scales and Capacity Effect.—Formerly barometer scales were graduated in units of length (inches or millimetres) and pressure was specified in inches or millimetres of mercury. In modern instruments the scale is graduated in millibars. Zero is reckoned from the level of the mercury in the cistern. This level is affected by the rise and fall of the mercury in the tube so that the position of the zero is variable. This is known as the capacity

effect. Suppose, for example, that the area of the horizontal section of the cistern is 10 times that of the tube. If the level of the mercury in the tube rises one inch that in the cistern will fall 0·1 inch and the difference in the two levels will change by 1·1 inches. A length of 1 inch on the scale will therefore correspond to a difference of pressure of 1·1 inches of mercury. If the same scale is to be used for all pressures its graduations must therefore be " contracted " in a certain proportion depending upon the relative diameters of tube and cistern. In very accurate barometers such as the Fortin, a constant zero, marked by the tip of a small ivory pointer, is used. Before each reading the level of the mercury in the cistern is adjusted to touch the pointer by means of a screw plunger.

Mounting the Barometer.—A bracket carrying gimbals is supplied for mounting the barometer. The gimbal ring is carried on the outer brass tube of the instrument. The bracket should be fixed to the bulkhead in the selected position, the barometer removed carefully from its box, brought gradually into an upright position and shipped. If the mercury does not immediately leave the top of the tube the cistern should be gently tapped with the fingers. No reading should be made until the instrument has been allowed time to take up the temperature of its surroundings. The readings may not be regarded as entirely free from the temperature effect until the lapse of a few hours after mounting.

The barometer should be mounted where it can swing freely and is not liable to interference from passing traffic. It may be necessary to provide guard rails. A position which might at any time be in direct sunshine, or one where the barometer would be subject to any form of artificial heating is to be avoided, the aim being to secure a position of as nearly as possible uniform temperature. Preferably the light should fall from behind the observer when a reading is being taken.

Care of Barometers.—The barometer is an extremely delicate and sensitive instrument and on no account should it be subject to violent jars or concussion. It should therefore always be unshipped while heavy guns are being fired. The barometer should be lifted out of its mounting, inclined gradually so that the mercury flows very gently to the top of the tube and then laid in its box in the horizontal position. If the barometer is inclined quickly the mercury will strike the top of the tube with considerable violence, causing possible breakage, since there is no cushion of air in the space above the column of mercury. The barometer should be carried with the cistern end upwards. If the instrument is packed for transport the box containing it should be placed in a packing case surrounded by soft packing material. The lid of the case should be screwed and not nailed down. Boxes and packing cases should always carry the label at the cistern end with the words " Delicate

instruments. Keep lying flat or this end upwards." The handle of a barometer box is so placed that when carried the box is slightly inclined with the cistern end uppermost. A barometer with a broken tube should never be packed until all the mercury has been collected and put into a strong glass or stone bottle which should be securely corked and packed with the barometer. Apart from the question of economy, mercury will attack the brass work of the barometer and the silvered scales if allowed to come into contact with them.

Barometer Corrections.—A barometer reading does not give directly the true pressure at the position of the instrument. Certain corrections have to be applied on account of :—

(*a*) Index error.
(*b*) Variations in the temperature of the barometer.
(*c*) The value of gravity.

Finally in order that the pressures at different stations may be comparable they must be reduced to a common level. Mean sea level has been selected. The variation of pressure with altitude amounts to 1 millibar in 30 feet while the horizontal variation is normally about 1 millibar in 30 miles. A synoptic chart on which true pressures were plotted would therefore exhibit little else than the variations of altitude of the stations and the horizontal distribution of pressure would be completely masked. Hence the remaining correction to be applied to a barometer reading is :—

(*d*) Reduction to mean sea level.

(*a*) *Index Error.*—Almost all barometers are subject to small errors on account of faulty graduation. Other errors arise from the fact that the barometer tube exerts a downward pull on the top of the mercury column due to the surface tension of the mercury, causing the mercury to be very slightly depressed. This effect is known as capillarity. The two errors are usually considered together under the name index errors.

Before a barometer is issued for use its index error at different parts of the scale is determined by comparison with an accurate standard barometer at the National Physical Laboratory (N.P.L.). A certificate showing the correction for index error is issued with each tested barometer. Alternatively the certificate may show the standard temperature at different points of the scale. The N.P.L. certificate should always be kept with the barometer, and returned with it.

(*b*) *Temperature of the Instrument.*—The temperature of a barometer affects its reading in two ways :—

(i) by causing variations in the density of the mercury and therefore in the height of the mercury column for a given pressure.

(ii) by causing variations in the length of the scale.

It is possible to construct a barometer so that it shall read correctly at any one specified temperature under standard gravity (at M.S.L. in latitude 45°). For, given the temperature and the value of gravity, the height of the mercury column can be calculated for each pressure, and the brass scale, at that temperature, can be graduated accordingly. The temperature at which a barometer reads correctly in latitude 45° at M.S.L. is called its " Standard Temperature." Millibar barometers are now constructed to have a standard temperature (apart from small variations due to index error) of 285°A. (see page 115).

If the temperature of a barometer which reads correctly at 285°A. changes to 286°A. (for example) while the pressure remains constant, the changes in the lengths of the mercury column and the brass scale can be calculated using the values of the coefficient of expansion of mercury in glass and of brass. The new reading of the instrument is thereby determined, and from it by subtraction of the true pressure, the correction to be applied. The correction at any other temperature can be calculated in the same way and a table drawn up showing corrections at different temperatures and indicated pressures. For temperatures above 285°A. under standard gravity the barometer reading is higher than the true pressure and *vice versa*. The temperature corrections of the barometer are given in Table I.

Table I.

Temperature Correction of the Kew Pattern Barometer.

(Millibar graduations.)

Corrections to be applied to the readings of *Kew Pattern Barometers* to reduce them to 285°A.

If the temperature of the attached thermometer is {above / below} 285°A. {subtract / add} the correction.

Attached thermometer (**add** correction).	BAROMETER READINGS (mb.).										Attached thermometer (**subtract** correction).
	860	880	900	920	940	960	980	1000	1020	1040	
°A.	Corrections (mb.)										°A.
284	·15	·15	·15	·16	·16	·16	·17	·17	·17	·18	286
283	·30	·30	·31	·32	·32	·33	·34	·34	·35	.36	287
282	·44	·45	·46	·47	·48	·49	·50	·51	·52	·53	288
281	·59	·61	·62	·63	·64	·66	·67	·68	·70	·71	289
280	·74	·76	·77	·79	·81	·82	·84	·86	·87	·89	290
279	·89	·91	·93	·95	·97	·99	1·01	1·03	1·05	1·07	291
278	1·04	1·06	1·08	1·11	1·13	1·15	1·17	1·20	1·22	1·24	292
277	1·19	1·21	1·24	1·26	1·29	1·32	1·34	1·37	1·39	1·42	293
276	1·33	1·36	1·39	1·42	1·45	1·48	1·51	1·54	1·57	1·60	294
275	1·48	1·51	1·55	1·58	1·61	1·64	1·68	1·71	1·74	1·78	295
274	1·63	1·66	1·70	1·74	1·77	1·81	1·85	1·88	1·92	1·95	296
273	1·78	1·82	1·86	1·90	1·93	1·97	2·01	2·05	2·09	2·13	297
272	1·93	1·97	2·01	2·05	2·10	2·14	2·18	2·22	2·27	2·31	298
271	2·08	2·12	2·17	2·21	2·26	2·30	2·35	2·39	2·44	2·49	299
270	2·22	2·27	2·32	2·37	2·42	2·47	2·52	2·57	2·61	2·66	300
269	2·37	2·42	2·48	2·53	2·58	2·63	2·68	2·74	2·79	2·84	301
268	2·52	2·57	2·63	2·69	2·74	2·80	2·85	2·91	2·96	3·02	302
267	2·67	2·73	2·78	2·84	2·90	2·96	3·02	3·08	3·14	3·20	303
266	2·82	2·88	2·94	3·00	3·06	3·13	3·19	3·25	3·31	3·37	304
265	2·97	3·03	3·09	3·16	3·22	3·29	3·35	3·42	3·49	3·55	305
264	3·11	3·18	3·25	3·32	3·39	3·45	3·52	3·59	3·66	3·73	306
263	3·26	3·33	3·40	3·48	3·55	3·62	3·69	3·76	3·83	3·91	307
262	3·41	3·48	3·56	3·63	3·71	3·78	3·86	3·93	4·01	4·08	308
261	3·56	3·63	3·71	3·79	3·87	3·95	4·03	4·10	4·18	4·26	309
260	3·71	3·79	3·87	3·95	4·03	4·11	4·19	4·28	4·36	4·44	310
259	3·86	3·94	4·02	4·11	4·19	4·28	4·36	4·45	4·53	4·62	311
258	4·00	4·09	4·18	4·26	4·35	4·44	4·53	4·62	4·71	4·79	312
257	4·15	4·24	4·33	4·42	4·51	4·61	4·70	4·79	4·88	4·97	313
256	4·30	4·39	4·49	4·58	4·68	4·79	4·86	4·96	5·05	5·15	314
255	4·45	4·54	4·64	4·74	4·84	4·93	5·03	5·13	5·23	5·33	315

(*c*) *Gravity.*—The weight of a column of mercury of given height and therefore the pressure required to balance it depends upon the value of **g**. There is a variation of **g** with latitude from a maximum at the poles to a minimum at the equator. The value in latitude 45° has been selected as standard for which barometers are constructed to read correctly at their standard temperatures. In latitudes below 45° the barometer reads too high and *vice versa*. The amount of the correction may be calculated, and is given in Table II.

TABLE II.

CORRECTIONS FOR REDUCING BAROMETRIC READINGS TO STANDARD GRAVITY IN LATITUDE 45°.

For latitudes 0° to 44° the correction is to be **subtracted.**
For latitudes 46° to 90° the correction is to be **added.**

Latitude N. or S.		Pressure in millibars.			Pressure in mercury inches.				Latitude N. or S.	
		950	1000	1050	28	29	30	31		
°	°	Corrections (millibars)			Corrections (mercury inches)				°	°
45	45	·00	·00	·00	·000	·000	·000	·000	45	45
44	46	·09	·09	·09	·003	·003	·003	·003	44	46
43	47	·17	·18	·19	·005	·005	·005	·006	43	47
42	48	·26	·27	·28	·008	·008	·008	·008	42	48
41	49	·34	·36	·38	·010	·010	·011	·011	41	49
40	50	·43	·45	·47	·013	·013	·013	·014	40	50
39	51	·51	·54	·57	·015	·016	·016	·017	39	51
38	52	·60	·63	·66	·018	·018	·019	·019	38	52
37	53	·67	·71	·75	·020	·021	·021	·022	37	53
36	54	·76	·80	·84	·022	·023	·024	·025	36	54
35	55	·85	·89	·93	·025	·026	·027	·027	35	55
34	56	·92	·97	1·02	·027	·028	·029	·030	34	56
33	57	1·00	1·05	1·10	·029	·031	·032	·033	33	57
32	58	1·08	1·14	1·20	·032	·033	·034	·035	32	58
31	59	1·16	1·22	1·28	·034	·035	·036	·038	31	59
30	60	1·24	1·30	1·37	·036	·038	·039	·040	30	60
29	61	1·30	1·37	1·44	·038	·040	·041	·043	29	61
28	62	1·38	1·45	1·52	·041	·042	·043	·045	28	62
27	63	1·44	1·52	1·60	·043	·044	·046	·047	27	63
26	64	1·51	1·59	1·67	·045	·046	·048	·049	26	64
25	65	1·58	1·66	1·74	·047	·048	·050	·052	25	65

TABLE II.—*continued.*

Latitude N. or S.		Pressure in millibars.			Pressure in mercury inches.				Latitude N. or S.	
		950	1000	1050	28	29	30	31		
°	°	Corrections (millibars)			Corrections (mercury inches)				°	°
24	66	1·64	1·73	1·82	·049	·050	·052	·054	24	66
23	67	1·71	1·80	1·89	·050	·052	·054	·056	23	67
22	68	1·77	1·86	1·95	·052	·054	·056	·058	22	68
21	69	1·82	1·92	2·02	·054	·056	·058	·060	21	69
20	70	1·88	1·98	2·08	·056	·058	·060	·062	20	70
19	71	1·94	2·04	2·14	·057	·059	·061	·063	19	71
18	72	2·00	2·10	2·21	·059	·061	·063	·065	18	72
17	73	2·04	2·15	2·26	·060	·062	·064	·067	17	73
16	74	2·09	2·20	2·31	·061	·064	·066	·068	16	74
15	75	2·13	2·24	2·35	·063	·065	·067	·070	15	75
14	76	2·18	2·29	2·40	·064	·066	·069	·071	14	76
13	77	2·21	2·33	2·45	·065	·068	·070	·072	13	77
12	78	2·25	2·37	2·49	·066	·069	·071	·073	12	78
11	79	2·28	2·40	2·52	·067	·070	·072	·074	11	79
10	80	2·31	2·43	2·55	·068	·071	·073	·075	10	80
9	81	2·34	2·46	2·58	·069	·071	·074	·076	9	81
8	82	2·36	2·49	2·61	·070	·072	·075	·077	8	82
7	83	2·38	2·51	2·64	·070	·073	·075	·078	7	83
6	84	2·40	2·53	2·66	·071	·073	·076	·079	6	84
5	85	2·42	2·55	2·68	·071	·074	·077	·079	5	85
4	86	2·43	2·56	2·69	·072	·074	·077	·080	4	86
3	87	2·44	2·57	2·70	·072	·075	·077	·080	3	87
2	88	2·45	2·58	2·71	·072	·075	·078	·080	2	88
1	89	2·46	2·59	2·72	·072	·075	·078	·080	1	89
0	90	2·46	2·59	2·72	·073	·075	·078	·080	0	90

(*d*) *Reduction to Mean Sea Level.*—The pressure at M.S.L. can be found by adding to the true pressure at barometer level the pressure due to the weight of the air column between the instrument and M.S.L. This correction depends upon the height, temperature and pressure of the air column, the last two factors affecting its density. At a land station, where the barometer may be several

hundreds of feet above sea level, temperature and pressure of the air column must be taken into account, but for a barometer on board ship, where the height of the instrument is less than 100 feet they may be neglected. The maximum correction for 100 feet under the extreme conditions of high pressure and low temperature likely to be experienced over the sea is 3·9 millibars, while the minimum is 3·3 millibars. The extreme correction therefore never differs by more than 0·3 millibar from the mean of 3·6 millibars. Since owing to pumping, barometer readings on board ship are often reliable only to the nearest millibar, the mean correction of 3·6 millibars per 100 feet, or approximately 1 millibar per 30 feet is sufficiently accurate under all conditions.

In practice all barometer corrections may be conveniently applied by expressing them first as temperature corrections. Suppose for example that the index error of a barometer at 1,000 mb. is + 0·5 mb. Table I shows that at 282°A. and 1,000 mb. a barometer with zero index error reads 0·5 mb. low, and therefore for the given barometer the temperature correction annuls the index error at 282° A. and the barometer reads correctly. The standard temperature of the barometer at 1,000 mb. is therefore 282° A. Similarly the standard temperature may be found at other pressures. The standard temperature provides an alternative method of expressing the index error correction and N.P.L. certificates are sometimes given in this form.

Similarly the correction for latitude may be found from Table II, and Table I used to find the temperature at which the correction is equal and opposite to the latitude correction.

Example :—Latitude 60°, pressure 1,000 mb., correction (Table II) = + 1·30 mb., *i.e.*, the barometer reads 1·30 mb. low.

From Table I, correction for 293° A. = − 1·37 mb.

Therefore in latitude 60° the barometer reads correctly at 293° A.

The temperature at which a barometer reads correctly in any latitude is called its Fiducial Temperature. It may be calculated from the standard temperature using Table III.

TABLE III.

Latitude	0°	5°	10°	15°	20°	25°	30°	35°	40°	45°
Subtract from the Standard Temperature ..	15°A	15	14	13	11·5	9·5	7·5	5	2·5	0
Latitude	90°	85°	80°	75°	70°	65°	60°	55°	50°	45°
Add to the Standard Temperature	15°A	15	14	13	11·5	9·5	7·5	5	2·5	0

Reference to Table I shows that a correction of 1 mb. corresponds to a difference of 6° A. (approx.) from the standard temperature. But 1 mb. is also the correction for a height of 28 feet and so,

correction for 6° A. = correction for 28 feet
or correction for 1° A. = correction for 5 feet (approx.).

Height above M.S.L. may therefore be corrected for by adding to the fiducial temperature 1° A. for every 5 feet above M.S.L. The result so obtained is called the Adjusted Fiducial Temperature (A.F.T.) and is the temperature at which the barometer reads correct M.S.L. pressure although it may itself be above sea level.

Finally, the instrument will generally be at a temperature, shown by the attached thermometer, different from the adjusted fiducial temperature. Since a correction of 1 mb. corresponds to a temperature difference of 6° A., the difference between the reading of the attached thermometer and the adjusted fiducial temperature, divided by 6 gives the correction in millibars to be applied to the barometer reading.

The procedure for calculating the barometer correction may be summarized as follows :—

(1) Determine the fiducial temperature (F.T.) for the latitude using Table III and the standard temperature as shown by the N.P.L. certificate and stamped on the instrument.

(2) Increase the F.T. by 1° A. for every 5 feet above sea level to obtain the adjusted fiducial temperature.

(3) Take the difference A.F.T.—attached thermometer reading and divide by 6.

(4) Add the result (with due regard to sign) to the actual barometer reading, to obtain the pressure reduced to M.S.L.

Example :—Barometer reading, 1,012·5 mb.
Height, 30 feet above sea level.
Latitude, 52° N.
Attached thermometer reads 290° A.
Standard temperature, 286° A.

From Table III by interpolation, F.T. = 286° + 3·5° A.
= 289·5° A.
Adding 6° A. for height 30 feet, A.F.T. = 295·5° A.
A.F.T.—Attached Thermometer reading = +5·5
∴ Correction = $+\frac{5\cdot5}{6}$ mb.
= +0·9 mb.

The true pressure reduced to M.S.L. is therefore
1,012·5 + 0·9 mb.
= 1,013·4 mb.

The Gold Slide.—Barometer corrections at sea may be obtained very simply by means of the Gold Slide, so called after its inventor, Lt.-Col. E. Gold, D.S.O., F.R.S. The slide is illustrated in Fig. 3. It consists of a stock *S* with which is incorporated the attached thermometer *A*, and a slider *G*, which can be moved by means of a

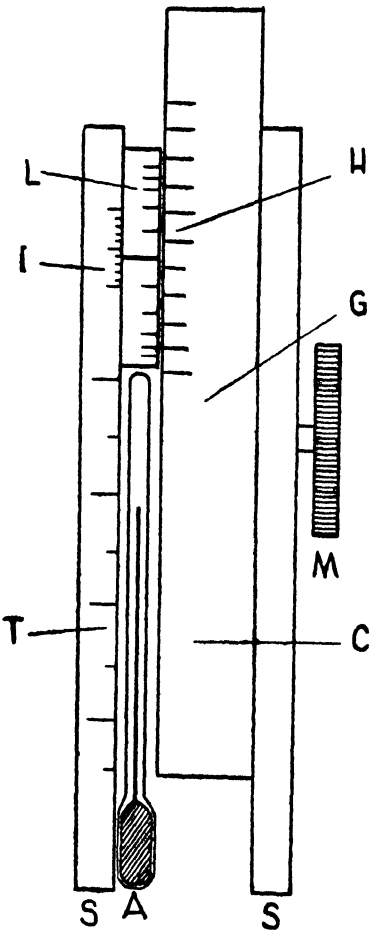

FIG. 3.—GOLD SLIDE.

rack and pinion actuated by a milled screw-head *M*. A pair of lugs with screws and a curved balance weight are provided for fixing the slide firmly to the barometer in the position usually occupied by the attached thermometer. The slide is constructed of brass, faced with silver to facilitate reading of the scales engraved upon it.

The left-hand part of the stock carries a temperature scale *T* alongside the stem of the thermometer. It is essential that the

graduations on this scale coincide exactly with those on the thermometer itself. Above the temperature scale near the top of the stock is a small scale *I* marked "Index" graduated from + 0·5 to — 0·5 millibars. Adjacent to the index scale is a small adjustable strip of metal carrying on its right hand edge a latitude scale *L*, graduated from 0° to 90°. Opposite to latitude 45°, on the left-hand edge of the metal strip is a small red mark. The slider *G* moves between the strip of metal and the right-hand side of the stock. Alongside the latitude scale it carries a height scale *H* graduated for heights 0-100 feet above the water-line. Below the height scale, alongside the thermometer stem is a correction scale *C*, graduated in tenth millibars from + 8·0 to — 7·0 millibars.

The use of the instrument depends upon the fact that all barometer corrections whether for index error, latitude or height are equivalent to certain temperature corrections which can be determined as shown on page 5. The slide is set for index error before issue, by the Meteorological Office. The short red mark opposite to latitude 45° on scale *L* is set opposite to the mean value of the index error, on scale *I*, and this setting is fixed by means of two screws at the back of the strip of metal. The strip should not be moved so long as the Gold slide is used with the same barometer, since it is permanently set for the index error of the particular instrument to which it is attached. To take a reading, the slide is set by means of the milled screw-head *M* so that the height of the barometer cistern above the water-line on scale *H* comes opposite to the latitude on scale *L*. The barometer correction is then read off on scale *C* in line with the top of the mercury column in the thermometer stem. A red figure on scale *C* indicates that the correction is to be subtracted from the barometer reading, a black figure that it is to be added. After the slide is set for latitude it is advisable to remove the milled screw-head so that the adjustment cannot be tampered with.

The Gold slide effects the corrections with varying degrees of exactness. The index correction is the mean for the whole range of pressures, the latitude correction is exact only at 1,000 mb. while the correction for height takes no account of the temperature and pressure of the air column between the barometer and sea level. The correction for temperature of the instrument is effected exactly under all conditions.

Care and Adjustment of the Gold Slide.—The scales should be kept clean by using a soft cloth to remove any dust or dirt. Metal polish should never be applied. A very little clock-oil may occasionally be used for lubrication. If the rack and pinion become very stiff they should be overhauled as follows. Remove the slide from the barometer, place it face downwards and by taking out the four screws remove the brass block securing the pinion in position. Wipe the pinion and its bearing with a soft rag to remove dust and

old oil ; apply a little fresh clock-oil. Now remove the four small screws, two at each end of the rack. The slider can then be taken out. Wipe off all dirt and old oil from the rack and bearing surfaces. Put a little fresh clock-oil on the rack and on the back of the slider. Re-assemble, taking care that the pinion is properly engaged in the rack before tightening the screws. The screws which secure the latitude scale should not be touched during this operation.

The Gold slide should be tested for correctness of adjustment as follows :—Bring the zero on the height scale opposite to 45° on the latitude scale, the zero on the correction scale should then be opposite to the standard temperature (mean for all pressures) on the temperature scale.

Reading the Barometer.—A light should be fitted where it will illuminate the white enamel of the bulkhead behind the barometer thus affording a contrasting background so that the top of the mercury column can be clearly seen while the barometer is being read ; the light should be switched off after the reading has been made.

The correction should be read from the Gold slide, to the nearest tenth of a millibar, before the barometer itself is read. If no Gold slide is fitted, the correction then being worked out as on page 10, the attached thermometer should be read first. Care must be taken to keep the eye at the same level as the top of the mercury thread in the thermometer stem, to avoid the error of parallax. If the reading of the correction is delayed the thermometer is likely to be affected by the proximity of the observer's body, more readily than the barometer itself, and will no longer give the temperature of the instrument as a whole.

To read the barometer, first tap the outer case very gently (steadying the barometer meanwhile so that it does not swing) until tapping no longer affects the shape of the top of the mercury column in the tube. Then turn the milled head actuating the vernier until its lower edge and the lower edge of the rear sliding piece which moves with the vernier, appear in line with the uppermost part of the domed surface of the mercury. This procedure automatically avoids the error of parallax. The barometer should then be read to the nearest tenth of a millibar, and the correction added or subtracted as the case may be to give the pressure reduced to M.S.L.

Care should be taken to avoid the common errors of 10, 5 or 1 millibar.

If the barometer is pumping, tapping the instrument, which is a precaution taken to avoid the effects of the possible sticking of the mercury in the tube, is unnecessary. Pumping arises from two causes. The first is the fluctuation of pressure inside the compartment in which the barometer is mounted, due to the gustiness of the wind. It may be avoided by closing all doors and scuttles of the

compartment while the barometer is being read. It is just as important to shut the lee door as the weather door. The second cause of pumping is rolling and pitching which affects the barometer in three ways :—

(*a*) Owing to the mobility of the mercury.
(*b*) Owing to the swinging of the barometer.
(*c*) Owing to changes of pressure due to the up and down movement of the ship.

No entirely satisfactory method of dealing with pumping has yet been devised, but reasonable accuracy may be obtained by taking the mean of the highest and lowest positions reached by the top of the mercury column during a period of one minute.

Checking a Barometer.—For the success of the scheme for weather reports from H.M. Ships to shore services, it is essential that all barometers used should be reading accurately. In order to provide facilities for checking barometer readings, standard barometers have been installed at the following ports :—

Home Station	..	Chatham, K.H.M's office.
		Sheerness, K.H.M's office.
		Devonport, K.H.M's office.
		Portsmouth, K.H.M's office.
Mediterranean	..	Gibraltar, K.H.M's office.
China, Singapore	..	K.H.M's office.
Shanghai	..	Resident Naval Officer.
American and West Indies		Bermuda, K.H.M's office.
		Halifax, Captain of the Dockyard.
		Esquimalt, Captain of the Dockyard.
Africa ..	..	Simonstown, Captain of the Dockyard.

Standard barometers are also installed at the following ports and corrected M.S.L. pressures will be supplied at any time by the shore authorities indicated, upon application.

Mediterranean	..	Malta, Meteorological Office, Malta.
East Indies	..	Colombo, The Observatory.
China ..	..	Hong Kong, Royal Observatory.

The check should consist of the comparison of the ship's barometer with the shore barometer over a series of simultaneous readings taken when pressure is not altering rapidly and when the barometer is not pumping.

In ports where there is a large rise and fall of tide the comparisons should be carried out at half-tide, as, otherwise, a correction will be necessary for the difference between mean sea level and the height of the tide at the time of observation. This correction amounts to 0·3 mb. for each 10 feet difference in the two heights.

A useful though less accurate check is afforded by comparing the reading of the ship's barometer at the time local observations are taken, with the isobar drawn through the ship's position in the local synoptic chart. Thus in Great Britain, a reading on board at 0700 G.M.T. can be compared with the Daily Weather Report which contains a synoptic chart for that hour. The synoptic charts for 1800 G.M.T. reproduced in the morning newspapers on the following day also afford a ship an opportunity of checking the barometer. In this method interpolation is necessary unless an isobar actually runs through the port at which the ship is lying.

The barometer should be replaced if the mean of at least three differences between the ship and shore barometers is 1 mb. or more.

The Aneroid Barometer (*Pattern No.* 501 and *Pattern No.* 502). —The aneroid barometer is an instrument which, though less reliable as a pressure indicator than the mercury barometer, has the great advantages of compactness and portability, and can therefore be used in positions in which it would be impossible to set up the latter instrument. The sensitive element is a partially exhausted disc-shaped metal box, which undergoes changes of shape resulting from variations of pressure of the external air. The ends of the box are usually corrugated to give greater flexibility. A control spring prevents complete collapse of the box under the external pressure. The small movements of the box are transmitted by a system of magnifying levers to a pointer which, moving over a graduated dial, indicates the air pressure to which the instrument is subject. A movable pointer operated by means of a milled head may be set to coincide with the aneroid pointer so that subsequent changes of pressure may be read directly. The instrument should be gently tapped before reading.

The Pattern No. 501 aneroid is graduated in millibars and inches and is certified by the National Physical Laboratory; the Pattern No. 502 instrument is graduated in millibars only and is not certified. The range of pressure is from 925 to 1,055 millibars.

The expansion with rise of temperature causes flexural deformations of the circular ends of the aneroid box so that they curve inwards. In the absence of any compensating device the indications of the instrument would therefore change with the temperature even though the pressure remained constant. The instruments supplied to H.M. Ships are compensated for temperature, the method usually adopted being to make the long arm of the lever system of two metals brazed together longitudinally. Owing to the different expansions of the two metals, changes of temperature cause bending, or in effect, changes in the length of the bi-metallic arm. The change of length is so arranged as to annul the effect of distortion of the aneroid box. This form of compensation is exact only at a particular pressure and if the aneroid box is completely exhausted. It is, however, sufficiently accurate for all practical purposes. A

simpler but less accurate method of compensation consists in leaving in the aneroid box a small amount of air, whose increase of pressure with rise of temperature counteracts the tendency of the box to distort.

Similarly constructed aneroid boxes behave differently when subject to the same changes of pressure and temperature, so that every instrument must be separately calibrated. During use the aneroid barometer is subject to hysteresis lag. That is, if the instrument is subjected to a large increase or decrease of pressure, after which the pressure is brought back to its original value, a considerable time elapses before the barometer returns to its original reading. There is also a secular error due to slow changes in the metal of which the aneroid box is made. This error can only be allowed for by comparison at intervals with a mercury barometer and adjustment of the aneroid pointer by means of a screw, access to which is given by a hole in the back of the case. The aneroid should be set to read M.S.L. pressure as determined in the ordinary way by means of the mercury barometer. It will continue to read sea-level pressure so long as it is not moved from the position in which the comparison was made. It responds almost immediately to changes of pressure. No correction for temperature or latitude is necessary.

The Barograph (*Pattern No.* 503) **and Open Scale Barograph** (*Pattern No.* 515).—The Barograph is an aneroid barometer which records the pressure by means of a pen moving over a revolving drum around which a chart is wrapped. It is a valuable supplement to the mercury barometer as it not only provides a continuous record of pressure for reference purposes but also shows minor fluctuations which are seldom noticeable in the action of the mercury barometer. It also enables the barometric tendency to be more readily obtained.

The instrument should be mounted athwartships on sorbo rubber, in a position where it is least likely to be affected by concussion, vibration or movement of the ship. A position as close as possible to the centre of pitch appears to give the best results; it has been found that the thickening of the trace observed at sea is mainly attributable to the effect of pitching. If the barograph is mounted with the recording pen athwartships there is less likelihood of the pen leaving the chart when the ship is rolling and of the ink splashing from the pen.

The Open Scale Barograph is illustrated in Fig. 4 and is similar to the ordinary barograph except that the time scale is slightly larger and the pressure scale much larger.

A vacuum pile *B*, which may consist either of a number of aneroid boxes, similar to those used in an aneroid barometer, or of a long deeply corrugated tube of thin metal, and which has a

concertina action under changes of atmospheric pressure, operates, through a system of magnifying levers, a long arm *A* carrying a pen *P* filled with special ink. The pen arm moves in a vertical plane and the pen traces out the record of pressure as a continuous line *T* on a printed chart wrapped round a revolving drum driven by clockwork. The pen arm is carried by means of a gate suspension *G*, which may be adjusted so that the pen rests lightly on the chart, and also permits the pen to be withdrawn from the chart when required by means of the rod *R* actuated by the lever *L*. A barograph, since it is simply a recording aneroid barometer, must be compared at intervals with a mercury barometer. Any adjustment necessary should be made by means of the milled head *M* above the aneroid chamber. A push rod *H* actuating another rod *J* bringing it into contact with the lever system at *K* causes the pen to make a vertical time-mark on the chart. The mark should be made at some fixed hour daily (usually 0900), to serve as a check on the time kept by the clock. The times at which the marks are made (to the nearest minute) should be noted and entered on the chart against the time marks when the chart is removed for insertion in the meteorological log. The clockwork may be regulated by moving the pointer which controls the spring of the balance wheel. The four rods *D* at the corners of the base of the instrument act as guides when the case is being removed or replaced.

The charts for both instruments are graduated from 950 to 1,050 millibars. The charts are changed weekly, on Mondays, when the clock should be wound and the record set to correct time. The chart for the open scale barograph, from which the barometric tendency is read, is ruled at 2-hourly intervals, which enables the barometric tendency to be easily read. For the Pattern No. 503 barograph the charts are Pattern No. 514; those for the open scale barograph, Pattern No. 515, are Pattern No. 538. A spare pen is supplied in the drawer in the base of the ordinary barograph. A spare pen arm, Pattern No. 2701, and two spare pens, Pattern No. 2702, are supplied with the open scale barograph.

The notes on the care and upkeep of recording instruments (page 35) apply to the barograph except where it is otherwise stated.

The dimensions and weights of the two types of barograph are as follows :—

	Length (inches)	Width (inches)	Height (inches)	Weight (lbs.)
Barograph, Pattern No. 503	12½	6½	8	9
Open Scale Barograph, Pattern No. 515	15	9	10½	19½

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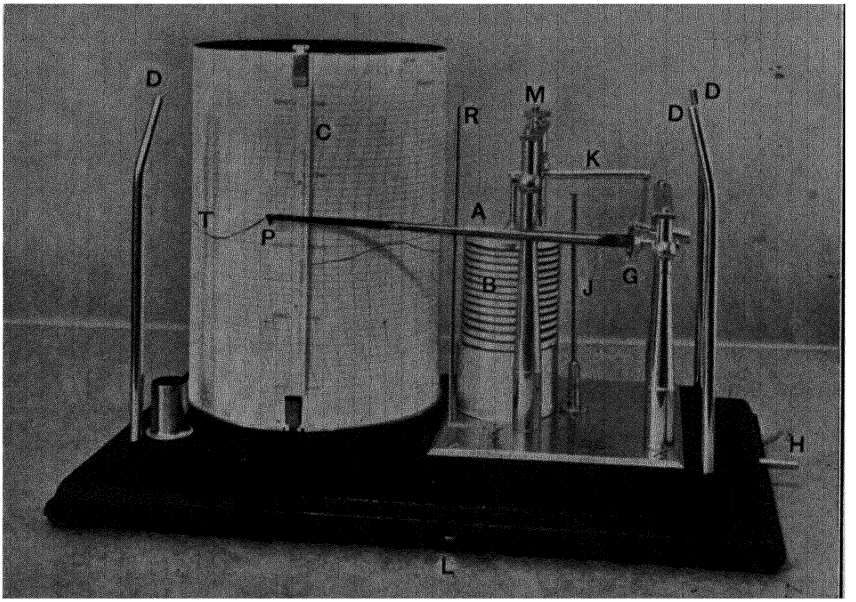

FIG. 4.—OPEN SCALE BAROGRAPH

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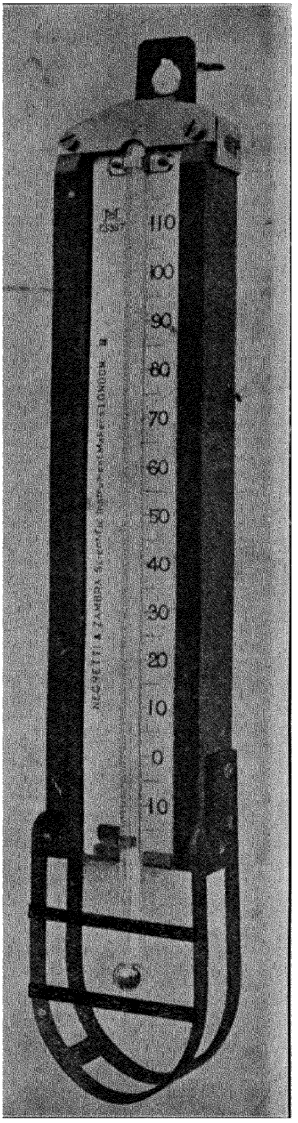
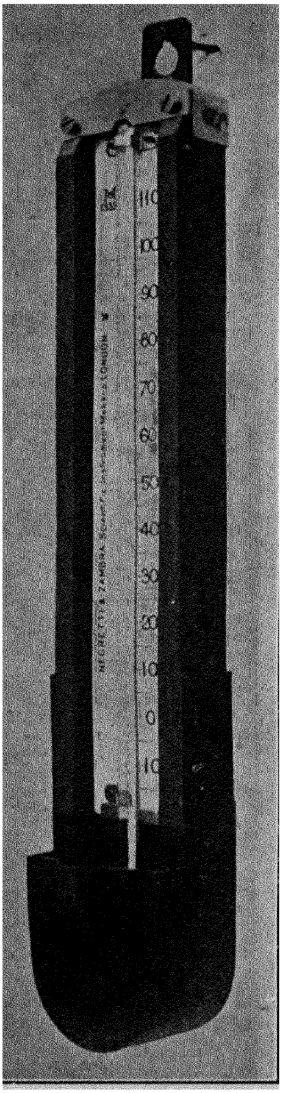


AIR PROTECTOR.



SEA PROTECTOR.

FIG. 5.—MAHOGANY PROTECTORS FOR THERMOMETERS.

TEMPERATURE AND HUMIDITY OF THE AIR.

The Thermometer (*Pattern No.* 504).—Air temperature is usually measured by means of a mercury thermometer. For use in temperate climates the stem of the thermometer should be graduated to read from —15° F. to 115° F. In hot climates an extension of the range up to 140° F. is necessary. Owing to the fact that mercury freezes at —38° F. the mercury thermometer is unsuitable for use in cold climates. Instead, a thermometer filled with ethyl alcohol is used, and can be employed down to —130° F., some 40° F. below the lowest temperature ever recorded on the earth's surface. The thermometer scale should be sufficiently open to allow of readings being made to 1/10° F., and for this reason should not contain more than eighteen 1° divisions to the inch. The graduations are marked on the stem of the thermometer itself, but, for ease in reading, the instrument is fixed to a porcelain base on which also the 5° and 10° graduations are marked. Since exact coincidence of the graduations on the stem and on the porcelain base is difficult of attainment, readings to one-tenth of a degree should be made using the scale on the stem. To reduce the risk of breakage a protector (Fig. 5) consisting of a mahogany frame into which the thermometer can be fitted, together with a metal guard to protect the bulb, is now supplied with each thermometer for use on board H.M. Ships. This is known as an air-thermometer protector, Pattern No. 541. The thermometer is fitted into the protector by unscrewing the metal plate at the top of the protector.

Thermometers should be kept clean, and, in particular, no deposit should be allowed to accumulate on the bulb. The presence of moisture on the bulb of a thermometer will cause it to behave as a wet bulb; the moisture should be carefully wiped off some 15 minutes before a reading is taken. If the graduations on the stem of a thermometer have become indistinct they should be restored by rubbing a soft dark coloured crayon along the stem, the excess of the substance being afterwards removed.

Thermometers issued by the Meteorological Office to H.M. Ships are first subjected to a test at the National Physical Laboratory. An N.P.L. certificate, showing the corrections to be applied at various points of the scale is issued with each tested thermometer. The corrections specified on the certificate should be applied to all readings.

Thermometers should be read as rapidly as possible to minimize the effects of the proximity of the observer, and of radiation from surrounding objects. They should be read with the top of the mercury thread at the same level as the observer's eye to avoid errors of parallax.

Screening of Thermometers.—The measurement of true air temperature is a matter of extreme difficulty owing to the fact that a thermometer freely exposed to the air is affected by radiation of

heat to and from surrounding objects. Heat is radiated along straight-line paths between any objects whose temperatures differ, so that a thermometer will be affected by any object within sight of it, whose temperature differs from its own. The passage of radiant heat through the air, however, leaves its temperature practically unaffected. It is therefore necessary to expose the thermometer in some form of enclosure or screen such as will allow the air to circulate freely past the bulb of the thermometer and yet shield the thermometer from the radiation emitted by its surroundings. The usual form of screen is a wooden box, painted white, with louvered sides. The air can circulate freely between the louvres, and the thermometers within the screen are effectively shielded from external radiation.

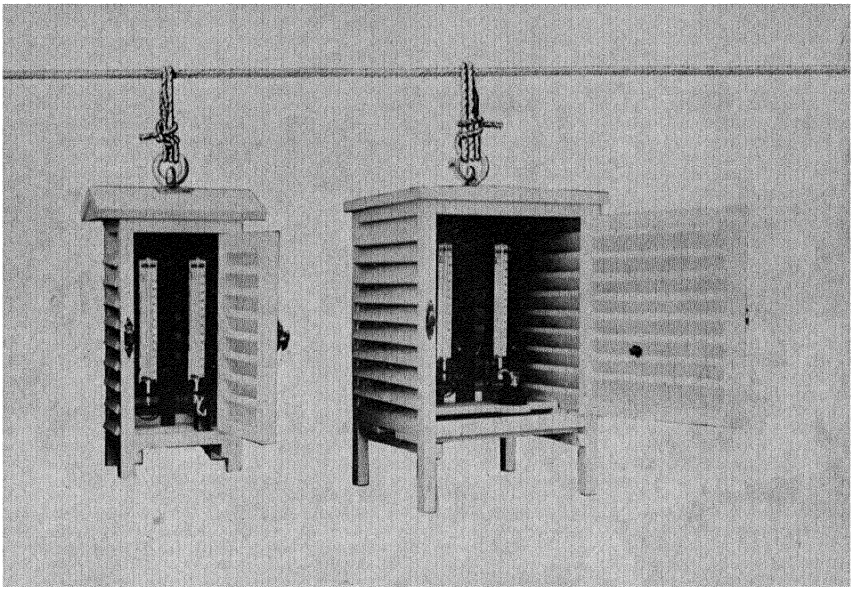
The screen used on board H.M. Ships is the Modified Shipboard Screen, Pattern No. 539. It is illustrated in Fig. 6. The thermometers (dry and wet bulb) are mounted side by side on a wooden support inside the screen. The front of the screen is hinged so as to give access to the interior. The screen is fitted with a ring which enables it to be slung up in a suitable position before an observation is taken ; it should not be permanently fixed.

At least fifteen minutes before an observation is taken the screen should be slung on the windward side of the upper deck, to enable its interior to take up the true air temperature. A position in the path of air which has been heated by contact with the funnels or any portion of the ship must be avoided.

The Wet-Bulb Thermometer.—For measuring the humidity of the air a wet-bulb thermometer is employed in conjunction with an ordinary or dry-bulb thermometer. The wet-bulb thermometer is an ordinary thermometer whose bulb is enclosed in a muslin cap from which a few strands of cotton wick dip down into a small vessel containing pure water. Water passes along the wick to the muslin from which, unless the air is saturated, it evaporates into the air. The latent heat of evaporation is drawn from the air immediately surrounding the wet bulb, causing its temperature to fall. The reduced temperature is that indicated by the wet-bulb thermometer. The lower the relative humidity of the air the faster is the rate of evaporation from the muslin, and the lower does the wet bulb read. Only if the air is saturated will the readings of the dry and wet bulb be the same. Normally the difference between the two readings is a few degrees, but in very dry weather it may amount to 25° F.

Special tables known as hygrometric tables (M.O. 265) are supplied, from which the dew-point, relative humidity and vapour pressure may be determined from the dry-bulb reading and the difference between it and the wet-bulb reading.

The water vessel should be placed below and somewhat to the side of the wet-bulb thermometer in such a way that from 3 to 6



Modified, with
Thermometer Protectors.

Portable, with
Thermometer Protectors.

FIG. 6.—SHIP'S THERMOMETER SCREENS.

inches of the wick is exposed to the air. If the exposed portion of the wick is longer than 6 inches the water will, in dry weather, have completely evaporated from the wick before it reaches the bulb of the thermometer. The wick should hang straight down into the water. If it be allowed to hang in a loop, water will drip from the lowest point of the loop and the water vessel will be rapidly emptied. The neck of the water vessel should be small, or a cover provided through which the wick passes by a small hole, to prevent moistening of the air inside the screen by evaporation of the water in the vessel. The screen should always be exposed so that the air cooled by evaporation from the wet bulb and reservoir will not blow across the dry bulb, *i.e.*, the dry bulb should be to **windward** of the wet bulb.

Pure distilled water or rain water should always be used for the wet bulb. The muslin and wick and, during stormy weather, the water in the vessel may become contaminated with salt from spray. If this is suspected the muslin, wick and water should be changed at once, since the presence of salt is almost certain to invalidate the wet-bulb reading. In any case the muslin and wick should be renewed once a week. Before fitting, the muslin and wick should be washed in water to which a little ammonia has been added, to remove grease. No reading should be taken until at least fifteen minutes have elapsed after fitting. Normally the wet-bulb reading will be lower than or equal to the dry-bulb reading, but occasionally the wet-bulb thermometer may read the higher. This occurs when the temperature is falling rapidly. The dry bulb being freely exposed takes up the air temperature quickly, but the wet bulb, being coated with wet muslin, shows considerable lag. In such a case the readings should be logged as the same.

When the reading of the wet bulb is below 32° F. the water on the wick becomes frozen and the supply of moisture to the muslin is cut off. The bulb becomes coated with a thin covering of ice, from which evaporation takes place just as from a water surface. So long as there is a coating of ice present therefore, the wet-bulb reading is valid, but ultimately the ice disappears. If this has occurred the muslin should be wetted, by means of a camel-hair brush or a feather, with ice-cold water, and a new film of ice allowed to form. Readings should be taken when the wet-bulb temperature has become steady. Very little water should be used in moistening the muslin cap, for if a thick ice coating is allowed to form the wet-bulb readings become invalidated.

If the wet bulb, still being moist, reads considerably below 32° F., the water has probably become super-cooled. No reading should be taken until the water on the muslin has been induced to freeze by touching it with a small portion of ice, snow or hoar frost. Freezing takes place at once, the wet-bulb temperature rises immediately to 32° F. and then falls slowly. When it has reached a steady value the readings should be taken.

The modified shipboard screen complete comprises :—1 Pattern No. 539 screen, 2 protectors, Pattern No. 541, 1 protector (sea), Pattern No. 542, 2 water bottles, Pattern No. 540, and 1 packet of muslin caps, Pattern No. 551.

In ship's weather reports the temperature of the air signalled should be the temperature obtained from the portable screen unless the ship is supplied with an electric psychrometer in which case the readings of the latter instrument should be used instead.

The Electric Psychrometer.—This is a form of Assmann Psychrometer supplied to ships where accurate readings of air temperature and humidity are required. It can be used to determine wet- and dry-bulb temperatures accurately to 0·1° F. under any conditions.

The principal difficulty experienced in measuring air temperature and humidity by means of wet- and dry-bulb thermometers housed in a screen is that of securing adequate ventilation. The rate of evaporation from the wet bulb depends upon the rate at which the air flows past it for speeds of less than about 10 knots, but is independent of speed when this exceeds 10 knots. It is often impossible to secure that the speed of the air past the thermometers in a screen shall be sufficient to overcome this difficulty.

In the electric psychrometer (Fig. 8) the dry- and wet-bulb thermometers are mounted side by side in a polished metal frame. Air is drawn past the thermometer bulbs by an electric fan housed in the dome *B*. The bulbs of the thermometers are doubly shielded from radiation by the polished metal tubes *E*. The air passes through the tubes *E*, into the main air duct *D* and out by the outlets *C*. A wind shield is provided for the purpose of preventing the aspiration being impaired by the wind blowing in through the air outlets. The muslin is in the form of a sleeve which fits over the bulb of the wet-bulb thermometer. The thermometers may be withdrawn from the case by unscrewing the dome containing the electric motor fan.

For using the electric psychrometer four brackets should be provided from which the instrument can be suspended, with a power point near each. The positions of these brackets should be :—

(i) One right forward in such a position that the psychrometer hangs a few inches forward of the stem of the ship and at least five feet above deck level.

(ii) One right aft in a similar position with the instrument hanging a few inches over the stern.

(iii) One on each side of the fore bridge. These latter two should be as far over the sides of the bridge as possible, consistent with the instrument being easily read.

Normally the forward point will be used, and occasionally with a following wind, the stern point. The bridge points are, to a certain extent, emergency points for use when heavy seas render the use of the forward point impracticable. The guiding principle in

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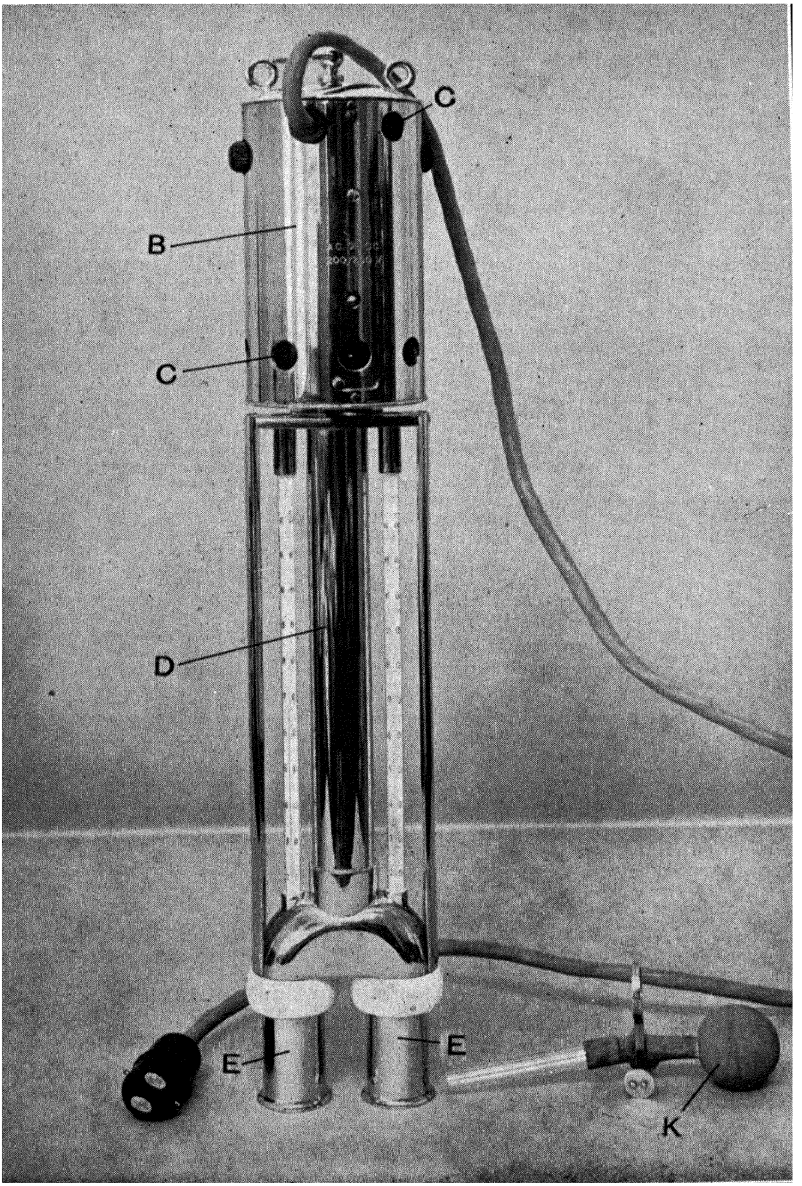


FIG. 8.—ELECTRIC PSYCHROMETER.

deciding on which point to use must be to ensure that the air passing over the psychrometer has not previously been in contact with the ship. Thus, if the forward point is out of action and the wind is from the starboard side, then the point at the starboard end of the bridge should be used.

In a wind which is nearly on the beam it is preferable to use either the forward or stern point, according to whether the wind is slightly ahead or astern.

In certain ships (*e.g.*, aircraft carriers) it may not be possible to provide brackets at all the points indicated above ; in such cases the most practicable points should be selected to satisfy the condition that the air flowing past the thermometer bulbs shall not have been previously in contact with the ship.

It is advisable to splice a lanyard to the psychrometer frame, so that the latter can be attached to the observer whilst shipping and unshipping, to guard against possible loss overboard.

Before taking an observation, make sure the mercury columns in the thermometers are unbroken and then suspend the psychrometer in the selected position. Preferably, but not necessarily, the instrument should be in the shade. If the psychrometer cannot be suspended in the shade, turn it so that the thermometer stems are shielded from the sun's rays by the frame.

Moisten the wet bulb. To do this fill the injector *K* (Fig. 8), consisting of a glass tube with rubber bulb and clip attached, with fresh distilled water. Holding the injector vertically with the bulb lowermost open the clip and slightly compress the rubber bulb until the water rises to a point just below the mouth of the tube ; then close the clip. Next introduce the glass tube into the inner radiation shield surrounding the wet bulb so that the muslin cover of the latter is completely immersed in the water. Hold the tube in this position until the muslin has become saturated and then open the clip before withdrawing the tube. This permits the surplus water round the muslin to flow back into the rubber bulb without settling between the thermometer bulb and the inner shield and thus obstructing the flow of air. Then withdraw the glass tube.

Plug in the lead to the psychrometer, switch on the current and stand well away from the instrument.

After an interval of two minutes read the two thermometers. It is of the utmost importance that the readings should be taken not earlier than two minutes and not later than six minutes after starting the motor. If, due to any cause, the readings are not taken within six minutes the wet bulb should be re-wetted and the instrument left for another two minutes before the readings are taken.

The thermometers should be read to the nearest 0·1° F., using for this purpose the reading eyepiece supplied with the instrument. The two forked ends of this should be placed on the thermometer which is being read and the eyepiece rotated until the slot in the top of

the eyepiece is perpendicular to the length of the thermometer. This slot should be directly over the top of the mercury column in the thermometer in order to avoid errors of parallax.

When reading the thermometer, the observer should be sure that :—

(*a*) He is not shielding the instrument from the wind.

(*b*) The instrument is not drawing in air which has been in contact with himself.

The wet-bulb reading should never be higher than the dry-bulb reading. Readings should always be examined immediately afterwards to see that this is not so. If the wet-bulb reading is higher than the dry, something is wrong and the cause should be investigated. Probably it will be due to a broken mercury column in the wet-bulb thermometer.

At each observation two sets of readings should be taken. These should agree within 0·2° F. If they do not, more readings should be taken until two consecutive sets of readings agree.

At sea the muslin on the wet bulb must be renewed at least every six weeks, and more often if it appears to be dirty. The wet-bulb thermometer must be taken out of the psychrometer at least once a week and the bulb with the muslin on it held under running water which is sufficiently pure for drinking purposes. Normally, this will keep the muslin sufficiently free from salt. It should also be washed after use in rough weather.

The polished tubes and frame should be kept bright by rubbing with the clean chamois leather provided. It is necessary to oil the bearings occasionally with a little good clock-oil. The parts requiring attention will be easily seen if the dome cover be removed. The instrument should be put away immediately after use.

Two spare Assmann thermometers (Pattern No. 2706), a spare rubber and glass filler (Pattern No. 2708) and a spare reading eyepiece (Pattern No. 2709) are provided.

Special psychrometric tables* are supplied for the determination of relative humidity and dew-point from the readings of the electric psychrometer.

The Whirling Psychrometer.—The whirling psychrometer (Fig. 9) may be used for the measurement of wet- and dry-bulb temperatures in positions in which it is difficult to set up the portable screen or the electric psychrometer. It consists of two thermometers mounted in a boxwood frame and provided with a handle and spindle by means of which the frame and thermometers may be whirled. The bulbs of the thermometers are therefore subject to a continuous current of air whose temperature is rapidly taken up by the dry-bulb thermometer. The effects of radiation may be regarded as

* Psychrometric Tables for obtaining the Vapour Pressure, Relative Humidity and Temperature of the Dew-Point (U.S.W.B., No. 235).

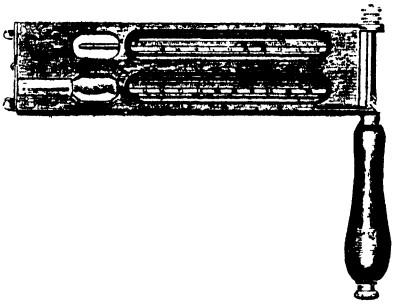


FIG. 9.—WHIRLING PSYCHROMETER.

negligible during the whirling. The frame carries a small vessel of distilled water into which dips a muslin sleeve which is drawn over the bulb of the wet-bulb thermometer. The instrument should be whirled for two minutes after which both thermometers should be read as quickly as possible, care being taken to avoid direct sunshine falling on to them. The psychrometric tables used with the whirling psychrometer are the same as with the electric psychrometer.

The Distant Reading Thermograph (Twin Pen) (*Pattern No.* 545).—The distant reading thermograph shown in Fig. 10 is an instrument designed to furnish in the meteorological office a continuous record of wet- and dry-bulb temperatures of the external air.

The sensitive element consists of two steel bulbs about six inches in length and one inch in diameter (one to be used as a dry bulb, the other as a wet bulb) housed in a small louvered screen painted white. The screen (Pattern No. 2700) is fixed in such a position in the ship, preferably on the foremast or fore superstructure, that there is as little chance as possible of the temperature of the air passing the bulbs being affected by the metal of the ship or by funnel smoke. The wet bulb is covered with a wick which dips into a trough containing distilled water. The wick consists of a sheet of material, one end being fitted with a sleeve which draws over the wet bulb. The other end of the wick is fitted with a smaller sleeve into which a rod is inserted, the rod acting as a weight to prevent the wick being blown out of the trough.

The type of wick shown in Fig. 10 is now obsolete.

Each bulb is connected with the recorder by a length of steel capillary tubing enclosed in lead of about $\frac{1}{4}$ inch diameter. At the recorder end the capillary is connected to a hollow flat spiral tube (Bourdon tube). The mercury in the bulb expands with rise of temperature and flowing along the capillary and into the Bourdon tube causes the latter to uncoil. One end of the Bourdon tube is fixed, the other actuates a pen-arm carrying a pen which traces a continuous record on a chart wrapped on a revolving drum, driven

by clockwork. The wet- and dry-bulb pens record on the same chart, but in order to avoid interference between them when the wet- and dry-bulb temperatures are the same, the wet-bulb pen is set to read permanently 10° F. too low. The range of the recorder is from 20° F. to 120° F. The clock is an eight-day one ; the drum makes one revolution in seven days, and the chart should be changed weekly, on Monday mornings.

It may happen that sections of the capillary tube passing through the ship are subject to artificial heating, which, causing expansion of the mercury in those sections will affect the recorder. To overcome this effect compensators are placed in the sections of the

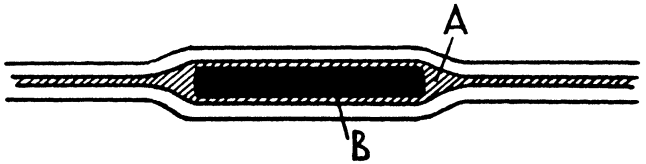


FIG. 11.—COMPENSATOR FOR D.R. THERMOGRAPH.

capillary tubing likely to be affected. A compensator takes the form of a widened portion of tubing *A* (Fig. 11) (external diameter about 1 inch) in which is placed a small rod of invar steel *B*. The volume of invar steel is practically unaffected by changes of temperature. As the temperature of the section rises the volume of the mercury in it and in the compensator increases. The volume of the tubing at *A* also increases, but the volume of *B* does not change. The difference between the volumes of *A* and *B* therefore increases and the increase in volume of the mercury is accommodated in the compensator, leaving the Bourdon tube unaffected.

Two spare pen-arms are supplied with each instrument. One, Pattern No. 2703, is for the dry bulb and the other, Pattern No. 2703A, is for the wet bulb ; the latter is the longer of the two. Two spare pens, Pattern No. 2704, are also supplied.

Installation of the Distant Reading Thermograph.—The recorder should be mounted with the pen-arms athwartships, in the ship's meteorological office, in a position as free as possible from vibration, moisture and dust. A cushion for the recorder should be made out of sorbo rubber or pieces of rubber tubing. The movement of the recorder on the cushion will lead to slight flexing of the capillaries and an adequate length of tubing must be left clear of the bulkhead to permit of such flexing.

The positions of the recorder and bulbs having been fixed the course to be taken by the capillary tubing should be decided upon. Any intervening decks or bulkheads should be drilled to permit the passage of the bulbs and compensators which are nearly 1 inch in

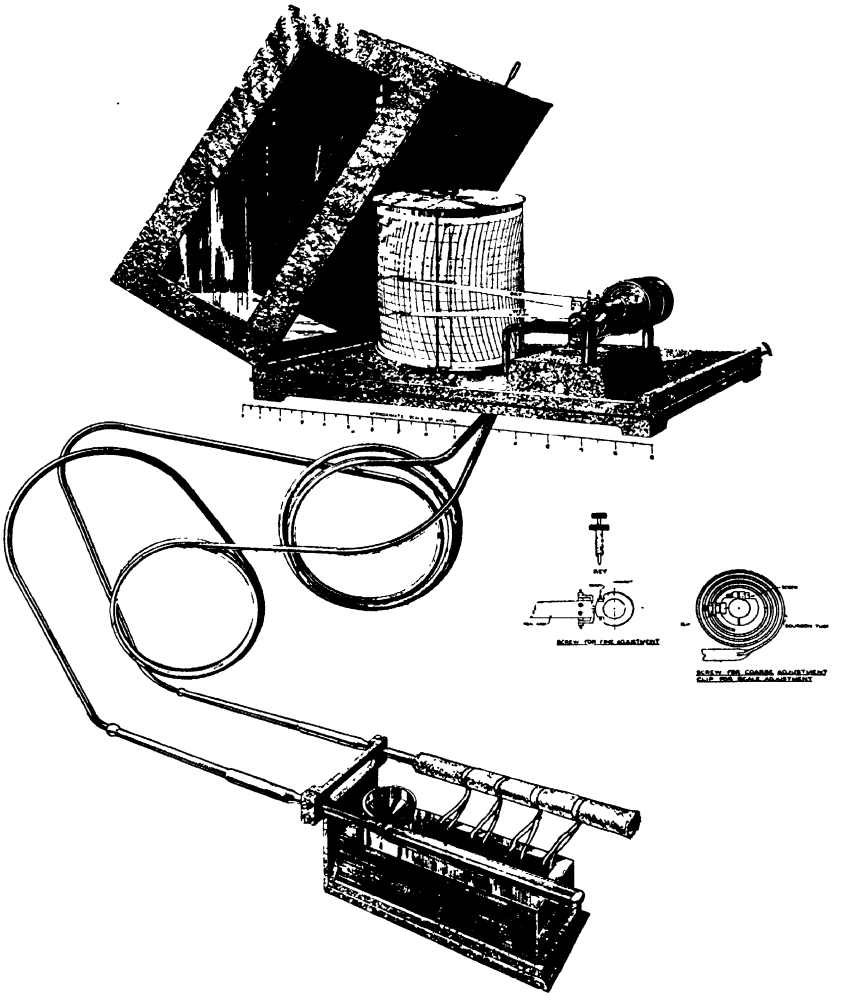


FIG. 10.—DISTANT READING THERMOGRAPH (TWIN PEN).

Care of the Distant Reading Thermograph.—The notes on the care and upkeep of recording instruments (page 35) apply to the recorder of the distant reading thermograph except where otherwise stated.

The bulbs are made of steel, nickel plated and then copper plated to protect them from corrosion. It is advisable to examine the bulbs carefully every two months to see if the copper plating has cracked in any way, as this would lead to corrosion of the bulb. The bulbs must not be scraped. The steel capillary is covered with a tube of copper brazed at both ends. In order to eliminate the possibility of corrosion lead tubing is drawn over the copper throughout its length. The capillary tubing should be examined periodically for defects in the covering. On no account must the bulb be subjected to a temperature in excess of the maximum shown on the chart of the recorder, and the tube should never be soldered. Any damage to the bulbs or capillary tubing which permits an escape of mercury necessitates return of the instrument to the makers.

The muslin and wick on the wet bulb should be changed weekly, or more often if there is reason to believe that salt spray has blown over it. Distilled water should be used for moistening the wet bulb.

Sea Temperature.

The Sea Thermometer.—An ordinary thermometer Pattern No. 504, as used in the portable screen, Pattern No. 539, but fitted with a sea protector (Fig. 5), is used for taking the temperature of the sea surface. The sea protector, Pattern No. 542, is similar to the air protector Pattern No. 541 except that the metal guard round the bulb is in the form of a cup which forms a reservoir for retaining a small quantity of sea water around the bulb while the temperature is being read. The thermometer should not be removed from the protector once it has been fitted.

To make an observation of the sea-surface temperature, a sample of water should be drawn in a canvas bucket, diameter not less than 5 inches, from over the ship's side forward of all outlets. Care should be taken that the sample of water is from the surface; the bucket should not be allowed to sink and so collect water which may not be at the surface temperature (*see* page 124). The thermometer bulb and protector should be held for a minute or so in the first sample drawn, then removed, the water contained in the metal guard being poured back into the sample, and the sample thrown over the side. The object of this procedure is to bring the thermometer and the bucket approximately to the temperature of the sea surface. A second sample of sea water should then be taken in the same manner as the first; the thermometer bulb and metal cup should be dipped in this sample and the temperature read as

soon as it becomes steady. It is essential that the bulb of the instrument should be under water at the time of reading; hence the reservoir around the bulb. If it is not covered, evaporation will take place from the drops of water adhering to it and it will tend to act as a wet-bulb thermometer and give a reading below the actual temperature of the sea water.

If a motor boat is used for work in shallow water the thermometer and protector might be lowered from the bows directly into the water (just low enough to fill the cup of the protector) care being taken to shield the thermometer from direct sunshine. The temperature of the second sample should be taken. If this direct method is used a note should be made against the observation in the meteorological log.

The canvas bucket should be stowed in the shade when not in use.

The Distant Reading Sea Thermograph.—A Distant Reading Sea Thermograph of similar construction to the Distant Reading Thermograph is supplied to some ships. The recorder is placed in the chart house and the bulb is fitted in the condenser intake or in a special recess in the ship's side constructed to permit a flow of sea water past the bulb. The temperature recorded is the temperature of the sea at a depth of several feet, which may differ by a few degrees from the actual sea-surface temperature, especially in calm weather. The thermograph indicates very conveniently, however, a change from a warmer to a colder current of water or *vice versa*, which is of importance, for instance, in forecasting visibility at sea.

Wind Speed and Direction.

The instruments supplied to H.M. Ships for recording the speed and direction of the wind, record the relative wind only. To obtain the true wind it is necessary to combine the relative wind with the course and speed of ship. This can be done by means of specially constructed tables or by a triangle of velocities.

The Munro Wind-Speed Recorder (*Pattern No.* 2714 and *Pattern No.* 549).—The instrument consists of a Dines pressure head and vane connected by two tubes, one transmitting pressure, the other suction to the recording apparatus. The head and vane are constructed on the principle of the Pitot tube and should be fitted in a position where the wind will be as free as possible from eddies caused by the superstructure of the ship or by the funnel gases. The foretruck is the most suitable position, but when this is not available the head and vane should be fitted to a short mast or spur in the best position available.

The head and vane are shown in Fig. 12. The upper part is free to rotate and the vane V thus keeps the horizontal tube A facing the direction from which the wind is coming. The wind blowing into this open tube A produces an increase of pressure

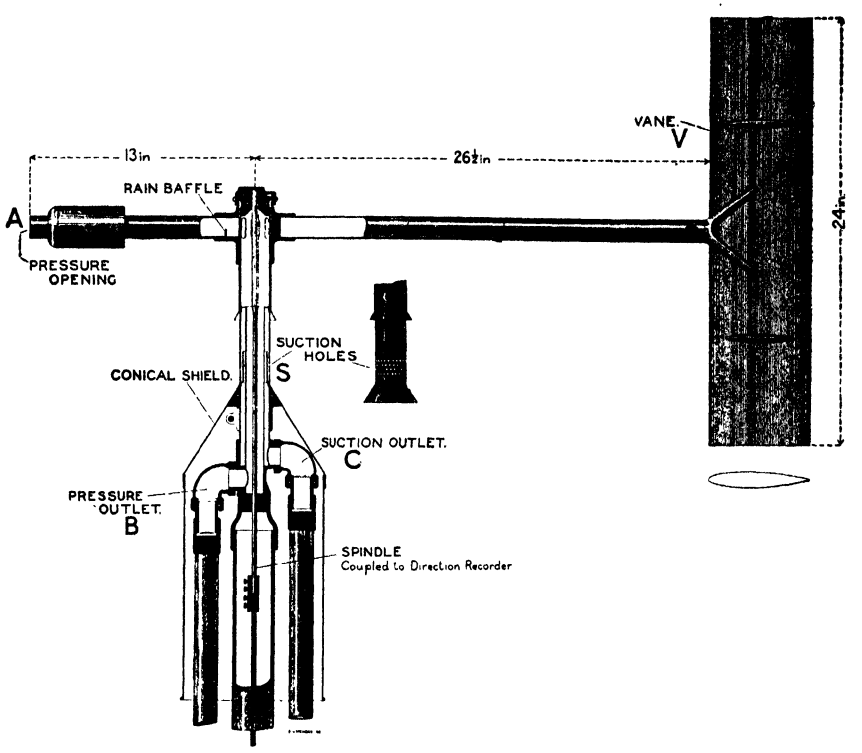


FIG. 12.—HEAD OF DINES ANEMOMETER (Mark II).

inside it, which is transmitted through the central tube of the head to the recording apparatus by means of tubing attached to the tube *B*. The fixed part of the central tube is surrounded by an outer tube *S* in which there are four rows of holes. Wind blowing past these produces a diminution of pressure in the annular space between the two tubes and this is transmitted to the recorder by means of a pipe attached at *C*. The pipes may be of considerable length, say 40 feet or 50 feet, so that the recorder can be placed a long way from the head. The pressure and suction outlets at the base of the head are covered by a cylindrical shield with a conical top, the purpose of which is to eliminate variations in the suction effect due to the disturbance in the wind stream which would otherwise arise from the presence of the projecting outlets.

The recorder shown in Fig. 13 consists of a static box containing a diaphragm which operates the recording pen by means of a spindle and lever.

The static box is connected to the suction pipe and the diaphragm to the pressure pipe and so variations of wind speed cause the spindle to move in or out of the static box, the spindle moving out when wind speed increases.

To face page 31.

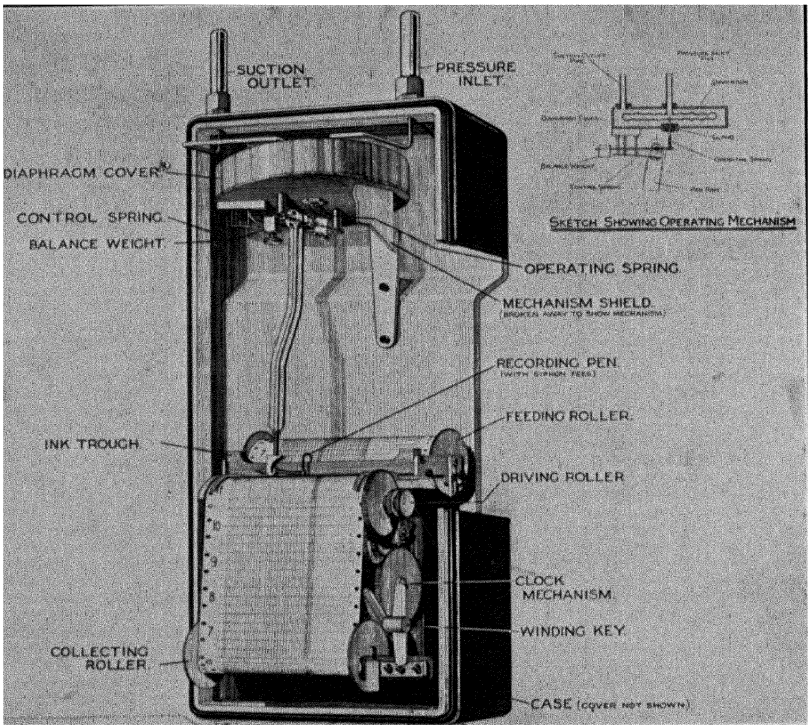


FIG. 13.—MUNRO WIND-SPEED RECORDER.

A leaf spring, at the upper end of the pen-arm is set in conjunction with a number of stop screws of varying lengths, as shown in the inset diagram, Fig. 13. As the wind speed increases the spring makes contact with the successive screws ; its effective length is thereby shortened and its stiffness increased, thus controlling the upper readings of the instrument. In the absence of the control spring the displacement of the pen would be proportional to the square of the wind speed ; with the spring in operation an approximately uniform scale is secured.

Below the static box is fitted a clockwork drum which unwinds a chart wrapped around a spool placed in the instrument. Between the drum and the spool is fitted a pen which siphons special ink from a trough and traces a record of the speed of the wind on the chart. A left-handed siphon pen (Pattern No. 555) (*i.e.*, one in which, looking along the pen towards the writing point, the siphon tubing bends to the left), is used with the Pattern No. 549 recorder and a right-handed pen (Pattern No. 2716) with the Pattern No. 2714 recorder.

The clock used is an eight-day one and gives a time scale of one inch of chart per hour. The chart is supplied in spools, each of which carries a length sufficient for 30 days. The charts are graduated from 0 to 80 knots in the case of the Pattern No. 2714 recorder and from 0 to 70 m.p.h. in the case of the Pattern No. 549 recorder. The Pattern No. 549 recorders are being replaced by the Pattern No. 2714 as opportunity permits.

Installation of the Munro Wind-Speed Recorder.—The base of the head is fitted with a $1\frac{1}{4}$ inch gas thread so that it can be screwed into a suitable socket.

The pressure and suction tubes should be constructed of $\frac{1}{2}$ inch copper or composition piping, and should be secured by metal clips. These tubes should be joined to the unions in the head and recorder by means of soldered joints. The unions are stamped with the letters " P " and " S " (pressure and suction) to ensure that the tubes are not cross connected.

The pressure and suction tubes should both be led from the head to the recorder through the water traps supplied with the instrument. Great care must be taken to ensure that the leads from the head to the water traps and from the recorder to the water traps are continuously downward ; if this is not possible additional drains should be fitted close to the recorder.

The case of the recorder should be secured to a fore and aft bulkhead in the chart house or the meteorological office so that the clockwork drum is horizontal.

If the head and vane are fitted on a separate mast extending well above the superstructure, suitable arrangements must be made to protect the instrument from damage by lightning.

The following particulars of dimensions and weights will be found useful in arranging for the installation of a wind recorder :—

Head and vane—

(*a*) Weight with shield .. .. 23 lbs.

(*b*) Overall dimensions—

Height, top of fin to bottom of shield .. .. 39 inches.

Length from pressure orifice to tail end of fin .. .. .. 46 ,,

Diameter of shield .. .. 8 ,,

Recorder—

(*a*) Weight .. .. 44 lbs.

(*b*) Overall dimensions—

Height .. .. 21 inches.

Width .. .. 8 ,,

Depth .. .. 8 ,,

Care and Adjustment of the Munro Wind-Speed Recorder.—In order to start the siphon when it is first fitted, it should be filled with methylated spirit, forced through it with the aid of the special filler (Pattern No. 555A) provided. After the pen is replaced in the stirrup at the end of the pen arm, the methylated spirit will siphon away and the ink will flow after it from the trough already filled with ink.

The pen should be flushed out with methylated spirit if it is to be kept out of the ink for more than a few minutes or it will become clogged. If it does become clogged, use the spare pen provided and place the defective one in a bath of methylated spirit until the latter can be forced through it by means of the special filler.

If the zero of the instrument needs adjustment it should be made by means of a sliding weight on a wire passing through the spindle carrying the pen-arm. Before this adjustment is made, the pressure and suction taps should be turned off and the instrument tapped until the pen has settled down. This may take some time. The pen-arm should be held firmly and carefully in the right hand while the milled screw is released and the weight adjusted along the wire. The weight should be shifted to the left if the pen is to the left of the zero and *vice versa*. After this adjustment the instrument should again be tapped and if necessary further adjustments made until the pen remains on the zero line when the instrument is tapped. The pen must never be pulled across the chart, against the diaphragm.

The following precautions should be taken in the upkeep of the instrument :—

The head should not be painted or the small suction holes in the vertical part of the vane may become clogged. On no account should one blow down the pressure or suction tubes or the diaphragm will be ruined.

The cocks in the drain traps should be opened for a few minutes once a month to allow any water that has accumulated to drain away; also make sure that the cocks have not become choked with paint.

The ball bearing inside the Dines head is lubricated before it leaves the makers and does not require further lubrication over a long period.

The pressure and suction tubes should not be dented in any way.

The lock nuts holding the unions to the recorder case must not be moved as this would interfere with the anchorage of the diaphragm and upset the calibration.

Oil must not be used on any part of the recorder.

Air Speed Indicator (*Portable*) **Mark I** (*Pattern* 556).—This instrument is supplied to ships for measuring the wind from any position in which a wind speed indicator or recorder is not permanently fitted, such as on the flight deck of a carrier or at a catapult. It is shown in Fig. 14.

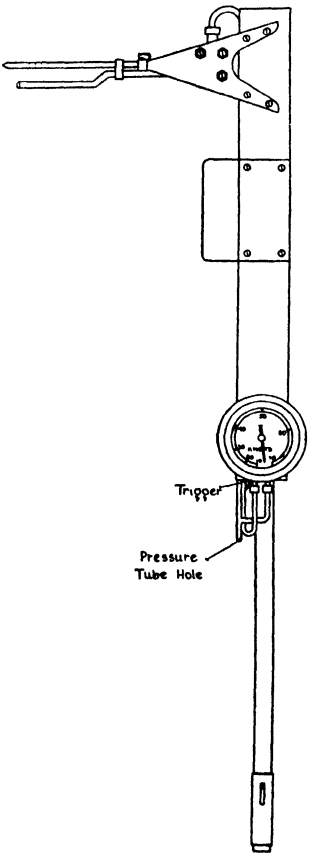


FIG. 14.—PORTABLE WIND-SPEED INDICATOR.

It consists of a pitot tube head of folding type and a dial graduated from 10—70 knots, both mounted on a staff 5½ feet long with a spring buffer fitted in the base to lessen any jarring due to careless handling of the instrument.

The pressure tube is fitted with a hole close to the gland through which it enters the dial. This hole must be closed with a finger while a reading is being taken.

At the back of the staff is fitted a trigger by means of which the observer can keep the diaphragm and pointer from sticking without removing a hand from the staff to tap the dial.

The pitot head of the instrument must be kept pointing *into the relative wind* while a reading is being taken.

The instrument should be tested from time to time with the special calibrator supplied to ships.

Observation of True Wind.—The direction and strength of the wind to be entered in the meteorological log and signalled in ships' weather reports may be obtained from the instrumental records, allowance being made for the course and speed of the ship. The direction and speed records take the form of a rather broad ribbon with wider excursions on either side of the mean representing the gusts and lulls. An imaginary line drawn along the mid points of the speed ribbon gives, with respect to the scale on the chart, the mean speed of the wind relative to the ship at any time. The direction of the relative wind is obtained in a similar way from the direction record. The true wind, at the level of the pressure head, may be found as follows in Fig. 15.

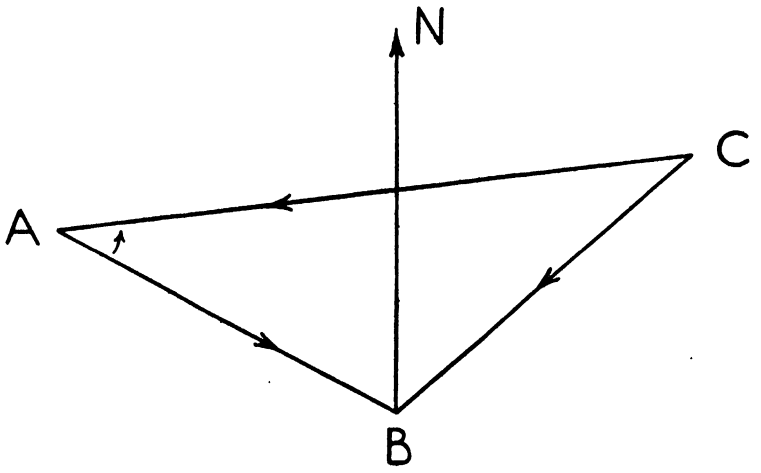


FIG. 15.—VECTOR TRIANGLE RELATING TRUE AND RELATIVE WIND.

On a suitable scale lay off a line AB representing the course and speed of the ship. From A lay off AC, making the angle BAC equal to the angle between the ship's head and the relative wind;

that is the angle read off from the direction recorder. If the wind is on the port side the angle BAC should be measured counter-clockwise, and *vice versa*. The length of AC must be equal, on the same scale as AB, to the speed of the relative wind. The line CB then represents the true wind, on the same scale as AB. The direction in degrees or compass points is given by the bearing of C from B with respect to true North.

For signalling in ships' weather reports the speed of the true wind should be converted to force on the Beaufort Scale, according to Table IV page 40.

GENERAL INSTRUCTIONS REGARDING AUTOGRAPHIC INSTRUMENTS.

The following instructions refer to barographs and thermographs generally and to wind-recording instruments where applicable.

Dating of Charts.—The date (time, month and year) of commencement and end of the record should be entered on each chart, either before it is fixed on the instrument or immediately after it is taken off. Should a record be missed in consequence of the pen not marking, or other causes, the chart should be filed with the successful records and not destroyed or used again.

Fixing the Charts on the Drum.—The chart is placed round the drum containing the clock where it is held in position by a spring. When fixing on the drum, care must be taken that the horizontal lines printed on the chart are parallel to the flange at the base of the drum. If the chart is carefully cut so that its lower edge is parallel to the horizontal lines, this will be the case when the edge of the chart is in contact with the flange all round the drum. If the chart is not accurately cut, allowance should be made for the fact by seeing that the horizontal lines are continuous where the two ends of the chart overlap while one end of the lower edge of the chart touches the flange.

Time Scale.—It is important that the time when the pen is at any point on the record should be accurately determinable. Three main causes of error exist :—

(*a*) Backlash between the clock and the spindle on which it turns. This may be reduced to a minimum if the cylinder be turned on its spindle so as to bring the pen **back** to the required time from a point in advance of its proper setting.

(*b*) Error of the clock. It may gain or lose. The adjustment should be made by means of the clock regulator, An error in the ruling of the chart may, however. produce the same result.

(*c*) Errors due to the chart becoming too damp.

In order that any errors of these kinds may be detected, it is essential that accurate time marks should be made on the records themselves. In the case of charts changed weekly, a time mark should be made every day. Although it is preferable to have the mark made at about the same time of day on each occasion, this is not essential. The essential thing is to know the exact time at which the mark was made. This object is secured if the observer enters in his meteorological log, or in a notebook kept beside the instrument, a note of the actual times at which the marks are made each day, and copies these times on the record when it is taken off at the end of the week. He should also write on the record the exact time at which the pen is lifted off the paper before removing the record. If there is backlash in the gearing the clock drum will not begin to rotate until the backlash has been taken up. The time mark should not therefore be made within half an hour of the commencement of the record.

The method of marking the time marks in the case of the barograph is described on page 18. In the case of the thermograph, the time mark can be made by moving the pen through a small vertical arc by gently pressing the Bourdon tube. For the Munro wind-speed recorder the stop cocks in the pressure and suction pipes should be opened until the instrument indicates zero. The recession of the trace to zero constitutes the time mark.

If the instrument is as free from friction as it ought to be no discontinuity should appear in the trace at a time mark. The absence of discontinuity in the trace furnishes good evidence that the instrument is working well.

Friction.—Friction between the working parts of the apparatus must be avoided as far as possible. The bearings should be cleaned occasionally and oiled with a good clock-oil, care being taken to remove excess of oil. The most serious friction generally occurs between the pen and the paper on which it writes. The pressure of the pen on the paper should be reduced to the minimum consistent with a continuous trace for which simple contact with the paper will suffice. Instruments are so arranged that the pen, or the style which carries it, is suspended like a gate, and arrangements are made for the slope of the gate bearings to be adjustable. In this way it is possible to regulate the pressure of the pen on the chart. In certain instruments this adjustment of the pressure of the pen on the paper is performed by means of a screw at the spindle end of the pen arm.

General.—The pen should be well washed from time to time in methylated spirit. A thin clear trace should be aimed at, but the point of the pen must not be so fine as to scratch the paper. A new pen may frequently be improved by drawing the point once or twice along an oil-stone but any trace of oil should afterwards be carefully removed.

Excess of ink should be avoided. Special care must be taken not to let ink come in contact with the metal arm which carries the pen as this will cause the metal to become brittle and break, or may cause the pen to adhere so firmly to the arm that it cannot be removed for cleaning. Should the arm become accidently inked, it should be immediately washed and slightly oiled.

The observer should satisfy himself that his instrument is working properly by frequent comparison of its indications with the readings of standard or control instruments. In the ideal case the difference between the readings of the standard and the value deduced from the curve is zero. In practice small discrepancies always occur; they must be reduced to a minimum by careful attention to the instrument

THE SUPPLY AND RETURN OF INSTRUMENTS USED IN H.M. SHIPS.

Manufacture.—Meteorological Instruments used in H.M. Ships are made by various commercial firms according to specifications issued by the Meteorological Office.

Before issue the instruments are tested either by the National Physical Laboratory at Teddington or by the Instrument Division of the Meteorological Office, Air Ministry.

Supply.—The instruments are supplied by the Meteorologica Office on repayment to Director of Stores, Admiralty, for issue to ships on demand through N.S.O.

Installation.—If the fitting of an instrument in a ship entails dockyard assistance, application should be made for the installation to be carried out as an Alteration and Addition during the ship's next refitting period.

Upkeep.—If any defects are developed which cannot be remedied by the ship's staff, application should be made to the Hydrographer, Admiralty, London, who will arrange for an Officer from the Naval Meteorological Branch to visit the ship to advise on the necessary action.

Return.—If it is found impossible to repair an instrument on board it should be returned to the N.S.O. and a new one demanded in its place.

Ships paying off into dockyard hands should return all meteorological instruments to the N.S.O. for return to the Meteorological Office. Certain of these instruments will be taken back by the Meteorological Office at a valuation, whilst others will remain Admiralty property but will be overhauled and stored until again required for Naval service.

Charts.—Blank charts for instruments should be demanded from N.S.O. at the ship's storing port, who obtains them from the makers of the instrument in the case of charts for distant reading thermographs and wind-speed recorders, and from the Meteorological Office in the case of other instruments.

Ink.—Ink for self-recording instruments should be demanded from N.S.O. An emergency supply may be made as follows :—

Ingredients—

Aniline colour	1 part.
Glycerine	10 parts.
Sugar	1 part.
Water	25 parts.

To mix.—Warm the water and dissolve the sugar in it. Add the glycerine in small quantities, stirring the mixture thoroughly until such added portion is dissolved. Beat up the aniline colour in a small quantity of the solution and stir the resulting mixture into the remainder of the solution. Allow the mixture to cool and the ink is ready for use.

CHAPTER II.

Non-instrumental Observations and Coding of Messages

WIND.

In ships not equipped with a wind recorder the direction and force of the wind must be estimated.

The direction is specified in degrees from true north or as that point of the compass from which the wind blows. Difficulty may arise owing to the fact that the wind is never steady even for a short period. Fluctuations through four points may occur in a few seconds. The mean direction should be estimated as nearly as possible. When the wind changes in the same direction as the hands of a watch, that is N.—E.—S.—W. it is said to veer; when it changes in the reverse direction it is said to back. These terms are to be used both in the Northern and Southern Hemispheres.

The Beaufort Scale.—Wind force is conveniently expressed by means of the Beaufort scale, a numerical scale devised in the year 1808 by Admiral Sir Francis Beaufort. It was based originally on the amount of canvas that a man-of-war of the period could carry, with different winds. The numbers on the scale ranged from 0 representing calm to 12 representing a hurricane "such that no canvas could withstand." With the disappearance of sailing ships the scale as originally devised became unsuitable, but it was still in use until the end of the last century when the need for its revision became acute. In 1905 the scale was revised by Dr. G. C. Simpson, having regard to the effects of the wind on fishing smacks for use at coastal stations and on objects such as trees for use at land stations, but no scale based upon effects in the open sea appears to have been drawn up. The use of the Beaufort scale had also extended to the Continent, and during the early part of the present century it became apparent that the specification of the scale was not sufficiently definite, in that differences arose between the estimates of wind made in various countries. Accordingly the International Meteorological Committee asked Dr. Simpson to draw up a revised scale based on the actual speed of the wind as measured by an anemometer. The true speeds were compared with the most reliable estimates of wind according to the Beaufort scale. Each Beaufort number was then assigned a certain range of wind speeds based on these comparisons. The revised scale was approved by the International Meteorological Committee in 1926 and has been in general use since then. The scale with the corresponding wind speeds, together with the descriptions of wind are given in Table IV.

TABLE IV.

Beaufort Number. International.	Meteorological Wind Scale: Limits of Speeds. Determined at Coast Stations for a height of 33 feet above sea level. Knots.	Meteorological Wind Scale: Limits of Speeds. Determined at Coast Stations for a height of 33 feet above sea level. Feet per sec.	Description of Wind.	Sea Criterion, 1939 (Provisional).
0	Less than 1	Less than 2	Calm	Sea like a mirror.
1	1—3	2—5	Light air ..	Ripples with the appearance of scales are formed, but without foam crests.
2	4—6	6—11	Light breeze	Small wavelets, still short, but more pronounced, crests have a glassy appearance and do not break.
3	7—10	12—18	Gentle breeze	Large wavelets. Crests begin to break. Foam of glassy appearance. Perhaps scattered white horses.
4	11—15	19—27	Moderate breeze	Small waves, becoming longer ; fairly frequent white horses.
5	16—20	28—36	Fresh breeze	Moderate waves, taking a more pronounced long form ; many white horses are formed (chance of some spray).
6	21—26	37—46	Strong breeze	Large waves begin to form ; the white foam crests are more extensive everywhere. (Probably some spray.)
7	27—33	47—56	Moderate gale	Sea heaps up and white foam from breaking waves begins to be blown in streaks along the direction of the wind. (Spindrift begins to be seen.)
8	34—40	57—68	*Fresh gale ..	Moderately high waves of greater length ; edges of crests break into spindrift. The foam is blown in well-marked streaks along the direction of the wind.
9	41—47	69—80	Strong gale ..	High waves. Dense streaks of foam along the direction of the wind. Sea begins to roll. Spray may affect visibility.
10	48—55	81—93	Whole gale ..	Very high waves with long overhanging crests. The resulting foam in great patches is blown in dense white streaks along the direction of the wind. On the whole, the surface of the sea takes a white appearance. The rolling of the sea becomes heavy and shocklike. Visibility is affected.
11	56—65	94—110	Storm.. ..	Exceptionally high waves. (Small and medium-sized ships might be for a long time lost to view behind the waves.) The sea is completely covered with long white patches of foam lying along the direction of the wind. Everywhere the edges of the wave crests are blown into froth. Visibility affected.
12	Above 65	Above 110	Hurricane ..	The air is filled with foam and spray. Sea completely white, with driving spray ; visibility very seriously affected.

* The Gale Warning Signal in the British Isles is hoisted for winds which may reach force 8 or above.

To make the specification of wind speed definite it is necessary to state at what height above the sea (or ground) the wind is measured. In the lowest layers of the atmosphere the wind speed varies rapidly with height increasing upwards from a value which, very close to the earth's surface, is practically zero. A height of 33 feet (10 metres) in an open situation has been adopted internationally as standard. It is then to be understood that the wind speed corresponding to any Beaufort number according to Table IV is the wind speed at 33 feet above the surface in an open position.

With no anemometer available the estimation of the wind is very difficult and the ability to form a reliable judgment of the wind is acquired only after long experience. It is possible to obtain the true wind by means of a vector diagram (*see* page 34) using for the relative wind that estimated by the observer as if the ship were stationary; alternatively Form H.162 may be used. The Sea Criterion will be found to be a useful and reliable guide in estimating the force of the wind, and the general run of the waves indicates its direction. To acquire experience the observer should frequently compare his estimates of true wind with the anemometer records, due allowance being made for the movement of the ship. It is often misleading to base the estimate of wind on an observation of funnel smoke.

Extreme Winds and Gales.—The observer should enter in the " Remarks " column of the meteorological log the direction and force of the strongest wind experienced since the last observation if the force reaches or exceeds force 6, and the time of occurrence of the extreme should be noted. If the wind attains the force of a gale (force 8) a note should be made of the time of commencement and end of the gale, as well as of the direction and force and time of recurrence of the strongest wind.

In ships where no wind recorder is in operation, the information regarding the occurrence of gales and strong winds cannot be expected to show the same precision as the observations made at fixed hours. The duration of gales may be subject to considerable uncertainty, partly because many gales rise or subside gradually so that it is difficult to specify a definite time for the commencement or end of wind force 8, and partly because observers cannot keep an uninterrupted look-out. It should, however, be borne in mind that such notes as " during early part of night," " during afternoon " are better than blanks in the log, when the observer feels unable to put down a definite hour for the occurrence. The notes as to gales and strong winds should be added to the signalled weather reports.

Weather.

Under the term " weather " are included observations of the state of the atmosphere usually referring to phenomena which cannot be expressed numerically, *e.g.*, whether rain, snow, etc., is actually

falling, the presence of obscuring matter in the atmosphere giving rise to fog, mist, haze, etc., the appearance of optical phenomena such as rainbows and halos. For the most part these phenomena indicate modifications in the condition of the water vapour in the atmosphere ; they are therefore known in some countries under the generic term " Hydrometeors."

(i) **Beaufort Letters.**—The state of the weather is recorded by means of the Beaufort notation (page 320). The phenomena are with few exceptions indicated by their initial letter and the notation therefore affords a simple and concise means of representing by a group of letters either the actual state of the weather at the time of observation (present weather) or a general summary of the conditions during the interval since the last observation was made (past weather).

The modification of the Beaufort notation for denoting intensity and continuity of phenomena and the various combinations of Beaufort letters are explained on page 321.

Caution.—Such errors as the use of *d.* for " dew," *f.* for " fine " or *sh.* for " showers" should be carefully avoided. *d.* would be read as " drizzle," *f.* as " fog " and *sh.* as " snow and hail."

(ii) **Appearance of the Sky.**—The letters *b.*, *c.*, *o.*, *g.*, *u.* and the combination *bc.* are used to describe the general appearance of the sky.

b., *bc.*, *c.* and *o.* refer to the proportion of the sky covered with cloud. *o.* is used whenever the sky is completely covered with a uniform layer of thick or heavy cloud (*St.* or *Nbst.*, *see* Chapter X). *c.* denotes cloudy sky and should be used when the sky is more than three-quarters covered with cloud and when *o.* is inappropriate. *c.* may therefore be used when the sky is completely covered provided that the cloud is not stratus or nimbostratus. *b.*, denoting blue sky is used when less than one quarter of the sky is covered with cloud. For cloud amounts between one quarter and three quarters the letters *bc.* are used in combination. Except for *o.* the use of these letters is not in any way dependent upon the form of cloud present. A sky more than three-quarters covered with a thin, almost imperceptible sheet of cirrostratus would be described by the letter *c.* even though the intensity of sunlight or moonlight were hardly affected by it.

When the total cloud amount is recorded in tenths of the sky covered the Beaufort letters should be used as follows :—

b.	0	—	2/10 total cloud amount.
bc.	3/10	—	7/10 ,, ,, ,,
c.	8/10	—	10/10 ,, ,, ,,

except *o.* 10/10 Stratus or Nimbostratus.

The letter *g.* (gloom) should be used when, owing to dense cloud or smoke accumulation **above** the surface, daylight is very much reduced without the visibility at the surface being much impaired. These conditions are described as high fog. *u.* (ugly threatening sky) is used generally when the appearance of the sky threatens an imminent thunderstorm.

(iii) **Precipitation.**—*Continuous precipitation* (Beaufort letters *rr.*, *ss.*, etc.).—In reports of past weather a break of only ten minutes' duration in a period of continuous rainfall can be disregarded, but a break of half an hour must be taken into account. Thus a period of moderate continuous rain with a break of ten minutes in it would be described as *rr.*, but a period of moderate continuous rain with a break of half-an-hour in it would be described as *orr.*, *o.*, *orr.*, if *o.* were the appropriate letter for the interval. The duration of precipitation necessary to justify the use of the word " continuous " in reports of past weather cannot be rigidly fixed. If it rained without a break for two hours it should undoubtedly be described as " continuous rain." If it rained for only half-an-hour it would not be called continuous rain. If it rained for an hour in the middle of a period otherwise without rain, it would not be necessary to describe it as continuous, *e.g.*, the description in Beaufort letters, *o.*, *or.*, *o.*, would be appropriate. If, however, the hour's rain came at the beginning of a period and was the continuation of continuous rain in the previous period it would still be reported as continuous rain in the new report, *e.g.*, the weather would be described as *orr.*, *o.* The same would apply to rain which began an hour or more before the report and of which there was no evidence of cessation at the time the report was made.

Occasional precipitation—showers.—There is a difference in the type of weather indicated by *ir.* (intermittent rain) and *p.* (passing showers). In general, showers are of short duration and the intervals between the showers are usually characterized by definite clearances of the sky. The clouds which give the showers are isolated clouds, and are often found with a W. or NW. wind

(W. or SW. in the southern hemisphere), a rising barometer and improving weather, behind a depression which has passed eastwards. The precipitation does not usually last more than fifteen minutes, although it may sometimes last for half an hour or more.

Occasional precipitation on the other hand, usually lasts for a longer time than showers and the weather between the periods of precipitation is usually cloudy or overcast. It should, however, be realised that weather between showers may also be cloudy or overcast and the weather between periods of occasional precipitation may be clear. The experienced observer will usually decide from the type of cloud which description to use.

Drizzle—slight rain.—Drizzle is not " rain in small amount " but " precipitation in which the drops are very small." Slight rain, on the other hand, is precipitation in which the drops are of appreciable size (they may even be large drops) but are relatively few in number. Observers should decide from the size of the drops whether the precipitation is drizzle or rain ; and from the combined effect of the number and size of the drops whether the precipitation is slight, moderate or heavy.

There are some occasions when rain is observed to be falling through drizzle ; both the drizzle and the rain should be noted in the meteorological log. In the case of signalled reports, if both cannot be reported, the drizzle should be reported as a rule because the rain on such occasions is generally slight and of short duration.

Hail.—A distinction is drawn between true hail and soft hail. In the former the stones are hard, and occasionally of considerable size ; in the latter they are small and soft, resembling little snow pellets. Both forms are denoted by the Beaufort letter *h*.

Ice Crystals in the Air.—On some occasions, usually in cold clear weather, small ice crystals are observed in the air down to ground level, and they may under suitable conditions give rise to halo phenomena. The symbol ←→ should be entered in the meteorological log on such occasions. The phenomenon is distinct from an ordinary fall of snow and may, indeed, occur when the sky is apparently free from cloud. The symbol ←→ should not be used merely to indicate that a regular crystalline structure is observed in flakes forming part of an ordinary snowfall, the proper symbol for which is *.

(iv) **Fog, Mist, Haze.**—The terms fog, mist and haze are used to indicate a diminution in the transparency of the atmosphere due to the suspension in the lower atmosphere of small particles. The particles may be either solid or liquid. The terms fog and mist are generally regarded as applicable to occasions when the obscurity is due mainly to water particles, the word haze being applicable when dust particles are the chief cause of lack of transparency. The observer has no definite criterion for determining whether he is observing dust or water particles, but the humidity of the lower atmosphere will afford some guide. If the relative humidity is high it may be assumed that water droplets predominate, so that the terms mist or fog are appropriate; on the other hand when the relative humidity is low the term haze is appropriate. The distinction between fog and mist adopted by the Meteorological Office is one of degree, the term fog being reserved for occasions of more marked obscurity. It has been found necessary to use observations of the actual visibility for classifying occasions of mist, fog or haze (page 218). Fog is reported only when the visibility is less than $\frac{1}{2}$ mile (1 kilometre) and mist or haze when the visibility is greater than this but less than 1 mile (2 kilometres). This does not prevent the use of such expressions as " slight haze " or " slight mist " when the visibility is 1 mile or more, but the unqualified words " mist " or " haze " should not be used in such circumstances.

Abnormal visibility.—On some occasions the atmosphere has an abnormal clearness and transparency; distant objects stand out in full relief from the background with great hardness and distinctness (resembling the " objects " in a stereoscopic view). Outline and details which appear softened or even indistinct under ordinary conditions, are sharply defined on these occasions. At sea the stars may be seen to rise and set at night with the naked eye and the topmast of a steamer hull down may be seen on the horizon by day. The term " visibility " is used to indicate this abnormal clearness and is denoted by the Beaufort letter *v*. It should be noted that " visibility " (*v*.) defined in this way is unrelated to the ordinary scale of visibility (page 47). The Beaufort letter *v*. if used, should come in " weather " and should not be used as a contraction for " visibility 9 " on the visibility scale.

Wet fog.—A fog in which water is deposited copiously on exposed surfaces should be noted by means of the letters *fe*.

(iv) **Fog, Mist, Haze**—*continued.*

State of the sky, and fog.—When the sky is discernible through the fog the cloud is recorded and reported as if no fog were present. If only a small portion of the sky near the zenith is visible and the portion is clear of cloud the state of the sky is reported as *b.*; the letters *bf.* should be used. If cloud can be seen through the fog the amount should be estimated as well as possible and the letters *b.*, *bc.*, *c.* or *o.* used as appropriate. If the sun, moon or stars can be seen through the fog, and there is no evidence of cloud above the fog, the amount of cloud should be entered as 0, irrespective of the horizontal thickness of the fog. If the vertical thickness of the fog is so great that it is impossible to tell whether there is cloud above it, the cloud amount should be entered as 10, and the state of the sky described by the Beaufort letter *o.* This applies, for example, to all cases of fog in the daytime when the sun is quite invisible.

(v) **Thunder**: **Lightning**: **Thunderstorm.**—The term thunderstorm (Beaufort letters *tl.*) should be used to indicate a thunderstorm in which both thunder and lightning are observed. On occasions when thunder alone is heard *t.* should be used. The letter *l.* when used alone stands for sheet lightning or lightning without thunder. The times at which thunderstorms occur should be noted in the remarks column of the meteorological log. It is desirable to note the time of commencement of a thunderstorm, as given by the first thunder and the duration of the thunderstorm to the nearest minute. The direction from which the thunderclouds approach should also be noted.

(vi) **Miscellaneous phenomena.**—The occurrence of the various optical phenomena described in Chapter XIII, and miscellaneous phenomena such as dew, hoar frost, rime as well as gales and squalls should be noted in the remarks column of the meteorological log. The time of commencement and cessation of gales and squalls, and the general weather conditions (wind, pressure, temperature, humidity and cloud) accompanying halos and other optical phenomena should be carefully recorded.

In using the Beaufort notation to describe present and past weather the letters denoting the various phenomena should be preceded by a letter denoting the state of the sky at the time. Thus :—

bcpr. = sky partly cloudy with shower of rain.
bF. = clear sky with thick fog.
c/tlr. = cloudy after thunderstorm with rain in the last hour.

If there are no phenomena to record the weather should be denoted by *b.*, *bc.*, *c.* or *o.* as appropriate.

A summary of past weather consists of a sequence of such complete groups of Beaufort letters, separated by commas. Thus :—

bc., c., rr. denotes partly cloudy conditions followed by an increase of cloud with continuous rain later.

These forms should be used for entering present and past weather in the meteorological log.

Visibility.

The terms " fog," " mist " and " haze " used for describing the degree of atmospheric obscurity under the heading " weather " are not sufficiently precise to meet modern requirements. A scale of visibility has accordingly been devised, based on the maximum distance at which an object is visible with different degrees of atmospheric obscurity. In practice a number of objects are selected as nearly as possible at standard fixed distances from the observing station, the distances increasing roughly in such a way that each is nearly double the next smaller distance. The determination of the most distant object of the series which is visible on any given occasion constitutes the observation of visibility.

Fog and Visibility Scale (*Specification for use at Sea*).—

0 Dense fog	..	Objects not visible at 50 yards.				
1 Thick fog	..	Objects visible at 50 yards	but not at	1 cable.		
2 Fog		,, ,,	1 cable	,,	,,	2 cables.
3 Moderate fog	..	,, ,,	2 cables	,,	,,	½ mile.
4 Mist or haze or very poor visibility		,, ,,	½ mile	,,	,,	1 mile.
5 Poor visibility	..	,, ,,	1 mile	,,	,,	2 miles.
6 Moderate visibility		,, ,,	2 miles	,,	,,	5 miles.
7 Good visibility	..	,, ,,	5 miles	,,	,,	10 miles.
8 Very good visibility		,, ,,	10 miles	,,	,,	30 miles.
9 Excellent visibility		Objects visible at more than 30 miles.				

It is necessary to have a criterion as to what is meant by an object being visible. It is often possible to see that there is " something " without being able to see what it is, unless one knows beforehand what it is ; in such a case the object is not visible according to the Meteorological Office convention. An object is therefore to be regarded as " visible " if it can be distinguished ; if the object is a ship and it can be distinguished as such, it is to be regarded as visible.

The measurement of visibility at a shore station by this method s a comparatively simple matter. At sea it is often difficult or

impossible owing to lack of objects and of knowledge of their distances when objects are available. When coasting and when fixes can be obtained, the distance of points when first sighted or last seen may be measured from the chart.

In a long ship the lowest numbers of the scale 0 and 1 can be determined by direct measurement. 2 and above indicate conditions of obscurity such that visibility extends beyond the length of the ship. If there is any obscurity or abnormal refraction the visible horizon may be very misleading as a means of judging distance, particularly when the height of the eye is great, as in the case of an observer on the bridge of a battleship. Cases have occurred where such attempted measurements are believed to have contributed to stranding.

In the case of observations made at night the visibility scale will be used to denote as nearly as possible the same degree of atmospheric obscurity as in daylight observations. This part of the subject presents difficulties and a good deal must be left to the observer, as the aids available for indicating atmospheric obscurity at night vary so much from one place to another. Stationary lights at known distances will in general provide the basis of estimation, but it is usually difficult to get a selection of fixed lights at the appropriate distances. Care must also be taken due to the difficulty raised by the fact that a very bright light such as that of a powerful lighthouse may cause an appearance of brightness in its direction, or show light on the sky, in a state of atmospheric obscurity which would not permit the observer to see the actual source of light in the same sense that he can see one of his daylight objects. The observer should therefore use his knowledge in such a way as to bring observations made with lights at night into line with the daylight scale, to the best of his ability. Apart from the use of lights a careful observer can derive a considerable amount of information as to night visibility from a general inspection. It is surprising how much can be seen even on a fairly dark night, *e.g.*, a distant range of hills or an elevated coast-line can often be made out against the sky line in circumstances which indicate that in daylight an object at that distance would be " visible."

The entry to be made for visibility in the meteorological log is the figure on the fog and visibility scale corresponding to the observed (or estimated) distance of visibility, followed by the estimated visibility in miles in brackets.

Clouds.

Cloud observation may be considered under four headings :—

Form of cloud.
Amount of cloud.
Height of cloud.
Movement of cloud.

Form of cloud.—For guidance in identifying cloud forms the meteorological observer is referred to Chapter X and the International Atlas of Clouds, where descriptions and photographs of the various types are given.

Each form of cloud present should be entered in the meteorological log under the heading high, medium or low cloud, the appropriate code figures being entered.

Amount of Cloud.—The degree of cloudiness is to be given by the figures 0—10, in which 0 represents a sky quite free from cloud, and 10 an entirely overcast sky in which no patches of blue sky are visible. It is required to estimate the number of tenths of the area of the sky which would be covered by the clouds present, supposing them moved up to each other so as to form a continuous sheet, with the proviso that, as stated above, 0 is used only when there is no cloud and 10 only when there is no clear sky. The entry " trace " should be made when there are small amounts of cloud present and these small amounts do not cover as much as one twentieth of the sky; and the entry " 9 +" when there are small breaks in an otherwise overcast sky, though the breaks do not amount to one-twentieth of the sky's area. The numbers given are to refer solely to the amount of the sky covered and not to the density, height or other quality of the cloud.

In estimating cloud amount the observer should sub-divide the sky into quadrants by means of diameters at right angles to each other. An estimate (on the scale 0—10) is then formed for each quadrant separately, and the figure finally entered in the log is the mean of the four numbers so obtained. The directions of the dividing diameters should be selected to give convenient sub-divisions of the prevailing cloudy canopy.

The amount of low cloud should be separately estimated by imagining that every other visible form of cloud is replaced by blue sky, and by determining the amount of low cloud in the manner already described.

When fog is present the instructions on page 45 (state of the sky and fog) should be followed.

If the hand is held out at arms length and the wrist bent so that the palm of the hand faces the observer, the hand from the wrist to the tips of the fingers covers roughly one-hundredth of the sky. The hand must be open but the fingers should be kept together and not spread apart.

Height of cloud.—The only method at present available to the observer on board ship for the measurement of cloud height is by means of pilot-balloon ascents (Chapter III). The balloon is filled to rise at a known rate and it is necessary only to note the time that elapses between the release of the balloon and its disappearance in the lower surface of the cloud. Care must, however, be taken to

distinguish between cases when the balloon is seen to enter a cloud and cases in which the balloon is obscured by cloud at some lower level drifting across the field of view.

When in harbour, or near the coast, in hilly and mountainous districts the heights of the bases of clouds which are lower than the tops of visible hills can be determined with some accuracy by inspection, when the clouds are banked-up against the hills, provided the heights of the different features of the hillside are known.

In most cases, however, it will be necessary for the observer to obtain the cloud height by estimation. The ability to estimate cloud height satisfactorily can be acquired only after long experience. Observers should take every opportunity of checking their estimates by measurement with pilot balloons and by comparing notes with the pilots of aircraft. Before a pilot-balloon ascent is made, if cloud is present its height should be estimated and the estimate compared afterwards with the measured height

Meteorological Officers in ships carrying aircraft may compare the heights of cloud with those of aircraft in flight. They may, by arranging with the pilots to fly at certain heights, make themselves proficient in judging the height of various types of aircraft by their appearance and apparent size. Some guidance in forming such estimates can be obtained from the following table giving the heights between which the bases of the different forms of cloud are usually observed (in temperate latitudes).

Form of cloud.	Usual Height.	Remarks.
Stratus (St.)	500—2,000 ft.	Sometimes practically down to the surface; sometimes as high as 4,000 feet.
Stratocumulus (Stcu.)	1,500—4,500 ft.	Sometimes as low as 500 feet; or as high as 8,000 feet.
Cumulus (Cu.)	2,000—5,000 ft.	Sometimes as low as 1,000 feet or as high as 8,000 feet.
Cumulonimbus (Cunb.)	2,000—5,000 ft.	As Cumulus. The top of Cunb. frequently extends to over 10,000 feet and may reach 25,000 feet.
Nimbostratus (Nbst.)	500—2,000 ft.	Sometimes practically down to the surface; sometimes as high as 4,000 feet.
Altostratus (Ast.)	Usually between 6,500 and 20,000 ft.	
Altocumulus (Acu.)		
Cirrus (Ci.)	Usually between 20,000 and 40,000 ft.	In the tropics the upper level may be 50,000 feet or more.
Cirrocumulus (Cicu.)		
Cirrostratus (Cist.)		

It is not anticipated that observers who have not had experience with balloons will be able to make very close estimates and they should not aim at getting nearer than the nearest 1,000 feet except for clouds below 1,000 feet.

Movement of cloud.—Whenever possible the meteorological observer in a ship should record the movement of cloud. particularly of medium and high cloud. It will not generally be practicable to obtain any other than a very rough estimate of cloud movement. When the ship is at anchor the observer may place himself in a stationary position so that he is in line with the cloud and a fixed point such as the masthead. The movement of the cloud should be observed relative to the fixed point. The form of cloud and the direction from which it comes should be noted and its speed may be described as rapid, moderate, slow or stationary.

At land stations more precise observations are made with the aid of nephoscopes.

Sea and Swell.

The state of the sea should be reported according to the Sea Disturbance Scale (Code XIII), which is here reproduced with a table of height of waves corresponding to the code figures. A correlative wind and sea scale is shown here in order to assist in estimating the state of sea from the force of the wind in the open sea with a long fetch and in deep water (*see also* p. 458). It will be noted that for wind forces 3 to 9 (inclusive) the code figure for state of sea is one less than the corresponding Beaufort wind scale number.

Beaufort Wind Scale and Correlative Sea Disturbance Scale.

Beaufort Number.	Description of Sea.	Mean height of waves in open sea, in feet.	Code figure.
0	Calm—glassy	—	0
1	Calm—rippled ..	½	1
2, 3	Smooth—wavelets ..	1, 2½	2
4	Slight	5	3
5	Moderate	9	4
6	Rough	14	5
7	Very rough	19	6
8	High	25	7
9, 10	Very high	31, 37	8
11, 12	*Phenomenal	45 or over	9

* As might exist at the centre of a hurricane.

For the purpose of recording and reporting swell, the Swell Scale (Code XIV) is used. Careful distinction should be made between sea and swell, sea being the waves caused by the wind at

the place and time of observation, while swell is wave motion due to past wind or wind at a distance. The direction from which the swell comes should be noted to the nearest compass point.

Hours of Observation.

Greenwich Time should be used throughout the year and in all parts of the world in making and recording observations. The four principal hours of observation for ships are 0000, 0600, 1200 and 1800 G.M.T. Observations should be made, recorded and signalled at as many of these hours as possible. Readings at other hours may be called for on special occasions.

Punctuality in taking the observations and in despatching the messages is of the greatest importance. The observations at fixed hours should normally be completed at the exact hour. Sufficient time should therefore be allowed for the somewhat protracted procedure in observing sea temperature, and air temperature when the electric psychrometer is used. If the portable screen is used for air temperature it should be suspended in the appropriate position at least 15 minutes before the reading is to be taken. The observation of the element which is varying most rapidly should as far as possible be made last, *i.e.*, nearest in time to the exact hour.

The Coding of Weather Reports.

The code is the New International Code (1929) which is in conformity with the recommendations of the International Meteorological Conference which met at Copenhagen in September, 1929.

The messages are made up in groups of five figures and care should be taken in writing out the messages to see that the proper grouping of the figures is maintained.

In order to help the observer in understanding and remembering the code in which the messages are sent the forms of the messages are represented below by letter symbols, each letter or group being replaced in actual messages by the corresponding figure or figures. Whenever an observation cannot be made for any reason, so that the corresponding figure or figures are not available, an X will be written in the place of each missing figure in the actual message.

Ships' weather reports are coded in one of the following forms :—

F 232.—YQLLL IIIGG DDFww PPVTT $3C_LC_MC_HN$
$T_dKD_KWN_h$ d_sv_sapp

F 261.—YQLLL IIIGG DDFww PPVTT $6KD_KCN$
$T_dd_sAWC_H$

F 291.—YQLLL IIIGG DDFww PPVTT $9SKD_KW$
CNN_hAT_d

F 232 is intended for use in temperate and high latitudes where the weather changes are of an irregular nature and are associated with the passage of depressions and anticyclones. F 261 is intended

for use in lower latitudes, where the weather has often pronounced diurnal variations, and barometric changes are slight and regular except when a tropical revolving storm may be in the neighbourhood.

F 291 is in use on the China station and may be adopted on other stations, its advantage being that it includes the state of the sea.

The following is an index to the symbols used to denote the different elements. The references to code numbers refer to Table III in A.L.W.S., Vol. II, 1937, pages 45-53. For convenience of reference the codes are also set out on a " Code Card," Form H.353.

		F 232, F 261 and F 291.	Code No.
First Group	*Y*	Day of the week	—
	Q	Octant of the globe	XX
	LLL	Latitude	—
Second Group	*lll*	Longitude	—
	GG	Time of observation (G.M.T.)	—
Third Group	*DD*	Wind direction	—
	F	Wind force	—
	ww	Present weather	V
Fourth Group	*PP*	Corrected reading of the barometer	—
	V	Visibility	VIb
	TT	Temperature	—
		F 232.	
Fifth Group	3	Identification figure for F 232	—
	C_L	Form of low cloud	I
	C_M	,, ,, medium cloud	II
	C_H	,, ,, high cloud	III
	N	Total amount of cloud	IV
Sixth Group	T_d	Difference between sea and air temperature	XXI
	K	Swell	XIV
	D_K	Direction from which swell comes	—
	W	Past weather	VIII
	N_h	Amount of low cloud	IV
Seventh Group	d_s	Direction of ship's movement	—
	v_s	Speed of ship	XXII
	a	Characteristic of barometric tendency	X
	pp	Amount of barometric tendency	X
		F 261.	
Fifth Group	6	Identification figure for F 261	—
	K	Swell	XIV
	D_K	Direction from which swell comes	—
	C	Form of predominating cloud	XVI
	N	Total amount of cloud	IV
Sixth Group	T_d	Difference between air and sea temperature	XXI
	d_s	Direction of ship's movement	—
	A	Barometric tendency	XI
	W	Past weather	VIII
	C_H	Form of high cloud	III

F 291.

			Code No.
Fifth Group	9	Identification figure for F 291	—
	S	State of sea	XIII
	K	Swell	XIV
	D_K	Direction from which swell comes	—
	W	Past weather	VIII
Sixth Group	C	Form of predominating cloud	XVI
	N	Total amount of cloud	IV
	N_h	Amount of low cloud	IV
	A	Barometric tendency	XI
	T_d	Difference between air and sea temperature	XXI

The groups *YQLLL lllGG* indicate the position of the ship and the time of observation.

Y = day of the week.
1=Sunday, 2=Monday, 3=Tuesday, 4=Wednesday, 5=Thursday, 6=Friday, 7=Saturday.

Q = Octant of the globe in which the ship is (Code XX).

LLL = Latitude of ship in degrees and tenths. Whether the latitude is North or South is determined by the figure for *Q*.

lll = Longitude of ship in degrees and tenths. For longitudes greater than 100° the hundreds figure is omitted. Whether the longitude is East or West, and whether it is greater or less than 100° is determined by the code figure for *Q*.

GG = Time of observation to the nearest hour, G.M.T.

INDIVIDUAL ELEMENTS IN WEATHER REPORTS—F 232.

Pressure *PP*.—The corrected barometer reading is sent with the hundreds and the decimal figure omitted. The corrected pressure is rounded off to the nearest whole millibar and the tens and units figures coded for *PP*, *e.g.*, 1,015·3 mb. is rounded off to 1,015 and coded as 15. If the decimal figure of the corrected reading is ·5 the reading will be rounded off to the odd one of the two whole millibars between which it lies.

e.g.—982·5 is rounded off to 983 and coded as 83.
981·5 is rounded off to 981 and coded as 81.

This practice is called "throwing to the odd."

Barometric Characteristic *a* **and Tendency** *pp*.—The characteristic is obtained from the shape of the barograph trace during the three hours preceding the observation. The appropriate figure for *a* is obtained from Code X. It must be noted that if one of the figures 0—4 of Code X is used the pen must be at least as high on the chart as it was three hours previous to the time of observation. It may be either on the same level as it was three hours ago, or above that level. If, on the other hand, it is below the level of three hours previously then one of the figures 5—9 must be used. For example,

if the pen has been rising then falling but is at a higher level than, or the same level as, it was three hours previous to the time of observation the figure to be reported is 0. If, however, it has been rising then falling, but is lower than it was three hours before the time of observation the figure to be reported is 9.

Typical shapes of the barograph trace during the preceding three hours and the appropriate code figures for *a* are given in Fig. 16.

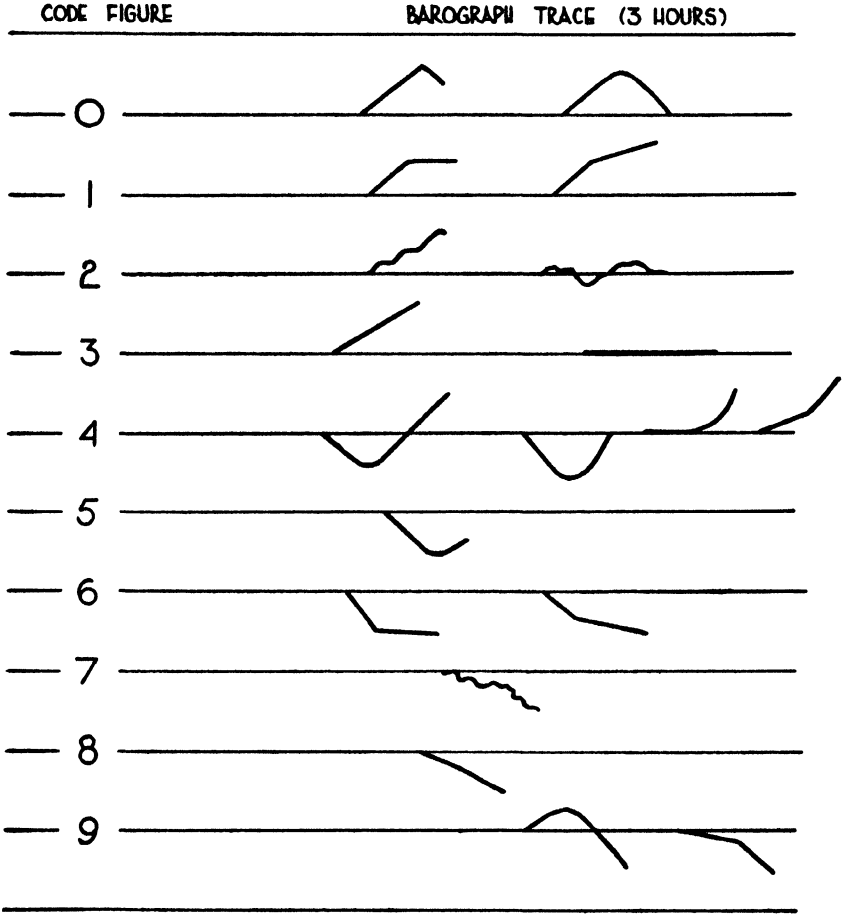


FIG. 16.—TYPICAL SHAPES OF BAROGRAPH TRACE AND CORRESPONDING CODE FIGURES FOR CHARACTERISTICS.

The amount of the tendency is read from the barograph chart by comparing the level of the pen at the time of observation with its level three hours earlier as indicated by the trace. The vertical interval between consecutive horizontal lines on the chart represents 1 millibar on the open scale barograph and the tendency should be

expressed in fifths of a division, *i.e.*, in fifths of a millibar. The observer must, therefore, estimate by eye the amount of change to fifths of a scale division.

Two figures (*pp*) will be used for reporting tendency. Tendencies from 0 to 9 inclusive will be preceded by a 0, *e.g.*, a tendency of 8 (fifths of a millibar) is reported as 08. Should the tendency exceed 99 the figures 99 will be reported for *pp* and a note will be added to the end of the message indicating the actual value; thus "Tendency 113."

Temperature *TT*.—The temperature is reported by two figures but the thermometers must be read to the nearest tenth of a degree for use in computing the humidity. The two figures to be reported for the temperature will be obtained by taking the reading of the dry bulb and rounding it off to the nearest whole degree Fahrenheit. If when reading the thermometer the mercury falls exactly half-way between two degree-marks the reading will be thrown to the odd (*see* above). Temperatures from 0 to 9 inclusive will be preceded by a 0. Temperatures below 0° F. will be reported by subtracting the numerical reading from 100. Temperatures greater than 100° F. are reported by omitting the hundreds figure, thus:—

27·3° F. is reported as 27.
4·7° F. is reported as 05.
−13·5° F. is reported as 87.
104·5° F. is reported as 05.

Wind Direction *DD* **and Force** *F*.—The direction of the wind is expressed in compass points and should be reported by two figures on the scale (01—32) in which 32 = North, 01 = North by East, 08 = East, 16 = South, etc. A calm is reported by 00.

The occurrence of unusual gustiness during the past hour should be reported by the addition of 33 to the number giving the wind direction on the scale (01—32); while the occurrence of a definite squall or line squall during the past hour should be reported by the addition of 67 to the number giving the wind direction on the scale 01—32.

The force of the wind is given by the Beaufort number. If the force exceed 9 at any hour of observation the number 9 is entered as the code figure and the words "Storm 10," "Storm 11," or "Storm 12" are added at the end of the message.

If the records of wind force are obtained from a wind recorder the conversion from wind speed (after allowing for the course and speed of the ship) to Beaufort numbers necessary for adapting the results to the reporting code should be made by the scale given in Table IV. When determining the direction or force of the wind from a wind recorder the mean direction or speed, as indicated by an imaginary line drawn through the middle of the trace, should be used, and not the extreme excursions of the pen on either side of the

mean. For observations at fixed hours the mean direction and speed for a period of 10 minutes should be taken from the records.

Sudden changes of wind.—A difficulty may arise in deciding upon the speed which should be reported if the hour of observation coincides with a sudden change of wind. The normal rule that the mean wind force over the past ten minutes is to be taken is not applicable when there has been a sudden change from one speed to another which differs materially from it. In this case the observation should be continued for a few minutes and if the speed remains at the new level this latter should be reported. If, however, the increase proved to be temporary only and due to a squall, the mean speed over a period of ten minutes centering at the time of observation would be taken. In all cases of this kind a note should be added at the end of the message and in the meteorological log to supplement the coded report. Thus in the first case given above the note might read " sudden increase wind just occurred 8 to 25 knots." In the second case the note would give the highest speed in the squall as " squall to 37 knots just passed." Where no wind recorder is available similar reports should be made as far as possible from estimates of the wind force, the values being given in Beaufort numbers.

Similarly, in the case of a sudden change in wind direction occurring at the time of observation, if the change proves to be other than temporary the new direction will be reported in the message and a note indicating the change added at the end of the message. Thus " Wind just veered, 18 to 24."

Clouds.—*Cloud Forms* C_L, C_M, C_H.—Instructions for determining the form of the cloud, with special reference to the use of Codes I, II, III are given in the International Observer's Cloud Atlas. The code figures to be used are those for the specifications which most nearly represent the state of the sky at the time of observation. If there is no cloud of the main group represented by C_L the appropriate code figure is 0 : similarly for the cases of no cloud of either of the main groups represented by C_M, C_H.

If owing to darkness, fog, dust-storm or other surface phenomenon it is impossible to determine the form of the cloud, the code figures for C_L, C_M and C_H should be replaced by X. This should also be adopted for medium and high cloud if there is a complete layer of low cloud which prevents any medium and high cloud being seen.

It should be noted that the expression " low cloud " in Code I refers to cloud of which the actual height above the surface of the sea is low. Cloud seen low down towards the horizon is not necessarily cloud of Code I. The matter should be decided by the form of the cloud.

Cloud Amount N_h, N.—For entry in the messages the cloud amounts will be coded in accordance with Code IV. It should be carefully observed that special code numbers are allotted to entries of 0, trace, 1, 9, 9 + and 10 while the remaining entries 2 to 8 are divided into three groups for purposes of coding. For example, if there is only a trace of low cloud visible the entry for N_h will be " trace " but the figure for the message will be 1 ; while if only a small patch of blue sky can be seen the entry for N will be 9+ but the figure for the message will be 7.

There is also a special code number 9, for the case when the sky is obscured by fog, dust-storm or other phenomenon, the entry for cloud amount being 10, both for N_h and N.

Present Weather *ww*.—A table showing the figures corresponding with the different weather conditions at the time of observation and the general character of the weather is given in Code V. As a general rule the largest number in the table which describes appropriately the present weather and the general character of the weather should be selected. The appropriate entry for *ww* in the meteorological log may be found from Table XIII which gives the Beaufort letter abbreviations corresponding to the weather types of Code V.

In selecting the appropriate number for *ww* the observer should take account not only of the weather during the ten minutes preceding the time of observation, but also, in certain circumstances which are clearly shown in Code V, of previous weather. The weather of more than an hour ago should not be reported under *ww* except in the cases of code numbers 15, 16.

Notes on the Code.—(*a*) *General.*—For ships' weather reports an abridged form of the code is used as shown on code card H353, the numbers being marked with an asterisk (*) in Code V. Experienced observers should use the full code.

The code numbers 00 to 19 should only be used when none of the later numbers is applicable. The code number 10 will be reported on occasions when precipitation is observed to be actually occurring in the neighbourhood of the ship though not actually at the ship itself. It should only be used when the precipitation can actually be seen falling at the time of observation by the observer. The entry in the meteorological log on such occasions will consist of Beaufort letters, indicating the state of the sky, followed by *jp* : thus, *bcjp*, or *cjp*. When possible a letter *r*, *s* or *h* will be added to show the form of the precipitation, thus *bcjps* will mean fair with adjacent snow falling. In the same manner *cjpr* or *ojpr* will be entered when the weather is cloudy or overcast with rain within sight.

(*b*) *Fog or Thick Dust-Haze.*—The numbers 40 and 43-49 must only be used when visibility is less than ½ mile at the time of observation. Number 41 should be used when visibility has been between 1 cable and ½ mile at some time in the past hour (but is

greater than ½ mile at the time of observation). Number 42 should be used when visibility has been less than 1 cable at some time in the past hour (but is greater than ½ mile at the time of observation). When code numbers 43, 45 and 47 are used (*i.e.*, sky discernible through fog) the cloud is reported as if no fog were present; but when code numbers 44, 46 and 48 are used (*i.e.*, sky not discernible) the cloud amount is reckoned as 10 and the height of the base of the low cloud as zero. When fog accompanies precipitation at the time of observations or within the preceding ten minutes one of the code numbers 57, 67 or 77 should be used.

(*c*) *Precipitation at the Time of Observation.*—If there is precipitation at the time of observation, or if there has been precipitation in the last ten minutes, one of the numbers 50-99 should be used. The numbers 80-89 will be used only when the precipitation is of the shower-type, and when precipitation is actually falling at the time of observation. The clouds which give showers are isolated clouds, and the showers are therefore always of short duration. Between the showers there is a definite clearance unless stratiform clouds are filling the interstices between the shower clouds.

(*d*) *Precipitation during last hour.*—If there has been occasional or continuous precipitation at the ship during the last hour, but none during the ten minutes preceding the time of observation and there is no fog, one of the numbers 20-29 should be used. Showers during the last hour which have ceased before the time of observation should be reported by one of the numbers 25-27. The figures 20-29 should not be used if there is precipitation at the time of observation.

(*e*) *The numbers* 20, 30, 40, 50, 60, 70, 80, 90.—These numbers are to be used in the abridged code, but not as a rule in the full code. A higher number in the appropriate decade can generally be selected to report the weather more definitely.

Past Weather *W.*—By past weather is meant the general character of the weather during the last six hours.

In selecting the figure for past weather (Code VIII) account must be taken of the number used for present weather (*ww*.). If the present weather indicates that precipitation has occurred (*ww* = 20-29) it may be necessary to emphasize some other feature by the past weather figure, but care should be taken on every occasion to indicate whether precipitation has occurred in the preceding interval. If this is not shown by the present weather figures it must be shown by the past weather figure. When fog has occurred this also should be indicated either in the present or past weather figures.

Visibility *V.*—The visibility is reported by a single figure according to Code VIb.

Difference between Air and Sea Temperature T_d.—The air and sea temperatures and their difference should be taken to 0·1° F. and the appropriate code figure selected from Code XXI.

Swell K **and Direction of Swell** D_K.—The type of swell should be reported by the appropriate figure selected from Code XIV. The direction **from** which the swell comes should be reported by a single figure on the scale (0-9) in which 1 = NE., 2 = E. and 3 = SE. . . . 8 = N, 0 = no swell, 9 = confused swell.

Course d_s **and Speed of Ship** v_s.—The ship's course should be reported by a single figure on the scale (0-8) as for direction of swell. If the ship is stationary 0 should be used. The speed of the ship is reported by the appropriate single figure, selected from Code XXII.

INDIVIDUAL ELEMENTS IN WEATHER REPORTS—F 261.

With the following exceptions the elements should be coded as in F 232.

Form of cloud C, C_H.—The predominating form of cloud present whether high, medium, or low cloud is reported by a single figure selected from Code XVI. If in addition there is high cloud it should be reported by the appropriate figure from Code III.

Course of ship d_s.—The course is reported as above. No provision is made for reporting the speed of the ship.

Barometric Tendency and Characteristic A.—The amount and characteristic of barometric tendency are reported by a single figure selected from Code XI. Code figures 1-4 are to be used when the barometer has risen by more than $\frac{3}{5}$ mb. ; 5-8 when it has fallen by more than $\frac{3}{5}$ mb. and 0 when it is steady or has changed by not more than $\frac{3}{5}$ mb.

INDIVIDUAL ELEMENTS IN WEATHER REPORTS—F 291.

The elements should be coded as in F 261. The state of sea which occurs only in F 291 should be coded according to the Sea Scale, page 51.

CHAPTER III.

The Measurement of Upper Winds over the Sea

The great increase in aerial activity consequent on the development and expansion of the Fleet Air Arm has made this branch of meteorology one of paramount importance. A knowledge of upper-air conditions for the operation of aircraft either with detached units or with the fleet is obviously of inestimable value in carrying out reconnaissance flights, bombing operations and navigational exercises generally. Furthermore the results obtained from observations of upper winds can be and are collated and utilized in the construction of charts of upper winds over the ocean. The value of such charts for strategical, operational and commercial use is self-evident.

THE MEASUREMENT OF UPPER WINDS BY MEANS OF PILOT BALLOONS.

Principles of Method.—The measurement of upper winds by means of pilot balloons depends upon the observations of a small balloon at successive intervals of time. The balloon, after being inflated with hydrogen until it can support a known weight, is set free at the earth's surface and rises at a rate that is nearly constant. At the same time the balloon is carried along horizontally by the wind and changes its position in accordance with the direction and speed of the wind in the layers through which it passes. The balloon is so light in comparison with its size that it may be considered to adjust itself almost immediately to the direction and speed of the air in which it is floating at any particular instant. If the position of the balloon can be determined relative to the earth's surface at successive intervals of time, it is then possible to calculate the horizontal speed with which it is moving and so obtain the wind speed and direction. The method has been in use on land for many years, a theodolite being employed for the necessary observations. At sea, owing to the motion of the ship a theodolite of normal pattern is unsuitable and greater difficulty is experienced in making the observations. A special type of theodolite mounted on gimbals may, however, be used and various other methods have been developed during the past few years. When the balloon is released, its elevation above the horizon, and azimuth, are usually read at intervals of one minute. In addition to knowing the elevation and azimuth of the balloon it is necessary before its movement over the surface of the earth can be traced, to determine its vertical height at the time of each

observation. The following methods of obtaining the height are in common use :—

(1) by assuming a constant rate of ascent ;

(2) by observation of the angle subtended at the point of observation by a tail of known length hanging below the balloon ; and

(3) by simultaneous observations from the ends of a base line.

In method (3) the base line must be at least ½ mile in length, so that the method is inapplicable at sea and will not be discussed here.

Method (2) which involves a tail hanging beneath the balloon, has not been found convenient for use afloat (brief particulars of the method are given below). Reliance has therefore been placed on method (1) owing to its simplicity ; in this method the balloon is filled to give a known lift and its rate of ascent is then assumed to be constant and of a known value. The height being known at the end of each minute, the horizontal distance of the balloon from the observer is found from the elevation reading and its position over the surface of the sea can be determined from the azimuth.

If the ship is stationary, the position of the balloon can in this way be plotted from minute to minute and the speed with which it has moved horizontally, *i.e.*, the speed of the air current in which it is travelling, can readily be calculated. If the ship is moving, allowance must be made in the calculations for course and speed.

Rate of Ascent.—The rate of ascent of a balloon is governed by the free lift which is the force driving the balloon upwards, and the resistance to its motion imposed by the air. With the method of filling now used, in which the balloon with a certain weight attached is inflated until it just floats, the free lift is equal to the weight removed from the balloon after inflation has been completed. Thus, if the balloon is filled with the standard filler (weighing 238 gm.), the free lift given to the balloon by removal of the filler is 238 gm. If after removal of the filler a 100 gm. weight is attached to the balloon, the net weight that has been removed will be (238—100) gm. and therefore the free lift will be 138 gm. If L be the free lift of a balloon, W the weight of the rubber balloon and of any equipment which may be attached to it, and v the rate of ascent, it can be shown that v is given by the following formula :—

$$v = \frac{q\, L^{\frac{1}{2}}}{(L + W)^{\frac{1}{2}}}$$

where q is a constant. The value of the constant q varies somewhat with different sizes of balloon, but for balloons measuring 150 inches in circumference when inflated, the size in most common use in the Navy, it is 310 when L and W are measured in grams and v is given in feet per minute. This has been found from a large number of ascents made at Shoeburyness and followed by two theodolites at the ends of a base line. For balloons less than 100 inches in

circumference when fully inflated (*see* page 64) q may be taken as 276.

It is not always possible to adjust the free lift of a balloon with great accuracy, particularly if the balloon has to be filled with hydrogen in a position exposed to the wind; the following table, which shows the variation in v for different values of L is, therefore, of interest:—

Free lift L =	200	220	238	260	280	gm.
Rate of ascent v =	671	686	700	716	730	ft./min.

In calculating this table, the balloon is assumed to weigh 80 gm., the mean weight of 150-inch balloons. The free lift normally used is 238 gm., which gives a rate of ascent of 700 ft./min. The table shows that a variation of 4 per cent. in the free lift produces a change of 1 per cent. in the rate of ascent.

It is found in practice that balloons, even if they are filled to the same free lift, do not always ascend at the same rate, the variations being due to the presence of upward and downward currents in the atmosphere. If the balloon is caught in one of these currents, it will rise at a greater or lesser rate than if it were rising in still air. Unless one of the methods for measuring the height of the balloon indicated above is adopted, it is not possible to say whether the balloon has been caught in one of these up or down currents and inaccuracies may accordingly be introduced into the results obtained. While it is very desirable that the height should be measured whenever possible, the inaccuracies referred to are not, as a rule, of sufficient magnitude to vitiate the results obtained. Upward and downward currents in the atmosphere are generally caused by the irregular heating of the earth's surface due to solar radiation, the air in contact with the heated areas rising owing to its increased buoyancy. Cumulus clouds are generally an indication of such rising currents, so that the results obtained when a known rate of ascent is assumed will be of less accuracy on days when cumulus clouds are present than when the clouds show stratification in the atmosphere. The assumption of a uniform rate of ascent will, further, be more accurate in night ascents, when there is no solar heating, than on sunny days, and will be more accurate at all times over the sea than over the land.

When determinations are made of the height of the balloon, it will certainly be found that in individual cases the rate of ascent differs markedly from that given by the formula. It may even be found that the mean rate of ascent obtained from a series of such ascents also differs somewhat from that given by the formula. It would be undesirable in such a case for the individual ship to adopt a revised mean rate of ascent, for use in future ascents in which the height is not directly observed. When a sufficient number of accurate observations have been obtained over the sea, it may be necessary to reconsider the value of q; but for the present, the value 310

(*i.e.*, a rate of ascent of 700 ft./min.) should be used for all daytime ascents with 150-inch balloons, in which the height is not directly observed. When the height of the balloon is observed, the average rate of ascent of the balloon should be determined from the observations and this rate of ascent should be used in the calculations of this particular ascent.

Balloons.

Sizes, Weights and Rates of Ascent.—Balloons are made in four sizes, the circumferences when inflated being 48, 70, 90 and 150 inches, and the average weights 9½, 20, 30 and 80 gms., respectively.

Since a variation of 12 per cent. in the weight of the balloon alters the rate of ascent by only 1 per cent., some variation in the weights of balloons can be allowed without detriment to the accuracy of the results. Thus, individual 150-inch balloons may weigh anything between 70 gm. and 90 gm. without any appreciable difference in their rates of ascent.

The 48-inch balloons are used mainly on land for determining the height of low cloud, the balloons being filled to ascend at a known rate and the time observed until they enter the cloud. The 70- and 90-inch sizes are commonly used on land for obtaining upper winds, but it has been found that at sea the larger size, 150-inch, is most suitable for general use. The reason for this is that it is much more difficult to follow a balloon at sea than on land and a balloon of the larger size can be kept in view for a longer period. Further, owing to the fact that it rises at a greater rate than a smaller balloon it ascends to a greater height in a given time, and so the horizontal distance travelled from the observer is proportionately less. There is, therefore, a double advantage in using the larger size of balloon. The average rates of ascent of the four sizes of balloon when inflated to their normal diameters are as follows:—

48-inch	70-inch	90-inch	150-inch
400 ft./min.	500 ft./min.	500 ft./min.	700 ft./min.

At the present time balloons of two sizes (150-inch and 90-inch) are supplied to H.M. Ships. The 150-inch balloon, with a rate of ascent of 700 ft./min. in day ascents and the same or a slightly reduced rate in night ascents (*see* page 72) is generally used; the 90-inch is intended for use in those day ascents for which a smaller rate of ascent is required. In the day time the *higher rate of ascent* (*i.e.*, the 150-inch balloon) will be required, for example, on occasions:—

(*a*) of normal conditions in clear weather;
(*b*) of strong winds, in order to gain the maximum height possible before the balloon is lost to view owing to its great (horizontal) distance away;
(*c*) when the ship is steaming up wind at high speed or into fresh winds.

The *lower rate of ascent* (*i.e.*, 90-inch balloon) on the other hand might be used :—

(*d*) in misty weather, with light or average winds (a balloon can usually be kept in sight to a greater height in misty weather if moving away slowly than if moving away rapidly) ;

(*e*) when the sky is covered with detached clouds (the slower rate of ascent increases the chances of keeping the balloon in sight in the cloud gaps and this may enable the wind above the clouds to be determined) ;

(*f*) when the upper-wind conditions are suspected to be variable (the variations can be more easily observed with the slower rate of ascent).

Colours.—Balloons are made in three colours, white or untinted, dark blue and dark red. Experience has shown that in bright sunny weather, white balloons are found to be best, as the sun's rays are reflected by the balloon and the spot of light thus created is visible long after the outline of the balloon has lost its definition. Blue balloons are more opaque than red and appear black when viewed from a distance. They therefore show up well against white skies, while red balloons are usually more suitable on days of broken cloud when the background varies between blue sky and cloud. Blue and red balloons can therefore be used successfully on dull cloudy days but the choice of colour is largely a matter of individual preference.

Care and handling of balloons.—Balloons used for upper wind measurements are usually made of rubber. They are liable to deteriorate rapidly especially if exposed to light, heat or air. They should, therefore, be kept covered in a cool place remote from any artificial source of heat and in tropical regions should be placed in cool storage when possible. After removing them from cool storage the balloons should be heated gradually to soften the rubber. It is in general a good plan if a balloon is found to have hardened, to warm and soften it by rubbing it between the hands or placing it in warm water, care being taken in the latter case that water does not enter into the balloon. If these precautions are taken the risk of bursting during inflation will be considerably reduced and furthermore balloons will not usually assume an irregular shape. Any balloons that do assume an irregular shape should be discarded. Balloons are sometimes found on inflation to have developed pinholes. Such balloons should be placed on one side and reserved for measuring cloud heights when the cloud is estimated to be below 1,000 feet. The hole may be patched by sticking a piece of gummed paper over it (when the balloon is fully inflated) and it is advisable to patch similarly any thin portions of the balloon which may be observed as these are liable to develop into pinholes. A

careful examination of the balloon before release will often ensure that time is not wasted in following a balloon that bursts before sufficient observations have been obtained.

Balloons should not be kept for more than twelve months as they are liable to deteriorate, especially in hot weather.

Hydrogen and hydrogen cylinders.

Hydrogen for filling balloons is supplied in cylinders of 100 of 200 cubic feet capacity. The cylinders are usually filled to a pressure of 120 atmospheres, but this pressure is reduced for use in H.M. Ships to 100 atmospheres so that a 100 cubic feet cylinder will contain only 83 cubic feet of hydrogen. The cylinders supplied to ships are covered with coir matting, marked " Admiralty " and a steel cap is fitted to protect the valve. This cap should always be in position when the cylinder is not in use and in warm and damp weather should be inspected periodically for rust and the threads greased. If the cap has been rusted in the thread a few light blows with a piece of wood should suffice to loosen it. Cylinders should be stowed on the upper deck or above, in a cool dry place, in racks clear of the heat from funnels and other sources of artificial heat and protected from the rays of the sun. **No smoking** should be allowed in their vicinity at any time. When hydrogen is being discharged from the cylinders it should be done slowly so as to avoid ignition through the friction caused by dust, etc. in the nozzle. Generally, the instructions contained in the Engineering Manual regarding gas flasks, are to be observed, so far as they are applicable.

The dimensions of the cylinders are :—

100 cubic feet.—4 ft. 6 ins. by 8¾ ins. Weight 134 lbs.
200 cubic feet.—6 ft. 6 ins. by 8 ins. Weight 256 lbs.

Filling and Release of Balloons.

Pilot-Balloon Shelters.—Owing to the difficulty of filling balloons in a strong wind it is desirable that a permanent roofed-in pilot-balloon shelter should be provided in ships carrying aircraft, in order that the wind finding from the ship may be carried out as rapidly as possible. This difficulty is particularly marked on occasions when a candle lantern is used to illuminate the balloon during a night ascent (*see* page 72), owing to the liability of the candle to be blown out. Without a suitable shelter it is usually impossible to adjust the free lift accurately and there is also a danger of both balloon and filler accidentally being allowed to escape. If a shelter be provided, it is possible to fill a balloon in the evening and keep it in the shelter until it is required for the ascent made just before dawn. (When this procedure is adopted, the balloon should be slightly overfilled and the free lift finally adjusted just before release). The shelter should be large enough to hold a hydrogen cylinder and an inflated balloon with the man

who would normally be engaged in filling it. As it is awkward to carry inflated balloons from one place to another, the shelter in which the balloon is filled should be as close as possible to the position from which the balloon would normally be released. The choice of the latter is governed by the conditions that the balloon when released should rise clear of the superstructure, aerials, rigging etc. (especially in night-ascents).

Apparatus for Filling.—The apparatus required for filling the balloon consists of the hydrogen cylinder with key, a left-hand fine-adjustment valve, a length of rubber tubing and the balloon

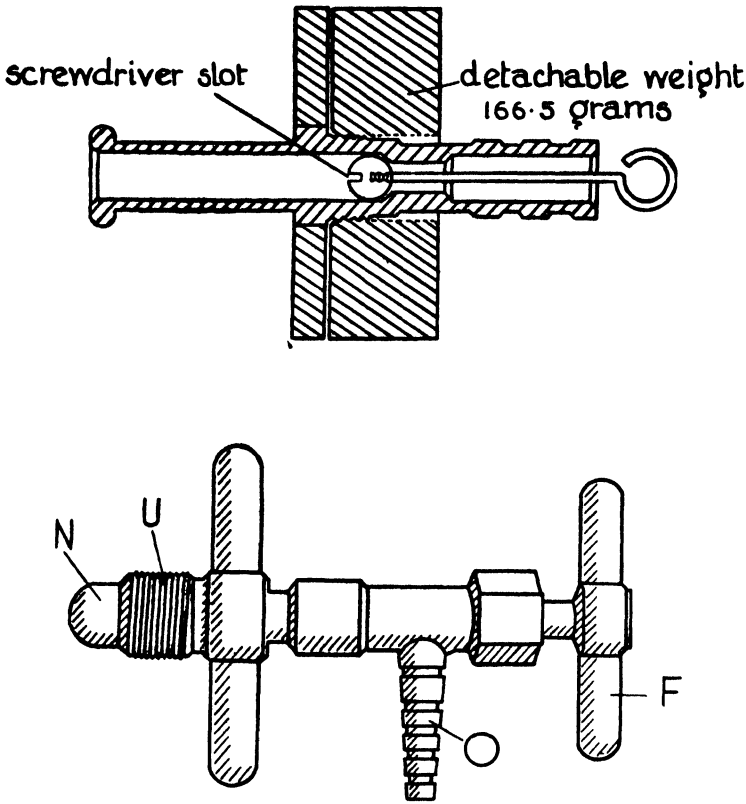


FIG. 17.—PILOT-BALLOON FILLER AND FINE-ADJUSTMENT VALVE.

filler. The filler is designed for use with both 150-inch and 90-inch balloons. The brass weight of the filler, Fig. 17, is made in two portions, the outer one of which may be screwed off. The weight of the complete filler is 238 gm., and when used with a 150-inch balloon gives a rate of ascent of 700 ft./min. When the heavier outer portion is screwed off the weight is reduced to 71·5 gm. and gives a rate of ascent of 500 ft./min. with a 90-inch balloon. In the most

recent type of balloon filler (Mark V) the valve consists of a brass ball and plunger in the form of a piece of brass wire bent into a loop at its end to prevent its passing through the filler. The ball has a threaded hole into which the plunger screws and a screw-driver slot. To remove the ball for cleaning purposes it is held in position by means of a small screw-driver fitting into the slot and the plunger unscrewed. To reassemble, the balloon filler is held vertical with the detachable weight lowermost and the ball dropped into the tube. The ball is manipulated so as to bring the screw-driver slot centrally uppermost and is then held in position by means of the screw-driver while the plunger is screwed into it.

Method of Filling.—In order to fill a balloon, the cap should first be unscrewed from the hydrogen cylinder and the left-hand fine-adjustment valve screwed firmly into the cylinder. In Fig. 17 *N* is the nozzle of the fine-adjustment valve, *U* is the union provided with a left-hand thread, *O* is the outlet for admitting hydrogen to the balloon, and *F* is the handle for operating the fine-adjustment valve, a right-hand screw. The rubber tube is connected to the outlet *O* and to the end of the balloon filler through which the small plunger of the valve projects. The neck of the balloon is placed on the other end of the filler, and should be tied on with a piece of soft twine. The main key *K*, Fig. 18, is then inserted in the main valve *V* of the hydrogen cylinder and the handle *F* eased until it is just finger tight. The main valve *V* is then opened very slightly by *K*, great care being taken that no rush of hydrogen takes place. The handle *F* is used to control the flow of gas and if the rubber is in good condition the filling can be proceeded with quickly, but if it shows any signs of having perished it is better to proceed slowly. Should the balloon burst while being filled no harm will be done to the personnel engaged, provided that there is no naked light in the vicinity. The filling should be continued until the balloon just lifts the filler and the attached tube, the main hydrogen valve *V* then being closed and the rubber tube disconnected from the filler. Care is necessary at this point to hold the balloon firmly or it may carry away the filler with it. Gas should be liberated slowly by pressing up on the plunger in the bottom of the filler until the balloon and filler just float freely in the air. When the hydrogen valve has been closed the fine-adjustment valve can be unshipped and the cylinder cap replaced. Having adjusted the free lift of the balloon the neck should be gripped firmly in the hand to prevent the escape of gas, the filler removed and a knot tied in the neck of the balloon itself. Alternatively, the neck of the balloon may be tied with soft twine and removed from the filler, when the balloon will be ready for the ascent. During filling, the balloon should be examined for weak patches, pinholes, etc., and these should be repaired in the manner described above.

It may, on some occasions, be desirable and possible to turn the ship to steam down wind during the ascent, but if not, the balloon

To face page 68.

FIG. 18.—EQUIPMENT FOR FILLING PILOT BALLOONS.

should always be released on the lee-side. Stop-watches should be started at the instant of release of the balloon or a few seconds before to allow for the height of position of release above the water-line. The average rate of ascent will be either 700 ft./min. or 500 ft./min. according to whether it is a 150-inch or a 90-inch balloon. If, owing to poor visibility, difficulty in following the balloon is anticipated, two balloons may be used instead of one, giving a larger object to be sighted. The two balloons should be filled independently and attached to the opposite ends of a light line about 1½ fathoms long before release.

Method of following the balloon.

It has already been explained on page 61 that in order to calculate the upper-wind currents it is necessary to know the position of the balloon in elevation and azimuth at intervals of one minute from the start of the flight. In order that sufficiently accurate results may be obtained, the observer should aim at measuring elevation and azimuth at these intervals. He should also endeavour to take these observations as accurately as possible, remembering however, that it is generally a waste of time to measure one with great accuracy, if the other can be measured only approximately. (The errors introduced by inaccuracies in the measurement of bearing and elevation are discussed on page 84). The height to which the balloon should or can be followed will depend upon the particular circumstances, but in general it is not necessary to follow it more than 15,000 feet.

Method 1. By sextant and compass.—The simplest method of following balloons is by the use of sextant and compass. The accuracy obtained by this method is not very high, and, as the balloon has to be followed by the naked eye, no long flight is possible, but owing to the simplicity of the equipment required, it is the method most readily available for use. The elevation and bearing of the balloon are observed by means of a sextant and compass respectively, the readings of the latter, if magnetic, being corrected to give true bearings. The first observation for bearing and elevation is taken at one minute, the ship's head (true) being recorded at the same time. A warning to "stand by" should be given to the observers 5 or 10 seconds before each observation. The observations are noted on Table I or Table II of the appropriate balloon-ascent form (H 358). In obtaining the elevation, it is more convenient to hold the sextant inverted and reflect the horizon to the balloon, rather than the balloon to the horizon, which is the usual way; the advantage is that the balloon becomes the direct image. It is also advantageous to keep the eye continuously on the balloon when it becomes faint, an assistant taking the elevation readings. An erecting (star) telescope may be used with the sextant but the desirability of using any telescope

at all is a matter of individual preference. For obtaining the bearing, the ship's compass or a gyro-repeater can be used. If using the latter, care should be taken to ensure that the instrument is correctly aligned to the master compass before commencing observations. This method of observing the bearing has two considerable disadvantages, one being that the balloon cannot be followed with the naked eye for any great distance, and the other that it is difficult to observe the balloon at high elevations. If using the magnetic compass, the compass error must be applied before noting the readings on the form.

Method 2. The tail method of determining the height of a balloon.—This method is used regularly on land, the necessary observations being taken with a theodolite. Until some suitable type of shipboard theodolite has been developed the method cannot satisfactorily be employed at sea. It is considered, however, that the brief description given below may be useful. The principle of the method is to utilize the relation between the height of the balloon and the angle subtended at the observer's eye by a tail of known length attached to the balloon. It is useful when the wind is fairly strong so that the elevation of the balloon does not exceed about 40°, but is not applicable when the elevation is large. From the observations obtained the height of the balloon at various intervals can be calculated. In Fig. 19, if E = elevation of the

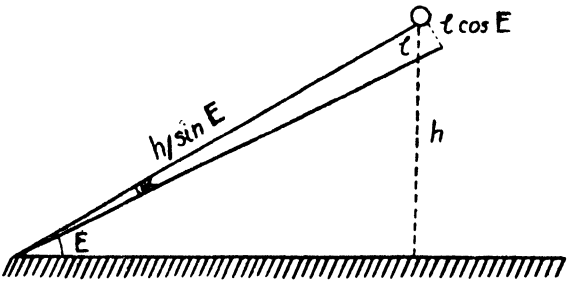


FIG. 19.

balloon, h = its height, l = the length of the tail attached to it and α = the angle subtended by the tail at the observer's eye we have :—

$$\alpha = \frac{l \cos E \sin E}{h} = \frac{l \sin 2E}{2h}$$

$$\text{Therefore } h = \frac{l \sin 2E}{2\alpha}$$

Thus, if α can be measured, we have a means of determining the height of the balloon. This can be done by using a graticule (with scale divisions every 0·01 inch, say) in the eyepiece of a suitable telescope, sextant or theodolite.

The tail consists of a length of thread with a piece of white, silver, red or black paper attached. The colour of the paper to be used will depend upon the cloud conditions ; the paper should be thin and about the size of a piece of foolscap. The tail is observed with the telescope and its apparent length, in terms of the graticule divisions, is determined at intervals, usually at the end of each minute. If the apparent length of the tail is m divisions, and K divisions on the graticule subtend an angle of 1 radian, $\alpha = m/K$

$$\text{and } h = \frac{Kl \sin 2E}{2m}$$

K is a constant of the optical instrument and can be determined once and for all by fixing a foot rule vertically at a distance of 100 feet from the object glass of the instrument and determining its length in terms of graticule divisions. The angle subtended by the foot rule at the observer's eye is $\frac{1}{100}$ radian.

If the rule appears to be x divisions long in the field of view, then $K = 100x$. Thus knowing K and l, we can determine the height of the balloon from the observation of E and the apparent length m of the tail in graticule divisions.

The tail is usually about 100 feet long, but for convenience in calculation, lK should be equal to $1{\cdot}2 \times 10^5$, and l should be chosen to give this or some similar simple factor with the fixed value of K. When l has been decided upon, fix two nails in a wall at a convenient distance, say, $\frac{1}{20}\,l$ apart, and measure off the tail by winding 20 such lengths of thread on the nails.

It may be found that in the early stages of the ascent all the tail is not visible in the field of view, so that the apparent length cannot be obtained. This difficulty can be overcome by fixing a second sheet of paper one-quarter of the way down the tail, below the balloon. Measure the apparent length of this secondary tail, and multiply these lengths by four to obtain the apparent length of the long tail which is not completely within the field of view. As soon as the long tail has come completely into the field of view observations should be made upon it instead of upon the secondary tail.

The apparent length of the tail should be estimated to tenths of a graticule division and the tail should be vertical when the observation is made. If the apparent length of the tail temporarily increases after the observation has been made, this indicates that the tail is swinging and the height calculated from the previous observation will be in excess of the true height of the balloon at that minute. Usually, however, the tail remains approximately vertical ; in any case, the smoothing of the height-time curve should eliminate any occasional errors due to oscillation of the tail.

There is no necessity that observations of E and m should be made by the same instrument and better results will probably be

obtained if one observer concentrates on measuring m, the values of E being obtained with those of A in the usual manner. There is no need either to take observations of m every minute; until practice has been gained m can be read at two- or three-minute intervals, enabling the heights at two- or three-minute intervals to be calculated.

Night-Ascents

When making night-ascents it is necessary to show a small light below the balloon, in order that its position may be visible in the sky. This may be done by suspending from the balloon a small electric torch of dry-battery type; these torches are supplied to all ships carrying pilot-balloon equipment. The batteries are packed in sealed tins, which should be stowed in a cool place and kept unopened until the contents are required for use. The torch is provided with a white bulb but if difficulty is experienced in distinguishing the light from a star a shade of light red paper should be placed round the torch. Attached to the torch is a cloth becket by which it can be suspended from the balloon by a piece of twine.

An alternative and more economical method of illumination is to suspend from the balloon a small paper lantern containing a lighted candle. Owing to practical difficulties this method cannot be used to any great extent at sea. As it is employed at shore stations, however, a description of the method is given here. These lanterns are cylindrical in shape, about $3\frac{1}{2}$ inches in diameter and 5 inches high, the top and bottom being cut out of cartridge paper and the cylindrical part made of tissue paper. A hole 2 inches in diameter is cut in the top of the lantern to provide ventilation. A length of about 2 inches of Christmas candle is stuck down to the bottom of the lantern and provides the illumination. A suspending loop of thin wire 6 inches in length is fitted to the top of the lantern. A length of sailmaker's twine, about 6 feet, is used to attach the lantern to the balloon. When ready, light the candle; to do this press the two ends of the lantern together so that the candle is easily accessible and when alight open the lantern out again. When releasing the balloon one of the following procedures is recommended: either, hold the balloon up until the lantern hangs suspended by the twine and then let go—this can usually only be done in light winds; or, attach a stray-line of twine, about 5 feet long, to the balloon and ease the balloon up on this until the weight of the lantern is taken and then let go. It is essential that the line attaching the lantern to the balloon should be taut before releasing and the lantern should be steadied to avoid swinging. Provided the lantern is not subjected to a sudden jerk there is no reason why the candle should not remain alight. It has been found that if a piece of bare elastic about $2\frac{1}{2}$ feet long and 3/4 mm. cross-section is used in the suspension between the lantern and the balloon the effect of jerking is reduced considerably.

Rate of Ascent in Night-ascents.—The torch or lantern hanging

below the balloon will affect the rate of ascent unless additional lift is given. Balloons of 150-inch should always be used for night-ascents, and when an electric torch is used for illumination, should be filled to lift torch and filler together, the filler being detached after inflation. The rate of ascent will then be reduced and may be determined from the following table :—

Weight of torch attached (grams).	Rate of ascent of 150-inch balloon with torch (ft./min.).
0	700
25	683
50	667
75	653
100	640

For weights of torch intermediate between the above values the rate of ascent may be determined by interpolation. With a lantern of the candle type the balloon may be filled to lift two lanterns in addition to the normal filler, one lantern only being used in the ascent. With the lantern attached the rate of ascent will then be the standard rate of 700 ft./min.

Method of following.—For following balloons at night, the ordinary sextant cannot be used owing to lack of an horizon. A bubble sextant can be used instead, with a gyro repeater for bearings.

METHODS OF COMPUTATION OF WIND SPEED AND DIRECTION.

When the rate of ascent (usually 700 feet per minute), the bearing and the elevation of a balloon at minute intervals, and the course and speed of the ship are known, there are two methods which can be employed to find the wind speed and direction :—

(*a*) By calculation, using a special pilot-balloon slide-rule.

(*b*) Graphical method.

An example of each method is given on pages 80-1.

Method (a)—By slide-rule.—The principle of this method can be understood by reference to Fig. 20 and Fig. 21.

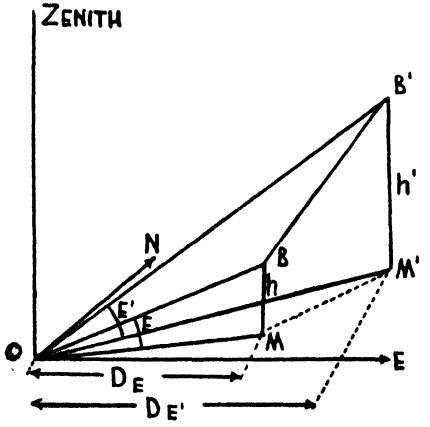


FIG. 20.

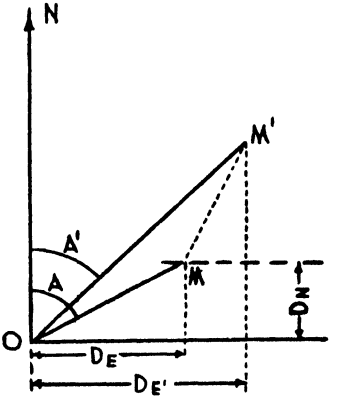


FIG. 21.

In Fig. 20—

B is the position of the balloon relative to the observer, O, at the end of the nth minute.

B' is its position at the end of the $(n + 1)$th minute.

M and M' are the projections of these points on the horizontal plane through the point of observation.

h and h' are the heights, observed or assumed, of the balloon at these times.

E and E' are the corresponding elevations.

In Fig. 21, which is the plan of Fig. 20, the symbols have the above meanings and, in addition, A and A' indicate the bearings. D_E denotes the distance eastward of M from O at the end of the nth minute, *i.e.*, the component of the displacement OM in an easterly direction, or the departure. (If M is to the westward of O, then D_E is reckoned negative). D_N denotes the distance northward of M from O, *i.e.*, the component of the displacement OM in a northerly direction, or the d.lat. (If M is to the southward of O, then D_N is reckoned negative.)

From Fig. 20 we have :—

$$OM = h \cot E$$
$$OM' = h' \cot E'$$

From Fig. 21 we have, further,

$D_E = h \cot E \sin A$: $D_N = h \cot E \cos A$
$D_E' = h' \cot E' \sin A'$: $D'_N = h' \cot E' \cos A'$

If the eastward displacement of the balloon relative to the ship during the $(n + 1)$th minute is represented by V'_E we have

$$V'_E = D_E' - D_E$$

Similarly, if we represent the northward displacement of the balloon relative to the ship during the $(n + 1)$th minute by V'_N we have

$$V'_N = D'_N - D_N$$

If v_E and v_N are the eastward and northward components of the ship's displacement during the $(n+1)$th minute (obtained from Table V) and V_E and V_N are the components of the true horizontal displacement of the balloon in that minute,

$$V_E = V'_E + v_E$$
$$V_N = V'_N + v_N$$

The average speed and direction of the wind in the layer traversed by the balloon in the $(n + 1)$th minute can now be found by entering Table VI with the values V_E and V_N as explained below.

In order to perform the calculations involved in the determination of V_E and V_N a special pilot-balloon slide-rule is provided.

The pilot-balloon slide-rule (Mark I) has a double slider, the two scales of which can be clamped in any position relative to each other.

This double slider is particularly useful in calculating when a constant rate of ascent is assumed. The one (1) mark on its lower or time scale is set opposite the rate of ascent in hundreds of feet per minute (7) on its top scale. The two scales of the slider are then clamped in this position.

On the bottom of the stock of the slide-rule is marked a tangent scale for elevation (E). On the top of the stock are marked cosine and sine scales for azimuth (A).

To calculate the magnitudes of D_E and D_N corresponding with any particular minute, say the third, with the balloon in the north-east quadrant; (1) Set the 3 of the time (or bottom) scale of the slider opposite the observed value of E on the elevation scale. (2) Read off on the top scale of the slider the graduation corresponding with the azimuth reading (for the third minute) on the cosine scale; this gives D_N. (3) Read off the graduation corresponding with the azimuth angle on the sine scale; this gives D_E.

Referring to Fig. 22, if the balloon is in the south-east quadrant

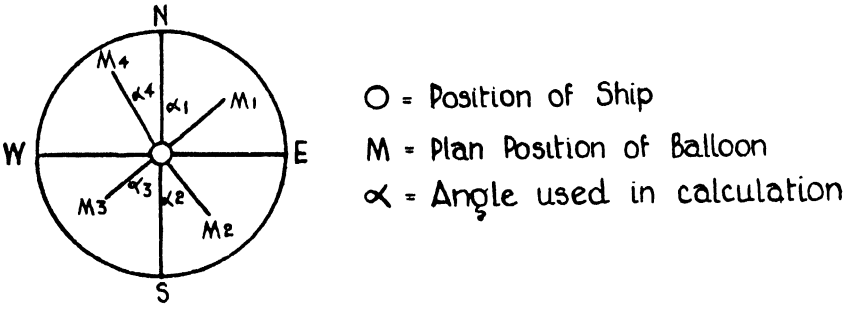


FIG. 22.

read off D_N and D_E against the supplement of the azimuth, *i.e.*, $\alpha_2=(180°-A)$ on the cosine and sine scales. If it is in the south-west quadrant read off D_N and D_E against the angle $\alpha_3 = (A - 180°)$ on the scales, and if it is in the north-west quadrant read off against the angle $\alpha_4 = (360° - A)$ on the scales.

These rules with regard to the azimuth angle appear rather confusing when expressed in words but they can easily be understood from Fig. 22, and the observer performing the calculations is recommended to develop the habit of visualizing this diagram in each particular case as the calculations are carried out. Note that D_N is always determined by means of the cosine scale, D_E by means of the sine scale.

If the height of the balloon is actually observed at each minute interval, the two scales of the slider must be clamped together so as to correspond everywhere. Then if the observed value of h (in hundreds of feet) on the lower scale of the slider is set opposite the corresponding value of E on the elevation scale, the values of D_E and D_N are read off on the top scale of the slider.

It should be noted that in this second method of using the slide rule, if for example, the five on the bottom scale of the slider is used for five hundred feet, then the five on the upper scale of the slider (which should be opposite the five on the lower scale) also represents five hundred feet when used to give D_E and D_N.

The above calculation will give the magnitudes of D_E and D_N for successive minutes. In order to determine the signs of D_E and D_N the computer should note in which quadrant relative to the observer the balloon lay at that particular minute and then refer to this table.

Balloon in	D_E.	D_N.
NE. quadrant	+	+
SE. quadrant	+	−
SW. quadrant	−	−
NW. quadrant	−	+

The table is represented diagrammatically in Fig. 23.

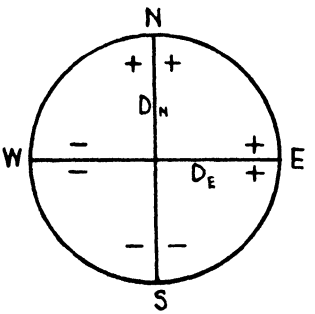


FIG. 23.

After a time the observer will acquire the habit of visualizing this diagram as the calculation proceeds and will find it unnecessary to refer to the table. Note that D_N changes sign only when A passes through 90° or 270°; D_E changes sign only when A passes through 180° and 360°. It is advisable to cast the eye down columns 5 and 6 of the balloon-ascent form when they have been completed, keeping this rule in mind.

Having determined the values of D_N and D_E at the end of each minute, and entered them in columns 5 and 6 of Table I of the form, subtract algebraically from D_E' the preceding value of D_E and enter the results V'_E in column 7. Repeat for the values of D_N and enter results V'_N in column 8.

When A differs from 180° or 360° by an amount less than 2° the limitations of the slide rule prevent the calculation of D_E by the

above method. In such cases, calculate D_N by means of the slide rule and use the formula $D_E = \frac{x}{60} D_N$ (where x = difference in degrees between A and 180° or 360°) for calculating D_E.

Similarly, when A differs from 90° or 270° by less than 2°, D_N cannot be calculated with the slide rule. In this case calculate D_E and use the formula $D_N = \frac{x'}{60} D_E$ where x' = difference in degrees between A and 90° or 270°. These formulae can be deduced

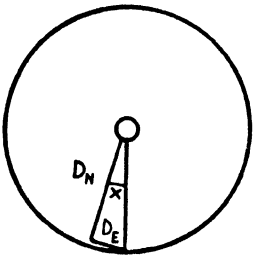


FIG. 24.

from a diagram (Fig. 24); for instance, when A is nearly 180° we have:—

$$\frac{x}{360} = \frac{D_E}{2\pi D_N}$$

$$\therefore D_E = \frac{xD_N}{60} \text{ (taking } \pi = 3 \text{ approximately).}$$

From the ship's course and speed obtain with the aid of Table V (page 82) the values of v_E and v_N and enter in columns 9 and 10. In order to determine the signs of v_E and v_N apply the rule given above (for D_E and D_N). If the ship steers the same course at the same speed throughout the ascent, v_E and v_N will be the same for each minute.

Add algebraically corresponding values of V'_E and v_E and enter result V_E in the column 11. Repeat for V'_N and v_N to obtain V_N (column 12).

To obtain the wind direction and speed from the wind components V_E and V_N, refer to Table VI (page 476).

From this table, take the larger component of the pair from the top horizontal row of figures (1 to 30) and run down the column below it. Also travel along the horizontal line of figures opposite the smaller component (on the diagonal or vertical line of figures). The figures in **clarendon** at the meeting place of the selected line and column give the wind speed, and the figures in *italics* an angle θ

which is actually the angle between the true direction of travel of the balloon in the minute in question and that of the larger of the components, V_E and V_N of its path. θ is directly related to the wind direction ϕ, as shown in the following table. These relations can be readily seen by drawing diagrams, remembering that ϕ, the wind direction, is the direction from which the wind is coming.

V_E	V_N	If V_E greater than V_N.	If V_E less than V_N.
+	+	$\phi = 270 - \theta$	$\phi = 180 + \theta$
+	−	$\phi = 270 + \theta$	$\phi = 360 - \theta$
−	−	$\phi = 90 - \theta$	$\phi = \theta$
−	+	$\phi = 90 + \theta$	$\phi = 180 - \theta$

If $V_E = 0$, $\phi = 180°$ or $360°$, according to the sign of V_N, similarly, if $V_N = 0$, $\phi = 90°$ or $270°$. When either V_E or V_N is zero, the wind speed in knots is numerically equal to the value of V_N or V_E in hundreds of feet per minute, *i.e.*, if $V_N = 300$ ft./min., wind speed = 3 knots.

Instead of using Table VI the wind speed and direction can be obtained from traverse tables, taking V_E as "Departure" and V_N as "Diff. Lat."

Method (b)—Graphical method.—In this method the results may be obtained entirely by plotting. A specially prepared Radial Chart Misc. 70a is convenient for this purpose the scale used being 3 inches = 4,000 feet. In the top left-hand corner, *i.e.*, the *NE.* quadrant of the chart the degree graduations provide the scale for plotting angle of balloon elevation. Circles are drawn at intervals of 100 feet and the vertical axis of the elevation scale is divided in distances of 700 feet (the ascent of the balloon in one minute). When other rates of ascent are used the axis should be divided accordingly. Through the points thus obtained height-lines are drawn, in red, parallel to the horizontal axis. Each line then corresponds to the height of the balloon at the time after release, in minutes, and is marked by its appropriate number.

To find the wind direction and speed for the first and subsequent minutes (Surface—700 feet, 700 feet—1,400 feet, etc.) proceed as follows :—

Using the prepared elevation scale, Fig. 25, lay off the elevation at 1 minute and mark the point at which the parallel-ruler cuts the 1-minute height-line, *A*. Similarly, for each succeeding minute, the elevation for that minute is laid off and the points at which the parallel-ruler cuts the appropriate height-lines are marked *B.*, *C.*, etc. The distance from the vertical axis to the point so obtained, measured horizontally is therefore the plan range of the balloon at

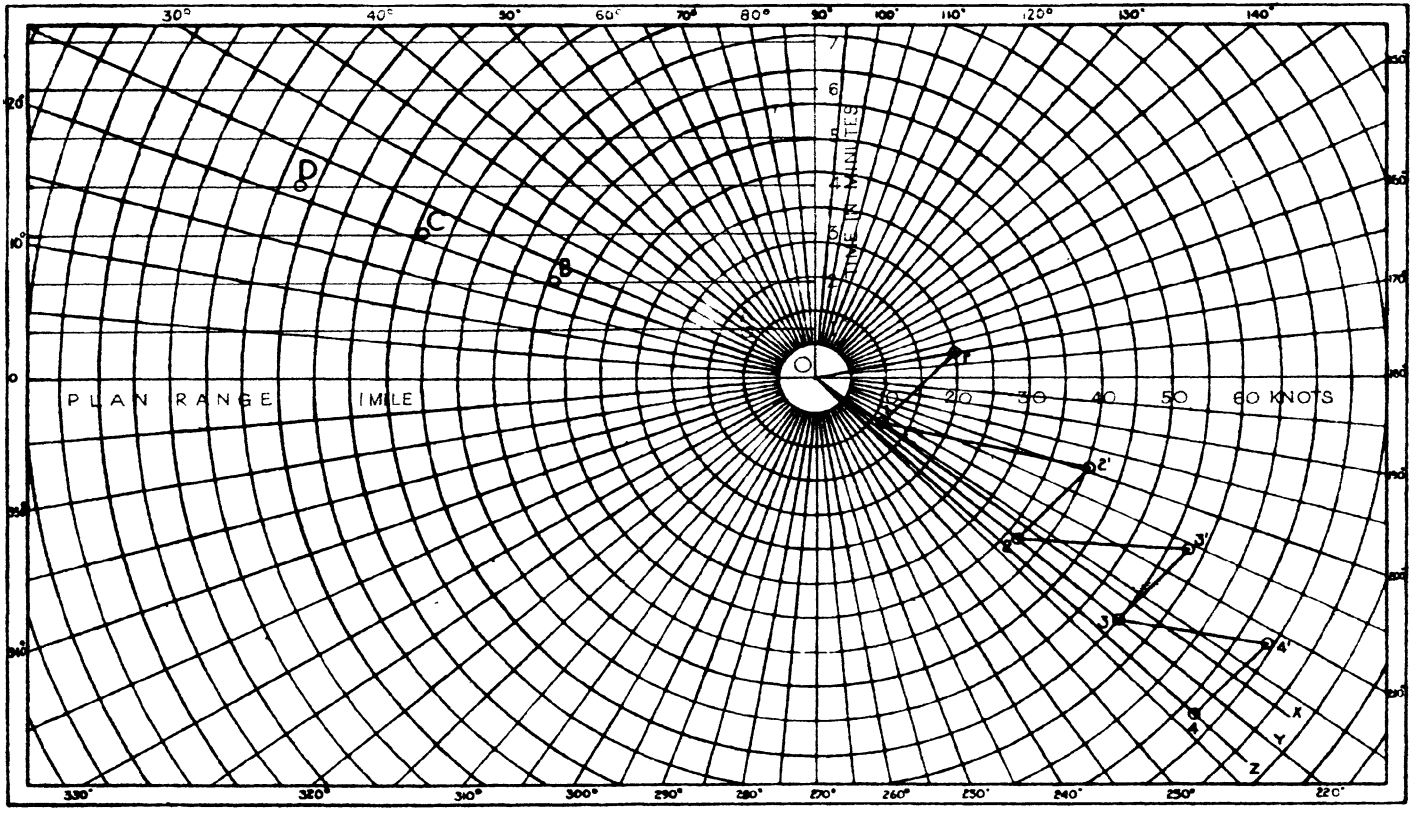


FIG. 25. GRAPHICAL METHOD OF COMPUTATION OF PILOT-BALLOON ASCENT.

To face page 78

the end of each minute, *e.g.*, the distance $OA = h \cot E$ in Fig. 25.

An alternative (and possibly quicker) method of determining the values of $h \cot E$ is by using the pilot-balloon slide-rule. Having set the one (1) mark on the lower or time scale of the slider opposite the rate of ascent in hundreds of feet per minute (7) on the upper scale, the two scales of the slider are then clamped in this position. The figure on the time scale corresponding to the appropriate minute (say 3) is then set opposite the elevation at that minute on the tangent scale. The value of $h \cot E$ is then obtained by setting the cursor to 90° on the sine scale (0° on the cosine scale) and reading off the number opposite to it on the upper scale of the slider.

Having obtained the values of $h \cot E$ by one of the above methods, these should be entered in the appropriate column of the balloon-ascent form. Next, from the centre of the radial chart lay off the bearings of the balloon at each minute (X, Y, Z). These should be drawn in with light lines and numbered to avoid confusion. With dividers, measure the plan range found for each minute on the elevation scale and lay off this distance, from the centre, on the corresponding bearing lines on the radial chart. The points so obtained should be clearly marked 1, 2, 3, etc.

From each point (1, 2, 3, etc.) lay off the course and speed of the ship; these points to be marked 1′, 2′, 3′, etc. The wind direction and speed for each successive minute is then found by joining the centre to point 1′, point 1 to point 2′, point 2 to point 3′, etc., etc.

The wind direction during the first minute is from 0 to 1′, during the second minute from 1 to 2′.

The speed of the wind during each minute is represented by the length of the lines 0–1′, 1–2′, etc. on the speed scale.

The wind direction and speed has now been obtained for each minute (and corresponding 700 feet of ascent) that the balloon has been under observation.

Example of computation of wind speed and direction by graphical method (Fig. 25).

Position { Lat. 56° 15′ N. / Long. 01° 08′ W.

Pilot Balloon Ascent from H.M.S. *X*. Course and Speed, 135°, 14 knots.

Date, 9th July, 1936. Time (Z.T.) 1110 (B.S.T.) Surface Wind { Direction, 345° / Speed, 18 knots.

Rate of Ascent of Balloon, 700 ft./min.

t	*h*	*E*°	Ship's		Relative Bearing.		True Bearing *A*°.	*h* Cot *E*.	Wind.		Remarks.
			Speed (knots).	Head (true).	Red	Green			Direction (true).	Speed (knots).	
1	2	3	4	5	6	7	8	9	10	11	12
1	7	32·6	14	135	—	82	217	1603	350°	19	Balloon lost in cloud.
2	14	20·9	14	135	—	85	220	3666	015°	29	
3	21	20·9	14	135	—	86	221	5500	004°	24	
4	28	21·2	14	135	—	88	223	7219	010°	21	

Example of computation of wind speed and direction by slide-rule method.

Position { Lat. 50° 16′ N. / Long. 1° 28′ W.

Pilot Balloon Ascent from H.M.S. *X*. Course and Speed, 315°, 14 knots.

Date, 10th July, 1936. Time (Z.T.) 1015 (B.S.T.) Surface Wind { Direction, 175° / Speed, 18 knots.

Rate of Ascent of Balloon, 700 ft./min.

Time from start mins. t	Azimuth Degrees (true). A.°	Elevation. E°.	Height 100's of feet. h	Apparent.		Apparent.		Ship's.		True.		Wind		Remarks.
				Distance East. D_E	Distance North D_N	Speed.		Speed.		Speed.		Direction (Degrees true). ϕ	Speed (knots). V	
						W—E V_E	S—N V_N	W—E V_E	S—N V_N	W—E V_E	S—N V_N			
1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
1	037	23·6	7	+ 9·7	+12·8	+ 9·7	+12·8	−10·0	+10·0	− 0·3	+22·8	179	23	Balloon lost in cloud.
2	040	20·9	14	+23·5	+28·0	+13·8	+15·2	−10·0	+10·0	+ 3·8	+25·2	189	25	
3	041	20·9	21	+36·0	+41·5	+12·5	+13·5	−10·0	+10·0	+ 2·5	+23·5	186	24	
4	043	21·2	28	+49·5	+53·0	+13·5	+11·5	−10·0	+10·0	+ 3·5	+21·5	189	22	

TABLE V.—TABLE FOR OBTAINING SHIP'S $D_E D_N$.

In Hundreds of Feet per Minute.

Course	10 knots.		11 knots.		12 knots.		13 knots.		14 knots.		15 knots.		Course
	N.	E.	N.	E.	N.	E.	N.	E.	N.	E.	N.	E.	
0	10·1	0	11·1	0	12·2	0	13·2	0	14·2	0	15·2	0	90°
5°	10·1	0·9	11·1	1·0	12·1	1·1	13·1	1·1	14·1	1·2	15·1	1·3	85°
10°	10·0	1·8	11·0	1·9	12·0	2·1	13·0	2·3	14·0	2·5	15·0	2·6	80°
15°	9·8	2·6	10·8	2·9	11·7	3·1	12·7	3·4	13·7	3·7	14·7	3·9	75°
20°	9·5	3·5	10·5	3·8	11·4	4·2	12·4	4·5	13·3	4·9	14·3	5·2	70°
25°	9·2	4·3	10·1	4·7	11·0	5·1	11·9	5·6	12·9	6·0	13·8	6·4	65°
30°	8·8	5·1	9·6	5·6	10·5	6·1	11·4	6·6	12·3	7·1	13·2	7·6	60°
35°	8·3	5·8	9·1	6·4	9·9	7·0	10·8	7·5	11·6	8·1	12·4	8·7	55°
40°	7·8	6·5	8·5	7·2	9·3	7·8	10·1	8·5	10·9	9·1	11·6	9·8	50°
45°	7·2	7·2	7·9	7·9	8·6	8·6	9·3	9·3	10·0	10·0	10·7	10·7	45°
	E.	N.	E.	N.	E.	N.	E.	N.	E.	N.	E.	N.	

Course	16 knots.		17 knots.		18 knots.		19 knots.		20 knots.		21 knots.		Course
	N.	E.	N.	E.	N.	E.	N.	E.	N.	E.	N.	E.	
0	16·2	0	17·2	0	18·2	0	19·2	0	20·3	0	21·3	0	90°
5°	16·1	1·4	17·2	1·5	18·2	1·6	19·2	1·7	20·2	1·8	21·2	1·9	85°
10°	15·9	2·8	17·0	3·0	18·0	3·2	19·0	3·3	19·9	3·5	21·0	3·7	80°
15°	15·6	4·2	16·6	4·5	17·6	4·7	18·6	5·0	19·6	5·2	20·5	5·5	75°
20°	15·2	5·5	16·2	5·9	17·1	6·2	18·1	6·6	19·0	6·9	20·0	7·2	70°
25°	14·7	6·8	15·6	7·3	16·5	7·7	17·4	8·1	18·4	8·6	19·3	9·0	65°
30°	14·0	8·1	14·9	8·6	15·8	9·1	16·7	9·6	17·5	10·1	18·4	10·6	60°
35°	13·3	9·3	14·1	9·9	14·9	10·5	15·8	11·0	16·6	11·6	17·5	12·2	55°
40°	12·4	10·4	13·2	11·1	14·0	11·7	14·7	12·4	15·5	13·0	16·3	13·7	50°
45°	11·5	11·5	12·2	12·2	12·9	12·9	13·6	13·6	14·3	14·3	15·0	15·0	45°
	E.	N.	E.	N.	E.	N	E.	N.	E.	N.	E.	N.	

Course	22 knots.		23 knots.		24 knots.		25 knots.		26 knots.		27 knots.		Course
	N.	E.	N.	E.	N.	E.	N.	E.	N.	E.	N.	E.	
0	22·3	0	23·3	0	24·3	0	25·3	0	26·3	0	27·3	0	90°
5°	22·2	1·9	23·2	2·0	24·2	2·1	25·2	2·2	26·2	2·3	27·2	2·4	85°
10°	22·0	3·9	23·0	4·0	23·9	4·2	24·9	4·4	25·9	4·6	26·9	4·8	80°
15°	21·5	5·8	22·5	6·0	23·5	6·3	24·5	6·5	25·4	6·8	26·4	7·1	75°
20°	21·0	7·6	21·9	8·0	22·8	8·3	23·7	8·6	24·7	9·0	25·7	9·3	70°
25°	20·2	9·4	21·1	9·8	22·0	10·2	22·9	10·7	23·8	11·1	24·7	11·5	65°
30°	19·3	11·1	20·2	11·7	21·1	12·2	22·0	12·7	22·8	13·2	23·7	13·7	60°
35°	18·3	12·8	19·1	13·3	19·9	13·9	20·7	14·5	21·6	15·1	22·4	15·7	55°
40°	17·1	14·3	17·8	14·9	18·6	15·6	19·4	16·3	20·2	16·9	20·9	17·5	50°
45°	15·8	15·8	16·5	16·5	17·2	17·2	17·9	17·9	18·6	18·6	19·3	19·3	45°
	E.	N.	E.	N.	E.	N.	E.	N.	E.	N.	E.	N.	

The above table shows the distance in hundreds of feet which the ship travels per minute in a northerly and easterly (or southerly and westerly) direction for different courses.

Calculation of Winds at High Levels.—Above 10,000 feet the observations are not in general sufficiently accurate to justify the computation of wind speed and direction at minute intervals, *i.e.*, for 500-feet or 700-feet layers. Between 10,000 feet and 16,000 feet the mean wind in layers approximately 2,000 feet thick should be computed. Subsequently, the wind at the 20,000 feet level and above should be computed by steps of 5,000 feet. For balloons ascending at 700 ft./min. the following table indicates the layers to be taken for the 10,000 feet level and above :—

Feet.			Feet.	Min.
*10,000	as mean	wind from	9,100 to 11,200,	13–16 (3 min. run)
12,000	,,	,,	11,200 ,, 12,600,	16–18 (2 ,,)
*13,000	,,	,,	11,900 ,, 14,000,	17–20 (3 ,,)
14,000	,,	,,	13,300 ,, 14,700,	19–21 (2 ,,)
*16,000	,,	,,	15,400 ,, 16,800,	22–24 (2 ,,)
*20,000	,,	,,	17,500 ,, 22,400,	25–32 (7 ,,)
25,000	,,	,,	22,400 ,, 27,300,	32–39 (7 ,,)
30,000	,,	,,	27,300 ,, 32,200,	39–46 (8 ,,)

* = International heights.

In each case the differences between the apparent D_E and D_N at beginning and end of period must be divided by the number of minutes in the period (given in brackets at end of lines) and then the ship's v_E and v_N added to it. Lastly, the resultant is found in the usual way.

Observations at minutes intermediate to those actually used in calculation of winds at 10,000 feet and above need not be worked out, but they should be made and recorded in case a check is required on the observations actually used.

Averaging Wind Speeds and Directions.—It should be borne in mind throughout that the resultant for any minute is not the wind at the height reached at the end of that minute, but is the mean wind in the layer traversed by the balloon during that minute. Thus, the first minute, gives not the wind at 700 feet, but the mean wind from 0 to 700 feet, *i.e.*, approximately that at 350 feet. If the wind at 5,000 feet be required, the 7th min. wind should not be used, as that is the mean wind from 4,200 to 4,900 feet ; the mean of 7th and 8th min. should be taken as this gives the mean for the layer from 4,200 to 5,600 feet, *i.e.*, for layer centering near 5,000 feet.

Note that to obtain the mean value of two minutes already worked out, the arithmetic mean of the two individual values should not be used, but the means of the V_E and V_N for the two minutes should be found and the resultant then looked out.

Cases of sudden variations in speed or direction or of both together should be carefully examined. It will often be found that one minute gives a speed markedly above and the next minute gives a speed markedly below the general level of the preceding

and following minutes (or *vice versa*). This is almost certainly due to an inaccurate elevation reading and only the mean value for the two minutes concerned should be used. Similar cases of changes in direction may be due to inaccurate azimuth readings.

Accuracy of Observations.

In view of the great difficulty in obtaining really accurate values of A and E in ships, it is in general sufficiently accurate :—

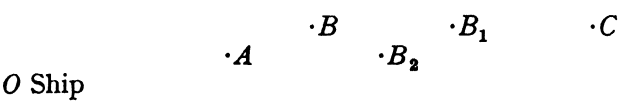
(1) to enter apparent D_E and D_N to nearest ·5 for first 3 or 4 minutes ; subsequently only to nearest whole number ;

(2) to treat ship's v_E and v_N as above.

(If ship's D_E and D_N are being used, then ship's v_E and v_N must be taken to nearest 1/10 and multiplied by the number of minutes, the nearest whole number to this product being entered, otherwise a cumulative error will be introduced) ;

(3) to enter ϕ to nearest 5° (nearest 10° when $V < 5$ knots) and V to nearest knot only.

Errors.—It has already been stated that the observations on the pilot balloon should be taken with the greatest possible accuracy. It will be profitable to consider the kind of errors introduced in the calculated wind by errors of stated magnitude in the observations. In general, the wind deduced from the pilot-balloon ascent should change more or less regularly from one minute to the next, and any sudden change which is found and which is not supported by the subsequent observations, though it does not necessarily mean an error of observation, yet suggests the desirability of considering this as the possible cause. In particular a change in wind either of speed or direction, which is reversed by the next reading, is almost always due to an error. The explanation of this is plain from the following conditions :—



In this diagram the true position of the balloon at three successive minutes is indicated by the points marked A, B and C ; let it be assumed that while A and C are correctly observed, an error is made in calculating either the distance or the bearing of the point B so that the position of the balloon is plotted at B_1 or B_2 respectively. In the case where the error is made in calculating the distance it is clear that the calculated speed relative to the ship for the minute A–B_1 will be in excess of the actual relative speed, while that from B_1–C will be below it. In the case where the error is made in bearing the calculated relative wind from A–B_2 will be veered from the true direction AB, while that from B_2–C will be backed from BC.

In each case an error in the intermediate reading causes a sudden change of wind to be reversed in the following minute.

It is necessary to form an idea as to how accurately the position of the balloon must be known at each minute in order to give fairly accurate results. A wind of 1 knot corresponds with a movement of about 100 feet per minute, so that if the position is accurately observed at one minute and is 100 feet in error at the following minute, the error in the deduced wind over that minute will be 1 knot.

Errors which may occur are as follows :—

(1) Errors in elevation ;
(2) Errors in azimuth ;
(3) Errors in height, owing to the rate of ascent differing from the assumed rate of ascent.

(1) **Errors in Elevation.**—These become increasingly important at low elevations and at great distances. Thus, if the balloon is at an elevation of 50° at the end of 5 minutes, the height being then 3,500 feet, an error of 1° in the observed elevation will only lead to an error in position of 100 feet and is, therefore, not of importance. If, however, the elevation is 10° at the end of 5 minutes, the true distance of the balloon will be 19,800 feet, and if the elevation is inaccurately observed as 11°, the calculated distance will be 18,000 involving an error of 1,800 feet.

(2) **Errors in Azimuth.**—The effect of an error of 1° in the azimuth is dependent solely upon the distance of the balloon from the ship. At a distance of 6,000 feet, the effect of such an error will be that the balloon is inaccurately placed by 100 feet while at 20,000 feet distance, the error in position will be 350 feet and thus becomes of some importance.

It will be observed that as regards both the observations of elevation and azimuth, errors of the order of 1° are of little importance when the balloon is close to the ship, but may become of considerable importance towards the end of the flight, particularly in a strong wind when the balloon is at a low elevation and 4 or 5 miles distant.

(3) **Errors in Height.**—As the height of the balloon is normally determined on the assumption that the rate of ascent is 700 feet per minute, the errors introduced by departure from this rate will be considered. If the ship is stationary the effect of an error in the rate of ascent which persists throughout the flight is simple. Such an error means that all the calculated wind speeds are subject to a proportional error, wind directions being unaffected. Thus, if the balloon actually ascends at 600 feet instead of 700 feet per minute all speeds will be too high in the proportion 7 to 6. The heights to which the winds are ascribed will also, of course, be in error. A wind which is found 5 minutes after the start is ascribed to a height of 3,500 feet, whereas it occurs, in the case mentioned, at a height of 3,000 feet. It is unusual for a balloon to ascend at a

markedly different rate from the normal over the whole of the flight. A more common case is for the balloon to be caught in an ascending or descending current at some stage, so that while it may ascend at 700 feet per minute for, say, the first 3 minutes, in the fourth minute it may be drawn up 800 feet or ascend only 600. If the wind is uniform the same rule applies as that just stated, *viz.*, the wind speed calculated in this minute will have an error in proportion to the error in the rate of ascent. If, however, the wind changes with height, this rule will cease to be strictly true, although it may still be taken as a rough guide, except in cases where, for example, there is a reversal of wind above, when no simple rule of this kind can be given. The rules above apply to a stationary ship. If the ship is moving they cease to be applicable to the actual wind deduced from the observations though they have some application to the speed of the wind relative to the speed of the ship.

In general, pilot-balloon ascents should be made regularly while at sea, but if ships are proceeding in company one ship should be detailed to make such observations; other ships should take them when they require them for operational purposes. Observations should not be made within 50 miles of a shore meteorological station at which routine pilot-balloon observations are made, unless they are required for operational purposes in the ship. The positions of shore meteorological stations making routine pilot-balloon observations are shown on the chart of "Distribution of observations of upper winds and upper-air temperatures over the ocean." (Chart No. 5075.)

The measurement of Upper Winds by observations of smoke-shell.

A simple and effective method of determining the wind speed and direction at a certain height is that of firing a smoke-shell to burst at the required height. The plan position of the smoke-cloud thus formed can be determined, by means of the method described earlier in this chapter, at successive intervals as it is borne along by the wind. From this information the wind speed and direction can be determined.

Coding of reports of Upper Winds.

If possible, a report of the results of the most recent pilot-balloon ascent should be added to the routine weather reports transmitted by H.M. Ships at sea to the appropriate meteorological centre. This report should be in the following form :—

Ships Pilot YQLLL lllGG $H_1ddv_1v_1$. $H_1ddv_1v_1$. $H_1ddv_1v_1$.—Individual elements in above report.

(1) YQLLL lllGG, *see* Chapter II, page 52.

N.B.—GG (= time of observation, G.M.T.) is the time of commencement of the ascent, to the nearest hour, and the position reported is the position of the ship at this time.

(2) $H_1ddv_1v_1$: Each group in this form gives the direction of the wind (dd) in degrees from true North to the nearest ten degrees (*i.e.*, 270° is coded as 27), and the speed of the wind (v_1v_1) in knots, at the height indicated by the first figure of the group (H_1) ; the code for H_1 is given in Code XVII.

Specimen message.—Monday, 2nd August, ship was in position lat. 50°40′ N., long. 23° 18′ W. Ascent commenced 0915 G.M.T. The observed winds were :—

Height (feet).	Wind.	Coded as.
Surface	235°, 9 knots	02309
1,000	277°, 10 knots	12810
2,000	293°, 15 knots	22915
3,000	290°, 15 knots	32915
5,000	254°, 26 knots	42526
7,000	267°, 17 knots	52717
10,000	263°, 15 knots	62615

The coded report of upper winds to be added to the ship's weather report is therefore :—

Ships Pilot 20507 23309 02309 12810 22915
32915 42526 52717 62615.

In order to provide a more detailed representation of the wind structure in the upper air, the Meteorological Conference held at Salzburg in September, 1937 decided to adopt a new International Code for reporting upper winds.

If this code should be adopted in the Navy, the code described above for reporting upper winds will become obsolete, and such reports will then be made in the form :—

Ships Pilot YQLLL lllGG HHddv$_5$ HHddv$_5$ etc.

Individual elements in above report.

(i) *YQLLL lllGG*, as for old code (*see above*) ;
(ii) *HHddv*$_5$: in this group :—

HH = the height of the layer in which the wind is measured, in hectometres above sea level ; the corresponding values in feet are shown in Table VIIA. *Note.*—The code for *HH* is self-evident. Thus :— *HH* = 00, 05, 10, 15, etc. correspond to heights of 0, 5, 10, 15, etc. hectometres respectively.

Normally, the winds reported should be that for sea level and for subsequent heights corresponding as closely as possible to figures in the code where the wind is *directly* observed. Thus, with a balloon rising at a rate of 700 feet per minute, the wind observed during the 1st—2nd minute of the ascent will be that of the layer centering at 1,050 feet, *i.e.*, where *HH* = 03, and this

wind should be reported. Similarly during the 4th—5th minute the wind observed will be that of the layer centering at 3,150 feet, *i.e.*, where $HH = 09$. Again, the wind observed during the 6th—7th minute will be that of the layer centering at 4,550 feet, *i.e.*, where $HH = 14$. In general, then, the winds to be reported are those observed at height intervals of approximately 5 hectometres, *i.e.*, at heights differing by approximately 05 in the code.

dd = direction of the wind on the scale 01-36, *i.e.*, degrees from 360° divided by 10 and rounded off to the nearest whole number, (00 = calm) ; thus 312° will be coded as 31, 167° as 17, 014° as 01, 007° as 01, etc.

Note.—001°, 002°, 003° and 004°, are coded as 36 and not as 00.

v_5 = speed of the wind according to the code given in Table VIIB. *Note.*—The approximate wind speed in km. per hour up to the value of 45 km. per hour (*i.e.*, 24 knots) is obtained by multiplying the code figure by 5. Similarly the speed in knots is approximately equal to the speed in km. per hour divided by 2. (For wind speeds in excess of 45 km. per hour, *see below*). Thus $v_5 = 2$ corresponds to a range of 8 — 12 km. per hour, *i.e.*, a range whose mean value is 10 km. per hour. The values in knots corresponding to the code figures are shown in Table VIIB. For wind speeds between 26 knots and 52 knots (inclusive), 50 is added to the code figures *dd* (*see* Table VIIB.)

For wind speeds of 53 knots to 78 knots (inclusive) at any height for which a value is reported, the code figure for v_5 is obtained from Table VIIB. and an index group 44444 is added to the end of the message, followed by one or more 5-figure groups of the form $m_1m_2m_3m_4m_5\ m_6m_7m_8XX$ where m_1 refers to the first group of the form $HHddv_5$ in the message, m_2 to the second group and so on. m is coded as 0 to indicate the corresponding groups in which the wind speed is less than 53 knots. m is coded as 1 to indicate the corresponding groups in which the wind speed is 53 knots or above. For example :—44444 00011 $100XX$ at the end of a message would indicate that the wind speed is less than 53 knots in groups 1, 2, 3, 7 and 8 of the form $HHddv_5$ in the message, and 53 knots or greater in groups 4, 5 and 6.

For wind speeds above 78 knots, 50 is added to the appropriate wind direction, the supplementary groups 44444 and $m_1m_2m_3m_4m_5$ etc. indicating the groups in which the wind speed is greater than 52 knots.

When there are marked discontinuities in the wind at certain levels, one additional group should be given for each change and should give the height, wind direction and speed of the first observation which clearly shows the change. This group should be given in its proper place in the message.

Specimen message :—Thursday, 15th July, 1937, ship was in position 51° 25′ N., 19° 14′ W. Ascent commenced 1340 (+1). The observed winds were :—

Minute	Height	Wind	Coded as
	Sea level	314° 3 knots	00311
1–2	1,050 ft. (03 hectometres)	309° 18 ,,	03317
4–5	3,150 ,, (09 ,,)	285° 30 ,,	09791
6–7	4,550 ,, (14 ,,)	260° 33 ,,	14762
9–10	6,650 ,, (20 ,,)	235° 22 ,,	20238
*10–11	7,350 ,, (22 ,,)	187° 25 ,,	22199
11–12	8,050 ,, (24 ,,)	162° 16 ,,	24166

* Extra group showing commencement of change in air current.

The coded report of upper winds is therefore :—

Ships Pilot .. 50514 19215 00311 03317 09791
14762 20238 22199 24166

If it is considered impracticable to report the winds at the above-mentioned heights the values computed for the surface and 1,000 feet should be reported, together with any marked changes that may be observed subsequently, with their corresponding heights. The code figures for all such heights and for wind speeds should be obtained from Tables VIIA and VIIB as described above, by entering the table with heights in feet and wind speeds in knots.

TABLE VII.A.—TABLE FOR CODING AND DECODING *HH* IN FEET.

HH	0	1	2	3	4	5	6	7	8	9
00	0	300	600	1,000	1,300	1,600	2,000	2,300	2,600	3,000
10	3,300	3,600	3,900	4,300	4,600	4,900	5,200	5,600	5,900	6,200
20	6,600	6,900	7,200	7,500	7,900	8,200	8,500	8,800	9,200	9,500
30	9,800	10,200	10,500	10,800	11,200	11,500	11,800	12,100	12,500	12,800
40	13,100	13,500	13,800	14,100	14,400	14,800	15,100	15,400	15,700	16,100
50	16,400	16,700	17,100	17,400	17,700	18,000	18,400	18,700	19,000	19,400
60	19,700	20,000	20,300	20,700	21,000	21,300	21,700	22,000	22,300	22,600
70	23,000	23,300	23,600	24,000	24,300	24,600	24,900	25,300	25,600	25,900
80	26,200	26,600	26,900	27,200	27,500	27,900	28,200	28,500	28,900	29,200

TABLE VII.B.—TABLE FOR CODING AND DECODING v_5 IN KNOTS.

v_5		0	1	2	3	4	5	6	7	8	9
	Wind Speed in Knots.	0	3	5	8	11	13	16	19	22	24
If 50 is added to wind direction		27	30	32	35	38	40	43	46	49	51
If group 44444 applies		54	57	59	62	65	67	70	73	76	78
If 50 is added to wind direction *and* group 44444 applies		80	83	85	88	91	94	96	99	102	105

CHAPTER IV.

The Measurement of Upper Air Temperatures

Observations from Aircraft.

A knowledge of temperature and humidity conditions in the upper air is of immense value to the forecaster. Observations are normally made by aircraft operating from land stations but, if and when possible, reports of upper-air temperatures and certain other observations should be made by aircraft attached to H.M. Ships. The reports of these observations should be recorded in the meteorological log and, in addition, transmitted in the appropriate code to the nearest meteorological headquarters.

In many countries daily routine flights are carried out (*e.g.*, at Mildenhall, in England) and the observations are utilized by the national forecasting service. Furthermore, a knowledge of the temperature and humidity distribution in the upper air being of great assistance in the process of analysing air masses it is therefore of considerable value in single observer forecasting.

Observations of Temperature.—The temperature and humidity observations are obtained by using an air psychrometer, strut type, in conjunction with an aeroplane aneroid barometer graduated in millibars or with an ordinary altimeter when no aneroid is available.

When the results of observations are to be signalled to the Meteorological Office, London, flights should be carried out in summer at about 0600 and 1200 G.M.T. and in winter at 0800 or 0830 and at 1300 G.M.T. If these times are inconvenient records of observations made at other periods should be sent.

The flight should be made if possible to a height corresponding to a barometric pressure of 550 millibars (about 16,000 feet), somewhere in the vicinity of the aerodrome or of the ship. The aircraft should be prepared to fly through cloud. The normal equipment consists of :—

(i) an air psychrometer consisting of wet-and dry-bulb spirit thermometers, graduated to −35° C., with open scales to facilitate reading at a distance.

(ii) Aneroids graduated in millibars.

(iii) In addition to the above a distant-reading thermometer may be carried.

Observations should be made at the surface and when the altimeter reads 1,000 feet above the surface : thereafter at 950, 900, 850, 800, 750, 700, 650, 600 and 550 millibars, respectively. The surface temperatures should be read from screen thermometers or preferably an electric psychrometer at the beginning and end of the

flight. The aircraft observations should be made both on the ascent and descent, both dry and wet bulb being read to the nearest half degree centigrade ; but with suitable apparatus, to the nearest tenth degree. At each height where observations are being made the aircraft should be flown level for two minutes before the readings are taken, in order that the thermometer may register the current temperature. With high-speed aircraft it is important that the air speed should be reduced to 100 knots on flattening out and should not exceed this figure during the 2 minutes of level flight before the readings are taken. This precaution is necessary because at higher speeds the effect of the speed of the aircraft is appreciably to increase the reading of the thermometer. At temperatures below 0° C. just before the water freezes on the wet bulb, it is sometimes found that the wet bulb reads higher than the dry. Great care must be taken on these occasions not to confuse the dry bulb with the wet bulb. (The wet-bulb reading is ignored in these cases.) After freezing, the wet bulb will usually continue to fall below the dry for some time until the coating of ice on the muslin has evaporated. The readings cannot be regarded as true wet-bulb temperatures after the first two or three minutes from the time of freezing but they should be reported, as they will be at least as high as the true wet-bulb readings and often give valuable information as to the humidity.

It is sometimes found that the temperature rises rapidly for a few hundred feet instead of falling in the usual manner. This is especially the case above clouds with level tops or above haze with a sharply defined upper surface. The amount of these " inversions," with the pressure and temperature at their upper and lower limits, should be recorded, and to do this accurately the temperature should be read at the bottom of the inversion and then the engine should be throttled down and the aircraft climbed slowly till the temperature ceases to rise. On the descent the pressure and amount of the inversion should again be noted, the aircraft descending slowly as the inversion may not be exactly on the same level as during the ascent.

Observations of Cloud and Visibility.—In addition to the information regarding temperature and humidity, a knowledge of the type, quantity and description of cloud, together with the visibility encountered during the flight, is useful to the forecaster. Consequently the record should contain as full a description as possible of all clouds observed and also of haze, or any other observation of interest. It should be borne in mind in this connexion that, since the usual cloud classification is based on the appearance of clouds from the ground, it is sometimes more convenient for an aerial observer to give simple descriptions without using names, *e.g.*, " continuous layer," " broken layer," " thin patches," " series of thin layers," etc.

When cloud types are described by name it must be remembered that a flat rippled cloud sheet is normally denoted by "stratocumulus" when it lies below 10,000 feet and "altocumulus" above that level. The cloud known as "altostratus" is a layer at middle heights (normally between 10,000 feet and 20,000 feet, or 700 and 450 millibars) which appears uniform from the ground and is usually not a definite layer but an ill-defined cloud, thousands of feet thick. If possible a camera should be carried in the aircraft to photograph the clouds, a hand-camera for oblique photographs being best. Sheets of cloud should be photographed from a height of about 2,000 feet above them, holding the camera horizontal or pointing slightly downwards, but as a rule with the horizon in the photograph.

Photographs should be taken within about 45° of the sun in azimuth, but large masses of cumulus cloud may also be photographed away from the sun. Slow plates should be used, with a short exposure if the sun is shining. Good results are obtained with Process plates and no filter, but with an expert photographic staff available, panchromatic process plates with a K.3 filter or (preferably) a red filter might be tried for still better results. Special attention should be paid to unusual cloud effects, or to wave formations. A record should be kept of times, places and heights at which photographs were taken, the directon in which the camera was pointing and, if possible, the height, character and thickness of the clouds.

Detail of Meteorological Instruments employed for observations from aircraft.

Thermometers.—The Air Psychrometer, strut type, consists of two large thermometers mounted side by side on a duralamin frame provided with fixing straps. The bulbs of the thermometers are covered by a sun-shield to protect them from the direct rays of the sun. The shield is of aluminium sheet formed into a box to surround the two bulbs. The ends of the box are partly closed with baffles which deflect an air stream over the bulbs. A water container is fixed below the sun-shield. To avoid the possibility of water reaching the dry bulb from the wet bulb, the wet bulb must always be placed abaft the dry bulb. The psychrometer should be securely strapped to a convenient vertical strut (it is only intended for use in biplanes) clear of the slip-stream, and facing directly towards the position from which it will be read. (Full details of the construction, operation and maintenance of this instrument are contained in Air Publication 1275, Vol. I. General Instrument Equipment for Aircraft.)

Aneroids.—The millibar aneroid should be set at the same reading as a standard mercury barometer, corrected for index error, latitude and temperature. The adjustment is carried out by turning the dial, after raising the clamp at the top. If left undisturbed the

aneroid should keep the correct setting, but this should be checked at frequent intervals.

Altimeter.—The ordinary altimeter, graduated in feet, should always be set to read zero before taking off.

Distant-reading thermometer.—The thermometer is of liquid expansion type in which the bulb and the indicator are connected by a capillary tube filled with mercury. The bulb should be fixed in a position at least six or eight feet from the fuselage where it will be free from hot air from the engine. The bulb is clamped to a wing spar by means of the holding down bolts and back piece. The dial is fixed in the cockpit in a convenient position for reading. In fixing the dial and the capillary tube, care must be taken that the tube is not bent into a sharp angle. The readings from the thermometer should be compared once weekly with a simultaneous reading of an ordinary mercury thermometer. Care should be taken that the sun is not shining on the bulbs.

Meteorograph.—A meteorograph is sometimes supplied to ships carrying aircraft in order that a continuous record of the pressure, temperature and humidity changes may be obtained throughout the flight.

Observations by Sounding Balloon and Meteorograph.

The meteorograph may also be used in conjunction with a registering or sounding balloon, or a kite, whereby an autographic record of pressure, temperature and sometimes humidity in the upper air may be obtained. A description of this instrument together with instructions for its use at sea is given below.

The registering or sounding balloons which carry the meteorograph are of larger size than those used in pilot-balloon observations. They are sent up in pairs, in tandem, and carry, in addition to the meteorograph, a sea-anchor float. The balloons are followed by the ship as long as they are in sight and an estimate is made of where they are likely to fall. On attaining a great height, say 40,000 or 50,000 feet, one of the balloons will burst: the remaining balloon, being unable to support the instruments and sea-anchor float, descends until the latter reaches the water. The balloon is now able to support the meteorograph some feet above the surface of the water, and is prevented from drifting by the sea-anchor float until recovered by the ship which has followed it. The balloons take about an hour to ascend and an hour to descend. In order to minimize the chance of losing the apparatus, and in cases where sea room is limited, a device can be attached by which the upper balloon is released at a pre-arranged height. The device consists of an aneroid box which expands as the pressure on it decreases with height, and by so doing releases a catch which holds the upper balloon. The apparatus is shown in Fig. 26.

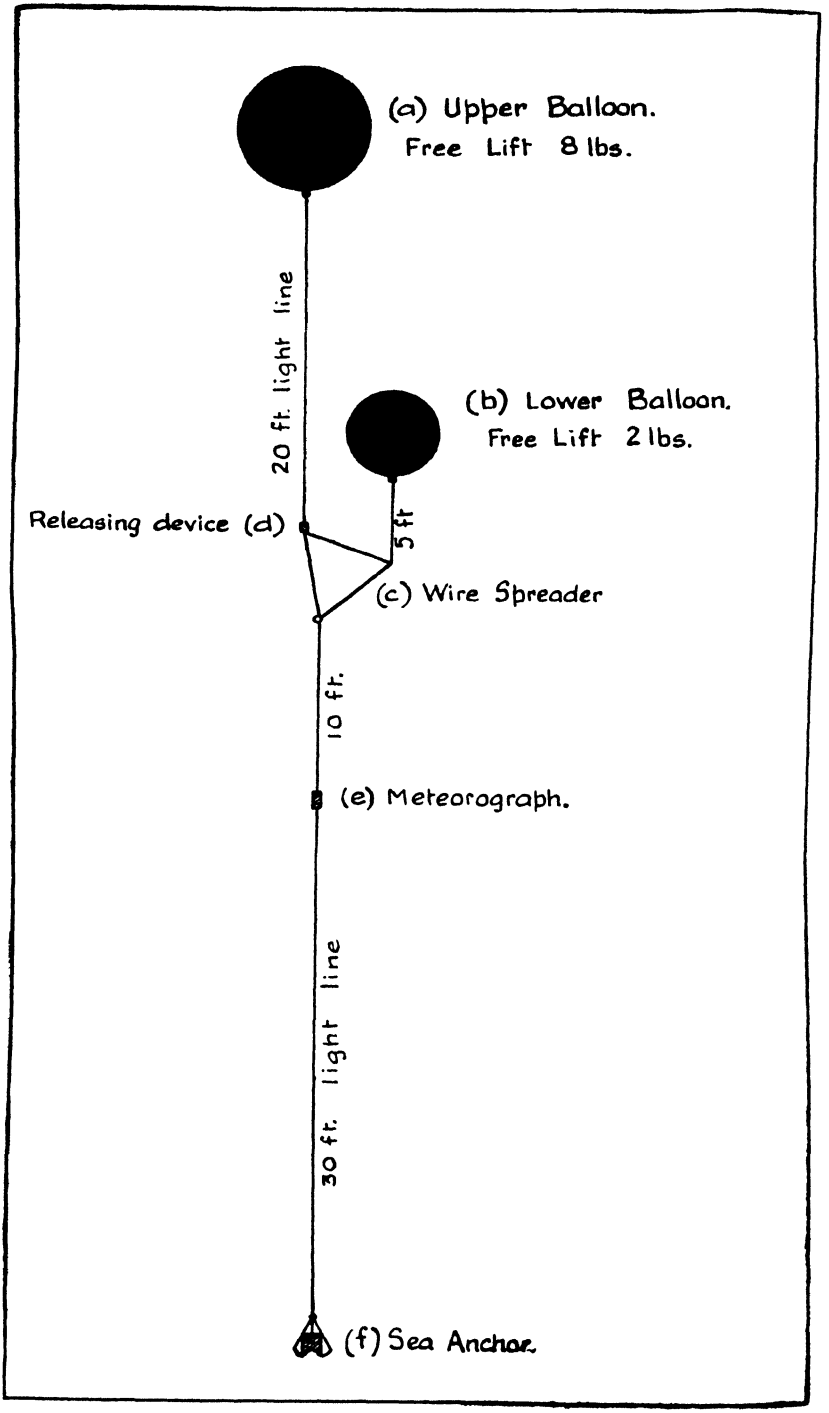


FIG. 26.—BALLOON AND SEA METEOROGRAPH.

The upper balloon (*a*) is inflated with hydrogen to give a free lift of 8 lbs. (3·63 kg.): its diameter will then be about 7 feet. The lower balloon (*b*) is inflated to give a free lift of 2 lbs. (0·91 kg.). The balloons are attached to a wire spreader (*c*) with a releasing device (*d*) (when used), set to release the upper balloon at 30,000 feet. The sea-anchor float (*f*) consists of a cylinder-shaped wicker frame covered with canvas, three feet in circumference. The canvas is perforated with holes, and is attached to the balloon by three lines passing through beckets on the upper part of the cylinder, thus making it collapsible during ascent. When in the sea, the base fills with water and acts as a sea-anchor, the cylinder preventing it from sinking. Its weight is 3¼ lbs.

The self-recording instrument is a *Dines Meteorograph* and is attached to the line securing the sea anchor to the spreader. The instrument is light and simple, its weight being about 2 ozs.

Fig. 27 is a photograph of the instrument which consists of an ordinary aneroid box (*aa*) which, being only partially exhausted of air, expands under reduced pressure as the apparatus ascends, and the two parts of the frame (*bb*) which are held together by the spring (*cc*) move apart. One side carries the plate (*dd*), on which is already engraved a small pressure and temperature diagram, and the other the scratching point (*ee*) so that when they move a scratch is made on the diagram. Changes in temperature cause a movement perpendicular to this scratch, and are recorded by the expansion and contraction of the slip of german silver (*ff*). The instrument is protected by an aluminium cylinder. When a record has been obtained, the plate (*dd*) is removed and forwarded to the Meteorological Office where the record is worked out.

Coding of Reports of Upper-Air Temperatures, etc.

When observations are made from aircraft operating with H.M. Ships at sea and reports are transmitted by W/T, the following code should be employed:—

(*a*) Observations made in conjunction with an aeroplane aneroid, graduated in millibars.

THUM YQLLL IIIGG $P_0P_0P_0P_0P_0$ $T_0T_0T_0W_0W_0W_0$ $T_1T_1T_1W_1W_1W_1$ —— $T_{10}T_{10}T_{10}W_{10}W_{10}W_{10}$ $P_{10}P_{10}P_{10}$

(*b*) Observations made in conjunction with an altimeter graduated in feet.

THUM FEET YQLLL IIIGG $P_0P_0P_0P_0P_0$ $T_0T_0T_0W_0W_0W_0$ $T_1T_1T_1W_1W_1W_1$—— $T_{10}T_{10}T_{10}W_{10}W_{10}W_{10}$ XXXXX

Individual elements in above reports.

THUM is the index word indicating that the message gives upper-air temperatures at intervals of 50 mb.

To face page 96.

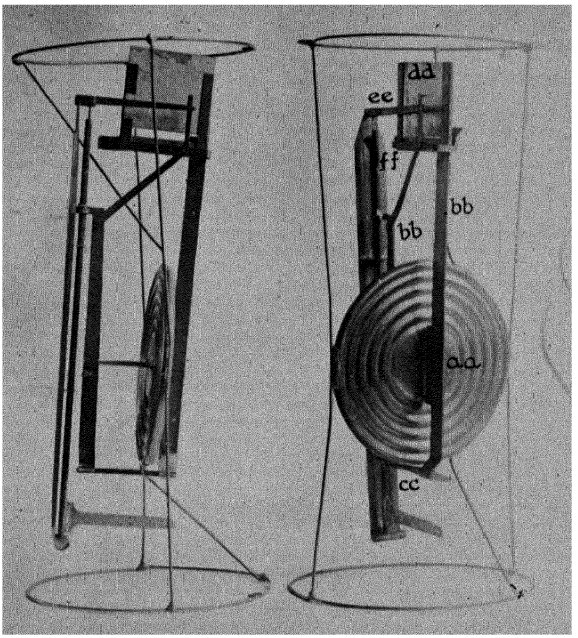


FIG. 27.—DINES METEOROGRAPH.

YQLLL IIIGG *see* Chapter II, page 52.

$P_0P_0P_0P_0P_0$. This gives the mean sea-level pressure by mercury barometer at the beginning of the flight, in millibars and tenths of a millibar (*e.g.*, 1,019·5 mb. is reported as 10195 and 997·5 mb. is reported as 09975).

TTT. This is the dry-bulb temperature in degrees C. or F. to tenths of a degree and WWW the corresponding wet-bulb. The suffixes 0, 1, 2, —9, 10 refer to sea level, 1,000 feet, 950 mb., 900 mb.—600 mb. 550 mb. respectively. $P_{10}P_{10}P_{10}$ is normally the aneroid reading at the tenth observation (550 mb.) after leaving the surface, but if a greater or less height is reached the aneroid reading at the highest level where a temperature reading is taken, is inserted instead. This final aneroid reading is included in the message as a check. In order that some reading may be obtained between the surface and the 950 mb. level under all conditions, the 1,000 feet level is incorporated in the message : the 1,000 mb. level cannot be used as there are occasions when the surface level pressure is below 1,000 mb. A temperature of 3·6° is coded as 036 ; 0·2° as 002, etc. For negative temperatures add 50, *e.g.*, —3·2° is coded as 532. If the reading is made to the nearest whole degree the last figure should be 0, *e.g.*, 40° is coded as 400. The temperature at surface level should be that obtained from a screen thermometer or electric psychrometer.

THUM FEET : when these index words are used in the message they indicate that the message gives surface pressure and temperature, and upper-air temperatures at heights of 1,000 feet, 2,000 feet, 4,000 feet and so on at intervals of 2,000 feet. The altimeter reading at which the final temperature observation ($T_{10}T_{10}T_{10}W_{10}W_{10}W_{10}$) is made, is added to the message as a check (XXXXX). It is important that the altimeter should be set to read zero at sea-level immediately before the flight.

Reports of inversions, clouds, visibility.

Inversion groups.—These should be sent after the usual groups in the form PPPTTT $P^1P^1P^1T^1T^1T^1$ where PPP is the pressure in millibars and TTT the temperature at the base of the inversion to the nearest tenth of a degree. $P^1P^1P^1T^1T^1T^1$ are the corresponding figures at the top of the inversion, *i.e.*, at the point where the temperature ceases to rise. The word " inversion " should precede the groups.

When observations are made in conjunction with an altimeter the inversion groups are coded giving the altimeter heights to the nearest hundred feet, the first figure being an 0 for heights below 10,000 feet. If preferred the inversion report may be given *en clair*, *e.g.*, inversion 36 to 39 from 4,600 to 5,300 feet.

Cloud groups.—After the group already enumerated, a short note should be inserted giving a description of the clouds and the

pressures at which they occur (the cloud height is deduced at headquarters from the pressure and temperature). The pressure at the level of the lowest clouds should always be given, and if these clouds are detached the observer must note carefully at what pressure he is level with the base of the clouds. In sending the messages the cloud names may be abbreviated, *e.g.*, ci., st., ast., stcu., etc. Typical descriptions of complex cloud arrangements would be: (1) nearly continuous layer 870 to 830 mb., irregular clouds below, lowest at 980 mb., broken layer 640 to 630 mb. and ast. from 600 mb. upwards. (2) Cu. and cunb., base 920 mb., highest tops well above 500 mb., with false ci. Broken layer about 800 to 750 mb., and thin patches about 620 mb. (If the tops of cunb. are very high, they may be estimated as above 500 or 400 mb., even if the ascent only reaches 600 mb.). With an altimeter, cloud heights in feet should be given.

Visibility groups.—These should follow the cloud groups and consist of 2, 4 or 6 figures according to the height reached. The group should be preceded by the word "visib." The visibility should be observed when the aneroid reads 800, 700 and 600 mb., and observations should be taken looking downwards and also horizontally. The visibility should as far as possible refer to haze only, not to cloud. If the clouds are continuous the downward visibility is zero and should be reported by the letter "X" but if there is a gap the visibility should be judged from the clearness of the sea through the gap. In judging the horizontal visibility when there are no distant clouds at the same level, and the haze has not a sharply defined top, care must be taken that the observation is made horizontally and not slightly downwards.

Visibility reports should be sent in the following code:—

0 Very bad.
1 Poor.
2 Indifferent.
3 Good.
4 Very good.

Symbolically, the report is made:—visib. $V_8V_8V_7V_7V_6V_6$, where V_8V_8 represent the downward and horizontal visibility (in that order) at 800 mb., V_7V_7 the downward and horizontal visibility at 700 mb. and so on.

A note should be added if the haze has a sharply defined top, or if there is undoubtedly a haze high up (which is not common). When using an altimeter, visibility should be observed at 6,000, 10,000 and 14,000 feet.

Specimen messages.

(i) Using aeroplane aneroid.

Friday 16th August ship was in position 49°20'N., 8°12'W.

Flight commenced 0715 G.M.T. The following observations were made :—

Height or Pressure.	Dry Bulb.	Wet Bulb.	Visibility.		Clouds, etc.
			Down.	Horizontally.	
Surface (Bar. 1,014·6 mb.)	67·3	61·3	—	—	Continuous Stcu. 925 to 875 mb. Inversion 47·3 to 55·8, 875 to 860 mb.
1,000 feet	65·7	57·6	—	—	
950 mb.	59·4	55·4	—	—	
900	50·2	50·1	—	—	
850	55·5	38·7	—	—	
800	53·0	37·1	0	3	
750	45·8	32·4	—	—	
700	38·3	24·2	0	4	
670	33·7	—	—	—	

This should be coded :—Weather London (or appropriate telegraphic address).
THUM 60493 08207 10146 673613 657576 594554 502501 555387 530371 458324 383242 337XXX 670mb. Inversion 875473 860558 Clouds continuous Stcu. 925 mb. to 875 mb. visib. 0304.

(ii) Using altimeter.

Friday 25th March ship was in position 53°04′N., 0°52′E Flight commenced 0750 G.M.T. The following observations were, made :—

Height.	Dry Bulb.	Wet Bulb.	Visibility.		Clouds, etc.
			Down.	Horizontally.	
Surface (Bar. 1,008·2 mb.)	45·3	44·0	—	—	
1,000 feet	44·0	42·2	—	—	A few Frcu. base 1,500 tops up to Stcu. in places. Broken Stcu 3,800 to 4,500 with inversion 30·3° to 40·0° 4,500 to 4,900 ft. Patches Acu. 13,200 feet with inversion 11·0 to 14·5 13,200 to 13,800 feet.
2,000	39·5	38·5	—	—	
4,000	32·3	32·0	—	—	
6,000	36·5	27·3	2	4	
8,000	29·0	26·0	—	—	
10,000	21·7	20·2	2	4	
12,000	14·3	—	—	—	
14,000	12·5	—	2	3	
15,000	4·0	—	—	—	

This should be coded :—

THUM FEET 63531 00908 10082 453440 440422 395385 323320 365273 290260 217202 143XXX 125XXX 040XXX 15000 Inversion 045303 049400 also 132110 138145 Clouds broken Stcu. 3800 to 4500 ft. also Frcu. base 1500 some tops reaching Stcu. level patches Acu. 13200 visib. 242423.

SECTION II

GENERAL METEOROLOGY

CHAPTER V

The Atmosphere—Radiation

Introductory.

The earth is surrounded by a gaseous envelope called the atmosphere. Meteorology treats of those physical properties of the atmosphere whose changes are manifest as weather and climate, and which exercise a profound influence upon the life and activities of man. Among these properties are atmospheric temperature and movement (wind), two factors which are directly experienced and upon which the practicability and effectiveness of naval, military and air operations are often dependent. Another property, the condition of the water vapour in the atmosphere has an influence which is even more potent, for to it are due rain and other forms of precipitation, fog and cloud. This factor is itself, however, to some extent controlled by temperature. Finally, the variations of atmospheric pressure, although unperceived without the aid of instruments, are so closely related to wind and weather, that the distribution of pressure is the basis upon which weather maps used for forecasting purposes are studied and classified.

General meteorology deals firstly with the three fundamental atmospheric properties, temperature, pressure and humidity (water-vapour content) and secondly with how variations of these properties with time and place are effective in producing weather and climate. Temperature, however, is largely controlled by solar and terrestrial radiation, and it will be necessary as a preliminary briefly to consider these factors.

Composition and Extent of the Atmosphere

The atmosphere consists of a mechanical mixture of a number of gases called the permanent gases of the atmosphere, of which the principal two are oxygen and nitrogen. Near the earth's surface the proportions by volume to the total volume of the permanent gases are oxygen 21 per cent., and nitrogen 78 per cent., the remaining 1 per cent. consisting of carbon dioxide, hydrogen and the inert gases, argon, helium, neon, krypton and xenon. That the air is a mixture of these gases can be shown by the absorption of one or more of its constituents by means of chemicals, leaving the proportions of the remaining constituents unaltered. The amounts of oxygen and carbon dioxide vary slightly with time and place, since these gases are involved in the life processes of the animal and vegetable worlds. The proportions of the other gases remain remarkably constant in the lower layers of the atmosphere up to an average height of 10 kilometres. In this region the temperature decreases with height, which, as will be seen later, gives rise to

convection or vertical currents in the air and consequently the lower layers become well mixed and homogeneous in composition. Above this region the temperature remains practically constant through a large range of height. Theoretical considerations indicate that under this condition each of the constituent gases would obey the law of Dalton, that is, as regards its distribution with height it would behave as if all the other gases were absent. The heavier gases would be present in comparatively large amounts in the lower layers of this region, but their proportions would fall off rapidly with height; the lighter gases would be present in comparatively small amounts in the lower layers, but their amounts would decrease slowly with height. The lower layers would therefore consist mainly of the heavier gases, oxygen and nitrogen; the upper layers mainly of the lighter gases, hydrogen and helium. There is some slight evidence that this settling out does actually occur. On the other hand, the latest experimental evidence, from observations of the spectra of the aurora and meteors, indicates that oxygen and nitrogen predominate to the outer regions of the atmosphere.

In addition to the permanent gases, the atmosphere contains a number of variable constituents, the principal of which is water vapour. Its amount varies up to 4 per cent. by volume, and is constantly changing. It is increased by evaporation from the surface of the sea, rivers and lakes, and from plant and animal life. As vapour it is invisible, but on condensation it gives rise to cloud, fog, rain, snow, dew, hoar frost—in fact to most of the phenomena connected with weather. Owing to the general decrease of temperature with height, it is present in its greatest amounts in the lowest layers, but it is carried up by vertical currents to the limits of convection, above which there is diffusion, but the quantities of water vapour are very small.

There are also present solid constituents in the form of extremely minute particles of dust, smoke and salt from sea spray. These are the centres or nuclei on which water vapour condenses in the form of droplets in cloud or fog—without them condensation would not readily take place. They are of special importance in relation to fog. They occur mostly in the surface layers of the atmosphere, but some are always present even at very high levels. It is impossible to state the height of the atmosphere, but observations show that there is sufficient gas present at 200 kilometers to cause the volatilization of meteors and at 1,000 kilometers to give rise to the aurora. But if the whole atmosphere were compressed to the same density as that at the surface, its height would be only 8 kilometres. This is called the height of the homogeneous atmosphere.

Radiation.

The warming of the earth's surface, the evaporation of water vapour into the atmosphere and the movement of the atmosphere

as wind, all necessitate a constant supply of energy from some outside source. Practically the whole of the energy received reaches the earth from the sun in the form of radiation. Other possible sources of energy might be considered, but their contributions are extremely minute. It is known, for example, that the interior of the earth is very hot, and therefore a constant flow of heat must take place from the interior to the surface. It has been calculated, however, that the heat energy thus reaching the earth's surface from below is sufficient only to raise its temperature by 0·1° C. Similarly, the radiant energy received from the moon and the stars is extremely small; indeed, apart from the sun, all other possible sources of energy make only negligible contributions to the earth's supply.

Radiant energy can take a number of apparently different forms, but fundamentally these forms are the same, consisting of electromagnetic waves, and the differences of form are due to differences of wave-length. Wireless waves, heat rays and light are some of the well-known forms in which radiation is manifest. Wireless waves are the longest, having wave-lengths of a few metres and upwards. Light and heat waves are very much shorter, and to specify their wave-lengths a special unit known as the micron, denoted by the Greek letter μ and equal to 10^{-4} centimetres is used. Rays of wave-length between 3 μ and 80 μ (approximately) are termed heat rays. If the wave-length lies between 0·4 μ and 0·7 μ, the radiation becomes visible; these are the light rays. The colours of the spectrum range between these limits of wave-length, that of the violet being 0·4 μ and of the red 0·7 μ. Infra-red rays lying beyond the red of the visible spectrum, having wave-lengths somewhat greater than 0·7 μ, have the power of penetrating mist and thick haze, and are used with specially prepared plates for long-range photography under conditions of bad visibility. Ultra-violet rays lie beyond the visible spectrum at the violet end, and an ordinary photographic plate is sensitive to them. Apart from the effect of refraction, radiation is propagated in straight lines. The intensity of radiation is measured by the amount of energy which crosses an area perpendicular to its direction of propagation in a given time. In the C.G.S. system the unit of intensity is one erg per square centimetre per second. This is too small a unit for meteorological work in which one gram-calorie per square centimetre per minute is used.

Radiant energy is constantly emitted by all bodies, the intensity depending upon the temperature and the nature of the surface of the emitting body. Experiment shows that there is an upper limit to the intensity of radiation that can be emitted by a body at a given temperature. A body which for every wave-length emits the maximum intensity of radiation corresponding to its temperature is called a **Perfect Radiator.**

A perfect radiator is also called a **Black Body** and the radiation

it emits, **Black Body Radiation.** There are no perfect radiators among known substances. Bodies which emit a large proportion of the black-body radiation corresponding to their temperature are termed good radiators ; bodies which emit only a small proportion of the black-body radiation are termed bad radiators.

At a given temperature, even a black body does not emit radiation of all wave-lengths with the same intensity. There is a certain wave-length for which the intensity is a maximum, and practically the whole of the energy is distributed over a comparatively small range of wave-lengths extending on either side of this maximum. The distribution of energy in the scale of wave-lengths emitted by a black body may be represented by a curve *PQR* (Fig. 28). Wave-length is represented along the horizontal axis

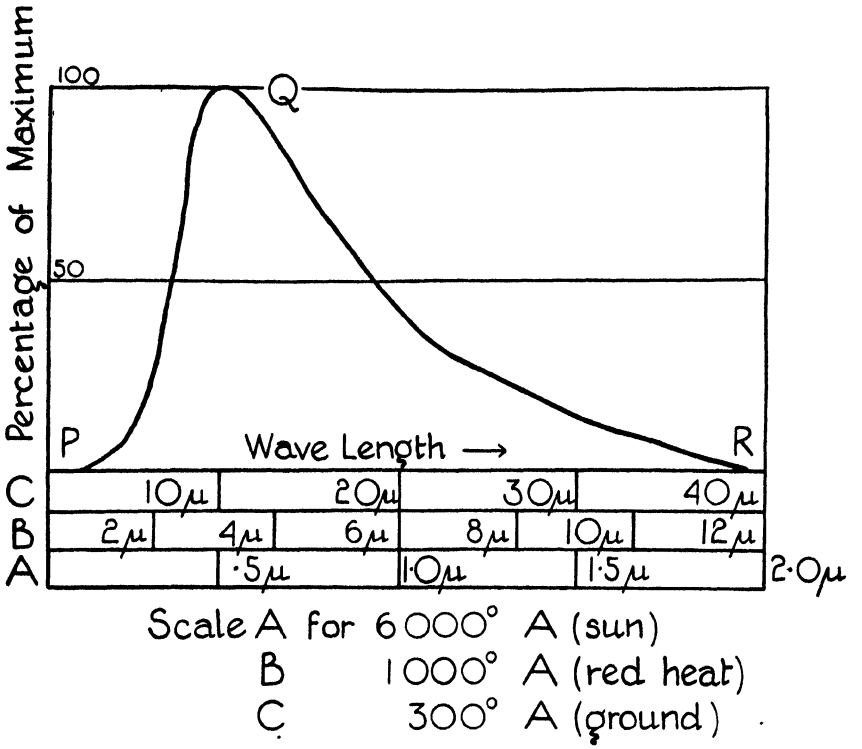


FIG. 28.—INTENSITY OF RADIATION AND WAVE-LENGTH.

and intensity of radiation, expressed as a percentage of the maximum along the vertical axis. At the point *Q*, corresponding to the wave-length of maximum intensity the relative intensity is 100 per cent. Practically the whole of the radiant energy is confined to the range of wave-lengths *PR*. The curve *PQR* has the same form whatever be the temperature of the radiating black body, but the wave-length at which the maximum intensity *Q* occurs

varies inversely with the temperature. The same curve *PQR* may be utilized for all temperatures, provided that the wave-length scale is suitably adjusted. In Fig. 28 three scales of wave-length are provided; *A* corresponding to the temperature of the sun (6,000° A.), *B* red heat (1,000° A.), and *C* the average temperature of the earth's surface (300° A.). The wave-lengths of maximum intensity are seen to be approximately 0·5 μ, 3 μ, and 10 μ, respectively. In general, the wave-length λ_m in microns, of maximum intensity for a given (Absolute) temperature T is given by Wien's Law, $\lambda_m = 2940/T$. Conversely, if the wave-length of maximum intensity can be found experimentally, the temperature of the emitting body can be found, provided that the body emits as a black body. In the case of the sun there is independent evidence that this condition is fulfilled, and its temperature is accordingly determined as 6,000° A. (approx.). The wave-length of maximum intensity of the sun's radiation falls within the visible spectrum, and about half the emitted energy is in the form of light. For a body at red heat, the range of wave-lengths *PR* slightly overlaps the visible spectrum at the red end, and the body therefore appears dull red in colour. As the temperature of the body increases the range of emitted wave-lengths *PR* shifts towards the short-wave end of the spectrum, ultimately embracing the whole of it, and the body passes through the successive stages of red and white heat. At ordinary temperatures (300° A.) the emitted radiation is invisible; nevertheless, the fact must not be overlooked that the earth's surface is constantly emitting radiation of the appropriate wave-lengths, and this outgoing radiation is as important in determining the earth's surface temperature as the incoming radiation from the sun.

The rate at which a body emits radiation per unit area of its surface is called its **Emissive Power.** It clearly varies with temperature and wave-length.

Absorption.

If radiation of any wave-length falls upon a black body it is completely absorbed and the whole of the energy becomes manifest as heat, raising the temperature of the absorbing body. A black body may, therefore, be called a **Perfect Absorber.** Only part of the energy is absorbed by an ordinary body, the remainder being reflected at its surface or transmitted through it. In the latter case, the body is said to be partially transparent to the radiation falling upon it. A body's **Absorptive Power** is defined in the same way as its emissive power, that is the rate of absorption per unit area of the energy falling upon it. The ratio of the amount of radiation absorbed to that incident is called its Absorption Coefficient. For a given body the absorption coefficient is equal to the Emission Coefficient. It follows that if a body at a given temperature emits radiation of a certain wave-length, it will also absorb radiation of that wave-length. A good radiator is therefore

a good absorber, and *vice versa*. A polished surface, since it reflects most of the light falling upon it, is a poor absorber from which it follows that when raised to a temperature sufficiently high to emit waves within the visible spectrum, it is also a poor radiator of those wave-lengths. This is true, however, only if the high temperature does not destroy the nature of the surface.

Many substances have the power of absorbing only radiation of certain wave-lengths and of transmitting radiation of other wave-lengths unchanged through them. They are then said to be transparent to the wave-lengths they transmit and opaque to those which they absorb. Ordinary glass is transparent to light, but opaque to heat waves. It is therefore used for greenhouses, admitting the incoming radiation from the sun, but largely preventing the loss of heat by radiation from the earth within, although the fact that the glass prevents the loss of warm air also plays a part.

Reflection and Scattering.

When radiation falls upon a highly-polished surface of a non-absorbing body, it is reflected in a particular direction, depending upon the angle at which it strikes the surface. If the polished surface is flat, reflection changes the direction of all the rays without altering the angle between any two of them, and if the radiation is in the form of light, the observer sees a reflected image of the object viewed. If, however, the surface is unpolished, each ray is reflected independently of the remainder by the small portion of surface which it strikes ; consequently the rays are reflected in all directions at random, and the observer sees only a more or less uniformly illuminated surface. This phenomenon is known as diffuse reflection. In general, objects are rendered visible by diffuse reflection of the light falling upon them.

When a beam of light passes through a transparent medium containing very small particles in suspension, a process similar in its effects to diffuse reflection takes place. Some of the light from the beam is sent out or "scattered" in all directions, and the particles are rendered visible by the scattered light that reaches the eye of the observer, although they are too small to be seen directly. True scattering takes place only if the diameter of the particles is less than the wave-length of the incident beam. The shorter the wave-length the greater is the amount of scattering, and for white light the blue is therefore scattered to a much greater extent than the red. A cloud of tobacco smoke, consisting of minute liquid drops of diameter $0{\cdot}2\ \mu$ (approx.) is rendered visible owing to the scattering of light of all wave-lengths, but since the blue is scattered more than the red the cloud appears blue in colour. The blue of the sky is due to the scattering of white light by the individual molecules of the atmosphere.

Solar Radiation. Insolation.

In its passage through the atmosphere the solar beam suffers losses by absorption due to various atmospheric constituents, by scattering due to solid and liquid particles in the air and by reflection at the surfaces of clouds. In determining the intensity of solar radiation by experiments at the earth's surface, account must be taken of these losses. They are difficult to measure and to estimate, and therefore the best determinations of the intensity of solar radiation are those made on high mountains where, owing to the reduced thickness of atmosphere to be traversed, losses are at a minimum. Allowance must be made for the residual losses in the upper regions of the atmosphere. When this is done, the result obtained is the intensity of the solar beam at the outer " boundary " of the atmosphere. The intensity is found to be very nearly constant and equal to 1·94 gram-calories per square centimetre per minute. An alternative value is 135 kilowatts per square dekametre. This is called the **Solar Constant.**

The solar radiation received at the earth's surface is called **Insolation.** For measuring its intensity a Pyrheliometer is used. One of the simplest forms of instrument is Angstrom's pyrheliometer, in which the insolation is received on a strip of platinum coated with lamp-black and exposed normally to the sun's rays. The lamp-black is an almost perfect absorber and as a result of the absorption the temperature of the platinum strip is raised. An electric current is passed through a second platinum strip similar to the first except that it is shielded from the sun's rays. The two platinum strips are brought to the same temperature by adjusting the strength of the current, which supplies energy to the second platinum strip at the same rate as the solar radiation supplies energy to the first. The strength of the current therefore gives a measure of the intensity of insolation.

The marked differences in intensity of insolation are due to a number of factors which are here considered. The first is the inclination of the direction of the beam to the earth's surface. In Fig. 29, let AB represent a portion of the earth's surface and the shaded belts two beams of solar radiation of the same cross section and of the same intensity, but at different inclinations to the surface. The amount of energy is the same for both beams but the areas over which they are spread, represented by PQ and RS clearly differ. If the angle between the direction of the beam and the normal to the surface is called the angle of incidence, the larger the angle of incidence the larger is the area over which the beam is spread and the smaller is its heating effect. If I is the intensity of the solar beam and θ its angle of incidence, the rate at which energy falls on the surface, per unit area, is $I \cos \theta$. The angle of incidence varies firstly on account of latitude, being least in the tropics and greatest at the poles, secondly on account of the daily passage of the sun across the sky and thirdly with the

seasonal variation of the sun's declination. The solar rays at large angles of incidence also suffer the greatest losses due to absorption and scattering on account of the greater thickness of atmosphere which they traverse.

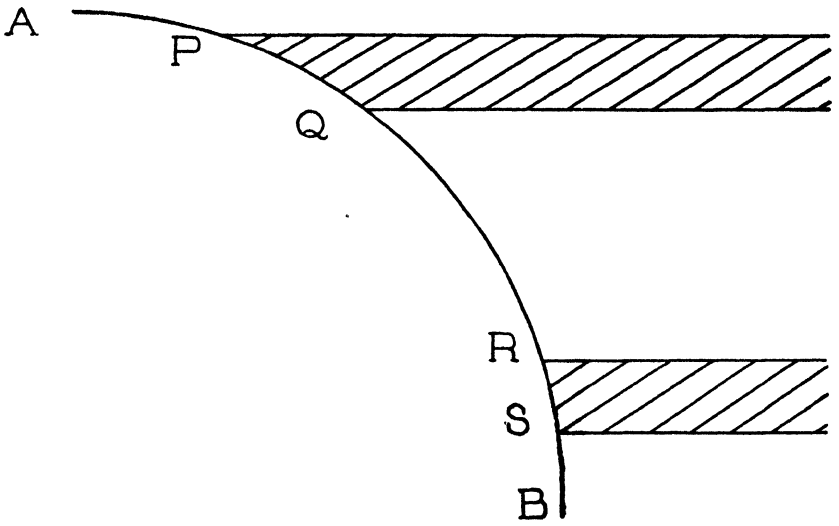


FIG. 29.—INTENSITY OF RADIATION WITH ANGLE OF INCIDENCE.

The second factor on which the heating effect of insolation depends is the nature of the surface upon which it falls. Water is moderately transparent to the sun's rays and is penetrated by them to a considerable depth so that their heating effect is distributed through a comparatively large volume of water and is therefore small. Moreover, a certain amount of the radiation is reflected by a water surface, the amount being large at large angles of incidence. Soil and rock, on the other hand, are good absorbers but poor heat conductors. The changes of temperature within them are therefore confined to thin layers next the absorbing surface and are correspondingly large. Snow- and ice-covered surfaces reflect about 80 per cent. of the incident radiation, so that over large areas in the polar regions only about one fifth of the energy of insolation, already depleted owing to the large angle of incidence, is available for heating purposes.

HISTORY OF RADIATION REACHING THE EARTH.

Since all bodies emit and absorb radiation which is propagated in straight lines, any body is affected by all other bodies "within sight." The rate of emission from a given body depends upon its own surface temperature but not directly upon the presence of other bodies. These have an indirect effect, however, in that the radiation they themselves emit is partly absorbed by the given

body, raising its temperature and therefore increasing its rate of emission. The changes of temperature experienced by a body depend upon the difference between the energy it emits and absorbs; only when these are the same does the body remain at a steady temperature. The temperature of a portion of the earth's surface is controlled in this way. It emits radiation constantly, but during the day receives more radiation than it emits and its temperature rises. At night it receives no solar radiation to make good the loss by emission and its temperature falls. The outgoing radiation from the earth is ultimately lost to space. Since the mean temperature at the surface of the earth does not appear to be changing it is concluded that the whole of the incoming solar radiation is re-radiated to space, so that on balance there is no loss or gain of heat at the earth.

The following brief account explains qualitatively the history of radiation at the earth from the time it reaches the outer limits of the atmosphere to its final loss to space. The figures given, which are approximate only, represent on a percentage basis relative to the original solar beam, the amounts of energy involved in each process. They apply to average conditions of cloudiness and nature of the surface of the earth as a whole.

In Fig. 30 let the inner circle represent a portion of the earth's surface and the outer circle the upper limit of the atmosphere. The pecked lines indicate short-wave (solar) radiation, the full lines long-wave (terrestrial) radiation. The incoming beam *A* (100 per cent.) traverses the atmosphere and suffers losses by absorption, reflection and scattering. Absorption is small, amounting only to about 7 per cent. of the radiation. The absorbed energy is represented by *C* (7 per cent.) and it is utilized in raising the temperature of the atmosphere by a small amount. Scattering and reflection at clouds and at the earth's surface send back a proportion of the radiation denoted by *B* (43 per cent.) to space. This radiation would enable the earth to be seen by observers on another planet and its amount, 43 per cent., would determine its brightness when so viewed. The latter figure is known as the earth's albedo. Some of the reflected and scattered radiation passes on to the earth's surface and combined with the direct radiation gives *D* (50 per cent.) which is absorbed and utilized in raising the earth's surface temperature.

The earth gives out its heat firstly in the form of long-wave radiation and secondly through warming of the atmosphere by contact. Now the long-wave radiation is readily absorbed by water vapour, consequently if there is appreciable water vapour present in the atmosphere the long-wave radiation is prevented from passing out directly to space, and is utilized in raising the temperature of the atmosphere. In Fig. 30, *E* represents the outgoing radiation, some of which, *G*, passes directly to space and the remainder, *F*, is absorbed by the water vapour in the atmosphere.

H represents the heat energy acquired by the atmosphere by contact with the warmed surface of the earth. Some of this energy takes the form of latent heat absorbed in evaporating water from the

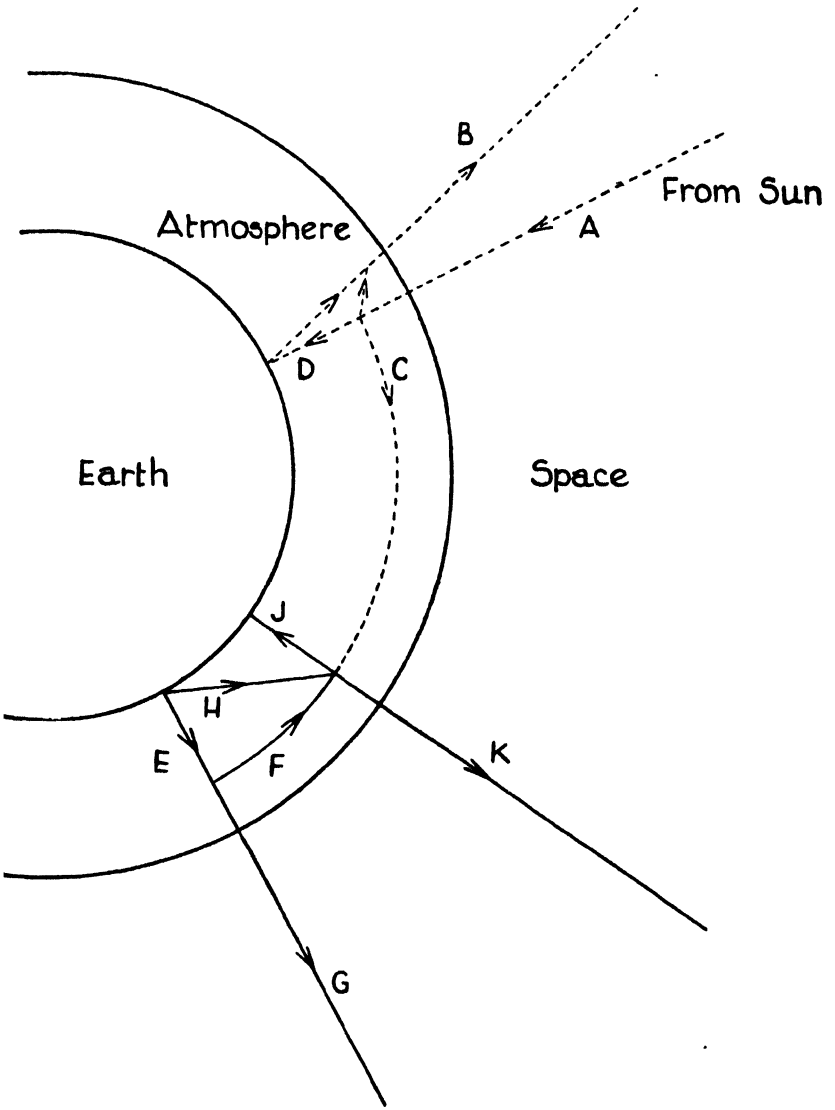


FIG. 30.—SOLAR AND TERRESTRIAL RADIATION.

surface. The temperature of the atmosphere is therefore raised by absorption of short-wave radiation, *C*, of long-wave radiation, *F*, and by " contact " heating, *H*. The atmosphere itself re-radiates *J* to the earth's surface and *K* to space. The total incoming radiation *A* is balanced by the outgoing radiation $B + G + K$. The

amounts of E, F, H, and J vary considerably and it is impossible to assign to them average values.

The solar radiation received at the earth's surface consists of the direct beam and some of the reflected and scattered rays. The former may be called sun radiation, the latter, sky radiation. Sky radiation may be temporarily increased over a limited area by reflection, from clouds, of radiation which would otherwise have been lost to space or have fallen on an adjacent part of the earth's surface.

An overcast sky reflects back to space about 70 per cent. of the incident radiation and so by day considerably reduces the energy received at the ground. At night a cloud radiates more heat downwards than does a clear sky and therefore, to some extent, compensates for the loss of heat at the earth's surface by terrestrial radiation. Water vapour permits the full passage of solar radiation but absorbs and radiates back much of the terrestrial radiation. This is known as the " greenhouse effect " of the atmosphere.

CHAPTER VI.

Air Temperature, Humidity and Pressure at the Surface

Temperature.

Temperature Scales.—For meteorological purposes temperature is generally measured by means of a mercury or alcohol thermometer. Two scales of temperature, the Fahrenheit and Centigrade scales are in common use, the former in English-speaking countries, the latter in most other countries. A third scale, the Réaumur, is now practically obsolete. To specify a scale of temperature, two fixed points are required, and the interval between them is uniformly divided into a number of degrees. The fixed points are the same for all scales and are called the **Freezing Point** and **Boiling Point** respectively. Freezing point is the temperature of a mixture of pure water and melting ice under standard conditions, *i.e.*, pressure 760 millimetres of mercury at mean sea level in latitude 45°. Boiling point is the temperature of pure boiling water under the same conditions. On the three temperature scales, the fixed points and the divisions of the interval between them are as follows :—

Scale.	Freezing Point.	Boiling Point.	Interval between Fixed Points.
Fahrenheit	32°	212°	180°
Centigrade	0°	100°	100°
Réaumur	0°	80°	80°

Since 180 degrees on the Fahrenheit scale are equivalent to 100 degrees on the Centigrade scale, the magnitude of the Fahrenheit degree is 5/9 that of the Centigrade degree. Since also the freezing point occurs at 32° and 0°, respectively, the relation between the two scales is given by :—

$$F. = 9/5\ C. + 32.$$
$$\text{or } C. = 5/9\ (F. - 32).$$

where F. denotes a given temperature on the Fahrenheit scale and C. the same temperature on the Centigrade scale.

With the increase in upper-air soundings, the use of these scales frequently involves negative temperatures with resulting inconvenience in calculation in dealing with physical problems in meteorology. The Absolute scale of temperature has therefore

come into use. Absolute temperature is physically significant, since various properties of a gas such as its volume under constant pressure and its pressure at constant volume are directly proportional to it. The fixed points on the Absolute scale are :—

Freezing Point, 273°. Boiling Point, 373°.

The interval being 100°, the degree is of the same magnitude as on the Centigrade scale. Absolute temperatures are expressed thus : 373° A., it being then understood that the degrees are of Centigrade magnitude. Absolute zero occurs at — 273° C. Modern physics shows that this is the lowest temperature at which matter can exist, so that all temperatures on the Absolute scale must be positive.

On the Fahrenheit scale, Absolute zero occurs at — 459° F. If Absolute temperatures are expressed in Fahrenheit degrees from this zero, the fact should be explicitly stated.

Measurement of Temperature.—The methods and instruments employed for the measurement of temperature on board ship are described in Chapter I, and the necessity for screening and adequate ventilation of the instruments in order to obtain accurate measurements of the true air temperature are there pointed out. Methods similar in principle to these are employed at land stations, but there are certain differences of detail, owing to the fact that on land the thermometers may be exposed permanently in one position. In most countries, suitably screened mercury thermometers are used, although in cold climates spirit thermometers are necessary. In the British Isles and in most countries of the Empire, the thermometers are exposed in a Stevenson screen. This is a wooden box, painted white, with double-louvred sides, mounted on a stand so that the thermometers in position inside it are 4 ft. above the ground. The screen opens towards the North (South in the southern hemisphere), so that the thermometers shall be affected as little as possible by the direct rays of the sun while they are being read. Normally, a maximum and minimum thermometer are housed in the screen together with the dry- and wet-bulb thermometers.

For obtaining a continuous record of temperature, a bi-metallic thermograph is housed in the Stevenson screen (large pattern). The sensitive element and the recorder are combined in one instrument. The sensitive element consists of two strips of different metals welded together along their flat surfaces and coiled into the form of a flat spring. This replaces and occupies a similar position to the Bourdon tube of the distant reading thermograph. One end of the spring is fixed to the base of the instrument, the other is connected to a pen-arm and pen, which traces a record on a chart wrapped round a revolving drum driven by clockwork. Changes of temperature cause the two metals of the spring to expand or contract by different amounts, so that the spring coils and uncoils and the pen traces a record of temperature on the chart.

Maximum and Minimum Thermometers.—Ordinary thermo-

meters suitably modified are used to measure the maximum and minimum temperatures reached during a given period.

The Maximum Thermometer is a mercury thermometer, and should be graduated to read from 0° F. to 130° F. in temperate climates. About 1 in. from the bulb there is a small constriction in the bore of the stem. As the temperature rises, mercury is forced past the constriction and the thread of mercury advances along the stem, as with an ordinary thermometer. As soon as the temperature begins to fall, however, the mercury in the bulb contracts, but the thread of mercury breaks at the constriction, leaving its upper end in position at the maximum temperature reached. The maximum thermometer is mounted in a position slightly inclined to the horizontal with the bulb lowermost.

The maximum thermometer is read at the morning and evening observations, and after reading must be re-set. Setting is performed by whirling the thermometer at arm's length with the bulb outwards, the centrifugal force causing the mercury to pass the constriction until the reading is the same as that of the ordinary thermometer.

The Minimum Thermometer is an alcohol thermometer and should be graduated to read from — 30° F. to 100° F. in temperate climates. The most common type of instrument is a spirit thermometer having a small index immersed in the spirit in the stem. As the temperature falls the index is carried towards the bulb by the receding spirit, but if the spirit subsequently expands, in consequence of a rise in temperature, it flows past the index, which is left in position to indicate the lowest temperature reached. To set the instrument, it is tilted with the bulb uppermost, and the index then falls along the stem as far as the end of the spirit column. The minimum thermometer is mounted in the same way as the maximum thermometer. It is read and set at the morning and evening observations.

Isothermal Charts.—The distribution of temperature over the earth's surface or any portion of it is most conveniently represented by a system of isotherms or lines connecting places of equal temperature, drawn on a map of the area to be considered. The most useful aids to a study of the general distribution of temperature over the surface of the earth are isothermal charts of the monthly mean-temperatures for January and July, normally the warmest or coldest months of the year in both hemispheres. An isothermal chart may show either actual temperature or temperature reduced to sea level. Normally, temperature decreases by about 1° F. for each 300 ft. of height. This variation is many times more rapid than the normal horizontal variation of temperature, so that altitude is the ruling factor which determines the temperature at any place. A map with isotherms of actual temperature would, therefore, emphasize the orographical features of a region, but would mask almost completely the horizontal distribution of

temperature. It is customary, therefore, to correct actual temperatures to sea level by the addition of 1° F. for each 300 ft. of altitude, and to construct isothermal maps of the corrected temperature. The effect of altitude must, however, always be borne in mind in using a map of sea level isotherms when studying the climate of any particular part of the world.

World charts of mean temperature reduced to sea level for the months of January and July are shown in Figs. 31 and 32.

Factors Affecting Temperature.—The principal factors affecting temperature are :—

(*a*) Altitude.
(*b*) Latitude.
(*c*) Land and sea distribution.
(*d*) Presence of snow or ice covering.
(*e*) Prevailing winds.
(*f*) Slope of the ground.
(*g*) Nature of the soil.

Of these (*f*) and (*g*) are merely local factors, and consideration of them will be deferred until later. The other factors may be regarded as general, exercising an influence over areas at least of continental dimensions.

(*a*) *Altitude.*—As has already been stated, owing to the rapid decrease of temperature upwards, altitude is the predominating factor which determines the temperature at any place. As a result, the variation of temperature from the foot to the summit of a high mountain in the tropics is equal to that at sea level from the equator to the poles.

(*b*) *Latitude.*—The angle of incidence of the sun's rays at the earth's surface increases from the tropics polewards, and therefore the amount of heat received on a given area diminishes in the same direction. Temperature therefore is normally highest in tropical and lowest in polar regions. There is, however, a seasonal variation of temperature owing to the annual changes in the sun's declination between 23½ degrees North and South. Not only is the inclination of the solar beam involved, but also the lengthening of the day in summer in high latitudes, which to some extent compensates for the reduction in intensity of insolation. Nevertheless, the highest temperatures occur in latitudes where the sun is overhead, about 20 degrees North in July and 20 degrees South in January, as may be seen by reference to Figs. 31 and 32.

(*c*) *Land and sea distribution.*—The distribution of land and sea has a marked effect upon temperature, the interior of a land mass being subject to extremes of heat and cold, while the oceans and

adjacent land areas enjoy more equable conditions. The difference between land and sea arises from a variety of causes :—

(i) The specific heat of water is greater than that of soil or rock, so that the gain or loss of a given amount of heat brings about smaller changes of temperature of the water than of the land surface.

(ii) Soil is a poor conductor of heat, so that changes of temperature are confined to a thin layer near the surface, and are correspondingly large for a given amount of heat gained or lost. The sea, on the other hand, is fairly transparent to the sun's rays, so that their heating effect penetrates to a considerable depth. The movement of the sea brings about a mixing of the water near the surface, and the heat supplied, being distributed through a thick layer of water is comparatively ineffective in raising the temperature.

(iii) Much of the incident heat is used up in evaporating water from the surface of the sea, a process which takes place with little change of temperature.

(iv) A land surface is an almost perfect absorber of radiation, whereas at large angles of incidence a water surface reflects much of the radiation falling upon it. Therefore a land area gains much more heat than a sea surface similarly exposed to the sun's rays.

Being a good absorber, a land surface is also a good radiator, and at night the radiative loss of heat from the land is much greater than from the sea. In high latitudes during the long polar night, this loss of heat goes on almost without interruption. The effect of cooling by radiation is cumulative, since once a mass of cold air becomes established in the lower layers of the atmosphere it cannot, owing to its density, be easily displaced, except by a denser and therefore colder air mass.

(v) The abundant water vapour over the sea exercises a blanketing influence by absorbing the outgoing radiation and re-radiating some of it back to the surface.

(*d*) *Presence of snow or ice covering.*—A snow surface reflects about 80 per cent. of the solar radiation falling upon it, and is also a good radiator of the terrestrial radiation wave-lengths. Consequently the temperature of a snow surface tends to sink to very low levels. Even if sufficient heat falls upon the snow surface to melt it, this process takes place at constant temperature, and so long as snow remains the temperature cannot rise above freezing point. The effect of freezing over of a sea surface, by destroying its mobility, is to make it similar in its behaviour to a land surface.

Extremely low temperatures may therefore be met with over the ice and snow in polar regions.

(*e*) *Prevailing Winds and Ocean Currents.*—Air moving over the surface of the earth tends to carry with it the temperature of its place of origin, modified by its subsequent history, and the wind therefore arrives at any place as a warm or cold wind according to the track it has followed. Areas not too far removed from the seaboard of the great land masses in temperate and high latitudes are particularly affected by this "imported" temperature, for it is between the oceans and the continents in these latitudes that the greatest contrasts of temperature occur. Generally, prevailing winds from sea to land exercise a tempering influence on the climate of the area over which they blow, whereas winds from land to sea carry with them extreme conditions. Over the sea, temperature is similarly affected by the great ocean currents and if the prevailing winds are favourable their influence may penetrate far inland.

World Distribution of Temperature—Annual Variation.—The general distribution of temperature over the surface of the earth may be studied with the aid of Figs. 31 and 32 in relation to which the influence of each of the factors above may be considered in turn. Altitude is excepted because allowance has been made for it in reducing the temperature to sea level.

The effect of latitude is at once evident, both charts showing a general decrease of temperature from the tropics polewards. The highest temperatures not only in January and July but at all times of the year tend to occur not at the equator itself but in latitudes 10 to 20 degrees North and South. The equatorial belt is a region of much cloud and rain, which restrict the supply of heat from the sun. The influence of land and sea distribution is shown in two ways. Firstly, the extremes of temperature occur over the land areas. Particularly noteworthy is the very low January temperature over Asia with the "cold pole" in north-east Siberia, whereas in July this region is slightly warmer than the average for its latitude. Similar cold centres appear in north Canada and Greenland. The highest temperatures, too, are found over the land, in South Africa and Australia in January, in the Sahara in July. Secondly, the regular run of the isotherms in the southern as compared with the northern hemisphere is to be attributed to the more uniform nature of the earth's surface, almost completely ocean, in the former.

Greenland is the one large region shown on the charts which is permanently covered with snow or ice. In winter the cold is almost as intense as that of north-east Siberia. In summer, the temperature, mostly below 32° F. is distinctly less than the average for the latitude shown by the southward bend of the 40° F. isotherm over Greenland in July.

The effect of the prevailing wind is exemplified by the distribution of temperature over the northern land masses of temperate

latitudes in January. There the prevailing winds are the Westerlies which bring the tempering influence of the oceans to western Europe, western North America, while to eastern Canada, eastern U.S.A., north China and eastern Siberia they bring the extreme cold of the continental interiors. The effect is well shown by the course of the 32° F. isotherm which crosses the Pacific coast of North America in latitude 56 degrees North and the Atlantic coast in latitude 41 degrees North; it enters Europe as far north as latitude 67 degrees North and, trending southwards across Asia re-enters the Pacific in latitude 35 degrees North. The North Atlantic in January furnishes the best example of the effect of ocean currents, where the Gulf stream and its north-eastward extension bring mild conditions to the British Isles, Scandinavia and the neighbouring sea and land areas. The north-eastern Atlantic is, for its latitude, the warmest region on the face of the earth. The transport of cold conditions by ocean currents is shown by the equatorwards bending of the isotherms off the coasts of south-west Africa, Chile and Morocco, which are washed by cold currents.

The difference between the summer and winter temperatures of any region is called its **Annual Range of Temperature.** Types of climate may be classified according to the annual range as follows :—

(*a*) *Oceanic Climate.*—The annual range is less than 20° F. This type occurs over the oceans and over land areas within the tropics. In the latter region the equality of day and night and the constantly high altitude of the midday sun favour uniform conditions.

(*b*) *Maritime Temperate Climate.*—The annual range lies between 20° F. and 35° F. This type occurs on the western sides of the continents in the northern hemisphere, the oceanic influence tempering otherwise continental conditions.

(*c*) *Continental Climate.*—The annual range is greater than 35° F. This type occurs in the interior regions of the great land masses of the northern hemisphere and on their eastern seaboards. The interior of Australia is the only area in the southern hemisphere with a climate of this type.

In most land areas of the world the highest and lowest temperatures of the year are reached in January and July, about one month later than the solstices. The summer maximum is reached at the same time in all latitudes, but the winter minimum, probably on account of the fact that the ground is covered with snow, becomes later with increasing latitude. Over the sea the extremes of air temperature occur in February and August, later than over land on account of the fact that the sea tends to conserve its heat and is therefore less rapidly affected than the land by changes in the

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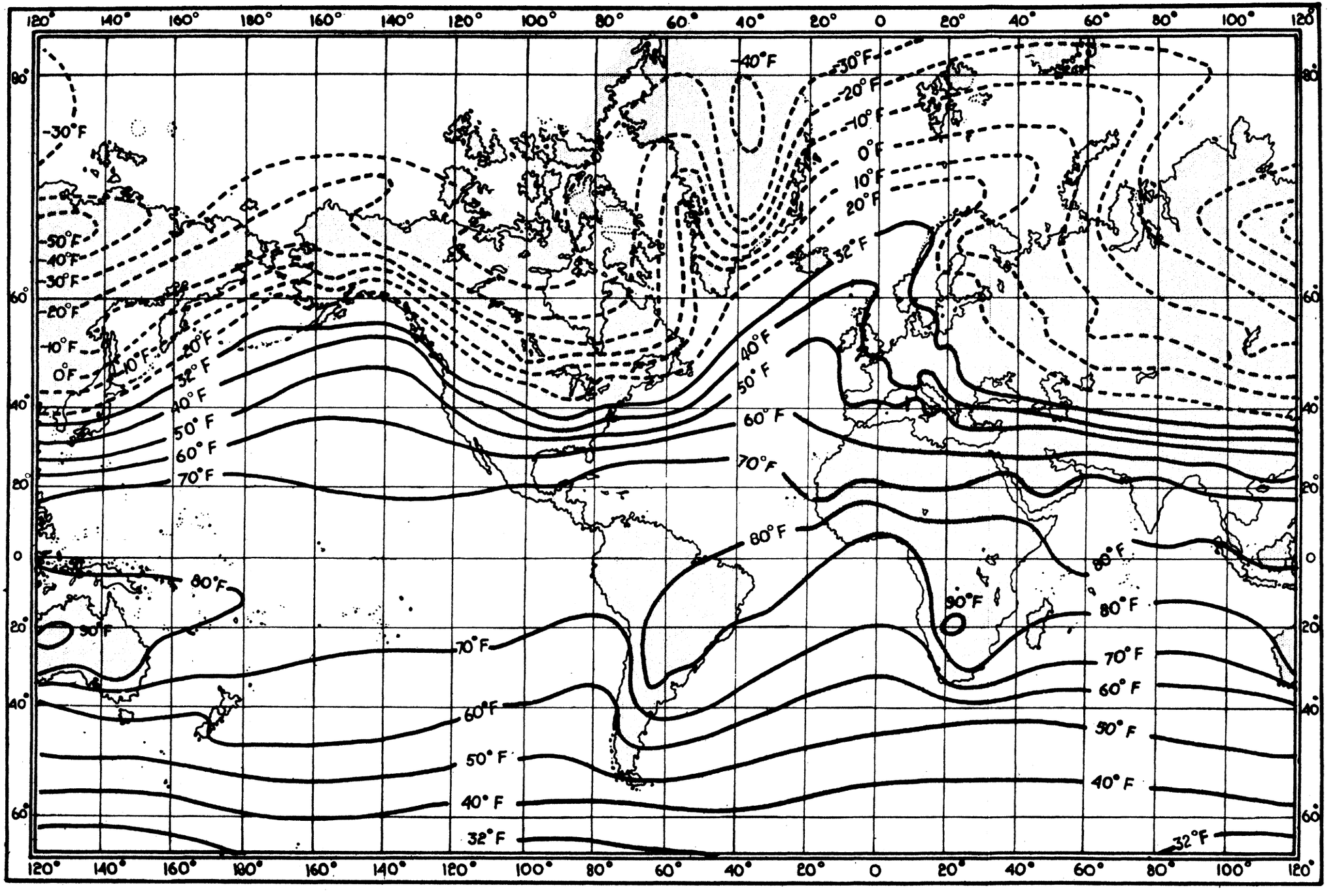


FIG. 31. ISOTHERMS OF M.S.L. TEMPERATURES FOR JANUARY.

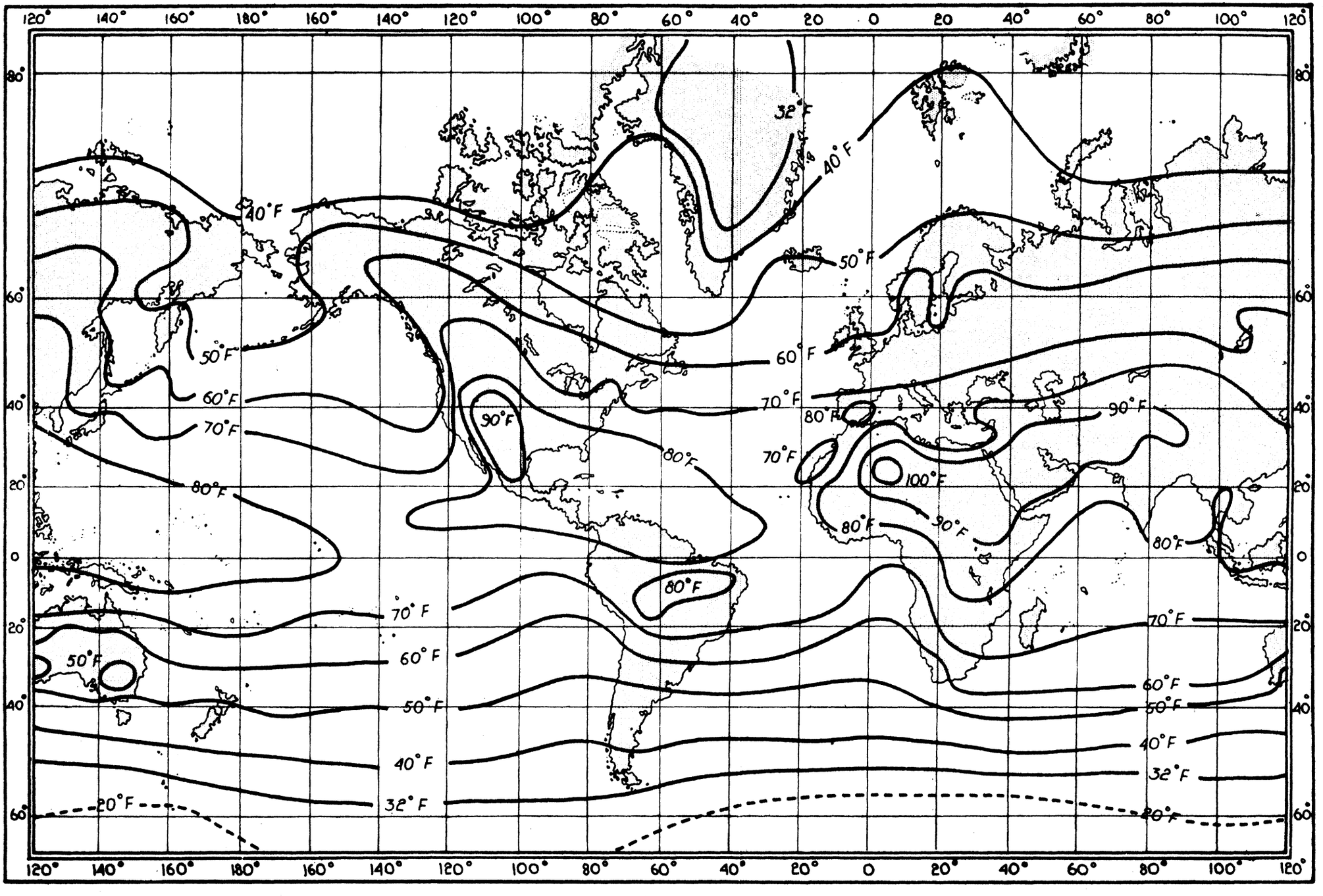


FIG. 32. ISOTHERMS OF M.S.L. TEMPERATURES FOR JULY.

amount of energy received due to the seasonal variations in the sun's declination.

Diurnal Variation of Temperature.—Except possibly at the poles there is normally at the earth's surface a daily rise and fall of temperature. The variation is generally much greater over land than over sea areas since the same factors which are operative in the case of the annual variation of temperature apply also to the diurnal variation. In the open sea the diurnal variation amounts to not more than 2° F., but even in a small land area such as Great Britain a variation of more than 40° F. in 24 hours has been recorded. In the great desert regions the daily variation is often much greater than this, intense heat by day being followed by night frosts. At a coastal station the diurnal variation is affected by land and sea conditions and takes intermediate values. The stronger control by the sun leads to a greater variation in summer than in winter. The following table shows the mean diurnal variation for the years 1920–9, in July and January, for an inland station (Vienna) and a coast station (Valencia).

Diurnal variation.	January.	July.
Vienna ..	7·9° F.	16·2° F.
Valencia ..	7·0° F.	8·8° F.

Clouds, which intercept the incoming and the outgoing radiation tend to smooth out the diurnal variation, keeping the temperature low by day and comparatively high by night. Clear weather, by permitting free radiation, enhances the diurnal variation. The clouding over of a previously clear sky during the night may bring about a small rise of temperature owing to radiation back to earth from the clouds.

The nature of the soil and soil covering have an important bearing on the diurnal variation. Bare soil and, in particular, sandy soil are good radiators and absorbers, but poor conductors of heat. The changes of temperature are, therefore, confined to the surface layers, and are correspondingly large. Vegetation, on the other hand, has a blanketing effect, reflecting much of the incoming radiation by day and preventing radiation from the ground to the clear sky at night. The diurnal variation is greater in valleys than in flat country. Air, cooled by radiation at night, becoming relatively dense, drifts downhill, and the coldest air is found in the valley bottoms. Moreover, it is in valleys that the surface of contact between the air and the ground is greatest, and the diurnal curve of air temperature most nearly follows that of the ground. The effect of slope is to raise or lower daytime temperatures according to its direction. In the northern hemisphere, for example, temperatures are higher on southward than on northward slopes. The night temperature is little affected.

Normally, the maximum temperature is reached about two hours after noon. and the minimum at about sunrise. It might

appear that the temperature should go on increasing between noon and sunset, while insolation is still being received, but a consideration of the loss of heat from the earth which goes on by day and night will show why this is not the case.

In Fig. 33, the curve *ABC* represents the insolation received at

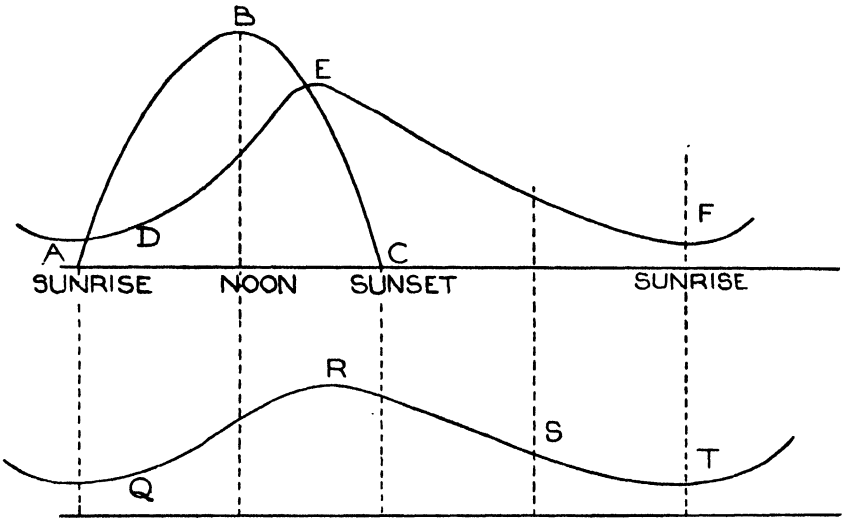


FIG. 33.—SOLAR AND TERRESTRIAL RADIATION AND DIURNAL VARIATION OF TEMPERATURE.

any place between sunrise and sunset. Assuming a cloudless sky, it is approximately a sine curve. In the lower half of the diagram, *QRST* represents the diurnal curve of temperature. During the forenoon, the temperature rises, *QR*. Now the outgoing radiation is proportional to the 4th power of the absolute temperature of the surface, and its rate of increase is relatively much greater than that of the temperature and that of the insolation which caused it. The curve of outgoing radiation, *DEF*, therefore slopes upwards more rapidly than *AB*, and crosses it at *E* shortly after noon. Now, so long as *ABC* lies above *DEF*, there is a net gain of heat at the earth's surface and the temperature rises, and *vice versa*. The maximum temperature therefore occurs at *E*, shortly after noon. The remaining portions of the curves *EF* and *RST* may be constructed from similar consideration, and it will be seen that the time of minimum temperature is at about sunrise.

The normal time of maximum temperature is advanced at many coastal places by the sea breeze which springs up during the forenoon as a result of the differential heating of land and sea. The sea breeze brings in relatively cool air, causing a sudden drop of temperature, and the maximum is reached from one to two hours before noon. In temperate latitudes in winter the normal diurnal variation may

be disturbed and even reversed by the arrival of warmer winds during the night and colder during the day. In summer, owing to the predominating influence of the sun, the diurnal variation is almost always normal, apart from the effect of the sea breeze above mentioned.

Ground Temperature—Sun Temperature.—So far only air temperature as measured by properly screened thermometers has been considered. If, however, a thermometer is freely exposed to the sun by day and to the open sky at night, the temperature which it indicates often differs considerably from the true air temperature. Any solid object exposed to the sun absorbs radiation to an extent depending upon the nature of its surface, dark matt surfaces being the best absorbers and light-coloured polished surfaces the worst. Solid objects exposed to the sun, as, for example, the various parts of a ship, become heated to different extents in a manner unrelated to the true air temperature. The degree of heating possible in the direct solar beam may be measured by means of a solar radiation thermometer, which has its bulb coated with lamp-black and is enclosed in an evacuated glass tube to prevent loss of heat from the bulb by conduction. The instrument is mounted in a position freely exposed to the sun's rays. It is no longer in use at official British stations, since its readings were found to be unreliable, two similar instruments similarly exposed not being in mutual agreement. Nevertheless, as indicating the possible heating due to absorption of solar radiation, the following high temperatures recorded with the black-bulb thermometer are interesting: 184° F. at Coolgardie (Australia), 178° F. at Greenwich, and 154° F. in the Antarctic.

A minimum thermometer laid horizontally on two pegs about 1 inch above the ground and freely exposed, known as the grass minimum thermometer, is used to indicate the lowest temperature occurring very close to the earth's surface at night. The grass minimum temperature is of considerable importance, in that it affects crops in the young and tender stages of growth. On a clear night it may be several degrees below the air temperature in the screen. If it falls below 30° F. while the air temperature remains above freezing point, there is said to be a ground frost.

Sea Temperature.—The general variation of sea temperature is decrease from the tropics polewards, but irregularities in the distribution are produced by the prevailing ocean currents, since the water tends to retain its own temperature. Poleward flowing currents are usually associated with sea temperatures relatively warm for the latitude, and *vice versa*. The Gulf Stream is the cause of the abnormally high sea temperatures of the north-east Atlantic, while the cold Labrador current brings low sea temperatures to the coasts of Newfoundland, Labrador and the eastern U.S.A. In the open ocean, however, temperatures tend to become equalized by mixing,

and it is in the narrow seas that the extremes of temperature have been recorded. The highest sea surface temperatures, over 90° F. occur in the Red Sea. The lowest, 28° F., which is approximately the freezing point of sea water, occur in the polar regions, and in all seas which become frozen in winter. In all parts of the world sea temperature suffers reduction through evaporative cooling. Inland seas are liable to be affected by the temperature of the water of rivers flowing into them. Near the coasts, where the prevailing winds are off-shore, the surface water is carried away seawards, and is replaced by cold water upwelling from below. Such conditions occur along the coasts of south-west Africa, Chile, Peru, Morocco and California.

The diurnal variation of sea surface temperature is small, amounting to about 1° F. in equatorial regions, where it is greatest. The annual variation is greatest in temperate latitudes, amounting on the average to 10° F. in the waters adjacent to the British Isles, and 40° F. off New York. It is, however, subject to wide variations, and is small in the tropics and in polar regions. The maximum and minimum are reached in September and March in temperate latitudes, but later in higher latitudes.

The surface water of the sea is generally well mixed, owing to the constant movement, and the distribution of temperature with depth is uniform near the surface. A spell of fine weather with light winds may, however, raise the temperature of the surface layer by a comparatively large amount, with a sharp surface of discontinuity at a depth of from one to two inches. It is found that this condition tends to persist and may survive several days of stormy weather. The necessity for using only the surface water in taking sea surface temperatures is therefore apparent.

The temperature of sea water normally decreases with increasing depth, the lowest temperatures occurring at the bottom of the sea in the polar regions and in the great ocean deeps. In tropical regions evaporation increases the salinity, and consequently the density of the warm surface water, causing it to sink and resulting in abnormally high temperatures at considerable depths.

Humidity.

Water Vapour in the Atmosphere.—Atmospheric air always contains mixed with it a small quantity of water in the form of invisible vapour. In warm damp climates the proportion of water vapour is relatively large and may amount to 4 per cent. by volume. It is small in cold climates and in desert regions. The water vapour mixed with the air behaves like any other gas. In particular it contributes its own pressure to the total pressure of the atmosphere. The pressure of a sample of air containing water vapour is the sum of the partial pressures of the air and vapour. These are the same as would be exerted by the air and the water vapour occupying the whole volume separately.

The proportion of water vapour to air is constantly changing. It is increased by evaporation from the surfaces of the sea, rivers and lakes, from snow and ice surfaces, from vegetation and from living animals and decreased by condensation in the form of cloud, fog, rain, snow, etc. The rate at which it occurs depends upon the nature of the evaporating surface, and other factors being equal, it is greater for a surface covered with vegetation than for a flat water-surface. Other factors affecting the rate of evaporation are temperature, wind speed, the amount of water vapour already present in the air (humidity) and, to a less extent, the pressure of the atmosphere. High temperature, strong wind, low humidity and low pressure aid evaporation and *vice versa*. The rate of evaporation is measured by means of an evaporimeter, which consists of a large tank sunk in the ground, about 2 feet deep and filled almost to the top. The evaporation can be determined by taking measurements of the depth of water in the tank, due allowance being made for rainfall as measured by a near-by rain-gauge.

Dry and Moist Air. Saturation.—Air which contains no water vapour whatever is called dry air. The term " dry " is, however, somewhat loosely applied to air which contains only a small proportion of water vapour, although the true meaning can usually be understood from the context. Air containing some water vapour is called moist air or damp air, the term " damp " usually signifying that the air is nearly saturated. For a given temperature there is an upper limit to the proportion of water vapour that air can contain. The limit increases with increasing temperature. Thus, for example, air at 32° F. is capable of holding water vapour in the proportion 6·1 grams per cubic metre ; at 50° F. the proportion is 9·5 grams per cubic metre. Air which contains the maximum possible amount of water vapour appropriate to its temperature is said to be **saturated** or " in the state of saturation." Volume for volume, water vapour is lighter than dry air in the ratio of their molecular weights (62/100). Saturated air and moist air are therefore always lighter than dry air at the same pressure and temperature. If a water surface is available, evaporation constantly takes place from it provided that the air in contact with it is not already saturated. It is possible for very pure air to contain a larger proportion of water vapour than is necessary for saturation ; such a sample of air is said to be **supersaturated.**

Absolute Humidity and Vapour Pressure.—The Absolute Humidity of a sample of air is the mass of water contained in a given volume of the moist air. It is usually expressed in grams per cubic metre. It can be measured by passing a known volume of air through a series of tubes containing chemical drying agents which extract the water vapour from the air. Weighing of the tubes before and after the experiment determines the amount of water vapour originally contained in the air. The vapour pressure is that part of the atmospheric pressure exerted by the water vapour.

At a given temperature it is directly proportional to the absolute humidity and provides an alternative means for the specification of that quantity. It is usually expressed in millibars. If, in the experimental determination of absolute humidity, as above described, the air, after drying, is brought to its original volume and temperature, it is found that its pressure has undergone a reduction. The difference of pressure before and after drying is the vapour pressure. The general formula connecting vapour pressure and absolute humidity is :—

$$e = \frac{T\,\delta}{216{\cdot}7}$$

where e = vapour pressure in millibars,
δ = absolute humidity of the air,
T = absolute temperature of the air.

In considering the physics of the atmosphere it is customary, for convenience in calculation, to use two other quantities to specify the amount of water vapour present. The first, the **Specific Humidity,** is the amount of water vapour in grams, contained in one gram of **moist** air. The second, the **Humidity Mixing Ratio,** is the amount of water vapour in grams mixed with one gram of **dry** air. These quantities are closely related to the absolute humidity and vapour pressure.

The variation of the absolute humidity for saturation at different temperatures may be shown by means of a curve (Fig. 34).

Relative Humidity and Dew-Point.—Absolute humidity and vapour pressure are not conveniently measurable as a matter of routine practice and therefore other related properties of moist air are used for meteorological observations. A given sample of moist air contains a certain fraction of the maximum possible amount of water vapour corresponding to its own temperature. This fraction, expressed as a percentage is called the **Relative Humidity** of the air.

Let M, e be the saturation values of absolute humidity and vapour pressure, respectively.

Let m, e' be the actual absolute humidity and vapour pressure of a given sample of air.

Then the relative humidity is given by :

$$\text{R.H.} = \frac{m}{M} \times 100 = \frac{e'}{e} \times 100.$$

For example, suppose air at 60° F. contains 10 grams of water vapour per cubic metre. From Fig. 34, $M = 13{\cdot}0$, also $m = 10$;

$$\text{therefore, R.H.} = \frac{10}{13{\cdot}0} \times 100$$
$$= 77 \text{ per cent.}$$

Now suppose that the air in this example undergoes a reduction of

temperature from its original 60° F. To a first approximation, neglecting the change of volume of the air, the amount of water

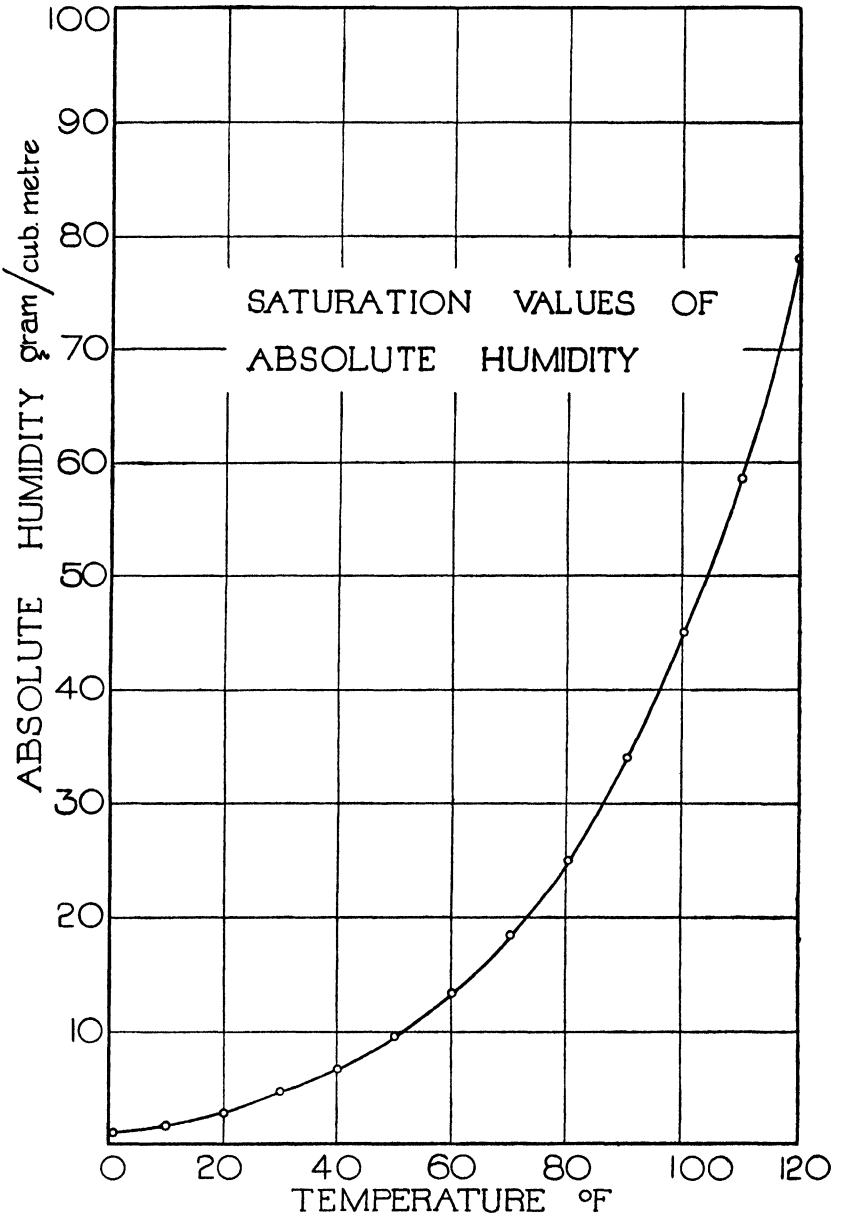


FIG. 34.—CURVE SHOWING RELATION BETWEEN TEMPERATURE AND ABSOLUTE HUMIDITY FOR SATURATION.

vapour present remains unchanged for a time at 10 grams per cubic metre. Reference to Fig. 34 shows that this amount of water is

sufficient to saturate the air at 52° F. Therefore, if the air is cooled to 52° F. it will become just saturated. If it is further cooled, say, to 32° F., it will be capable of holding only 6·1 grams of water vapour per cubic metre, and the surplus, 3·9 grams, will condense out as liquid water. Condensation will commence as soon as the saturation temperature is passed. This temperature is called the **Dew-Point.**

The above figures are not strictly accurate since if the air is cooled at constant pressure its volume will decrease and the original 10 grams of water vapour will occupy less than 1 cubic metre *i.e.*, the absolute humidity of the air will rise merely on account of the change of volume. For this reason it is necessary to define the dew-point with reference to vapour pressure and not absolute humidity. The dew-point of a mass of air is the temperature for which the saturation vapour pressure is identical with the actual pressure of the vapour in the air.

The dew-point may be measured directly by means of the Dew-Point Hygrometer (Fig. 35).

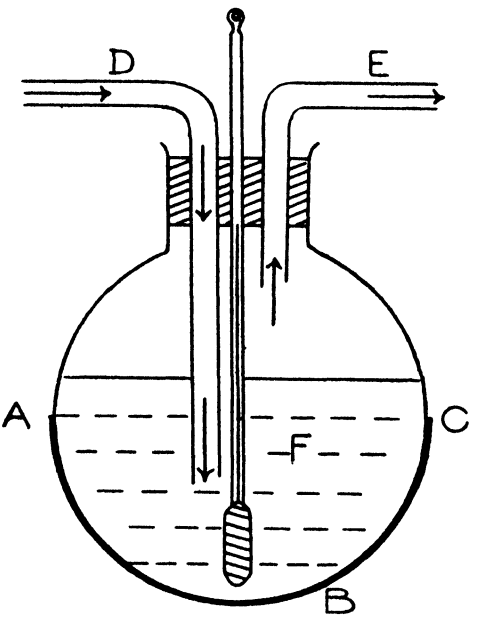


FIG. 35.—DEW-POINT HYGROMETER.

A small flask, the lower half of the bulb of which has a polished metal coating *ABC*, is partially filled with ether or some other volatile liquid, *F*. By means of a suitable arrangement of tubes, *D* and *E*, and an aspirator (not shown), air is sucked through the apparatus, and bubbling through the liquid causes it to evaporate. Now, the loss of heat consumed in evaporating the liquid leads to

a reduction of temperature of the polished metal coating and the air immediately in contact with it. As soon as the dew-point of the air is reached, moisture begins to condense on the polished metal cap, dimming its surface. The temperature registered by the thermometer when the dimming is first observed, is the dew-point of the air. A check is given by stopping the aspiration and noting the temperature at which the dimming disappears.

Measurement of Humidity. The Wet-Bulb Thermometer.—For meteorological purposes the humidity and other associated properties of the air are most conveniently determined by means of a wet- and dry-bulb psychrometer, consisting of two similar thermometers mounted side by side, in the screen, the electric psychrometer or the whirling psychrometer. (*See* Chapter I.). One of these, known as the wet-bulb thermometer, has its bulb kept permanently moistened by means of a muslin cap closely wrapped round it, from which an absorbent wick dips into a vessel containing pure water. Air flowing past the wet bulb, provided it is not saturated, evaporates some of the water from the muslin. The latent heat required to effect the evaporation is supplied by the air, whose temperature falls in consequence. In a perfect instrument, the air immediately surrounding the wet bulb becomes saturated and its temperature is determined by the fact that the heat given up by the air in consequence of its fall of temperature is equal to that required to effect evaporation sufficient for saturation at the reduced temperature. It is this temperature which is registered by the wet-bulb thermometer, and is in consequence called the wet-bulb temperature. Its value is intermediate between the dew-point and the dry-bulb temperature. If the air is saturated to begin with, no further evaporation takes place from the wet bulb and the two thermometers read alike. If the air is warm and dry, a large amount of water vapour must be evaporated from the wet bulb to bring about saturation, the reduction of temperature to supply the necessary heat is large and the wet-bulb temperature differs considerably from that of the dry bulb, hence the difference between the dry- and wet-bulb temperatures increases as the relative humidity decreases, and *vice versa*. The same applies when the air is cold, but the difference between the wet- and dry-bulb temperatures for a given humidity is less.

If the dry- and wet-bulb temperatures are measured, the relative humidity, vapour pressure and the dew-point may be obtained from specially constructed tables known as Psychrometric (or Hygrometric) Tables. The reading of the wet-bulb thermometer is independent of wind speed past the bulb if the latter exceeds 10 knots. This is often difficult to secure with thermometers mounted in a screen in the ordinary way. This arrangement therefore requires a set of hygrometric tables based on average wind conditions and different from that appropriate to the electric psychrometer for which the ventilation is sufficient.

A continuous record of relative humidity is furnished by the Hair Hygrograph. At many land stations this instrument is housed in the large Stevenson screen. The sensitive element is a bundle of human hair from which the oil has been extracted by washing in alcohol. The length of such a bundle of prepared hair is affected by changes in the relative humidity of the air. In the hair hygrograph the bundle of hair is fastened to rigid supports at its ends, and is kept taut in the middle by means of a hook. The hook is connected to a pen-arm carrying a pen which moves over a chart wrapped on a revolving drum. Changes in the length of the hair cause the pen to trace a continuous record on the chart. If the hairs are kept clean by frequent washing with pure water, the readings of the instrument are reliable to within 5 per cent. at all temperatures, both above and below the freezing point.

Variation of Humidity over the Earth's Surface.—A distinction must be drawn between absolute and relative humidity.

Absolute humidity.—The absolute humidity of the air is limited to the saturation value appropriate to its temperature. The tendency of air, which is not saturated, is constantly towards saturation, provided that moisture is available for evaporation. High temperature is favourable to high absolute humidity, since the capacity of air for water vapour increases with its temperature. Hence the general variation of absolute humidity is a decrease from the tropics polewards, but this simple distribution is modified somewhat by the amount of moisture available for evaporation. In the equatorial regions, the absolute humidity is constantly high, owing to the high temperature coupled with abundant moisture. Over deserts, despite the high day temperatures, the absolute humidity is low, moisture being scarce. On account of low temperature, the absolute humidity is low over the continents in winter, and in high latitudes throughout the year.

Relative humidity.—The variation of relative humidity is complicated, since it depends not only upon the amount of water vapour actually present, but also upon the dry-bulb temperature. For example, air at 32° F. containing 6·1 grams of water vapour per cu. m., has a relative humidity of 100 per cent. The same amount of water vapour with the air at 50° F. gives a relative humidity of only 64 per cent. Owing to its small capacity for water vapour, cold air is in general more nearly saturated than warm air. The distribution of relative humidity over the earth is accordingly almost the reverse of the distribution of temperature and absolute humidity. The chief exception is the equatorial belt where, owing to the abundant moisture, the relative humidity is high notwithstanding the high temperature. Relative humidity is lowest over desert regions, where high temperature is associated with low absolute humidity, and highest in the polar regions.

Little is known at present regarding the variation of humidity

over the sea. At first sight it would appear that conditions approaching saturation obtain everywhere except near the coasts, owing to the long fetch of the air over the sea surface. Consideration shows, however, that comparatively low humidities are possible under two conditions. The first is the polar air stream, or more generally any air current flowing from a cold to a warm sea surface. The air is warmed by contact with the sea but evaporation cannot keep pace with the rise of dry-bulb temperature. So long as the temperature of the air continues to rise, low humidity is to be expected but when the air temperature becomes steady saturation conditions are again approached. Much depends upon the time for which the air is in contact with the sea surface—that is, upon the wind speed and the fetch. Secondly, low humidity occurs in subsiding air, that is, air descending from upper levels to the surface. It will be shown in Chapter VII that this process leads to heating of the air, so that in virtue of its rise of temperature it arrives at the surface relatively dry. Now subsidence occurs on a large scale in the sub-tropical high-pressure belts (Chapter VIII) and low humidity is to be expected both within these belts and in the trade-winds which flow out of them towards the equator. On the other hand, poleward flowing currents are generally subject to a fall of temperature owing to the decreasing temperature of the sea surface over which they flow, and even if they are not saturated to begin with they very rapidly become so.

Near the coasts the humidity conditions depend upon whether the wind is on- or off-shore. In the latter case the humidity over the sea will be that over the adjacent land, modified by the sea conditions as the air travels further from the coast.

Diurnal Variation of Humidity.—*Absolute Humidity.*—At sea and coast stations evaporation is limited only by the attainment of saturation. The variation of absolute humidity therefore follows that of temperature very closely. The maximum occurs in the early afternoon, the minimum shortly after sunrise. When the temperature begins to fall in the afternoon evaporation still goes on, but water vapour is constantly being diffused into the higher levels so that the reduced afternoon evaporation cannot make good the deficiency. There is a similar variation at inland stations in winter, but in summer there are indications of two maxima, at about 0900 and 2100 local time, and two minima, at about 1500 and shortly after sunrise. The early morning minimum is due to the low temperature. The rise of temperature during the forenoon is accompanied by evaporation but by about 0900 another factor, convection, becomes operative. (*See* Chapter VII.). The heated layers of air at the surface ascend and are replaced by drier air from above. The absolute humidity therefore falls, reaching a minimum when convection is strongest in the afternoon. From then onwards convection weakens, the air remains at the surface long enough to take up more water vapour and the absolute humidity

rises. The fall during the night is due to diffusion of water vapour upwards and to the fall of temperature, and the excess moisture is condensed out as fog or dew.

Relative Humidity.—The diurnal variation is determined by absolute humidity and temperature, the latter being the controlling factor. Although the absolute humidity usually rises during the forenoon, temperature rises, by comparison, much more rapidly, and similarly for the fall at night. The variation of relative humidity is therefore reversed from that of temperature, the maximum occurring at sunrise and the minimum in the early afternoon.

As in the case of temperature the diurnal variation of absolute and relative humidity are modified by the arrival of a drier or damper air mass associated with a change of wind.

Pressure.

Atmospheric Pressure.—Although it is a well-known fact that the atmosphere exerts a pressure, the exact implications of this statement often are only somewhat vaguely understood and it may therefore be desirable as a preliminary to recall some of the fundamental facts in relation to pressure generally. By virtue of its elasticity any gas exerts a force upon a surface with which it is in contact. The pressure of the gas is equal to the magnitude of this force per unit area of the surface. Thus, if A is the area of the surface and F the total force exerted upon it by the gas, the pressure P is given by the relation :—

$$P = F/A.$$

This holds for an area of any magnitude only if the pressure is the same throughout the whole volume of the gas. When the pressure varies it may still be defined at any point as a force per unit area, thus :—

$$P = \delta F/\delta A,$$

where δA denotes a small area surrounding the point, δF the small force acting upon it. This force is always perpendicular to the surface and always has the same magnitude in whatever direction the surface may be orientated about the point in question. Accordingly, the pressure at any point is not associated, as is a force, with any particular direction, but is said to act uniformly in all directions.

Consider a small thin metal disc at rest horizontally upon the ground, together with the vertical column of air which, resting upon the disc as base and having everywhere the same cross section as the disc, extends upwards to the limit of the atmosphere. The whole weight of this air column is supported by the disc which therefore, sustains a certain pressure whose magnitude is equal to the weight of the air column divided by its cross-sectional area.

That is, the pressure is equal to the weight of a vertical column of air of unit cross section, extending from the disc to the upper limit of the atmosphere, per unit area. This is simply expressed by saying that the pressure at any point is that due to the weight of the air above it.

Pressure may be expressed in various units according to the nature of the pressure involved, *e.g.*, boiler pressures are expressed in pounds per square inch, wind pressures on buildings in pounds per square foot. In physical problems the C.G.S. unit, dynes per square centimetre, is used. The meteorological unit is the millibar, equal to 1,000 dynes per square centimetre. The average value of the atmospheric pressure at the earth's surface is approximately 15 pounds per square inch, or 1,000 millibars. This pressure is often referred to as " one atmosphere." A pressure of 1,000 millibars is also called one bar.

The Simple Barometer.—That the atmosphere exerts a pressure was first demonstrated by Torricelli in 1640. Torricelli's experiment, involving the basic principle of the mercurial barometer, which is now used almost universally to measure atmospheric pressure, is performed as follows :—

A strong glass tube, *A* (Fig. 36), about 3 feet in length, closed

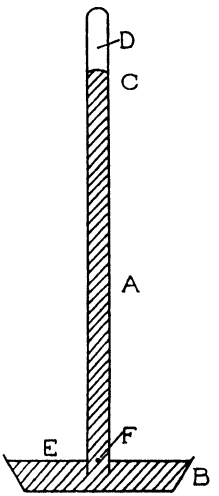


FIG. 36.—TORRICELLI'S EXPERIMENT—THE SIMPLE BAROMETER.

at one end, is filled with mercury. The open end being temporarily sealed with the finger, the tube is inverted and its lower end immersed in mercury contained in a shallow open vessel, *B*. The finger is now withdrawn and the tube fixed in a vertical position. A column of mercury then remains standing in the tube to the level *C*, about 30 inches above the surface of the mercury in *B*. The space *D* above the mercury column is an almost perfect vacuum, containing

only a little mercury vapour. The weight of the mercury column in the tube is supported by the pressure of the atmosphere on the surface of the mercury in the open vessel. Let E denote any point in this surface, and let F be a point at the same level within the tube. The pressure at E is equal to atmospheric pressure; that at F is the pressure due to the weight of a column of mercury of height FC, and since E and F are at the same level within the mercury, these pressures are equal. The weight of the mercury column (per unit cross section), therefore, affords a measure of the pressure of the atmosphere. Now, the weight of the column depends upon (*a*) its density, (*b*) the value of gravity at the place (*c*) its height. The density is known if the temperature is known, so for given values of temperature of the mercury and of gravity, the pressure is determined by the height of the column. The mercurial barometer is an adaption of this simple principle to afford a ready means of measuring the height and temperature of the column of mercury which the atmosphere can support. It is described in detail in Chapter I.

Pressure Units.—Formerly, owing to the fact that it is the height of a mercury column which is actually measured, pressure, as determined by a mercury barometer, was expressed in units of length, *i.e.*, in inches or millimetres of mercury. It was necessary, at the same time, to specify the temperature of the mercury and the value of gravity. Standard values of these quantities were selected, viz. :—

Temperature of mercury ..	32° F.
Value of gravity	That at M.S.L. in latitude 45°.

Thus, pressure was specified in terms of inches or millimetres of mercury at 32° F. at M.S.L. in latitude 45°. Corrections had to be applied to observations taken under conditions other than standard. It was, however, recognized as illogical to express pressures in units of length, and moreover the mode of specification was cumbersome.

Therefore, the British Meteorological Service in 1914 decided to adopt a new unit, the millibar, which is a unit not of length, but of pressure. The millibar has since been brought into use internationally. Normally, in temperate and high latitudes, pressure varies between 970 and 1,030 millibars approximately, but readings as high as 1,050 millibars and as low as 925 millibars may occur occasionally. The highest and lowest pressures (reduced to M.S.L.) on record are respectively 1,076·2 millibars at Irkutsk (Siberia), and 886·8 millibars experienced on board a ship 400 miles east of Luzon (Philippines).

Although barometers are now graduated in millibars, what is immediately observed is still the height of a column of mercury and, before the true pressure can be obtained, the reading must be corrected to standard conditions of temperature and gravity. In taking the reading, the millibar graduations on the scale must,

in the first instance, be regarded as measures of length, and the corrections specified in terms of "millibar lengths" also. For example, if the top of the mercury column stands opposite the 1,000 millibars graduation, all that is involved is that the height of the column is equal to the length of the scale between zero and this graduation. For convenience this is expressed by saying that "the uncorrected pressure is 1,000 millibars." Similarly, the correction is worked out in "millibar lengths," called for convenience "millibars." When added to or subtracted from, the uncorrected pressure, as the case may be, it gives the true pressure in millibars. It is only after the correction has been applied that it is strictly correct to speak of a pressure of (say) 1,000 millibars. Further, the use of an expression such as "a pressure of 1,000 millibars" implies that the pressure referred to is a true, corrected pressure.

The method of correcting barometer readings to obtain true pressures is described in Chapter I.

Reduction of Pressure to Mean Sea Level.—The pressure at any point, being due to the weight of the air above it, diminishes with increasing altitude. Near to sea level the average rate of decrease of pressure with height is $3\frac{1}{2}$ millibars per 100 feet. The normal variation of pressure in the horizontal direction is, however, of the order 1 millibar in 30 miles. The variation of actual pressure between stations therefore emphasizes their variation of altitude and masks the horizontal pressure variation. For this reason it is necessary to reduce pressures to a common level, mean sea level being selected.

To reduce barometric pressure to M.S.L. there must be added to it the pressure due to a column of air extending vertically from barometer level down to sea level. Such a column of air has, of course, no real existence. It would be provided by a vertical mine shaft sunk to sea level from the station. The correction depends upon the length of such an air column, *i.e.*, upon the altitude of the station. To a less extent it depends upon the temperature and pressure of the air column and a value must be assumed for the former of these quantities. The M.S.L. pressure may be obtained by means of the formula :—

$$\log p_o = \log p + \frac{h}{221T}$$

where p_o = pressure at M.S.L.
p = pressure at station level,
h = altitude of station in feet,
T = mean absolute temperature of the air column between station level and M.S.L.

Logarithms are to the base 10.

For a station not more than about 1,000 feet above the sea level, T in the formula may be taken as the temperature at the

station itself. For a station at a higher altitude than 1,000 feet this method may involve considerable errors. If at such a station the temperature is abnormally high, the correction term $\frac{h}{221T}$ is too small, resulting in a value of M.S.L. pressure which is too low. Similarly, if the temperature is abnormally low the corrected pressure is too high. The very high winter M.S.L. pressures over Siberia and the relatively low pressures occurring on hot days over the elevated regions of Spain and Africa are to be attributed in part at least to the method of reduction to M.S.L. It would, perhaps, be more satisfactory to correct all pressures to a common level of (say) 2,000 feet above M.S.L. The majority of stations are at a lower altitude than this, and the temperature in the air column between station level and 2,000 feet could be observed directly by means of an aircraft carrying thermometers.

Isobaric Charts.—The distribution of pressure over the earth's surface or any portion of it is conveniently represented by a system of isobars drawn upon a map of the area considered. An isobar is a line drawn through all places at which the M.S.L. pressure has the same value. A chart of the isobars over the British Isles at 1300 G.M.T. on 17th June, 1935, is shown in Fig. 138. The values of pressure, to the nearest tenth of a millibar (with the hundreds figures omitted) are written in the positions of the stations. Isobars are drawn at intervals of two millibars and the value of the pressure is written alongside each isobar. In many respects isobars are similar to the contour lines on an orographical map. They follow approximately parallel curves, can never cross one another and must either form closed curves or terminate only at the edges of the map. The arrangement of isobars in Fig. 138 shows an area of low pressure to the west of Ireland with an area of high pressure over France

On an orographical map the slope or gradient of the land surface is indicated by the spacing of the contour lines. Closely spaced lines represent steep slopes, lines far apart gentle slopes. The gradient is measured along the line of greatest slope, that is, along the shortest distance between consecutive contours and approximately perpendicular to them. Similarly, on an isobaric chart the spacing of the isobars indicates the pressure gradient, which is measured by the shortest distance between consecutive isobars. If the isobars are close together the gradient is said to be strong or steep; if far apart, weak or slack. In Fig. 138, the pressure gradients are relatively steep over the south-west coast of Ireland and slack over most of England. Quantitatively, the pressure gradient may be expressed as the variation in pressure over a given distance, *e.g.*, in millibars per 100 miles, or as the distance apart of consecutive isobars, *e.g.*, distance apart of consecutive isobars at 2 millibar intervals = 50 miles. For purposes of calculation

the gradient must be expressed in a consistent system of units, *e.g.*, in the C.G.S. system as dynes per square centimetre per centimetre, in the British system as pounds per square foot per foot.

Relation Between Surface Wind and Isobars.—The strength of the wind near the earth's surface may be specified by its average speed (*e.g.*, in knots or feet per second) or by its force on the Beaufort scale (Chapter II). The direction of the wind is specified as the compass point from which the wind blows. The distribution of wind at any particular time over a given area is conveniently represented by wind arrows drawn upon a map of the area considered. A wind arrow is a line terminating at the position of the station and drawn outwards from it in the direction from which the wind blows. The arrow carries a number of feathers representing the force of the wind on the Beaufort scale. Each full feather represents two points on the scale, the odd point, if required, being represented by a half feather. A calm is shown by a circle drawn round the position of the station. For example, in Fig. 138 the wind at Scilly Islands is SSE. force 5; at Pembroke, SSW. force 3; and calm at Aldergrove. A brief study of Fig. 138 suggests the following conclusions. Firstly, the wind blows nearly parallel to the isobars in its neighbourhood, crossing them at a small angle in the direction from the higher to the lower pressure. Secondly, the direction of the wind is such that to an observer facing the wind the lower pressure is on his right. This is a particular case of a law first enunciated by Professor Buys Ballot of Utrecht in 1857. The formal statement of Buys Ballot's law is :—" If the observer stands with his back to the wind the lower pressure will be on his left if in the northern hemisphere, and on his right if in the southern hemisphere." Seamen are, however, accustomed to face the wind and the law in the form preferred by them is as follows :—Face the wind; the lower pressure will then be to the right if in the northern hemisphere, and to the left if in the southern hemisphere. Alternatively, the law may be expressed by saying that the air circulates clockwise round centres of high pressure and anticlockwise round centres of low pressure in the northern hemisphere and *vice versa* in the southern hemisphere. Thirdly, the wind is strongest where the pressure gradient is steepest and *vice versa*. The speed of the wind is roughly proportional to the pressure gradient. It will be seen later that these rules relating the wind to the pressure distribution apply generally and arise as a result of the forces which act on air moving under the influence of a variable pressure distribution over the surface of the rotating earth.

The Diurnal Variation of Pressure.—In temperate and high latitudes, the atmospheric pressure is subject to large, rapid and irregular changes which are associated with the development and movement of depressions and anticyclones. In the tropics, however, the pressure changes little from day to day, but is subject

to small regular fluctuations having a twelve-hourly period. The maxima and minima occur at the same local time everywhere, the maxima at 1000 and 2200 local time, the minima at 0400 and 1600. The diurnal pressure wave is therefore analogous to a double tidal wave sweeping round the earth once in each 24 hours.

The range of the diurnal variation is greatest at the equator, and becomes insignificant above latitude 60°. In temperate latitudes it can only be detected from the average hourly values of pressure taken over a large number of years, but in the tropics it is the most noteworthy feature of the pressure variation, and appears as such on the barogram. A typical barogram for tropical waters is shown in Fig. 37.

The average range of the diurnal change and its variation with latitude is shown in the following table :—

Latitude.	Range of diurnal variation.
Equator	3 mb.
30°	2·5 mb.
35°	1·7 mb.
45°	1·2 mb.
50°	0·9 mb.
60°	0·4 mb.

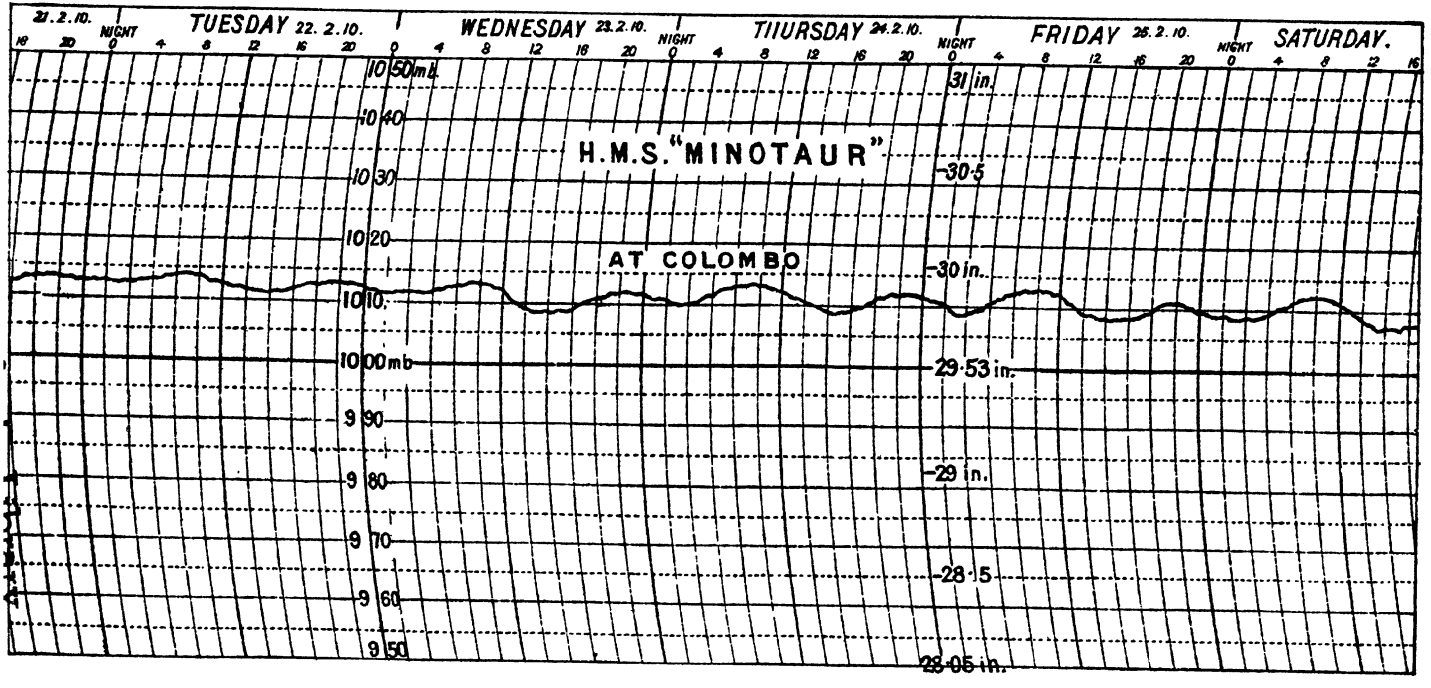


FIG. 37.—ILLUSTRATING DIURNAL RANGE OF ATMOSPHERIC PRESSURE IN THE INDIAN OCEAN.

CHAPTER VII.

Pressure, Temperature and Humidity in the Upper Air

Pressure.

The Pressure-Height Formula.—In the previous chapter it has been shown that the pressure of the atmosphere at any point is the pressure due to the weight of air above the point. Pressure in the atmosphere therefore diminishes with increasing height. Near the earth's surface pressure decreases at the rate of $3\frac{1}{2}$ millibars for 100 feet of ascent. Consider two horizontal levels in the atmosphere, the height of the lower being h_0 and of the upper, h, the corresponding pressures being p_0 and p, respectively. The pressure difference between the two levels is equal to the pressure due to the weight of the air column between them. The weight of the air column depends upon :—

(*a*) its height h–h_0
(*b*) its temperature
(*c*) its pressure

the last two factors affecting density. The relation between pressure and height is usually written in the form :—

$$h - h_0 = 221 \cdot 1 \; T \; (\log_{10} p_0 - \log_{10} p),$$

where T is the mean absolute temperature of the air between the two levels. The height difference, h–h_0 is expressed in feet, but it is immaterial in what units p_0 and p are expressed, since $\log p_0 - \log p = \log (p_0/p)$, *i.e.*, only the ratio of the pressures is involved. This formula is known as the **Pressure-Height Formula.** When heights are expressed in metres it becomes :—

$$h - h_0 = 64 \cdot 7 \; T \; (\log_{10} p_0 - \log_{10} p).$$

The formula is strictly applicable only when the temperature is constant throughout the whole of the air column. If, however, the height difference does not exceed 2,000 feet, it is sufficiently accurate to take T as the arithmetic mean of the temperatures at the two levels. For greater height differences, the mean temperature should be found from observations taken at regular intervals, not exceeding 2,000 feet between the two levels.

The effect of humidity is usually negligible for most practical purposes, and is therefore ignored in the pressure-height formula. Since damp air is of lower density than dry air, the general effect of the presence of water vapour is to increase the height interval corresponding to given values of pressure.

The Altimeter.—The pressure-height formula finds its most useful application in the altimeter—the instrument in general use

as the height indicator in an aircraft. Height in the atmosphere is measured through the associated pressure changes, for the rapid indication of which a portable aneroid barometer is suitable. The aneroid may be graduated to read pressure directly, in which case, if the vertical temperature distribution is known, the height may be determined by applying the pressure-height formula. Generally, however, the instrument is graduated so as to read height directly. It is then known as the altimeter. It is subject to the general defects of aneroids mentioned in Chapter I, and in particular to hysteresis lag. After a rapid descent from 20,000 feet (say) the pointer may, instead of indicating zero, indicate 300 feet. After a few minutes this may have changed to 100 feet, but it may take hours for the pointer to return to zero. The altimeter must also be compensated for temperature. It is usually constructed so that the height scale is uniformly graduated.

Although the altimeter indicates height it responds only to changes of pressure and the height interval corresponding to a given change of pressure depends upon the temperature of the air. But the given change of pressure is always indicated by the same movement of the pointer, that is by the same height interval on the dial. It follows that to design an instrument to record height correctly at all temperatures would require an expanding scale. In practice, the instrument is designed to record heights correctly for one specified temperature distribution. For temperatures other than this, corrections must be applied to the altimeter readings. The law connecting temperature with height for which the altimeter reads correctly is known as its calibration law.

The Isothermal Altimeter.—The most common type of altimeter is designed to indicate height correctly when the air temperature is uniform and equal to 10° C. (283° A. or 50° F.). It is therefore known as the Isothermal Altimeter, sometimes also as the Trade Scale Altimeter. Its chief advantage is that the corrections for deviations of temperature from the isothermal law can be easily calculated mentally. Since, however, the specified conditions of temperature often differ somewhat widely from those which obtain in practice, the corrections are often large.

The pressure-height relation for the isothermal law is :—

$$h - h_0 = 62{,}570\ (\log_{10} p_0 - \log_{10} p)$$

$h - h_0$ being expressed in feet.

Temperature Corrections for Isothermal Altimeter.—(*a*) *When the temperature distribution between ground level and the aircraft is known.*—The mean temperature of the air below the level of the aircraft is determined from the readings of the strut thermometer, (Chapter IV), taken at regular intervals not exceeding 2,000 feet. If the mean temperature differs from 10° C., a correction must be

applied to the altimeter reading. The correction is calculated in the following way :—

Take h_o = 0, *i.e.*, h_o is at ground level.

Let p_o = pressure at ground level,

p = pressure at aircraft level,

h = true height of aircraft,

h_1 = height indicated by the altimeter,

T = mean (Absolute) temperature below the level of the aircraft.

Then $h = 221 \cdot 1T (\log p_o - \log p)$ ..... (1)

The indicated height at pressure p will be the same as if the temperature were everywhere 283° A.

i.e., $h_1 = 221 \cdot 1 \times 283 (\log p_o - \log p)$ ..... (2)

Dividing (1) by (2), $\frac{h}{h_1} = \frac{T}{283}$ ..... (3)

The ratio of the true height to the indicated height is therefore equal to the ratio of the true mean absolute temperature below the aircraft to 283° A. With negligible error, the round number 300 may be substituted for 283 in equation (3), giving

$$h = \frac{h_1 T}{300} \quad \ldots\ldots\ldots\ldots\ldots \quad (4)$$

The correction is given by Airy's Rule, which follows from (4) :—

" When the mean temperature of the air between the ground and the aircraft differs from 10° C. (283° A.) the indicated height must be $\left\{\begin{matrix}\text{increased}\\\text{decreased}\end{matrix}\right\}$ by its $\frac{1}{300}$ part for each degree by which the mean temperature $\left\{\begin{matrix}\text{exceeds}\\\text{falls short of}\end{matrix}\right\}$ 10° C."

Example.—Indicated height, 7,500 feet. Mean temperature, 20° C., *i.e.*, 10° in excess of 10° C. The indicated height must be increased by

$$\frac{10}{300} \times 7{,}500 \text{ feet},$$

i.e., 250 feet.

The true height is therefore 7,750 feet.

Airy's Rule may be stated in a similar form for Fahrenheit temperature. The standard temperature for the isothermal altimeter is 50° F., which is equal to 509° Absolute Fahrenheit. Equation (3) must therefore be replaced by

$$\frac{h}{h_1} = \frac{T'}{509} \quad \ldots\ldots\ldots\ldots\ldots \quad (5)$$

where T' is the mean temperature of the air column in degrees Absolute Fahrenheit. The round number 500 may be substituted for 509, giving

$$h = \frac{h_1 T'}{500} \quad \ldots\ldots\ldots\ldots\ldots \quad (6)$$

Airy's Rule then takes the form :—

"When the mean temperature of the air between the ground and aircraft level differs from 50° F., the indicated height must be $\left\{\begin{matrix}\text{increased}\\\text{decreased}\end{matrix}\right\}$ by its $\frac{1}{500}$ part for each degree by which the mean temperature $\left\{\begin{matrix}\text{exceeds}\\\text{falls short of}\end{matrix}\right\}$ 50° F."

Example.—Indicated height, 5,000 feet. Mean temperature, 30° F., *i.e.*, 20° short of 50° F. The indicated height must therefore be decreased by

$$\frac{20}{500} \times 5{,}000 \text{ feet},$$

i.e., 200 feet.

The true height is therefore 4,800 feet.

(*b*) *When the temperature at the ground only is known.*—Unless a strut thermometer is carried it is not possible to determine the mean temperature below the level of the aircraft. The temperature at ground level may be read before the commencement of the flight and it may be assumed that below aircraft level the temperature decreases at the rate 1·6° C. per 1,000 feet of indicated height. The mean temperature between the ground and the aircraft is then the temperature at half the indicated height, and may be used in applying Airy's Rule.

Example.—Indicated height, 4,000 feet. Temperature at ground level, 6° C.

Mean temperature = temperature at 2,000 feet,
= 6 − 2 × 1·6° C.,
= 2·8° C.

This falls short of 10° C. by 7·2° C.

The indicated height must therefore be decreased by

$$\frac{7{\cdot}2 \times 4{,}000 \text{ feet}.}{300}$$

= 96 feet,

i.e., true height = 3,904 feet.

(*c*) *When the temperature at the level of the aircraft only is known.*—Even if a strut thermometer is carried it is not always expedient to take the temperature at intervals in order to determine the mean temperature up to the aircraft level. But the temperature at that level can be read immediately and the same assumption as in (*b*) may be used to determine the mean temperature, *i.e.*, the temperature is assumed to increase from the aircraft downwards at the rate 1·6° C. per 1,000 feet of indicated height. The correction may then be worked out as in (*b*).

Isothermal Altimeter—Corrections for Variations of M.S.L. Pressure.—(*a*) *Zero adjustment before commencement of flight.*—The reading of an altimeter at rest on the ground changes continually

owing to the time variation of M.S.L. pressure. The height scale is therefore adjustable so that its zero can be brought into coincidence with the pointer at any position within the range corresponding to the normal variation of barometric pressure. If the instrument is required to indicate height above the ground at the station it is set to zero before the aircraft leaves the ground. If it is required to read height above M.S.L. the scale is set so that the pointer indicates the altitude of the station. This adjustment is possible owing to the fact that the height scale is uniformly graduated.

(*b*) *Zero adjustment during flight for M.S.L. pressure at destination.*—The zero of the altimeter in an aircraft in flight will change on account of the variation of M.S.L. pressure between the departure and destination stations. Assume that, as in the case of aircraft operating from ships, the stations are situated at M.S.L., the altimeter being set to zero before the commencement of the flight will continue to read zero whenever it is subject to a pressure equal to the M.S.L. pressure at the point of departure. If the pressure at the destination is lower than that at the point of departure the altimeter will, on arrival read higher than at departure, the error amounting to 100 feet for each $3\frac{1}{2}$ millibars pressure difference. Similarly, if the pressure at the destination is higher than at departure the altimeter will, on arrival, read low by 100 feet for each $3\frac{1}{2}$ millibars of pressure difference.

If the aircraft is equipped with W/T the pressure at the destination may be requested shortly before arrival, and the adjustment made. In other cases the pressure at the destination may be ascertained from the current synoptic chart, allowance being made, if necessary, for the changes of pressure due to the movements of depressions and anticyclones during the time of flight.

Example:—Pressure at departure station = 1,020 mb. Estimated pressure at destination on arrival = 1,010 mb. The zero adjustment should, therefore, be such as to make the altimeter read lower by:—

$$\frac{1{,}020 - 1{,}010}{3\frac{1}{2}} \times 100 \text{ feet.}$$

$$= 286 \text{ feet.}$$

The more important case is that in which the aircraft is flying towards the lower pressure. Not only does the altimeter read too high, but weather conditions are often adverse, low cloud, rain and bad visibility being frequently associated with low pressure.

The I.C.A.N. Altimeter.—Some altimeters are calibrated according to the I.C.A.N. (International Commission for Air Navigation) law. This law represents very closely the average variation of temperature with height in temperate latitudes. The corrections to the I.C.A.N. altimeter are therefore very much smaller than in the case of the isothermal altimeter, but their calculation is more complicated.

The I.C.A.N. law is based on a temperature of 15° C. at the ground, decreasing uniformly with height at the rate of 1·98° C. per 1,000 feet to — 56·5° C. at 36,090 feet, above which level it remains constant at — 56·5° C.

The International Standard Atmosphere.—The I.C.A.N. law defines an international standard atmosphere for the comparison of aircraft performance in different countries. The following assumptions are made in defining the international standard atmosphere :—

(*a*) The air is dry and its chemical composition is the same at all altitudes.

(*b*) The value of gravity is uniform and equal to 980·62 cm./sec/sec

(*c*) The temperature and pressure at M.S.L. are 15° C. and 1,013·2 mb. (760 mm. of mercury at 0° C.).

(*d*) At any altitude Z (metres) measured above M.S.L. and between 0 and 11,000 metres the temperature of the air is equal to $15 - 0{\cdot}0065Z$ degrees Centigrade.

(*e*) For altitudes above 11,000 metres the temperature of the air is constant and equal to — 56·5° C.

The following table gives the pressure and temperature for a selection of heights above M.S.L. for the international standard atmosphere.

Height. feet.	Pressure. mb.	Temperature. °F.	Temperature. °C.
0	1013·2	59·0	15·0
2,000	942·1	51·8	11·0
4,000	875·1	44·8	7·1
6,000	811·9	37·6	3·1
8,000	752·5	30·4	—0·9
10,000	696·9	23·4	—4·8
12,000	644·3	16·2	—8·8
14,000	595·1	9·1	—12·7
16,000	549·0	1·9	—16·7
18,000	505·7	—5·3	—20·7
20,000	465·3	—12·3	—24·6
25,000	375·8	—30·1	—34·5
30,000	300·8	—47·9	—44·4
35,000	238·2	—65·7	—54·3
36,090*	226·1		
40,000	187·4	—69·7	—56·5
45,000	147·3		
50,000	115·9		

* Tropopause.

Temperature.

Methods of Observation.—The fact that the temperature of the atmosphere near the earth's surface decreases with increasing altitude has long been known, but it is only since the development of flying that daily observations of temperature in the upper air have become available. During the past twenty years observations have been made with increasing regularity in many countries, and our knowledge of the upper-air conditions over a large part of the land areas of the globe is reasonably complete. Over the sea, however, the information at present available is very scanty. At the present day, three principal methods are employed for the observations, the aeroplane ascent, the registering or sounding balloon and the kite. In the aeroplane method (described in detail in Chapter IV), wet- and dry-bulb thermometers are carried and readings are made at selected levels. The sounding balloon and the kite ascents employ an instrument known as a meteorograph designed to give autographic records of pressure, temperature and sometimes humidity in the upper air. The instrument used in the British Meteorological Service is the Dines Balloon Meteorograph. A description of the instrument, together with instructions for its use at sea, is given in Chapter IV. On land the meteorograph is attached to a single balloon which, when fully inflated, has a diameter of about 6 feet. It ultimately returns to earth, owing either to the balloon bursting at a great height or sinking gradually as a result of leakage of gas. The meteorograph is protected from damage by a light bamboo framework known as the "spider," within which it is suspended. A label attached to it requests the finder to return the instrument to the Meteorological Office. A high percentage of records is ultimately recovered, but often not until days or even weeks have elapsed since their release. The method therefore does not permit of the results of the ascents being made available to the daily forecasting service, but it provides much of the information upon which present-day knowledge of the upper air is based.

The kite method employs a meteorograph similar to that used with the sounding balloon, but the ascents are limited to 10,000–15,000 feet, as compared with the 30,000 feet, or more, often attained by the balloon. Aeroplane ascents are generally carried out to 20,000–25,000 feet.

The General Distribution of Temperature in the Upper Air.—The results of the numerous observations of upper-air temperature now available show that up to a certain height temperature generally decreases at an approximately uniform rate from the surface upwards. The fall of temperature with height is called the **Lapse**; the rate of fall, the **Lapse-Rate**. The mean value of the lapse-rate is 3° F. per 1,000 feet, with maximum values at about 20,000 feet and near the surface. The lapse-rate may, however, deviate

considerably from its mean value in atmospheric layers of limited thickness.

Individual soundings show that at a certain height the lapse of temperature terminates abruptly, and above that height the temperature remains practically constant. The height at which the change occurs varies from one ascent to another at the same place, so that if mean values of upper-air temperature are computed from a large number of ascents they tend to mask the abrupt nature of the transition. The lower region of the atmosphere, characterized by a lapse of temperature is called the **Troposphere,** the upper isothermal region the **Stratosphere** and the boundary between them, the **Tropopause.** The height of the tropopause varies with latitude, being greatest at the equator and least at the poles. Since the lapse of temperature is continued up to the level of the tropopause, a high tropopause corresponds to low temperature in the stratosphere, and *vice versa.* At the equator the height of the tropopause is approximately double that at the poles, and therefore the lowest temperatures in the atmosphere occur in the stratosphere of equatorial regions despite their high surface temperatures. The following table shows the mean height of the tropopause and the mean temperature of the stratosphere, indicating also their variation with latitude.

	Height of Tropopause.	Temperature in Stratosphere.
	feet	°F.
Equator	50,000	−100
Temperate latitudes (British Isles)	35,000	−70
Poles	20,000	−60

The stratosphere is generally higher and somewhat warmer in summer than in winter, but its variation of temperature is comparatively small. It is higher and colder over anticyclones than over depressions, the difference of height amounting to 5,000–10,000 feet. Consequently, the height is subject to irregular variations associated with the large fluctuations of pressure of the anticyclones and depressions of temperate latitudes. In the troposphere the lapse-rate is slightly greater in summer than in winter, but the difference is very small, so that the annual variation of temperature is nearly the same at all levels. It tends to a maximum at 20,000 feet at coastal and sea stations ; at continental stations owing to the predominating influence of radiation it is greatest at the surface. The diurnal variation of temperature is confined to the lowest layers. It diminishes rapidly with height and disappears

between 3,000 and 6,000 feet. The effect of height on the diurnal range is indicated by the following table :—

Eiffel Tower.	Diurnal range of temperature.	
	January.	July.
	°F.	°F.
Base	6·5	16·5
Top	2	10

The formation of cloud, precipitation and most weather phenomena take place entirely within the troposphere.

Changes of Temperature in Unsaturated Air Subjected to Vertical Displacements.—In order to explain the observed lapse of temperature in the troposphere, as well as many weather phenomena, it is necessary to consider the behaviour of a small isolated parcel of air when moved upwards or downwards through the surrounding atmosphere. The parcel of air is considered to be so isolated that it does not lose its identity through intermixture with its environment. This condition is normally not satisfied in the atmosphere, but it can be regarded as applying to the interior of a large mass of air such as forms a detached cumulus cloud. Also no interchange of heat is assumed to take place between the parcel of air and its environment. This condition is never exactly fulfilled, for the air will lose or gain heat by intermixture, contact and radiation. These factors are, however, neglected, a step which is justified by the applicability of the results of the argument which follows, in everyday practice. Changes in the condition (temperature, pressure and volume) of the air which take place without heat transfer are said to be **adiabatic,** or, the air mass is subject to adiabatic conditions. It is evident that the displaced air mass immediately and automatically takes up the pressure of its surroundings.

Firstly, consider a small mass of perfectly dry air to be displaced vertically upwards. Its pressure diminishes and its volume increases in consequence. The surrounding air must be pushed aside necessitating the expenditure of mechanical energy. Since the air is subject to adiabatic conditions it can acquire no energy in the form of heat from external sources for the performance of the mechanical work and the expansion can take place only at the expense of its own internal energy. The loss of internal energy is manifest as a fall of temperature of the displaced air mass, and its amount depends upon the change of pressure. If the original pressure and temperature (Absolute) of the air are p_1 and T_1 and the final

pressure and temperature p and T respectively these quantities are related by the formula

$$\log T = \log T_1 - 0{\cdot}286 \log \frac{p_1}{p} \quad \ldots\ldots\ldots\ldots \quad (1)$$

It is, however, simpler and more convenient to express the change of temperature in terms of height. The relation is practically a linear one, that is, the rate of change of temperature with height is almost constant and the graph of temperature with height is practically a straight line. The rate of change is 5·4° F. per 1,000 feet so that if T_1, h_1 represent the original temperature in degrees Fahrenheit and height in feet respectively, and T, h the final temperature and height:

$$T = T_1 - 0{\cdot}0054\,(h - h_1) \quad \ldots\ldots\ldots \quad (2)$$

This relation is exact, however, only if the rising air has everywhere the same temperature as its immediate surroundings, a condition which does not usually obtain owing to the difference between the normal lapse-rate of 3° F. per 1,000 feet and the above value for the rate of change of temperature in the rising air mass. Nevertheless, so long as the temperature of the rising air does not differ by large amounts (say, 20° F. or more) from its surroundings the above relation is very approximately satisfied, and may be applied to changes of height up to about 10,000 feet. For greater displacements it is necessary to use the pressure-temperature relation (1) above converting heights to pressures by means of the formula on page 140. In the discussion which follows changes of temperature in a rising air mass will be related to height and represented graphically with respect to axes graduated uniformly for temperature and height as in Fig. 38. Such a graph is called a temperature-height graph. The adiabatic changes of temperature in a rising mass of dry air are represented by straight lines, such as AB in Fig. 38, having a constant backward slope of 5·4° F. per 1,000 feet. The limitations of the accuracy of this representation must however, be borne in mind. The straight lines are called **Dry Adiabatics.**

If the small mass of dry air is displaced downwards it undergoes compression, and mechanical work is performed upon it. The energy thus acquired cannot be disposed of by heat transfer, so that the internal energy of the air increases, that is, its temperature rises. The rate of change of temperature is equal to that in the case of rising air, viz., 5·4° F. per 1,000 feet, exactly for small displacements, approximately for large displacements. The same dry adiabatics on the temperature-height graph therefore represent the changes of temperature of dry air moving either upwards or downwards.

If the air contains water vapour its changes of temperature consequent upon vertical displacement are the same as if the air were perfectly dry, provided that at no stage of the process does the air become saturated. Dry adiabatics therefore represent the

temperature changes in unsaturated air, that is, ordinary atmospheric air so long as it remains unsaturated.

If the actual temperature distribution in the atmosphere is such that a rising mass of unsaturated air has everywhere the same temperature as its environment, the lapse-rate of the atmosphere is said to be dry adiabatic. The dry adiabatic lapse-rate is therefore 5·4° F. per 1,000 feet. The change of temperature due to vertical movement of air is often called dynamical heating or cooling.

Potential Temperature.—The changes of temperature of a mass of unsaturated air whose pressure p and temperature T correspond to the point C on the temperature-height graph of Fig. 38 are

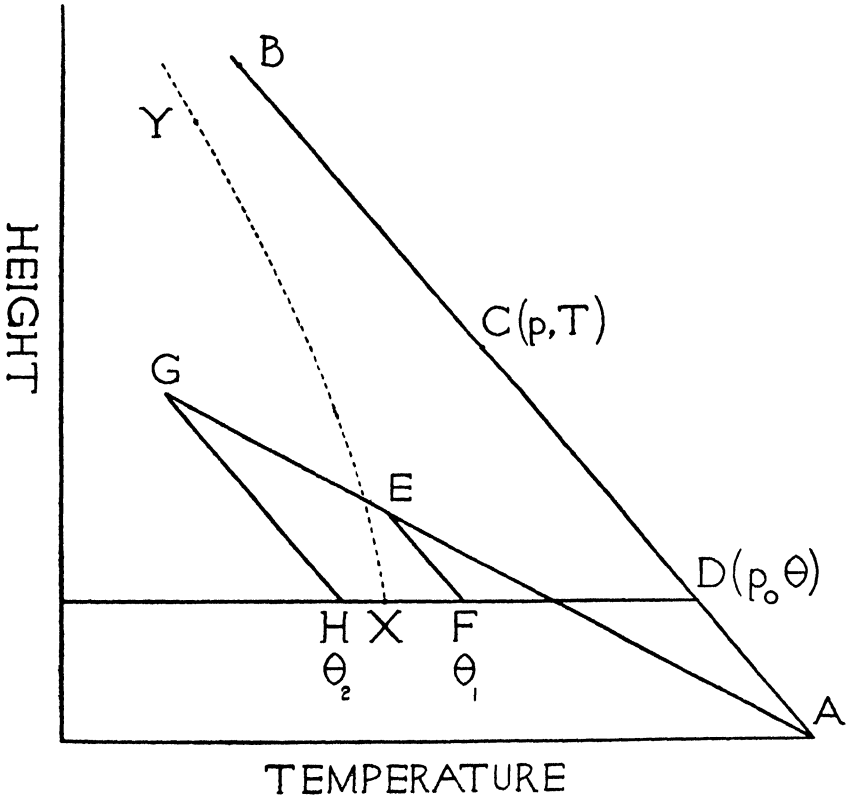


FIG. 38.—POTENTIAL TEMPERATURE.

represented by the movement of C along the dry adiabatic AB drawn through C. Let D be the point on AB corresponding to a certain standard pressure p_0 and temperature θ. Then, however the air is moved, up or down, whenever it is brought to the pressure p_0 it simultaneously takes up the temperature θ, which is called its **Potential Temperature.** In practice the standard pressure p_0

is taken as 1,000 mb. and the definition of potential temperature is therefore: "The potential temperature of a mass of unsaturated air is the temperature which it would assume if brought adiabatically to the pressure 1,000 mb."

Let p, T be the initial pressure and temperature of the air.
,, p_0 be the standard pressure.
,, θ be the potential temperature of the air.

Then $\log \theta = \log T - 0{\cdot}286 \log (p/p_0)$ ..... (1)

Logarithms are taken either to base e or base 10.

In the latter case, taking $p_0 = 1{,}000$ mb.

$$\log \theta = \log T - 0{\cdot}286 \log p + 0{\cdot}858 \quad \ldots\ldots\ldots (2)$$

Temperatures are in degrees Absolute.

The potential temperature of a mass of unsaturated air remains constant whatever be the changes of pressure (or height) to which it is subject, so long as these changes are adiabatic. For small height intervals the calculation of potential temperature may be carried out using height instead of pressure thus:—

Let h, T be original height and temperature of the air expressed in feet and degrees Fahrenheit.

Let h_0 be the height at which the pressure is 1,000 mb.

Then $\theta = T + 0{\cdot}0054\,(h - h_0)$ ..... (3)

This formula is subject to the same limitations as (2) on page 149, and moreover necessitates the calculation of h_0 by means of the pressure-height formula.

If the existing lapse-rate in the atmosphere is the dry adiabatic (D.A.L.R.), the line AB of Fig. 38 represents the temperature distribution, and the potential temperature is constant at all heights. If the existing lapse-rate is greater than the D.A.L.R. the vertical distribution of temperature is represented by a line AG, whose slope is less than that of AB. Let E, G correspond to two samples of air at different levels. Their potential temperatures, θ_1 and θ_2 respectively, are found by drawing dry adiabatics EF, GH to the 1,000 mb. level. θ_2 is clearly less than θ_1, so that the potential temperature decreases with height. Similarly it may be shown that if the lapse-rate is less than the D.A.L.R., which is the normal condition in the atmosphere, the potential temperature increases with height.

Changes of Temperature in Saturated Air Subject to Vertical displacements.—The changes of temperature consequent upon vertical displacement of saturated air differ according as the air descends or rises, that is, according as its pressure increases or decreases. If the air is displaced downwards, being heated on compression, its capacity for water vapour increases and it becomes no longer saturated. On displacement upwards, the air is immediately cooled below its dew-point, and the excess water vapour condenses with the production of latent heat. This heat is taken

up by the air and partly makes good the loss of internal energy due to expansion, so that the fall of temperature is less than if the air were unsaturated. For a given change of pressure (or height) the fall of temperature, however, varies with the original temperature of the air. This may be explained by reference to Fig. 34, which shows that if, for example, one cubic metre of air undergoes a reduction of temperature of 20° F. from 80° F. to 60° F., 12·7 grams of water vapour condense. For the same reduction of temperature, 20° F. from 40° F. to 20° F. only 4·5 grams of water vapour condense. That is, a given reduction of temperature is accompanied by an amount of condensation which increases with the temperature at which the process takes place. The amount of latent heat released is proportional to the amount of water vapour condensed. It follows that if a given volume of saturated air expands, its resulting fall of temperature is smaller the higher the temperature of the air, and *vice versa*. At low temperatures the rate of fall differs little from the dry adiabatic. Its value varies from 2·5° F. per 1,000 feet at a temperature of 60° F. to 5° F. per 1,000 feet at — 40° F., the average value throughout the atmosphere being 3° F. per 1,000 feet.

The changes of temperature in a rising saturated air mass are said to be saturation adiabatic. If the existing variation of temperature with height is such that a rising saturated air mass would have everywhere the same temperature as its surroundings, the lapse-rate is said to be **Saturation Adiabatic.** The saturation adiabatic lapse-rate (S.A.L.R.), unlike the fixed D.A.L.R., varies with the temperature, being least at high temperatures and therefore low altitudes, and greatest at low temperatures and high altitudes. On the temperature-height graph the variations of temperature of rising saturated air are represented by saturation adiabatics, which are curved lines, more vertical than the dry adiabatics at high temperatures and becoming almost parallel to them at low temperatures. A typical saturation adiabatic is the pecked line *XY* of Fig. 38.

When a mass of saturated air rises, the condensed water appears first as cloud and later as some form of precipitation. If it is assumed that all the water falls out of the air as it is condensed, it is possible for air to be heated to a considerable degree by being carried upwards and then restored to its original level. Suppose the air is originally saturated and carried up 10,000 feet. The reduction of temperature at the rate of 3° F. per 1,000 feet is 30° F. If, now, the air is brought back to its original level, its temperature increases at 5½° F. per 1,000 feet, that is by 55° F., and the net rise of temperature is 25° F. The source of the heat thus acquired is the latent heat of the precipitated water.

Stable and Unstable Atmosphere.—We proceed to examine the behaviour of a small mass of air which, originally at rest in the

atmosphere, is given a small upward displacement. In the first instance the air will be considered to be unsaturated. In Fig. 39

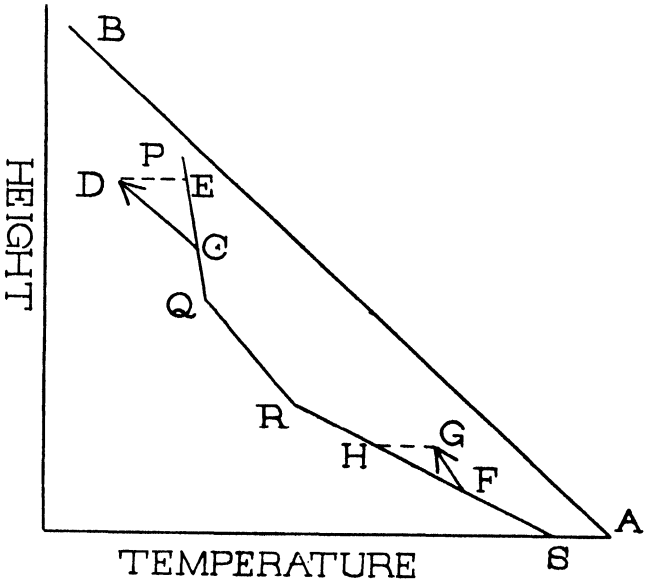


FIG. 39.—LAPSE-RATE IN STABLE AND UNSTABLE ATMOSPHERE.

let *PQRS* represent the existing distribution of temperature and let *AB* be a dry adiabatic.

(*a*) *Lapse-rate less than D.A.L.R.*—In the layer represented by *PQ* of Fig. 39, let the lapse-rate be less than D.A.L.R., so that *PQ* is more nearly vertical than *AB*. Let a small mass of air represented by *C* in the layer *PQ* receive a small upward displacement to the level *D*. Its change of state is represented by *CD*, a dry adiabatic, and therefore at *D* its temperature is lower than the temperature of the air surrounding at its own level, represented by *E*. The air automatically takes up the pressure of its surroundings; its density must, therefore, be greater, so that if left to itself the displaced air will tend to fall back to its original level. Similarly, if the air at *C* receives a small downward displacement it finds itself amid surroundings colder and denser than itself, and therefore, owing to buoyancy, rises to its original level. Any other small mass of air within the layer *PQ* will behave similarly. If slightly displaced in the vertical it tends to return to its original level and the layer *PQ* is said to be a stable layer. The criterion for stability is that the lapse-rate shall be less than the D.A.L.R.

(*b*) *Lapse-rate greater than D.A.L.R.*—In the layer represented by *RS* of Fig. 39, let the lapse-rate be greater than the dry adiabatic, so that *RS* is more nearly horizontal than *AB*. Let a small mass

of air represented by *F* in the layer *RS* receive a small upward displacement to the level *G*. At *G* its temperature is higher than that of the surrounding air at the same level *H* and therefore, owing to buoyancy, the air, if left to itself, will continue to rise. Similarly, if the air at *F* is displaced downwards it becomes colder and heavier than its surrounding, and if left to itself will continue to sink further from its original position. The atmospheric layer *RS* is said to be an unstable layer, and the condition for instability is that the lapse-rate shall be greater than the D.A.L.R.

(*c*) *Lapse-rate equal to D.A.L.R.*—In this case, *e.g.* the layer *QR*, air displaced either upwards or downwards is at the same temperature as its surroundings, and there is no tendency for it to move in either direction when left to itself. The layer *QR* is therefore said to be in neutral equilibrium. The above criteria for atmospheric stability apply to unsaturated air, and the critical lapse-rate is the D.A.L.R. They may be extended to saturated air if the S.A.L.R. is taken as the critical lapse-rate and only upward motion is considered. An atmospheric layer is therefore unstable, stable or in neutral equilibrium to saturated air according as the lapse-rate is greater than, less than, or equal to the S.A.L.R.

In the foregoing discussion the disturbance imposed on the air has been assumed to take the form of a small vertical displacement, but it will be readily seen that the same stability criteria apply if the disturbance takes the form of a slight rise of temperature of a small mass of air, as compared with its surroundings.

Convection.—The ascent of air is known as convection. It may be either forced as, for example, when an air stream ascends in crossing a mountain barrier, or spontaneous as in an unstable atmosphere. In either case the air is usually unsaturated to begin with, and during the first stages of its ascent follows a dry adiabatic. Sooner or later, however, as a result of the lowering of its temperature, it becomes saturated, and thereafter follows a saturation adiabatic with condensation of the contained water vapour. The curve on the temperature-height diagram representing the temperature changes of rising air is known as the **Path Curve** and generally consists of a dry adiabatic below the level at which condensation begins and a saturation adiabatic above that level. The curve representing the actual variation of temperature with height in the atmosphere is known as the **Environment Curve** and its form varies with time and place.

The path curve in conjunction with the environment curve indicates to what extent convection, once started, will go on spontaneously, or to what level the rising air will freely ascend. The necessary condition for free convection is that the temperature of the rising air is higher than that of its surroundings; in other words that the path curve lies to the right of the environment curve. In Fig. 40 let *ABC* be the environment curve and *AXY* the path curve for the air at the surface, consisting of the dry

adiabatic AX below the condensation level X and the saturation adiabatic XY above that level. Below the level C the path curve lies to the right of the environment curve and the air at A,

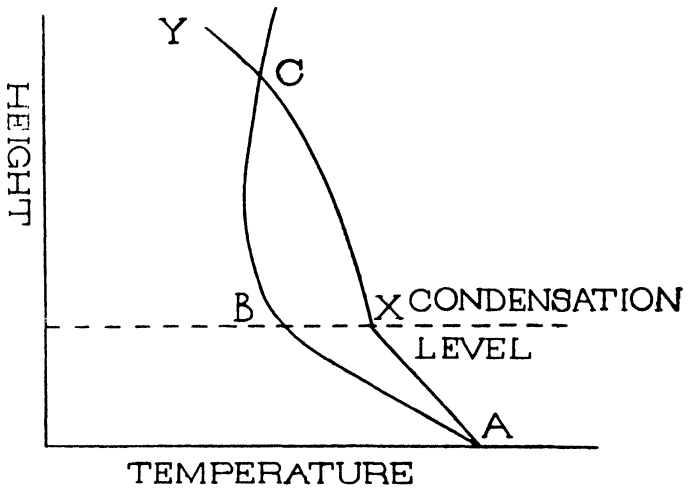


FIG. 40.—ENVIRONMENT CURVE AND PATH CURVE.

if disturbed, would ascend spontaneously to that level. If it penetrated beyond C, finding itself amid warmer surroundings, it would soon come to rest. Practically then, convection is limited to the atmosphere below the level of C.

Under the most unstable conditions that might exist in the troposphere, convection could never penetrate far above the tropopause, for in the stratosphere the environment curve becomes vertical and must sooner or later be crossed by the path curve of any air rising by convection from the lower levels. It is for this reason that weather phenomena in which convection plays an important part, are confined to the troposphere.

The condition for instability in the atmosphere is that below the condensation level the lapse-rate shall be greater than the D.A.L.R., and above that level greater than the S.A.L.R. An atmospheric layer in which the lapse-rate lies between these two is said to be conditionally unstable since the instability is conditional upon the presence of saturated air.

The initial disturbance necessary to bring about convection in an unstable atmosphere usually takes place at the earth's surface, either through heating by contact of the air with a warm land or sea surface, or by the forced ascent of air in crossing mountains or other surface irregularities. Sometimes the initial ascent is brought about owing to the undercutting of warm air by an adjacent mass of colder, heavier air, as at a cold front (Chapter XVIII.). These processes are known as " trigger action."

Convection normally takes the form of localized air currents rising more or less vigorously, while the compensating downward flow is spread over a large area as a gradual sinking or settling of the atmosphere. The upward currents are the cause of the impulsive uplift experienced by aircraft flying into them, known as " bumpiness." Above the condensation level the rising currents are rendered visible by the cumulus clouds which form within them. These clouds are usually detached, with flat bases at the condensation level and rounded " cauliflower " or " woolpack " tops marking the upper limit of the rising air as it penetrates the surrounding atmosphere. If convection goes on to great heights, the cloud takes the form of the towering cumulonimbus, or thundercloud, and the condensed moisture, becoming in excess of the amount that the air can hold in suspension as cloud, is precipitated as rain, snow or hail. Since the precipitation usually falls from detached clouds, it is of the localized shower type. If the upward movement is sufficiently violent, the water-drops are broken up, a process which results in the separation of electrical charges and gives rise to the thunderstorm. Forced convection also gives rise to cloud and precipitation, but since it is not localized, but involves the ascent of a large air mass over a large area, the cloud is formed in a more or less continuous layer with widespread and persistent precipitation. These various convection phenomena receive fuller discussion in the appropriate chapters.

Inversions.—The normal decrease of temperature with height is sometimes reversed in atmospheric layers of limited thickness. In such a layer there is said to be an inversion of temperature or, simply, an inversion. On the temperature-height diagram an inversion is represented by the environment curve sloping upwards towards the right. Inversions may occur from ground level upwards as at *PQ* in Fig. 41, or in a layer situated above the earth's surface with the normal lapse of temperature above and below it as at *BC*. The latter type often extends over a large area, and the layer is horizontal or nearly so. The thickness of effective inversions varies from a few hundred to a few thousand feet. A layer in which the temperature remains constant with height, represented by a vertical portion of the environment curve, is called an **Isothermal Layer.** Its properties are similar to those of the inversion, so that what follows in connection with inversions may also be regarded as applying to isothermal layers. The most important feature of an inversion is its pronounced stability. On page 153, atmospheric layers have been regarded simply as stable or unstable, without reference to the degree of stability or instability. In Fig. 39 *QP* is a stable layer, and air within it, at *C*, on rising to the level *D* becomes cooler than its surroundings by an amount *DE*. Now, the smaller the lapse-rate in *QP* the greater does *DE* become ; if *QP* becomes an inversion (sloping upwards towards the right), *DE* increases still further. Since *DE* measures the tendency of the

lifted air to return to its original position, this tendency increases as the lapse-rate diminishes, and becomes very pronounced within an inversion. Consequently, an inversion is a very persistent state of an atmospheric layer, disturbances within it being quickly damped out and vertical motion much restricted.

An inversion is also effective in restricting convection originating in unstable layers below it. A path curve for rising air, such as *AXY* of Fig. 41, if it lies to the right of the environment curve in the lower layers, is almost certain to cross it within the inversion which is therefore generally impenetrable by freely rising air. The

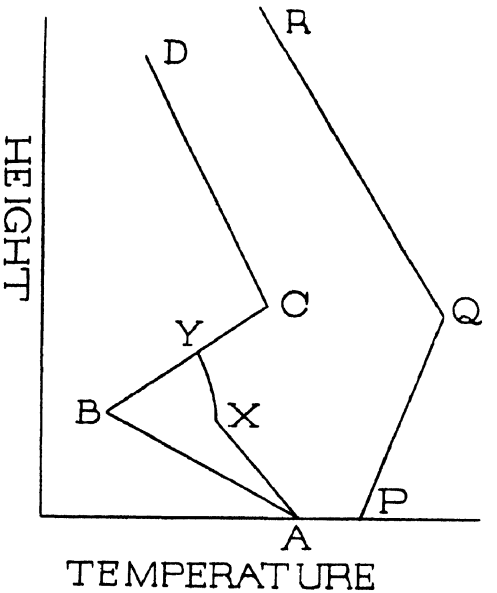


FIG. 41.—DIAGRAM ILLUSTRATING INVERSIONS.

rising air currents spread out horizontally below the inversion and if bearing cloud the latter spreads out as a flat sheet. Such cloud sheets often occur immediately below inversion layers. Over cities and industrial areas where atmospheric pollution by smoke is frequent, an inversion prevents the escape and dissipation of smoke into the upper levels, tending rather to concentrate it near the surface and giving rise to bad visibility and gloom. Because it restricts convection in this way, an inversion is sometimes called a "convective lid."

Inversions are formed by the heating of the upper or cooling of the lower layers. Surface inversions form principally over land on calm, clear nights, when the ground is cooled by radiation and the cooling effect is communicated to the air by contact and radiation from the air to the ground. Above the surface, inversions may be formed by the horizontal flow, or advection, of warm air

over cold, or by cold air below warm air, as at warm and cold fronts. In anticyclonic weather they are also formed by the process of subsidence or at the top of a turbulent layer.

Subsidence.—Subsidence is the gradual sinking or settling down of a large body of air over a wide area. A localized form of subsidence occurs in places where the barometric pressure is rising rapidly, the subsidence rate being of the order 200 feet per hour; but in general subsidence, such as occurs in anticyclones, the rate is much smaller (1,000—1,500 feet per day). The sinking causes dynamical warming and if, as is usual, it occurs in middle and high levels, the temperature of the subsided air becomes higher than that of the air below it, forming an inversion. As distinct, however, from those formed between subsiding and non-subsiding air, inversions and more frequently, isothermal layers, are formed within the mass of subsiding air. The process of their formation may be indicated by the following example, in which the values of heights and temperatures have been selected merely for purposes of illustration, and have no special significance. In Fig. 42, let *ABC* represent

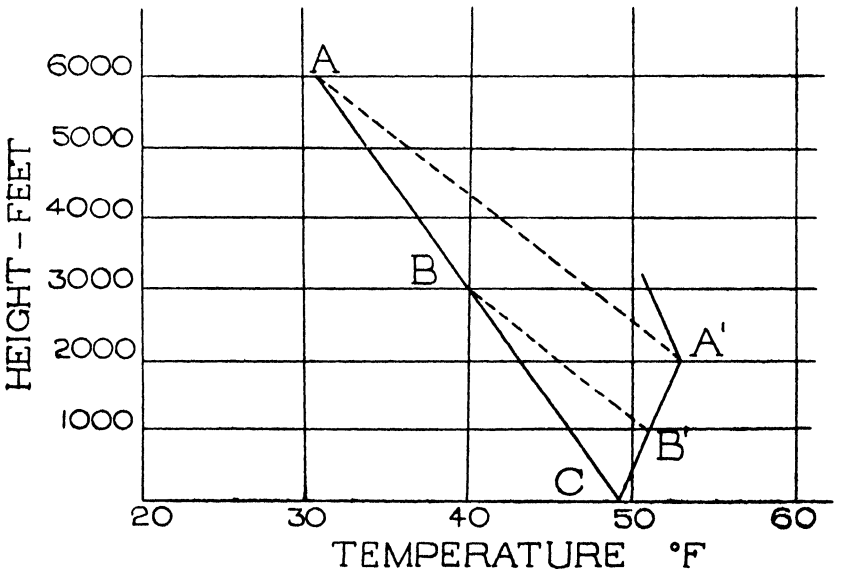


FIG. 42.—DIAGRAM ILLUSTRATING SUBSIDENCE.

the normal distribution of temperature in the lowest 6,000 feet, the lapse-rate being assumed to have the constant value 3° F. per 1,000 feet. Let *AB* represent the temperature distribution in a horizontal layer of air 3,000 feet thick with its base at 3,000 feet. Let the temperature at its base be 40° F., so that the temperature at the top is 31° F. Now let the layer subside until its base is 1,000 feet above the surface. Owing to the increase of pressure

the thickness of the layer will diminish. Since the layer is approaching the earth's surface, at which there can be no downward movement, it will tend to spread out laterally, resulting in a further decrease of thickness. Assume that in the new position the layer is 1,000 feet thick. The temperatures at the base and top may be calculated by adding to the respective original temperatures 5½° F. for each 1,000 feet of descent. Thus :—

Final temperature at base = 40 + 2 × 5½° F. = 51° F.
,, ,, ,, top = 31 + 4 × 5½° F. = 53° F.

In Fig. 42 the temperatures after subsidence are represented by A' and B' at 2,000 and 1,000 feet respectively, AA' and BB' being drawn as dry adiabatics. An inversion is therefore set up in the layer. Even if there is no lateral spreading of the layer, the decrease of thickness due to compression causes the lapse-rate to diminish, *i.e.*, has a stabilizing influence, provided that the original lapse-rate is less than the D.A.L.R.

Inversions may also be formed in subsiding air in the lower part of which there is a layer of cloud, since the cloud-laden air warms at the saturation-adiabatic rate with evaporation taking place, while the cloud-free air above warms at the dry-adiabatic rate. Only slight inversions can be formed by this process, however, because the cloud rapidly disappears, after which all the air warms at the same rate.

Cause of the Observed Vertical Distribution of Atmospheric Temperature.—If the atmosphere were subject to no external influences it would settle down in a steady state with the heaviest, and therefore coldest, air at the bottom, so that the temperature, increasing with height, would have the opposite distribution to that actually observed. But the atmosphere is in fact subject to heating through insolation, and this, as has been seen on page 111, takes place from below by contact with the heated surface of the earth, and by absorption of terrestrial radiation by the water vapour in the lowest layers. Heating at the base increases the lapse-rate. But stability considerations limit the lapse-rate to the dry-adiabatic value for unsaturated air and the saturation-adiabatic for saturated air. As soon as the critical lapse-rate is exceeded, convection begins, and goes on until equilibrium is once more established. The average lapse-rate in the atmosphere therefore is very near to the critical value, and since all air contains moisture, and on rising becomes saturated sooner or later, the critical lapse-rate is that appropriate to saturated air, *i.e.*, the S.A.L.R., whose mean value, 3° F. per 1,000 feet, is the average lapse-rate observed. Below the condensation level, however, the critical lapse-rate is the D.A.L.R., which is approached and occasionally slightly exceeded when the surface heating is intense, as over desert countries and land areas on summer afternoons, and may be very greatly exceeded close to the surface of the ground.

The cause of the existence of the stratosphere is not yet completely understood. Radiation is the controlling influence in the higher levels, and the isothermal conditions in the stratosphere are most probably due to equality of absorbed and emitted radiation. There is some evidence that the temperature undergoes a considerable increase in the higher stratosphere.

Humidity.

Variation of Humidity in the Upper Air.—Absolute humidity decreases with increasing height in the atmosphere, not only on account of the fact that the source of the water vapour is at the earth's surface, whence it is carried by convection and diffusion to the upper layers, but also because of the decreasing capacity of the air for water vapour resulting from the general diminution of temperature with height. The rate of decrease is very irregular and somewhat rapid, and on the average the amount of water vapour is reduced to 1/10th of its surface value at a height of 4,000 feet. The distribution of relative humidity is even more irregular, since it depends upon the temperature of the air as well as its water-vapour content. Within a cloud the relative humidity is equal, or nearly equal to 100 per cent. It is very low in subsidence inversions, since these are formed by dynamical heating without evaporation.

Since moist air is lighter than dry air at the same temperature and pressure, the distribution of water vapour in the atmosphere has some bearing upon its stability. The air in a cloud, with relative humidity 100 per cent., may be slightly colder and yet lighter than its relatively dry surroundings, so that convection goes on rather beyond the level at which the temperature of the rising air becomes equal to that of its environment. Above a cloud sheet a small inversion of temperature is necessary for equilibrium, since otherwise the moist air of the cloud layer would be lighter than the relatively dry air immediately above it.

CHAPTER VIII.

Wind

Introductory.—Wind is the horizontal movement of air over the earth's surface. It is completely specified if its speed and the direction from which it blows are stated, but since these are subject to rather wide and rapid fluctuations, exact specification is difficult. Meteorology is concerned both with the wind at the earth's surface and the wind at various heights above the surface, known as the upper winds. At the surface the direction of the wind is expressed in compass points or in degrees measured from true North ; the speed, either in knots (miles per hour, feet per second, etc.), or as the force on the Beaufort Scale. Upper winds are always expressed in units of speed and degrees from true North, the direction of the wind being always understood as that *from* which it blows.

The speed of a steady wind may be measured by the force (or pressure) which it exerts on a plane surface exposed at right angles to it. Experiments in wind tunnels have led to the following formula for the pressure on a flat plate at right angles to a steady stream of air :—

$$P = A\rho v^2$$

where P = pressure due to the air stream, in pounds per sq. ft.
ρ = density of the air,
v = speed of air stream in m.p.h.,
A = a constant depending upon the shape and size of the plate.

For a circular plate about one square foot in area and for air of normal density, P is equal to $0{\cdot}003\ v^2$.

Expressed in terms of the Beaufort force F the formula is :—

$$P = 0{\cdot}0105\ F^3.$$

This leads to a relation between the Beaufort force and the corresponding mean wind speed in m.p.h., viz.,

$$v = 1{\cdot}87\sqrt{F^3}.$$

The Measurement of Wind.—The methods of wind measurement at sea are dealt with in Chapter I of this manual, and it is sufficient here to describe briefly the methods in use at land stations. An instrument used for measuring wind is called an Anemometer.

(*a*) *The Dines Pressure Tube Anemometer.*—This is the land counterpart of the Munro wind-speed recorder used in H.M. ships. It consists of a Dines pressure head and vane (Fig. 12), but differs from the ship's instrument in the type of recorder used. This is shown diagrammatically in Fig. 43.

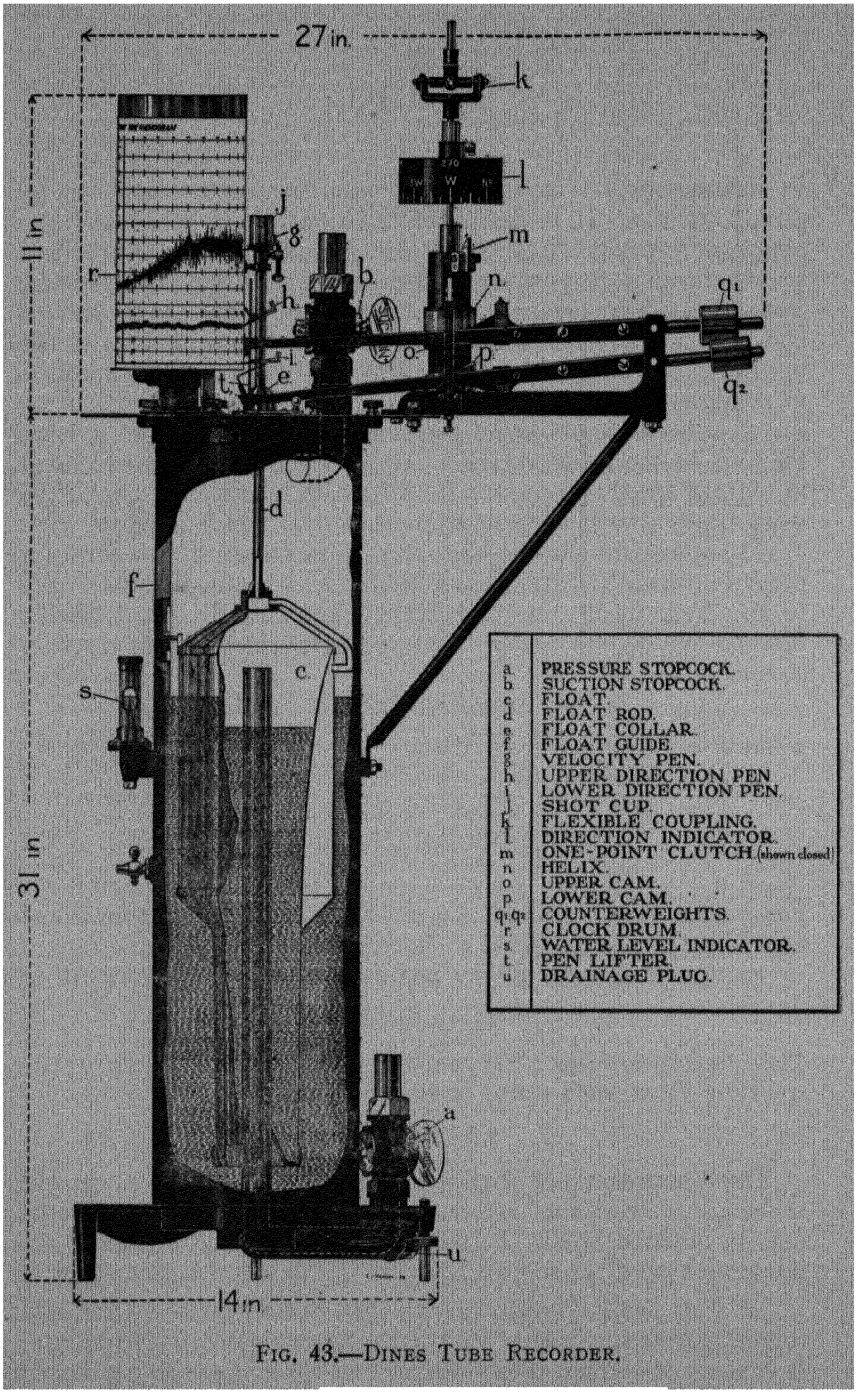


FIG. 43.—DINES TUBE RECORDER.

The recorder consists of a float *c*, which is placed mouth downwards in a closed tank partially filled with water. The pressure and suction tubes at *a* and *b*, from the head, lead into the air spaces below and above the float respectively. The increase and decrease of pressure on the two sides of the float combine to raise the float, and a record of the motion of the latter is thus a record of the wind blowing past the head. To the top of the float is attached a vertical rod *d* which passes through an air-tight collar *e* in the cover of the tank; this rod carries a pen which records on a drum *r*, rotated by clockwork. To the float is attached a small wheel which moves in a vertical guide *f* and prevents the float from rotating.

The shape of the inner surface of the float is of great importance. It is so constructed that the displacement of the recording pen from the zero of the scale is proportional to the speed of the wind at the head. The combination of the pressure and suction effects in the manner described increases the motion of the float, but its primary object is to eliminate the effects of changes in the pressure of the air in the room in which the recorder is placed.

For recording wind direction, the movements of the vane are transmitted through the supporting mast by means of a rigid rod to the stand of the recording drum, vertically beneath. By an apparatus designed by Mr. W. H. Dines, F.R.S., the motion is recorded on the paper which takes the speed record. On the recording sheet the directions are set out between two horizontal lines, both of which are marked N., and which form the top and bottom lines of the direction chart. The variation of the direction of the vane is transformed into motion up or down the sheet by means of a cam. Two pens are employed, one or other of which moves with the vane, even if the vane makes more than a complete revolution.

(*b*) *The Cup Anemometer.*—The cup anemometer usually consists of four hemispherical cups attached to the ends of two crossed metal arms. The cross is pivoted at its central point in such a way that it is free to rotate in a horizontal plane. The difference of pressure of the wind on the convex and concave surfaces of the cups causes the cross to spin round. In some modern instruments three cups are used instead of four. The number of revolutions is recorded on a dial showing miles run of wind. The run of the wind during the interval between two successive readings may therefore be found by subtraction. By depressing a lever, a subsidiary counting apparatus, reading in tenths of a mile, and therefore suitable for measuring the mean speed in a short interval such as 5 or 10 minutes, can be brought into play. Experience shows that an instrument of this pattern tends to read too low at low speeds and too high at high speeds. Tables are supplied showing the corrections to be applied to wind speeds, based on the number of revolutions of the cups in a given time.

In the electrical form of instrument, shown in Fig. 44, the cups turn a contact maker which completes the circuit and rings a bell, or buzzer, once in every 25 turns of the cups. The mean wind speed may be obtained by timing with a stop watch the interval between successive rings on the bell.

Owing to its inertia, the cup anemometer does not respond to the short-period fluctuations of wind as does the pressure-tube anemometer, whose records show the gusts and lulls in great detail. When the wind drops suddenly, the cups tend to go on rotating, and when it rises they lag in gathering speed. The cup anemometer therefore measures the average wind speed of a short period, say, half a minute, and not instantaneous values.

Wind-Roses.—Since wind is a vector quantity it cannot be averaged for statistical purposes like temperature, pressure and other meteorological elements, and some other method of summarizing winds at a particular station is therefore necessary. The most convenient is that expressing the relative frequency of winds from each direction. The information may be given in tabular form, but for representation on pilot charts the graphical form, that is the wind-rose, is more suitable. A wind-rose is a diagram showing, for a definite locality or district, and usually for a more or less extended period, the proportion of winds blowing from each of the main points of the compass. As a rule, the "rose" indicates also the strength of the wind from each quarter, and the number or proportion of cases in which the air was calm.

The simplest form of wind-rose is shown in Fig. 45, in which

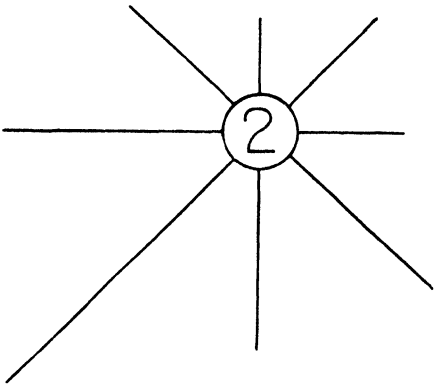


FIG. 45.—SIMPLEST FORM OF WIND-ROSE.

the number or proportion of winds blowing from each of the principal eight points of the compass is represented by lines converging towards a small circle, the proportion of winds from each direction being represented by varying length of the lines. The figure in the circle gives the number or percentage of cases in which the air was calm.

To face page 164.

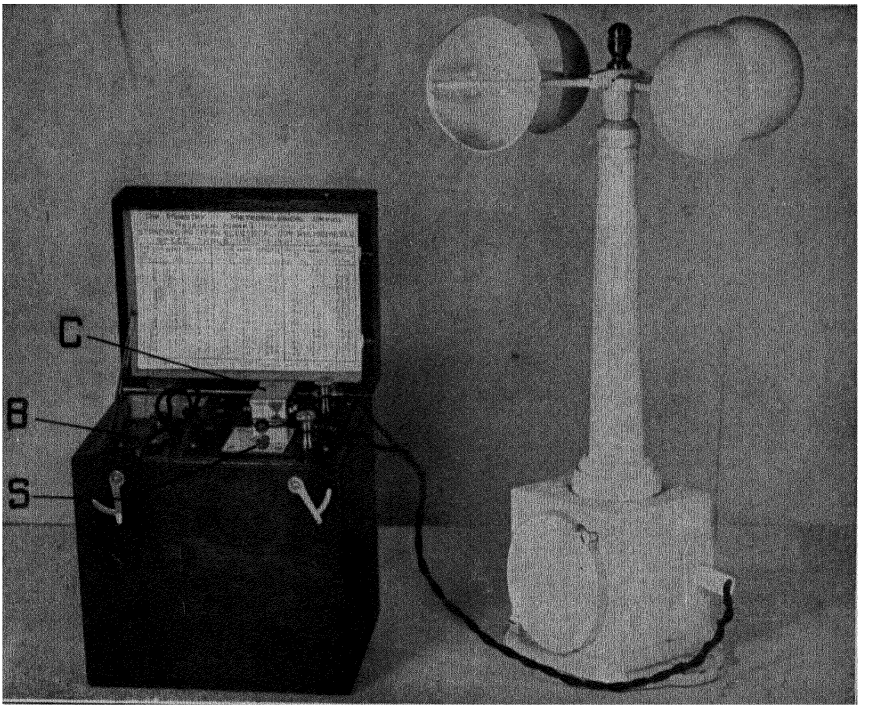


FIG. 44.—ELECTRIC CUP ANEMOMETER AND PORTABLE RECEIVER.
(B = Battery, C = Buzzer, S = Switch).

The Baillie wind-rose was introduced by Nav.-Lieut C. W. Baillie in 1892. In this form of wind-rose, the arrows which fly with the wind show by their length the frequency of winds from various directions and by their thickness the frequencies of the various forces. This principle has been utilized in the octagon wind-rose.

The octagon wind-rose based on the design of Sir Napier Shaw and Captain L. G. Garbett, R.N., shows the percentage frequencies of winds from eight points of the compass, and for different ranges of speed at the surface and at heights of 1,500 feet above the surface and 3,000 feet, 6,000 feet and 10,000 feet above mean sea level. A specimen wind-rose is shown in Fig. 46. The percentage of

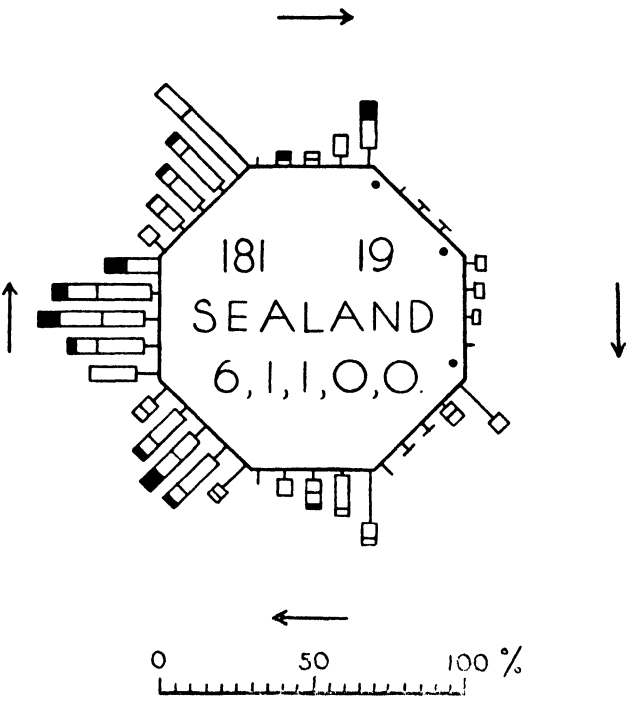


FIG. 46.—OCTAGON WIND-ROSE.

frequencies of winds from each direction are shown by the lengths of the arrows on such a scale that the distances between the opposite sides of the octagon are equal to 100 per cent.

The different ranges of wind speed are indicated as follows :—

———	4—15 m.p.h., 3—13 knots.
▭	16—31 m.p.h., 14—27 ,,
▭	32—47 m.p.h., 28—41 ,,
▬	48 m.p.h. and above, 42 knots and above.

Ranges of both 14—27 knots and 28—41 knots are indicated by the same convention, viz., an open rectangle. The two ranges are distinguished by a transverse line across the rectangle, the frequency of the lower range being represented by the length of that portion of the rectangle nearest the octagon.

The arrows on any one side of the octagon indicate the percentage frequencies at the heights specified above, the height increasing in a clockwise sense ; that is, the first arrow refers to the surface and the last arrow, reading in a clockwise sense to 10,000 feet above mean sea level.

In cases where no observations are recorded from a particular direction at a height, a dot is placed on the inside of the octagon to mark the position of the absent arrow.

The figures above the name of the station (printed in the centre of the octagon) indicate the number of observations at 1,500 feet and 10,000 feet, respectively, on which the rose is based. A dash (—) indicates that less than 15 observations were available.

The figures below the name of the station indicate the percentage frequencies of wind speeds less than 3 knots, irrespective of direction at each of the five levels.

The Relation between Wind and Pressure Distribution

The Motion of the Wind under Balanced Forces.—It has been indicated in Chapter VI that a relation exists between the wind at the earth's surface and the mean sea level distribution of pressure. This relation, however, is only approximate on account of the effect of friction between the air and the surface of land or water over which it moves. For air at a sufficient altitude to be beyond the influence of friction at the surface, that is above about 2,000 feet, the relation between the wind and the pressure distribution takes a similar but more precise form, which will now be investigated. Its modification for air at the surface will be discussed at a later stage.

It is assumed that the air moves horizontally under the influence of the pressure distribution at its own level, and that the latter changes only very slowly. The air then settles down to a state of steady motion, that is, the wind speed and direction at a particular place remain constant. The movement of the air then takes place under balanced forces—it is neither accelerated nor retarded. A simple example of motion under balanced forces is afforded by a cyclist riding at constant speed along a horizontal curved track. On rounding the bend, the machine automatically inclines itself towards the centre at an angle which adjusts itself to the curvature and the speed. Equilibrium is maintained under the action of two forces, the centrifugal force tending to drive the cyclist towards the outside of the bend, and the force of gravity tending to make him fall inwards towards the centre. The angle of inclination is

adjusted so that these forces balance. In a somewhat similar way, the speed and direction of the wind are adjusted so that it moves steadily under the action of three balanced forces.

These forces are :—

(i) The force due to the pressure gradient, tending to drive the air across the isobars in the direction of the lower pressure.

(ii) The geostrophic force, or that force which, owing to the rotation of the earth, continually tends to deflect a moving body at right angles to its line of motion.

(iii) The cyclostrophic force, or the centrifugal force which acts on the air owing to its motion along a small-circle path.

The Force due to the Pressure Gradient.—In Fig. 47, let *Ox*, *Oy*,

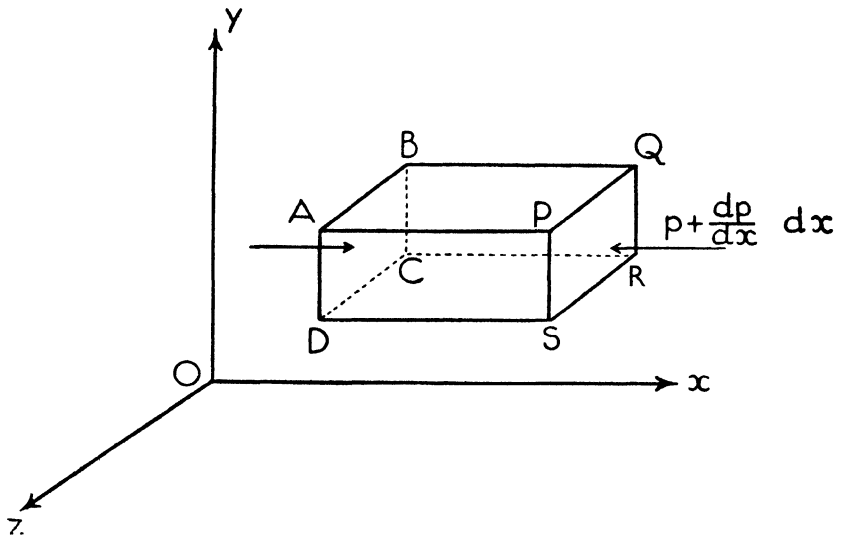


FIG. 47.—FORCE DUE TO PRESSURE GRADIENT.

Oz be a set of rectangular axes, *Ox* being horizontal and the direction of the local pressure gradient, *Oy* vertical, and *Oz* horizontal. Let *ABCDPQRS* be a small rectangular volume of air whose sides, of length *dx*, *dy* and *dz* are respectively parallel to the co-ordinate axes.

The pressure forces over the horizontal faces are balanced by the weight of the air, but owing to the variation of pressure in the direction *Ox*, there is a resultant force acting on the air in that direction, the resultant of the pressure forces acting on the faces *ABCD* and *PQRS*.

Let p be the pressure on the face $ABCD$.

Then the pressure on the face $PQRS$ is $p + \frac{dp}{dx} dx$.

The force on each face is equal to the pressure multiplied by its area $dydz$.

Force on $ABCD = pdydz$.

,, $PQRS = \left(p + \frac{dp}{dx} dx\right) dy\, dz$.

The resultant force, resolved in the direction Ox is therefore

$$p\, dy\, dz - \left(p + \frac{dp}{dx} dx\right) dy\, dz$$

$$= - \frac{dp}{dx} dx\, dy\, dz$$

$$= - V \frac{dp}{dx}$$

where V is the volume of the air. The pressure gradient is $\frac{dp}{dx}$; the negative sign shows that the resultant force acts in the direction of the lower pressure.

The magnitude of the pressure gradient is usually denoted by γ. The pressure force is therefore equal to γV, acting parallel to the pressure gradient in the direction of decreasing pressure.

The Geostrophic Force.—The air, in its motion, tends to follow a straight-line path in space. The earth's surface over which the air moves is, however, rotating and, to an observer on the rotating earth, the path of the air appears to be curved. The air appears to be acted upon by a force which continually tends to deflect it to one side. This force is called the geostrophic force, and may be investigated by considering the motion in space of a small area of the earth's surface.

It may be shown that at any place the earth's rotation may be resolved into two components.

(1) A turn-table movement of the horizontal plane about the vertical axis of the place. The sense of this movement is West—South—East, or anticlockwise, in the northern hemisphere, and West—North—East, or clockwise, in the southern hemisphere.

(2) A tilting movement about a North—South line, through the earth's centre, parallel to the tangent to the meridian at the place. The sense of this movement is West—Zenith—East in both hemispheres.

Since the wind blows horizontally, its direction is affected by the turn-table movement only, which may be regarded as the

tendency of the earth's surface to slide horizontally beneath the air. The tilting movement being entirely vertical, produces no such tendency.

In Fig. 48, let the circle with centre C represent the earth, CT

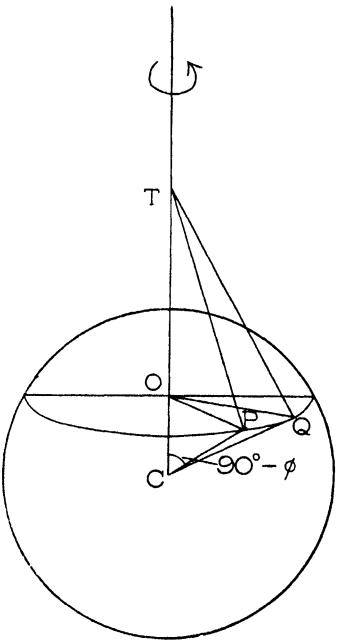


FIG. 48.—"TURN-TABLE MOVEMENT" OF THE EARTH'S SURFACE.

the earth's axis and the arrow, the direction of the earth's rotation. PQ are two points on the same parallel of latitude, centre O latitude ϕ, such that the earth's rotation carries P to Q in a very short interval of time t.

Let ω be the angular velocity of the earth's rotation.

Let Ω be the angular velocity of the turn-table movement.

Draw PT and QT tangential to the earth's surface to meet CT in T. These are the horizontal North—South lines at P and Q, respectively. Then PTQ is the angle through which PT turns in the time t, and therefore :—

$$\Omega t = \angle\, PTQ.$$

But also

$$\omega t = \angle\, POQ,$$

so that

$$\Omega = \frac{\angle\, PTQ}{\angle\, POQ}\omega$$

Now, if t is sufficiently small, so that PQ may be regarded as a straight line :—

$$\angle PTQ = \frac{PQ}{PT} \text{ and } \angle POQ = \frac{PQ}{OP}$$

$$\text{and } \frac{\angle PTQ}{\angle POQ} = \frac{OP}{PT}$$

$$= \sin\phi, \text{ since } \angle PTO = \phi.$$

Therefore $\Omega = \omega \sin\phi$.

The rate of the turn-table movement is therefore $\omega \sin\phi$.

Any small portion of the earth's surface may be regarded as a flat disc rotating with angular velocity $\omega \sin\phi$. In Fig. 49 let the circle with centre O represent such a disc.

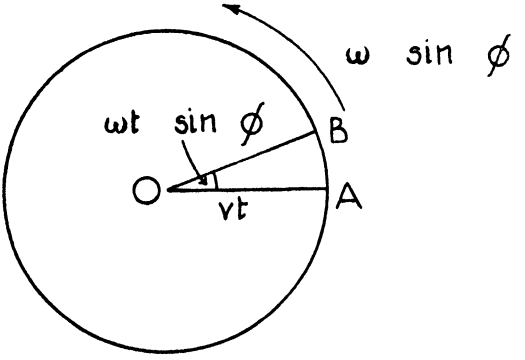


FIG. 49.—DEVIATION OF AIR MOVING ON A ROTATING DISC.

Suppose that air starting from O with uniform velocity v reaches a point A on the circumference of the disc in time t. Then $OA = vt$. But when the air was at O, B was at A, and therefore to an observer at O the air appeared to be moving in the direction OB. During the time t taken by the air to reach A the disc turns through an angle $\omega t \sin\phi$. To an observer on the moving earth, the air therefore appears deflected towards its right by an amount $\omega OA\ t \sin\phi = \omega\ vt^2 \sin\phi$.

Substitute this displacement in the well-known dynamical formula :—

$$S = \tfrac{1}{2}at^2,$$

where S is the displacement in time t and a the acceleration, the initial displacement being zero :—

Then

$$vt^2\omega \sin\phi = \tfrac{1}{2}\ at^2,$$

or

$$a = 2v\omega \sin\phi.$$

The effect of the earth's rotation is therefore equivalent to an acceleration $2\omega v \sin\phi$ at right angles to the path of the wind. In

the northern hemisphere the acceleration is directed towards the right. In the southern hemisphere, since the sense of the turn-table movement is clockwise, the acceleration is directed towards the left.

If m is the mass of the moving air, the geostrophic force is equal to

$$2m\omega v \sin\phi.$$

The Cyclostrophic Force.—If a body of mass m moves with uniform velocity v in a circle of radius r, it appears to be acted upon by a centrifugal force, in the plane of the circle, and acting radially outwards, of magnitude $\frac{mv^2}{r}$.

If the body is moving on the earth's surface the centrifugal force is present, except when the motion is along a great circle. In Fig. 50, let A be the pole of the small circle PRQ along which

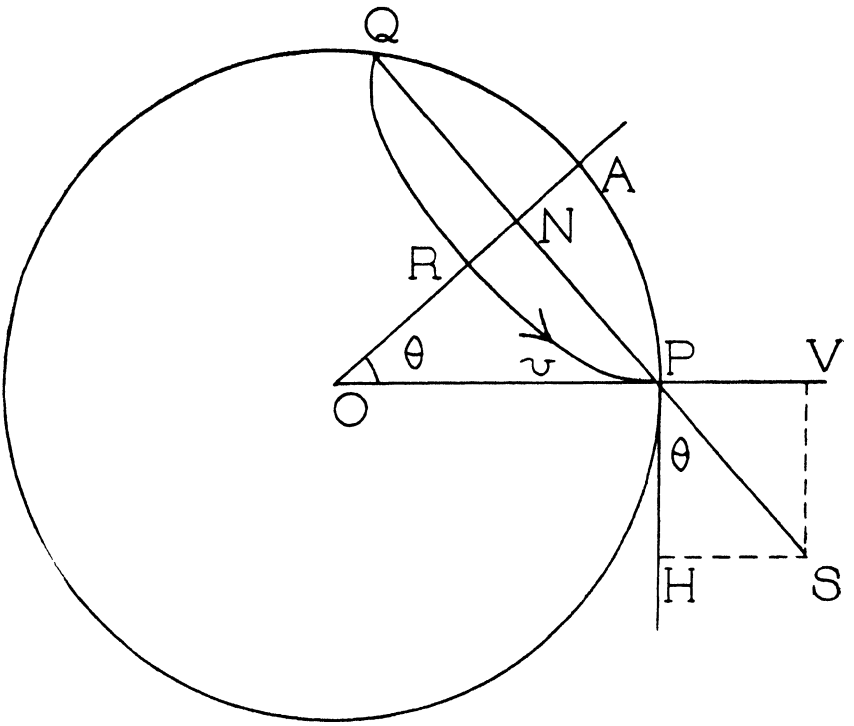


FIG. 50.—CYCLOSTROPHIC FORCE.

the air is moving with velocity v. Join PQ cutting the radius of the earth OA in N. Let θ be the angular radius of the small circle and E the radius of the earth. Then

$$NP = E \sin\theta,$$

and the magnitude of the centrifugal force is $\frac{mv^2}{E \sin\theta}$.

This acts along PS (NP produced), and therefore can be resolved into horizontal and vertical components along PH and PV, respectively.

The horizontal component only affects the horizontal motion. Since $\angle SPH = \theta$, its magnitude is

$$\frac{mv^2 \cos\theta}{E \sin\theta}$$

$$= \frac{mv^2 \cot\theta}{E}$$

and it acts at right angles to the direction of motion of the air, and outwards from the centre of curvature of its path.

Flow of the Wind along the Isobars.—Let the geostrophic, cyclostrophic and pressure-gradient forces on a mass of moving air be denoted by G, C and B, respectively. If the air is moving steadily, these forces mutually balance. Now G and C, being at right angles to the direction of motion, act in the same straight line, and therefore have a resultant equal to the algebraic sum $G + C$ in the same straight line. In Fig. 51 let OP represent the

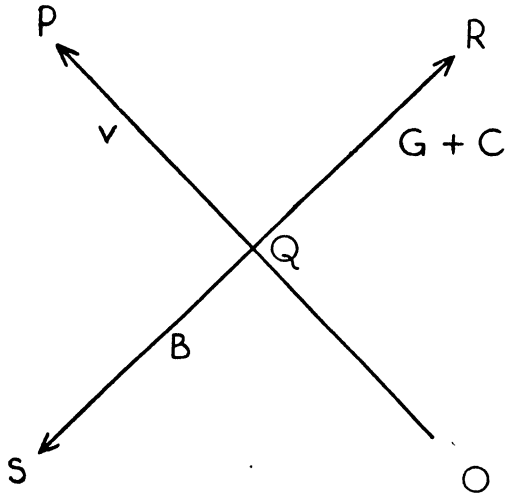


FIG. 51.—FLOW OF WIND ALONG THE ISOBARS.

velocity of the wind v, and let QR represent the resultant force $G + C$. This must be balanced by the remaining force B, which must therefore act in the same straight line as $G + C$, *i.e.*, perpendicular to OP. Since B is itself perpendicular to the isobars, OP must be parallel to them, that is, the wind flows along the isobars.

For straight isobars the force C is zero, and the wind flows under the action of the balanced forces B and G. In the northern

hemisphere, G is directed towards the right of the wind, and therefore B is directed towards the left of the wind, *i.e.*, the wind blows so as to have the lower pressure on its left. In the southern hemisphere the wind has the lower pressure on its right.

Cyclonic and Anticyclonic Winds.—By assuming that the three forces B, G and C mutually balance, an equation connecting the wind velocity with the pressure gradient may be derived. Two cases arise according as to whether the isobars are concave to the lower or higher pressure.

(*a*) *Cyclonic curvature of the isobars.*—The isobars are concave to the lower pressure about which the circulation of the wind is anti-clockwise in the northern and clockwise in the southern hemisphere. Accordingly, in Fig. 52, which has been drawn for

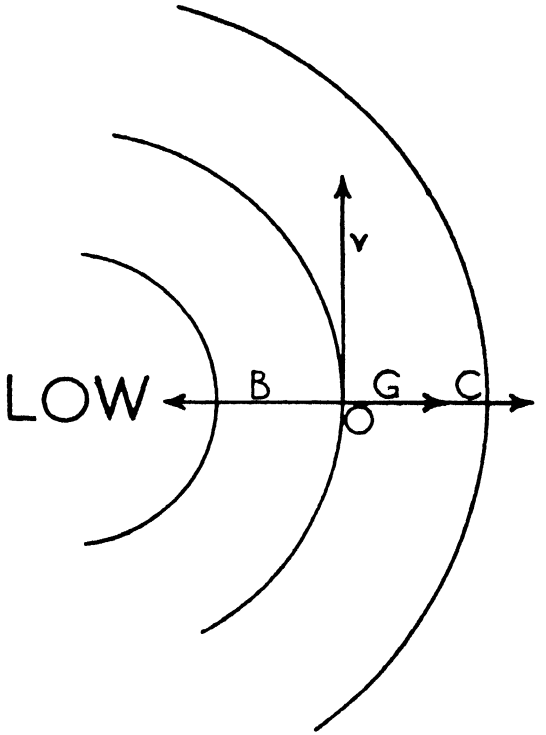


FIG. 52.—CYCLONIC CIRCULATION OF WIND (NORTHERN HEMISPHERE).

the northern hemisphere, let v represent the wind circulating anti-clockwise about the centre of low pressure enclosed by the curved isobars; v is tangential to the isobars. Of the forces B, C and G, B is directed inwards towards the low-pressure centre, C outwards, and since G is directed towards the right it also acts outwards. The arrangement of the forces is as shown in Fig. 52

The forces will balance only if,

$$B = G + C.$$

Substituting the values which have been derived for these forces, we obtain,

$$\gamma V = 2m\omega v \sin\phi + \frac{mv^2 \cot \theta}{E}$$

Divide throughout by V and write ρ for the density $\frac{m}{V}$. Then

$$\gamma = 2\omega\rho v \sin\phi + \frac{\rho v^2 \cot \theta}{E} \quad \ldots\ldots\ldots\ldots (1)$$

(*b*) *Anticyclonic curvature of the isobars.*—The corresponding equation for anticyclonic winds may be derived by assuming (for the northern hemisphere) a clockwise circulation about the centre of high pressure. It may, moreover, be shown that with anticyclonic isobars only a clockwise circulation is possible. For if the circulation were anticlockwise, B, C and G would all be directed outwards from the high-pressure centre, and therefore could not mutually balance.

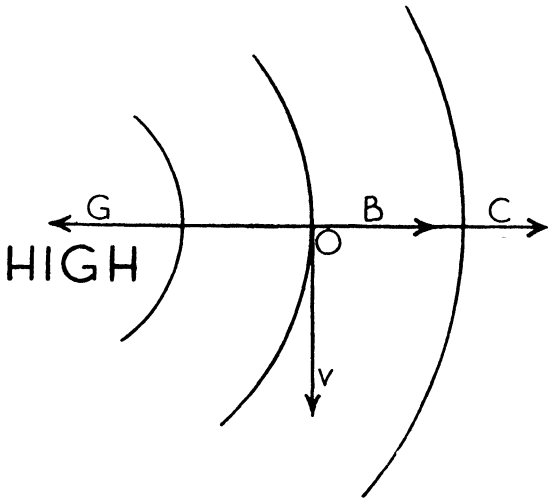


FIG. 53.—ANTICYCLONIC CIRCULATION OF WIND (NORTHERN HEMISPHERE).

The actual arrangement of the forces is as shown in Fig. 53, and the equation expressing the balance of the forces is

$$B = G - C.$$

Substituting the derived values, and as before, writing $\frac{m}{V} = \rho$, we obtain :

$$\gamma = 2\omega\rho v \sin\phi - \frac{\rho v^2 \cot \theta}{E} \quad \ldots\ldots\ldots\ldots (2).$$

At the equator, $\phi = 0$ and (2) becomes

$$\gamma = -\frac{\rho v^2 \cot \theta}{E}$$

$$\text{or} \quad v^2 = -\frac{E\gamma}{\rho \cot \theta} \quad \ldots\ldots\ldots\ldots\ldots\ldots (3)$$

Since E, γ, ρ and cot θ are positive, (3) requires that v^2 shall be a negative quantity, which is impossible. It follows that a local anticyclone cannot continue to exist at the equator. In effect this means that, were such a localized centre of high pressure to develop, the forces B and C, both acting outwards, would cause a direct flow of air across the isobars, with consequent collapse of the high pressure due to the withdrawal of air.

As the velocity increases, the predominating term in (2) becomes $-\frac{\rho v^2 \cot \theta}{E}$ which is negative. This makes the right-hand side of (2) negative, which is impossible, since γ must be positive. The equation therefore sets an upper limit to the value of v, and therefore to the value of γ, since $\gamma = 2\rho\omega v \sin \phi - \frac{\rho v^2 \cot \theta}{E}$. This is especially true in the central regions of an anticyclone, where θ is small and cot θ large. A small value of v then suffices to make γ negative. Therefore, in the central part of an anticyclone, the pressure gradients must be weak and the winds light, a conclusion which is borne out by experience.

The Gradient Wind.—The equations for cyclonic and anticyclonic winds express the wind speed in terms of the pressure distribution. The quantities γ, ρ, ω, θ, E and ϕ being known, the equations may be solved as quadratics in v. It is essential for the calculations of v by this method that the pressure distribution shall be known to a considerable degree of accuracy. The M.S.L. pressure distribution is the only one for which isobars can be drawn to a reasonable degree of precision, but the corresponding winds (at the surface) are affected by friction, and equations (1) and (2) are not applicable to them. On the other hand, at levels beyond the surface influence where equations (1) and (2) hold, the pressure distribution cannot be directly determined.

There is, however, justification for assuming that in the lowest 2,000 feet the pressure distribution changes little from that at sea level. The actual value of the pressure is reduced from M.S.L. to 2,000 feet, but the reduction is approximately the same everywhere, so that the isobars at 2,000 feet are coincident with those at M.S.L., only representing different pressures. In particular. the pressure gradients at 2,000 feet do not differ appreciably from those at M.S.L., and the M.S.L. isobars may therefore be used to determine the wind at 2,000 feet by means of the equations (1) and

(2), page 174. This has obvious advantages, since 2,000 feet is a convenient flying height for aircraft. The M.S.L. isobars may be used for determining the wind at 2,000 feet when cloud, darkness or other circumstances make direct observation impossible. Moreover, the wind at 2,000 feet may be determined for any part of the region for which the isobars have been mapped. This procedure has been justified in an investigation by Gold, who obtained direct confirmation that the isobars give a reasonably close approximation to the wind at 2,000 feet, by comparison of observed and computed winds. The wind so obtained is called the Gradient Wind, and the equations (1) and (2), page 174, the Gradient Wind Equations.

The Geostrophic Wind.—The calculation of the gradient wind is complicated owing to the presence of the cyclostrophic term $\pm \frac{\rho v^2 \cot \theta}{E}$. The curvature of the path of the air is difficult to determine in practice, and often it is not the same as the curvature of the isobars. In many cases, however, the cyclostrophic term is small in comparison with the other two, and may safely be neglected. For straight isobars it is zero. The principal cases in which it becomes important are :—

(*a*) In low latitudes where $\sin\phi$ is small, so that the geostrophic force is small.
(*b*) With very strong wind when v^2 becomes large.
(*c*) near the centre of a low-pressure system where, the isobars being sharply curved, θ becomes small and $\cot\theta$ large.

These cases apart, the modified form of both equations,

$$v = \frac{\gamma}{2\omega\rho \sin\phi}$$

gives a good approximation to the wind at 2,000 feet. It is called the Geostrophic Wind Equation, and the wind which is so determined is called the Geostrophic Wind.

The Geostrophic Wind Scale.—The speed of the wind at 2,000 feet may be read directly from an isobaric chart by means of the Geostrophic Wind Scale. Assume that ρ, ϕ and the scale of the chart are known. The geostrophic wind equation then determines the wind speed v in terms of the distance apart of consecutive isobars drawn (say) at intervals of 1 millibar.

The wind speeds may therefore be marked out along a line AB (Fig. 54) at distances from A equal to the corresponding distances apart of consecutive isobars. The scale thus constructed is a reciprocal scale, since the geostrophic wind equation shows that v is proportional to γ, and therefore inversely proportional to the distance between the isobars. If, then, the scale is placed at right angles to the isobars with one isobar coincident with A, the intersection of the next consecutive isobar with the scale gives the geostrophic wind.

For practical use, the geostrophic wind scale is engraved on a sheet of transparent celluloid, as shown in Fig. 55. The standard pressure, temperature (affecting density) and latitude are shown

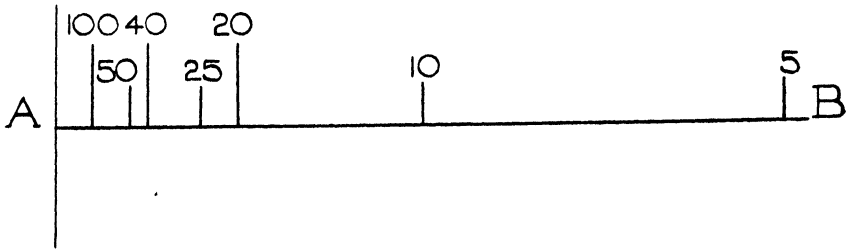


FIG. 54.—PRINCIPLE OF GEOSTROPHIC WIND SCALE.

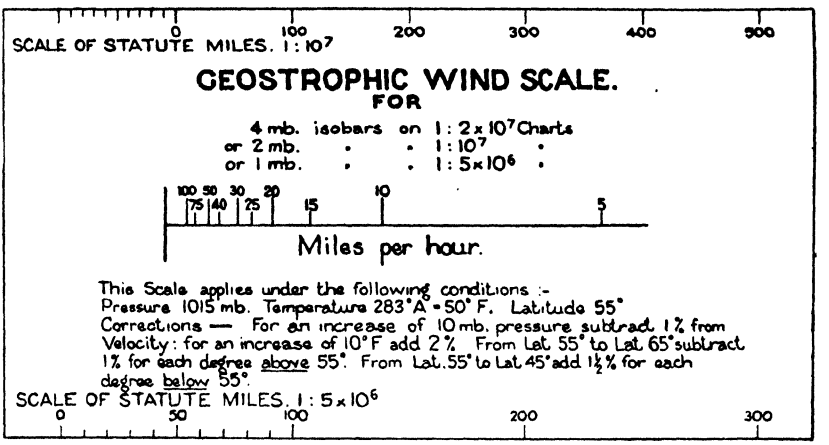
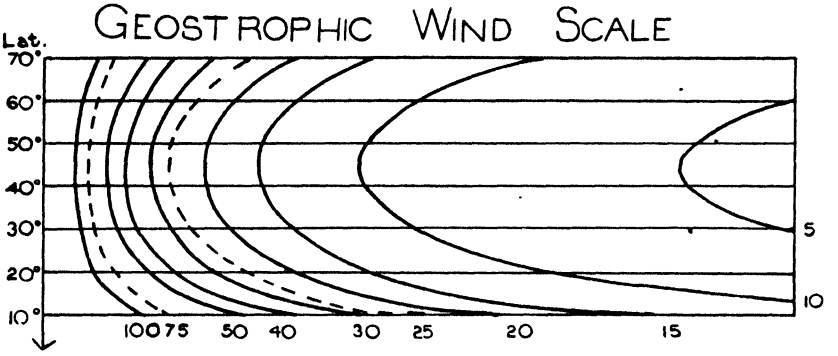


FIG. 55.—GEOSTROPHIC WIND SCALE.

on the scale, and for conditions other than these the specified corrections should be applied. The isobaric intervals corresponding to different scales of the chart are also specified. For example, with a chart on the scale $1 : 2 \times 10^7$, the geostrophic wind scale should be used with isobars drawn at intervals of 4 mb. in order that the interval between consecutive isobars shall give the correct wind. If an isobaric interval other than that appropriate to the scale of the chart is used, the wind as read from the scale must be multiplied by the appropriate factor. For example, if the scale is $1 : 5 \times 10^6$, the appropriate isobaric interval is 1 mb. If, however, the isobars are drawn at intervals of 2 mb., the wind speed obtained by placing the scale across the consecutive isobars must be multiplied by 2.

For use with charts, the scale of which varies considerably with

the latitude, and also to allow for the variation of sin ϕ, composite wind scales are constructed as shown in Fig. 56.



Gives geostrophic wind in Knots when used with isobars at 2 mb. intervals and air at pressure 1000 mb. temperature 290° A. or 63° F. Each 10 mb. excess pressure decreases wind by 1%; each 5° F. excess temperature increases wind by 1% and vice versa. Allowance is already made for variation of scale of chart with latitude.

NOTE.—This scale is for use with Mercator charts with natural scale 1 : 10,000,000 at latitude 15°.

FIG. 56.—COMPOSITE GEOSTROPHIC WIND SCALE.

A number of geostrophic wind scales appropriate to the different latitudes are placed parallel to one another in order of latitude, and the points on the scales corresponding to the same wind speeds joined by curved lines. To use the composite wind scale, dividers are placed across the isobars at the point where it is required to measure the geostrophic wind, and are then transferred to the scale at the appropriate latitude. For latitudes other than those corresponding to the horizontal lines on the scale, it is necessary to use intermediate positions found by interpolation.

The Relation between Geostrophic and Gradient Wind.—(*a*) *Anticyclonic Wind.*—For anticyclonic circulation the gradient-wind equation is

$$G = B + C \quad \ldots\ldots\ldots\ldots\ldots\ldots (1)$$

The geostrophic wind is given by

$$G = B \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots (2)$$

Now C is positive, since it involves v^2, and G is proportional to v. The value of G given by (2) is therefore less than that given by (1) from which it follows that the geostrophic wind is less than the gradient wind. The geostrophic wind is therefore an under-estimate of the wind in an anticyclone.

(*b*) *Cyclonic Wind.*—The gradient-wind equation is

$$G = B - C \quad \ldots\ldots\ldots\ldots\ldots\ldots (3)$$

Comparing (2) with (3), it is seen that the geostrophic wind is an over-estimate of the wind in a depression.

These conclusions are frequently borne out by synoptic charts, the same pressure gradient being associated with stronger winds for the anticyclone than for the depression.

The Relation of Surface Wind to Isobars.—The wind is parallel to the isobars only if the force due to the pressure gradient and the geostrophic and cyclostrophic forces mutually balance. At the surface an additional force, namely, that due to friction, must be taken into account. Its effect is to destroy the balance between the other forces, with the result that the wind at the surface is not parallel to the isobars. Friction may be regarded as a force acting parallel to the velocity of the air, but in the opposite direction, and its first effect therefore is to reduce the speed of the wind. The geostrophic and cyclostrophic forces, being proportional to the wind speed and its square, are correspondingly reduced, and no longer able to balance the pressure force, so that a residual force acts in the direction of the pressure gradient. The winds are therefore turned towards the lower pressure and flow across the isobars at an angle which depends upon the magnitude of the frictional forces. There is then a net flow of surface air across the isobars into a low-pressure centre and a net flow outwards from a high-pressure centre. The effect of friction is greatest over land of irregular relief, or where there are many obstructions to the flow of air, such as trees and buildings. It is less over gently undulating country, and smallest over the sea. The angle at which the wind crosses the isobars and the reduction of its speed increase with increasing friction, so that for a given pressure gradient the wind is stronger and more nearly parallel to the isobars over the sea than over land.

The Effect of Changing Pressure Distribution.—The gradient- and geostrophic-wind equations have been derived on the basis of the assumption that the pressure distribution remains unchanged or changes only slowly. The effect of changes in the pressure distribution has been investigated by Brunt and Douglas. Their results show that when the pressure distribution is changing the winds are no longer geostrophic, and the more rapid the change of pressure the greater is the difference between the actual and the geostrophic wind. If, on a synoptic chart, lines are drawn through places having the same barometric tendency, a series of curves closely resembling isobars is obtained. These are called **Isallobars**. Areas within which the barometer is rising are called isallobaric highs; those within which it is falling, isallobaric lows. The distance apart of consecutive isallobars is a measure of the isallobaric gradient. Brunt and Douglas found that the effect of changing pressure distribution can be allowed for by adding to the geostrophic wind a component blowing across the isallobars, at right angles,

into the low values of isallobars with a magnitude proportional to the isallobaric gradient. That is, there is convergence of winds from all sides into a region where the barometer is falling and a divergence of winds outwards from a region where it is rising. The isallobaric component of wind was found frequently to amount to 10 knots. The effects of the divergence and convergence on the weather will be discussed in a later chapter.

The Variation of Wind in the Lowest 2,000 Feet.

Turbulence.—It is a fact of everyday observation that the wind at a particular place on the earth's surface is never steady, but blows in a series of gusts and lulls with accompanying fluctuations of direction, the period of the fluctuation being usually a few seconds. The variation of wind from point to point is also very rapid, and differences of 10 knots in wind speed, and 30° to 60° in direction may occur momentarily between positions only a few yards apart. Moreover, these fluctuations are extremely irregular, and appear to bear no definite relation to one another. The movement of the air, being thus exceedingly complicated, is incapable either of exact specification or mathematical analysis. The atmosphere is, in fact, usually in a state of turmoil comparable with that observed in a swiftly-flowing river or tidal stream. In flowing water the nature of the motion is rendered visible by the vortices, whirlpools or eddies which appear as dimples on its surface. These eddies constitute the characteristic feature of the motion. They are constantly being generated and dying away, their places being taken by others. The less vigorous may become caught up in the circulations of the stronger, and absorbed by them, while other eddies may break up into a number of smaller ones. Although the motion is visible only on account of the irregularities it produces on the surface, it is evident that it extends throughout the whole volume of the water, the axes of the eddies being prolonged as irregular curves, either to the bed of the stream or back to the surface. Observation shows that all motion of liquids and gases, except when the velocity is very small, is of this eddying nature. It is known as Turbulent Motion, or simply as Turbulence.

The distinction between steady (stream-line) flow and turbulent motion was first demonstrated by Osborne Reynolds in a series of experiments on the flow of water in glass tubes of various diameters. The tubes had bell-mouthed entrances, into which a little aniline dye was introduced and the motion of the coloured water observed. At low velocities the motion of the coloured water was drawn out into a fine straight band, but as the velocity of flow was increased, a limit was attained at which the motion became turbulent. The band of coloured fluid oscillated wildly from one side of the tube to the other and soon lost its identity amid the generally confused motion.

Owing to its irregular nature the motion in a turbulent fluid

cannot be studied in detail as by tracing the history of any individual particle. It is, however, possible to form a reasonable picture of the effect of a large number of eddies. The process is similar to that of molecular diffusion. If a small quantity of colouring matter is introduced at one point into a vessel of clear " still " water it gradually diffuses throughout the whole volume of the water, which becomes uniformly coloured in consequence. Similarly, a small quantity of smoke introduced into a chamber of " still " air becomes uniformly diffused throughout the whole volume. Other physical properties of the air such as its water-vapour content, linear momentum in a given direction, heat and dust content are subject to the same diffusive process. The general effect of eddy motion also is to diffuse the properties of the air throughout its volume and the process is known as eddy diffusion. Diffusion by eddies, however, takes place at a far greater rate than that achieved by molecular diffusion, and this difference of rate constitutes the essential difference between the two processes.

Almost all fluid motion is turbulent, but not equally so. The turbulence in a slowly-flowing river, for example, is widely different from that in a mountain torrent. The difference lies, not in the fundamentally turbulent nature of the motion, but in what may be called its degree of turbulence. In the atmosphere, too, there are varying degrees of turbulence. There is obviously less turbulence, or a lower degree of turbulence, in the " still " air of a closed room, where movement is shown by cigarette smoke, or by the dust particles in a beam of sunlight, than there is in the atmospheric movement rendered visible by the smoke issuing from a ship's funnels in a strong wind. Other manifestations of atmospheric turbulence, intermediate in degree between these cases, are the motion of fog particles in calm air, the shimmering of the air over heated land on summer afternoons, the whirling and drifting of falling snowflakes, and catspaws on the surface of the sea.

In all these types of motion it is evident that there are present eddies of varying sizes, and that the largest eddies are associated only with the highest degrees of turbulence. The largest eddies shown by cigarette smoke appear to be only a few inches in diameter, whereas in the funnel smoke, the maximum eddy diameter appears to be several feet, although the smaller eddies may also be present. Clearly, too, with the larger eddies the diffusive process is much more rapid than with the smaller. It may be described as a thorough churning or mixing of the air involved.

Degree of turbulence is really a mathematical conception, but to avoid the difficulties involved in considering it from that aspect, it will be sufficient to confine attention to the qualitative considerations above. Generally, it may be said, that in the atmosphere the largest scale turbulence is associated with the largest eddies, and the largest and most rapid fluctuations of the wind, and *vice versa*.

Factors Affecting Turbulence.—The principal factors affecting turbulence in the atmosphere are :—

(*a*) The speed of the wind.

(*b*) The nature of the surface over which the wind blows and the obstructions that it offers to the wind.

(*c*) The lapse-rate.

With regard to the first of these factors, observation and the considerations of the preceding paragraph show that the degree of turbulence increases with the speed of the wind.

The effect of the surface and its obstructions is illustrated by the records of wind reproduced in Figs. 57 and 58. The former is

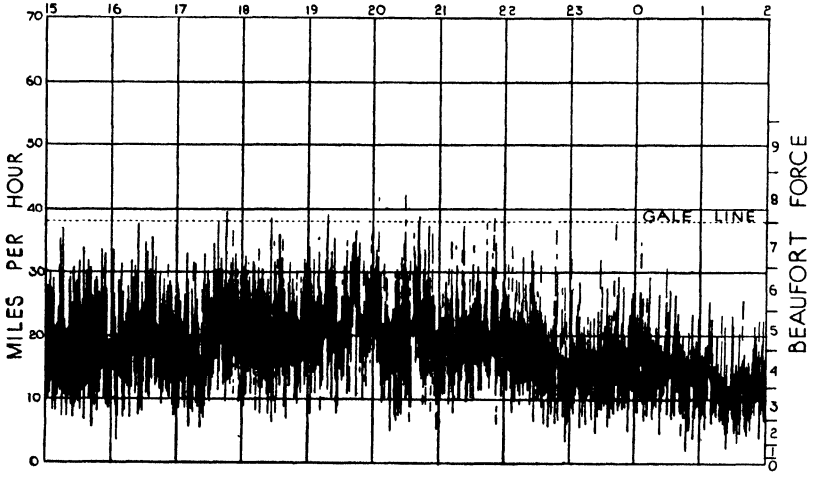


FIG. 57.—ANEMOGRAM—AIR MINISTRY, FEBRUARY 28TH, 1937.

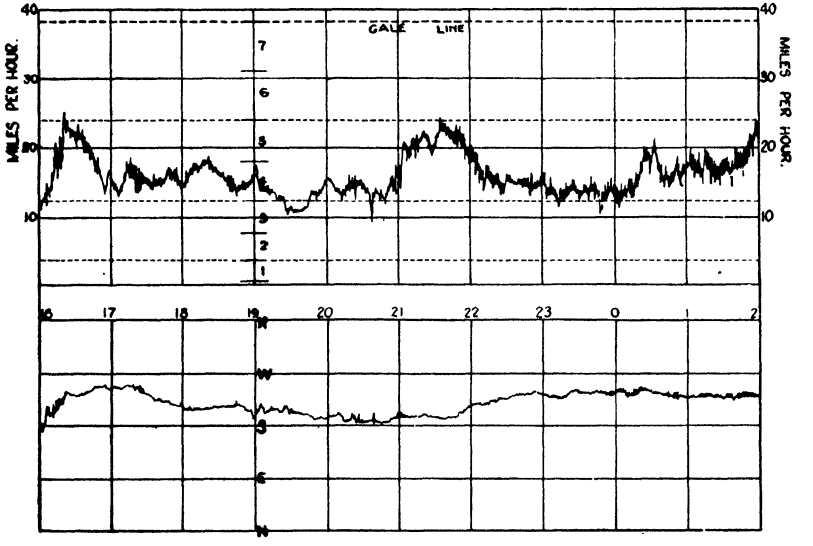


FIG. 58.—ANEMOGRAM—BELL ROCK LIGHTHOUSE.

a record of wind from the anemometer on the roof of the Air Ministry, London. The surrounding buildings provide a very irregular surface, with numerous obstructions, and the wind is turbulent to an extreme degree. The record of wind speed consists of a broad ribbon with wide excursions on either side of the mean, representing the gusts and lulls, (while the direction record would also show wide fluctuations).

The second record reproduced is from the Bell Rock Lighthouse, which has an open exposure to the sea in all directions. The comparatively steady nature of the wind is at once apparent. There are still gusts and lulls, but they are much less prominent than in the London record, both traces taking the form of a narrow ribbon of uniform width. The Bell Rock record is typical of those obtained at sea. At open situations in level country the gustiness lies between the two extremes represented by the Air Ministry and Bell Rock records. An uneven surface with many obstructions is associated with a high degree of turbulence, and a level, even surface with a low degree of turbulence. The turbulence is evidently due, in part at least, to the friction between the air and the surface over which it blows.

The importance of the lapse-rate lies in the fact that the eddy motion set up in a current of air blowing past obstructions has vertical as well as horizontal components. If the lapse-rate is large, vertical motion, once started, tends to persist, owing to instability. If, however, the air is stable, as with a small lapse-rate or an inversion, the vertical motion is damped, and the more stable the lapse-rate the greater is the damping effect. Now vertical motion brings about interchanges of air between different levels, the displaced air tending to retain its own horizontal momentum. If the speeds at different levels vary considerably among themselves, vertical interchange brings air masses with widely different speeds together, with consequent further increase of turbulence. A high degree of turbulence is therefore associated with a large lapse-rate, and *vice versa*. This conclusion is supported by observation, for, when the lapse-rate is large the wind is often gusty, but a small lapse-rate or an inversion is often associated with quiet conditions as in the still air of a fog.

The Variation of Wind in the lowest 2,000 Feet.—The variation of wind in the lowest layers is governed by the turbulent interchange of momentum between the surface air and that at a sufficiently high level to be beyond the influence of friction at the surface. The latter is called the free atmosphere, and its base may be taken to be at a height of approximately 2,000 feet above the surface. For convenience, the wind at 2,000 feet will be referred to as the upper wind (in the lowest 2,000 feet).

In the absence of turbulence the surface air and the upper air would pursue their natural tendencies unhindered, the surface air to be retarded and turned towards the lower pressure (page 179).

or even to come to rest under the influence of surface friction, the upper air to acquire the speed and direction of the gradient wind appropriate to the M.S.L. isobars. The momentum of the surface air is therefore less than, and in a different direction from, that of the upper air. Owing to turbulence, the momentum tends to become uniformly distributed through the layers of air involved and the surface air, in particular, gains momentum at the expense of the upper. The momentum thus acquired is being constantly consumed in overcoming friction at the surface, so that the uniform distribution with height is never attained, but there is a constant transfer from the upper to the surface layers. It is chiefly by this means that the surface wind is maintained.

Owing to the tendency towards equalization of momentum the surface wind and the upper wind constantly tend to equality of speed and direction. The effect is greatest in the case of the surface wind. Equality is never actually attained owing to friction at the surface, but the greater the degree of turbulence, that is the more rapid the transport of momentum, the more nearly is equality approached. The relation between the surface wind and the upper wind for a high and low degree of turbulence are illustrated in Fig. 59a, which applies to the northern hemisphere. The parallel straight lines represent the isobars, the heavy arrow the speed and direction of the upper wind, the full light arrow and the broken arrow the surface wind with a high and low degree of turbulence, respectively. In both cases, the speed of the surface wind is less than that of the upper wind, and its direction is backed from that of the upper wind, the reduction in speed and the amount of the backing being greater for the lower degree of turbulence. Since it is the vertical transport of momentum which is important, the effective turbulence is governed principally by the lapse-rate of temperature, and not by the other two factors. For this reason, on cloudy days, which are associated with little surface heating, and therefore a small lapse-rate, the surface wind is comparatively light in relation to the upper wind, and its direction is backed considerably from the run of the isobars. The variation of wind from the surface to 2,000 feet is correspondingly rapid. On clear days, with pronounced surface heating and therefore a large lapse-rate, the speed of the surface wind is more nearly equal to the gradient-wind speed, and is backed only a little from the run of the isobars. There is little variation of wind between the surface and 2,000 feet. Always, however, from the surface up to the 2,000 feet level, the wind veers and increases in speed to a greater or less degree, depending upon the turbulence.

Fig. 59b shows the corresponding variations for the southern hemisphere. The relation between the speeds of the upper and surface winds is unaltered ; that between their direction is reversed, the wind backing from the surface up to the 2,000 feet level.

Attempts have been made from time to time to discover some

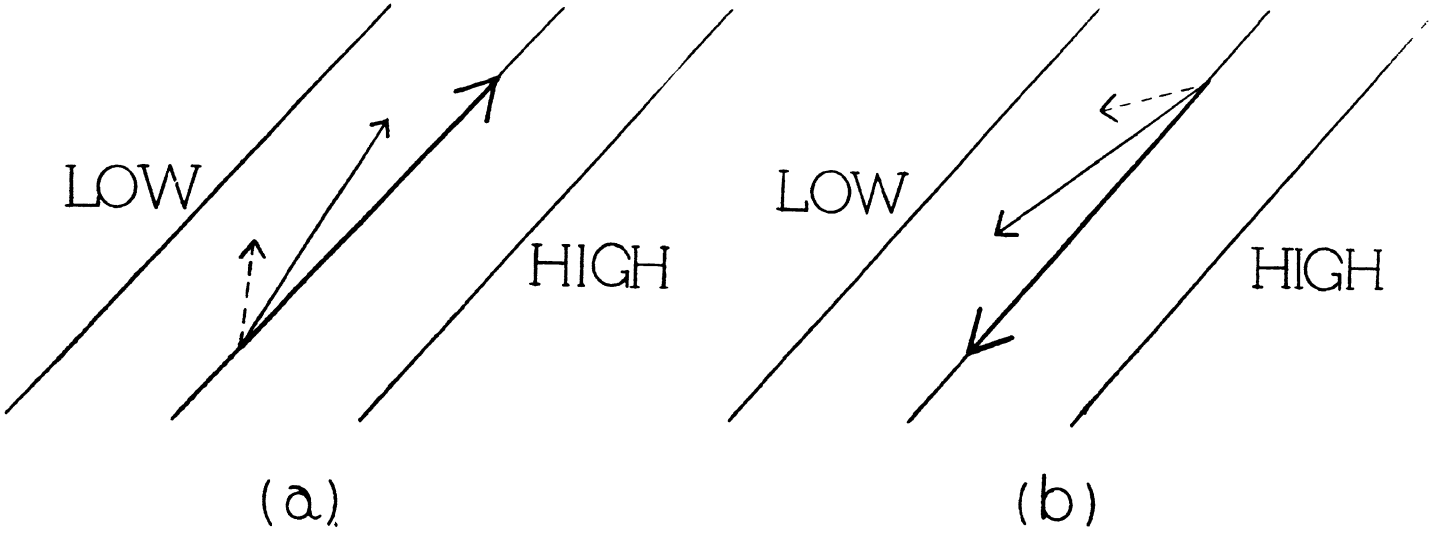


FIG. 59.—VARIATION OF WIND IN THE LOWEST 2,000 FEET—(*a*) NORTHERN (*b*) SOUTHERN HEMISPHERES.

simple formula that will express the variation of wind with height in the lowest layers. One such empirical formula has been put forward by E. H. Chapman :—

$$V = a \log H + b,$$

where V is the speed of the wind at height H, a and b are constants which vary from place to place and from time to time at the same place.

If the speed of the wind at the surface ($H = 0$) and at 2,000 feet are known, it is possible to determine a and b by substitution in Chapman's formula. The wind speed at any other height can then be found.

The Diurnal Variation of Wind.—At many land stations there is a well-marked tendency for the wind to increase in speed and veer (in the northern hemisphere) between sunrise and mid-afternoon, and to decrease and back towards evening. This is due to the increase of turbulence associated with the normal increase of lapse-rate which results from the heating of the land surface. In Fig. 59a the dotted arrow is representative of the wind by night, and in the early morning when the turbulence is small, the light arrow the afternoon wind, when the turbulence has increased. The diurnal variation of wind is most marked in clear summer weather, when the surface heating is considerable. Increases of wind from calm or force 1 at sunrise to force 4 in the early afternoon are then not uncommon.

On cloudy days the diurnal variation disappears. It may also be masked by changes in the general pressure distribution, as, for example, the approach of an intense low-pressure system with its accompanying strong winds. With an inversion of temperature at the earth's surface, the downward transport of momentum becomes negligible, and the surface wind may fall calm or very light, and bear no relation to the upper wind. This accounts for the light winds and calms which are often reported from land stations in the early morning after a clear night, even though the pressure gradient may be appreciable. Particularly in valleys, into which the cold air gravitates, do the surface winds often bear no relation to the pressure gradient, the upper wind gliding over the pools of cold air without disturbing them. On cloudy nights there is no tendency for inversions to form and the surface wind is often maintained.

The upper wind is subject to a diurnal variation which is the reverse of that at the surface. Momentum gained by the surface air must be lost by the upper air ; hence the upper wind tends to change its speed and direction towards those of the surface wind when turbulence is greatest. That is, the upper wind decreases and backs (in the northern hemisphere) by day, increases and veers by night. From an analysis of the results of pilot-balloon ascents, J. Durward has reached the conclusion that the diurnal backing of the upper wind may amount to 20°. The diurnal variations of

speed are illustrated in Fig. 60, which shows the average winds at different times of the day at a station near the ground and at the top (1,000 feet) of the Eiffel Tower.

In the southern hemisphere the diurnal variation of wind is similar to that in the northern hemisphere as regards speed, but as regards direction the term veering must be substituted for backing, and *vice versa*.

Effect of the Surface. Winds in the lowest layers over the Sea.—Over the sea, turbulence and surface friction are operative, but the magnitude of their effects differs from that over the land :—

(*a*) Despite the lower degree of turbulence over the sea as compared with the land, surface friction being also very much less, the surface wind approaches the gradient wind more nearly over the sea than over the land.

Approximate rules are :—

Over the sea, surface wind speed = $\frac{2}{3}$ geostrophic wind.

Over the land, surface wind speed = $\frac{1}{3}$ geostrophic wind.

These rules are often borne out by the winds on the east and west coasts of the British Isles. With uniform pressure gradient for west or south-west winds across the country, the west coast stations may report winds of force 8, the east coast stations only force 5. The exposures of the former stations are equivalent to open sea exposures, those of the latter to land exposures. The effect of land and sea exposures is shown in Fig. 61.

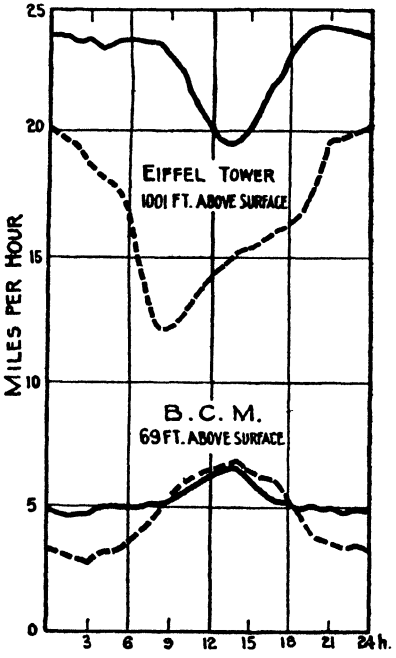


FIG. 60.—DIURNAL VARIATION OF WIND SPEED

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which indicates the relation between the surface wind and the geostrophic wind at Great Yarmouth for different wind directions.

The outer circle represents 100 per cent. of the geostrophic wind, and the inner circle 50 per cent. The full line represents the percentage ratio of actual to geostrophic wind for winds of force 6 on the Beaufort scale; the dotted line the ratio of winds for force 4. For on-shore winds, in the eastern semi-circle, the winds are between 50 and 100 per cent. geostrophic; for off-shore winds, in the western semicircle, the winds are mainly about 50 per cent. geostrophic for force 6 and somewhat less for force 4.

(*b*) The winds tend to follow the isobars much more closely over the sea than over the land. The angle at which the wind cuts the isobars seldom exceeds 20° over the sea.

(*c*) There is little variation of the air temperature at sea between day and night, and therefore no diurnal variation of wind. In particular there is no lull of the wind at night.

(*d*) Variations of turbulence are caused by changes in the relation between the temperature of the sea surface and that of the air in immediate contact with it. Air flowing over and towards warmer water has its surface temperature, and therefore its lapse-rate increased. The increased turbulence tends to bring about closer agreement between the surface and the upper winds. The lapse-rate in air flowing over and towards colder water decreases with consequent decrease of turbulence. For these reasons, winds flowing over the sea to lower latitudes are usually in closer agreement with the pressure gradient than winds flowing polewards. If, however, the air passes over land at any time the temperature in the lowest layers will be affected in a complicated manner, and the rules given above do not apply to air which has recently crossed a land mass. For example, in the neighbourhood of the British Isles the rules may be applied to winds from between SSW. through W. to NE., but not to others. It is found that NW. winds agree more closely with the gradient than do SW. winds.

The Variation of Wind with Height in the Free Atmosphere.

Introductory.—In the free atmosphere, beyond the influence of turbulence arising from surface friction, it may be assumed that the wind at any level is equal to the gradient wind appropriate to the pressure distribution at that level. This assumption has

already been made in relating the wind at 2,000 feet to the M.S.L. pressure distribution, 2,000 feet being taken as the base of the free atmosphere. The wind at 2,000 feet may be conveniently regarded as the basis of reference for the variation of wind with height in the upper levels, and for brevity it will be referred to as

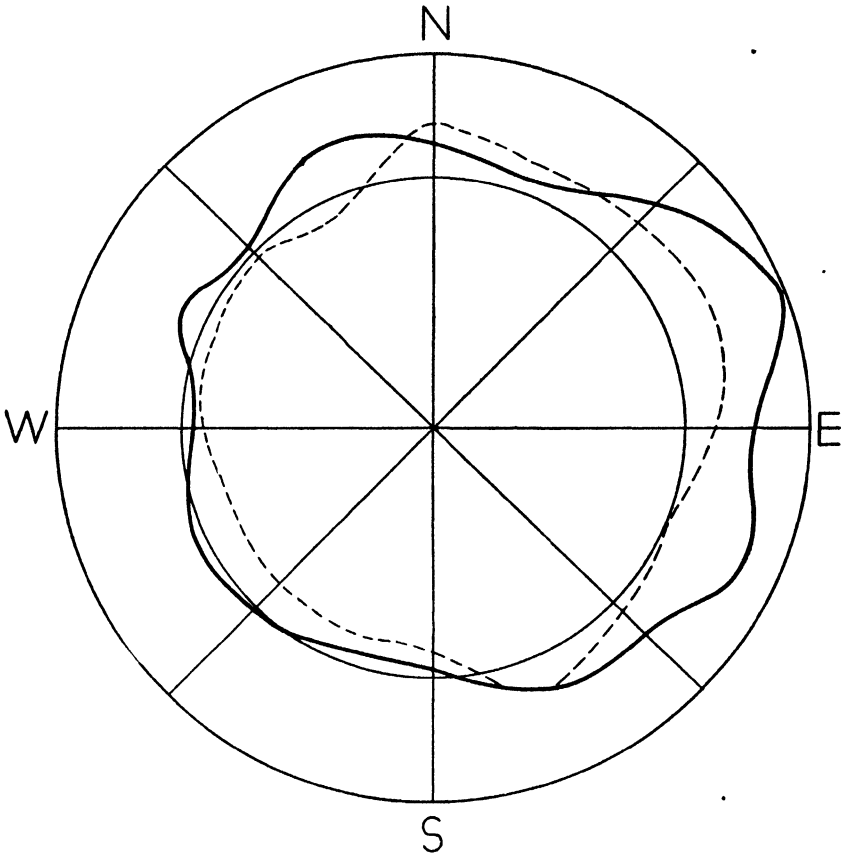


FIG. 61.—RELATION BETWEEN GEOSTROPHIC AND SURFACE WIND AT YARMOUTH.

the "lower wind." Above 2,000 feet the pressure distribution generally changes from one level to another, leading to a variation of wind with height. The changing pressure distribution affords a convenient means of investigating the nature of the change of wind with height, and depends, in its turn, upon the distribution of temperature in the atmosphere below that height, and also upon the M.S.L. pressure distribution. The variation of wind with height, in some special cases, will now be discussed.

Uniform M.S.L. Pressure.—In Fig. 62 let AB, CD represent two air columns of equal height h. Let B and D be at M.S.L., where the pressure is uniform and equal to p_o. Let T and T' denote the mean temperatures in the two air columns, p and p' the pressures at A and C, respectively, and p_d and p'_d the differences of pressure between the tops and bases of the air columns.

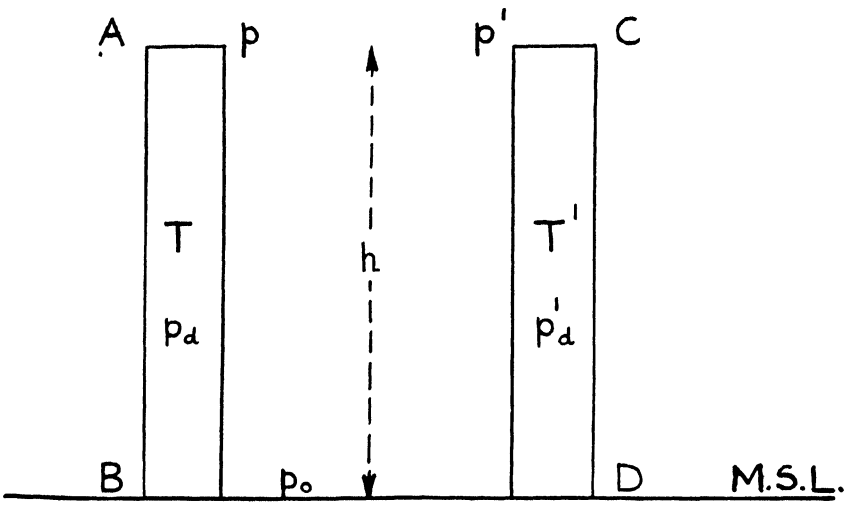


FIG. 62.

Let T be greater than T'.

Now p_d is the pressure due to the weight of the air column AB, and p'_d is that due to the weight of CD. Since T is greater than T', the air in AB is less dense than that in CD, consequently p_d is less than p'_d.

Then, for AB, $p_o = p + p_d$.
and for CD, $p_o = p' + p'_d$.
Therefore $p' + p'_d = p + p_d$.
and p is greater than p'.

The pressure at the top of the air column having the higher mean temperature is therefore the greater.

This result may be stated as follows:—

If the M.S.L. pressure is uniform, then at any upper level the higher pressure is associated with the higher mean temperature between the surface and the level in question. The lower pressure similarly is associated with the lower mean temperature.

It should be noted that the effective temperature is neither that at M.S.L. nor at the selected upper level, but the mean temperature of the atmosphere between these two levels. The effect of difference of mean temperature in the horizontal is to bring about pressure differences and therefore winds in the upper levels above regions where the M.S.L. pressure is uniform with no lower wind.

The difference of pressure at the tops of two air columns gradually accumulates as the height of the columns increases and the variation of wind with height is therefore gradual.

The pressure-height formula (Chapter VII) may be used to indicate the magnitude of the pressure difference in a typical case.

Let $p_o = 1{,}000$ mb., $h = 10{,}000$ feet, $T = 280°$ A., and $T' = 270°$ A.

Then, for AB, $h = 221T\ (\log_{10} p_o - \log_{10} p)$,

$$i.e.,\ \log_{10} p = \log_{10} p_o - \frac{h}{221T} = 3 - \frac{10{,}000}{221 \times 280}$$

Similarly for BC,

$$\log_{10} p' = 3 - \frac{10{,}000}{221 \times 270}.$$

Evaluating by logarithms, we obtain $p = 689$ mb., $p' = 680$ mb., giving a pressure difference of 9 mb. This difference, if assumed to be spread uniformly over a distance of 400 miles, would give a wind of 25 knots at 10,000 feet although there is no wind at M.S.L.

Variation of Temperature Across the Isobars.—The nature of the change of wind with height may be similarly investigated when the M.S.L. pressure is not uniform. It is convenient to consider firstly the case in which the mean temperature up to any specified height is constant for all points on the same M.S.L. isobar. There is then said to be a variation of temperature across the isobars. In Fig. 63

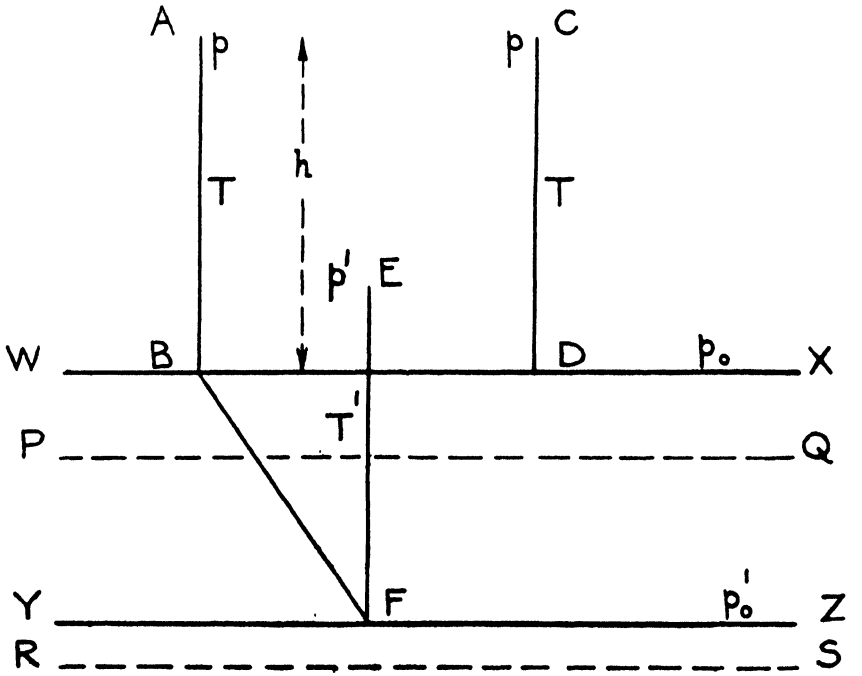


FIG. 63.

let the straight lines WX, YZ represent two consecutive M.S.L. isobars. Then, for the assumed variation, the distribution of mean temperature may be represented by isotherms PQ, RS parallel to the isobars. Let AB, CD represent two air columns of equal height h at points on the isobar WX along which the M.S.L. pressure is p_o. Let the mean temperature up to height h in both air columns be T.

The pressure at A is equal to that at C, since the two air columns are of equal weight. Let this pressure be denoted by p. If a system of isobars is drawn in the horizontal plane at height h, the isobar corresponding to the pressure p passes through A and C and is therefore parallel to the M.S.L. isobar through B and D. The same result holds, under the given conditions for the isobars at any other height. Since the wind at any level is parallel to the isobars at that level, it follows that the direction of the wind remains constant at all heights.

To investigate the effect of the assumed temperature distribution on the wind speed, consider a third column of air EF on the isobar YZ. Let the pressures at E and F be denoted by p' and p'_o, respectively, and let T' be the mean temperature of the air column. Let BF be perpendicular to WX and YZ, so that the difference of pressure between B and F is a measure of the M.S.L. pressure gradient and that between A and E is a measure of the pressure gradient at the upper level. Assume that the M.S.L. pressure gradient is directed from F to B, *i.e.*, $p'_o > p_o$, so that the lower wind (in the northern hemisphere) blows in the direction WX. Two cases arise according as to whether the mean temperature increases or decreases in the direction BF. In the former case the higher mean temperature is on the side of the high pressure ; in the latter of the low pressure.

(*a*) *Higher mean temperature associated with the higher pressure.*— In this case $T' > T$ and the weight of the air column EF is less than that of AB. The reduction of pressure from F to E is less than that from B to A.

Therefore $p'_o - p' < p_o - p$

Adding $p' - p_o$ to both sides of this inequality,

$$p'_o - p_o < p' - p.$$

The difference of pressure between E and A is therefore greater than that between F and B, so that the pressure gradient increases with height. Since the density of the air normally decreases with height, the speed of the wind increases with height.

(*b*) *Higher mean temperature associated with the lower pressure.*— In this case, $T' < T$ and the weight of the air column EF is greater than that of AB. The reduction of pressure from F to E is greater than that from B to A.

Then, $p'_o - p' > p_o - p$, from which there follows as before,

$$p'_o - p_o > p' - p.$$

The difference of pressure between E and A is therefore less than that between F and B. The pressure gradient decreases with height and at a certain level, depending upon the temperature and M.S.L. pressure distribution, disappears. At this level the conditions are similar to those discussed on page 190, and at higher levels a reversed gradient gradually builds up with increasing height, with the higher pressure now associated with the higher temperature. The wind undergoes corresponding variations, decreasing up to the level of zero gradient, where it becomes zero, and then slowly increasing from the reverse direction.

The foregoing results may be summed up as follows :—

(i) Variations of temperature across the isobars cause the speed of the wind but not the direction to change with height.

(ii) The speed of the wind increases with height if the higher mean temperature occurs in association with the higher pressure.

(iii) The speed of the wind decreases with height, sooner or later becoming zero and then increases from the reverse direction if the lower temperature occurs in association with the higher pressure.

Variation of Temperature Along the Isobars.—In Fig. 64 let

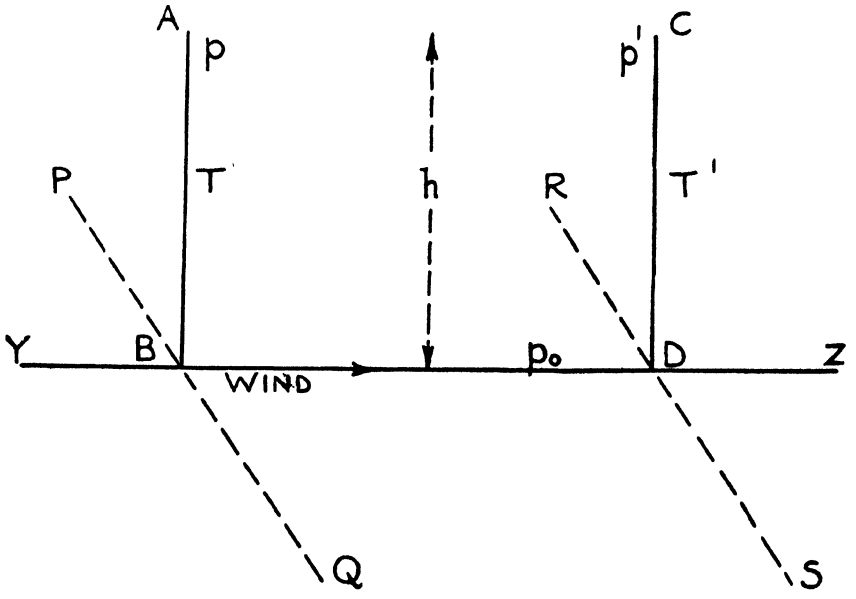


FIG. 64.

YZ represent the isobar at M.S.L. for the pressure p_0, and let the pressure distribution be such that, in the northern hemisphere the lower wind blows in the direction YZ. For a variation of temperature along the isobars the isotherms PQ, RS run across YZ. Let AB and CD represent two columns of air of equal height h, B and D lying on YZ, and denote the temperatures in the air columns by T and T', respectively. Let p and p' denote the pressures at A and C.

There are two cases, according as to whether the mean temperature increases or decreases in the direction YZ.

(*a*) *Lower wind blowing from the area of lower temperature.*—The mean temperature increases in the direction YZ so that $T' > T$. It follows that, since the pressures at B and D are equal, $p < p'$.

The isobars at the level of A and C are therefore inclined to the isobars at M.S.L. in such a way that the pressure increases from A to C, and the upper wind has a component across AC in the sense that for an observer facing the wind C would be situated to his left. In Fig. 65, which is drawn in a horizontal plane for the

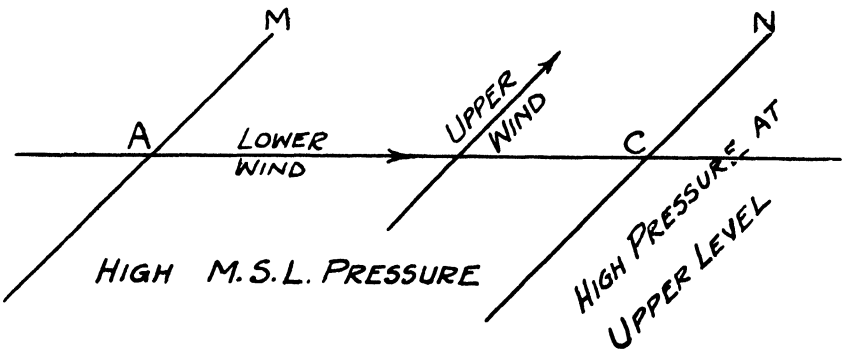


FIG. 65.

upper level, AM and CN represent the isobars at height h, while the M.S.L. isobars are parallel to AC. The upper wind, in the direction AM (in the northern hemisphere) is therefore backed from the lower wind. The angle of backing depends upon the rate of increase of temperature along the isobars and also upon the height h.

(*b*) *Lower wind blowing from the area of higher temperature.*—It may be shown in a similar manner that (in the northern hemisphere) the wind veers with height when the lower wind blows in the direction of decreasing temperature.

These results may be summed up as follows :—

(i) Variations of temperature along the isobars cause the direction of the wind to vary with height.

(ii) In the northern hemisphere the wind backs with height if it blows towards higher temperature and veers with height if it blows towards lower temperature. Hence the rule " Warm to windward upper wind veers. Cold to windward upper wind backs."

(iii) For the southern hemisphere " veer " should be substituted for " back " in (ii), and *vice versa.*

The usual variation of temperature is neither exactly along nor across the isobars, and consequently the normal variation of wind with height depends upon both types of temperature distribution. That is, in general there are changes in both the wind speed and direction with increasing height.

Some Further Effects of Temperature Distribution.—Mathematical analysis leads to the following additional results, which are here stated without proof :—

(*a*) The larger the horizontal temperature gradient the more effective it is in bringing about changes of wind with height. It is relatively more effective when the M.S.L. pressure gradient is weak than when it is strong. For example, low temperature occurring in association with high M.S.L. pressure will produce a reversal of wind at a comparatively low level if the M.S.L. pressure gradient is weak.

(*b*) If there is no horizontal temperature gradient, the wind at any height is proportional to the absolute temperature at that height, and its direction does not vary with height. Owing to the normal decrease of temperature with height, the wind would, under the condition stated, also decrease with height. The fact that it normally increases shows that horizontal temperature gradients usually exist.

(*c*) In a region of the atmosphere in which the temperature is uniform, both vertically and horizontally (an isothermal layer), the wind is constant at all heights.

(*d*) If the distribution of temperature persists up to great heights, the winds ultimately tend to become parallel to the isotherms of mean temperature. In the northern hemisphere, the winds at great heights circulate clockwise round regions of high mean temperature and anticlockwise round regions of low mean temperature. These circulations are reversed in the southern hemisphere.

(*e*) If the temperature is fairly uniform over a wide area, the winds in the free air will conform closely to the M.S.L. isobars.

(*f*) If the horizontal temperature gradients are steep at the surface, then the winds in the free air very quickly show decided shifts from the M.S.L. isobars.

Thermal Wind.—The upper wind may be regarded as a combination of the lower wind and the effect of the temperature distribution. For this reason the vector difference between the upper and lower winds is called the Thermal Wind. In Fig. 66, let *PQ*

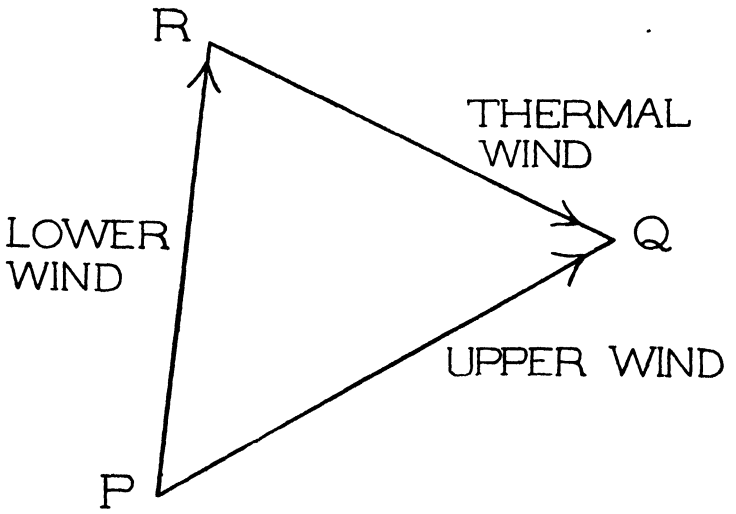


FIG. 66.—THERMAL WIND.

represent the upper wind and *PR* the lower wind. The thermal wind is then represented by the vector *RQ*. The thermal wind is parallel to the isotherms and has the higher temperature on its right in the northern and on its left in the southern hemisphere. Its magnitude is proportional to the temperature gradient. For example, a difference of 3° F. in mean temperature between Spithead and Torbay (100 miles) gives rise to a thermal wind of 12 knots at 6,000 feet. The thermal wind therefore bears a similar relation to the mean temperature distribution as the gradient wind does to the pressure distribution.

Normal Variation of Wind with Height.—Normally atmospheric temperature decreases from the equator polewards. Pressure gradients directed towards the pole, having the higher pressure associated with higher temperature, normally increase with height. Such gradients give westerly winds in both hemispheres. Therefore, normally, westerly winds increase with height. On the other hand, easterly winds which occur when the higher pressure is on the poleward or low-temperature side normally decrease with height, and sooner or later undergo reversal.

These conclusions are supported by observations which show that in general, westerly surface winds persist to the highest levels of the troposphere frequently with great increase of speed. Easterly surface winds are usually shallow, and are replaced by westerlies

in the first few thousand feet. The somewhat rare occasions of persistence of easterlies to great heights must be attributed to an abnormal temperature distribution in which the warmer air lies to the poleward side.

Deduction of Temperature Distribution from Observed Winds.—In practice it is simpler to observe the upper winds by pilot balloons or other means, than to observe the distribution of temperature from which the upper winds may be calculated. Rather it is more convenient to utilize the observed variations of wind with height for deducing the temperature distribution. Study of the upper-wind reports will often help to locate the warmer and colder air in a particular situation. For example, if the wind is observed to increase with height with little change of direction, then it may be inferred that the atmosphere is on the whole warmer on the high-pressure side than on the opposite. If the wind decreases with height the warmer air lies to the low-pressure side.

If the wind veers with height (in the northern hemisphere) through a range of height in the lower levels, the lower wind must be blowing from a region of higher to one of lower temperature. Warmer air will, therefore, be gradually brought in, increasing the lapse-rate at the top of the layer concerned. In an extreme case there may be a change from stability to instability, with accompanying cloud formation, and if the general lapse-rate is favourable, showers and possibly thunderstorms. If the wind backs with height, stability will be increased at the top of the layer in which the backing occurs. These conclusions will be true for the southern hemisphere if backing is substituted for veering, and *vice versa*.

Nephoscope observations (see page 317) are of assistance in determining the general temperature distribution since winds at great heights tend to flow perpendicular to the gradient of mean temperature. The stronger the wind, the more pronounced is the temperature gradient. Therefore rapid movement of high cloud indicates the existence of markedly high temperature to the right of the wind in the northern, to the left in the southern hemisphere.

The General Distribution of Pressure and Winds.

Prevailing Winds.—Although at most places on the earth's surface the wind is never constant in speed and direction, yet if the observations at any given place over an extended period such as one year are considered, they show that the wind has a pronounced tendency to blow most frequently from one direction, which varies in different parts of the world. In the British Isles, for example, the wind blows from between south-west and north-west on about two-thirds of the days of the year. The wind which is observed most frequently at any place is called the prevailing wind. In certain regions, such as the trade-wind belts the prevailing wind blows with remarkable steadiness; in others, such as the

temperate zones, it is intermittent, owing to the fluctuations of wind associated with the irregular incidence of cyclonic depressions ; in the equatorial regions and the sub-tropical belts of high pressure the winds are so light and variable that there can be said to be no prevailing wind. The prevailing winds of the earth's surface constitute a general circulation, which is related to the general average distribution of pressure in a manner precisely similar to that in which the actual winds at a given epoch are related to the simultaneous distribution of pressure, the relation being expressed by Buys Ballot's law. For example, in winter the average pressure becomes very high over the continent of Asia and the prevailing winds circulate around the high-pressure centre in a clockwise sense, so that, in China they blow from north-west or north and in India from north-east. These winds blow on most days throughout the winter, although they may be absent on any particular day, due to short-period fluctuations of the pressure which occur from time to time, such as the depressions of the Yangtze valley, on the east sides of which the winds are, of course, southerly. The relation between the prevailing winds and the general distribution of pressure is so close that the two can only be discussed together, which it is proposed to do in the succeeding paragraphs. The subject is best treated by considering firstly the pressure and wind systems that would exist if the earth's surface were level and uniform in nature, that is, consisting entirely of ocean or similar land surface. The resulting system is known as the planetary system of pressure and winds. Secondly will be considered the manner in which the planetary system is modified by the great land and sea areas of the earth's surface thus leading to the distribution of pressure and wind which is actually observed.

The Planetary System of Pressure and Winds.—No completely satisfactory explanation of the planetary system has yet been given, and the following only accounts in a general way for its principal features. On a uniform globe the effective factor for the production of pressure gradients and therefore of winds is the distribution of temperature, which in its turn is due to the variation with latitude of the solar radiation received at the surface. The effect of a horizontal gradient of temperature may be considered as follows. In Fig. 62 let AB and CD represent two columns of air of equal height, but with different mean temperatures, that of AB being the greater. Assume that the pressures at B and D are equal and that B and D are both at the surface level. It is shown on page 190 that under these conditions the pressure at A is higher than that at C, as a result of which air flows from A to C. This transfer of air in the upper level brings about an increase in the total weight of the air above D and a decrease in the total weight of the air above B. At the surface, therefore, the pressure rises at D and falls at B. In other words, high temperature gives rise to low pressure at the surface and *vice versa*. Now on a uniform earth the temperature

would diminish steadily from the equator, polewards. At the surface, therefore, under the influence of temperature alone, the pressure would be lowest at the equator and highest at the poles and vary gradually with latitude. This simple variation is, however, modified by the rotation of the earth, the effect of which may be indicated in the following way. In Fig. 62 assume that AB is situated on the equator and that CD is due north of AB, thus fulfilling the condition that AB shall be warmer than CD. Fig. 67

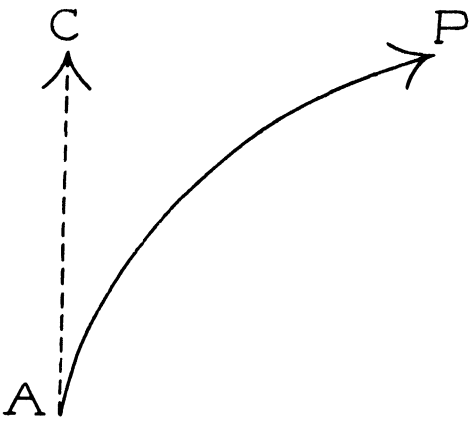


FIG. 67.

has been drawn in a horizontal plane at the level of A and C and it has been shown that air will tend to flow from A to C as indicated by the dotted arrow. As soon as the air leaves the equator, however, it becomes subject to the geostrophic force which deflects it towards the right of the direction in which it is moving, so that it is gradually deflected towards the north-east and finally towards the east, following a curved path AP. At P the tendency of the geostrophic force is to turn the air equatorwards, but this tendency is opposed by the gradient of pressure and the air settles down to move steadily from west to east with such a speed that the geostrophic force and the force due to the pressure gradient balance one another. The air moving outwards from over the equator therefore accumulates in the latitude corresponding to P, and at the earth's surface in this latitude the pressure accordingly becomes high. It may be shown that there will be a similar accumulation of air with high pressure in the corresponding southerly latitude. The two belts of high pressure are found to exist on the earth's surface and are centred in about 35° north and south latitude. There is a belt of relatively low pressure in the equatorial region. On the poleward sides of the high-pressure belts the pressure falls again, but since the temperature effect causes relatively high pressure at the poles, a minimum of pressure must occur in middle latitudes in each hemisphere. Observation shows that on the average, the minima occur

in latitude 60°. The M.S.L. pressure distribution on a uniform rotating globe therefore consists of :—

(*a*) an equatorial belt of low pressure,

(*b*) belts of high pressure centred in 35° north and south latitude,

(*c*) belts of low pressure centred in about 60° north and south latitude,

(*d*) caps of high pressure at the poles.

The winds corresponding to this pressure system, in accordance with Buys Ballot's law blow at an angle of from 20° to 30° to the isobars crossing them from higher to lower pressure. The system of pressure and winds is shown in Fig. 68.

The weak pressure gradients of the equatorial low-pressure belt give rise to a region of calms known as the Doldrums. Blowing in towards the doldrums from the high-pressure belts on either side are systems of steady winds known as the Trade-Winds. Owing to the deflectional effect of the earth's rotation these winds blow from north-east in the northern and south-east in the southern hemisphere and are therefore known as the north-east and south-east trades, respectively. The central parts of the high-pressure belts are regions of calm or light variable winds, known as the Horse Latitudes. On their poleward sides the winds are mainly from a westerly point, but owing to the fact that the low-pressure belts of temperate latitudes are not permanent but consist of a frequent succession of detached depressions the westerlies are more variable than the trades. The low-pressure belts are largely the result of the depressions which form through the interaction of the warm air of the westerlies and the cold air flowing outwards from the poles—they cannot be adequately explained on the basis of the planetary system alone. The prevailing winds of the westerlies are south-west in the northern and north-west in the southern hemisphere. The almost uninterrupted belt of ocean which surrounds the globe in southern temperate latitudes renders the westerlies very strong and persistent, and they have in consequence become known as the Roaring Forties. Little is known of the prevailing winds in the polar regions ; they are probably north-east in the Arctic and south-east in the Antarctic. The pressure- and wind-belts tend to swing north and south through about 5°—10° of latitude with the seasonal changes in the sun's declination.

Effects of Land and Sea Distribution.—The wind systems of the earth are the result of the modifications of the planetary pressure system due to the difference of temperature between sea and land. Over large land areas the temperature becomes very high in summer and very low in winter. Over the sea, the variation is much less than over land, so that the oceans are relatively warm in winter and cool in summer. This leads to relatively high pressure over

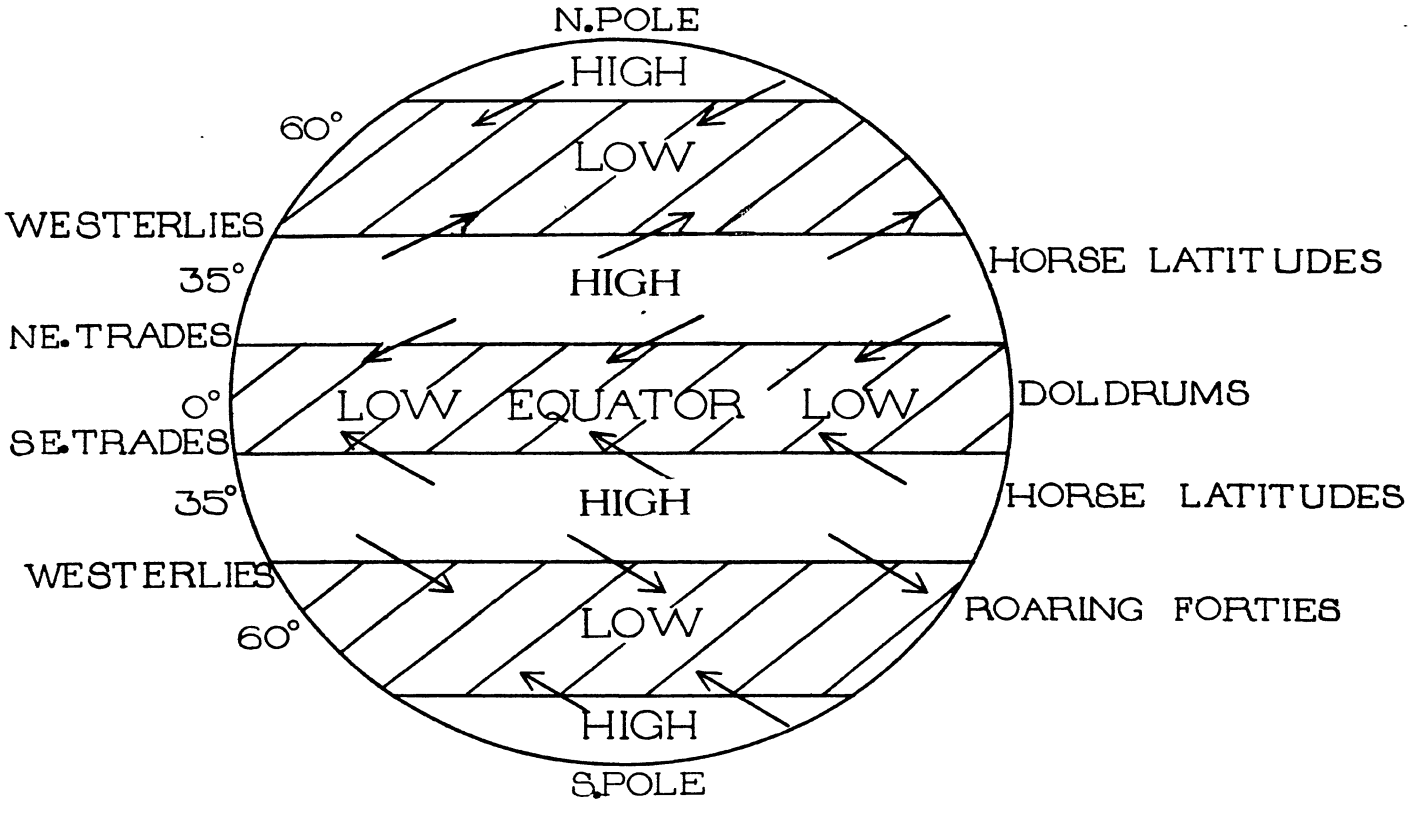


FIG. 68.—PLANETARY WINDS.

land in winter and low pressure in summer; over the sea the pressure is relatively high in summer and low in winter. The seasonal variations superposed on the planetary system modify it, resulting in the systems shown in Figs. 69 and 70, which show the pressure and winds for January and July, respectively.

Northern hemisphere.—Since the northern hemisphere contains the great land masses its winds are the more disturbed. In summer, gaps appear over the land in the sub-tropical belt of high pressure, leaving detached anticyclones over the oceans, which, owing to the relatively low sea temperature, become intensified. The trades do not therefore extend completely round the globe, but appear only on the eastern and southern sides of the oceanic highs. The large low-pressure system which develops over Asia so modifies the winds of the Indian Ocean and the China Seas that the north-east trades disappear from these regions, and give place to the south-west and south-east monsoons respectively. The low-pressure belts of the temperate zone are much enfeebled over the oceans, so that the westerlies are on the whole moderate, and often light and variable.

In winter the most pronounced feature of the pressure distribution is the intense anticyclone which develops over Asia, with its westward extension across Europe and north Africa. It exercises control over the winds of the Indian Ocean and the China Seas, and gives rise to the winter monsoon. Owing to the relative warmth of the oceans, large areas of low pressure known as the Icelandic and Aleutian lows develop over the North Atlantic and North Pacific respectively. They are due not so much to permanent low pressure, but to the fact that depressions most frequently develop in these regions. For limited periods, even in winter, Iceland and the Aleutian Islands may be covered by anticyclones, but on the whole depressions predominate and the average pressure is accordingly low. In conjunction with the continental highs, the oceanic lows so steepen the pressure gradients as to produce conditions of extreme storminess over the oceans and also at times over the adjacent land areas.

Southern hemisphere.—Owing to the greater uniformity of the surface in the southern, as compared with the northern hemisphere, the winds are much more nearly in accord with the planetary system and show comparatively little seasonal fluctuation. In summer, gaps appear over land areas in the sub-tropical high-pressure belt, but as compared with the Asiatic low they are insignificant. The roaring forties persist throughout the year, with greater violence in winter than in summer (southern hemisphere).

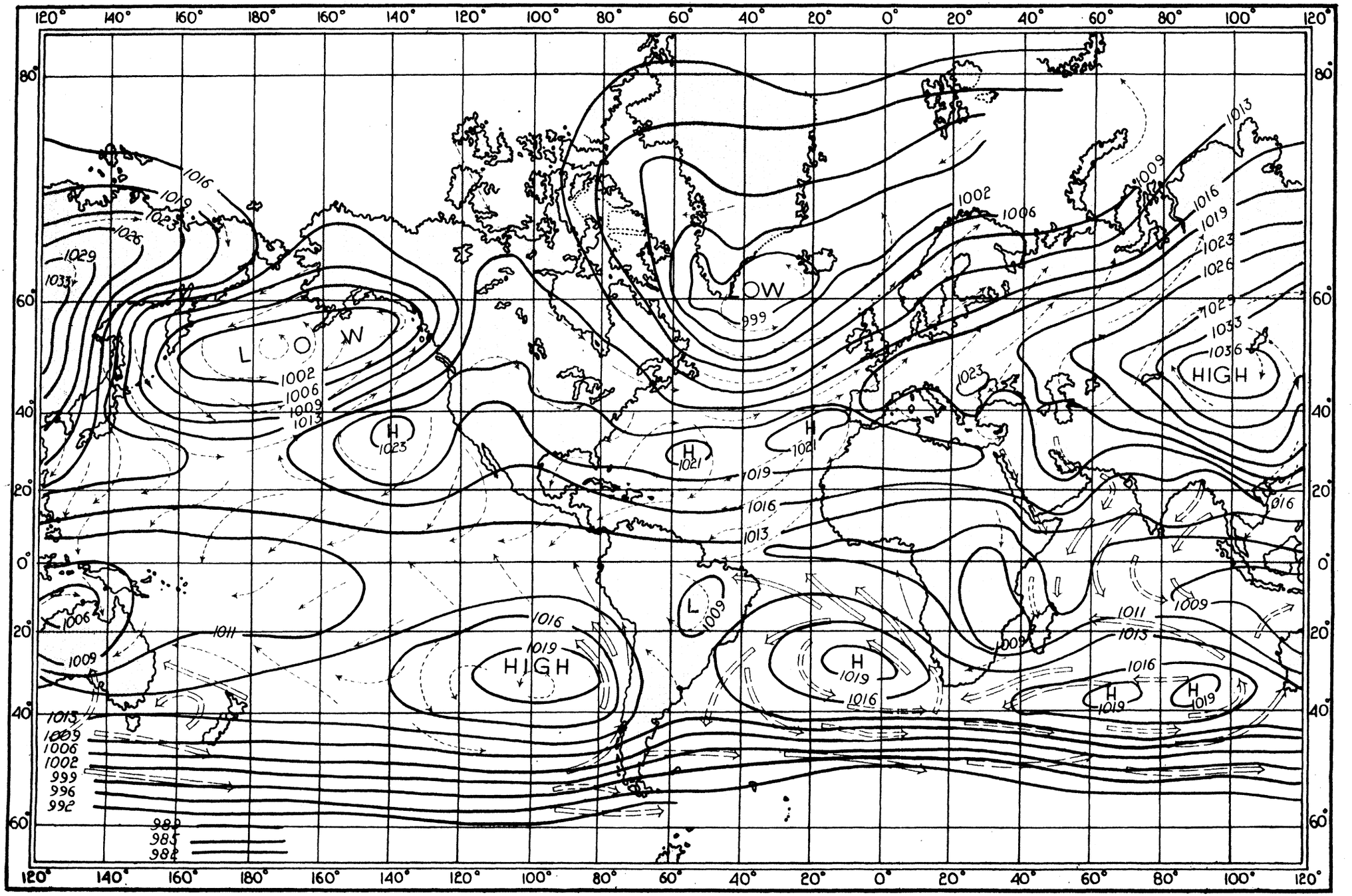
The Principal Wind and Pressure Belts.

The principal winds and the regions in which they occur are shown in Figs. 71 and 72.

FIG. 69.

PRESSURE AND WIND —— JANUARY.

To face page 202.



5064 WIND { STEADINESS: Frequency less than 50% of all observations. Frequency 50–75%. Frequency above 75%. Direction only.
FORCE { 1–3
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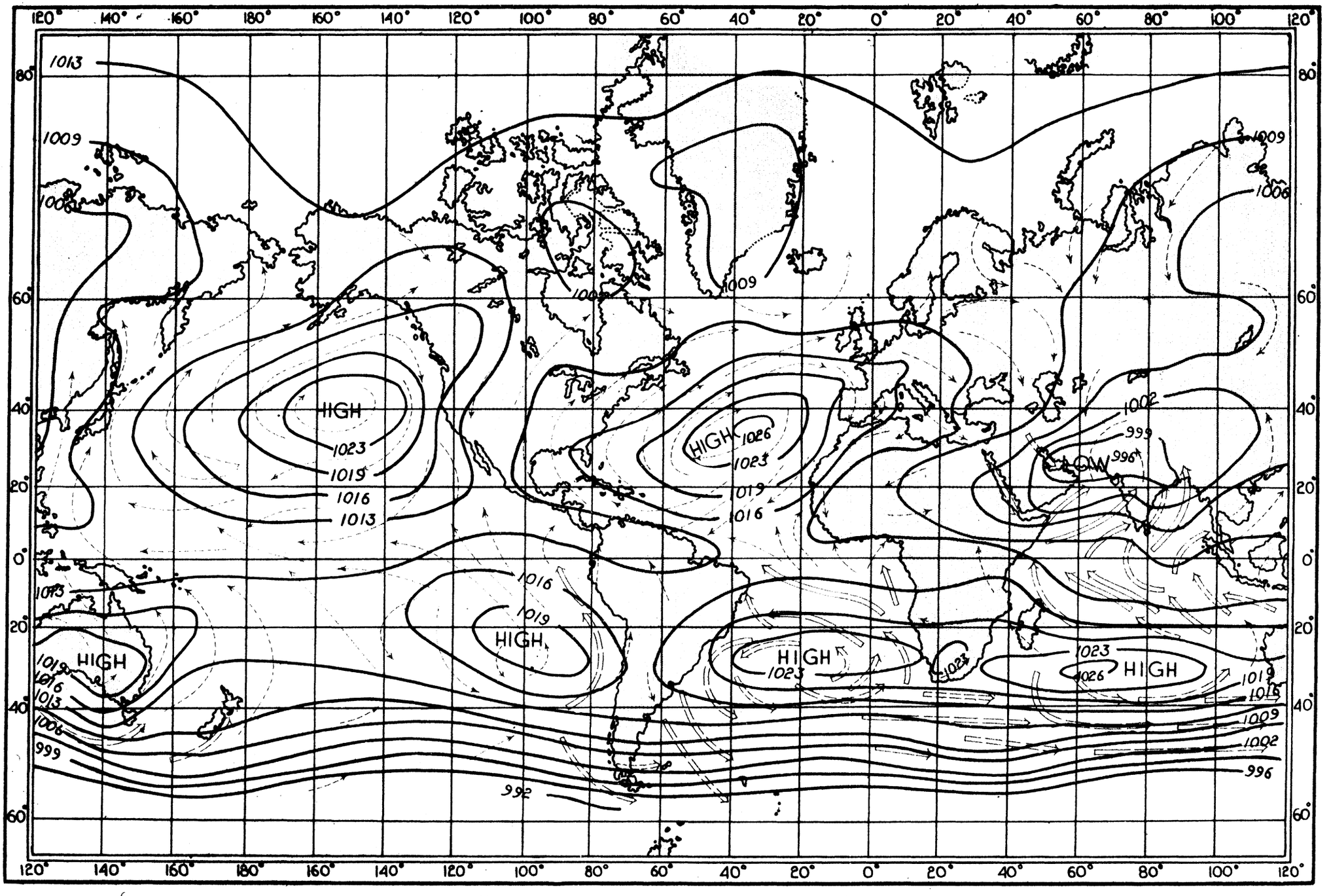
FIG. 70.

PRESSURE AND WIND. —— JULY.



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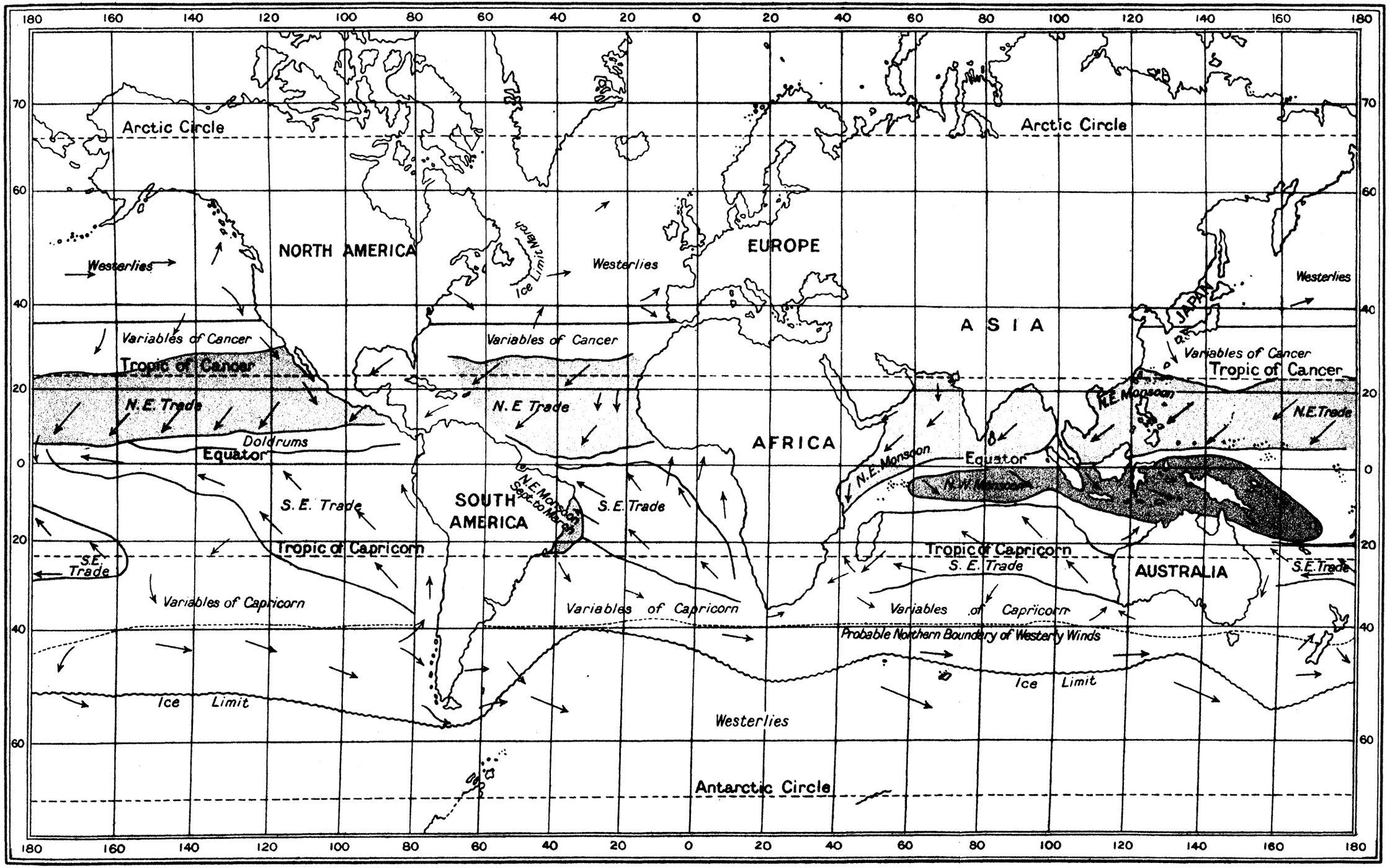


FIG. 71 — WIND CHART OF THE WORLD — NOVEMBER — APRIL.

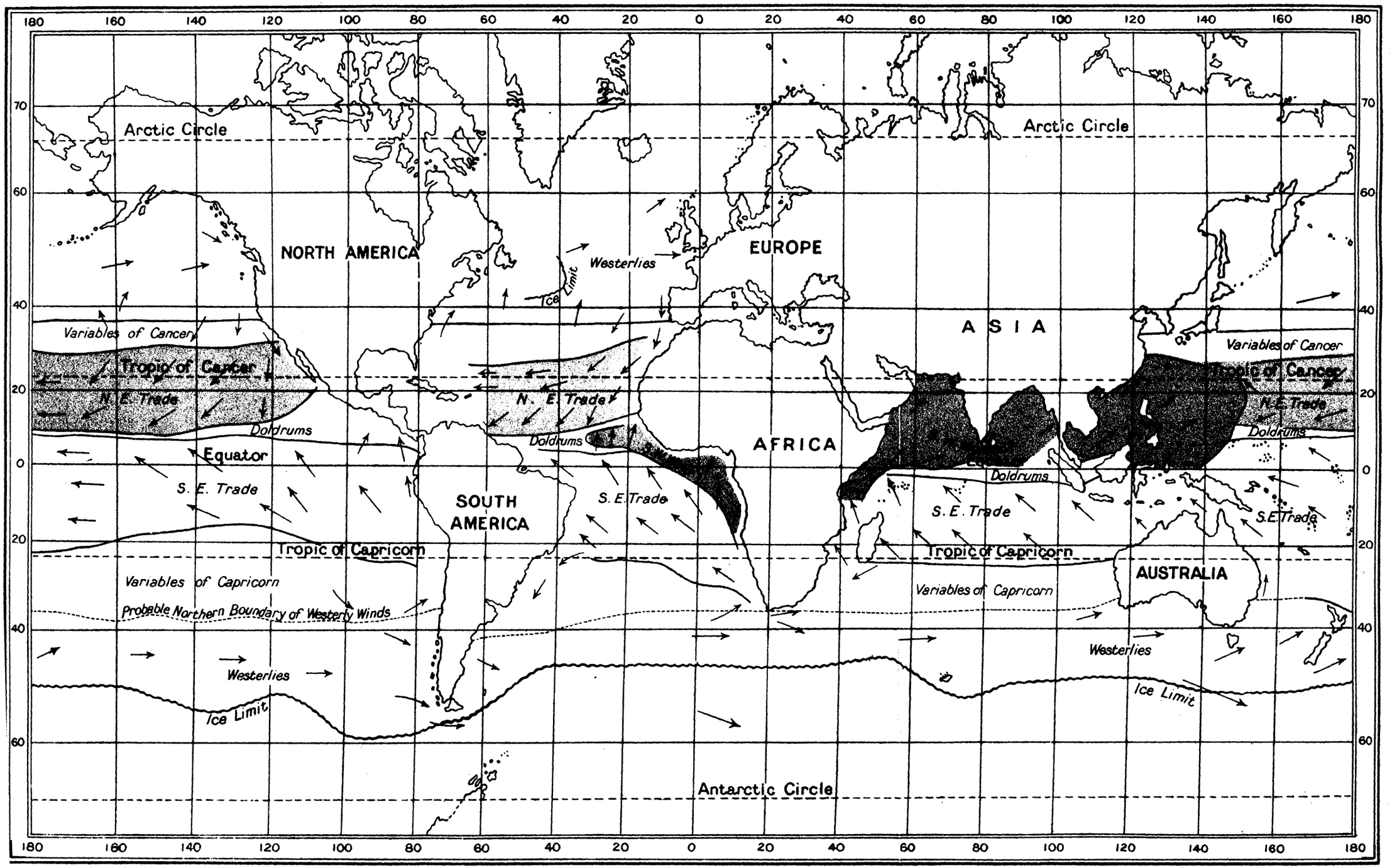


FIG. 72 — WIND CHART OF THE WORLD — MAY–OCTOBER.

The Doldrums.—The main pressure belts being primarily due to differences in the solar heating of the earth's surface, the doldrums tend to lie along the thermal and not the geographical equator, and are always to the north of the latter. In common with all the pressure- and wind-belts they tend to swing north and south with the seasonal changes in the sun's declination. Their limits are ill-defined ; approximate average values are given in the following table :—

	Limits of Doldrums.	
	January.	July.
Atlantic Ocean	Equator–2° N.	5° N.–10° N.
Pacific Ocean	4° N.–8° N.	8° N.–12° N.

In the northern summer, when the doldrums are farthest north, the south-east trades cross the equator and, owing to the change in direction of the deflectional force due to the earth's rotation, become south-westerly. In the Indian Ocean the predominating winds are the Asiatic monsoons and since their influence extends to south of the equator the doldrums proper do not exist there.

The region of the doldrums is one of ascending air currents fed by the trade-winds which converge on either side. The high temperature and abundant moisture are favourable to vigorous convection with copious rainfall, which is accompanied by thunder-storms of great intensity. Waterspouts (*see* Chapter XII) occur frequently, and it is probable also that tropical cyclones (Chapter XIX) originate within the region. On account of the uniformity of pressure, winds are generally calm or light. In the days of sail, ships often lay becalmed for days on end and frequently made long detours in order that they might pass through the doldrums at their narrowest part. On the coasts, land and sea breezes are prominent. The diurnal variation of pressure is greatest in the doldrums, and apart from it the time variation of pressure is small in that region.

The Trade-Winds.—The approximate limits of the trade-wind belts are as follows :—

Area.	Wind.	January.	July.
N. Atlantic Ocean..	NE. Trades	2° N.–25° N.	10° N.–30° N.
S. Atlantic Ocean ..	SE. Trades	Equator–30° S.	5° N.–25° S.
N. Pacific Ocean ..	NE. Trades	8° N.–25° N.	12° N.–30° N.
S. Pacific Ocean ..	SE. Trades	4° N.–30° S.	8° N.–25° S.
Indian Ocean ..	SE. Trades	15° S.–30° S.	Equator–25° S.

There is no north-east trade-wind in the Indian Ocean.

The trade-winds proper occur on the eastern and equatorial sides of the sub-tropical oceanic anticyclones.

Towards the western sides of the anticyclones the winds turn polewards, becoming south-easterly in the northern and north-easterly in the southern hemisphere, at the same time becoming less steady in speed and direction. In these areas, as for example the West Indies and the China Seas, tropical cyclones are frequent.

The trade-winds are remarkable for their steadiness, particularly in the Atlantic Ocean. At St. Helena, in the trade-wind belt of the South Atlantic, the wind is south-east on the average on 24 days in January and 19 days in July. The trades attain their maximum strength in each hemisphere in spring; the north-east trades are strongest in April and the south-east trades in September. The following table shows the average speed month by month of the trade-winds of the Atlantic Ocean.

Average Speed of Atlantic Trade-Winds (Knots)

Wind.	Jan.	Feb.	Mar.	Apr.	May.	June.	July.
NE. Trades	9	10	10	11	10	9	8
SE. Trades	12	11	11	11	10	11	11

Wind.	Aug.	Sept.	Oct.	Nov.	Dec.	Year.
NE. Trades	6	7	5	7	9	8·2
SE. Trades	13	15	13	14	13	12·1

The south-east trades are considerably the stronger. The trade-winds of the Pacific are somewhat less strong and more variable in direction than those of the Atlantic.

The trades bring air from higher to lower latitudes, and consequently the weather is brisk and invigorating, with the sky often cloudless or only partly covered by broken cumulus. Locally, in the trade-wind belts, there are exceptions to the usual weather. An example is furnished by the south-west coast of Africa. Over the heated land a semi-permanent area of low pressure develops and the south-east trades become deflected landwards. Owing to the low sea temperatures off this coast fog often occurs and is carried landwards by the deflected trades. The fog extends a hundred miles or so seawards and is thickest a few miles off shore. Similar foggy conditions are frequent in the trade-wind belts off other "cold-water coasts," notably off Chile, Peru, Morocco, and California.

Anti-Trades.—In the higher levels of the atmosphere the trade-winds undergo complete reversal, the upper currents being known as the anti-trades. On the Peak of Teneriffe (12,000 feet), for example, the prevailing wind is south-west, but at sea level the north-east trades blow almost continually. The height at which the reversal takes place appears to vary with season and latitude. Pilot-balloon observations made on board H.M.S. *Repulse*, in the South Atlantic in 1925, showed reversal of the south-east trade at about 3,000 feet At the Canary Islands, in 1911, the north-easterly surface wind was found to persist to the highest levels attained by observation in winter, but in summer the south-west anti-trades appeared at 7,000 feet. The source of the anti-trades is probably the rising air of the doldrums, which flows away polewards in the upper levels, but the flow seems also to be associated with the usual reversal of easterly winds with height, which arises on account of the general temperature gradients from the equator to the poles.

Monsoons.—The monsoons are so called on account of the fact that they are characteristic of the seasons, persisting from the same direction for long periods and undergoing complete reversal with change of season. They occur chiefly in south and east Asia, and are closely connected with the great changes of pressure which take place between summer and winter over the land mass. In the northern winter the pressure distribution is dominated by the intense Asiatic anticyclone, the resulting monsoon winds being north-westerly in north China, northerly in central China, and north-easterly in India. In summer the centre of lowest pressure occurs over north-west India, and there is a continuous gradient of pressure from the anticyclone of the south Indian Ocean to the Himalayas. Consequently the south-east trades cross the equator and, owing to the change of the earth's rotational deflection, reach India as a south-west wind. In the China Sea the same general current is experienced from south-east.

Monsoons of India.—Originating in the continental interior, the north-east monsoon is cold and dry, but considerable shelter is afforded to India by the mountain ranges of the north. It blows from October to March, during which period India has its cool dry season. It extends over the Indian Ocean, penetrating somewhat to the south of the equator, and appears in Australia as the north-west monsoon. From March until June, India experiences its hot weather, but by the latter month the pressure gradient for the south-west monsoon is fully established and the latter appears suddenly. Being moisture-laden, owing to its long passage over the warm sea, it brings abundant rainfall. The monsoon arrives at a given place at approximately the same date in each year and advances steadily north-eastwards, reaching Bombay early in June and the Punjab about a month later. The rains continue until the

reversal of wind in October. Numerous tropical cyclones occur in the India Ocean and the Bay of Bengal, more particularly at the beginning and end of the south-west monsoon, while feeble disturbances pass along the Ganges valley, so that the wind direction is not steady as in the trades. Marked changes of wind, amounting at times to complete reversal take place in the upper levels, the winter monsoon of India extending on an average to 6,500 feet and the summer monsoon to 16,000 feet. East Africa is, to some extent, affected by the monsoons which extend over the Indian Ocean southward and westward almost to the Mozambique Channel.

Monsoons of China.—Owing to its continental origin, the winter monsoon is a cold dry wind. It is not entirely free from cyclonic disturbances which bring snowstorms to north China. Dust-storms, originating in the deserts of the interior, are frequent. In summer the prevailing wind is south-easterly, and is accompanied by frequent typhoons.

Other Monsoons.—There are minor monsoon effects in other parts of the world, as, for example, in the Gulf of Guinea, where from June to September there is a south-west monsoon extending from Sierra Leone to Walvis Bay. In eastern Brazil a north-east monsoon blows from September to March. In the south-eastern part of the United States of America the prevailing winds are north-westerly in winter and south-easterly in summer. A slight monsoon effect occurs in Spain and Portugal.

Westerlies.—In the zones bounded approximately by latitudes 40° and 65° in the northern, and 35° and 65° in the southern hemisphere, the prevailing winds are westerly at all seasons of the year. They are stronger in winter than in summer, for in the former season the warm oceans are conducive to the development of vigorous depressions. The westerlies may extend beyond these limits with the seasonal swing of the wind- and pressure-belts, the Mediterranean, for example, falling within the westerlies in winter and the sub-tropical high in summer. Owing to the numerous depressions of temperate latitudes, the westerlies are extremely variable in direction, mainly from between south-west and north-west, but easterly winds persisting over long periods occur from time to time. The frequent changes of wind and the travelling depressions with which they are associated make weather conditions of the temperate zones very changeable from day to day, especially in winter, when the westerlies are strongest. South-westerly winds in January may raise temperatures in the British Isles to the neighbourhood of 55° F., followed after a few days by easterly winds which may keep the temperature continually below freezing point. The belts of the westerlies are zones of much cloud and rain, north-west Europe being one of the cloudiest regions of the earth's surface.

It is in temperate latitudes that the greatest contrasts of temperature arise between land and sea, and therefore that the

climate of the land areas is most affected by the direction of the prevailing winds, whether from sea to land or *vice versa*. This is especially the case on the western seaboards of the land masses, where the direction of the winds is from sea to land. The westerlies of the North Atlantic carry the oceanic influence far into the interior of western Europe, ameliorating the severity of winter and tempering the heat of summer. Regions which are reached by the westerlies only after a long passage over land, as for example, Labrador, are, however, conspicuous for the coldness of their winters, which may be contrasted with the mildness of the British Isles in approximately the same latitude. The distribution of land and sea in the belts of the westerlies, in determining the climate, is as important a factor as the prevailing wind itself.

The Horse Latitudes.—These are the regions of high pressure and variable winds which lie between the trade-winds and the westerlies. From their association with the tropics they are sometimes called the Variables of Cancer and the Variables of Capricorn. Owing to the persistent high pressure, weather in these latitudes is mostly cloudless and dry, and the winds very light. Over land, the absence of precipitation has given rise to the great desert belts, including the Sahara and California deserts in the northern, and the Kalahari and Australian deserts in the southern hemisphere.

The Polar Regions.—(*a*) *The Arctic.*—In summer the lowest temperatures and consequently the highest pressures at the surface occur near the pole. Such observations as are available show that the maximum of pressure at the pole is not well marked. The prevailing wind appears to be light north-easterly. In winter the highest pressures occur not at the pole but over north-east Siberia and to a less degree over northern Canada. A belt of high pressure joins these anticyclones, passing on the Bering Strait side of the pole. The winds conform to this pressure distribution and it would appear that the south-westerlies of the North Atlantic penetrate at times to the Arctic Basin between Greenland and Spitzbergen.

(*b*) *The Antarctic.*—The south polar region consists of a large elevated land mass covered by an ice cap and, owing to the large correction on account of altitude, such observations of pressure as have been taken on the ice cap are somewhat uncertain. The high frequency of south-east and south winds experienced by Antarctic expeditions points to a polar cap of high pressure, but the winds may to some extent be accounted for by the gravitational outflow of cold heavy air from the plateau. These winds attain considerable violence and cause storms of driving snow—the blizzards of the Antarctic.

The polar anticyclones are confined to the lowest layers of the atmosphere, since, owing to the low temperature, pressure falls off very rapidly with height. At a few thousand feet pressure is probably lowest at the poles themselves with westerly circumpolar

wind circulations. Observations of the smoke from Mount Erebus (13,000 feet) show a prevalence of north-westerly winds although the prevailing direction at sea level is south-east.

Special Winds.

Land and Sea Breezes.—Near the coast on warm days an on-shore breeze frequently sets in during the forenoon, rising to a maximum in the early afternoon and dying away in the evening. At night, particularly if the sky be clear, the opposite, an off-shore breeze, may spring up but it is usually less pronounced than the day-time wind. These winds are known as land and sea breezes. They are due to the different degrees of heating and cooling experienced by the adjacent land and sea under the influence of solar heating by day and radiation to the sky at night. It has been seen on page 190 that high temperature leads to high pressure in the upper levels, and low temperature to low pressure, resulting in a flow of air from land to sea in the upper levels during the day. This transfer of air causes the surface pressure to rise over the sea and fall over the land; consequently, there is a flow of air from sea to land, which is experienced as the sea breeze. At night the higher surface pressure is over the land and the flow of air is reversed.

Land and sea breezes occur chiefly in tropical countries where the solar heating is powerful. They are experienced in the British Isles and other countries in temperate latitudes during warm summer weather. The sea breeze may reach force 4 or 5 in tropical and Mediterranean countries; in the British Isles it rarely exceeds force 3. Only a narrow strip some 10–15 miles on either side of the coast line is affected and the breezes disappear at a height of 500–1,000 feet. If the general pressure gradient is steep the land and sea breezes may be entirely masked.

The local pressure gradient which gives rise to land and sea breezes is directed across the coast-line whence it might be thought that, in accordance with Buys Ballot's law the wind should be along shore. Being of a transitory nature, however, the land and sea breezes do not adjust themselves readily to the pressure gradient and consequently blow more or less directly from sea to land and *vice versa*.

Katabatic Wind.—On clear nights there is often a downward flow of air on mountain and hill slopes, and particularly along valley bottoms, known as a katabatic wind. It is caused by the gravitational flow of air which is cooled by contact with the radiating land surface.

Let AB, Fig. 73, represent the slope of a hillside, C any point on the hill and D a point in the free air at the same level as C. The ground loses heat freely to the clear sky at night and the fall of temperature is communicated by contact to the air at C. This air, becoming denser than that at D, tends to sink, and since

vertical sinking is impossible the flow takes place along the slope.

The descending air is warmed dynamically 5½° F. for each 1,000 feet of descent, and since the lapse-rate in the free air is generally less than this the descending air may be expected to become warmer than the free air at the same level and be brought to rest. But it also continues in contact with the cold surface and the consequent

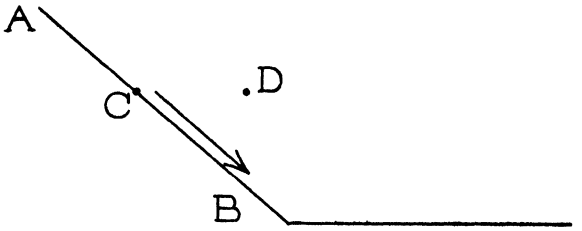


FIG. 73.—KATABATIC WIND.

cooling renders the dynamical warming largely ineffective so that the katabatic wind is maintained. Normally katabatic winds do not attain great force. If, however, the slope is covered with snow or ice the descending air is strongly cooled and may attain a considerable speed. Particularly if the slope is steep as on the coasts of Greenland and Antarctica, the flow takes the form of a cataract of cold air. Katabatic winds are experienced at sea near to all mountainous coasts by day and night.

Anabatic wind.—The anabatic wind is the gentle upward flow of air on hill slopes on fine warm days. It is caused by the warming of the air by contact with the heated surface. The ascent cools the air dynamically, but near to the ground the continued contact prevents the temperature from falling as rapidly as in the free air, so that the ascending motion persists. The dynamical cooling may ultimately bring about condensation with cloud formation on the summit of the mountain.

REGIONAL WINDS.

The following is a list and also a brief description of the principal local winds that occur in different regions of the earth. With the exception of one or two winds which occur in the interior of the great continents, such as the "buran" of Russia and Siberia, the "seistan" of eastern Persia, and the "tornado" and "blizzard" of North America, they are practically all coastal or maritime winds, and are therefore appropriate to this book. Fuller details of these regional winds are contained in the various handbooks of local weather for each of the naval stations.

Mediterranean Sea and Southern Europe.

Levanter.—The levanter is an easterly wind which is associated especially with the Strait of Gibraltar and with the channel between

Spain and Morocco. On the east coast of Spain the wind may be from the north-east. The associated pressure distribution generally shows high pressure over central Europe and a depression over the south-west Mediterranean or African coast. The levanter is usually a light or moderate wind, and seldom blows with gale force, but it brings with it excessive moisture, heavy dews, local cloud, a thick muggy atmosphere and sometimes rain. It is most frequent during the period July to October and in March ; it may, however, occur at any time of the year.

Vendavales.—The vendavales are strong south-westerly winds which occur in the Strait of Gibraltar and the region off the east coast of Spain. They are associated with advancing depressions of late autumn to early spring ; they are usually introduced by, and are liable to be interspersed with, thunderstorms and violent squalls.

Leveche.—This is a dry, scorching, sand- and dust-laden wind from between south-east and south-west, peculiar to the south-east coast of Spain between Almeria and Valencia ; it extends only a few miles inland, and occurs in front of an advancing depression.

Llevantades.—Gales from between north-north-east and east-north-east are the most important gales of the east coast of Spain. They are known locally as llevantades, and are an intense form of the llevante or levante, *i.e.*, north-easterly winds of long fetch, as opposed to diurnal coastal breezes. These gales are most frequent and dangerous in spring and autumn (February to May and October to December), and are generally associated with slow-moving depressions crossing the Mediterranean between France and Algeria.

Bize.—The bize is a cold, dry wind which blows in the winter in the mountainous regions of southern France from the north, north-east or north-west. Its origin is similar to that of the mistral, but it is frequently accompanied by heavy clouds.

Leste.—The leste is a hot, dry, southerly wind which occurs in the region between the Strait of Gibraltar and Madeira, and along the north African coast in front of an advancing depression.

Mistral.—Mistral is the name given to strong northerly winds which occur in the Gulf of Lions and over the adjacent coastal land regions, and in particular in the Rhone valley. The mistral is usually associated with high pressure over north and west France and low pressure over the Tyrrhenian Sea, but the wind due to the pressure distribution is probably reinforced by the katabatic outflow of cold air from the mountains. In spite of the warming due to descent, the mistral is cold in comparison with the warm Mediterranean, and it may bring freezing temperatures to the south of France. The air is, however, usually dry, so that the weather is generally bright and clear. The mistral often reaches gale force

in winter, and gales of 80 to 85 knots may occur at sea. The wind speed is often far in excess of the gradient wind, and the direction may be perpendicular to the isobars. The strongest mistral is experienced in the vicinity of the Rhone delta, and there is usually a rapid decrease seawards, but occasionally it may extend to Malta and north Africa. Its power of raising a heavy sea in a short time constitutes it a serious danger to navigation.

Bora.—The bora is a wind, in some respects similar to the mistral, which occurs along the eastern shore of the northern Adriatic. It is a north-easterly wind occurring chiefly in winter, when it may attain gale force. It may occur :—

(i) behind the cold front of a depression moving south-east over the Adriatic ;

(ii) with persistent high pressure over Europe and low pressure over the Mediterranean ;

(iii) with high pressure over the Balkans.

In cases (i) and (ii) the bora is only violent if the Mediterranean " low " is well developed and appears to be due to the reinforcement of the normal north-easterly wind by katabatic outflow from the mountains of Dalmatia. In case (iii) the bora is local, being confined to a coastal strip, and it weakens rapidly with increasing distance over the sea. In Italy the bora may bring cold dull weather with squally showers of snow or sleet, or the weather may be clear and cold. Violent gusts and squalls sometimes reaching 100 knots occur on the eastern side of the Adriatic, and these constitute a danger to shipping, especially as they often set in with little or no warning. The name bora is given to a similar wind which is experienced off the mountainous north-east coasts of the Black Sea.

Scirocco.—This is the name given to the southerly wind in advance of a low-pressure system moving eastwards in the Mediterranean. It originates in the desert regions of north Africa, and reaches the sea as a very hot and dry wind, often heavily dust-laden. On crossing the sea it picks up moisture readily owing to its high temperature and reaches Malta hot and excessively humid —conditions which are particularly disagreeable. At night, cooling produces stratus cloud-formation and copious dew deposit. The scirocco often extends to the Adriatic, where it may reach gale force, and is often accompanied by heavy rain. Owing to continued cooling by passage over the sea, fog frequently forms in the northern Mediterranean.

Gregale.—Gregale is the strong north-east wind of the central and west Mediterranean, occurring chiefly during the cool seasons. It is particularly important at Malta, owing to the northerly aspect of the harbours there and the heavy seas which it is capable of raising. The accompanying weather may be fine or showery. with

rain and hail. The gregale is associated with pressures above normal to the north-west or north, and is most frequent during the winter. It may blow continuously up to 5 days, but most commonly lasts only 1 or 2 days.

Föhn.—The föhn is a warm dry wind which is experienced to leeward of high ground and, in particular, in northern Alpine valleys. Moisture-laden southerly winds blowing against the Alps are forced to ascend their southern slopes, soon becoming saturated and undergoing reduction of temperature at the saturation adiabatic lapse-rate of about 3° F. per 1,000 feet. Cloud is formed and the excess moisture precipitated on the southern slopes. Having attained the summit the winds descend on the far side and undergo increase of temperature at the dry adiabatic lapse-rate of about 5½° F. per 1,000 feet. They therefore appear as warm winds in the northern valleys, and having precipitated much of their moisture, are, in addition, very dry. In Fig. 74, let *A B C* represent

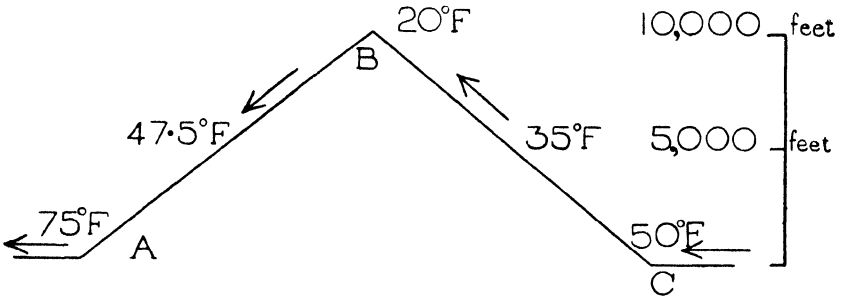


FIG. 74.—FÖHN WIND.

the mountain ridge, height 10,000 feet, and assume that the air arriving at *C*, on the southern side, is saturated, having a temperature of 50° F. The diagram shows the temperature of the air at the summit and on both slopes at the base, and half way between base and summit. From *C* to *B* the air cools at the rate of 3° F. per 1,000 feet; from *B* to *A* its temperature rises at the rate of 5½° F. per 1,000 feet.

The föhn is of great importance in the Alpine valleys on account of the damage it may bring about by the rapid melting of snows. The föhn effect is, however, extremely common in many parts of the world, and on the eastern side of the Rockies of Canada and the U.S.A. it is known as the Chinook. Even in the British Isles there are slight föhn effects, shown by the relatively high temperatures experienced at places like Aberdeen and the resorts of the North Wales coast on days of westerly or south-westerly winds.

Solano.—The solano is an easterly or south-easterly wind which brings rain to the south-east part of the coast of Spain and the Strait of Gibraltar.

Marin.—This is a strong wind in the Gulf of Lions blowing from a south-easterly direction, and is next in frequency and importance to the mistral in this region. It is generally warm, moist and cloudy, with rain and thick weather, and is associated with depressions which enter the Gulf of Lions from the west or south-west after traversing southern France and northern Spain.

Libeccio.—The libeccio is the westerly or south-westerly wind which predominates in northern Corsica all the year round; it frequently raises high seas and may give violent westerly squalls. In summer it is most persistent, but in winter it alternates with the tramontana (north-east or north).

Tramontana.—This is a local name for a north-easterly or northerly wind which in winter is prominent on the west coast of Italy and fairly prevalent off the north of Corsica. It is a fresh wind of the fine-weather mistral type, and does not often reach gale force. It is associated with a depression over the Adriatic simultaneously with an anticyclone further west.

Maestro.—The maestro is the name given to a north-westerly wind which blows in summer in the Adriatic when pressure is low over the Balkan peninsula. It is a fresh breeze accompanied by fine weather and light clouds.

Meltemi (Etesians).—During the summer in the Aegean Sea, the prevailing winds of the main circulation are due chiefly to the deep continental depression centred over the north-west of India. These winds are known as meltemi by the Turks and etesians by the Greeks. They blow from a direction which may be anywhere between north-east and north-west according to the character of the country surrounding the region concerned; meltemi weather is ordinarily fine and clear, the northerly winds tempering the fierce summer heat of the region.

Khamsin.—This name is generally understood to apply to dry, dusty and mostly hot southerly winds in Egypt, and the Red Sea, similar to the dry scirocco further west. The khamsin is not limited to any single wind direction, but is associated, rather, with the dryness, the heat and the dust-laden atmosphere. Khamsin winds blow in front of depressions moving eastward over the Mediterranean or through northern Africa, and the name is frequently applied both to the depressions themselves and to the characteristic weather which accompanies them. They are most frequent in the spring, from February to June.

Red Sea, Persian Gulf, and Arabian Sea.

Kharif.—During the period of the south-west monsoon, a strong south-westerly wind gets up daily about 2200, local time, and blows until noon on the following day on the southern shore of the Gulf of Aden in the vicinity of Berbera. This wind, which is known

locally as the kharif, is frequently of gale force during the months of June, July, and August.

Simoom.—This is the name given to the stifling, dry, dust-laden scirocco wind which originates in the Arabian desert and penetrates into Palestine and Syria as a south-easterly or easterly wind. The simoom is most frequent from April to June and from September to November inclusive ; in the Sahara desert it is also encountered in the form of a dust-storm and the air is filled and darkened by the whirling dust clouds. A similar type of dust-storm which occurs in the northern and central Sudan is known as a **Haboob.**

Belat.—The belat is a strong north-westerly wind which blows seawards on the south coast of Arabia. It is most prevalent in winter and early spring.

Shamal.—The prevailing winds which blow in summer down the Mesopotamian corridor to the low-pressure system of southern Asia are known as shamal. These north-westerly winds are remarkably constant and especially strong during the day, when they carry clouds of dust ; during their occurrence the sky is cloudless, the air very dry, the sunshine fierce and the heat intense. There is, however, a marked lull in the winds at night.

Seistan.—This is a strong northerly wind which blows in summer in the province of Seistan in eastern Persia. It continues for about four months and is therefore known as the " wind of 120 days." It sometimes reaches gale force.

Kaus.—In winter the prevailing winds in the region of the Persian Gulf are south-easterly and are known as kaus. They are most frequent between December and April. Sometimes they are followed for a few hours by a very strong wind from the south-west, called **Suahili,** which is dangerous for small craft.

Indian Coast.

Elephanta.—This is the name given to a strong southerly or south-easterly wind which blows on the Malabar coast of India during the months of September and October and marks the end of the south-west monsoon.

Malacca Strait.

Sumatra.—The sumatras are strong squalls, accompanied by violent thunder, lightning and rain, which occur in the Malacca Strait during the period of the south-west monsoon. The warm seas and the many mountains of this region cause the rainfall to be very heavy and the region as a whole is one of the rainiest on earth.

Australia.

Brickfielder.—The south of Australia is sometimes visited by troughs of low pressure moving from west to east, with the " V " of the depression pointing north. In advance of the depression

the wind is northerly, and blows from the interior of the Australian continent carrying heat and dust like the scirocco of the Mediterranean. Such winds are known as brickfielders.

Southerly Buster.—Closely associated with the brickfielder is the southerly buster. In the rear of the trough of low pressure the winds are southerly and appear very cold by contrast with the warm northerly winds in advance of the depression, although, as the southerly air stream comes from the ocean its temperature is well above freezing-point. The southerly buster is particularly prominent on the coast of New South Wales, especially about Sydney. After a day or more of hot, sultry weather with a northerly wind there is a short lull, and then very suddenly, when the trough has passed, a strong, often violent, wind sets in from the south. A striking roll of cumulus cloud may accompany the buster, and there is usually heavy rain. The phenomena which are similar to the line-squalls of our latitudes, are most frequent in spring and summer.

West Africa.

Tornadoes.—On the Congo and Guinea coasts of West Africa violent storms occur known as tornadoes. These storms are in the nature of thunder squalls; they often start very suddenly, last but a short time, sometimes only a quarter of an hour, and may do considerable damage on land and sea. They almost always travel from east to west, and are especially frequent at the beginning and end of the rains, *i.e.*, in the late spring and autumn. They are generally accompanied by exceedingly heavy rain and blinding lightning, but dry tornadoes also occur. The tornadoes of this region must not be confused with the storms of the same name in the U.S.A. These latter have, in general, less rain, and far more violent winds revolving rapidly within a short radius.

Harmattan.—The harmattan is a dry wind associated with the north-east trades and occurs during the dry season (November to March) on the west coast of Africa in the neighbourhood of the Cape Verde Islands and the Gulf of Guinea. The wind direction is such that it comes from the Sahara, carrying with it fine dust which may be deposited on ships at sea. Being both dry and relatively cool, it affords a welcome relief from the damp heat of the tropics, and from its health-giving powers it is known on parts of the littoral as the Doctor, in spite of the fact that it carries large quantities of dust from the desert. This dust is often carried in sufficient quantity to form a thick haze which may seriously impede coastal navigation.

South America.

Williwaws.—This is the name given to the very violent squalls encountered in the Strait of Magellan at all seasons, but notably during the winter months.

Pampero.—Pampero is the name given in the Argentine and Uruguay to a severe storm of wind, with rain, thunder and lightning. It is a line-squall with the typical arched cloud along its front. The pampero is associated with depressions passing across the Argentine from west to east. In advance of the depression the wind is northerly in direction, warm and rainy, and is known as the **Zonda.** As the depression passes the wind shifts suddenly to a southerly or south-westerly quarter, which brings with it a much lower temperature. The pampero may be compared with the southerly buster of Australia, which appears to be of precisely the same type.

Central and North America.

Norther.—The norther of the Gulf of Mexico and Central America is a cool dry wind associated with the advanced portion of an anticyclone over the southern United States and the rear of a depression over the western Caribbean Sea. Northers are most frequent during the colder months of the year, or during that portion of the year when the intense anticyclones of the United States move farthest southward. In force, the norther sometimes attains a severity which renders it dangerous not only to ships in the open sea and in some of the continental harbours, both Atlantic and Pacific, but also, though less frequently, even in the north coast harbours of some of the West Indian Islands, and as far to the southward as the Isthmus of Panama. In the Pacific the norther is known locally as the **Tehuantepecer** in the Gulf of Tehuantepec, and as the **Papagayo** on the north-west coast of Costa Rica.

Santa Ana.—This is the name given to the hot and occasionally violent winds sometimes experienced in southern California. These winds blow towards the Pacific coast from the deserts to the east of the Sierra Nevada and acquire additional heat by their descent; they are most frequent in winter, and are exceedingly hot, dry and dusty. They are akin to the khamsin of Egypt and Arabia.

Tornadoes.—The whole of the United States east of the Rockies is liable in summer to severe storms of wind called tornadoes. They are described in Chapter XII. Tornadoes are most frequent in the valleys of the upper Mississippi and Missouri, and are associated with depressions where the interaction of winds with a large difference in temperature gives opportunity for violent vertical convection; they occur also in Australia, and are not unknown in Europe, although less intense.

Blizzard.—The name blizzard is applied to any strong wind accompanied by great cold and drifting or falling snow. Blizzards are frequently experienced over the great cold regions of the earth, viz., the polar regions, and the interior of the North American and Asiatic continents in winter; the most severe blizzards are probably those of the Antarctic. In Russia and Siberia such storms are known as **Burans,** while in the tundras they are called **Poorgas.**

CHAPTER IX.

Visibility and Fog

Factors Affecting Visibility.—Visibility in the atmosphere depends upon the object looked at, its background, its illumination and the degree of atmospheric obscurity between the observer and the object.

Atmospheric obscurity, the factor with which the meteorologist is primarily concerned, is produced by the presence of a large number of minute particles held in suspension in the air. These particles may be separated into two main classes :—

(*a*) Solid particles of smoke, wind-blown sand or dust of volcanic origin.

(*b*) Water droplets or ice particles.

Large numbers of solid particles are always present in the lower layers of the atmosphere. Typical figures (for London) are, on a clear day, 300 particles per cu. cm. and on a foggy day 60,000 particles per cu. cm. The average particle diameter is about 10^{-4} cm. and the volume occupied by the particles is extremely small. The maximum of 60,000 particles quoted above occupy only 10^{-9} cu. cm. The mass of solid impurity varies up to 1 milligram per cu. m.

Water droplets are not readily produced in a mass of pure air even if the air is supersaturated. The presence of minute particles of certain substances which have the power of attracting moisture is essential. These, known as hygroscopic particles, act as nuclei on which condensation of the water vapour in the air may take place. They consist chiefly of certain compounds of sulphur and other products of combustion present in smoke from factory and domestic chimneys, and of salt from sea spray. Condensation may take place on hygroscopic nuclei when the relative humidity is as low as 75 per cent. but for extensive fog formation saturation of the air is usually necessary.

As the humidity approaches 100 per cent. the nuclei are swollen to perhaps twice their original diameters. When this limit is surpassed a few of the nuclei absorb water rapidly and grow to perhaps 100 times their original diameters.

The number of particles present in the air may be estimated by means of a nucleus counter, which is an instrument for counting the particles in a known volume of air. In Aitken's nucleus counter the known volume of air is passed into an expansion chamber and adiabatically cooled below its dew-point. Condensation of the excess water vapour takes place, the water droplets forming on the hygroscopic nuclei that are present. The droplets settle on a glass reticle and may be counted with the aid of a microscope. In Owens' dust counter a jet of damp air is forced through a narrow slit in front of the microscope cover glass. The fall of temperature due to expansion of the air after passing the slit causes the formation of a film of moisture to which the dust particles adhere. They are

then counted under the microscope. Differences in the results obtained with the two instruments, together with the fact that in Aitken's instrument water droplets are counted while in Owens' instrument dust particles are counted, have led to the conclusion that the water droplets condense on invisible nuclei and not on dust particles. Fogs are therefore due to the presence of hygroscopic nuclei in the air. The frequent and dense fogs experienced in large cities where, in the smoke, hygroscopic nuclei are abundant, support this conclusion. Solid dust particles on the other hand are usually associated with the hazy conditions of dry weather. They do not of themselves cause dense fog, but they may intensify fog caused by water droplets.

The particles or droplets which cause fog are so minute that they have no appreciable velocity of fall but virtually remain in suspension in the air for indefinite periods. Rain, consisting of comparatively large water drops, snow, and hail do not as a rule, unless falling heavily, reduce visibility below about one mile ; frequently in falling rain visibility is as much as five miles. Drizzle, which consists of numerous small water droplets, sometimes so small as to be scarcely distinguishable from fog particles, often causes very bad visibility. The smaller and more numerous the droplets the more effective do they appear to be in impairing visibility.

Illumination and background are important. An object which contrasts with its background is more easily seen than one that does not. On a sunny day visibility is better looking away from the sun than towards it ; on a cloudy day visibility is usually the same in all directions except in so far as it is affected by local artificial causes such as the drift of smoke from large cities.

A large lapse-rate of temperature and a strong wind are favourable to good visibility. The former is associated with vigorous convection which carries the particles upwards from their source at the surface and reduces their concentration in the lower levels. Over land the best visibilities occur on summer afternoons when normally the lapse-rate is highest. High wind also reduces the concentration of particles in the lowest levels by scattering them horizontally. At inland stations visibility improves with increasing wind speed and winds above 20 knots are rarely accompanied by poor visibility. At coast stations and in the open sea, however, the higher winds raise quantities of spray which diminishes the visibility near the surface. A small lapse-rate, especially. an inversion, and calm or light wind, because they do not dissipate the particles, are favourable to bad visibility.

Definitions of Fog, Mist, Haze.—Fog is said to be present when the horizontal visibility is less than 1,100 yards. Visibilities between 1,100 and 2,200 yards are described either as mist or haze. The former, due to water drops, is associated with high relative humidity and the latter due to dust or smoke particles usually with low relative humidity.

Conditions for Fog Formation.—Experiments with dust counters have shown that there are always sufficient hygroscopic nuclei present in the air to cause condensation of water vapour to the extent necessary for fog formation if other conditions are suitable. Therefore, if air is cooled below its dew-point condensation will take place and water droplets or ice particles will be formed. The condition for fog formation is therefore that a sufficiently large volume of air shall be cooled below its dew-point.

The cooling of a large volume of air is generally brought about by two distinct processes. Firstly, a limited portion of the air is cooled by proximity to a cold surface of land or sea over which it passes. Secondly, the cooling is diffused throughout the entire mass by turbulent mixing (*see* Chapter VIII). The first process is aided by calm or light winds which allow the air to remain for long periods over the cold surface. For the second process a light wind, since it involves the necessary degree of turbulence, is more favourable than a flat calm. On the other hand, strong winds, involving a high degree of turbulence, diffuse the cooling through too great a volume of air thus rendering it ineffective.

Under certain conditions condensation may be brought about by the mixing alone of two air masses which are nearly but not quite saturated.

In Fig. 75 let AB represent the curve of saturation values of

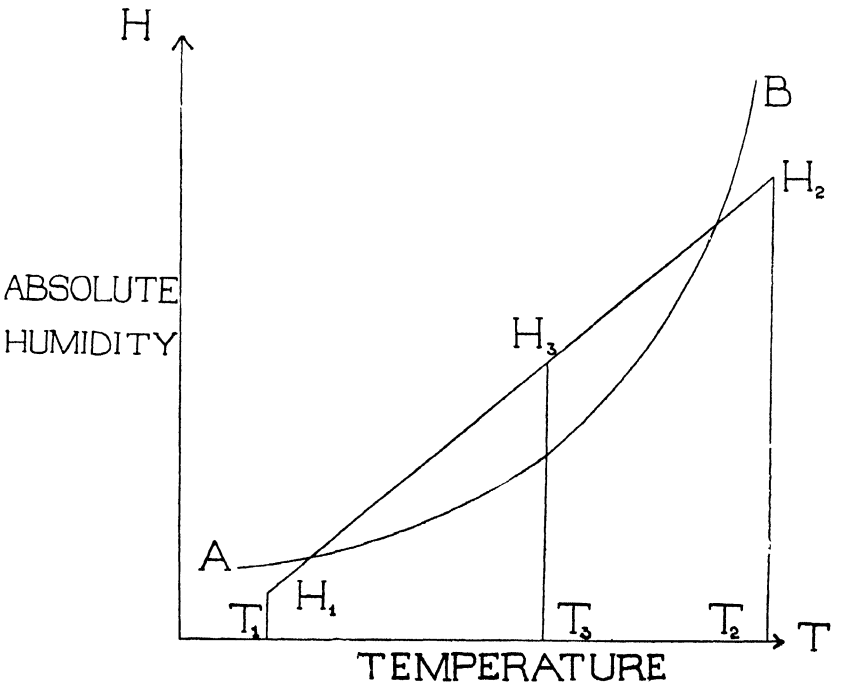


FIG. 75.—FORMATION OF FOG BY MIXING.

absolute humidity for air plotted against temperature (Chapter VI). Let a mass of air whose temperature is T_1 and absolute humidity H_1 be mixed with another mass whose temperature is T_2 and absolute humidity H_2. If the two air masses are not quite saturated the points H_1 and H_2 lie just below the curve AB. It may be shown that after mixing, the resulting air mass has a temperature T_3 between T_1 and T_2 and absolute humidity H_3, such that H_3 lies on the straight line joining H_1H_2. Since the curve AB is convex to the temperature axis, the line H_1H_2 may lie above AB for part of its length, and H_3 may therefore be greater than the saturation value corresponding to T_3. The excess water vapour accordingly condenses out on mixing of the air masses.

To sum up, the conditions for the formation of fog are :—

(*a*) A supply of sufficiently moist air ;
(*b*) A cold surface of land or sea ;
(*c*) Turbulent mixing within the air mass.

The last factor is aided by a light wind and small lapse-rate or inversion of temperature.

Types of Fog.

Radiation (or Land) **Fog.**—Land fog is the most common of all fogs. It forms as a result of proximity of moist air to a land surface cooled by radiation to a clear sky at night. The surface cooling gradually builds up an inversion of temperature from the ground upwards, which increases in depth as the higher layers of air become affected. The lowest layers in time become cooled below their dew-point, and fog forms in them. Further cooling takes place by radiation from the fog particles, and the fog therefore steadily increases in depth. The distribution of temperature with height is usually similar to that shown in Fig. 76 in which the inversion of

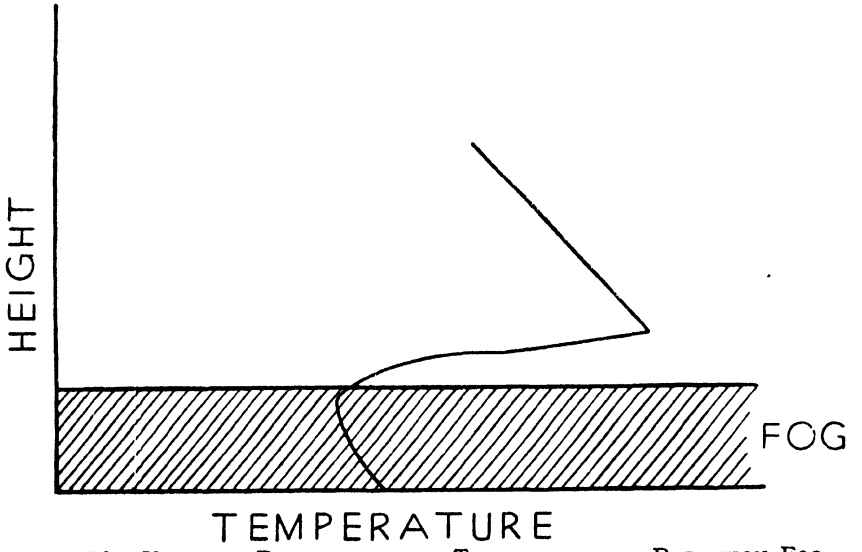


Fig 76.—Vertical Distribution of Temperature in Radiation Fog.

temperature begins within the fog layer near to the top, and just above the fog there may be almost a discontinuity. The inversion permits just sufficient turbulence to diffuse the fog throughout the lowest layers without dissipating it. It also favours the lull in the wind at night, which is essential to fog, for a moderate or strong wind would, through turbulence, bring down warm air from above the inversion and the land surface would become warmed by contact. The height of the fog layer varies, on an average, between 500 and 1,000 feet. Above the fog there is usually clear sky, with bright sunshine by day, and owing to the inversion, the temperature is distinctly higher than at the surface.

Radiation fog occurs chiefly over low-lying flat country and in valleys. It is especially frequent over marshland, where the abundant moisture maintains the dampness of the air while the vegetation provides the radiating cooled surface. On the other hand, rivers and lakes in flat country often give rise to localized clear areas, for, as compared with a land surface, water is very little cooled by radiation. Pilots of aircraft often take advantage of these clear areas to follow the courses of rivers during fog.

The principal parts of the world where land fog is prevalent are :—

(*a*) The lowlands of Europe from the British Isles and northern France to the Baltic.
(*b*) Siberia.
(*c*) The eastern part of the U.S.A.

It is especially frequent near large cities, particularly if they are situated in valleys. London and the lower Thames valley furnish a good example.

Land fogs are most frequent in autumn and winter, for in these seasons the night radiation period is of long duration and the air is dampest. It forms during the night, attaining its maximum intensity shortly after sunrise. The heating power of the sun is usually sufficient to disperse it during the day, except in mid-winter, when the insolation is feeble and of short duration, and the nocturnal radiation period correspondingly prolonged. The inversion of temperature at this period becomes intense and, owing to the layer of cloud which frequently forms, it is not easily broken down by solar heating. Consequently, winter fogs may persist for several days. In valleys the cold air accumulates by gravitational inflow, giving rise to reservoirs of cold fog-laden air which are very persistent, since once the cold heavy air has settled in the valleys it is not easily removed.

In summer the short nights restrict the cooling period to a few hours, and only shallow ground fogs are formed usually over low ground and in valleys. They are rapidly dispersed in the early morning by the sun.

There is often a pronounced increase of fog intensity shortly after sunrise. For example, in the London area, with the visibility

before dawn about 1,000 yards, a rapid decrease to 200 yards or less often takes place shortly after sunrise. The most satisfactory explanation of this phenomenon is that the sun's heating brings about a slight increase of turbulence in the foggy air mass, with consequent mixing and intensification of fog.

Land fogs occur in anticyclones and wedges of high pressure which are associated with the favourable light winds and clear skies. They may also occur in the cool air on the rear side of a depression which has passed away eastwards. Both the cloud and the wind decrease at night under these circumstances, and the moisture is usually available on the ground owing to the rain associated with the previous passage of the depression.

Radiation fog formed over land may drift over and envelop the neighbouring waters. The narrow waters, such as the Strait of Dover and the Thames estuary are especially liable to be thus affected, the drift of fog taking place from both shores. All coastal waters are subject to drifting land fog, except in the tropics where fogs of all types are infrequent.

Sea Fog.—True sea fog is formed in warm damp air which passes over a cold sea surface. The temperature of the sea surface must be lower than the dew-point of the air, for then the air becomes cooled below its dew-point and condensation results. The physical processes involved are similar to those in the formation of land fog; the main difference lies in the nature of the cooling surface. The sea surface temperature is only slightly affected by solar heating and night radiation, and varies little from day to day. The occurrence of sea fog therefore depends neither on the time of day nor on the state of the sky. A small lapse-rate is necessary, for if the temperature falls off rapidly with height, so also does the dew-point, and only for a shallow layer next to the surface is the dew-point higher than the sea temperature. The continued contact of the air with the cold sea surface, however, usually brings about an inversion of temperature with saturation in the lowest layers, and the condition that the lapse-rate shall be small is automatically satisfied. Sea fogs may form and persist with moderate or even strong winds, for, owing to its high specific heat and stirring by the wind, sea water retains its low temperature when brought into contact with the warmer air carried down from the upper layers by the increased turbulence associated with the stronger winds.

Sea fog is comparatively shallow and often the topmasts of ships may be seen protruding through it. At times its base does not rest on the sea and there are a few feet of clear air immediately above the water line, hence the necessity for placing look-outs as high up and as low down as possible when steaming in a fog.

In temperate and high latitudes sea fog occurs chiefly in spring and early summer when the sea is cold and the air passing over it is relatively warm owing to its origin in lower latitudes or its recent passage over heated adjacent land masses.

The principal parts of the world in which sea fog is prevalent are :—

(*a*) *The Polar regions* in summer.

(*b*) *The Grand Banks of Newfoundland.*—The sea temperature is low owing to the Labrador Current, and the area adjoins warmer ocean to the east and south and (in spring and summer) the heated land mass to the west. Except with northerly winds the air supply therefore is warm. Sea fog occurs on more than 50 per cent. of days in summer.

(*c*) *North-West Pacific.*—Conditions are similar to those off Newfoundland. The sea temperature is low owing to the cold Kamchatka Current, while to the south-east of Japan occur the highest sea surface temperatures in the Pacific.

(*d*) *Cold water coasts.*—The western coasts of the continents lying within the trade-wind belts, notably California, Chile, south-west Africa and Morocco, are washed by cold currents. The prevailing winds are off-shore and tend to bring about the drift of surface water seawards with upwelling of the colder deeper water near the coasts. Fog is formed along the coasts by the warm winds blowing over the cold sea surface.

(*e*) *British Isles.*—Fog is frequent in the English Channel approaches in spring and early summer and occurs when warm damp air is brought up from the vicinity of the Azores by south-westerly winds. Slight changes of wind direction modify the fog distribution. The Eastern Channel may be affected with WSW. winds, the Irish Sea by SSW. winds. In the North Sea and the sea between the Faeroes and Scandinavia fog often accompanies easterly winds blowing off the heated continent. Warm south-westerlies bring fog to the vicinity of the Faeroes.

Fogs are practically unknown in tropical seas and comparatively rare in the Mediterranean.

Sea Smoke or Warm Water Fog.—This is the name given to a type of fog which may occur when cold air blows over relatively warm sea. It occurs chiefly in polar regions over the open water appearing at the gaps in the ice pack, and it is also experienced in the Norwegian fjords. Evaporation takes place from the water surface at the rate corresponding to its own temperature. The air being at a much lower temperature is unable to take up the whole amount of water vapour, some of which therefore immediately condenses forming a thick fog. The air temperature being generally below freezing point the fog is usually composed of ice crystals. It is often transitory because the warm water surface steepens

the lapse-rate causing convection which disperses the fog vertically giving it the appearance of rising smoke. A similar phenomenon is the " steaming " of rivers and lakes in cold weather.

Warm-Front Fog.—The passage of a warm front (Chapter XVIII) is occasionally accompanied by widespread fog, especially if the temperature of the air in advance of the front is very low. The fog is due to mixing of the warm and cold air on the two sides of the front. Usually, however, the deterioration of visibility at a warm front is due to low cloud, which envelops the higher ground and occasionally falls to sea level.

Fog Due to Mixing between Two Air Currents.—Fog sometimes occurs when an easterly current of air from Europe unites with a south-westerly current from the Atlantic to form a single broad southerly current across the British Isles. In winter and spring the south-westerly current is warm and moist, the easterly cold and dry. Mixing takes place across the boundary between the two branches of the united current and if temperature and humidity conditions are suitable a belt of fog forms along the boundary. Such fog belts are liable to occur wherever air streams of widely different origins unite and flow side by side.

Fog Formed by the Drifting of Warm Moist Air over Land Previously Cooled and Over Fallen Snow.—Fog may occur in a current of warm moist air passing over land that has been previously cooled. The processes involved are the same as those in the formation of sea fog. It is usually necessary for the land surface to have been thoroughly cooled, as by a prolonged spell of cold winds or radiation frost, since its temperature is raised rapidly by the warm air blowing over it. For this reason the fog is confined to the time of onset of the warmer winds and does not persist. In the British Isles it occurs in winter with the replacement of easterly by south-westerly winds. In the U.S.A. it is associated with the flow of warm air from the Gulf of Mexico northwards after a " cold wave."

High Fog.—Occasionally condensation due to mixing may take place at an upper level while visibility at the surface is little affected. The mixing is usually the result of the juxtaposition of air currents with differing temperatures and humidities. Conditions are calm at the surface, but the two air currents exist at a height of one to two thousand feet. The warmer rises above the colder, forming an inversion of temperature. Normally, the mixing gives rise to pure water-fog, which at the high level is simply stratus cloud. Over cities and industrial areas, however, there is often dense accumulation of smoke just below the inversion. This pall of smoke, although not affecting surface visibility appreciably, may cut off practically all daylight and cause gloom or even darkness, as of night.

Smoke Fog.—Smoke alone is seldom responsible for fog, but if meteorological conditions are suitable it may transform what would otherwise be a white water fog into the obnoxious brown fog of cities. The smoke may be carried by the wind to considerable distances from the source of pollution. London smoke may, for example, seriously affect visibility in the Thames estuary when the wind is west, and its effects are noticeable as far away as Spithead and the Strait of Dover. Apart from those meteorological factors which have been considered in this chapter as contributing to the causes of fog *in situ*, the immediate past history of an air mass, particularly as regards its passage over sources of pollution by smoke, is the most important factor affecting the visibility within it.

Dust Fog.—In desert countries the squally winds accompanying the passage of a cold front (Chapter XVIII) often raise dense clouds of dust to considerable heights and visibility is seriously impeded. The finer dust particles remain in suspension in the air and may be carried by the wind to great distances. Saharan dust is often encountered in the form of a fine all-pervading mist or fog up to 100 miles from the west coast of Africa, carried out to sea by the harmattan. The brickfielder, the hot dry wind from the interior which occurs in south-east Australia, is often accompanied by dust fog.

Orographic Fog.—When a warm moist air current meets a range of mountains or an elevated coast-line, it is forced to ascend and becomes dynamically cooled, soon reaching its dew-point. The resulting condensation leads to the formation of cloud, or, if the air be nearly saturated, mist or fog at the surface. The fog so formed often extends for a short distance seawards. The conditions for its formation differ but little from those for true sea fog, but it is not necessary for the sea temperature to be below the dew-point of the air, since the cooling is dynamical and not by contact.

Visibility from Aircraft.—The distance at which an object on the earth's surface may be seen from an aircraft in flight is often very much less than from a point on the ground. The diminished visibility from the air is due to the presence of dust haze in the lower atmosphere. The haze usually has a well-defined top, which occurs immediately below an inversion of temperature. Above the haze visibility is generally good. The inversion forms a convective lid above which the haze cannot rise ; it therefore remains concentrated in the lower layers. Over land, haze is most likely to occur in the early morning in clear weather, when the temperature inversion is most pronounced owing to nocturnal cooling of the earth's surface. It occurs over sea and land under anticyclonic conditions, the inversion then being due to subsidence (Chapter VII).

If the haze layer is very thick, visibility from the air may improve with height.

In Fig. 77 let AB represent the ground, and let the haze layer be represented by the shaded belt. Let C and D be two positions of an aircraft, C just above the haze layer and D vertically above C

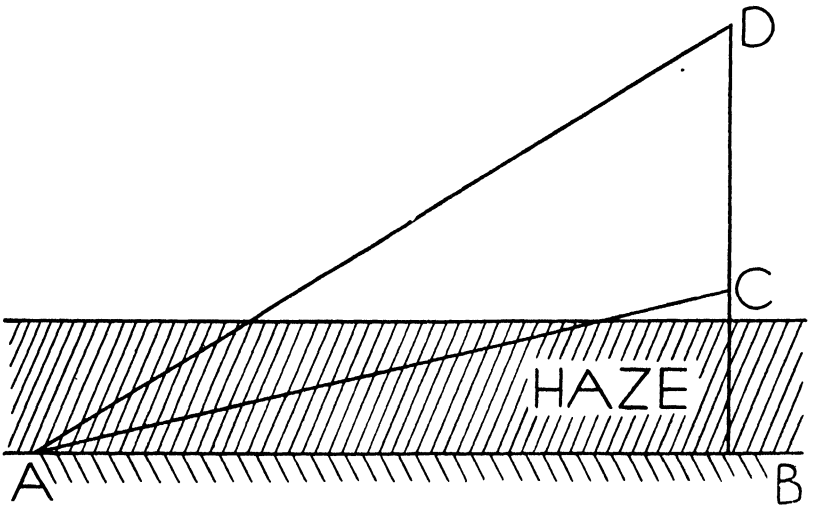


FIG. 77.—VISIBILITY FROM AIRCRAFT.

and considerably higher. In so far as the visibility of an object A on the ground is affected by the haze layer, it will be better from D than from C, owing to the greater thickness of haze along the line of sight CA than along DA.

CHAPTER X.

Clouds

Cloud consists of minute particles of water or solid ice in suspension in the air, and is formed by the condensation of water vapour consequent upon the cooling of air below its dew-point.

The process of condensation is similar to that described in connexion with the formation of fog (Chapter IX), where it is stated that the presence of hygroscopic nuclei is necessary. These nuclei are always present in sufficient quantities in the atmosphere to permit the formation of water-droplets from the vapour if the conditions of temperature and humidity are suitable. It may therefore be assumed that cloud (or fog, if at the surface) will form in a mass of air which is cooled beyond the point of saturation. The amount of water present in cloud, either in the form of droplets or ice crystals has been estimated as varying between 0·1 and 5 gm. per cu. metre. If the amount of condensed water exceeds the carrying power of the cloud, the excess is precipitated as rain or snow.

Classification of Clouds.

It is not possible in any classification to separate clouds into distinct forms; there is a gradual transition between the various types. For this reason the observer sometimes experiences difficulty in deciding to which type a particular cloud belongs.

Clouds are most easily classified according to their appearance, and on this basis, a British Meteorologist, Luke Howard, in 1803, first distinguished four main groups as follows:—

Cirrus	Cloud having a fibrous, feathery appearance.
Cumulus ..	Cloud having a heaped, lumped or rocky appearance.
Stratus	Clouds which appear to be in a level sheet.
Nimbus	Ragged low clouds of bad weather from which rain or snow usually falls.

Another simple classification is based upon the height of the cloud. There are three classes—high, medium and low cloud. Since the highest clouds are usually of the cirrus type they are denoted either by the name cirrus or the prefix cirro- in combination. Similarly, medium clouds are denoted by the prefix alto-. The three groups and the average heights at which they occur are as follows:—

High cloud, cirrus, or cirro- ..	Above 20,000 feet.
Medium cloud, alto-	6,500 to 20,000 feet.
Low cloud	Below 6,500 feet.

The heights given refer to temperate latitudes and are mean

heights, above sea level, or above the general level of land in the region. The actual heights may vary considerably from these mean values; low cloud may occur up to 8,000 feet while in the polar regions cirrus may be found almost down to the surface.

The International Cloud Classification.

In 1930 the International Commission for the Study of Clouds adopted a classification based on height and appearance. The following is a list and description of the ten different cloud types or genera of this classification; the abbreviations by which they are known are also given :—

1. **Cirrus (Ci.).** Fig. 78.—Detached clouds of delicate and fibrous appearance, without shading, generally white in colour, often of a silky appearance.

Cirrus appears in the most varied forms such as isolated tufts, lines drawn across a blue sky, branching feather-like plumes, curved lines ending in tufts, etc.; they are often arranged in bands which cross the sky like meridian lines, and which, owing to the effect of perspective, converge to a point on the horizon, or to two opposite points (cirrostratus and cirrocumulus often take part in the formation of these bands). Cirrus clouds are composed of ice crystals and are so thin that the sun's disc retains its sharp definition when seen through them; they cause very little dimming of sunlight. Owing to their great height they appear illuminated some time before sunrise and after sunset. Some forms of cirrus are popularly known as "mare's tails."

2. **Cirrocumulus (Cicu. or Cc.).** Fig. 79.—A cirriform layer or patch composed of small white flakes or of very small globular masses, without shadows, which are arranged in groups, or lines, or more often in ripples resembling those of the sand on the sea shore.

In general cirrocumulus represents a degraded state of cirrus and cirrostratus both of which may change into it. In this case the changing patches often retain some fibrous structure in places.

Real cirrocumulus is uncommon. It must not be confused with small altocumulus patches on the edges of altocumulus sheets. Cirrocumulus is popularly known as "mackerel sky."

3. **Cirrostratus (Cist. or Cs.).** Fig. 80.—A thin whitish veil, which does not blur the outlines of the sun or moon, but gives rise to halos. Sometimes it is quite diffuse and merely gives the sky a milky look; sometimes it more or less distinctly shows a fibrous structure with disordered filaments. When very diffuse it is known as "cirronebula."

4. **Altocumulus (Acu. or Ac.).** Fig. 81.—A layer, or patches composed of laminae, or rather, flattened globular masses, the smallest elements of the regularly arranged layer being fairly small and thin, with or without shading. These elements are arranged in

groups, in lines or waves, following one or two directions and are sometimes so close together that their edges join.

The thin and semi-transparent edges of the elements often show irisations (Chapter XIII) which are rather characteristic of this class of cloud. Coronae also appear in altocumulus.

5. **Altostratus (Ast. or As.).** Fig. 82.—Striated or fibrous veil, more or less grey or bluish in colour. This cloud is like thick cirrostratus but without halo phenomena; the sun or moon shows vaguely, with a faint gleam, as though through ground glass. Sometimes the sheet is thin with forms intermediate with cirrostratus (altostratus translucidus). Sometimes it is very thick and dark (altostratus opacus), sometimes even completely hiding the sun or moon. In this case differences of thickness may cause relatively light patches between very dark parts; but the surface never shows real relief, and the striated or fibrous structure is always seen in places in the body of the cloud. Precipitation may fall from altostratus. It is popularly known as "watery sky."

6. **Stratocumulus (Stcu. or Sc.).** Figs. 83-4.—A layer or patches composed of laminae or globular masses; the smallest of the regularly arranged elements are fairly large; they are soft and grey, with darker parts. These elements are arranged in groups, in lines, or in waves, aligned in one or in two directions. Very often the rolls are so close that their edges join together; when they cover the whole sky, as on the continent, especially in winter, they have a wavy appearance.

7. **Stratus (St.).** Fig. 85.—A uniform layer of cloud, resembling fog, but not resting on the ground. When this very low layer is broken up into irregular shreds it is designated **Fractostratus (Frst. or Fs.).** Fig. 86.

8. **Nimbostratus (Ns. or Nbst.).** Fig. 87.—A low, amorphous and rainy layer of a dark grey colour and nearly uniform; feebly illuminated seemingly from inside. When it gives precipitation it is in the form of continuous rain or snow.

But precipitation alone is not a sufficient criterion to distinguish the cloud which should be called nimbostratus even when no rain or snow falls from it.

There is often precipitation which does not reach the ground; in this case the base of the cloud is always diffuse and looks "wet" on account of the general trailing precipitation, "virga," so that it is not possible to determine the limit of its lower surface. With this is associated **Fractonimbus,** ragged low clouds of bad weather ("scud"), frequently seen below altostratus or nimbostratus.

9. **Cumulus (Cu.).** Fig. 88.—Thick clouds with vertical development; the upper surface is dome-shaped and exhibits protuberances while the base is nearly horizontal.

When the cloud is opposite to the sun the surfaces normal to the observer are brighter than the edges of the protuberances. When the light comes from the side, the clouds exhibit strong contrasts of light and shade ; against the sun, on the other hand, they look dark with a bright edge.

True cumulus is definitely limited above and below ; its surface often appears hard and clear cut. But one may also observe a cloud resembling ragged cumulus in which the different parts show constant change. This cloud is designated **Fractocumulus (Frcu. or Fc.)**. Fig. 89.

Cumulus is formed by a rising mass of air in which condensation is going on (thermal convection) ; a distinction should be drawn between the small fair-weather cumulus which almost always forms over land on fine summer days and the large cumulus which may give rise to local showers.

10. **Cumulonimbus (Cunb. or Cb.)**. Fig. 90.—Heavy masses of cloud, with great vertical development, whose cumuliform summits rise in the form of mountains or towers, the upper parts having a fibrous texture and often spreading out in the shape of an anvil, Fig. 91.

The base resembles nimbostratus, and one generally notices " virga." This base has often a layer of very low ragged clouds below it (fractostratus, fractocumulus).

Cumulonimbus clouds generally produce showers of rain or snow and sometimes of hail or soft hail, and often thunderstorms as well.

If the whole of the cloud cannot be seen the fall of a real shower is enough to characterize the cloud as a cumulonimbus.

The anvil-shaped tops arise as a result of the convection currents penetrating to high levels where, as a result of the low temperature, the cloud particles become frozen and are carried away by strong upper winds. On account of their appearance they are known as " false " cirrus or " anvil " cirrus.

There are three commonly occurring varieties of cloud which are not included in the above classification :—

Lenticularis. Fig. 92.—Clouds of an ovoid, lens or airship shape with clean-cut edges, especially common on days of föhn, scirocco and mistral winds (Chapter VIII). This form exists at all levels from cirrostratus to stratus, the most common being lenticular altocumulus ; it frequently shows irisation.

Mammatus. Fig. 93.—This description is given to all clouds whose lower surfaces form pockets or festoons. The form is found especially in stratocumulus and cumulonimbus either at the base, or even more often on the lower surface of anvil projections. It is also found, though rarely, in cirrus clouds, probably when they have originated in the anvil of a dispersing cumulonimbus. It is commonly called mammatocumulus.

Castellatus. Fig. 94.—Added to the name altocumulus to indicate the turretted or crenellated appearance sometimes assumed by individual cloudlets of medium height, usually in summer. This type frequently precedes thundery conditions.

Illustrations of the above types of cloud are shown on the following pages. A more comprehensive set of illustrations is published in the " International Atlas of Clouds and States of the Sky," copies of which are supplied to certain of H.M. ships.

Physical Processes of Cloud Formation.

A study of the physical processes of cloud formation is concerned chiefly with the ways in which condensation is brought about. By far the most important factor is the dynamical cooling due to vertical motion, but radiation also plays a part. It is possible broadly to classify clouds according to the process of cooling principally involved in their development, although such a classification cannot be rigid, since in most cloud formations more than one of the processes operate. The various processes are considered below :—

(1) **Mixing of two Masses of Nearly Saturated Air at Different Temperatures.**—The process of mixing has been described in Chapter IX. If it occurs at the surface, it gives rise to fog ; if at some upper level, to stratus cloud.

(2) **Radiation.**—Water vapour is an efficient absorber of the heat waves radiated by the earth's surface, and is therefore also an efficient radiator ; but dry air is not. Changes of temperature resulting from radiation therefore take place chiefly in moist layers of the atmosphere in which cloud may most readily form. The effects of gain or loss of heat by radiation in such a moist layer depend upon the lapse-rate in the environment ; if it is a stable one a gain or loss of heat leads to heating or cooling of the moist air without change of height. The latter process may be sufficient to bring about the formation of stratiform cloud. If, however, the lapse-rate is large, heating of the moist layer may lead to its ascent with dynamical cooling and formation of convection cloud. On the other hand, radiation cooling in such an environment may lead to sinking with dynamical heating. This may lead to the evaporation of cloud already present since cloud is itself an efficient radiator and absorber of heat waves. The well-known tendency for cumulus to disappear at night is attributable to this cause. The loss of heat by radiation from the upper surface of a level cloud sheet in a stable environment may lead to the formation of an inversion of temperature above the cloud, and this in its turn renders the cloud more persistent. The incoming solar radiation passes with little absorption through a clear atmosphere, whether moist or not ; but in

passing through cloud its path is lengthened many times, owing to repeated reflections from the droplets. Its heating effect inside the cloud may therefore be considerable. The ultimate result of this heating again depends on the stability or instability of the environment. Stratified cloud, associated with a stable atmosphere, will be simply warmed and will tend to evaporate; hence the frequent disappearance of stratus and stratocumulus during the forenoon in summer. On the other hand, cumulus tends to grow on being heated, consequent upon the increased tendency to convection.

Stratified cloud, because it is affected directly by radiation, tends to form at night and disappear by day. Cumulus, owing to the additional effects of vertical motion in the unstable environment, tends to increase by day and disappear by night.

(3) **Thermal Convection.**—By thermal convection is meant the ascent of a limited mass of air through its environment. The slightest rise of temperature will cause saturated air to ascend in an environment in which the lapse-rate exceeds the saturation adiabatic, with the formation of cumulus or cumulonimbus. Usually, however, air is unsaturated, and it is necessary for convection that the lapse-rate shall exceed the dry adiabatic below the condensation level. The requisite lapse-rate is most rapidly established in the lower layers over land owing to solar heating of the surface, and some cumulus is almost always formed over land on summer afternoons. The passage of air of polar origin to lower latitudes increases the lapse-rate owing to contact with the land or sea, which becomes progressively warmer, and cumulus cloud is therefore characteristic of polar air streams. In this case convection may begin either by day or night, especially over the sea, whose temperature does not fall off at night, as does that of the land surface. The extent to which the cumulus grows depends upon the lapse-rate in the upper levels; large cumulus and cumulonimbus are associated with instability extending to great heights; small cumulus with limited instability. Cumulus usually has a flat base at the condensation level; the base of cumulonimbus is often ragged, owing to the turbulent motion due to the violent convection.

The temperature in cumulus cloud is occasionally found to be somewhat lower (up to 4° F.) than at the same level in the environment. Differences up to 2° F. may exist without destroying the buoyancy of the rising air, owing to the greater humidity and therefore lesser density within the cloud. It seems, however, that the upward movement of the air through an environment more than 2° F. warmer than itself can only exist as a result of the momentum acquired in ascending through unstable layers below. For convection to great heights it is generally necessary that the cloud shall, at all stages of the process, be at a slightly higher temperature than its environment. If the rising currents of cumulus encounter an inversion or stable layer, the vertical motion is arrested and the

Cirrus (**Ci.**).—Detached clouds of delicate and fibrous appearance, without shading, generally white in colour, often of a silky appearance.

The photograph illustrates the general characteristics included in the definition, but the variety of forms is very great. These include hair-like tufts, straight filaments with curled or tufted ends, branched filaments in feathery form and many others.

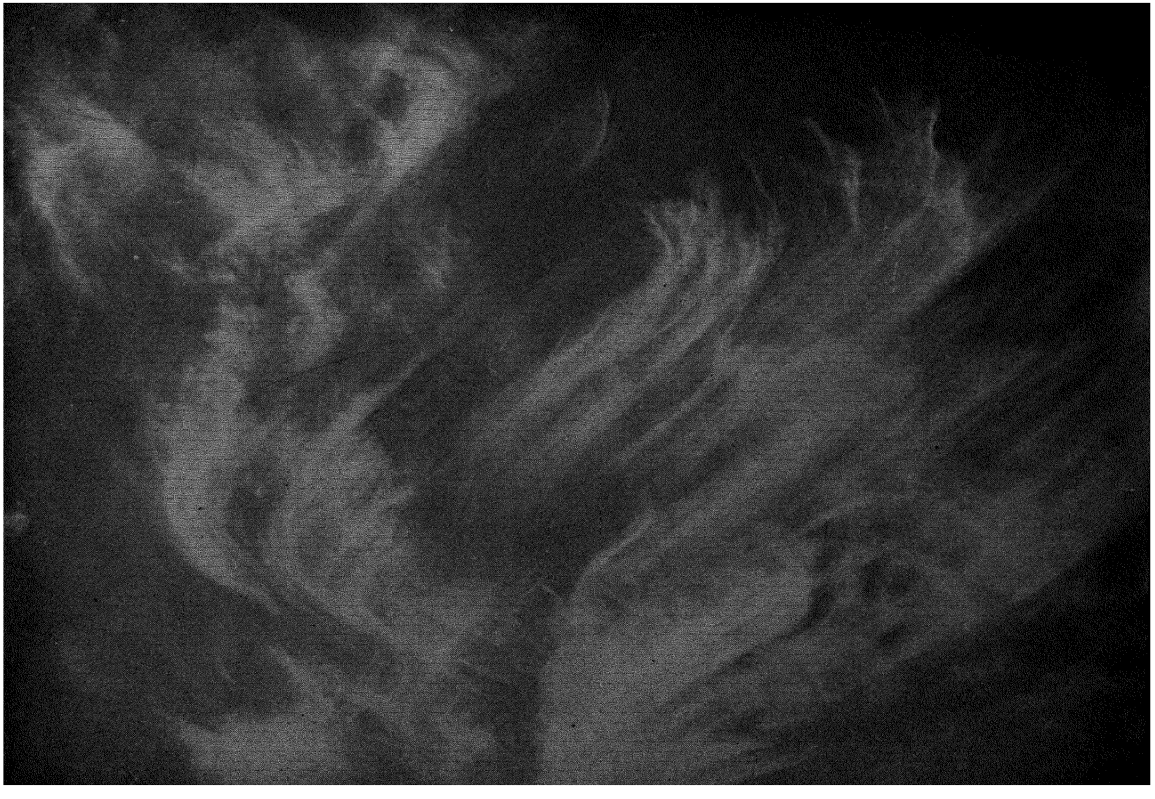


FIG. 78.—CIRRUS (Ci.).—Wisps or filaments of high cloud (20,000–40,000 ft.).

Figs. 78 to 94 to face page 232

Cirrocumulus (Cicu.).—A cirriform layer or patch composed of small white flakes or of very small globular masses, without shadows, which are arranged in groups or lines, or more often in ripples resembling those of the sand on the shore. As compared with altocumulus, the structure of cirrocumulus is exceedingly delicate and there is an obvious association with cirrus, as illustrated in this plate. In other examples, the cloudlets may be larger and more widely separated. In the cloud aggregates the lenticular structure may occur (cirrocumulus lenticularis, Fig. 92a).

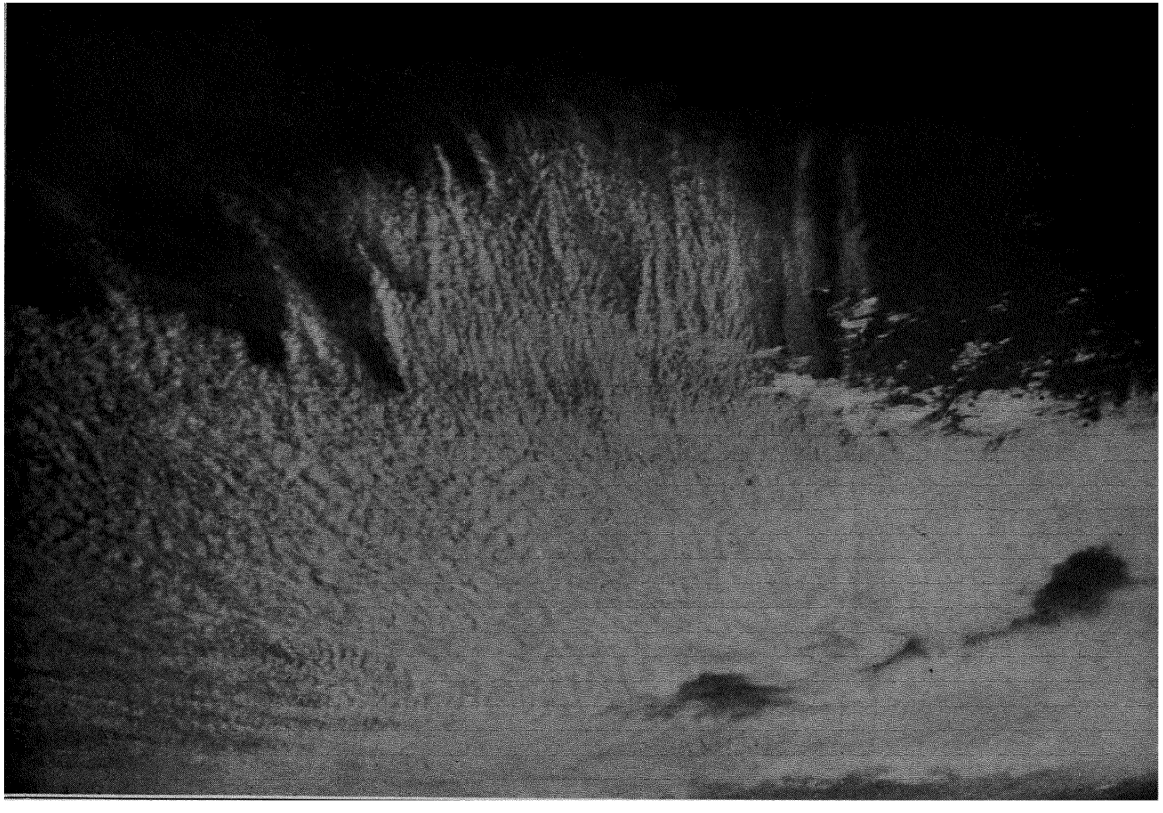


FIG. 79.—CIRROCUMULUS (Cicu.).—A high layer of small cloudlets associated with cirrus (20,000–40,000 ft.).

Cirrostratus (Cist.).—A thin whitish veil which does not blur the outlines of the sun or moon but gives rise to halos. Sometimes it is quite diffuse and merely gives the sky a milky look; sometimes it more or less distinctly shows a fibrous structure with disordered filaments.

In this illustration the cirrostratus sheet, of a rather dense type, occupies rather more than the upper half of the picture. Below, a number of small cumulus clouds are seen against the clear sky.

Cirrostratus is distinguished from altostratus by its more tenuous appearance and by the fact that halos are observable in cirrostratus alone, never in altostratus.

FIG. 80.—CIRROSTRATUS (Cist.).—Thin sheet of very high cloud 20,000–40,000 ft. (cumulus below).

Altocumulus (Acu.).—A layer, or patches composed of laminae of rather flattened globular masses, the smallest elements of the regularly arranged layer being fairly small and thin, with or without shading. The illustration shows a typical example in which the cloudlets are arranged in a fairly regular pattern with numerous clear interstices. In other cases the cloudlets may be widely separated and may show a turreted structure indicating considerable vertical development (Acu. castellatus, Fig. 94 a); or they may be arranged in bands. A special case of the latter structure is Acu. lenticularis (*see* Fig. 92 b).

FIG. 81.—ALTCCUMULUS (Acu.).—A layer of cloudlets of intermediate height (6,500–20,000 ft.).

Altostratus (Ast).—Striated or fibrous veil, more or less grey or bluish in colour. The disc of the sun or moon can usually be seen through altostratus, and coronae may be observed. Altostratus is readily distinguished from cirrostratus (the dense forms of which it sometimes resembles) by its lower apparent height and by the fact that it gives rise to no halo phenomena. Owing to the absence of definite details of structure altostratus is difficult to illustrate by a photograph, but this picture, in which there are a few small patches of lower cloud, gives a good idea of its normal appearance.

FIG. 82.—ALTOSTRATUS (Ast.).—A level sheet of intermediate height (6,500–20,000 ft.).

Stratocumulus (Stcu.). — A layer or patches composed of laminal or globular masses; the smallest of the regularly arranged elements are fairly large; they are soft and grey, with darker parts.

In this example the cloudlets are of soft and irregular form. Note the uniformity of height as revealed by the sharp lower edge of the sheet seen against the clear sky. There is no sharp distinction between the higher forms of stratocumulus and the lower forms of altocumulus, but the heavy shadows seen in this example indicate that it should be classified as stratocumulus.

FIG. 83.—STRATOCUMULUS (Stcu.).—A layer of large cloudlets at a relatively low level (1,500–4,500 ft.).

Stratocumulus (Stcu.).—A typical example of low, heavy stratocumulus. The individual cloudlets are of considerable size, are at a uniform height and appear to be of small vertical thickness.

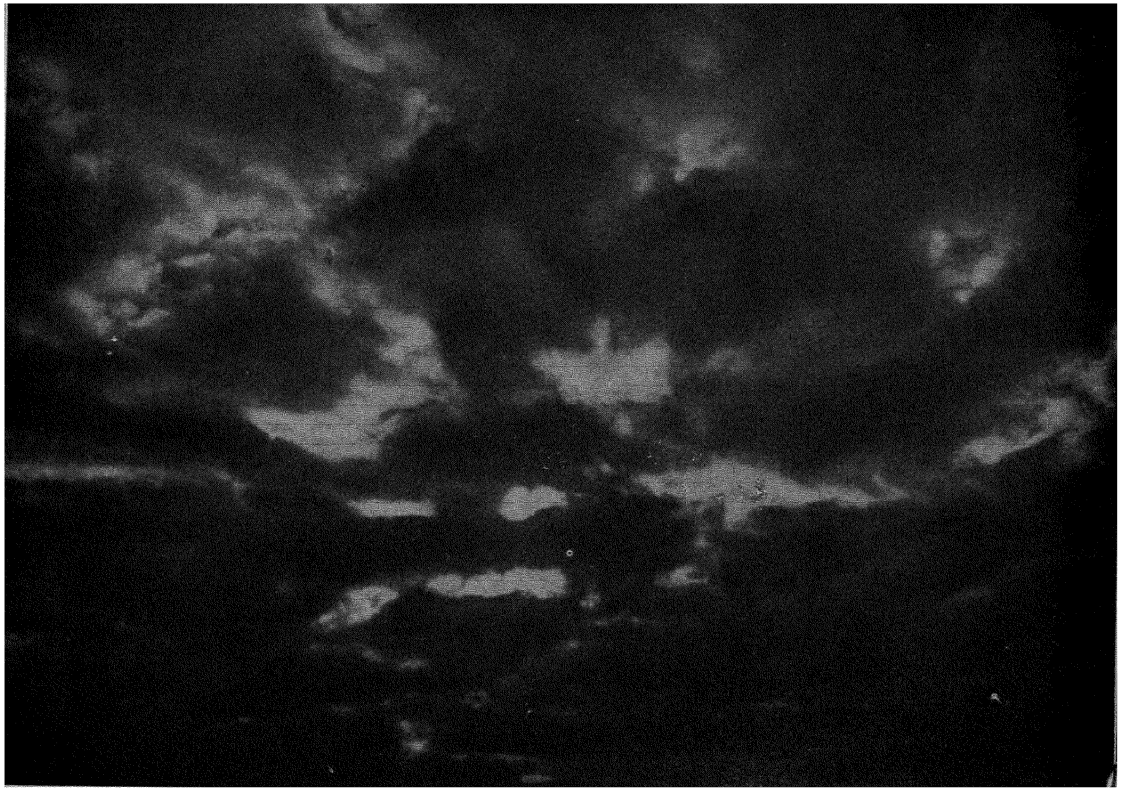


FIG. 84.—STRATOCUMULUS (Stcu.). —Lumpy clouds forming a layer at a low level.

Stratus (St.).—A uniform layer of cloud, resembling fog, but not resting on the ground.

In its most typical form, stratus shows no details of structure when observed from the ground. In this example the cloud hides the top of a hill and the line of the lower surface of the cloud can therefore be observed, passing round the hill from behind the tree on the right to the slope on the left.

FIG. 85.—STRATUS (St.).—Level sheet of low cloud : 500–2,000 ft.

Stratus (St.).—In this example the cloud is seen in the form of ragged masses, to which the name Fractostratus is applicable.

FIG. 86.—FRACTOSTRATUS (Frst.).—Ragged stratus. Drifting masses of low cloud.

Nimbostratus (Nbst.).—A low, amorphous, and rainy layer of a dark grey colour and nearly uniform. When it gives precipitation it is in the form of continuous rain or snow. The usual evolution is as follows: A layer of altostratus becomes thicker and lower until it becomes a layer of nimbostratus. Beneath the latter, as in the case here illustrated, there is usually a development of ragged, very low cloud (fracto-cumulus or fractostracto) which may eventually fuse together and obscure the nimbostratus.

FIG. 87.—NIMBOSTRATUS (Nbst.).—Low, rainy layer of dark grey cloud nearly uniform: 500–2,000 ft.

Cumulus (**Cu.**).—Thick clouds with vertical development ; the upper surface is dome-shaped and exhibits protuberances, while the base is nearly horizontal. When the light comes from the side, the clouds exhibit strong contrasts of light and shade ; against the sun, on the other hand, they look dark with a bright edge, the " silver lining."

FIG. 88.—CUMULUS (Cu.).—Detached cloud with flat base and rounded top. Height of base usually 2,000–5,000 ft.

Cumulus (Cu.).—True cumulus is well defined with hard and clear-cut outlines ; but one may also observe as in this photograph clouds belonging to the same system in which the outline is ragged and changing rapidly. To such clouds the name fracto-cumulus (frcu) is applied.

FIG. 89.—Fair weather cumulus (Cu.) and some fractocumulus (Frcu.).

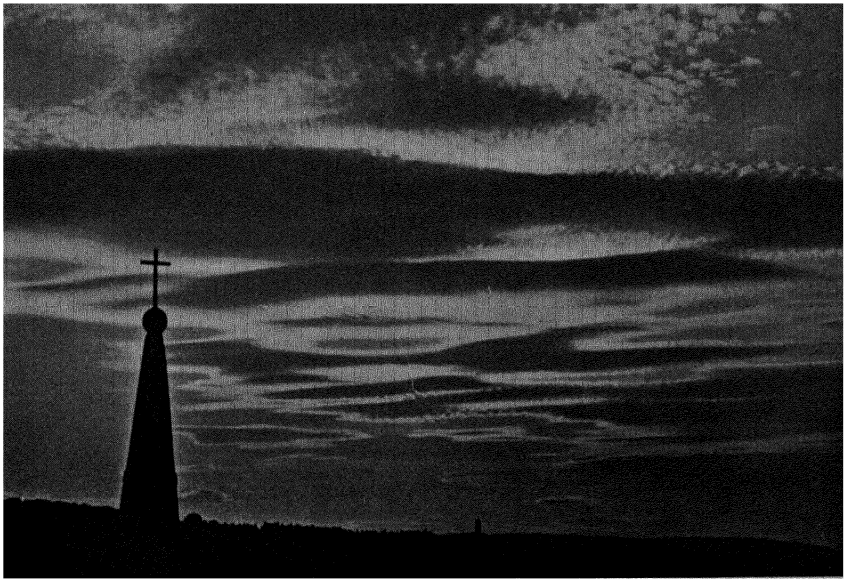
Cumulonimbus (Cunb.).—Heavy masses of cloud, with great vertical development, whose cumuliform summits rise in the form of mountains or towers, the upper parts having a fibrous texture and often spreading out in the shape of an anvil. This example shows in the lower left-hand corner a *virga* or trail of precipitation, but the upper parts are quite hard and clear cut.

FIG. 90.—CUMULONIMBUS (Cunb.).—" Thunderstorm cloud " without " anvil." Height of base usually 2,000–5,000 ft.

Cumulonimbus (Cunb.).—This photograph of a mass of cumulonimbus advancing towards the spectator, shows a remarkably perfect example of the "anvil" formation of the upper parts referred to in the definition. These upper parts are of cirriform structure and were formerly distinguished by the name "false cirrus." The modern name, as given in the International Atlas of Clouds (1932), is cirrus nothus.

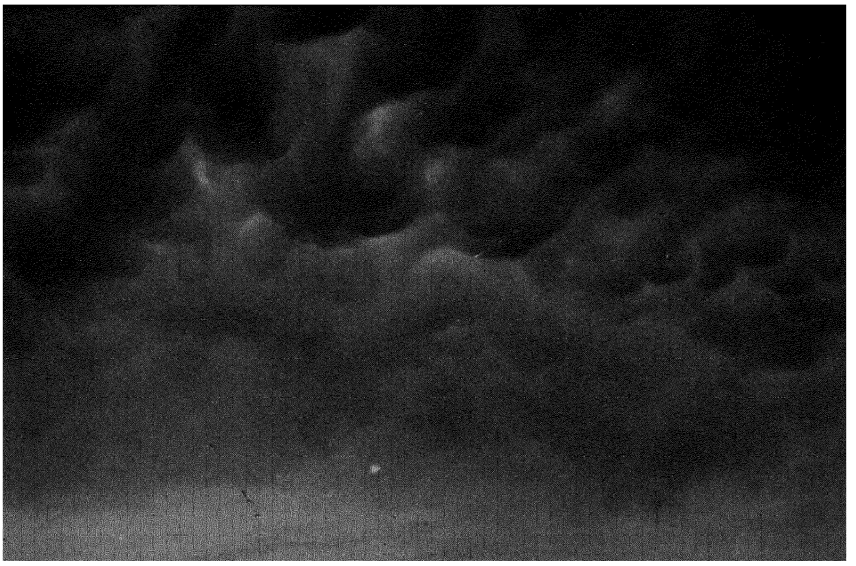
FIG. 91.—CUMULONIMBUS (Cunb.) with large "anvil" of "false cirrus" (*cirrus nothus*).

(a) **Cirrocumulus lenticularis.**



(b) **Altocumulus lenticularis.**—Two examples of cloud aggregates exhibiting the structure connoted by the adjective "lenticular" ("lenticularis"). The individual cloudlets are fused together into groups the general shape of which resembles the cross-section of a lens.

FIG. 92.



(a) **Mammatocumulus (Cu. mammatus).**—A typical example of "festoon-cloud," a low cloud whose lower surface shows numerous udder-like structures, suspended from the main sheet.

(b) **"Mammatus" structure in a cloud of intermediate height.**—Though reaching its most typical development in the "festoon-cloud," the mammatus structure is sometimes observed in clouds at higher levels, *e.g.*, on the underside of the "false cirrus" projecting from cumulonimbus. In this illustration a rather ragged mammatus structure is seen in a cloud of intermediate height.

FIG. 93.

(a) **Altocumulus castellatus** (turret cloud).—This special form of altocumulus is relatively infrequent. The individual cloudlets resemble miniature cumulus, are usually widely separated and appear brilliantly white against a deep blue sky. When seen low down on the horizon the appearance is that of many heads of small cumulus rising from the main sheet.

(b) **Stratus cumuliformis.**—An example of cumulus-like structure occurring in stratus cloud. The adjective "cumuliformis" is of general application and is added to the name of the cloud when a tendency to assume the shape of cumulus is observed.

FIG. 94.

cloud tops may spread out horizontally, forming stratocumulus, altocumulus or anvil cirrus.

(4) **Vertical Motion of Air over an Area of Great Horizontal extent, i.e., Frontal Cloud.**—Cloud formed as a result of vertical motion over a wide area often occurs with continuous precipitation, usually in association with fronts and occlusions. The process is described in Chapter XVIII. In the rain areas associated with clouds of this type, the atmosphere below the frontal cloud-sheet becomes saturated or nearly saturated up to high levels due to evaporation from the falling rain; the damp air favours the formation of clouds, particularly stratocumulus or altocumulus, often in a series of layers. Further development is often through the sequence—altostratus, nimbostratus, rain.

(5) **Orographic Effects.**—Cloud may be formed by the forced ascent of damp air over hills or elevated coasts. Such cloud usually takes the form of stratus with a horizontal base, enveloping the upper slopes. If the vertical distribution of temperature is unstable, the forced ascent may cause sufficient trigger action for the formation of cumulonimbus with showers and possibly thunderstorms.

(6) **Local Ascent of a Damp-Air Stratum.**—Lenticular and other clouds of smooth outline may be formed at an upward bend in the stream-lines of a horizontal current. If such a current flows over a range of hills an upward wave is imparted to the stream extending far above the summit. If at any level the air is almost saturated the upward movement may bring about condensation, resulting in the formation of a stationary cloud. The actual particles composing the cloud are constantly changing, being carried along in the general current and rendered visible by condensation only when passing through the upward bend in the stream-lines. Close observation shows that small cloudlets, forming outside the upstream edge are carried into the main body of the cloud, and similarly may become detached from the downstream edge and be carried a short distance away from it before evaporating.

(7) **Turbulence.**—The upward transport of water vapour by turbulence in a stable atmosphere often gives rise to a layer of stratus or stratocumulus, which may be either continuous or broken up; in consequence such clouds are often called eddy cloud sheets. It has been seen in Chapter VIII that the tendency of turbulence is to equalize the distribution of certain atmospheric properties. Water-vapour content is such a property and since normally the amount of water vapour in the atmosphere decreases with height, the effect of turbulence is to transport water vapour upwards. The air, which at the same time is carried upwards, therefore becomes cooler and damper, and sooner or later reaches its dew-point with the formation of cloud. Since the upward movement is uniform over a wide area, the cloud takes the form of a continuous

sheet. There is often an inversion of temperature above such a cloud sheet. The condition for the formation of sheet cloud, as opposed to detached cumulus, is a stable vertical temperature distribution; the upward movement is then eddy convection and not thermal convection. Stratocumulus forms readily over the sea where an adequate moisture supply is available and the lapse-rate often small. When it forms over land it is usually due to the drifting in of air from the sea. The passage of air across the North Sea often causes an extensive sheet of stratocumulus to spread over east and south-east England, even though the air is free from cloud when it leaves the continent. Moderately powerful sunshine causes such a cloud sheet to break up unless it is continually renewed by the turbulent transport of water vapour upwards. Stratocumulus, therefore, tends to dissipate at some distance inland in the daytime, except in winter, when solar heating is weak.

Eddy convection in the upper levels may give rise to layers of altocumulus, cirrocumulus and some forms of cirrus; sometimes a layer of cloud is formed in the turbulent region between two horizontal air currents moving with different speeds at adjacent levels. Waves developing at the surface of discontinuity are probably associated with the rippled and cellular patterns observed in some forms of cirrocumulus and altocumulus. Recent work by Walker and Mal has shown, however, that the cellular structure is to be attributed to the development of thermal instability within a cloud sheet, the upper surface of which is cooled by radiation to the clear sky and the lower surface warmed by radiation from the earth's surface. Experiments with a shallow layer of volatile liquid whose upper surface is cooled by evaporation have shown that the layer breaks up into a number of separate cells, roughly hexagonal in shape. The motion is upwards at the centre of the cells and downwards at their edges. In the atmosphere the upward motion would, by dynamical cooling and condensation, give rise to cloudlets having the same shape as the cells. The variety of shapes actually observed is attributed to a variation of wind, or shear, between the upper and lower surfaces of the cloud sheet.

Stratus often originates as surface fog formed during the night and raised by turbulence consequent upon the heating which sets in after sunrise.

EXTRACTS FROM THE INTERNATIONAL ATLAS OF CLOUDS AND OF STATES OF THE SKY.

The following notes on the observation of clouds from the air, states of the sky and tropical clouds, pages 234 - 259, *are taken from the International Atlas of Clouds and of States of the Sky.*

The Observation of Clouds from the Air.

The classification of clouds is based on their appearance from the ground, while the aerial observer sees them also from above.

It is very desirable that airmen employed on meteorological work should make a systematic study of clouds from the ground as well as from the air, in order that they may use the names in the orthodox manner. Nevertheless, even the best observers will find it necessary to supplement in plain language their descriptions of clouds seen from the air.

One may distinguish :—

(*a*) High clouds, of genera cirrus or cirrostratus.
(*b*) Clouds of genus altostratus, of slight density.
(*c*) Clouds in the form of a horizontal layer or sheet, very numerous groups, including forms which the classification from the ground arranges in different genera—stratus, stratocumulus, altocumulus—but whose structure observed from the air is practically similar.
(*d*) Haze.
(*e*) Clouds of convection currents (cumulus or cumulonimbus).

(*a*) **High Clouds.**—High clouds are not as a rule reached by aeroplane observers, and are thus only observed from below. Sometimes, however, when the low clouds cover the sky, the aerial observer can obtain valuable information as to the upper clouds.

Cirrus and cirrostratus clouds composed of snowflakes or ice crystals are occasionally encountered at quite moderate heights. One may also encounter very finely scattered ice crystals which do not form a cloud visible from the ground and are seen sparkling in the sun.

(*b*) **Clouds of the Genus Altostratus.**—The cloud known as altostratus (or nimbostratus) has a great vertical extension but no great opacity. At temperatures below freezing point it normally consists of ice crystals or small snowflakes. Falling snow often makes the base of the cloud indistinct and variable. During precipitation the cloud may extend quite low and merge into turbulent fractostratus clouds. The sun is usually visible from quite a moderate height in the cloud, but the clouds usually extend to a great height, the upper part being thin and of cirrostratus form. Occasionally altostratus is bounded above by a layer of altocumulus, or by cloud masses resembling the tops of cumulonimbus.

Sometimes there are thin layers or patches of clouds of altocumulus type about the lower part of the altostratus. A few examples of altostratus consist entirely of a succession of such layers. In such cases the cloud is not quite uniform when viewed from below, and if the sun is visible, cloud of varying texture can be seen driven across it. The aerial observer should describe such complicated cloud structure in plain language.

(*c*) **Clouds in a Layer or Horizontal Sheet.**—These clouds exist at very different levels, but when viewed from above they have a very similar structure, the upper surface having the characteristic rippled aspect.

In the rare cases where the air is unstable above the cloud layer, the clouds grow up irregularly and take on the character "cumuliformis"; when they are well developed they resemble an extensive group of cumulus as seen from above. Clouds of a structure essentially similar to altocumulus cumuliformis sometimes form at low altitudes and may eventually develop into cumulonimbus.

For an observer within the cloud, the cloud layer simply resembles a layer of fog. However, if it is thin, one may see the ground through it when one looks directly downwards. In this case the clouds appear grey when seen from above, except on the side where they reflect the light of the sun. When the layer becomes denser it reflects more light and appears white when viewed from above.

Horizontal cloud sheets may occur at any height, but they are most frequent in the first 10,000 feet Seen from the ground, clouds in the form of a horizontal layer, although their structure is similar, may be classified according to their altitude in different genera: stratus, stratocumulus, altocumulus, cirrocumulus. All the clouds of this type produce glories, coronae and irisation when the light is favourable.

Stratus seen from the ground shows no definite structure, because the observer is too near the cloud layer, but if the sun is visible through it and the clouds are drifting, it is easy to see that the layer is not really homogeneous. It is only very thick stratus (1,500 feet or thereabouts) which shows no structure when seen from below. From above, a structure resembling stratocumulus can always be observed. Even fogs may have ripples, and if one observes them from a great height they very much resemble altocumulus viewed from below.

At considerable heights, cloud sheets are usually very thin, but very thin layers are by no means rare, even in the lower layers of the atmosphere. Moreover, it sometimes happens that a thin layer develops rapidly over a vast area and becomes thicker with large ripples. One sometimes encounters also a damp layer of great vertical extent, containing a complicated series of thin cloud layers, often of limited area, one above the other.

A thin cloud layer appears—for an observer within it—like a fog a few feet thick, resembling a meadow mist in the evening. When one is above the cloud the structure again becomes visible. Seen from above, the thin cloud layers appear sometimes quite smooth, while viewed from below they show a delicate but definite structure if there is a blue sky behind them. On a background of cirrostratus or of altostratus they are invisible; when they are situated between the observer and the background of cumulus they appear grey against the white cumulus background. The thin cloud is usually above the level of the base of the cumulus.

(*d*) **Haze.**—There is no sharp division between thin clouds and thick haze. Haze is normally present in the lower layers, and has

often a sharply-defined top below an inversion. The light is reflected from the upper surface of the haze when one looks downwards on the side of the sun. In consequence it is often impossible to see anything on the ground, except perhaps the sunlight reflected from a water surface, although when one looks away from the sun, visibility towards the ground is moderate.

Haze of less thickness is often observed in damp layers higher up, forming a dark line round the horizon, and if one flies at this level, it is almost impossible to distinguish the haze from clouds which may exist in the layer, if they are very thin.

Moreover, there is no sharply defined limit between haze and clouds from the point of view of physical structure ; in fact, hygroscopic particles cause a certain amount of condensation before the air is actually saturated, so that there is nothing in the nature of a discontinuity associated with 100 per cent. relative humidity, but only a gradual transformation from haze to cloud.

(*e*) **Convection Clouds.**—Cumulus clouds present no difficulty to aerial observers, but it must be remembered that owing to the effect of perspective, cumulus clouds near the horizon may appear as a continuous layer with irregular tops.

The tops of large cumulonimbus whose base is hidden, can sometimes be seen from a distance of over 120 miles.

Large cumulus and cumulonimbus, whose base may be as high as 13,000 feet, sometimes develop and extend to the normal cirrus level. Properly speaking, these clouds should be called altocumulus cumuliformis, owing to the height of their base. In the case of true cumulus or cumulonimbus, the convectional movements should begin near the ground, whereas in the case of altocumulus cumuliformis, they only commence in the layer of altocumulus.

When cumulus clouds reach a stable layer they often form a cloud sheet with an irregular hummocky upper surface similar to the formation described in paragraph (*c*) above.

It frequently happens that the tops of the cumulus break through a cloud sheet. This development must be distinguished from that in which the cloud sheet itself erupts (type cumuliformis). Even if there is a large inversion, cumulus clouds often protrude some hundreds of feet from a cloud sheet. When numerous cumulus clouds break through in this way, the upper surface of the clouds assume an irregular hummocky form which has many varieties and must be described in plain language, as it corresponds to nothing observed from the ground. Occasionally, large cumulus clouds break through a cloud sheet and eventually develop into cumulonimbus.

STATES OF THE SKY

Classification of States of the Sky.

This classification is based not only upon knowledge of the physical processes involved in the formation of clouds but also upon the synoptic study of disturbances (Chapter XVIII).

The enumeration of the kinds of cloud which cover the sky at a given moment is not a sufficient characterization of the "state of the sky." What really characterizes it, is the aggregate of the individual clouds and their organization.

A distinction will be drawn between the "upper and middle sky" and the "lower sky." (The divisions of the space which may be occupied by clouds, or rather the cloud complexes contained in them, will be referred to repeatedly in the following pages, so that it is necessary to adopt abbreviated terms.)

The state of the upper and middle sky is determined by thermodynamical processes associated with big atmospheric disturbances.

It is most often a question of adiabatic expansion which accompanies a general fall of pressure. Turbulent processes also are always to be found to a greater or lesser extent and many aspects peculiar to the middle sky are due to turbulence. Even convection may sometimes have effect up to these altitudes. But in general these states of the sky depend almost entirely upon the synoptic situation; they correspond to definite sectors of big atmospheric disturbances and are like them, migrant or durable. They appear most clearly in middle latitudes, for in low latitudes, apart from tropical cyclones, variations of pressure are often feeble and local while in northern regions the upper and middle sky is often masked by a very overcast lower sky.

The state of the lower sky, on the contrary, is determined by the meteorological conditions existing in the low layer which is directly influenced by the surface of the earth. Cloud formations in that part are essentially caused by the processes of convection or turbulence and sometimes by cooling by radiation. These states of the sky are therefore understood to depend largely upon the nature of the surface of the earth whether sea or land, upon the relief of the land, its heating, etc. They are particularly abundant and heavy in maritime and northern climates.

Upper and Middle Skies.

These states of the sky are one of the principal manifestations of the life of large normal atmospheric disturbances, and characterize the various zones:—Front, central, rear and lateral.

In some of them it is advisable to distinguish between the case of a feeble disturbance, and the case of a typical well-developed disturbance.

Further, it is expedient to treat separately the big thundery disturbances of middle latitudes which exhibit characteristics peculiar to themselves and a degree of complexity unknown in ordinary disturbances. In such cases no useful purpose is served by distinguishing more than the front zones on the one hand and the central and rear zones on the other.

There are the following states of the sky :—

1.—Emissary sky.
2.—Front zone sky, typical.
3.—Front zone sky, weak.
4.—Lateral sky.
5.—Central sky, typical.
6.—Central sky, weak.
7.—Rear sky.
8.—Pre-thunderstorm sky.
9.—Thundery sky.

Lower Skies.

Simple Skies. *"Convection" Skies.*—The extent and intensity of convection depend upon the vertical instability of the atmosphere which results from the heating of the surface layer of a mass of air which is cold above. It also varies with the water content of the upper layers. Three states of the sky are to be distinguished according to the intensity of this convection :—

1. Cumuliform sky of fine weather.
2. Disturbed cumuliform sky (without cumulonimbus).
3. Disturbed cumuliform sky (with cumulonimbus).

"Turbulence" Skies.—When air is slightly disturbed over a large surface on a small scale, horizontal layers of cloud are produced. The following distinctions are drawn according to whether the area of disturbance is clearly bounded above by a much more stable layer, or whether it has no distinctly defined limits :—

4. Stratiform sky (stratus or stratocumulus).
5. Amorphous sky (fractostratus or fractocumulus of bad weather).

Mixed Skies.—Some combinations of two of these skies are evidently impossible. For example, the amorphous sky cannot exist at the same time as the cumuliform of fine weather because there is then not enough humidity ; the association of a stratiform sky with that of cumulonimbus is rare, the marked inversion of the former being hardly compatible with the powerful convection of the latter.

The following combinations are conceivable :—

6. Cumuliform sky of fine weather (1) and stratiform sky (4).
7. Disturbed cumuliform sky (2 or 3) and stratiform sky (4).
8. Disturbed cumuliform sky (2 or 3) and amorphous sky (5).

Combinations of Lower Skies and Upper and Middle Skies.

Many combinations of lower skies with upper and middle skies are possible, but not all, however. Thus, for example, beneath the middle layers there very often exists a marked subsidence of the air which tends to weaken convection so that this state of the sky is incompatible with a disturbed cumuliform or cumulonimbus sky.

Definitions and Descriptions of the States of Sky. Upper and Middle Skies.

Emissary Sky.—Sky of cirrus. Either isolated or forming little groups at some distance from each other.

The types of sky which may be directly replaced by an emissary sky are :—the cumuliform of fine weather, the stratiform, the rear zone (which it follows fairly often when there is over-lapping of two disturbances), or the thundery sky.

These clouds are the first indications of the presence, at a distance, of a disturbance from which the cirrus clouds emanate.

When the observer is on the extreme edge of the disturbance, with only these emissary cirrus in view, he cannot know whether he is in front or at the side of the depression.

This state of the sky is distinguishable from the beginning of the front zone sky by the fact that the cirrus have no tendency to coalesce.

Compatible lower skies.—All lower skies may be accompanied by emissary cirrus above.

Typical Front Zone Sky.—Sky of cirrus in fine filaments, regularly arranged, passing progressively into a transparent veil of cirrostratus or altostratus.

The upper and middle skies which may be directly followed by a typical front zone sky are :—The emissary sky, the rear zone sky (superimposition possible), and the thundery sky.

The typical front zone sky begins with cirrus sometimes accompanied by cirrocumulus. The cirrus is generally of the fibrous type in long transparent bands, never of the foamy or irregular type which is encountered in thundery disturbances.

The most rapidly moving cirrus is to be found in this state of the sky. In Europe the direction of the current in which it travels is generally between WSW. and NW. The cirrocumulus is generally in the classical form (little white globular masses rather transparent and close together).

The cirrus is soon replaced by an immense cirriform sheet which is very transparent at first. The most beautiful solar phenomena are to be observed with this type of cirrostratus ; the complete halo of 22° is an almost regular occurrence.

At the time when the first clouds of the front zone sky appear, it not infrequently happens that the last cloud elements of the preceding disturbance have not yet disappeared ; with the result that the overlapping of a rear zone or thunder sky with that of a front zone may be observed.

Compatible lower skies :—various convection and stratiform types.

Under the cirrus there is often at first a cumuliform type below (only over the land and provided the time of day is favourable). But as the veil of cirrostratus develops, these cumulus clouds " subside," flatten out and generally end by disappearing.

If a stratiform sky has prevailed before the arrival of the front zone, the low clouds can generally be seen to break into fragments and disappear, but there are exceptions to this rule. It may happen that the stratiform sky hides the whole evolution of the front zone sky.

Weak Front Zone Sky.—Sky composed of medium clouds (more rarely high), regularly arranged, often in parallel bands. Banks of altocumulus not very thick and with clear outlines, uniting gradually into a continuous layer.

The states of the sky which may be directly followed by a weak front zone sky are :—The emissary sky, the rear zone sky (superimposition possible), and the thundery sky.

The weak front zone sky begins with altocumulus which is as likely as not to be accompanied by cirrus. (There might sometimes be a tendency to confuse the altocumulus with cirrocumulus).

There is not much to be said about the cirrus. It is generally of the type described above under the typical front zone heading, but it is generally less abundant and less rapid.

The altocumulus clouds belong to two different varieties, which often exist together. The one kind covers the sky with an immense patchwork ; the rectilinear fissures which divide them call to mind the ornamentation of a crocodile's skin ; their matter is often diaphanous, and the blue of the sky which is seen through their mass blends with them at the edges in an almost imperceptible fashion. On the edge of these sheets or of the fragments which become detached, the cloudlets are much smaller and, in places, very similar to cirrocumulus. The other kind, on the contrary, congregate and form large, elongated bands, the front edge of which crosses the sky, clear-cut and rectilinear in shape ; their matter is white and opaque, while the cloudlets constituting the bands are in the shape of rolls or ovoid stones. To this variety belong usually the clouds which, on some summer evenings, hide the sunset with a horizontal violet barrier, and which are sometimes called in error "stratus of the horizon." The characteristic optical phenomenon occasioned by these clouds is that of the solar or lunar corona.

Compatible lower skies :—the lower types which may exist below a weak front zone sky are the same as in the case of a typical front zone sky, that is to say, convection or stratiform skies, and there is the same tendency for low clouds to disappear.

Lateral Sky.—Little, isolated banks of high and medium cloud, often lenticular, irregularly arranged and in perpetual transformation. This description of a lateral sky applies only to the lateral zone on the warm side of a disturbance (southern side in the northern hemisphere). On the cold side, the lateral part in a typical case presents exactly the same aspects as the front zone (cirrus, then complete veil of cirrostratus or altocumulus). Fine weather, however, may follow immediately, instead of the rainy weather which follows in the front zone.

The states of the sky which may be directly followed by the lateral types are :—cumuliform sky of fine weather, stratiform sky and emissary sky.

The lateral sky almost always fcllows the isolated cirrus, the distant forerunners of the disturbance, on the flanks as well as in the front.

After these cirrus clouds which, in some instances, however, may be wanting, the altocumulus appears ; these medium clouds should not be confused with the " shreds " of the rear zone sky, nor with the regularly arranged masses in the front zone sky. Here they appear in isolated banks of altocumulus, arranged irregularly. Lenticular altocumlus, of pearly whiteness, is of frequent occurrence in this type. In the general way, these banks of altocumulus are remarkably unstable in form. They are incessantly dispersing or developing afresh ; if any one looks at the sky after having ceased to do so for a few minutes he will find it completely changed.

The wind generally remains light, since the place of observation is far removed from the centre of the disturbance.

The sky is never completely overcast ; the amount of alto-cumulus reaches a maximum, then diminishes again, and the fine weather sky reappears without any more serious effect of the disturbance having been experienced.

Compatible lower skies :—these are of the convection (excepting disturbed cumuliform with cumulonimbus) and stratiform types.

Beneath the lateral sky a cumulus type can persist without subsiding much. Nevertheless, large cumulus are rarely found there (except in mountainous regions). A stratiform sky is frequently associated with a lateral zone sky especially in maritime climates.

Typical Central Sky.—Low ceiling, opaque veil of altostratus or of nimbostratus becoming covered by fractostratus or fractocumulus and perhaps giving continuous rain.

The only upper and middle state of the sky which can precede a typical central zone sky is the typical front zone sky (sometimes hidden by a stratiform sky).

The veil of cirrostratus in the front zone sky is soon followed by a thicker veil ; the halo disappears and the sun only remains visible as a silvery spot of ill-defined contour and fading brightness. This is the typical altostratus appearance. Altocumulus and cirro-cumulus, or rather clouds looking like the latter, but being perhaps only altocumulus, are still to be seen, but instead of appearing white against the blue of the sky, they stand out dark grey against the pearly grey of the altostratus veil. Often they are masses of little round spots ; more rarely they extend in close parallel lines reminiscent of the undulations which the sea forms on the sand of the sea-shore.

But the sun soon disappears and the ceiling of altostratus darkens

as it becomes gradually lower and passes into nimbostratus. Rain, generally of the continuous kind begins, and lasts for several hours. In some conditions the rainfall may be prolonged and last the whole day. Rainfall accompanying altostratus or nimbostratus is generally more intense than the light rain from altocumulus in a weak central sky type but it is not as abundant as in the downpours from the typical rear zone sky which follows. The atmosphere is very humid, another point of difference between central rain and rear zone showers. Visibility is rather bad. In general the sheet of altostratus or nimbostratus terminates shortly after the rain stops, its rear edge being generally fairly clear-cut, rectilinear in shape and of fibrous structure. In fewer cases it may be possible to witness a progressive thinning of the altostratus or nimbostratus for some hours after the rain.

Compatible lower skies :—The amorphous sky is necessarily associated with the typical central zone sky.

The complete passage of an altostratus system is always accompanied by low clouds. If the lower part of the sky is clear of clouds at the time of the arrival of the altostratus system, low clouds are seen to appear in the form of fractostratus, fractocumulus or stratocumulus ; but in the last case also the amorphous condition (fractostratus) dominates in the end. Even if there are still some cumuliform clouds when the central zone conditions set in, they are gradually transformed into fractostratus. The latter sometimes form eventually a continuous layer which entirely hides the altostratus or nimbostratus. When the rain has nearly ceased falling, however, the fractostratus is often transformed into fractocumulus.

Weak Central Sky.—Low ceiling. Opaque veil of altostratus formed from altocumulus which have coalesced, sometimes but not always becoming covered by fractostratus or by fractocumulus and being capable of giving more or less continuous though slight rainfall. Misty sky. Wind generally rather light.

The only upper and middle state of the sky which can be directly followed by a weak central zone sky is the weak front zone type.

The banks of altocumulus in the weak front zone sky are followed by a sheet of altocumulus. On the front edges altocumulus without shadows are clearly seen. But almost immediately shadows begin to form on the altocumulus and they develop into a uniform sheet, true altostratus, the altocumulus structure being almost indiscernible ; mist intervenes, an essential factor in the weak central zone sky and one which contributes towards the uniform appearance of the altostratus in such a fashion that if the edges had not been observed it would be impossible to distinguish it from an altostratus of cirriform structure. Visibility deteriorates rapidly, sometimes to only a few hundred yards.

Fractostratus intermingles with the mist in small shreds of little density. These clouds, which in a typical central zone sky are almost always associated with rain, often pass without giving rain

in this type. When rain does fall, it is fine and resembles at times a heavy drizzle. Altostratus of this type does not exhibit the same continuity as in typical central zone cases. Gaps or thin places may appear. Sometimes even a rift may divide it into two sheets; the mist then becomes momentarily invisible; only the whitish colour of the sky reveals it.

The rear edge of the sheet is formed of altocumulus without shadow, shiny white and very opaque. It stands out clearly against the sky and is often broken by well-marked fissures through which the blue sky is seen.

Compatible lower skies :—amorphous sky.

The lower sky is analogous to the corresponding one in a typical central sky. The quantity of low clouds is always less (in exceptional cases there is none at all) and the fractostratus is seldom sufficient to hide the sky above completely.

Rear Zone Sky.—Unstable weather, characterized by rapid alternations of pronounced improvements with exceptional visibility and menacing, disturbed skies which may be accompanied by showers and squalls. The upper and middle parts of the sky contain remains of dense and relatively low, medium or high clouds.

In the case of the rear zone sky the lower sky is inseparable from the upper and middle sky. It is characterized by more or less intense convection. If the lower sky contains cumulonimbus (heavy showers, squalls) the disturbance is a typical one. If there are only disturbed cumuliform clouds (showers and feeble gusts of wind) it is weak.

A description of a typical rear zone sky will be followed by that of a weak one but in both, the lower part of the sky plays a large part.

First it should be noted that the rear zone sky generally follows the central zone type. In some cases, unstable conditions which give rise to rear zone clouds may develop after fine weather without the previous appearance of the central zone type. On the other hand, in the rear of a central zone sky the rear zone sky may be lacking and be replaced by a stratiform type.

Typical rear zone sky.—It has just been shown that the dominating characteristic of a central zone sky is its uniformity. It is just the opposite in the rear zone.

It is not an uncommon occurrence for rain in a depression to be falling simultaneously over an area equal to half of France. In the rear zone, however, the most violent shower almost always leaves one corner of the horizon illuminated by the sun which is soon to reappear. Diversity obtains; an observer flying at a great height would see beneath him, following upon the impenetrable mantle of the central sheet, a heterogeneous collection of cumulus, fracto-cumulus and cumulonimbus. Squall clouds and extraordinary

transparency of the atmosphere are the two essential characteristics of the diverse aspects which the rear zone sky may assume.

The typical silhouette of cumulonimbus is not generally to be seen owing to the presence of cumulus or other low clouds (fracto-cumulus, fractostratus) which hide the cumulonimbus clouds when they are distant enough to be seen in profile.

Electrical manifestations are feeble in this zone, (a few peals of thunder, occasional lightning) but they do occur sometimes and even in winter the fall of temperature does not preclude their occurrence completely. Another characteristic of squalls associated with depressions is the regularity of their travel over large areas.

The cumulus which surround cumulonimbus are distinguishable from the cumulus of fine weather by the ragged appearance of their lower edge and the bulging of their upper part. This tendency for cumulus to bulge always exists in rear zone skies.

In fine-weather cumulus the ratio of the vertical height to the horizontal length is of the order of 1/10 but in the cumulus discussed above it may reach or surpass 1/2. Further, the usual typical rear zone skies do not exhibit the most beautiful examples of dome formation which are more clearly shown by the misty cumulus of the weak rear zone sky and the marble-like piles of showery cumulus.

In the clearances in the rear zone, visibility is very good and sometimes exceptional. The sky assumes a characteristically deep blue colour, very different to the pale blue of anticyclonic fine weather which always makes one think of a vivid colour seen as though through a thin veil of dust.

Finally, the last clouds withdraw and fine weather is established. When the passage of a depression is completed by the end of the day, the night air is favourable to the formation of fog. If it passes during the night the distant views are slightly veiled and the blue of the sky is less vivid.

The rear zone phase may last a long time (several days) owing to the persistence of a current of cold, unstable air in which cumulonimbus is formed. This is exemplified in the showery weather typical of maritime polar air, on the coasts of north-west Europe.

Weak rear zone sky.—The clouds to be found in the weak rear zone sky are nominally the same as those of the typical rear zone but their characteristics are different.

Although the lower sky generally does not degenerate beyond the disturbed cumuliform stage, cumulonimbus are not entirely lacking; although the tops bulge considerably they do not present any danger and frequently appear lost in the mist. Electrical discharges from them are practically unnoticeable, and the gusts of wind which accompany them are not at all strong. They frequently pass without a shower, and they never cause heavy rainfall. In weak disturbances it might almost be said that mist takes the place of rain.

The harmless cumulonimbus belonging to the passage of the type in question is generally associated with sheets of altocumulus, while that of thunderstorms and typical rear zone squalls is associated with cirrostratus and false cirrus. Moreover, isolated fragments of altocumulus may be seen trailing across the sky, quite independent of any cumulonimbus. This is of much rarer occurrence in the typical rear zone sky, but may be seen in the thundery sky. The aggregate of these fragments at the same height as the central sheet and following at some distance in its wake, gives a very peculiar appearance to the weak rear zone sky. It is also hardly possible to confuse these altocumulus clouds with those of the weak front zone sky. They are much more opaque, more brilliant and their irregularly arranged banks have silhouettes which are easy to recognise.

Finally, in the place occupied, as will be seen later, by stratocumulus in the rear of a thundery disturbance, there are generally found with the type in question, layers which may be called by the same name, but which exhibit a peculiar structure. The layer looks just like altocumulus in vast rolls or in patchwork formation, but the cloudlets are infinitely greater than in true layers of altocumulus, less geometrical and obviously lower.

The weak rear zone sky is likewise the type which exhibits in its clearer parts certain well-developed cumulus which are sometimes designated " silver-edged " cumulus. These are huge clouds of grey colour, tinged with violet, which when situated in front of the sun remain altogether dark, except for a thin streak of silver or gold which silhouettes them. In the atmosphere, always a little misty at such times, the sun's rays, owing to partial interception by these clouds, form an immense pencil of diverging rays which are clearly visible owing to the mist, and have a very fine, æsthetic effect.

Eventually fine weather returns, provided that the front zone cloud of a new disturbance is not very near.

Pre-Thunderstorm Sky.—Dense cirrus in very soft outlines. Partial veil of thick cirrostratus, domed altocumulus (cumuliformis) in separate tufts—since the beginning of the approach of the storm no thunder clouds in the strict sense of the word have yet arrived—very light wind.

The types of sky which may be immediately followed by the pre-thunderstorm sky are :—cumuliform fine weather sky and the rear zone sky.

The pre-thunderstorm sky begins with cirrus, generally of a very characteristic type. The forms are diverse and most of the varieties which experts have distinguished in the family of cirrus may be said to belong to thundery disturbances. But one characteristic common to almost all varieties is their opacity.

This cirrus is generally derived from the anvils of cumulonimbus clouds whose cumuliform parts have disappeared.

It is accompanied or preceded (sometimes by as much as 24 hours) by isolated altocumulus (cumuliformis) in a fine weather sky, which also indicates a high temperature-gradient in the upper layers.

Compatible lower skies :—the only lower type which can be associated with the pre-thunderstorm sky is the cumuliform fine weather sky which characterizes warm periods in summer, and which may develop considerably in mountainous regions.

Thundery Sky.—Sky characterized by its overcast aspect, chaotic, heavy and as if it were stagnant. The lower layers of cloud often having a mammatus appearance. A general absence of wind, except during the storms or thundery showers. The upper and middle sky is composed chiefly of dense cirrostratus and diverse forms of altocumulus.

The only state of the sky which can immediately precede the thundery sky is the pre-thunderstorm sky.

As in the case of the rear zone sky of an ordinary disturbance, the lower sky is inseparable from the upper and middle sky. Strong convection (cumulonimbus) is an essential characteristic of thundery skies except in the extreme rear, where layers of stratocumulus are often to be found.

The central sheet often produces no rain. It is sometimes difficult to identify and in many cases one is tempted to say that it is altogether lacking, since a thunderstorm system consists essentially of a multitude of cumulonimbus. Some thundery disturbances bring abundant and lasting rain which falls from a vast sheet of altostratus but strangely enough these altostratus clouds do not precede the cumulonimbus ; they follow or are intermingled with them. The name " thundery rains " is kept for these characteristic rains.

In the cirrus family are to be included the extensive sheets of fibrous structure, often fairly transparent and assuming at times the appearance of altostratus. A rather curious veil of cloud is to be seen in these conditions ; it is so vast that it seems to cover the sky and so light that its existence is conjectured rather than perceived. This immense, almost imperceptible gauze is chiefly revealed by the irregularities of its structure and by the effect it has upon the appearance of the stars.

Of the altocumulus family, altocumulus (cumuliformis) is to be seen, a characteristic type in thundery systems. It generally appears in the form of large tufts with ragged edges, without shadows, and set somewhat apart. The colour of the sky at the time of the passage of thundery altocumulus is remarkable, the blue becomes mingled with a violet hue as it loses its vividness. Sometimes the dome formations are accentuated and develop considerably (altocumulus castellatus). These clouds are often white, sometimes grey, and frequently merge into a greyish cirrostratus.

The thundery sky, especially near the end of its passage, is essentially composed of cumulonimbus with its attendant train of fractocumulus and fractostratus. In Europe, the most violent thunderstorms occur in this phase. A curious point about these clouds is their rapid development. The cumulus clouds show clearly the tendency to form domes. Visibility in the clear intervals admits of the same observations as in the rear zone skies of ordinary disturbances. The peculiar trait in the thundery type is the presence of two clouds, the mammatocumulus and a type of stratocumulus which does not appear with absolute regularity in other cases. The mammatocumulus is in a well-known form, but often somewhat different from its classic form ; in place of rounded protuberances it often exhibits rolls or species of festoons.

The thundery stratocumulus is formed of elongated clouds, dull in colour, of a dark bluish grey. It appears in the rear of thunderstorms and particularly in the following circumstances :—when a thunderstorm finishes in the evening and there seems some justification for expecting clear weather the next morning, it often happens that when the sun rises the sky is clouded with an extensive bank of this stratocumulus. It seldom lasts, however, beyond the first few hours of the day.

Isolated banks of cirrostratus may be seen at the same time as the cumulonimbus with, sometimes, lightning in them. These clouds are probably the cirrus anvils remaining from cumulonimbus clouds.

Lower Skies.

" Convection " Skies.

Cumuliform Fine Weather Sky.—Moderate " convection " sky, characterized by the presence of cumulus with horizontal bases, more or less dome-shaped but without excessive vertical development and exhibiting a clear diurnal variation which may result in altocumulus or stratocumulus.

This type of sky has a definite diurnal variation especially on the continent. The evolution of " settled fine " weather may often be observed in the morning from the more or less dome-shaped, fine weather cumulus which lasts during the day but disperses in the evening. In the tropics as well as in the mountainous regions of the temperate zone, fine weather cumulus may grow until it develops into cumulonimbus which gives heat thunderstorms. If the state of the sky reaches this stage it should be classed according to circumstances either as " thundery sky " or as disturbed cumuliform sky with cumulonimbus.

Isolated cumulus sky may develop and assume very varying aspects according to circumstances. It may happen that in the course of its vertical development the cumulus reaches a stable layer which prevents it from developing to any greater height.

The result is a spreading out of the tops of the cumulus which form a sheet resembling altocumulus or stratocumulus. If, subsequently, the convection currents weaken, the cumuliform part of the clouds disperses first while the altocumulus may survive. Thus a fine weather cumulus sky may be transformed, especially towards evening, into a sky characterized by banks of low altocumulus (altocumulus cumulogenitus).

If the stable layer is only a little above the level of condensation, the cumulus becomes extended quite near to its base and may coalesce into an almost continuous layer. A rather analogous phenomenon may frequently be observed in the evening when dome-shaped cumulus subsides and spreads at the level of its base (stratocumulus vesperalis).

Disturbed Cumuliform Sky Without Cumulonimbus.—Intense convection sky, characterized by thick cumulus, jagged and seething but with no cirriform structure at the summits.

When convection takes place in a sufficiently strong current the forms of dome-shaped cumulus are not so regular and they tend to become cumulus congestus. The weather is then more disturbed with rapid alternations of threatening skies and bright periods.

This motley sky of changing cumulus congestus is characteristic of cold, unstable currents which are warmed by contact with warm earth or sea.

Cumulus of this kind—especially that which is formed over the sea—shows much less diurnal variation than fine weather cumulus.

These conditions are especially likely to occur after the passage of a disturbance. In the latter case a cumulus congestus sky is also associated with fragments or broken sheets of medium cloud of altocumulus structure.

Disturbed Cumuliform Sky With Cumulonimbus.—Intense convection sky, characterized by the presence of cumulonimbus.

In the same meteorological conditions convection may increase sufficiently for some of the cumulus to become cumulonimbus. The sky then assumes the typical form associated with showers and bright periods. Sometimes there are thunderstorms too, but they are of short duration and of little importance compared with those of thundery systems.

"Turbulence" Skies.

Stratiform Sky.—Rather low ceiling layer of stratocumulus, stratus or fog; weather not rainy (excepting fine drizzle) but misty.

In those parts of the continents which are within the temperate zone, this state of the sky is mainly observed in winter in anticyclonic conditions; it is very rare in summer. In the intermediate seasons, autumn and spring, it exhibits a clear diurnal variation. First of all there is no low cloud, then fog or stratus is formed by cooling during the night and it is not dispersed until after the sun

rises the next day. In winter insolation is often too small to dissipate the fog or stratus which has formed during the night and the sky may remain overcast for several days in succession—above the stratus there is generally a clear sky, as may be seen from mountains of no great height.

The same skies of fog, stratus or stratocumulus are observable at sea as a result of warm air currents and they may thus extend to regions which are not anticyclonic.

Western Europe is often overrun by this maritime variety of stratus weather. On the northern edge of subtropical anticyclones there is often a humid current from west or south-west in which the stratus and stratocumulus often constitute a lateral connecting zone between two successive disturbances. It should be noted that stratocumulus or stratus of this type do not have a definite diurnal variation ; they sometimes spread over the sky in the middle of the day.

Stratus does not generally give any precipitation ; but if it is very dense it may give drizzle, which is accentuated if orographical effects are also present.

Amorphous Sky.—More or less continuous layer of fractostratus or of fractocumulus of bad weather.

This state of the sky somewhat resembles the preceding one. Nevertheless, it is distinguishable by its more ragged appearance and complete absence of regular undulations. On the other hand, low clouds are never the only ones in this type ; above them there is always either a sheet of altostratus (possibly of altocumulus structure) or cumulonimbus from which more or less continuous precipitation is falling.

Fractostratus or fractocumulus appear at first in isolated masses beneath the veil of medium cloud but they multiply rapidly and from before long a more or less continuous layer.

Mixed Skies.

Stratiform and Cumuliform Sky of Fine Weather.—A mixed stratiform and cumuliform sky may develop when a sheet of stratus or stratocumulus is dispersing and fractostratus is being transformed into fractocumulus. Cumulus is thus formed at the expense of stratus and at the same level where the stratus previously existed.

This type of sky is always transitory, for in the long run one of the two tendencies always triumphs over the other and forms a simple cumuliform or stratiform sky.

Another and much more lasting variety of a mixed stratiform and cumuliform sky may develop when cumulus is formed beneath the cloud sheet (which in this instance consists of stratocumulus, stratus being too low for cumulus to exist beneath it). In the case considered the cumulus is of the fine weather type, remains small and does not reach the sheet of stratocumulus.

Stratiform and Disturbed Cumuliform Sky.—If convection is very strong and if the cumulus clouds assume a disturbed aspect, their summits generally penetrate into or even right through the cloud layer above them, a phenomenon which cannot be seen from the ground but may be seen by an airman. In humid climates this type of sky may give showers.

Amorphous and Disturbed Cumuliform Sky.—This type only exists in showery weather, when fractostratus or fractocumulus may be formed beneath cumulonimbus (sometimes even beneath heavy, disturbed cumulus). In humid climates, fractostratus or fractocumulus may form a layer which fills the spaces between the cumulonimbus. In this case the individual cumulonimbus clouds cannot be recognized in their typical form; nevertheless their passage is indicated by a temporary darkening of the sky and showers.

Tropical Clouds.

The causes of the formation of clouds are the same in tropical as in temperature zones: upward movement, followed by dynamic cooling, cooling by radiation in inversions, and the mingling of layers of air at different temperatures. In the tropics, however, the vertical upward movements resulting from insolation are greater than in the temperate zone, whilst the differences in temperature between neighbouring layers of air in different conditions of motion are much less. Consequently the various phenomena corresponding to the upward movement along surfaces of discontinuity are almost completely eliminated.

Ordinary depressions, with their displacements and characteristic cloud passages, are absent from regions near the equator. For instance, in Indo-China and the Philippines, they are only to be found in the extreme north of the country. Sometimes, beyond 5° from the equator, tropical cyclones, which are certainly disturbances of turbulent character with clearly defined trajectories, are found, but these are very different from ordinary depressions and only appear in a small portion of the tropical zone. For this reason, the organization of clouds is not the same in the tropics as in the temperate zone, although the species of cloud found there are more or less the same. Convection clouds are much more common and play a predominant part. Instead of cloud systems, accompanying depressions, combinations of cloud are to be found, characteristic of a whole season, or certain parts of the day, or certain regions.

As an example of this, in temperate latitudes the association of altocumulus with altostratus is characteristic of the central part of a disturbance, whilst in the tropical zone this association is typical of sunrise.

Nimbostratus resulting from the lowering and thickening of a layer of altostratus is unknown in the equatorial zone. Although there the cloud systems do not show the typical passage from forward

sky to central and then to rear sky, the cloudiness has almost the same variability as in temperate regions. Periods of rain and dryness succeed each other in a similar manner, but they lack definite progression. It appears that, partially at any rate, they are indirectly connected with disturbances passing and developing at a great distance. The essential character of the clouds in these rainy periods is the premature formation of cumulus clouds, and their greater development in the course of the day, owing to the intensification of the convection resulting from the tendency to an upward movement of the whole.

Since the direct effect of migratory depressions is small and the influence of diurnal variations is preponderant, the value of the state of the sky for forecasting purposes is largely lost. This question has only been studied a little so far, but it can be stated, for instance, that the formation of altostratus is by no means an indication of rainy weather; on the contrary, by intercepting the solar rays, this cloud may sensibly reduce the convective motions which are the prime causes of afternoon storms.

Owing to the greater simplicity of meteorological conditions, observations in the tropical zone may help to explain the physical processes in the formation of clouds. The influence of radiation, especially, is strikingly manifested in the equatorial sky, while in countries further from the equator, the dynamic processes preponderate to such a degree that the effect of radiation is often concealed, and consequently there is a risk of its being underestimated.

Cooling by radiation is ordinarily accompanied by the disappearance of clouds: in fact, the cooled cloud-elements sink, become warmed by compression, and evaporate. Cooling by radiation only exceeds the dynamic heating in the case of layers of air resting on the ground or of sufficiently stable layers in the free atmosphere; thus condensations accumulate in stable layers.

Certain cloud formations seem to be due to heating by radiation, such as the formation of amorphous clouds below a dense layer of altostratus.

Particular Characteristics of Cloud Types in the Tropics.

(1) **Cumulus and Cumulonimbus.**—Cumulus and cumulonimbus are the predominant clouds in the tropics. Their diurnal variation, usually very pronounced, at any rate over land, gives the appearance of the sky its chief characteristic.

As the day advances, the vertical dimensions of the cumulus increase. Ascending currents, in an ordinarily light wind and intense insolation find circumstances relatively favourable to their development; hence the height of the upper limit of the troposphere is greater by many thousands of feet than that observed in the temperate zone. The vertical dimensions of cumuliform cloud may increase considerably; thus cumulonimbus reaching a height

of 50,000 feet often rises well above the average level of cirrus at 35,000 to 40,000 feet. In the equatorial zone these clouds have much more of the appearance of turrets than in the temperate zone.

Cumulonimbus is characterized by the formation of fibrous, frozen upper portions which change into cirrus and may last for a long time after the rest of the cloud has disappeared. Often the cirrus plume takes the form of an anvil which may become very large. The " mammatus " variety frequently occurs on the under surface of the wing of the anvil.

After a storm the sky often has a chaotic appearance, the undeveloped forms have for the most part disappeared, and stratiform clouds are observed extending at different levels, being the remains of condensations subsisting in the stable layers. In mountainous countries a layer of altostratus, from which fine rain is falling, is often observed.

In contrast to the sky covered with cumulus is the typical morning sky, with stratiform clouds.

The existence of a well-developed system of sea breezes and land breezes produces a marked contrast in the diurnal variation of cumulus on land and sea, in the neighbourhood of the coast. During the night they dissolve over the land, whilst they form over the sea which is relatively warm ; at sunrise they are observed over the sea, like an endless line along the seaboard.

(2) **Stratus.**—Stratus is a somewhat uncommon cloud in hot countries. It is a local cloud observed on plateaux, in sheltered valleys and above large rivers where mist is formed by radiation during the night. In the morning the solar heat draws up the mist from the ground, producing stratus, which it then evaporates. On the south-west coast of Africa, south of 8 degrees S., where the cold water along the sea coast promotes the formation of a dry mist, and in other similar cases, stratus occurs fairly frequently in the early hours of the morning. Stratus is also fairly common on the coast of northern Indo-China, where the " crachin " mist from the warm sea moistens the relatively cold earth and makes a break in the dry season during January and February ; as it rises this mist forms stratus clouds.

(3) **Stratocumulus.**—Stratocumulus occurs very frequently in the dry season, and all the year round in the region of the trade winds. Owing to anticyclonic conditions an inversion is then formed at a height of about 5,000 to 8,000 feet. The vertical currents in the lower layers coming into contact with the stable layer spread out their products of condensation below this layer, and the result is a more or less thick stratocumulus or flattened isolated cumulus, called trade-wind cumulus. In certain regions, such as the eastern part of the Dutch East Indies, in the neighbourhood of the Kei Islands, thick stratocumulus, producing slight rain, covers the sky from August to October.

Stratocumulus often occurs during the night and morning in mountainous countries where the effect of nocturnal radiation is emphasized by the forced uplift of air at the side of the mountain, raising the whole layer of inversion.

Under these circumstances, the sky is overcast in the morning—a frequent phenomenon on the Indo-China coast, for instance at Hué, during the north-east monsoon—and clears during the day when the strong breezes in the valleys disperse the humid layer and mingle it with the dry air of the higher levels. The diurnal variation of the cloudiness is then more or less the reverse of the normal variation.

On land and on seas where winds from large islands or neighbouring continents blow, the air is hazy in the lower layers but extraordinarily transparent above the level of the stratocumulus.

In the dry regions of the sub-tropical zone, the inversion in which stratocumulus forms may become very intense and consequently the flattening out of cloud very clearly defined.

For the temperate zone the cloud code distinguishes between two combinations of cloud, C_L 7 and C_L 8 (*see* Chapter XV), containing cumulus and stratocumulus. In the equatorial zone, this combination of cumulus and stratocumulus is so rare that as a general rule it may be considered as non-existent.

During the dry season the level of condensation is relatively high, near to the inversion; the products of condensation then spread out in the form of flattened cumulus (trade-wind cumulus) or stratocumulus. The base of these clouds may reach a height of 6,500 feet; with very dry conditions they are of very small dimensions; it is, however, proposed to designate these clouds, in the lower level of condensation, as stratocumulus and not as altocumulus.

During the rainy season, the level of condensation of convection currents is much lower, and stratocumulus at this level is comparatively rare. As a rule, therefore, between the lower level of condensation and an altitude of 10,000 feet no layers exist which are sufficiently stable for the formation of stratiform clouds. As a distinctive criterion between stratocumulus and altocumulus, it may be assumed that, for the tropics, the term stratocumulus should be used for the clouds at the lower level of condensation (convection) and altocumulus for those at higher levels. The name altocumulus will therefore be applied to the clouds which are above the cumulus in the wet season and are generally at an altitude of more than 10,000 feet.

(4) **Altocumulus.**—Altocumulus cloud occurs fairly frequently in the monsoon regions where it forms in the layer of separation between the monsoon with a westerly component and the upper current with an easterly component. At Batavia the west monsoon has an average thickness of 16,000 feet and the average height of the altocumulus is also about 16,000 feet. It is observed,

however, in all seasons, principally in the morning and evening. During the day time its presence is less noticeable, owing to the predominating development of cumulus and of the much less favourable light due to the high altitude of the sun above the horizon. It really occurs more frequently in the cooler part of the day, because its formation is promoted by the cooling by radiation of the air in the inversion. It is not a cloud peculiar to the dry season, like stratocumulus, and it is distinguished from it by its greater height and the smaller dimensions of its elements.

Altocumulus may appear at several levels simultaneously. Often the elements are organized according to certain directions. Frequently small flakes are produced, resembling cirrocumulus, which increase rapidly in size and end by assuming vague contours and dispersing.

Under favourable conditions of lighting, when, for instance, the sun is very low over the horizon, below the balls of altocumulus, small descending trails in the form of " virga " are frequently observed, giving the impression that the lower portions of the altocumulus have left the stable layer and drop into the unstable air and melt away.

(5) **Altostratus.**—Altostratus, which in temperate regions occurs as a typical genus of the central part of a disturbance, is less frequent in the tropical zone, because ordinary depressions do not occur in these regions. It is observed chiefly in the afternoon and evening after a storm, when the remains from the processes of condensation appear in the form of one or more layers of stratiform cloud with the appearance of altostratus or altocumulus. Sometimes when the cumulonimbus has melted away a thick altostratus remains, producing a fine rain. Under the action of nocturnal radiation, it may exist or come into being in the inversion during the night, and it is probably due to this cause that it occurs relatively frequently in the early hours of the morning. Generally, altostratus is accompanied by altocumulus. Altostratus may also form independently of large cumulus ; in this case it is probably a general ascending movement which gives rise to its formation in the humid layers.

(6) **Cirrocumulus.**—Real cirrocumulus is uncommon both in the tropics and in the higher latitudes ; it is easily confused with altocumulus. To distinguish it from the latter, it may be taken as a criterion, that it should be definitely connected with cirrus or cirrostratus, resulting from the evolution of cirrus or cirrostratus, or exhibit characteristics manifesting that it is of ice-crystal structure.

(7) **Cirrus and Cirrostratus.**—Cirrus is a cloud of frequent occurrence in all seasons. Very often an anvil cirrus is observed to grow by the transformation of the tops of cumulus of large vertical dimension. Whereas in other parts of the globe the upper limit of the troposphere only rises a short distance above the average

level of cirrus, in the equatorial zone it extends to a relatively high altitude, exceeding this average level by 10,000 ft. or more. Hence the high turrets of the cumulus often rise to a great height, tending to produce cirriform clouds, and to promote the formation of anvil cirrus.

Cirrostratus is a little more uncommon than cirrus. Its density may vary from that of a fine veil to that of a fibrous, whitish layer, shutting off a great part of the light of the sun. Sometimes this transition is seen to take several hours, obviously under the influence of a general rising movement of the atmosphere. The ice-crystal structure is frequently manifested by the formation of halos.

(8) **Dry Mist.**—In the course of and particularly towards the end of the dry season, a milky veil of mist covers the sky and renders stars of the 4th and 5th magnitudes invisible; under exceptional conditions it may become so dense that the sun only appears as a red disc.

Locally, in the course of very dry years, this mist may acquire the density of a greyish cloud. Low clouds of this genus, real stratus, are fairly common in the straits of Karimata and Casper between Borneo and Sumatra. This mist is formed by hygroscopic particles, a large proportion of which are derived from the products of combustion; these particles increase greatly in size by the absorption of water vapour when reaching the zone bordering on the rainy region, where humidity increases, but where the particles are not yet precipitated by rain. The upper limit of the mist, which lies at a height of roughly 6,500 feet, is generally very clearly marked; under very dry conditions, it appears in the form of a horizon of dust. Above the mist, the air is extraordinarily transparent and very dry, due to its descending motion. At the level of separation an inversion occurs. Convection is therefore limited to the lower mist layers where, owing to the absence of rain, the hygroscopic corpuscles accumulate. The mist then extends to the level of the stratocumulus, but the air is transparent below the altocumulus.

States of the Sky in the Tropics.

Morning Stratiform Clouds.—The contrast between the sky at sunrise and in the afternoon is generally very marked on land. During the night all the cloudlets have disappeared, the products of condensation of the daytime have melted away under the influence of nocturnal radiation; some remains still persist in the form of stratiform clouds, in the inversions, where, in addition, the radiation has caused a quantity of supplementary water vapour to condense. During the dry season, the stratocumuli, which float at the upper boundary of the misty layer are as a rule the only stratiform clouds in the lower and middle levels, but during the rest of the year, when moisture is carried in large quantities up to the high regions of the

troposphere, a number of layers is frequently observed, which together cover the whole of the sky.

Convection Sky.—In these regions, which are pre-eminently prone to thunderstorms, thundery skies, strictly speaking, are nevertheless very uncommon, because thundery systems do not exist near the equator.

Generally, premature formation of cumulus in the morning is sufficient to predict in all probability a thunderstorm in the afternoon. Obviously convection, often reinforced by a general rising motion, is powerful enough to bring cumulonimbus into being, even in cases when the sky exhibits none of the characteristics of the sky in temperate climates.

These thunderstorms are almost all real heat storms of limited duration and extent.

Cumuliform sky may be produced in a number of somewhat different typical ways :—

(1) *Ordinary formation*, produced by slight wind and relatively unstable state of the atmosphere : fine weather cumulus, often becoming transformed into clouds with towers and into cumulonimbus.

(2) *Large clouds of the rainy season*, in the regions and during the periods of relatively strong wind, generally blowing from the sea. The cumulus clouds then closely resemble those of the rear sector of an ordinary disturbance in the temperate zone.

The cause of the convection is partially the same : the wind is relatively fresh : the difference of temperature, however, being small, the convection is intensified by insolation.

The fact that insolation is no longer the principal agent is manifested by the diurnal variation of the rain, which is far less marked than in the former case. Under these conditions, thunderstorms are relatively uncommon.

(3) *Clouds of the dry season*, with downward movement of the air in the higher layers. The latter are dry and usually of great transparency. The convection extends on an average, to a height of 6,000 to 8,000 feet where it is arrested by a characteristic inversion. The layers of air situated below the inversion are very misty for two reasons ; absence of rain precipitating the hygroscopic particles, and convection of insufficient power to carry them to high altitudes. On certain very dry days the temperature of condensation of the ascending currents does not reach as far as the level of the inversion, and an intensification of the mist below the base of the inversion only is observed. In rather more humid

weather, when the level of condensation coincides with the inversion, very flattened cumulus formations appear; finally, with still greater humidity, more highly developed cumuli with flattened tops appear. Here and there, particularly in regions where mountains are favourable to convection, large cumulus clouds may penetrate the inversion, but they do not as a rule reach the stage of cumulonimbus.

Amorphous Sky.—An uncommon, but very remarkable phenomenon is sometimes produced when a dense layer of altostratus covers the sky. A cloud evolution is then observed bearing some resemblance to the transformation of altostratus into nimbostratus in the temperate zone; it differs from this, however, in that there is no passage of an inclined cloud exhibiting to the observer parts in successively lower stages. Fractocumulus and cumulus with fragmentary contours are seen to form, grow, develop in height, gradually fill the entire space below the altostratus and finally combine to form a single thick layer of amorphous structure. It is as though the clouds were sucked up by the higher layer. This phenomenon may be regarded as an effect of radiation. The altostratus is a source of fairly intense radiation, far more intense than that emitted by the clear sky. The fractocumuli are therefore heated under the influence of a two-fold radiation proceeding simultaneously from above and from below, and greater than that which they emit themselves. This effect acts primarily on the edge of the cloud and gives it its fragmentary appearance.

Local Clouds in the Tropics.

In the case of clouds due to ascending currents and low clouds, the dependence on the local conditions of the surface of the globe is manifested much more clearly in the tropics than in the temperate zone.

On land, the wind is relatively slight and the effects of diurnal heating and cooling are relatively large. Consequently, mountains, which are the seats of ascending currents during the day and of downward currents during the night, give rise to a highly characteristic formation of clouds, the same every day, except as regards the intensity of the phenomena.

In the morning, cumuli begin to form on the sides of the mountains, encircling the cones of volcanoes with wreaths of clouds increasing in height and density in the course of the day. Whilst the valley breeze feeds the rising currents, a return wind bends back the upper parts of the large cumulonimbus towards the plain and, when, in the afternoon, the rain refreshes and cools the ground on the sides, the mountain breeze begins to blow and the storm pursues its course to the plain. The diurnal variation of this process

is so marked, and the regularity of this phenomenon so great, that under favourable conditions, as at Buitenzorg (Java) thunder may be heard between 5 and 6 o'clock in the afternoon on 250 days of the year, whereas it is not heard once between 3 and 8 o'clock in the morning.

Another characteristic phenomenon, far more marked than in the temperate zone, is brought about by local differences in the formation of cumulus, due to the distribution of land and sea; there is an abundance of cumulus formation on the sea coast during the day, whereas over the sea, near the coast, the sky remains comparatively clear; at sea, cumuli form during the night whilst they disappear on the coast. Under favourable circumstances, when the trade-wind (or monsoon) blows landwards at night, cumuli are driven towards the coast, it is sometimes observed, at coastal stations, that there is a maximum occurrence of thunder-storms in the early hours of the morning, greater than that of the afternoon.

Another influence which imparts a highly permanent character to local clouds, is the regularity of the general currents in the air, the direction of which remains fairly constant throughout a whole season. Stratocumulus, which during the dry season forms mainly in districts where the wind is forced to rise along the slope of a mountain, is found every morning and the sky clears during the day. These phenomena are so common in such regions that their influence is very distinctly manifested in the average figures for the diurnal variation of cloudiness.

The diurnal variations are of great regularity. Apart from the variations, caused by the succession of the rainy and dry periods which give rise to some modifications in the intensity of the diurnal processes, they are practically the same day after day. Thus, for instance, the mist, which, during the night forms in the lowest lying parts of the district, is a very common phenomenon in certain localities.

The cloud caps on Teneriffe Peak and on the Soembing (Java) are a fairly common phenomenon which is produced on the rounded summits of volcanoes; obviously the forced uplift of the flow lines, caused by the mountain, suffices to produce condensation. The formation of cloud caps is a morning phenomenon; during the day, turbulence causes them to disappear.

Due to the effect of the forced rising of the air on the side of the mountain, the top surface of the cloud sometimes exhibits clearly marked contours, due to the inversion above it.

Although these are not clouds in the strict sense of the word, mention should be made of "Haboobs," sand- or dust-storms of the northern and central Sudan. Sometimes the sand is lifted to a height of more than 3,000 feet.

CHAPTER XI.

Precipitation

The term precipitation is used to denote any one of the results of the condensation of the invisible water vapour contained in the atmosphere. In practice the use of the word is limited to describe an appreciable deposit in either the liquid or the solid form, and hence the elements of mist, fog and cloud, while produced by the condensation of water vapour in the air, are not included in the practical application of the term. The various forms of precipitation will now be discussed.

Rain.—When a mass of air containing water vapour is made to ascend, it may reach a stage where it is cooled below its dew-point, and the water vapour will be partly condensed into fine particles which at first remain suspended in the form of cloud. If convection within the cloud becomes active, these minute water particles will be carried higher up, so that the process of cooling and of condensation is continued. The droplets coalesce to form larger droplets, and on reaching the stage where the upward currents are no longer able to sustain them, they begin to fall downwards. As they descend through the cloud they increase in size, due to the accumulation of more moisture. On emerging from the cloud base, however, they enter a layer of air which is no longer saturated and where the temperature is higher, so that evaporation takes place. If the rate of evaporation is great it may prevent their reaching the earth, but if the raindrops are sufficiently big and numerous, a certain number will reach the earth's surface. In this case rain is said to be falling.

The size of raindrops depends very largely on the temperature. In general, the drops are larger in warm countries than in cold climates, and larger in summer than in winter. This is due to the greater quantity of moisture present in the atmosphere under these conditions, and also on account of the greater activity of the convection currents.

Rain is classified in three main types, according to the manner in which the cooling mentioned above is brought about :—

(1) Convectional rain.
(2) Frontal rain.
(3) Orographic rain.

(1) **Convectional Rain.**—When the water-vapour content in the air is high, and the lapse-rate of temperature in the lower strata of the air is large, due to surface heating, and when, as a result, active convection currents are produced, rain of this type is experienced. Such conditions are associated especially with hot lands,

and the hottest hours of the day. The heavy rain of the tropical regions and much of the summer rainfall of the temperate latitudes, including that associated with many thunderstorms (page 276), is brought about in this way. On the equator the rainfall is heaviest at and just after the equinoxes, when the insolational heating is greatest. Between these two maxima there is less rain, and some months have very little. On each side of the equator there is a belt where the two rainfall maxima fall within the same half year corresponding approximately to the times of overhead sun, and are separated merely by a month or two of less rain, the other half year being almost rainless. The interval between the two periods of maximum rainfall becomes progressively shorter as the distance from the equator increases. This regime prevails between about latitudes 5 and 12 degrees north and south. Outside this is the belt between about latitude 12 and 16 degrees north and south, where there is a single maximum in the rainfall curve at and after the time of overhead sun, and a long dry season with no rain at all.

During the summer months the tropical waters of the great oceans are the scene of severe revolving storms within which the vortical ascent of air currents may be intensely active, producing thick cloud and heavy rain, the high temperature and humidity of the air conducing to the intensity of condensation.

Within the trade-wind belts the convection currents are feeble and the rainfall, if any, is intermittent and of slight intensity. In the calm regions of the horse latitudes, the air currents are mainly descending currents, and as a result the rainfall is insignificant. The poleward flowing air currents which emanate from these regions, in their passage over the sea, absorb at first large quantities of moisture, but as they are travelling from a warmer to a colder area, become cooled down, and so they reach a point where precipitation takes place. This precipitation decreases in amount with increase in latitude, and within the polar circle the amount becomes comparatively small. The general circulation of the atmosphere, therefore, tends to produce a distribution of rainfall, of the convectional type, showing a maximum near the equator, two minima in latitudes 20 to 30 degrees north and south, two maxima again in latitudes 40 to 50 degrees north and south, and thereafter a gradual decrease towards the poles. This general distribution, however, is considerably affected by the unequal division of land and sea.

Rain associated with thunderstorms (page 276) affords a striking example of the convectional type. Strong surface heating and the consequent development of a large lapse-rate of temperature induces active convection currents, and if there is an adequate supply of moisture available, towering thunderclouds may be formed, accompanied by rapid condensation and heavy rainfall.

Instability showers associated with polar air currents provide another example of the convectional type of rain. Such instability is caused by the warming of the lower layers of the air or by the

cooling of the upper layers or by a combination of both causes, and this takes place in maritime polar air currents in the course of their long journey over the ocean, giving rise to strong convectional activity and intermittent heavy showers. Over the sea these showers may occur at any season but they are more frequent in autumn and winter ; after the current has crossed a considerable land area they are most frequent on spring and summer days when the surface heating is marked, and they are rare in mid-winter.

(2) **Frontal Rain.**—Rain of this type is associated with depressions of the temperate regions and in general is more continuous than rain of the convectional type. On the other hand, frontal rain is seldom as heavy or as destructive as the convectional rain of the tropical regions. Frontal rain is formed when a warm, moist mass of air is forced to ascend over a colder and therefore denser mass of air, as at a warm front (page 348). In this case the rain band may be as much as two hundred miles wide, and during its passage over a station the rainfall will be more or less continuous, commencing from altostratus cloud and gradually developing into a steady downpour which ceases or diminishes to drizzle as the warm front passes. Again, when a cold current of air undercuts a warm moist current and causes it to ascend, rain will fall, as at a cold front (page 350). Owing to the steeper slope of the surface of discontinuity between the two air masses, the rainfall will not be of such long duration, but on account of the more rapid ascent of the warm air will be much heavier than in the case of a warm front. The rainfall associated with an occlusion (page 351) exhibits to a certain extent a combination of the characteristics of warm-front and cold-front rain, both in duration and intensity, so that rain may occur before and after the passage of the occlusion.

Unlike convectional rainfall, frontal rain is, in general, independent of the time of day and of the surface temperature at the place where the rain falls. Convectional influences, due to the surface heating of the lower layers of the atmosphere, may, however, be present, together with other diurnal effects, thereby stimulating the ascensional process of the warm air.

In the case of warm fronts there is some tendency for the rain to reach its maximum in the latter part of the night or early morning ; and, particularly in summer and over land, the rain may cease in the middle of the day and early afternoon, or at all events fall off considerably. In the case of cold fronts, on the other hand, there is a very marked tendency indeed for most rain to occur in the afternoon, and there is a slight tendency to a secondary maximum about 0400. Depressions of the temperate regions have their greatest frequency in the winter, and consequently the maximum rainfall of the frontal type occurs at that season of the year.

(3) **Orographic Rain.**—When a current of moisture-laden air strikes a mountain range it is forced to ascend ; the process of

ascent tends to cause condensation and precipitation, especially on the windward side of the mountain range. This type of rainfall, which is the heaviest on earth, is known as orographic rain. An example of this is afforded by the Western Ghats of India, where the south-west monsoon winds, saturated with moisture after their long journey over the Indian Ocean, have to rise to a height of over 4,000 feet. On the windward side of this mountain range, therefore, during the period of the south-west monsoon there is an abundance of rain which is almost continuous for two or three months, and only ceases when the south-westerly current is cut off. After depositing its moisture the air current descends to the Deccan and becomes dry through dynamical heating and consequent evaporation. As a result, the rainfall on the leeward side of these mountains is comparatively slight.

The rainfall of the west coast of Scotland is another example of rain caused by the process of ascent over mountains. Here the prevailing westerly winds, well supplied with moisture after their journey over the Atlantic, are forced to ascend over the west-facing edge of the Highlands, and in so doing, deposit their moisture content in the form of rain. There are many other striking illustrations of rain produced in this way. Whenever a mountain range or table land is exposed to winds, which, in the course of their journey have acquired a high moisture content, there is always considerable rainfall on the windward side, while on the leeward side, especially of high ranges, the rainfall may be slight, or even negligible.

The effect of forced ascension of air is also present when on-shore winds are blowing at a coastal station. All wind is impeded by friction with the surface over which it blows, the effect being least over the sea and greatest over a forested land. Hence the lower strata of on-shore winds are slowed down when they reach the land, and there is formed, as it were, a " block " in the circulation, which the air that continues to press on from the sea has to overrun. The heavy rainfall of the densely forested Niger delta is partly due to this.

Orographic influences may be present also in the case of frontal rain, the process of ascent of the warm air being rendered more active when the front passes over a range of mountains, while the phenomenon of a cloud-burst (page 280) is brought about by the sudden cutting-off of the supporting up-currents below a thunder-cloud as it passes over an area of high land.

Snow.—When water vapour condenses at a temperature below 32° F. it passes into the solid state and the minute ice crystals thus formed in the air are of a feathery or needle-like structure. When the temperature of the air is very low—under 8° F.—the snow falls in single crystals, but above this temperature the crystals unite to form snow-flakes which may be as large as an inch across. The air through which the snow descends to the earth must be at a temperature below or little above freezing-point (a surface tempera-

ture of 37° F. is a good criterion), or the snow-flakes will melt. Rain is often mixed with the snow, when the temperature of the air is somewhat above freezing-point, resulting in the formation of **sleet.**

Snow occurs in all latitudes from equator to pole, but at sea level it is very rare inside the tropics. The temperature of the air decreases with increase in height; consequently snow is quite common at high altitudes in tropical regions where there is abundant moisture in the air, and even perpetual snow is found on the Andes near the equator. Between latitudes 30 degrees and 40 degrees snow is of rare occurrence at sea level, but it does occur occasionally on the northern coasts of Africa, in Florida and Greece. Farther inland, and on high ground, it is quite common, as in Palestine, Spain and the southern states of North America. Between latitude 40 degrees and the poles, snow is common at all levels during the winter, with the possible exception of the northern seaboard of the Mediterranean, where it is seldom experienced. In the temperate regions, however, the lowlands have less snow than might be expected, for although in continental interiors the winters are certainly cold enough for snow formation, the air contains little water vapour. Over Europe the snowfall is heavier on the western shores than over the central regions, though on account of air temperature differences, the coastal regions are not covered with snow for the same length of time as central regions are. Over eastern Canada, on the other hand, the snowfall is a maximum on the Atlantic seaboard and decreases towards the western prairies. The polar regions, especially that of the southern hemisphere, are characterized by immense areas of ice and snow, where the cold in winter is intense and even in the warmest months the mean temperature is well below freezing-point on the coasts, while on the high snow-fields of the interior it is very much colder.

Dew.—When the earth's surface or the deck of a ship become cooled to a temperature which is below the dew-point of the surrounding air, moisture is deposited upon them in the form of dew. Such a reduction of temperature is caused by the nocturnal cooling of the surface by radiation of heat to the sky. Conditions are most favourable for the formation of dew on a clear night when the air is still, for, if the sky is overcast, there is little heat loss by radiation from the surface and much wind continually brings new air into contact with the radiating surface whereby its temperature is maintained above the dew-point. Dew is more abundant in valleys, where the air is cold and stagnant, than on high ground where the air is usually in motion. It forms most readily upon objects such as grass and leaves, which, having a large surface area relative to their volume lose heat freely by radiation and undergo a considerable reduction of temperature in consequence.

Hoar Frost.—If the temperature of the radiating surface falls below freezing-point, and at the same time below the dew-point of

the surrounding air, the vapour immediately passes into the solid state and a deposit of ice-crystals is formed. Such a deposit is known as hoar frost. Frost, like dew, is more prevalent in valleys than on high ground, since cold air, on account of its greater density, gravitates down the valley slopes and lies stagnant at the bottom, thereby increasing the cooling effect of the ground.

Rime.—When frost and wet fog occur together, the minute super-cooled particles of which the fog is composed, freeze on coming into contact with solid objects such as trees, telegraph wires, wind-screens, wings of aircraft, etc., and a deposit of white rough ice may grow out to windward from these objects. This deposit is known as rime.

Glazed Frost.—Glazed frost is generally caused by the falling of rain on to a surface whose temperature is below freezing-point. It may also be due to the effect of a warm moist current setting in suddenly over an area which has experienced intense cold for some time. Unlike hoar frost and rime, which are white and opaque, glazed frost is a transparent smooth coating of ice and may form on the ground, trees, and even on the wings of aircraft.

Hail.—Hail consists of hard pellets of ice of various shapes and sizes and more or less transparent, which fall from cumulonimbus cloud and are often associated with thunderstorms. Hailstones may attain a great size; stones as large as golf balls have been observed in Europe, and the recorded weights range up to over 2 lb. The rapidly ascending current of a cumulonimbus cloud is quickly cooled by adiabatic expansion and cloud particles and raindrops are formed at comparatively low levels. The rain does not immediately fall to earth, however, but is carried upwards by the rising current of air. If the rising current exceeds 25 feet per second all the condensed water is retained in the cloud; if the rate of ascent of the air is less than this the larger water drops fall to earth and the smaller drops only are carried upwards. At a level, not usually exceeding 13,000 feet in temperate latitudes, the temperature has fallen below the freezing point. Ice crystals are not immediately produced, but water drops, in the super-cooled condition. Near the top of the cloud, however, ice crystals appear and these will grow into soft white pellets by a process of condensation direct from vapour to solid. When the weight of a pellet is sufficient to overcome the resistance of the ascending air current it will commence to fall. It encounters the super-cooled water drops which freeze immediately on coming into contact with it, and imprison a quantity of air in the process thereby surrounding the pellet with a coating of opaque white ice. In the lower part of the cloud the hailstone encounters raindrops not in the super-cooled condition; these freeze on to the pellet comparatively slowly, surrounding it with a coating of clear ice. When hailstones are split open to show their internal structure one can nearly always see the inner soft white

mass of ice which was collected while the stones were in the super-cooled region, surrounded by a layer of clear ice formed by the freezing of the water deposited when the stone was passing through the non-super-cooled region.

Sometimes the hailstone is again caught in the ascending current and carried to the top of the cloud. On descent it acquires another coating of soft white ice in the super-cooled region and of clear ice in the lower region. This process may be repeated several times so that when the hailstone finally falls to the earth it consists of a number of alternate layers of soft white, and clear hard ice. This structure has been found on splitting some of the larger hailstones.

In winter the freezing point is reached at a very low level and consequently there is no region of non-super-cooled water drops in the lowest part of the cloud. The hailstones therefore acquire no depositions of hard clear ice but consist entirely of the white pellets known as soft hail or "graupel."

The Formation of Ice on Aircraft.

(From a paper "Ice accretion on Aircraft," by G. C. Simpson, K.C.B., F.R.S.).

The Physics of Ice Accretion.—Clouds consist of particles of water in either the solid or liquid state. In the solid state they are small crystals of ice in the form of flat hexagonal plates, like stars, or of long thin hexagonal rods. The size of these crystals varies greatly, and they are generally so small that they cannot be seen individually by the naked eye; but under favourable conditions they may grow until the plates are as much as an eighth of an inch in diameter and the rods of a similar length.

Generally speaking the size depends on the temperature in which they are formed, the larger sizes being found only when the temperature is just below the freezing point. At low temperatures the crystals occur singly and there is little in a cloud of small ice crystals by which a pilot can distinguish it from an ordinary cloud of water drops. At higher temperatures when the crystals are larger their branches become entangled and snow flakes are formed. Each snow flake consists of a number of ice crystals and with care a dry snow flake can be separated into its constituent crystals.

Cloud particles in the liquid state are always spheres, and they vary in size from nuclei of ultra-microscopic dimensions to raindrops having a diameter of a fifth of an inch, which is the largest drop of water which can hold itself together as it falls through the air.

On first thoughts one would expect that clouds would consist of ice particles when the temperature is below the freezing point and of water when it is above the freezing point. Unfortunately no such simple rule holds good. Water in the form of liquid drops can exist

for an indefinite time in an atmosphere many degrees below the freezing point. As the drops must have the same temperature as the air surrounding them we have the apparent paradox of a liquid at a temperature far below its freezing point. This, however, is a condition often met with in nature and such a liquid is said to be supercooled. A supercooled liquid is, however, in an unstable state and it needs little inducement to cause it to take up the solid state appropriate to its temperature. No supercooled drop of water could survive being broken up without the formation of ice, and it is only necessary for supercooled water to be touched with a particle of ice to induce freezing. There again we meet with an unexpected fact. If a supercooled drop is induced to freeze it does not usually turn completely into ice ; only a fraction of the water goes into the solid state, the remainder retaining the liquid state. This is due to the fact that when water turns into ice it releases a relatively large amount of heat which becomes available for raising its temperature. Laboratory experiments have shown that when a drop of water freezes, it releases sufficient heat to raise its temperature through 144° F. Suppose that we have a supercooled water drop at a temperature of − 112° F. and induce it to change into ice. Sufficient heat is released to raise its temperature by 144° F. ; thus, after turning into ice the whole drop is at a temperature of 32° F., that is at its freezing point. Now take another supercooled drop at a temperature of − 40° F. and induce it to change into ice. When half the drop has frozen sufficient heat will have been released to raise the temperature of the whole drop by 72° F. ; that is the temperature of the drop at that stage will be − 40 + 72 = 32° F., again the freezing point. It is clear that at this stage the freezing must cease for any further freezing would release heat which would only melt the ice already formed. In other words only 1/144th part of a supercooled water drop will freeze for each degree Fahrenheit its temperature is below the freezing point.

This solidification of a portion of a supercooled drop, with the raising of the temperature of the whole to the freezing point, takes place every time an aeroplane strikes a cloud particle when flying through a water cloud at a temperature below the freezing point ; but this is not the whole story.

To fix our ideas let us imagine an aeroplane flying through a cloud when the temperature is 28° F., that is 4° F. below the freezing point. A drop strikes the leading edge of the wing, it is disrupted, freezing starts and 4/144th part immediately turns into ice and the remaining 140/144th part becomes water at the freezing point. This water, however, cannot remain liquid long because it is in contact with the cold material of the wing which is at the temperature of the air, and also it is in a stream of air four degrees below the freezing point ; hence unless it is blown off it soon freezes and ice equivalent to the mass of the drop is attached to the wing.

Here we have an explanation of the different forms in which ice

accretion appears on an aircraft. If the temperature is very low and the drops are small the amount of liquid from each drop is so small that it cannot spread, but immediately freezes just where it strikes. In other words, each cloud particle becomes an ice particle attached to the wing. These ice particles remain individuals and a large amount of air is imprisoned between them. The ice accumulation is, therefore, of a light texture and grows out towards the air stream. The mechanical impact of the air, however, tends to consolidate the accumulated ice, and the result is an ice accretion confined to the leading edge of wings and struts which has a white opaque appearance and can easily be shaken off. It is really rime (which is formed in exactly the same way) but compressed by the impact of the air.

If the temperature is relatively high and the drops are large, the amount of water which remains unfrozen on impact is large and tends to spread over the wing; if at the same time the drops are abundant the water from neighbouring drops unites and the leading part of the wing becomes wet. The air stream causes this water to flow back over the wing freezing as it flows. Cases are on record in which, during ice accretion in dense clouds, water has been observed to stream off the wing. The ice in this case is formed on the wing from a sheet of liquid water and therefore has the transparent glassy appearance of ordinary ice. Clearly, ice formed in this way is tough and sticks closely to the fabric of the aircraft; it cannot easily be shaken off, and if it breaks off at all it comes away in lumps of an appreciable and sometimes dangerous size.

A form of ice accretion, similar to that just described but much more heavy and dangerous, occurs when the source of the water is rain instead of supercooled cloud particles. Temperature does not always decrease as one ascends in the atmosphere, but on occasions there is a layer of warm air with temperatures above the freezing point overlaying a layer in which the temperature is well below the freezing point. Precipitation which falls from the warm layer will be in the form of rain and when this enters the cold layer it will not freeze immediately. An aircraft flying through the cold lower layer will become wet from the rain and the water will freeze owing to the low temperature of the air. Great masses of clear glassy ice will form on all parts of the aircraft, the weight of which may force it down in a very short time. This form of ice accretion is called " glazed frost " in this country and " glaze " in America.

The three forms of ice accretion which we have just considered have all required the presence of water already precipitated into either cloud particles or raindrops. There is a fourth form of ice accretion which may occur in clear air due to the formation of ice on the aircraft directly from the water vapour in the atmosphere. This occurs only when the aircraft, after remaining sufficiently long in a cold region of the atmosphere to take up the low temperature, rapidly moves into a warm damp layer. In this case the temperature of the surface of the aircraft may well be below the dew-point of the

damp air and also below the freezing point. In consequence water vapour is condensed directly as ice on to the surface of the aircraft. The method of formation is similar to the formation of hoar frost on grass and shrubs during a clear frosty night, and the ice accretion on the aircraft has the white semi-crystalline appearance of hoar frost. Usually this form of ice accretion is not serious, but cases are on record in which aircraft, descending from great heights where the temperature was very low into a layer of damp air or even altostratus clouds, have encountered fairly rapid accretion where none was experienced during the ascent. This has been sufficient to cover the aerial and thereby cause it to alter its capacity.

The four main types of ice accretion may be summarized as follows in order of increasing danger :—

(*a*) A white semi-crystalline coating of ice which covers the surface of the aircraft. It may form in clear air when a cold aircraft enters warmer and damper air during a rapid descent. It has little effect on the flying of the aircraft, but may obscure vision if it forms on the wind screen and interfere with W/T communication by coating the aerial with ice.

Surface equivalent : hoar frost.

(*b*) A light white opaque deposit of ice which accumulates on the leading edges of wings, struts and wires of an aircraft. It forms when flying in filmy clouds consisting of small supercooled cloud particles and generally at low temperatures. The deposit has not great weight ; but its danger lies in an alteration of the aerodynamic characteristic of the wings and in the choking of the orifices of the carburettor and flying instruments.

Surface equivalent : rime.

(*c*) A transparent or translucent coating of ice which has a glassy surface appearance. It forms within dense cloud consisting of large cloud particles and generally at temperatures near the freezing point. Its danger is due to the weight of the accumulation and to vibration set up by the unequal loading of wings, struts and especially the blades of the propellers. When appreciable blocks of ice break off the vibration may become so intense that there is danger of fracture of the aeroplane structure.

Surface equivalent : intermediate between rime and glazed frost.

(*d*) A heavy coating of clear ice which forms all over an aircraft when rain falls on it from a warm upper layer while it is flying in a layer with temperature below the freezing point. The danger is great owing to the weight of the large mass of ice which may accumulate in a short time

Surface equivalent : glazed frost.

The Meteorology of Ice Accretion.—The physics of ice accretion show that for ice accumulation to become dangerous it is necessary that there should be an abundance of water in the liquid state at air temperatures below the freezing point. These conditions are met with only in clouds or in rain, and the meteorology of ice accretion is mainly concerned with the formation of clouds, the formation of precipitation and the temperature.

The Formation of Clouds.—Cloud is always formed by the cooling of air below the dew-point, that is below the temperature at which the air is saturated. At higher temperatures the water is present in the form of invisible vapour. Atmospheric air contains a large number of " condensation nuclei " which, except near large industrial centres, are almost exclusively particles of sea salt derived from sea spray during storms at sea. When the temperature of the air approaches the saturation point these nuclei attract water molecules and commence to grow in size. After saturation has been reached all the water condensed out of the air in consequence of further lowering of the temperature goes to increase the size of the drops formed around the nuclei. The cooling of the air which leads to condensation is normally due to ascending currents, for air cools by expansion as it rises. As soon as the air rises to the height where it reaches its saturation point cloud particles are formed on the nuclei as already described. In still air these would fall ; but in general the air is rising at a greater rate than that at which the particles would fall in still air ; they would therefore be carried upwards, growing in size all the time. An ascending current is therefore of considerable importance for ice accretion ; it supplies the vapour which is condensed into water and carries the water particles thus formed upwards into cold regions where they become supercooled. The greater the speed of the upward current, whether formed by warm air ascending or by the forcing upwards of air streams where they impinge on rising ground, the greater the accumulation of water in the cloud, and the greater the probability that the cloud particles will consist of supercooled water drops.

Precipitation.—A cloud of any great thickness may be divided into three parts. In the upper part where the temperature is well below 0° F., the cloud will be entirely ice, the cloud particles themselves consisting of minute ice crystals ; in the middle part with temperature between 0° F. and 32° F., the cloud may consist of either ice crystals or water particles or a mixture of both ; in the lower part where the temperature is above 32° F., the cloud particles are water.

Precipitation, *i.e.*, snow, sleet or rain, normally starts in the middle layers as snow, the individual ice crystals becoming interlocked into snow flakes which fall more rapidly than the ascending current of air can drive them up. When the snow descends into regions where the temperature is above the freezing point the snow flakes first partially melt, giving rise to sleet, and finally melt com-

pletely into raindrops. There is reason to believe that all appreciable rain, other than drizzle, is formed in this way.

In the upper part of the cloud ice accretion cannot occur for the ice crystals are dry, and if they strike the aircraft they fall away again, like any other kind of dry dust. If there is only dry snow in the middle part of the cloud, there will be little or no ice accretion for the snow which strikes the aircraft is quickly shaken off. It is however in this part of the cloud that supercooled cloud particles occur, and all the typical forms of ice accretion which have been described above may be encountered. In the lower part of the cloud ice accretion can only occur if there is sleet; for sleet being of a sticky nature adheres to aircraft and quite appreciable accumulation may be collected.

Temperature.—The part played by temperature in the process of ice accretion is primarily that of determining whether or not ice will form. If there are water drops present ice will be formed when they strike the aircraft whatever the temperature may be so long as it is below the freezing point. The temperature however plays an indirect role because generally speaking the temperature determines the amount of water present.

The total amount of water vapour which air can contain is strictly limited by the temperature, and, when condensation takes place in a mass of air, the amount of water deposited as cloud particles for a given fall of temperature is the greater the higher the temperature. Thus clouds formed in the cold air of the upper atmosphere are less dense than the clouds formed in the warmer air of the lower strata. We should therefore expect for this reason alone that ice accretion would be less the lower the temperature. This is borne out by actual experience, for 217 records of ice formation on aircraft in England during 1932–7 were distributed as regards temperature as shown in the following table :—

Temperature	Above 32° F.	32° F. to 28° F.	27° F. to 23° F.	22° F. to 18° F.	17° F. to 13° F.	12° F. to 8° F.	7° F. to 3° F.	2° F. to −2° F.	Below −2° F.	Total
Occurrence of ice formation	0	56	78	43	13	15	3	4	5	217

The numbers in this table have no very strict meaning, for the number of flights within each group of temperatures was not the same; but it does show clearly that ice formation is more frequently met with in this country at temperatures just below the freezing point than at temperatures much lower. Temperatures between 32° F. and 23° F. account for more than half the occasions of ice accretion in this country, and a similar result is found in other countries also. At lower temperatures the frequency falls off rapidly and practically ceases at 0° F. That icing should not occur

at lower temperatures than 0° F. is theoretically explained by this being the lowest temperature at which a solution of sea salt and water can remain liquid—water particles form on crystals of sea salt as nuclei. In other investigations occasional occurrences of icing at temperatures one or two degrees above the freezing point have been reported and generally ascribed to the lowering of the temperature of the deposited water due to rapid evaporation (wet-bulb effect). Ice accretion can also take place at temperatures slightly above the freezing point when sleet is encountered. Thus while, generally speaking, the freezing point is the temperature at which ice accretion ceases, one must be prepared to meet it occasionally at temperatures one or two degrees higher.

Annual Variation of Ice Accretion.—In order to exhibit the variations of the frequency of ice accretion from month to month the following table has been prepared based on the reports referred to above :—

Annual Variation of Ice Accretion.

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Average height (ft.) of freezing point.	1,800	1,800	3,300	4,900	6,600	8,400	9,800	10,400	9,800	8,000	5,500	2,600
Average height (ft.) of ice formation.	3,200	3,800	4,100	4,300	6,900	6,400	10,500	..	6,100	5,900	4,600	4,300
Total reports of ice formation. 1932–6.	34	46	20	20	6	1	2	0	3	7	29	27

The bottom line of this table shows that in this country there is very little ice accretion reported during the summer months—June to September—while the months of greatest accretion are January and February. The reason for this is clearly seen when one examines the first line which gives the average height at which the freezing point occurs in the different months. The greater part of the flying in this country is carried out at heights between 2,000 and 5,000 feet ; hence during the summer relatively few aircraft fly at a height where the temperature is below the freezing point, thus accounting for the absence of reports during the period June to September. In December, January and February the height at which freezing occurs is well below the height at which aircraft normally fly, hence the reports during that season are relatively numerous. In the second

line is given the average height at which ice formation was first noticed; this line of figures however is misleading if one uses it to find the height at which ice accretion is most likely to be met with. In January and February the average height of the freezing point is 1,800 feet; but the average height at which ice accretion was observed was between 3,200 and 3,800 feet. This is not because ice accretion is most likely to be met with at that height, but chiefly because the majority of the aircraft were flying at that height. If there had been an equal number of aircraft at all heights the most frequent layer of ice accretion would have been lower than that actually recorded. For practical purposes the first line gives the most valuable idea of the height at which ice accretion is likely to be met with at different seasons, as ice accretion is predominatingly confined to the five thousand feet above the height at which the freezing point occurs.

The result of this investigation is just what one would expect, both on theoretical grounds and on the results of experience. Ice accretion is most frequently met with in the winter, first because there is then more cloud and bad weather, and secondly because the temperature, at the height at which flying takes place, is round about the freezing point which is the temperature at which ice accretion is the most frequent and the most dangerous. During the summer there is less cloud, and the freezing point is not reached much below 8,000 or 9,000 feet; for both these reasons aircraft seldom become involved in ice accretion during the summer months.

General Rules for Avoiding the Dangers of Ice Accretion.—When ice accretion is observed in flight there is one and only one method of escaping its dangers, and that is to get out of the ice-forming layer at the first possible moment.

For all practical purposes the ice-forming layer is limited to the temperatures between 32° F. and 8° F. Unfortunately these temperatures represent normally a height difference of 7,000 feet, and it is not always possible for an aircraft to climb this distance especially if its climbing qualities have already been affected by ice accumulation. Still it is not always necessary to climb the full amount for the ice accretion ends with the cloud, so that, if the upper surface of the cloud is known to be within reach, one way of escape is to ascend into the clear air above the cloud where further ice accretion will cease and the ice already formed will evaporate or shake off.

The alternative is to descend to where the temperature is above the freezing point, and this is the most usual procedure in this country, for during periods of heavy cloud and precipitation even in winter the temperature at the ground is usually well above the freezing point, and there is on most occasions sufficient room for flight below the clouds.

If, however, the cloud is down on the hill tops, or the visibility is bad beneath the cloud owing to mist or precipitation, there is no

alternative but to climb in the hope that the ice accretion will decrease with lower temperatures or the top of the cloud will soon be reached.

In the winter another dangerous condition may arise in which it is not safe to descend. This occurs when glazed frost is observed to be forming, caused by raindrops falling on to the aircraft and freezing there. Luckily this form of ice accretion does not often occur in this country; but when it does, it is always due to a warm air current, in which the rain forms, mounting upwards over a layer of cold air, the temperature of which is below the freezing point. It is clear that in these circumstances there is no escape by going lower, for the cold air rests on the surface; the only hope of escape—unless one can land—is to climb until one passes out of the layer of cold air into the warm air from which the rain is falling.

One of the greatest difficulties which face a pilot when he meets with ice accumulation is to decide what to do. Ice accretion varies greatly in intensity, from a light deposit which can do no damage to a heavy deposit which makes the aircraft unmanageable in a very few minutes. Ice accretion may be met with in a climb and quickly left behind; on the other hand, if the aircraft is in level flight quite a small rate of ice accretion may produce in course of time dangerous accumulations. Experience and common sense are the best guides as to what action should be taken; but the following points might be made:—

(*a*) The lower the temperature and the thinner the cloud the less the danger of large accumulation of ice, and the more time required for the accumulation to build up.

(*b*) If the cloud is dense and clear ice is observed to be forming the danger is great, and no time should be lost in seeking security. The danger is greatest when ice is observed to be forming at the same time that rain, sleet or snow is falling.

(*c*) If the reserve power of the aircraft is great escape is best attempted upwards, unless it is definitely known that good conditions exist below.

(*d*) If the aircraft is heavily loaded and has little reserve power, an ice-free region should be sought at lower altitudes as soon as ice accretion, no matter how small, is observed.

(*e*) If action is taken quickly the reciprocal course is often the wisest.

If it is known or suspected that ice accretion may be encountered on a flight—and its possibility should always be borne in mind during the winter months—the meteorological officer should be consulted and if possible a synoptic chart examined. From the synoptic chart the regions of activity along a front can be seen and their motion determined. Regions where the wind impinges on high land, so causing increased danger, can be marked out. The meteoro-

logical officer will be able to calculate the height of the layer in which the freezing point occurs, and compare it with the height of the cloud base and the height of the hills. Unfortunately the top of the cloud layer can seldom be estimated; but on some occasions even this may be possible. With this information the pilot will be in a position to choose the route to take, to consider alternative routes if the one he has chosen proves impracticable and he will be prepared to decide quickly whether to ascend or descend when ice-forming conditions are experienced. If before starting the pilot decides to climb above the clouds if he meets ice accretion, it would be best to fly above the clouds from the start.

CHAPTER XII.

Thunderstorms Line-squalls and Tornadoes

THUNDERSTORMS.

Thunderstorms occur through marked instability in the atmosphere, and the phenomena of thunder and lightning which accompany these storms are not the cause of the storms but the result. Under suitable conditions of temperature and humidity, rapid convection and condensation take place, resulting in the formation of large cumulonimbus clouds. The explanation of thunderstorms is to be found in an examination of the convectional process that goes on within these clouds.

The conditions necessary for the development of thunderstorms are :—

(i) A lapse-rate of temperature in excess of the saturation adiabatic through a range of not less than 10,000 feet above the cloud base (and usually much more in the case of severe storms).

(ii) An adequate supply of moisture from below.

In temperate latitudes the requisite conditions enumerated above exist most frequently in cols or shallow depressions, and in general are neither so frequent nor severe as in the tropics. In the British Isles the main risk of thunderstorms occurs in summer and more especially in the east, south-east, north-east and midlands of England and in the east of Scotland. This is due to the fact that with the prevailing westerly winds most convectional activity occurs in these areas in summer. The hottest summers are not, however, the most thundery, for there are many hot summer days, and especially in anticyclonic areas when the above conditions are not fulfilled. Thunderstorms occur most frequently in the afternoons, when the insolation is at a maximum, but if a suitable degree of instability and humidity exists, and a suitable impulse or " trigger action " (*see* page 155) is provided (such as the passage of a cold front *see* Chapter XVIII) a thunderstorm may develop at any hour of the day or night.

There are, in the main, two classes into which thunderstorms may be divided according to the cause whereby the instability of the atmosphere arises. These are :—

(i) Heat thunderstorms and coastal thunderstorms.

(ii) Frontal thunderstorms.

Heat Thunderstorms.—In summer the layers of air in contact with the ground become heated during the early part of the day and if the lapse-rate of temperature is large up to sufficiently great

heights, rapid convection results. The air thus rising from the lowest layers is replaced by air flowing in from all sides, accompanied by a gradual sinking of the atmosphere in the region surrounding the ascending current. The heavy rain falling from the cumulonimbus cloud often, however, owing to frictional drag, gives rise to a descending current of cold air below the cloud, the low temperature being maintained, despite the adiabatic warming due to descent, by evaporation of the falling rain. This cold current has a forward movement in the direction of advance of the storm. The ascending current, on the other hand, has a small movement in the opposite direction, through the air above it being drawn backwards into the advancing storm-cloud. This accounts for the sudden shift in the wind direction which takes place as the thunderstorm passes over, and is the explanation of the phenomenon of "thunder coming against the wind." Two shifts of wind therefore take place, one in front of the storm and the other just as the storm breaks.

Warm, moist conditions and rapidly ascending air currents are necessary for the development of these thunderstorms and consequently they occur most frequently in the summer-time and during the afternoon or evening of warm sultry days when the wind is light, over land areas from which sufficient moisture can be drawn. Storms of this type are seldom experienced over the ocean outside the tropics because the surface temperature does not rise high enough to create a large lapse-rate, neither do they occur over deserts, because there the air is too dry for the formation of cumulonimbus clouds. Over inland districts where there is sufficient moisture they are much more frequent than on the coast. It has been stated above that the necessary conditions of humidity and temperature are fulfilled in cols and shallow depressions. The rapid rise in temperature is in itself productive of a small local depression, but these depressions are very shallow, seldom extending far into the atmosphere, so that the case is entirely different from that of thunderstorms due to regular depressions which will be explained later on in this chapter. Heat thunderstorms are generally of a local nature and occur with great frequency over mountainous districts. Over some of the high mountainous islands in the tropics they are of almost daily occurrence. At Buitenzorg in Java, for example, thunder may be heard on 250 days in the year. The heating at the surface, of the moisture-laden air, causes the development of huge cumulonimbus clouds during the forenoon, the thunderstorm breaks during the afternoon, and towards evening as the convectional process dies down and the sky clears through the descent of air, little or no cloud is visible in the starlit sky at night. Owing to the abundance of moisture present in the atmosphere the clouds form at comparatively low levels, the mountain tops often being visible above the storm-cloud. In districts such as the Alps where the amount of moisture is small, the clouds form at great heights as the ascending currents must reach a greater altitude before condensation takes place.

Coastal Thunderstorms.—These may be regarded as a class of heat thunderstorms. They occur with considerable frequency in the western coastal districts of the British Isles and France, both in summer and winter, day and night, but more especially in winter. The explanation of these thunderstorms is to be found in the large lapse-rate associated with deep maritime polar air currents. The instability and moisture come from the heat and water vapour imparted by the relatively warm ocean to the lower layers of initially cold currents of air in the course of their passage across the Atlantic to the western European coasts. In such cases the final impulse or "trigger action" is caused by the elevation of the air stream over the land, especially in mountainous areas. Such storms are usually of slight intensity and dissipate rapidly on moving inland.

Frontal Thunderstorms.—It has already been seen that marked instability in the atmosphere, coupled with an adequate supply of moisture, are the conditions necessary for the development of thunderstorms. Such conditions are sometimes fulfilled when a cold current undercuts a warm current and causes it to ascend (page 350), *i.e.*, at a cold front in a depression. If then, condensation along the line of discontinuity between the two air masses is very active, thunderstorms may arise. The instability becomes very marked if the cold and the warm air currents on the two sides of the front differ considerably in temperature and if their directions are considerably inclined to each other. Cold-front thunderstorms, unlike local thunderstorms, sweep broadside over the country or, as sometimes happens if the front is moving end-on, they may last for several hours at one place. When they occur over the land they are most common in the afternoon or evening when the temperature of the lower layers of the atmosphere has been increased by insolational heating. They are generally accompanied by a violent squall of wind and a sharp fall of rain, hail or snow, always by an abrupt rise of pressure and fall of temperature. They are much more frequent on the coasts than inland and as the incidence of depressions is greater in winter than in summer they have their greatest frequency in the winter.

Thunderstorms are also associated with warm fronts (*see* Chapter XVIII), but they are never so severe nor so common as cold-front thunderstorms, for the ascensional motion of the warm air at a warm front is much more gradual than it is at a cold front. The air in the warm sector being generally of tropical origin, is warm and moist and is therefore well supplied with the potential energy necessary for convective activity. The impulse or "trigger action" is provided when the warm air suffers an upward deflection as it passes over the colder air mass at the warm front, so that the process may result in the development of thunderstorms. As the major activity which produces warm-front thunderstorms takes place at high altitudes these storms are not so violent as those of the cold-front

type: furthermore, the cumulonimbus cloud is masked by the warm-front cloud sheets. In common with cold-front thunderstorms they have their greatest frequency in the winter and may occur with equal regularity by day or night.

Occluded fronts both of the cold and warm type are also productive of thunderstorms, but they appear to occur more frequently with the former. It is a common occurrence for occlusions even of comparatively feeble intensity to develop into surprisingly active cold fronts, especially when passing from sea to land, and to exhibit the characteristic features of these phenomena. In the British Isles such thunderstorms occur with remarkable frequency in summer.

Over the ocean, except in the regions of the doldrums, storms of the frontal type only are experienced. In high latitudes, depression thunderstorms are rare because the quantity of moisture in the air is small.

Thunderstorms of all types have the same fundamental characteristics. Condensation takes place in the ascending current as soon as the temperature of the ascending air is reduced to the dew-point. The drops of water thus formed do not immediately fall out but are carried upward by the rising air. Small drops of water fall very slowly through still air and can therefore be carried upward by a relatively slight ascending current. The largest drops (about 5·5 millimetres in diameter) which can persist in the air can be supported by an upward current of about 25 feet per second. When larger drops than these are formed they are unstable and immediately break up into smaller drops. In a thunderstorm the ascending current is not steady, but blows in a succession of gusts and lulls, so that the drops of water in the rain cloud rise and fall intermittently, sometimes coalescing into larger drops, and then breaking up again into smaller drops. The drops which get to the edge of the ascending current, or which reach the head of the current, and spread out horizontally, fall to the ground, giving the heavy rain of the early passage of the thunderstorm. The rain, however, does not fall continuously during thunderstorms but generally in very heavy showers. Frequently after almost ceasing, the rain renews itself with great violence after a lightning flash, large drops falling at first, followed by smaller drops. The velocity of a drop of water increases with its size and, therefore, although both large and small drops are formed simultaneously, the larger drops naturally reach the earth first. The rainfall in a thunderstorm is therefore distributed in an entirely different way from that associated with a depression. Instead of the rain commencing with small drops, and the drops increasing in size towards the centre of the storm, and then decreasing again, the heaviest drops in a thunderstorm fall at the commencement and the size decreases towards the end. The frequent renewal of very heavy rain-drops during the passage of the storm is a phenomenon peculiar to thunderstorms.

It sometimes happens that exceptionally heavy rain falls from a thundercloud especially in hilly and mountainous districts. As the storm moves over a mountain range the supply of warm moist air from below is cut off; convectional activity suddenly ceases and the mass of water and hail within the cloud is precipitated in a heavy downpour. Such a phenomenon is known as a **cloud-burst.**

Thunder and Lightning.—As a result of the violent convective movements of air which take place within a thundercloud, charges of positive and negative electricity accumulate in different parts of the cloud and the potential difference between them ultimately becomes so great that an electrical discharge takes place. The rapid passage of electricity through the air causes the latter to become heated and rendered visible, as in the case of an electric spark. Hence the visible phenomenon of discharge is known as a lightning flash. The flashes are popularly classified into two kinds, "forked lightning" and "sheet lightning," but physically there is no difference between the two. When forked lightning is seen, the discharge is viewed directly; in the case of sheet lightning the discharge is obscured by clouds, which partly reflect the flash and also become diffusely illuminated somewhat in the same way as a white sheet is illuminated when held up in front of, and at some distance from, a flashing light. The intense heating of the air in the path of the flash causes it to expand very rapidly, almost with explosive violence. The sound waves thus generated travel out in all directions and are heard as thunder. Since light travels almost instantaneously, while sound travels at about 1,100 feet (approx. one-fifth mile) per second, an appreciable interval of time elapses between seeing the lightning flash and hearing the thunder. The long continuance of the thunder is explained by the fact that, the lightning flash being of considerable length, the sound from its different parts does not reach the observer at the same time, but is spread over an interval of several seconds' duration. In mountainous countries the sound is still further prolonged by the numerous echoes which are set up.

Various theories have been advanced to account for the separation of electrical charges within a thundercloud and the most satisfactory (because it is supported by observations made within the cloud) is that recently put forward by Sir George Simpson and F. J. Scrase. The work of these investigators is described in full in a paper entitled "The distribution of electricity in thunderclouds," Proceedings of the Royal Society of London, Series A. No: 906, Vol. 161, pp. 309–352, August, 1937. The experimental method adopted and the conclusions reached will be briefly described here.

When two electrical charges of opposite sign are situated in two parts of a cloud, there is between them an electric field, or potential gradient. The potential increases in the direction of the positive charge. If, in particular, the charges are arranged vertically, the potential will increase in the upward direction, in a region between a lower negative charge and an upper positive charge. In such a case

the potential gradient is said to be positive. If, on the other hand, the negative charge lies uppermost, the potential gradient between the two regions is negative.

Simpson and Scrase devised an experimental method of determining the sign of the potential gradient, and hence the vertical arrangement of the charges. The method depends upon the fact that when a long pointed conductor is placed in an electric field, so that it lies parallel to the lines of force, an electric current flows along the conductor. The direction of the current is such that positive electricity flows away from the end directed towards the positive charge, *i.e.*, downwards when the potential gradient is positive, upwards when it is negative. To determine the sign of the potential gradient, therefore, it is necessary only to determine the direction in which the current flows. In the apparatus devised by Simpson and Scrase the conductor took the form of a long wire which was carriedinto the atmosphere by means of a hydrogen-filled balloon. A small lead weight on the end of the wire ensured that it hung vertically. To determine the direction in which current flowed in the wire, an instrument known as an alti-electrograph was used. This instrument consisted essentially of a disc rotated by clockwork, on which was mounted a circular piece of pole-finding paper, that is, paper impregnated with a chemical solution. Hence, when an electric current was passed through it a discoloration was produced at the electrode at which the current entered, thus determining the direction of its flow. Set to track on the paper were two iron pins, one of which was electrically connected to the wire hanging below the instrument. The other pin was connected to short wires projected upwards from the light bamboo framework in which the instrument was enclosed for protection. The two pins were set to describe circles of different radii on the chart, and a stain was produced on the paper at one or other of them, according to the direction of flow of the current. If the stain appeared at the outer circle the potential gradient was negative; if at the inner circle, positive. Mounted on the same axis as the disc and rotating with it was a second disc on which were traced records of pressure, temperature and humidity, the principle being similar to that of the Dines Meteorograph (Chapter IV). Since the discs rotated at the same rate the heights corresponding to different parts of the electrical record could be found, utilizing the pressure and temperature records. In all, 70 soundings were made, 51 of them in thundery or showery conditions, and 29 legible records were subsequently recovered.

The results showed that, in general, the distribution of electrical charges in a thundercloud is a positive charge in the upper part of the cloud, a negative charge in the lower part, with a local concentration of positive charges in the base of the cloud usually just in front of the heavy rain centre, but frequently in other parts of the base. The upper positive charge occurs in a region where the temperature is

well below the freezing point and the cloud particles are almost certainly ice crystals; near the base of the cloud, however, the temperature is above freezing point and the lower positive charge is therefore associated with water drops. This distribution of charges is shown in Fig. 95. The shaded area represents a vertical section of

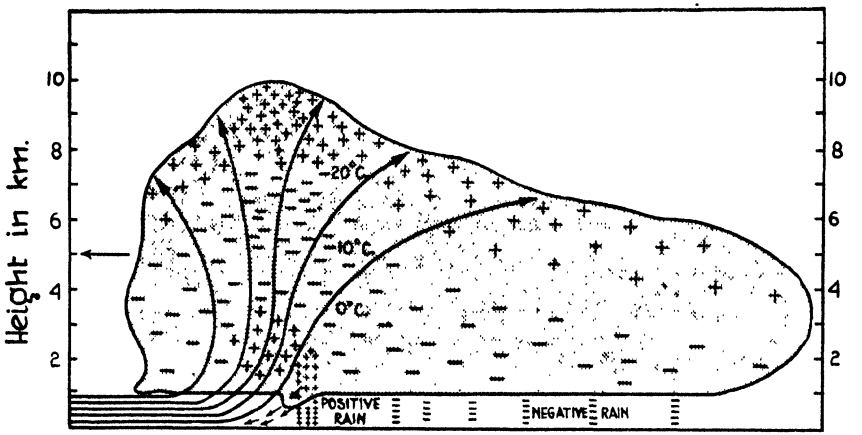


FIG. 95.—DISTRIBUTION OF ELECTRICAL CHARGES IN A THUNDERCLOUD.

the cloud and the distribution of charges is shown by the positive and negative signs. The continuous lines terminated by arrow-heads are the lines of air-flow; the closer these lines the stronger is the current of air.

There is no single physical process which will completely account for the observed distribution of charges. As regards the lower positive charge the mechanism of its production is satisfactorily described by Simpson's breaking-drop theory. According to this theory, a rain drop in a rapid current of air is broken up into smaller drops and in the process electrical separation takes place, the droplets acquiring a positive charge and the air a negative charge. It was mentioned on page 279 that if the speed of the rising current exceeds 25 feet per second, which is the speed of fall of the largest rain drops, all the condensed water will be retained in the cloud. There is therefore a concentration of water drops above the rapidly ascending current, which has its greatest speed near the base of the cloud. The negatively charged air will be carried away to the upper parts of the cloud; the positive charges will remain more or less concentrated near the base of the cloud, near to the rising current. The separation of charges by the breaking-drop process has been confirmed by laboratory experiments. No experimental confirmation has however been obtained for the explanation suggested by Simpson of the upper positive charge. From his observations in the Antarctic (1911–12) during blizzards, Simpson suggested that the impact of ice crystals causes them to become negatively charged

and the air to become positively charged. Now the upper part of a thundercloud consists of ice crystals in a state of violent movement owing to the convection, and in this way these may acquire negative charges. At the same time they are sinking through the air, so that there is a movement of negative charges downwards while positively charged air remains at the top of the cloud.

The charge on the falling rain depends upon the part of the cloud in which it originates. Hence, as in Fig. 95 positively charged rain is expected in the centre of the storm in the vicinity of the upward current, with negatively charged rain in the outer parts of the storm. Experimental evidence, on the whole, gives confirmation of this.

Line-Squalls.

It has already been seen in the discussion on thunderstorms that these storms may be associated with the passage of a cold front. There is another phenomenon known as the **line-squall,** which may be regarded as simply a well marked cold front, but which under favourable conditions gives a long roll of cloud extending along the front and a very strong squall of wind at the onset of the front. The passage of a line-squall is marked by characteristic happenings which include all or some of the following :—

(i) A line of low black cloud, often appearing as an arch.
(ii) A sudden or rapid rise of wind speed.
(iii) A sudden or rapid change of wind direction.
(iv) A sudden or rapid drop in temperature.
(v) A rapid rise in barometric pressure.
(vi) Heavy rain or hail (occasionally snow).
(vii) Thunder and lightning.

It is the first and second of these characteristic happenings that, combined, have given the phenomenon the name line-squall. Many thunderstorms occurring over Europe are of the line-squall type. The storm front makes its appearance in the west or north-west with threatening clouds and, at times, the long roll-like cloud, and beyond this a peculiar light grey uniform sky. The dark low band of cloud passes overhead and heavy rain commences as the light grey cloud comes along. The wind, which has fallen light from the south or south-west as the clouds begin to bank up, comes in a violent squall from the west or north-west about the time when the dark roll-cloud passes overhead, and falls again after the commencement of the heavy rain. The violent motions peculiar to the line-squall are limited to the first few thousand feet, while the intense convectional activity which leads to the development of thunderstorms takes place above this level.

The horizontal length of a line-squall in a characteristic case is well represented by that of a cold front, *i.e.*, anything up to a thousand miles, while its length of life may be twenty-four hours

or more. Similarly, its rate of advance can be predicted with considerable certainty for in active depressions any given part of a cold front advances with a speed which is approximately equal to the component, perpendicular to the front, of the gradient wind in the cold air in its rear. As regards the geographical distribution of line-squalls, cold fronts associated with depressions of the temperate regions may penetrate into quite low latitudes frequently extending within the tropical zone. In the United States line-squalls of great intensity are experienced over the central plains, the warm air in front being drawn from the Gulf of Mexico and so, being very humid, gives rises to heavy precipitation and thunderstorms along the squall fronts where the wind changes to north-west. The Mediterranean often experiences severe line-squalls, especially in winter, with cold air from higher latitudes pouring through the gaps in the mountain chains of southern Europe. On the southern shores of the Mediterranean, cold fronts are frequently rainless, for the warm current which, in general, provides the moisture, being drawn from the Sahara, is exceedingly dry. In this case the front may even pass without any clouds forming when it is termed a "white squall." The cold-weather storms of the Persian Gulf and NW. India are depressions associated with the penetration of cold air from the north and line-squall phenomena may be manifested in them. In the southern hemisphere, cold fronts occur in association with the eastward moving depressions of the Roaring Forties. Here again the penetration of polar currents into quite low latitudes may cause severe line-squalls there. Notable examples of such storms are the Southerly Buster of summer at Sydney, Australia, the Pampero of the Argentine and Uruguay and the line-squalls of South Africa. It has already been pointed out that the penetration of cold air from higher latitudes may extend to within the tropics causing similar effects in quite low latitudes. There are two other ways in which relatively warm and cold air masses of sufficient extent may come together and produce line-squall phenomena within the tropics. The first of these is concerned with the sea breeze in coastal districts. Under certain conditions the cool sea breeze does not arrive inland gradually but suddenly, as a cold front, and its arrival may be accompanied by thunderstorms. This effect may also occur in a marked way in the lower latitudes of the temperate zones but in higher latitudes it is less and less evident. The necessary general conditions are realized in lower Egypt and records have shown that at Abu Sueir near Ismailia during the hot part of the year, a cold front arrives practically every afternoon. It is usually a very slight affair, and in view of the extreme dryness of the air, is generally a white squall. The other way is the reverse case when air chilled by radiation at night on high land near the sea flows down and impinges on the warm air still lying over the sea. In the tropics the contrast of temperature between the two air masses may be considerable

and in the case of a long mountain range the resulting squalls may occur along a sufficiently long front to be regarded as line squalls. The Sumatras of the Malacca Strait which always occur at night, are probably of this character.

The Structure of a Line-Squall.—Some understanding of the phenomena of line-squalls may be gathered by reference to Fig. 96.

The horizontal and vertical scales in the figure are the same, so that a correct idea of the proportions is at once given. The stippled area is a section through the tip of the cold wedge which is supposed to be advancing from left to right. Owing to surface friction, the cold air is found further advanced some distance above the surface itself, and the upper limit of the cold air, *A B*, instead of continuing on until it cuts the surface, is doubled back and meets it at the point *C*. This point is to be considered as marking the cold front at the surface. The rate of advance of this point to the right in the figure, or in other words, the speed of the cold air at *C* in this direction, is the rate of advance of the front. The moment when the barometer begins to rise at the surface is when the point *B* comes overhead, and as the vertical thickness of the cold wedge of denser air increases, so the barometer continues to rise. Beyond the left of the figure the upper boundary of the cold air slopes less rapidly upwards and so the rise of pressure proceeds less rapidly. The arrow at *D* indicates that the air there is being overtaken by the cold wedge. When this air arrives at *E* it immediately comes under the influence of the strong pressure gradient and its relative motion is checked or even reversed. It is therefore on the passage of this point that the first sudden shift of wind takes place at a station ; this point also marks the commencement of the squall. The squall continues until a little after the passage of the point *C*, when a decrease of wind takes place. In a typical case, the passage of the whole portion of the cold wedge shown in the figure would only be a matter of ten minutes or so. The strong horizontal convergence created between *D* and *E* produces a strong upward current at *F*. If there is sufficient water vapour present, cloud will form here, and owing to the tendency of the cold air at *B* to fall, a rolling motion as indicated may be imparted to this cloud. This is the characteristic long roll cloud of a line-squall which is seen in cross section in the figure. The advance of the cold wedge having caused the initial rapid rise of the warm air at *F* and the formation of the roll cloud, further ascent takes place, gradually, along the surface *A B* with the formation of a light grey cloud sheet *N*. But the ascent of the warm air is not necessarily limited to the surface *A B*. For, if the warm air in the layer *J K* is unstable with respect to saturated air, *i.e.*, if the lapse-rate of temperature is in excess of the saturation adiabatic lapse-rate, then the warm air at *F* having started to rise vertically will continue to do so under its own buoyancy and a towering cumulus cloud will form in the space *M L* with the possibility of thunder, lightning and

heavy rain or hail. When this occurs, the roll cloud at *F* is frequently obscured by the massive clouds thus formed.

It should be borne in mind, therefore, that essentially the same conditions may produce the rain and squally winds of the passage of a cold front, the line-squall, the cold-front type of thunderstorm and, in addition, tornadoes and waterspouts.

Tornadoes.

A tornado is a violent atmospheric whirl characterized by a peculiar funnel-shaped cloud pendant from heavy cumulonimbus, and accompanied by heavy rain or hail, thunder and lightning and a loud rumbling sound. The direction of the whirl is counter-clockwise in the northern, clockwise in the southern hemisphere. Its diameter varies between about 50 feet and 1 mile, the average being 1,000 feet. The rate of travel is generally between 20 and 35 knots; the direction, NE. in the northern and SE. in the southern hemisphere. Since the tornado never exists for more than a few hours at the most the area affected is a comparatively small and narrow belt. The damage wrought by a tornado is nevertheless serious owing to the high wind speeds, estimated to exceed 200 knots, which occur within it. It appears also that there is a considerable reduction of pressure within the whirl, estimates varying from 100 to 500 millibars. Such a reduction of pressure over a short distance involves enormous pressure-gradients which can only be balanced by the centrifugal force on air moving rapidly in paths of small curvature; hence the large wind speeds. Buildings over which tornadoes pass, sometimes burst outwards with explosive violence owing to the sudden reduction of pressure outside them. Barograph records from near the path of a tornado show a sudden fall of pressure followed by an immediate recovery, indicated by an elongated V-shaped trace, the two sides of the V being almost indistinguishable. The whirl is rendered visible by the water vapour condensed as a result of the reduction of pressure within it. The upward currents within a tornado are exceedingly violent, and some idea of their speed may be afforded by the fact that heavy objects, timbers, roofs of buildings and farm animals have been lifted up and transported to distances of several hundreds of yards. Trees are uprooted and buildings destroyed and the tornado is, in fact, the most violent and destructive of all meteorological phenomena.

Tornadoes occur chiefly in Australia and the central plains of the U.S.A., in both of which areas their average number is 50 a year. In the latter region they are known also as Twisters. In Europe they are occasionally experienced but there they never attain an excessive degree of violence. They are most frequent in spring and summer in the late afternoon. Violent convection, such as occurs at a well marked cold front or boundary between two air masses of widely differing temperatures, is essential to their formation. They almost invariably occur on the trough line of a V-shaped

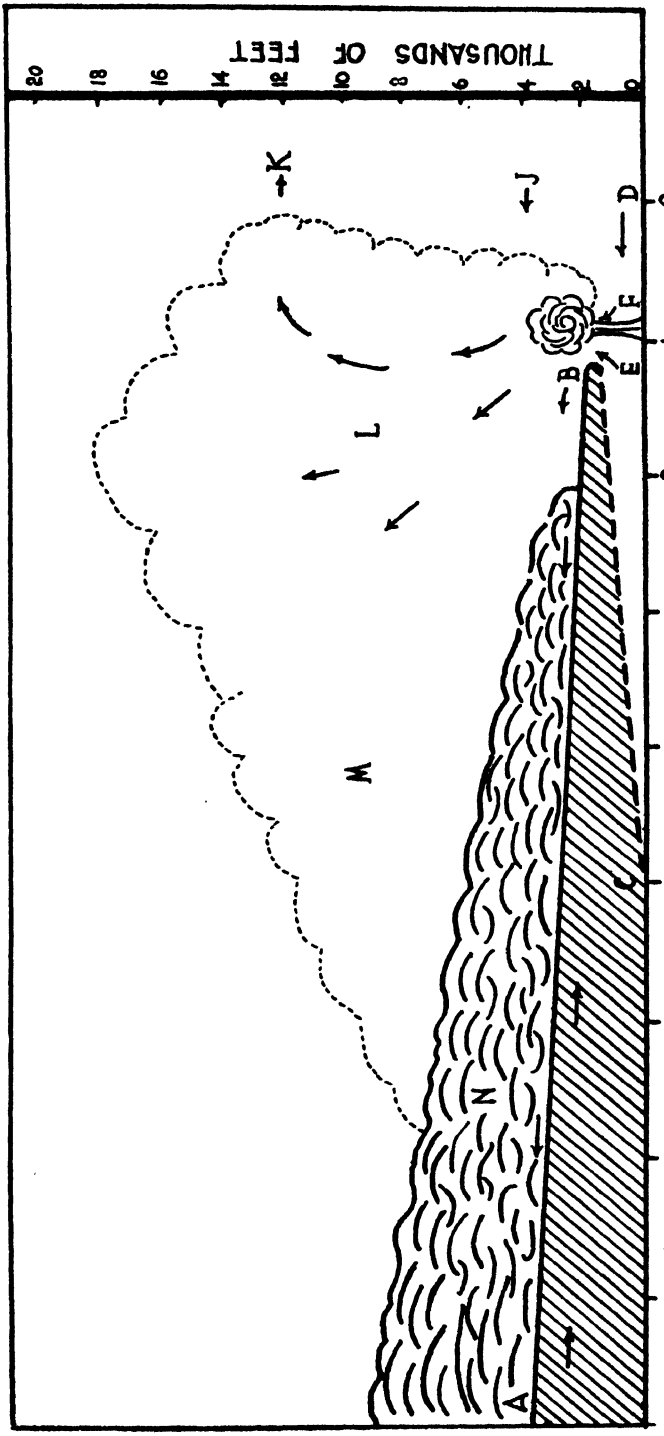


FIG. 96.—DETAILED STRUCTURE OF A LINE-SQUALL.

depression, *i.e.*, at a cold front. The air in advance of the trough is usually warm and very humid, characterized by a feeling of extreme closeness, while in the rear the air is much colder. The central U.S.A. is especially subject to tornadoes since the absence of an east to west mountain barrier permits the warm humid Gulf air to come in contact with the air moving southwards from northern Canada. The whirl appears to be the result of air rushing in horizontally to fill the place of that carried upwards by convection.

The name tornado is also given to the squall which frequently accompanies a thunderstorm in West Africa. This, however, is an entirely different phenomenon from the tornado proper of the U.S.A. and Australia.

Waterspouts and Whirlwinds.—The waterspout is the ocean counterpart of the tornado, but is usually much less violent. It occurs with cumulonimbus cloud from which a pendant funnel-shaped cloud appears and makes its way down towards the sea beneath. Shortly before the spout reaches the sea surface the water is thrown into violent commotion and is drawn up several feet, while the spray is drawn considerably higher appearing to join up with the spout, and a continuous writhing vortex between sea and cloud is formed. Some waterspouts do not reach the sea surface but appear to remain suspended from the clouds. The waterspout is of smaller extent and less violent than the tornado, the diameter of the whirl, judged by eye, appearing to be about 20 to 30 feet. Its effects are very local but the wind is often sufficiently violent to cause a rough and confused but not high sea. Lasting from 10 minutes to half an hour the waterspout usually becomes attenuated owing to the differing speeds of its base and upper parts. It finally breaks at about one-third of its height from the base and then quickly disappears. Like the tornado it is the result of convective instability, usually brought about by the juxtaposition of warm and cold air masses, but not necessarily at a front. Contrasts of temperature sufficient to form waterspouts may be produced within a uniform air mass through the cooling of air due to thunderstorm precipitation.

CHAPTER XIII.

Optical Phenomena

There are a great many optical phenomena which not only arrest the attention on account of their beauty, but also are more or less closely associated with the weather.

The following is a list and also a description of the more important optical and associated phenomena which may be observed in certain circumstances from ships at sea, and from aircraft :—

Aurora.—In Latin, the word "aurora" means "dawn." According to the Oxford dictionary the term "aurora borealis" was introduced by Gassendi in 1621. This term, the northern dawn, is apt enough for a faint glow seen at night near the northern horizon. The same name, however, is given to a wider class of luminous phenomena. On the recognition of similar phenomena in the Antarctic regions, these were called "aurora australis." The single word aurora is now in general use in both hemispheres.

Aurora takes many forms, the two most important being that of an arc, and that of a curtain. The quiet arc is the most symmetrical and stable form of aurora, sometimes persisting with little visible variation for hours ; frequently, however, streamers of light shoot out radially from the arc and the whole structure undergoes rapid changes. The lower edge, both of arcs and curtains, is usually much the better defined. Several forms may occur simultaneously or successively in any auroral display.

Aurora is very rare in southern Europe, and is but seldom seen in the south of England. But the frequency increases rapidly with latitude, and in the Orkneys and Shetlands aurora is a common phenomenon. In these regions it may occasionally be observed to the south as well as to the north. A zone of maximum frequency passes across the north of Norway, the south of Iceland and Greenland and surrounds the north magnetic pole ; the distribution in the Antarctic is more or less similar. Cape Evans, the headquarters of Scott's last expedition, was within the auroral zone ; aurora was seen frequently to east-north-east and comparatively rarely overhead. It is difficult to assign laws for the frequency of occurrence of aurora as weak displays are obscured by moonlight or twilight ; also, in daylight or on nights when the sky is heavily clouded observations of aurora are impossible. In the British Isles and similar northern latitudes it is most common in the late evenings and near the equinoxes ; but in the north of Norway and Greenland it appears to be most frequent near mid-winter.

Aurora is undoubtedly an electrical discharge, and when visible in England, is nearly always accompanied by a magnetic storm, but this is not the case when it is confined to high latitudes.

Halo Phenomena.—The term halo is used to denote a circle of light round a luminous body, such as the sun or moon, and is formed by the reflection and refraction of light by ice crystals. Many different kinds of halo have been observed; the most common is a luminous ring of about 22° radius surrounding the sun or moon, the space within the ring appearing less bright than that just outside. The ring, if faint, is white; if more strongly developed the inner edge is a faint red, outside which, yellow may be detected. The angle of 22° is the angle of minimum deviation for light passing through a prism of ice (refractive index 1·31) with faces inclined at 60°. Thus the occurrence of the halo of 22° radius indicates the presence of ice crystals with faces inclined at 60°. Alternate faces of an hexagonal prism are inclined at this angle, and as hexagonal prisms are frequently found amongst the ice crystals of which cirrus clouds are composed, the halo is probably due to the refraction of light through such prisms.

A halo of about twice the radius, *i.e.*, a halo of 46°, occurs more rarely; its luminosity is usually much less than that of the halo of 22°, while the arrangement of colours, if visible, is the same. This halo requires crystals with faces inclined at right angles.

The halo of 22° is sometimes within a circumscribed nearly elliptical halo, the points of contact being at the upper and lower points. The complete circumscribed halo is only seen when the elevation of the sun is 40° or more. With lower elevations, separate arcs of contact are seen, although the arc of lower contact is very rare. Contact arcs appear occasionally at the sides of the halos of 22° and 46°, but they are as rare as the arcs of lower contact. The ends of the arc of upper contact of the halo of 22° are frequently bent upwards.

All the optical phenomena produced by reflection and refraction of light by ice crystals are sometimes grouped together as halo phenomena.

Mock Sun Ring.—Occasionally a colourless white ring, which passes through the sun parallel to the horizon, may be recognized. This is called "parhelic circle" or "mock sun ring." The latter name has been given to it because the mock suns described below lie on or near it.

Parhelion.—The term parhelion or "mock sun" is used to describe the luminous or even brilliant image of the sun seen most frequently at or near the intersection of the halo of 22° with the mock sun ring. The mock suns of this halo are occasionally brilliant in colouring, with red on the side nearest the sun, and yellow, green and blue following in order. Frequently mock suns are seen without any of the rings being observed.

A white mock sun is sometimes to be seen immediately opposite to, and at the same elevation as, the sun; this phenomenon is known as an "anthelion" or "counter sun." A similar image

may also be observed at the same elevation as the sun and in azimuth about 120° or 140° from it; this image is called a "paranthelion." The bright patches seen at the points where the halo of 22° meets the arcs of contact (*see* above), are sometimes referred to as mock suns. The term "sun dogs" is popularly applied to all mock suns.

Paraselenæ.—Paraselanæ or "mock moons" have been observed occasionally; they are analogous to mock suns.

Sun Pillars.—A sun pillar is a vertical column of light above (and sometimes below) the sun, most easily observed during cold weather at sunrise or sunset. Sun pillars frequently extend about 20° above the sun, and generally end in a point. At sunset, they may be entirely red, but as a rule, they are white, and show a marked glittering. The phenomenon, which is due to the reflection of sunlight from small snow-crystals, may be seen over a wide area.

In addition to the halo phenomena described above, many others are known to occur; they are less common than those mentioned above, and will not be discussed here.

Coronæ.—Coronæ are seen most frequently round the moon. As their diameter is generally considerably less than that of the halo of 22° they are very near the luminary, and can thus only be seen around the sun under favourable circumstances. No doubt solar coronæ occur as frequently as lunar coronæ; the former may be observed by making use of a smoked glass to reduce the intensity of the light.

Coronæ are very different from halos. The latter are produced by refraction due to ice-crystals, whereas the former are due to the diffraction of light by water-drops. The positions and orders of the colours serve to distinguish the two sets of phenomena; coronæ usually show a brownish-red inner ring, which, together with the bluish-white inner field between the ring and the luminary, forms the so-called "aureole." Frequently the aureole alone is visible. The brownish-red ring is characteristically different from the red ring of a halo; the former is distinctly brownish, especially when the aureole alone is visible, and of considerable width, whereas the latter is beautifully red and much narrower. If other colours are distinguishable, they follow the brownish-red of the aureole in the order from violet to red, whereas the red in a halo is followed by orange, yellow and green; the order of the colours is thus reversed. As coronæ are diffraction phenomena they occasionally show the sequence of colour two or three or even four times over; this can never be the case with a halo.

Bishop's Ring.—A dull reddish-brown ring which is seen at certain periods round the sun in a clear sky. In the middle of the day the inner radius of the ring is about 10°, the outer 20°, but when the sun is low the ring becomes larger, the

brightest part being about 19° from the sun. After sunset the ring is lost in the warm colours of the sky. Bishop's ring was first observed after the great eruption of Krakatoa in 1883 and remained visible till the spring of 1886, and was no doubt due to minute particles shot out by the eruption and remaining in the atmosphere at great heights. Bishop's ring was seen again after the eruptions of the Souffrière in St. Vincent and Mount Pelée, in Martinique, in 1902. Like the brownish ring surrounding the aureole observed in clouds, the phenomenon is explained by diffraction.

Brocken Spectre.—It sometimes happens that an observer in an aircraft flying above the clouds sees the shadow of the aircraft, on the cloud below, surrounded by coloured rings. The whole appearance has been called "Brocken spectre," and the rings are usually known as a "glory." The colours are not caused by the shadow but are due to light diffracted backwards in the same way as the corona is due to light diffracted forwards. Outside the coloured rings, and somewhat removed from them, there frequently appears at the same time another ring of pure white, known as "Bouguer's halo" or "Ulloa's ring."

Iridiscence or Irisation.—Green and red colours are occasionally seen on the edges of cirrocumulus and altocumulus clouds, at a distance up to 25° or more from the sun or moon; they are also seen at times on the edges of fractocumulus or stratocumulus clouds. The colours occur in bands following the outlines of the cloud. They are due to diffraction by water-drops and may therefore be regarded as portions of coronæ. The arrangement of the colours is due to the regular gradual variation in the size of the drops, the larger drops occurring in the central parts and the smaller drops at the edges of the cloud.

Blue of the Sky.—The blue of the sky is due mainly to the scattering of sunlight by the individual molecules constituting the air. It was demonstrated mathematically by the fifth Lord Rayleigh that if light of different wave-lengths were to pass through a medium containing very small particles the proportion to be scattered would be greater the shorter the wave-length of the light. Thus the short waves composing the blue and violet end of the spectrum are scattered more than the long red and yellow waves. Hence light passing through a medium containing a great number of such particles is left with an excess of red, whilst light emerging laterally has an excess of blue. It is for this reason that soapy water looks yellowish when one looks through it at a source of white light and bluish when one looks across the direction of illumination.

That a column of air viewed across the direction of illumination also appears blue was demonstrated in the laboratory by the present Lord Rayleigh. Accordingly there is no need to assume the existence of grosser particles suspended in the air to explain the normal blue

of the sky. Where large particles due to smoke or dust are present the light scattered by them does not contain so high a proportion of blue, in other words the blue of the sky is diluted with white. The sky as seen from high altitudes is of a deeper but purer blue because there are fewer large particles in suspension than at lower altitudes.

Rainbows.—Rainbows are due to the refraction and reflection of sunlight (occasionally moonlight) in falling drops of rain. They are circular arcs of coloured light centred at the anti-solar point of the celestial sphere. The normal appearance of a bright well-developed rainbow is as follows: the chief or primary bow shows the sequence of colours, violet, blue, green, yellow, orange and red, the red being on the outside or top of the bow. Closely inside this bow are one or two supernumerary bows with the same colours in the same order, the first inner bow being much fainter than the primary bow and the second fainter still. In cases of exceptionally brilliant rainbows more than two supernumerary bows may be seen. Outside the primary bow and completely detached from it is the secondary rainbow in which the colours appear in the reverse order, red inside and violet at the top or outside. The primary bow is formed by means of the reflection of sunlight from the far side of each raindrop. The light which is reflected in this way does not come out in all directions but only in directions lying within about 42° from the direction of the sun. The reflected light is most intense near the limit. Accordingly an observer looking towards the raindrops receives a certain amount of light from all directions within 42° from the shadow of his head, but most light along rays which make an angle of about 42° with the central line. The limiting angle depends on the colour of the light, and in so much as white light is compounded of light of different spectral colours, the observer sees the concentric arches of different colours referred to above. Some of the light falling on a drop does not emerge until after it has been reflected twice; none of the twice reflected light which reaches an observer makes an angle of less than 50° with the line to the centre of his shadow. This is the explanation of the secondary bow. In very exceptional cases more than one secondary bow have been observed. From the above reasoning it will be seen that the radius of the arc of the primary rainbow is about 42°; that of the secondary, about 50°. It follows therefore that if the sun's altitude exceeds 42° no primary rainbow can be formed; consequently rainbows are mainly morning and evening phenomena.

Lunar rainbows are formed in precisely the same circumstances as solar bows, but are considerably rarer having regard to the comparatively short periods that a bright moon is above the horizon. In favourable circumstances a lunar rainbow is so much fainter than a solar one that colour is difficult to distinguish and the normal appearance of a lunar rainbow is whitish.

On rare occasions a white rainbow is seen opposite the sun in fog; this phenomenon is known as a "fog bow." Its outer margin has a reddish and its inner a bluish tinge but the middle of the band is quite white. A supernumerary bow is sometimes seen inside the primary bow and with the colours reversed. The bows are produced in the same way as the ordinary rainbow but owing to the smallness of the drops, the colours overlap and the bow appears white.

Mirage.—The term mirage is used for certain appearances produced by the refraction of light by the atmosphere; these appearances include the illusion of a sheet of water in the desert as well as cases in which definite distant objects can be seen in duplicate.

Over a heated desert the air very close to the ground is less dense than that above it so that the velocity of light is greatest close to the ground. Rays coming down from the sky at a gentle inclination may be bent up again to the observer, to whom the rays appear to come from a bright water surface. The banks, reeds, etc., are the images of various objects distorted by being viewed through layers of air of varying density so that a dark stone appears as though it were an upright stake. Hills situated a short distance away may appear as detached masses floating on this lake-like surface, their lower portions being invisible under the conditions prevailing. Mirages may often be seen over smooth road surfaces on calm hot days in England, especially over tarred roads.

These are all cases of inferior mirage. "Superior" mirage in which the light rays are bent downwards from a warm stratum of air resting on a colder is seen most frequently in polar regions. The distances involved are greater so that the details can hardly be observed without telescopic aid. With such assistance a distant ship may sometimes be seen in triplicate, one of the images being inverted. As the stratification which produces superior mirage is stable the images are clear and well defined in contrast to the shimmering images of the inferior mirage.

The images produced by superior mirage are not always exactly in the same vertical. The separation is explained by a slight departure of the strata of equal density from horizontal planes. Lateral refraction of this kind is to be distinguished from the phenomenon which can be observed sometimes near a heated wall when objects appear to be reflected in the wall.

Fata Morgana.—This is a complicated form of mirage in which multiple images of an object are produced; these images are frequently elongated. It is produced by several layers of air of different refractive indices.

Sunrise and Sunset Colours.—The general explanation of the variety of colours that are to be seen in the sky about the time of sunrise or sunset is as follows. White light such as that from

the sun may be regarded as composite, the constituents being light of all the colours of the spectrum. When the light waves meet obstacles in their course, such obstacles as the molecules of the atmospheric gases or larger obstacles such as particles of dust, the waves are broken and secondary waves proceed in all directions from the obstacles. The direct light is therefore reduced in strength and the further the light goes through an atmosphere of such obstacles the more the strength is reduced, the energy being used up in producing scattered light. This effect is more pronounced with blue light for which the wave-length is short than with red light for which the wave-length is longer. Accordingly, a beam of white light passing through air loses the constituents of shorter wave-length and becomes yellow, then orange and finally red.

This accounts for the changing colour of the sun as it nears the horizon. The scattering by air alone merely makes the setting sun yellow but if there is dust in the air or even the nuclei on which water vapour is condensed then the sun becomes orange or red before it sets.

Clouds which are illuminated by the light from the setting sun are also red whilst other clouds which are illuminated by scattered light in which the blue constituents are present are white or grey; higher clouds illuminated by light which has only passed through less dense and cleaner air may also appear white.

Twilight Arch.—This term has been used with more than one meaning. Probably the best practice is to confine it to the boundary between the illuminated segment and the dark sky above as seen in the later stages of twilight. In Von Bezold's nomenclature several twilight arches are recognized; the first western twilight arch is the boundary seen even before sunset between the extended tinted band or segment of light along the horizon and the blue sky above; this arch persists through the development and disappearance of the "purple light," and may be visible together with the second western or true twilight arch

Purple Light.—Shortly after the sun has set below the western horizon a brighter patch appears on the darkening sky about 25° directly above the position where the sun has disappeared. This patch appears brighter as the sky darkens and takes on a purple tone. The patch expands into a disc and when the sun is about 4° below the horizon it reaches its maximum brilliancy, when it may be so bright that white buildings in the east which are lit up by it, glow with a purple colour which corresponds to the afterglow seen on the peaks of snow-covered mountains. The disc of purple light sinks downwards at twice the rate at which the sun sinks while at the same time its radius expands and its light becomes less intense. It finally sets behind the bright segment of the twilight arch. Occasionally when the first purple light has passed below the horizon, the phenomenon repeats itself with less intensity.

The second patch of light appears at a slightly less altitude than the first but otherwise follows the same course.

Crepuscular Rays.—Occasionally soon after sunset the sky is divided up into lighter and darker areas by lines which diverge from the position (below the horizon) of the sun. The lighter areas indicate parts of the atmosphere illuminated by sunshine, the darker areas those from which the sunshine is cut off by intervening mountains or by clouds. The phenomenon is essentially the same as that seen in the daytime when the sun is hidden by cloud and " ladders " diverging from the sun appear where there are gaps in the cloud. The " crepuscular rays " are coloured, the illuminated areas being pink and the shadows appearing greenish by contrast. The divergence of the " crepuscular rays " is an optical illusion as light rays from the sun are practically parallel.

Under favourable circumstances the shadows are thrown right across the sky and to an observer with his back to the sun the rays appear to converge to a point which is a little above the horizon. These anti-crepuscular rays are generally ill-defined and may be mistaken for patches of cloud.

Zodiacal Light.—A cone of faint light in the sky, which is seen stretching along the zodiac from the western horizon after the twilight of sunset has faded, and from the eastern horizon before the twilight of sunrise has begun. In our latitudes it is best seen from January to March after sunset, and in the autumn before sunrise. In the tropics it is seen at all seasons in the absence of moonlight. The light is usually fainter than that of the Milky Way. The luminosity is probably due to the scattering of sunlight by an exceedingly rare gas which is rotating round the sun like an extended atmosphere.

Green Flash.—The green flash is a phenomenon which is observed fairly often at sunset, more rarely (because seldom looked for) at sunrise. At sunset the last glimpse of the sun which is seen is a brilliant green. The colour lasts but two or three seconds. The explanation is that the rays from the sun are refracted in their passage through the atmosphere, the blue and green more than the yellow and red so that when the sun is so low that no red light from it can reach the observer, green light from the upper limb can. The blue light is generally absorbed or diffracted by the atmosphere. The phenomenon requires a clear atmosphere; whether unusual refractive power is also necessary has not been settled.

Scintillation or Twinkling of the Stars.—Refraction by the atmosphere is responsible for the twinkling of the stars; it is due to irregularities in the density of the air, the irregularities being carried along by the wind. The unmistakable twinkling has been known for ages as distinguishing the stars from the planets which shine with equally bright but steady light. The apparent lack of

scintillation of a planet is due to the finite area of the planet's disc as compared with the point-source of light which represents a star.

Blood-rain.—Occasionally rain of a red colour is observed which leaves a red stain on the ground. The coloration is due to the drops containing small dust particles which have been carried by currents in the upper air, sometimes for long distances, from sandy regions.

St. Elmo's Fire.—Brush-like discharges of electricity sometimes seen on the masts and yards of ships at sea during stormy weather; it is also seen on mountains on projecting objects. It may be imitated by bringing a sharp pointed object, such as a needle, near a charged Leyden jar. According to Trabert's observations on the Sonnblick the character of St. Elmo's fire changes with the sign of the electricity which is being discharged into the air. The negative "fire" is concentrated so that an object like a mast is completely enveloped in fire. The positive fire takes the form of streamers some three or four inches long.

Ball Lightning.—An occasional incident during thunderstorms is the appearance of ball lightning. The circumstances vary considerably. The size of the balls varies greatly, from the size of nuts to spheres a foot or two in diameter. The most frequent balls are between 10 cm. and 20 cm. in diameter. Sometimes they occur immediately after a brilliant flash of ordinary lightning, at other times when there has been no flash at all. Sometimes only a single ball is seen, at other times the same observer sees two or three and several are reported in one locality. Sometimes the balls drift through the air and vanish harmlessly, but in some cases a ball has exploded on reaching the ground. Usually there is no sign of heating where a ball has passed yet there is one case where clean holes were bored through several window panes, the glass appearing to have been melted. Ball lightning has been observed in closed rooms, but whether it penetrates the walls or forms inside is not quite certain. The light is seldom brilliant, though occasionally observers have been dazzled. Cases are on record where a ball has broken up into smaller ones, but these are rare. In some cases the balls develop during very heavy rain; in other cases when there has been no rain for several minutes. A ball may last for a few seconds or several minutes. Its movement is never fast, generally the speed is comparable with a walking pace, but it is not clear how the ball lightning is propelled. Probably in most cases it is carried by air currents. Apparently ball lightning does not occur when there is much wind.

Pearl-necklace lightning.—Ball lightning has been associated occasionally with pearl-necklace lightning. This phenomenon is a development of an ordinary lightning flash. Immediately after the flash a number of bright lights are seen. These lights are of uniform

size and appear like pearls on a string. They last about a couple of seconds.

Phosphorescence.—The phosphorescence of the sea is one of the most remarkable of natural phenomena and various explanations of it have been attempted in the past. One theory held that sea water absorbed sunlight by day and emitted it by night. In 1749-50, however, a small animalcule giving a blue light was discovered by Vianuelli and Guixellani in the Mediterranean and subsequent investigation has shown that animal life is the origin of marine phosphorescence. Light production is due to a great variety of organisms, from microscopic ones up to many forms of deep sea fish, and the peculiarity of the light is that it is generated without heat, a process never yet achieved in the laboratory. Different types of light have been analysed and the colour is also found to vary through white, silver, green blue and lilac to red. The method of production has been shown to be some form of slow oxidisation which is entirely automatic in the lower forms of life, but is in some measure under control in the higher forms. Recent research has shown that while phosphorescence may occur anywhere it is most frequent in the warmer tropical seas and in particular in the Arabian Sea, where it exhibits a definite maximum in August. In the North Atlantic Ocean some regions show summer maxima, while others have spring maxima. It also appears that phosphorescence is more frequent in coastal regions than in mid-ocean. One remarkable type of phosphorescence is the diffused "milky sea" which may give light enough to read by and to illuminate clouds. This type has often been reported as exerting a calming effect upon the sea surface. More elaborate phenomena occur which have not yet been satisfactorily explained, such as phosphorescent bands and the great rotating phosphorescent wheels.

Atmospherics.—Electrical impulses of natural origin which cause crashing or grinding noises in a wireless receiving set. They are believed to originate in the earth's atmosphere, most probably in discharges of the same nature as lightning. Atmospherics of measurable strength are received in England at the rate of about 50 per second at night, and in the tropics about 50 per second during the day and a few hundred, rising occasionally to a few thousand, per second, during the night. By means of radio direction recorders, the sources of many atmospherics have been determined, and it has been found that some radiate from centres 4,000 miles away. The average strength of atmospherics received in England is consistent with their radiation from centres 2,000 miles away, which in turn is consistent with their origin in regions in which thunderstorms are frequent.

Meteor.—A meteor or shooting star is a fragment of solid material entering the upper regions of the atmosphere from outer space and visible by its own luminosity. The luminosity is attributed to

incandescence due to the compression of the air in front of the meteor. A large meteor may have a luminous trail that persists for half an hour or longer. The majority of meteors are exceedingly small, comparable with grains of sand, but large bodies—meteorites, aerolites or siderites—fall to the ground occasionally. There is no generally accepted theory of the origin of meteors. The appearance of a large meteor is followed occasionally by sounds usually known as "detonations." The sounds are probably due to the waves created by the meteor making its way through the air, not to the explosive breaking up of the meteor. The sounds can be heard occasionally at a distance of 50 miles from any part of the visible path of the meteor. In some cases a zone of silence and an outer zone of audibility have been found.

To explain the breaking up of a meteor it has been suggested that the air pressure acting on a surface which is probably quite irregular causes the meteor to spin. The meteor is not strong enough to withstand the centrifugal force generated by very rapid rotation and therefore breaks into fragments.

CHAPTER XIV.

Sea, Ocean Currents and Ice

The sea covers about three-quarters of the earth's surface and the proportion of sea to land is much greater in the southern hemisphere than it is in the northern hemisphere. The greatest sounding that has been recorded in the ocean is 5,703 fathoms, in the Emden Deep east of the Philippine Islands, so that the height of the atmosphere which is at least 130 statute miles is more than twenty times that of the greatest depth of ocean. The effect of solar insolation is to heat the land more quickly than the sea ; the land is only affected by this heat to a depth of a few feet while over the ocean the heat penetrates to a considerable depth, and its surface temperature is only slightly raised. Similarly, the land loses its heat more readily than the sea and so the changes of air temperature over land areas are much greater than those over the sea. As a result the climates of the oceans are more equable than those of the continents. Owing to the absence of surface irregularities the atmospheric circulation of air is more perfectly established over the ocean than it is over the land where obstacles are encountered in the form of mountain ranges, etc. Steady winds, therefore, are experienced over the ocean and they, in turn, control to a vast extent the climates of the globe.

Sea and Swell.

The wind acts on the sea surface and produces waves. These waves travel and develop into " swell " which may persist and be propagated through enormous distances from the source even after the original cause of the wave motion has ceased or passed away. When wave motion is set up in the sea there is no immediate horizontal translation of water due to the motion of the waves. The energy, however, imparted to the water by the action of the wind, is transferred through the succession of cycles described by the waves in the course of their journey, until it is finally expended as the waves are dashed to pieces on the sea-shore. In general, the state of the sea in different parts of the ocean varies or persists according to the wind, so that the wind circulation gives some indication of the sea which may be expected in the different wind belts. The length of the waves is increased when the length of the sheet of water to windward, that is, the " fetch " is great, although the wind force may remain the same. In the region of the Roaring Forties, the sea is more continuously heavy, for the absence of land affords an unobstructed passage for the winds and the waves right round the earth from the westward. Long waves travel faster

than short waves and with a given wind force the wave height increases quickly and soon attains its maximum height. The height also decreases more rapidly than the other dimensions of a wave when the wind drops. Waves set up by the wind pass on to great distances, the length and speed of travel remaining the same, the height diminishing as they proceed, so that the long low regular undulations characteristic of swell are formed. Their course is straight and not affected by the rotation of the earth as is that of the wind. The highest waves due to wind are encountered near the middle of the North Atlantic Ocean where wave heights up to 80 feet from trough to crest have been measured. Sea and swell are often present together and frequently travel in different directions. Cross seas and swell may be experienced in all parts of the oceans but in some regions they are of frequent occurrence and may give rise to a steep sea, a source of great danger to small vessels. Such conditions may also be experienced when a current sets in a direction across the wind. A striking example of this is provided in the Indian Ocean between 100 and 250 miles south of Sokotra, where, during the SW. monsoon season a very strong current sets in an ESE. direction across the wind and there is very often a dangerous cross sea. In July and August, when the monsoon and the current are usually strongest, on as many as 44 in 100 occasions, cross seas are encountered in this region. A more detailed description of sea and swell will be found in the Admiralty Manual of Navigation (1937), Vol. III, Chapter XX.

Ocean Currents.

Currents induced in the surface layers of the ocean may result in a transference of water over a range of many thousands of miles. Consequently, warm water may be conveyed from tropical regions into temperate latitudes, while water from the colder oceans may flow in the form of a current into the tropics. Ocean currents themselves therefore may have a marked effect on the weather over adjacent seas and land areas. Two notable examples of such currents are the Gulf stream and the Labrador current. The former flows in a north-easterly direction from the Gulf of Mexico, conveying warm water as far as the coasts of Europe, and influencing the climate of the western and north-western seaboard of the European continent, as well as that of the waters adjacent to its path. The latter current flows down the coasts of Labrador and Newfoundland, carrying water from the Arctic regions, and by virtue of its chilling effect on the surface air, gives rise to the sea fog which prevails over the Newfoundland banks in summer. Wind friction is one of the main forces which induce ocean currents, but there are, in addition, internal influences which play an important part in their creation. Of these the most important are the effects of temperature and salinity. Differences in temperature and salinity are due to the movement of water from different parts or from different depths

of the ocean, or from loss of water due to evaporation, or excess of fresh water due to rain, rivers and melting ice. Some or all of these generating forces, then, are responsible for the creation of ocean currents, and as a large proportion of them are set up in belts of steady winds, their direction of flow will tend to conform to that of the generating wind. Owing to the earth's rotation, however, the current suffers a deflection in its course which is directed to the right in the northern hemisphere and to the left in the southern hemisphere. The configuration of the bottom is also a factor which deforms currents once they are generated. The effect has been expressed in a general statement, viz., bottom configuration in the northern hemisphere tends to deflect currents to the right, clockwise, around islands and shoals (left, in the southern hemisphere), and to the left, anti-clockwise, around basins and deeps (right, in the southern hemisphere). As no single force can be held responsible for the creation of ocean currents, a current may therefore run across the wind or even against it. It is well known that even in well-defined main streams of the ocean, variable currents are encountered so that it is not possible to state definitely at any time how the current will run. An account of the distribution of ocean currents and their origin is beyond the scope of this book, but a study of the wind systems over the sea and the circulation of the surface waters will lead to the general conclusion that surface currents mainly conform to the wind circulation, but that they are deflected by the coasts and the configuration of the ocean bed.

Ice.

As has been seen above, cold water currents emanating from the polar regions play an important part in the development of weather conditions over the adjacent seas and land areas. Closely associated with these currents is the phenomenon of ice, for not only do cold currents originate in icy regions but they frequently provide the medium whereby ice in some shape or form is conveyed from the polar regions into lower latitudes, constituting thereby a serious menace to navigation. There are about ten different forms in which ice persists at sea according to the manner in which it was originally formed. The ice which drifts to the trade routes and which is a danger to navigation is, however, of two main types :— land ice or glacier ice, commonly called icebergs, and sea ice or pack. Other types, which are mainly different forms of the above, are :— pancake ice, hummocky floes, drift ice, growlers, brash, etc. (A complete list of ice terms and their definitions, in general use, is given in the Arctic Pilot, Vol. II).

Icebergs, on account of their size, are especially dangerous, as they may float with as much as eight ninths of their mass below water. The glaciers of the Arctic regions to which the majority of icebergs of the northern hemisphere owe their origin, occupy immense areas and over Greenland they extend for miles inland. Owing to

the configuration of the land and the increase of pressure due to continuous accumulation of ice and snow in its rear, the glacier is carried forward to the ocean, and, on reaching the shore, while still preserving its continuity, protrudes seaward, borne up by the sea. The process of disintegration resulting in the formation of icebergs is simplified when the glacier occupies a river-bed or ice "frith." Eventually its equilibrium is destroyed either by sea disturbance, by its own weight, or some other agency, a portion of the floating mass parts from the parent glacier and an iceberg is "calved". Pack ice is a term used in a wide sense to include any area of sea ice which is water-borne, while growlers are pieces of glacier ice so small as barely to show above the surface of the sea. The size of icebergs is often deceptive to the eye. The height of glacier bergs never exceeds 450 feet above the surface after calving and in the region of the North Atlantic trade routes those measured do not exceed 270 feet above the sea, or 1,700 feet from end to end above the water.

The majority of icebergs in the southern hemisphere are tabular in form and originally formed part of an ice barrier. Ice islands up to 70 miles in length have been seen in as low a latitude as 55 degrees south in the region near South Georgia, while heights of from 800 to 1,700 feet from water-line to summit have been recorded. The disruption of ice barriers apparently is due to the unequal expansion of the ice mass; for in the winter when the temperature of the air is probably 40 or 50 degrees below zero, and the sea from 28 to 30 degrees above, so great a temperature contrast cannot fail to produce the separation of large portions of the barrier, and these are driven to the north by the prevailing wind.

The movement of ice is controlled by ocean currents and the distribution of atmospheric pressure in the region of its formation. In the northern hemisphere the centres of many depressions that influence the weather conditions in Davis strait and Baffin bay, during their progress eastward, pass to the south of those regions, and the strong winds and gales associated with these disturbances, blowing as they do from eastward in their northern segment, drive the icebergs that have not left the shelter of the friths on the west coast of Greenland into open water. During the colder months of the year, the average distribution of atmospheric pressure to the eastward and westward of Davis strait and Baffin bay is conducive to the prevalence of northerly or north-easterly winds; pressure is relatively low over southern Greenland and a large area of low pressure is situated over the ocean immediately to the south-east or south of Greenland. It is relatively high over the Dominion of Canada between the 80th and 120th meridians of west longitude. These winds act with the prevailing current in carrying bergs, with other ice, to the southward, steering them at the same time towards the western shores of the strait, where it may be assumed some of them winter, while others are again

driven into the open by westerly gales that are experienced when the centres of depressions pass to the north of the strait. Ice that has broken away from the western side would also be carried to the eastward by these westerly gales and to the southward by the current. The result is that while some of the ice is driven on to the south-west coast of Greenland a considerable part of it clears the land and drifts into the Atlantic. This assemblage of ice is probably reinforced by a few icebergs and pack ice which has collected near the southern coasts of Greenland and Labrador and on being released by early winter gales resume their drift southward into the Atlantic. Ice may therefore reach the trans-atlantic trade routes as early as January or February and soon after come under the influence of the warm current from the south. Wind as well as current has a great influence on the drift of all forms of ice, but current is mainly responsible for the drift of icebergs. The Labrador current is greatly increased in speed during the spring and summer months by the melting of ice in the Arctic seas, and by the prevailing northerly and north-easterly winds. Consequently, the great southward movement of ice towards the Newfoundland banks commences at that period of the year. Some icebergs may be several years upon their journey to lower latitudes while others only take four or five months to make the passage, the quantity of ice that appears in the North Atlantic each year depending for the most part upon the meteorological conditions that have obtained in that ocean and in the adjoining Arctic region, during the previous summer. As has been seen, ice may drift into the trade routes as early as January or February but the amount generally increases in the late spring and early summer. During that period the International Ice Patrol, which operates on the Newfoundland banks, locates the ice and informs shipping by W/T as far as possible of its position. Bergs, growlers and pieces of ice have been sighted at one time or another over the greater part of the North Atlantic north of latitude 30 degrees north, but it is improbable that ice observed south of the latitude of Cape Farewell in Greenland has arrived in that position by any other way than that of the Labrador current and its branches. An account of the geographical distribution of ice is outside the compass of this book, but the description given above of the typical case of ice in the North Atlantic illustrates the important part played by meteorological conditions in its formation and movement.

When navigating in regions where ice may be encountered, no possible means for detecting its presence should be neglected. There are certain meteorological and other phenomena associated with ice, the recognition of which may prove a timely warning to the seaman, even in the case of thick weather. One of these is the phenomenon known as " ice blink." This is an effulgence reflected from ice, which is seen in the sky near the horizon in the direction of the ice, and which renders a berg that is snow-covered, distinguish-

able at some distance; at short distances this brightness may assume the appearance of a white cloud. In dense fog when the sun is shining, an iceberg cannot be seen at a greater distance than 100 yards, when it appears as a luminous white mass; if the sun is not shining it first appears close-to as a dark shape. In light fog or drizzle a berg is visible from one to three miles. The absence of swell in a fresh breeze may be a sign of sea ice to windward. Echoes of sound reflected from icebergs are uncertain, so that warning by echo of the siren is quite unreliable. Mirages are frequent under favourable atmospheric conditions on the southern and eastern sides of the ice regions of the northern hemisphere, where the cold and warm waters mix. Icebergs often appear inverted and assume fantastic shapes, causing them to appear much larger than they really are. The presence of several growlers and of smaller pieces of detached ice signify that a berg is in the locality and probably to windward. As a rule no appreciable change of air temperature can be detected near an iceberg, nor do sudden changes in sea temperature and salinity signify that there is ice at hand; in the neighbourhood of the Newfoundland banks, however, the sea temperature drops very suddenly on crossing the "Cold Wall" or boundary between the cold northern water and the warm tropical current of the Gulf stream. Not only is there a difference in temperature on either side of the wall, but there is a striking change of colour, the Arctic side of the wall being an olive or bottle green, while the Gulf stream side is indigo blue. Such a change in the sea water characteristics should therefore be taken as a warning that the ship has entered the area in which bergs are more liable to be met than southward in the warm water. Late in the season of their frequency, icebergs in the process of disintegration frequently emit cracking sounds like rifle shots. In warm water these manifestations are of hourly occurrence and can be heard during calm weather at a distance of half a mile or even a mile.

Such indications or warnings of the proximity of an iceberg may prove of assistance in its detection, but they are only supplementary to a good look-out, which is essential.

SECTION III

SYNOPTIC METEOROLOGY

CHAPTER XV.

Synoptic Messages

Introductory.—Day-to-day weather forecasting is carried out primarily with the aid of weather maps or synoptic charts. A weather map of any area is an outline chart on which are written, with the aid of conventional symbols, observations of weather made at a number of stations distributed over the area concerned. These comprise land stations, lightships, lighthouses and ships. Stations of the first three types, fixed in position, have been established in most parts of the world. Ships are for the most part confined to the regular cruising and trade routes, and their number and location vary from day to day. For convenience the various types of stations and ships mentioned above will be referred to collectively as "stations."

The stations of each country are organized under a central meteorological office to which they forward their observations as soon as completed, by wireless, telegraph or telephone. At the central office, the observations are (*a*) immediately plotted on the current weather chart, and (*b*) collected together and broadcast by W/T in the form of a single collective message. By picking up such of the collective messages as it requires any forecasting centre may construct its own weather chart of the area in which it is interested.

Information Contained in Synoptic Messages.

The information included in the station reports is given in Table VIII. Not all the observed elements are included in reports from stations of a particular class; for example, humidity and cloud height are not included in reports from ships. In Table VIII "L" denotes that the information is included in reports from land stations, "S" from ships, "C" from coastal (land) stations, and "V" from light vessels.

Table VIII.

Information.				
State of sky — Types of cloud	L	S	—	V
State of sky — Amount of cloud	L	S	—	V
State of sky — Height of lowest cloud	L	—	—	V
Wind — Direction	L	S	—	V
Wind — Force	L	S	—	V
Temperature	L	S	—	V
Humidity	L	—	—	V
Visibility	L	S	—	V
Visibility seawards	—	—	C	—
Weather at time :—Rain, snow, fog, etc.	L	S	—	V
Weather since last observation	L	S	—	V
Barometer reading	L	S	—	V
Barometric tendency (change in last 3 hours)	L	S	—	V
State of sea	—	—	C	—
Swell (type and direction from which it comes)	—	S	—	V
Sea temperature	—	S	—	V
Position of ship	—	S	—	—
Movement of ship (course and speed)	—	S	—	—
Day and time of report	—	S	—	—

At certain hours at land stations observations are made and included in the reports, of elements of climatological importance. These are :—

Amount of rainfall.
Duration of sunshine.
Maximum temperature.
Minimum temperature.
Grass minimum temperature.
State of the ground.

Times of Observation.—In order that the observations from the various stations may be properly co-ordinated, it is essential that they shall be made as nearly as possible at the same time. Accordingly certain hours have been fixed as times of observation, but in some cases departures of one hour from these fixed times are allowed. The times of observation for Europe and north Africa are shown below and in all cases are G.M.T.

Station.	Times.			
All land stations (standard hours)	—	0700	1300	1800
Ships	0000	0600	1200	1800
Certain land stations	0100	—	—	—
Certain land stations (for aviation)	—	—	1000	1600
Certain land stations (for aviation)	0400	—	—	2200

Other times of observation have been fixed for land stations in America, Australia and the East. For details see A.L.W.S., Vol. I.*

Meteorological Codes.

To transmit the observations in plain language would result in unduly long and cumbersome messages. Coded forms have therefore been devised. A coded message consists of a number of five-figure groups, the meaning to be attached to any particular figure, depending upon its position in the message, that is, upon the group in which it occurs and its position in that group. For example, in the most widely used form of report the last two figures in the fourth group always give the temperature, the third figure of the second group visibility, and so on. So familiar does the forecaster become with the coded form of message that at a glance it conveys to him all the details of the weather reported, much more readily than if the same information were written out in full.

Coded messages are most readily interpreted with the aid of symbolic forms. Each element in a message is represented by a

* Extracts from A.L.W.S. in this Chapter were correct in January, 1939, but are liable to be altered. From 1940 onwards they should be verified by reference to the A.L.R.S. Vol. III.

letter, or, when two or three figures are used in the message for a particular element, by the same letter once or twice repeated. For example, visibility, which is reported by a single figure, is represented by the single letter "V"; temperature which requires two figures for its specification in whole degrees (Centigrade or Fahrenheit) is represented by the two letters "TT." Elements of a quantitative nature, such as temperature, can be represented without further modification in a message consisting of numerals. Other elements of a descriptive nature, such as weather and form of cloud, can be represented numerically only by means of codes.

The forms and codes in general use are those adopted by the International Meteorological Conference held at Copenhagen in September, 1929. They are known as the New International Code (1929).

The symbols, forms and codes are given in the Admiralty List of Wireless Signals, Vol. II, and the details of messages issued from individual stations in A.L.W.S. Vol. I. This information is also given in full in "Secretariat de l'Organisation Meteorologique Internationale, (O.M.I.) Publication No. 9, Les Messages Synoptiques du Temps," and for British synoptic issues also in "Wireless Weather Messages" (M.O. 252).

NEW INTERNATIONAL CODE (1929).

The Symbols and their Meanings.—*See* Admiralty List of Wireless Signals, Vol. II (1937), Table I, pages 41, *et seq.*

Symbolic Form of Messages.—(i) *Reports from ships at sea*:—

F 232.—YQLLL IIIGG DDFww PPVTT $3C_LC_MC_HN$ $T_dKD_KWN_h$ $d_sv_s$app.

F 261.—YQLLL IIIGG DDFww PPVTT $6KD_KCN$ $T_dd_sAWC_H$.

F 291.—YQLLL IIIGG DDFww PPVTT $9SKD_KW$ CNN_hAT_d.

F 2.—YQLLL IIIGG DDFww PPVTT.

One or more of the later groups may be omitted in F 232, F 261, or F 291. In such cases the initial 3, 6 or 9 in the fifth group distinguishes the form employed.

(ii) *Reports from land stations and light vessels*:—

F 11.—$IIIC_LC_M$ wwVhN_h DDFWN PPPTT UC_Happ (RRjjj).

F 12.—$IIIC_LC_M$ wwVhN_h DDFWN PPPTT UURRt_W.

The groups in parentheses are sent at certain hours only. The group RRjjj is sent only at 0700 and 1800. In reports of all kinds, missing figures are replaced by "X" (one for each missing figure).

Collective messages are normally arranged in sections each containing a sequence of reports of the type indicated and preceded by an index group, as follows:—

Index Group.	Information contained in Section.
Synop	Ordinary reports from land stations.
Ship(s)	Ordinary report(s) from ship(s).
Pilot	Reports of upper winds from land stations.
Temp.	Reports of upper-air temperatures from land stations.
Neph.	Reports of upper and medium cloud movement (nephoscopes).
Ship Pilot ..	Reports of upper winds from ships.
Temp. ship ..	Reports of upper-air temperatures from ships.

Each index group refers to all that portion of the message between it and the next subsequent index group. In collective messages where reports are included in Codes F 11 and F 12 the reports in Code F 12 will be collected together and prefixed by the group TROP.

Specification of the Scales.—*See* Admiralty List of Wireless Signals, Vol. II (1937), Codes I—XXIII, pp. 45—53. Reference should be made to Chapter X, page 228, for an explanation of the abbreviations for cloud forms in Codes I, II, III and XVI.

Synoptic Organization in the British Isles.

The headquarters of the British meteorological service is the Meteorological Office, Air Ministry, London, where the principal work of forecasting weather for the British Isles and adjacent waters is carried on, together with the collection and re-distribution of synoptic reports. Observations are received regularly from British ships in the north-eastern Atlantic and the North Sea, and from forty-five land stations in the British Isles. The ships comprise :—

(*a*) H.M. ships.

(*b*) Merchant ships of British registry which are selected ships of the British Meteorological Office voluntary observing fleet.

The land stations fall into three main classes :—

(*c*) Distributive stations at Royal Air Force and civil aerodromes. These are staffed by trained Meteorological Office personnel, and are local forecasting centres, utilizing the collective messages broadcast by the Air Ministry. Observations of upper winds and occasionally of upper-air temperatures are made at these stations.

(*d*) Telegraphic reporting stations. These are maintained by the Meteorological Office but the observations are taken by coastguards, lighthouse keepers and private observers.

(*e*) The geophysical observatories of the Meteorological Office at Kew, Aberdeen Eskdalemuir and Lerwick.

Receipt of messages from these stations and ships begins at the Air Ministry immediately after the observations are taken and the plotting of the current synoptic chart commences forthwith.

The reports are re-issued in three collective messages, commencing (for the standard hours 0700, 1300 and 1800) at five minutes, thirty minutes and one hour after the time of observation. By international arrangement the reports from Iceland and the Færoes are included in these transmissions.

SYNOPTIC ORGANIZATION IN THE NORTHERN HEMISPHERE.

A procedure similar to that in the British Isles is carried out in each country. The number of countries in Europe alone is, however, too great for any one meteorological establishment conveniently to receive all the collective messages. A regional scheme has therefore been devised under which the collective reports from a region are received by the central office of one of the larger countries in that region, and a selection of stations re-broadcast. The information for a large part of the northern hemisphere is accordingly contained in six main transmissions. These are :—

(*a*) Paris (Eiffel Tower).—West and south-west Europe, the British Isles, Iceland, north Africa, and the ships of the countries in the region.

(*b*) Berlin.—North, central and south-east Europe and ships of the countries in the region.

(*c*) Rome.—Southern Europe and central and eastern part of North Africa.

(*d*) Moscow.—U.S.S.R.

(*e*) Rugby.—A selection of station reports from (*a*), (*b*) and (*c*) for utilization by Canada and the U.S.A.

(*f*) Arlington.—U.S.A., Canada, Greenland, West Indies (1st July-31st October) and ships in the western Atlantic and eastern Pacific.

MESSAGES USED IN H.M. SHIPS.

(i) **Fleet Synoptic Messages.**—H.M. Ships in which a qualified meteorological officer is borne rely largely on the collective messages from the Air Ministry and those enumerated in the preceding paragraph. An alternative and shorter form of message, known as a Fleet Synoptic message, is also broadcast on the Home and most foreign stations chiefly for the benefit of small ships. Fleet Synoptic messages contain reports from ships and from a selection of land stations, frequently also forecasts of weather in the area to which they apply and statements regarding the positions and probable movements of depressions, anticyclones and fronts. The following table of Fleet Synoptic messages is compiled from A.L.W.S., Vol. I. Reference should be made for further details to the current edition of this publication.

The form of report for land stations varies slightly from station to station, but all reports contain the groups DDFww PPVTT. Ship reports are given in the abbreviated form (F 2).

Fleet Synoptic Messages.

Station.	Transmitting W/T Station.
Home	Cleethorpes.
Mediterranean	Gibraltar and Malta.
East Indies	Bombay Fort.
China	Singapore. Hong Kong.
Australia	Point Cook (Melbourne).
New Zealand	Wellington, Suva, Apia.
America and West Indies	Halifax.

(ii) **Weather Shipping Messages.**—(*See* A.L.W.S., Vol. II, (1937), page 88.) The weather shipping message is a useful supplement to the Fleet Synoptic on the Home station. It contains station reports from a number of stations round the coasts of the British Isles, together with forecasts for adjacent waters. It is broadcast from Rugby. Used in conjunction with Karlsborg W/T Weather Bulletins, it provides complete meteorological information for shipping in north-west Europe. The form of report in both messages is :—IAPPV$_s$ DDFww.

(iii) **Messages Broadcast by Other Countries.**—These are often useful to ships in adjacent waters. The principal messages are given below, but for a complete list it is necessary to refer to A.L.W.S., Vol. I.

Station.	W/T Transmitting Station.
North America (West Coast)	San Francisco.
South America	Rio de Janeiro.
South Africa	Roberts Heights.
China	Cavite (Philippine Is.).

REPORTS OF UPPER WINDS.

Reports of wind at various levels in the upper atmosphere, usually obtained by means of pilot-balloon ascents, are included in many collective messages. They are transmitted in the New International Code (1937). The direction and speed of the wind at various heights are reported in the form :—

PILOT *IIIGG HHddv*$_5$ *HHddv*$_5$ etc.

(A description of the new international code is given in Chapter III, page 86, which should be consulted).

Individual elements in above report.

(i) *IIIGG* : *see below*. In ship reports this is replaced by the usual groups *YQLLL lllGG*.

(ii) *HHddv*$_5$: in this group :—

HH, *HH*, etc. = the heights in hectometres, at which upper winds are reported, usually (for reports issued by the Meteorological Office, Air Ministry, London) sea level, 5 hectometres, 10 hectometres, etc. The corresponding heights in feet are given in Table VII.A, page 89.

dd = direction of the wind on the scale 01-36.

v_5 = speed of the wind according to Table VII.B, page 90.

N.B.—If 50 is added to the wind direction *dd*, the corresponding wind speed is from 26 knots to 52 knots (inclusive). When supplementary groups 44444 $m_1m_2m_3m_4m_5$ etc. are added to the message the wind speed is greater than 52 knots. When 50 is added to the wind direction *and* groups 44444 etc. apply, the wind speed is greater than 78 knots.

Example of upper wind reports :—

Message (i)	PILOT ..	11806	00302	05318	10814	15826
(ii)	PILOT ..	16316	00275	05309	10810	15813
			20805	25291	44444	00000
			1XXXX			

The first of these reports is of an ascent at Renfrew (118) at 0600 G.M.T. The second, an ascent at Calshot (163) at 1600 G.M.T.

Referring to Tables VII.A and VII.B, pages 89-90, these reports are decoded as follows :—

Station.	Sea level.	1,600 ft.	3,300 ft.	4,900 ft.	6,600 ft.	8,200 ft.
Renfrew ..	300°/5 kts.	310°/22 kts.	310°/38 kts.	320°/43 kts.	—	—
Calshot ..	270°/13 kts.	300°/24 kts.	310°/27 kts.	310°/35 kts.	300°/40 kts.	290°/57 kts.

In some Fleet Synoptic messages upper winds are reported as follows :—

Land Stations.—The direction and speed of the wind at various heights is reported in the form :—

PILOT IIIGG $H_1ddv_1v_1$ $H_1ddv_1v_1$ etc.

The index word PILOT appears before each sequence of upper-wind reports.

III = station number from which the observation is made.
GG = time of ascent (G.M.T.) to nearest hour.
dd = direction *from* which the wind blows in degrees from true North, divided by 10 and rounded off to the nearest whole number.
v_1v_1 = speed of the upper wind.
H_1, H_1, H_1, is the series of International Heights at which upper wind is reported. *See* Code XVII.

Calm at any height is indicated by the figures 0000 for ddv_1v_1.

An upper-wind report may often be recognized from the increasing sequence formed by the first figures of successive groups.

If the wind is not available at any height the corresponding group is omitted from the message.

Ships.—The direction and speed of the wind at various heights is reported in the form :—

SHIP PILOT YQLLL IIIGG $H_1ddv_1v_1$ etc.

where the groups $H_1ddv_1v_1$ etc. are to be interpreted as in land station reports.

Examples of upper wind reports.—

Message PILOT 15106 12312 22419 32424 52628
23607 12220 22525 42636

The first report is of an ascent made at Croydon (151) at 0600 G.M.T., the second at Abbeville (236) at 0700 G.M.T. In the Croydon report the speeds are in m.p.h., which may be converted to knots by multiplication by 5/6 (approx.). In the Abbeville report speeds are in kilometres per hour, which, for conversion into knots, must be multiplied by $5/8 \times 5/6 = 1/2$ (approx.).

The two reports may therefore be tabulated as follows :—

Station.	Time.	1,000 ft.	2,000 ft.	3,000 ft.	5,000 ft.	7,000 ft.
Croydon	0600	230°/ 10 kts.	240°/ 16 kts.	240°/ 20 kts	—	260°/ 23 kts.
Abbeville	0700	220°/ 10 kts.	250°/ 13 kts.	—	260°/ 18 kts.	—

Nephoscope Reports.

These consist of reports of the movement of high and medium cloud. The form of report is :—NEPH III $Cddv_rv_r$.

NEPH is the index group which precedes each sequence of nephoscope reports in a collective message.
III = station number from which the observation is made.
C = type of cloud observed according to the code :—

1 Cirrus
2 Cirrostratus
3 Cirrocumulus
4 Altocumulus
5 Altostratus.

dd = direction *from* which the cloud moves, in degrees divided by 10 and rounded off to the nearest whole number.

$v_r v_r$ = angular speed of cloud in radians per hour. To obtain the true speed of the cloud in miles per hour multiply $v_r v_r$ by the assumed height of the cloud, in miles. For cirriform cloud (C = 1, 2 or 3) the height is assumed to be 5 miles and for alto- cloud (C = 4 or 5), 3 miles. Stationary cloud is reported by the figures 0000 for $ddv_r v_r$.

Examples of nephoscope reports.—

NEPH	126 11509	215 40218
Station	Holyhead	Le Bourget
Cloud	Cirrus	Altocumulus
Direction from which cloud moves ..	150°	020°
Speed of cloud ..	9 × 5 = 45 m.p.h.	18 × 3 = 54 m.p.h

REPORTS OF UPPER-AIR TEMPERATURES AND HUMIDITY.

Observations of temperature and humidity at various levels in the upper air are made by aircraft or sounding-balloon ascents. The observations from aeroplanes are included in the synoptic reports.

Land Stations.—Reports in collective messages are in one of the forms (i), (ii), (iii) or (iv) as under :—

(i) TEMP IIIGG PPP TTTUU PPP TTTUU, etc., where :—

TEMP is an index group denoting that the message gives observations of upper-air temperature.

IIIGG has the same meaning as on page 316.

Each pair of groups PPP TTTUU refers to a particular level for which :—

PPP denotes the pressure in whole mb. The initial figure is omitted if the pressure exceeds 1,000 mb.

TTT denotes the temperature to the nearest tenth-degree. For temperatures below 0° the sign is changed and 50·0 added to the numerical value. Thus 535 will denote — 3·5°if from the context it appears that the temperature is below 0°. TTT is expressed in °F. or °C., according to the country of origin of the message.

UU denotes the relative humidity expressed as a percentage, 00 meaning 100 per cent.

A specimen message and its interpretation are given herewith :—

TEMP. 14407 002 35000 970 42085
950 41585 900 38084 850 31591
800 25500 700 145XX 600 515XX 500 700XX

IIIGG = 14407. Station : Mildenhall. Time : 0700 G.M.T.

Pressure.	Temperature.	Humidity.
mb.	°F.	per cent.
1,002	35·0	100
970	42·0	85
950	41·5	85
900	38·0	84
850	31·5	91
800	25·5	100
700	14·5	—
600	−1·5	—
500	−20·0	—

(ii) TEMP IIIGG 00000 TTTUU TTTUU, etc., where 00000 indicates that the values of TTTUU are for fixed pressures, 1,000, 900, 800 mb., etc.

(iii) TEMP IIIGG 99999 TTTPP TTTPP, etc., where 99999 indicates that the values of TTTPP are for standard heights, 1,000, 2,000, 3,000 metres, etc. (strictly dynamic metres ; these may, with small errors, be taken as equivalent to ordinary metres).

PP denotes pressure in whole mb., with the hundreds figure omitted.

(iv) TEMP IIIGG H_dH_dPPP TTTUU, where PPP, TTT and UU have the same meanings as in (i) and H_dH_d gives the height in hectometres (strictly dynamic hectometres ; these may with small error be taken as equivalent to ordinary hectometres). The corresponding heights in feet are given in Table VII.A, p. 89. This is the form which is at present (1939) used in British Synoptic issues.

Ships.—(*a*) Reports in collective messages will be in the forms (i), (ii) or (iii) above, with the substitution of TEMP SHIP for TEMP, followed immediately by the groups YQLLL IIIGG having the same meaning as in ships' weather reports.

Example :—Form (i), to which preference will be given.

TEMP SHIP YQLLL IIIGG PPP TTTUU, etc.

(*b*) Upper-air reports made by individual ships to the Meteorological Office may be intercepted by other ships. They will then be in the form described on page 96.

CHAPTER XVI.

The Plotting of Synoptic Charts

Introductory.—From the information contained in the synoptic messages the current synoptic chart is plotted, either by using conventional symbols to represent weather, cloud etc., or by the method in which written abbreviations are used—this may be referred to as the Beaufort letter method. It is recommended that all officers performing meteorological duties shall make themselves familiar with the symbols and Beaufort letters as well as the various meteorological codes, so that they can readily interpret the synoptic messages and weather charts. This facility in the use of codes and symbols can be most readily and profitably attained by frequent practice in the plotting and study of synoptic charts.

The primary object of using symbols for the plotting of weather maps is to secure a uniform method in all countries. The meteorological officer on board ship or the pilot of an aircraft will then meet with the same system of plotting everywhere and confusion due to differences of method will be eliminated. The symbolic system of plotting used by the British Meteorological Office is that approved by the International Meteorological Conference held at Warsaw in 1935, and the system has been adopted for use in H.M. ships. The Beaufort letter method may for the present be used in cases where the symbols are unfamiliar.

Weather Symbols and Beaufort Letters.

(*a*) **Basic Symbols and Beaufort Letters.**—Table IX gives the basic symbols and corresponding Beaufort letters used in the plotting of synoptic charts.

Table IX.

Weather.	Beaufort letters.	Symbols.
Blue sky (0–¼ clouded)	*b*	
Sky partly (¼–¾ clouded)	*bc*	
Cloudy (more than ¾ clouded)	*c*	
Drizzle	*d*	
Wet air (without precipitation)	*e*	
Fog	*f*	
Fog over sea (coast station)	*fs*	
Fog over low ground (inland station)	*fg*	
Gloom	*g*	
Hail	*h*	△ or ▲
Precipitation in sight of station	*jp*	
Line squall	*kq*	
Storm of drifting snow	*ks*	
Sandstorm or dust storm	*kz*	
Lightning	*l*	
Mist	*m*	
Overcast sky (completely covered with St. or Nbst.)	*o*	
Passing showers	*p*	
Squally weather	*q*	
Rain	*r*	
Sleet	*rs*	
Snow	*s*	
Thunder	*t*	
Thunderstorm with rain or snow	*tr* or *ts*	
Ugly, threatening sky	*u*	
Unusual visibility	*v*	
Dew	*w*	
Hoar frost	*x*	
Dry air	*y*	
Haze	*z*	
Dust-devils		
Waterspout		
Ice crystals falling through the air		

(*b*) **Combinations of Beaufort Letters and Symbols.**—The basic letters or symbols of Table IX are combined in obvious ways to represent commonly occurring types of weather. The occurrence of rain and snow together, that is sleet, is so important that it is included in Table IX as a combination of letters or symbols. Other examples are :—

Weather.	Beaufort letters.	Symbols *See* Table XI.
Fog and drizzle	*df*	*ww* = 57
Rain shower	*pr*	*ww* = 81
Hail shower	*ph*	*ww* = 88
Thunderstorm with snow or rain	*tls* or *tlr*	*ww* = 93

(*c*) **Modification of Symbols and Beaufort Letters to denote Intensity and Continuity.**—(i) *Symbols.*—Continuity is generally represented by repetition of symbols horizontally ; co-existence or intensity by repetition vertically.

(ii) *Beaufort Letters.*—Continuity is represented by the repetition of letters. The letter *i* placed before a Beaufort letter denotes that the phenomenon indicated is intermittent. Capital letters are used to denote intensity. When the phenomenon is slight suffix *o* is used. For telegraphic purposes the letter *a* before a Beaufort letter indicates that that letter is to be read as a capital ; similarly, *n* before a Beaufort letter indicates that that letter takes suffix *o*.

Examples :—

Weather.	Symbols *See* Table XI.	Beaufort letters.	Beaufort letters (telegraphic).
Intermittent slight rain	*ww* = 61	ir_o	*inr*
Continuous slight rain	*ww* = 62	$r_o r_o$	*nrnr*
Intermittent moderate rain ..	*ww* = 63	*ir*	*ir*
Continuous moderate rain	*ww* = 64	*rr*	*rr*
Intermittent heavy rain	*ww* = 65	*iR*	*iar*
Continuous heavy rain	*ww* = 66	*RR*	*arar*

A heavy shower is denoted by the symbol ∀ ; heavy squalls by ⩓.

A solidus (/) placed in a group of Beaufort letters separates weather at the time of observation from weather in the preceding hour. Thus a cloudy sky at the time of observation after a shower of rain in the preceding hour is denoted by *c/pr*.

SYMBOLIC METHOD OF PLOTTING.

Symbols shown in Tables X and XI are used (page 476).

TABLE XII.

DIFFERENCE BETWEEN TEMPERATURE TT AND DEW-POINT T_sT_s.

U	Relative humidity.	Dry-Bulb temperature (° F.).									
Code figure.	Per cent.	10	20	30	40	50	60	70	80	90	100
		Values of TT—T_sT_s.									
0	97	1	1	1	1	1	1	1	1	1	1
9	92	2	2	2	2	2	2	2	3	3	3
8	85	4	4	4	4	4	5	5	5	5	5
7	75	6	7	7	7	8	8	8	9	9	9
6	65	×	10	11	11	11	12	12	13	13	14
5	55	×	14	14	15	15	16	17	18	18	19
4	45	×	×	19	20	20	21	22	23	24	25
3	35	×	×	24	25	26	28	29	30	31	33
2	25	×	×	×	33	34	36	37	39	41	42
1	10	×	×	×	×	×	57	59	62	64	66

Note.—The relative humidities and dew-points refer throughout to a water surface.

Select from the above table the row corresponding to the relative humidity U and the column corresponding to the dry-bulb temperature TT. The dew-point is obtained by subtracting from TT the number common to the row and column selected.

PLOTTING—LAND STATIONS.

F 11.—*III* C_LC_M *wwVhN*$_h$ *DDFWN PPPTT* UC_Happ (*RRSV*$_s$*E*).

Station Circles.—The position of each station on the chart is shown by a small blank circle, or in the case of a less important station by a dot. If it is desired to plot one of the latter stations, the circle should first be drawn in the appropriate position in black ink, 4 mm. in diameter.

The station number *III* is printed near the station circle for the purpose of rapid identification. In many cases the hundreds figure of *III* is omitted. This should in general lead to no difficulty in picking out a particular station, since the same hundreds figure is used throughout each country or group of countries. Some difficulty may, however, be experienced in the case of stations situated near to the boundaries of regions allotted different hundreds figures for the station numbers. The exact position of the station may then

be ascertained from the list of stations in A.L.W.S., Vol. II (1937), page 95.

Station Model.—The observations are plotted in black and red ink and are arranged relative to the station circle in accordance with the model shown in Fig. 97.

The nomenclature of symbols used in the model is that of the International Code (1929), except that :—

w = That part of ww which refers to the last hour, but not to the time of observation.

T_sT_s = Dew-point.

In special circumstances which may arise from time to time, a certain deformation of the station model is permissible, provided that no change is made in the position of the elements relative to one another.

As shown in the model **red** ink will be used for C_H, V, T_sT_s and W. **Red** ink will be used also for a and pp when pp is **negative,** *i.e.*, when the code figure for a is 5, 6, 7, 8 or 9. **Black** ink will be used for the remaining symbols shown, and for a and pp when pp is **positive,** *i.e.*, when the code figure for a is 0, 1, 2, 3, 4.

Order of Plotting.—The following order of plotting is recommended :—

$$DDF \quad N \quad C_L \quad C_M \quad ww \quad N_h \quad h \quad PPP \quad TT \quad a \quad pp \quad V$$
$$W \quad T_sT_s \; (U) \quad C_H \quad (V_s) \quad (S).$$

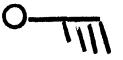
The wind should always be plotted first to permit the grouping of the other elements according to the position of the wind arrow.

In the above order only one change of pen is necessary from black ink to red, either immediately before or immediately after pp is plotted.

INDIVIDUAL ELEMENTS.

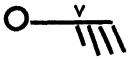
Wind (DDF).—To be represented by an arrow flying with the wind, the head of the arrow terminating at the circumference of the station circle.

The force of the wind to be indicated by feathers on the arrow, sloping slightly backwards. Each complete feather to represent two points of the Beaufort scale, an odd point being represented by a half feather. An odd half feather to be placed on the station side of the group. Thus an east wind of force 7 would be indicated by



The addition of 33 to a wind direction (unusual gustiness) to be

indicated by a small V on the wind arrow placed on the side opposite to the feathers, thus :—



The addition of 67 to a wind direction (definite squall or line squall during past hour) to be indicated by the squall symbol ⋀ immediately after the symbol for past weather (*see* below).

A calm to be indicated by a circle concentric with, and external to, the station circle.

Total Amount of Cloud *N*.—To be represented by a number of vertical lines drawn inside the station circle as shown in Table X. The greater the number of lines, the greater is the degree of cloudiness represented.

Forms of Cloud C_H, C_M, C_L, C.—The appropriate symbols as shown in Table X will be plotted in the positions indicated in the station model. If high cloud C_H, but no medium cloud C_M is reported, the symbol for C_H will be entered immediately above the station circle.

Note that **red** ink is to be used only for the symbols for C_H, and those symbols for C appropriate to high cloud.

The cloud symbols should be compared with the corresponding codes. To some extent the symbols are pictorial representations of the cloud forms to which they correspond. For example, stratified cloud is represented by symbols involving horizontal lines ($C_L = 5$, $C_L = 6$, $C_M = 1$, $C_H = 7$); cumulus or heaped cloud by curved lines convex upwards ($C_L = 2$, $C_M = 6$), and there are various combination symbols.

Present Weather *ww*.—The symbols and their positions relative to the station circle are shown in Table XI. The symbols will be plotted in black.

In the case of code figures 20—29, 41, 42, 91 and 92 the appropriate symbol for past weather will be plotted in place of the red *W*, shown against these numbers in Table XI.

In cases where an alternative is shown (*i.e.*, 10, 20, 80, 93, 95, 97) plot either the rain or the snow symbol whichever, judging by the temperature, seems the more likely.

The vertical stroke means " has increased " if placed before a symbol and " has decreased " if placed after a symbol.

The sign ( ) means " within sight of the station."

No symbols are plotted for *ww* = 00 to 03 for those messages in which *N* appears; the necessary differentiation is provided by the appropriate symbols for *N*.

The symbols in Table XI should be compared with the specification of present weather in Code V and with the basic symbols in Table IX, page 320.

Low Cloud Amount N_h; **Height** h.—The amount and height will be entered in code as received and separated by a solidus; thus, 4/5 means amount $N_h = 4$, height $h = 5$. The code figures for N_h and h should be interpreted with reference to Codes IV and VII.

Barometer *PPP*.—The three code figures received from land stations will be entered in the position shown in the model with a decimal point before the third figure: thus, 09·7.

Temperature *TT*.—Temperature will be entered in whole degrees Fahrenheit in black in the position shown on the station model.

British stations will report temperatures below 0° F. by subtracting the numerical reading from 100. Thus, — 5° F. will be reported as 95; — 13° F. as 87. This procedure will be adopted in coding all surface temperatures of less than 0° F. in all W/T issues (British synoptic, European collectives, Rugby collective, Fleet synoptics) from Air Ministry.

In reports from Continental stations the temperature is sent in degrees Centigrade. Conversion to degrees Fahrenheit is therefore necessary. 50 is added to negative values; thus — 3° C. is sent as 53.

An examination of the temperature distribution at adjacent stations will enable the plotter to determine whether in a particular instance the temperature is positive or negative.

Barometric Tendency and Characteristic *ppa*.—The symbols for *a* should be plotted in black or red, as shown in Table X. They should be compared with Code X. The symbol for *a* typifies the shape of the barograph curve during the preceding 3 hours. There is no symbol for $a = 3$ (barometer steady or rising steadily), nor for $a = 8$ (barometer falling steadily).

The amount of the tendency is reported in fifth-millibars, but plotted in tenth-millibars; *pp* should therefore be multiplied by 2, and plotted without the decimal point. Thus $pp = 13$ in message is entered on the chart as 26.

Note that for positive tendencies ($a = 0, 1, 2, 3, 4$) *pp* is entered in **black,** and for negative tendencies ($a = 5, 6, 7, 8, 9$) *pp* is entered in **red.**

Visibility *V*.—The code figure without the letter *V* will be entered in red in the position shown on the station model. The code figure for visibility should be interpreted by reference to Codes VI and VIB.

Visibility Seawards V_s.—(Coastal stations only at 0700 and 1800). The code figure for V_s preceded by the letter *V* should be written in red over the sea adjacent to the station circle, but well clear of the remaining entries.

State of Sea *S*.—(Coastal stations only at 0700 and 1800).—The code figure for *S* preceded by the letter *S* should be written in red immediately below the entry for V_s.

Past Weather *W*.—The symbols shown in Table X should be plotted in red in the position shown on the station model. They should be compared with the specification for past weather in Code VIII.

Dew-Point T_sT_s.—From the relative humidity *U* and the temperature *TT* the dew-point T_sT_s may be found using Table XII, (page 322), according to the instructions in the footnote. The dew-point is plotted in whole degrees Fahrenheit in red in the position shown on the station model.

Example.—Message 16062 64638 18568 97352 8X706 01371. The plot of this report is shown in Fig. 98.

Plotting—Ship Reports.

General.—Firstly check the date and time of the report (*Y*, *GG*) to ensure that it is appropriate to the chart being plotted. 0600 ship reports are plotted on the 0700 chart, 1200 reports on the 1300 chart, and 0000 reports on the 0100 chart. Draw the station circle in black in the position given by *LLL*, *lll* and *Q*. If the longitude is greater than 100° the hundreds figure is omitted from *lll*.

Station Model.—The station model is identical with that for land stations except that *tt* (sea temperature) replaces T_sT_s (dew-point).

F 232.—*YQLLL lllGG DDFww PPVTT* $3C_LC_MC_HN$ $T_dKD_KWN_h$ d_sv_sapp.

Individual Elements.—*DDF N* C_L C_M C_H *ww TT a pp V* and *W* are plotted as for a land station, (page 322).

Barometer *PP*.—The barometer is sent in whole millibars with the hundreds figure omitted. The entry will be the two code figures for *PP* followed by ·0 : thus, *PP* = 09, entry 09·0.

Low Cloud Amount N_h.—The height *h* is not given. N_h should be entered as for land stations followed by a solidus and a hyphen : thus, $N_h = 4$, entry 4/-.

Sea Temperature *tt*.—The sea temperature is obtained by adding to or subtracting from the air temperature *TT* the amounts shown in Table X. Amounts with plus sign are to be added ; the amounts with minus sign are to be subtracted.

Swell *K*. **Direction of Swell** D_K.—The symbols are shown in red in Table X. The symbol should be plotted to lie in the direction from which the swell comes and should be as near as possible to the position of the ship without interfering with the entries of the other observations.

The direction from which the swell comes is given in Fig. 99.

To face page 326.

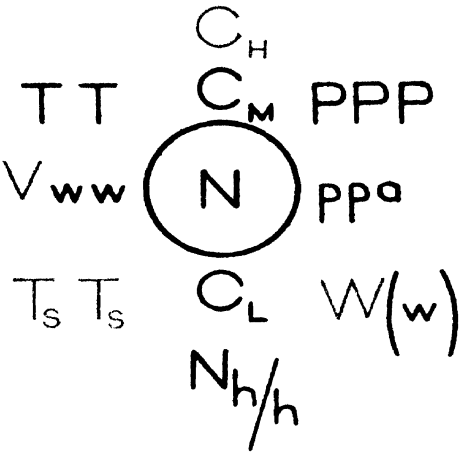


FIG. 97. STATION MODEL.

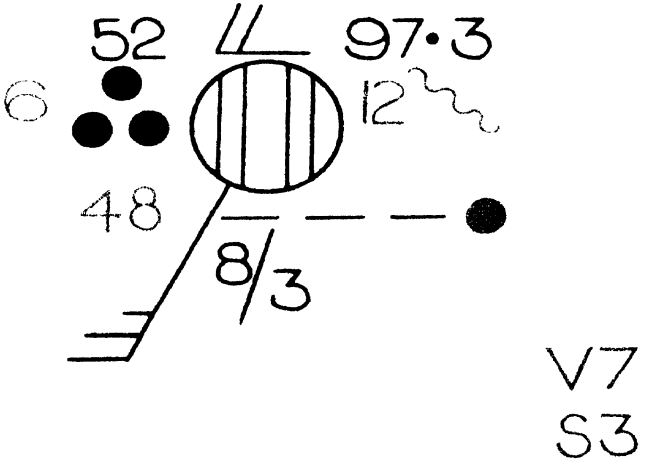


FIG. 98. SPECIMEN PLOT (LAND STATION).

Run of Ship d_s, v_s.—The direction in which the ship is travelling is indicated by a red arrow with reversed head terminating at the circumference of the station circle and lying in the direction from which the ship has come (Fig. 100).

The distance covered by the ship in the last three hours is given in sea miles according to the code figure for v_s in Table X. It should be entered in red within the reversed arrow head for d_s.

Example.—Message 33362 12518 04325 17768 39416 37885 16405. Day of report ($Y = 3$): Tuesday. Time ($GG = 18$): 1800 G.M.T. Position of ship ($Q = 3$): N. latitude, 0-90 degrees east longitude. The position is therefore 36·2 degrees N, 12·5 degrees E. The plot is shown in Fig. 101.

F 261.—*YQLLL lllGG DDFww PPVTT 6KD$_K$CN T$_d$d$_s$AWC$_H$.*

Individual Elements.—*DDF N ww TT V* and *W* are plotted as for land stations. *PP K D$_K$ T$_d$* are plotted as for ship reports, F 232.

Form of Cloud *C*, *C$_H$*.—The symbols for *C* are given in Table X. The code figures 6, 7, 8, 9, 0 denote low cloud, and for these the appropriate symbol should be written in the position for C_L. Code figures 1, 2, 3 denote high cloud and 4, 5 medium cloud, the symbols for which should be written in the positions for C_H, C_M, respectively.

C_H should be plotted independently of *C* unless the symbols for *C* and C_H are the same.

Direction of Ship's Movement d_s.—The reversed arrow head should be used to indicate the direction of the ship's movement as in Fig. 100. The distance covered is not available.

Barometric Tendency and Characteristic Expressed by a Single Figure *A*.—The appropriate numerical entries (tenth-millibars) for the *A* code are shown in Table X. The **red** figures denote falling, the **black** rising tendencies. The entry should be made in the position for *ppa* on the station model.

F 291.—Plot according to such of the instructions for F 261 as are appropriate. *S* should be plotted as for land stations (page 325).

Abbreviated Ship Reports.—

F 2.—*YQLLL lllGG DDFww PPVTT.* Plot according to such of the instructions for F 232 as are appropriate.

PLOTTING—FLEET SYNOPTIC AND WEATHER SHIPPING MESSAGES.

Ships.—Ship reports are always F 2 and should be plotted according to the instructions above.

Land Stations.—Fleet Synoptic: *jjjjj DDFww PPVTT.* Weather Shipping: *IAPPV$_s$ DDFww.*

The group *jjjjj* varies from station to station. All the elements in these messages are included in F261 for ships and the instructions on pages 326-7 should be followed as appropriate.

Since *N* does not appear in the 3-group Fleet Synoptic

and Weather Shipping Reports, the station circle is normally left blank. It may, however, be used for *ww* = 00 to 03, according to the following scheme :

ww		Station circle.
00	cloudless	
01	partly cloudy	
02	cloudy	
03	overcast	

This applies to all reports in which *ww* but not *N* appears.

Beaufort Letter Method of Plotting.

This differs from the symbolic method only in the replacement of the weather and cloud symbols by written abbreviations, and in the use of a simplified table of dew-points. The instructions for plotting by the symbolic method should be followed except that Table XIII should be used instead of Tables X, XI and XII.

The Plotting of Nephoscope Observations.

A nephoscope observation is plotted in red on the synoptic chart in the position of the station reporting it. The direction from which the cloud moves is indicated by a line drawn up to the station circle, similar to a wind arrow. A single line is used for "alto" clouds, a double line for cirrus clouds. The line carries a number of feathers indicating the speed of the cloud, each feather representing 10 m.p.h. They are placed in groups of 3, the odd feathers, if any, being on the side farthest from the station circle. The method of plotting the preceding examples is shown in Fig. 102.

The Drawing of Isobars.

When the observations from stations and ships have been plotted on the chart the next step is to draw the isobars, or lines joining places where the M.S.L. pressure is the same. They are drawn at intervals of 1, 2 or 4 millibars according to the scale of the chart. Generally the M.S.L. pressure at a station is not exactly equal to a whole number of millibars and the isobars must therefore be interpolated between stations. It may be possible to draw them using the observations of pressure alone in a manner similar to that in which contour lines are drawn on a topographical map using the height of various points, but the observations of wind should also be utilized, in accordance with the following rules :—

(*a*) The winds (at the surface) are inclined to the isobars at an angle which varies between 10° and 20° over the sea, and between 20° and 40° over the land, and blow across the isobars from the higher to the lower pressure.

To face page 328

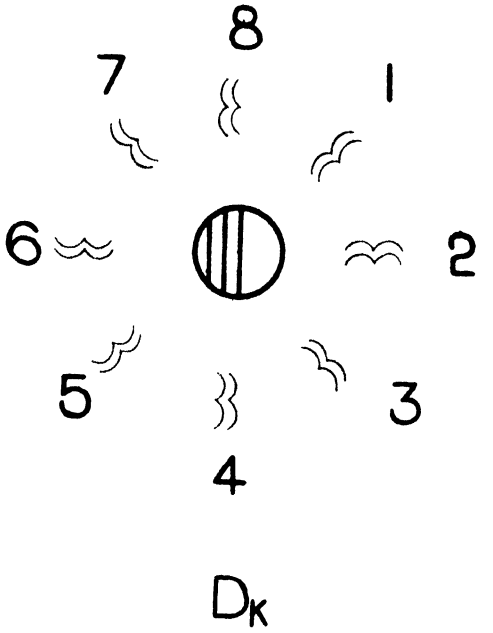


FIG. 99. METHOD OF PLOTTING SWELL.

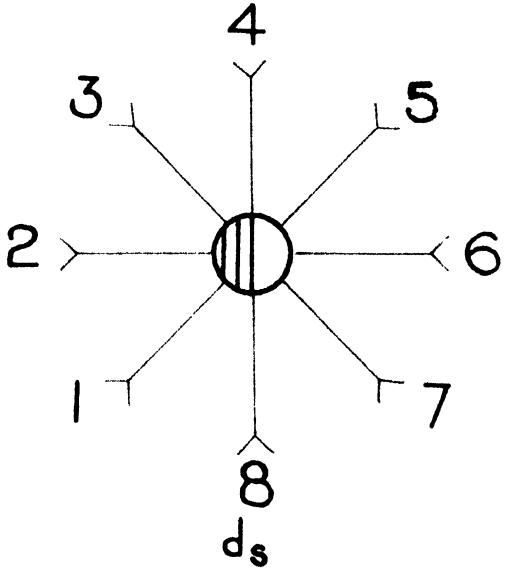


FIG. 100. METHOD OF PLOTTING RUN OF SHIP.

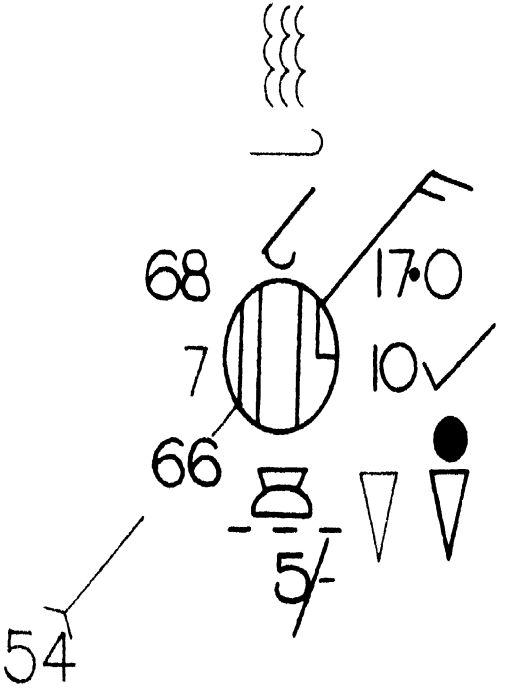


FIG. 101. SPECIMEN PLOT (SHIP).

NEPH 126 11509 (Cirrus from 150° at 45 m.p.h.)

NEPH 215 40218 (Altocumulus from 020° at 54 m.p.h.)

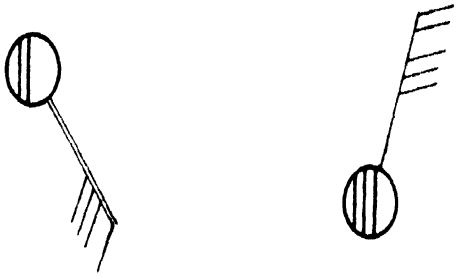


FIG. 102. PLOTTING OF NEPHOSCOPE OBSERVATIONS.

(*b*) If the observer faces the wind the lower pressure is situated on his right in the northern hemisphere and on his left in the southern hemisphere.

(*c*) The stronger the wind, the closer together are the isobars and *vice versa*.

The method of drawing isobars may be illustrated by reference to Fig. 153, in which the isobars over the British Isles and north France for 0700 on 31st October, 1936, are drawn at 2 mb. intervals. Select the lowest or highest pressure on the chart and commence with the isobar appropriate to that pressure. In Fig. 153 the lowest pressure is 1,010·0 mb. at the ship in lat. 54° North, long. 5° East, and the isobar of 1,010 mb. therefore passes through the position of the ship, with the lower pressure to the left of the wind, *i.e.*, to the NW. The isobar for the lowest pressure must in general form a closed curve, since it is surrounded by higher pressures ; if the lowest pressure occurs near the edge of the chart the isobar may, however, be open. The isobars of 1,012 and 1,014 mb. are drawn surrounding that of 1,010 mb. in accordance with the pressure and wind at Helder and Yarmouth. It is then noticed that Portland, whose pressure is 1,011·4 mb., lies outside both these lines, so that separate closed isobars of 1,012 and 1,014 mb. must be drawn enclosing this station. Consider next the isobar of 1,018 mb. There is one station reporting this pressure exactly, namely, Ross-on-Wye, and the wind there is due north. In accordance with rules (*a*) and (*b*) of this paragraph, the isobar of 1,018 mb. passes through Ross-on-Wye running approximately in the direction NE. to SW. with the lower pressure on its SE. side. The isobar must pass S. of Birmingham (pressure 1,019·2 mb.) and almost through Cranwell (pressure 1,018·1 mb.). Since the wind is NW. at the latter station the isobar there runs slightly W. of N. Between Spurn Head (pressure 1,016·9 mb.) and Catterick (pressure 1,020·8 mb.) the position of the line must be determined by rough interpolation. The exact procedure is as follows :— The pressure at Spurn Head is 1·1 mb. less, whilst that at Catterick is 2·8 mb. greater than 1,018 mb. The isobar therefore divides the distance between these stations in the proportion 2·8 to 1·1, *i.e.*, 3 to 1 approximately, and lies nearer to Spurn Head than to Catterick. With practice the position of the line may be judged rapidly by eye. The isobar of 1,016 mb. may then be drawn between the 1,018 mb. and the two 1,014 mb. isobars, deviating from the regular spacing only where required by the observations, *e.g.*, at the station in the extreme SE. of England. The remaining isobars are drawn in a similar manner. Where an isobar intersects a front (*see* Chapter XVIII) it changes direction abruptly. Consequently, as over NW. France in Fig. 153, the portions of the isobar on opposite sides of the front meet at a sharp angle. The lower pressure is always on the inside.

If there are large sea areas over which observations are scanty

it is advisable firstly to draw the isobars over the adjacent land areas, linking them up by as smooth and regular curves as are consistent with the observations over the sea.

The following rules are useful in checking the isobars after they are drawn :—

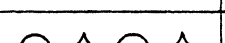
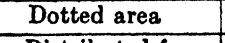
(*a*) Isobars never intersect one another.

(*b*) If a straight line is drawn between two centres of low pressure (or two centres of high pressure) it crosses an isobar for the same pressure twice consecutively. In Fig. 153, if a straight line is drawn between the centres of low pressure over Portland and in lat. 54° North, 5° East, it crosses the isobars in the following order :— 1,012 mb. near Portsmouth, 1,014 mb. SW. of London, and again near Yarmouth, 1,012 mb. NE. of Yarmouth and finally 1,010 mb. near the second centre. It crosses the isobar of 1,014 mb. twice consecutively. If, however, the line is drawn between centres of high and low pressure it crosses the isobars in correct ascending or descending order of magnitude.

Fronts.—The method of representing fronts on working charts and printed charts is given in Table XIV.

TABLE XIV.

SYMBOLS FOR FRONTS.

Character of front.	*on Working Charts*	*on Printed Charts.*	*Direction of Motion of Fronts.*
Cold front at the ground	Continuous blue line		↓
Cold front above the ground	Broken blue line		↓
Warm front at the ground	Continuous red line		↑
Warm front above the ground	Broken red line		↑
Occluded front at the ground	Continuous purple line		↑
Occluded front above the ground	Broken purple line		↑
*Areas of tropical air	Continuous red area	Dotted area	
*Areas of fog	Continuous yellow area	Distributed fog symbols	

The above table is that approved by the Warsaw Conference 1935.

*Note: In the Meteorological Office the normal practice is not to shade areas; if areas are shaded, the method of this table should be followed.

CHAPTER XVII.

Issue of Forecasts, Gale Warnings, Etc.

General.—In addition to the various synoptic reports collected and re-distributed as explained in Chapter XV, weather forecasts for particular areas are also issued at certain specified times daily from meteorological centres all over the world. These forecasts are based on the weather charts prepared from the synoptic information received from reporting stations and ships. They are sometimes incorporated, wholly or in condensed form, in the re-distributed synoptic report messages, in plain language; but for the general public, agricultural concerns and aviation, forecasts are issued as such through the medium of the press, wireless and printed Daily Weather Reports. Facilities are also available at most meteorological centres for the supply of special information by private enquiry. Thus, on receipt of a telegraphic or telephonic request, a special forecast of weather conditions in a certain area will be issued to a ship carrying out operations in that area. Again, a special route forecast may be issued to the pilot of an aircraft about to embark on a flight.

Preparation of Forecasts.—The times at which observations for synoptic meteorology are made have been fixed by international agreement. In order to meet public requirements, the hours adopted throughout Europe are 0100, 0700, 1300 and 1800, G.M.T., all of which, of course, fall within the limits agreed to internationally. The majority of observing stations observe at all four times, but a few omit the 0100 observations. A system of international exchange of selected observations at certain of these times between the principal forecasting centres of the northern hemisphere, provides a picture of the geographical distribution of meteorological elements over a vast area, thereby greatly facilitating the problem of forecasting. At the Meteorological Office, Air Ministry, London, synoptic information is plotted at the above-mentioned times on outline charts which cover the whole of Europe and the North Atlantic. In addition, large-scale charts covering NW. Europe are drawn for these hours as well as for the intermediate hours 1000, 1600, 2200 and 0400 G.M.T. (the last two in summer only). These charts contain, in addition to the usual information, observations of cloud and visibility necessary for aviation forecasts. Small-scale charts covering most of the northern hemisphere are prepared for 0100 and 1300 G.M.T. daily. Thus the forecaster is in a position to follow closely the sequence of weather conditions over any particular area, and an analysis of the meteorological elements therein enables him to prepare and issue a forecast.

Information Contained in a Forecast in the British Isles.—Normally a forecast is issued for 12, 24 or 36 hours, and includes :—

(i) A brief general statement of barometric pressure systems throughout the area affected, and their anticipated movements.

(ii) A statement of the anticipated state of the sky (as regards cloud), precipitation (rain, hail, snow or sleet), visibility and temperature, whether it is likely to be greater than, equal to or less than, the normal for the time of year, or higher or lower than at the time of making the forecast.

(iii) A statement of the anticipated direction and force of the surface wind and the changes therein which are expected during the period of the forecast. (See page 40). When verbal descriptions of the strength of the wind are given in forecasts they are those of the international scale, but the word ' breeze ' is usually omitted except for forces 2 and 3. Thus if the wind is forecast as force 1 it will be described as ' light airs,' if it is forecast as force 3 it will be described as ' gentle breeze,' if it is forecast as force 5 it will be described as ' fresh,' if it is forecast as force 7 it will be described as a ' moderate gale,' and if force 11 or 12 is forecast it will be described as a wind of ' storm force ' or of ' hurricane force ' respectively.

When winds of more than one Beaufort force are anticipated, a combination of the above terms will be used : thus if the wind is expected to be forces 5 and 6 the forecast will describe the wind as ' fresh or strong.' If the wind is expected to be of forces 9 and 10 it will be described as ' strong or whole gale.'

The only exceptions to this procedure will be (*a*) when winds of forces 1, 2 and 3 are grouped together, the group is described as ' wind light ' and (*b*) when winds of about force 8 are anticipated they are described as ' gale ' without any adjective.

(iv) A note as to the probability of such occurrences as night frost, fog or thunder.

(v) A statement under the heading *Further Outlook* giving the conditions likely to be experienced in the 24 hours or more following the period of the forecast.

The term *General Inference* is used for a description of the general pressure distribution and the changes of pressure in progress, together with a statement of the type of weather likely to be experienced through such changes. It usually precedes a series of more detailed forecasts for individual districts, and gives the foundations or framework on which these forecasts are formed.

Thus the daily weather forecasts issued by the Meteorological Office, London, contain detailed information regarding the probability of weather conditions throughout 20 different districts in the British Isles (*see* A.L.W.S., Vol. II).

Forecasts issued for special purposes contain special matter, additional to that already specified. For aviation, for example, the anticipated winds at selected or demanded heights are included, together with a statement concerning the height of low cloud, the visibility and any changes therein. Again, for shipping, statements may be included concerning visibility and the state of the sea. Visibility reports should be used with caution, for, at coastal stations the seaward and landward visibility may differ considerably.

Special Forecasts.—(For issue of special forecasts to shipping generally *see* A.L.W.S., Vol. II). In the British Isles or adjacent waters, special forecasts are issued to H.M. ships or aircraft on request. These forecasts cover the area shown on the map of the British Isles and adjacent waters published in the A.L.W.S., Vol. II, and, in addition, the Bay of Biscay and the Portuguese coast. As synoptic charts are prepared in the Meteorological Office, Air Ministry, London, based on observations taken at 0100, 0700, 1300 and 1800, G.M.T., forecasts required in connexion with special operations can be issued on these charts at the following times, 0230, 0900, 1430 and 2000, G.M.T. Forecasts based on the latest information will be given immediately at any time throughout the 24 hours in response to requests. but at hours intermediate to those given above, the forecasts will normally be based on less complete and up-to-date information than is available at the hours named. Applications for these special forecasts should specify the area, date, time and duration for which they are required, the particular meteorological element or elements upon which information is sought, together with the time at which they should be sent, having regard to the considerations enumerated above. They should be addressed to the Meteorological Office, Air Ministry (telegraphic address: "Weather Telex London") and repeated to the Admiralty. Applications transmitted by W/T should be passed via Whitehall. H.M. ships cruising in the eastern North Atlantic can obtain a general inference from the Meteorological Office, Air Ministry, London, the procedure for obtaining the information being the same as that described above for other special forecasts.

The following examples are given as a guide to the method in which a request for a special forecast should be rendered:—

(i) To Weather Telex London, repeated Admiralty via Whitehall.

From Ramillies.

Request special weather forecast regarding rain, temperature and state of sea for Thames estuary for five hours commencing 0900 at 0900 to-morrow, Tuesday.

(ii) To Weather Telex London, repeated Admiralty via Whitehall.

From C.-in-C., H.F.

Request weather route forecast from Firth of Forth to following aerodromes direct in each case :—Grantham, Catterick, for five hours commencing 0730 at 0730 to-morrow, Thursday.

Transmission of Forecasts.—In addition to the forecasts transmitted by W/T and incorporated in such messages as the Weather Shipping Bulletins, Fleet Synoptic Messages, etc., forecasts for the general public are issued in Great Britain, twice or three times daily by R/T. Aviation weather forecasts are also transmitted several times daily both in code and by R/T. In most countries of the world a similar procedure is followed. A complete list of times, frequencies used, areas affected and information issued in all countries is contained in the A.L.W.S., Vols. I and II.

Gale Warnings.

One of the first duties of the forecaster as soon as the observations are plotted on the charts is to examine the conditions to estimate the probability of strong winds which may be dangerous to shipping. In the majority of maritime countries arrangements exist whereby gale warnings may be issued to coastwise shipping when the wind is likely to increase to gale force. (The term " Gale " is used to denote a wind of force 8, or 34 knots and above, on the Beaufort scale). These gale warnings are conveyed in a number of ways as follows :—

(i) By visual signals.
(ii) By W/T.
(iii) By R/T.

In the British Isles gale warnings are initiated by the Meteorological Office and the information distributed through the co-operation of the Admiralty, the Post Office and the Board of Trade. (For details of gale warnings by W/T and R/T in all countries *see* A.L.W.S., Vols. I and II). They are also sent direct from the Meteorological Office to Staff Officers (Intelligence), and to the Commanding Officer, Coast of Scotland, who in turn forward them to the shore Naval authorities concerned, and to H.M. ships in port.

On receipt of a gale warning, storm cones are hoisted at certain points round the coast, so that they may be visible to shipping to indicate the probability of a gale being experienced in the near future. A description of the visual signals which serve as gale warnings in the British Isles, is given below :—

The warning signal by day consists of a cone 3 feet high by 3 feet wide at the base, and painted black. When it is hoisted

NORTH CONE.

By Day. **By Night.**

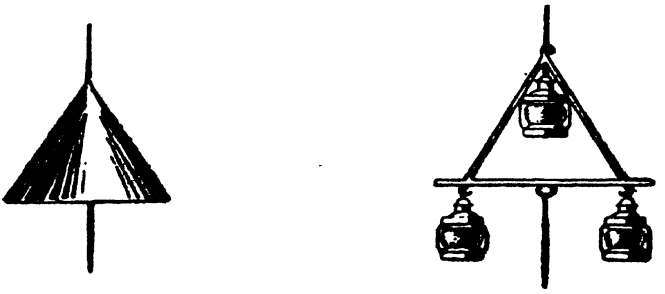


FIG. 103a.—NORTH CONE.

Hoisted for Gales.

From SE., E., or NE., backing to N.
,, NW., veering to N., NE. or E.
,, N., veering to NE. or E.
,, NE., veering to E.

SOUTH CONE.

By Day. **By Night.**

FIG. 103b.—SOUTH CONE.

Hoisted for Gales.

From SE., veering to SW., W. or NW.
,, SW., veering to W. or NW.
,, W., veering to NW.
And also from E., veering to S. or SW.

point upwards it is called a " North Cone " and indicates that the gale will commence from a northerly point. When the cone is hoisted point downwards it is called a " South Cone " and indicates that the gale will commence from a southerly point. For gales commencing from east or west the north cone is hoisted if the gale is expected to change to a northerly direction, and the south cone if it is expected to change to a southerly direction. These signals are also displayed by Fishery Protection vessels, the cone being hoisted inferior to the Fishery flag. By night, the warning takes the form of 3 lanterns, showing white or red lights, in a triangle of 4 foot base, but round the coast of the British Isles night signals are only displayed at 3 or 4 stations. Storm signals are hoisted when a gale is expected within a distance of 50 or 100 miles of the place where the signal is displayed, and within 10 hours of the time of hoisting. Depending on the rapidity of development of the gale, however, this period may be 20 or even 30 hours. The signal is lowered when the gale has passed and it is anticipated that there will be a period of not less than 12 hours free from gales or strong winds. The cone is left flying during a temporary abatement of the wind if a renewal of the gale is expected. The storm signals are illustrated in Fig. 103.

Warnings of Tropical Storms.—Although the generic term " cyclone " is applied to all regions of low barometric pressure, in the temperate latitudes such regions of low pressure are spoken of as depressions and the term cyclone is taken to refer only to a " tropical cyclone " (*see* Chapter XIX).

These storms are generally very intense and destructive in character and it is therefore desirable that areas affected should be given due warning of their approach. In the majority of countries adjacent to regions of cyclonic activity supplementary observing stations are opened during the months of greatest frequency, and there is an organization which deals with the issue of storm warnings. These warnings are transmitted by W/T and R/T either in plain language or in code (or both) and contain information regarding the position and intensity of the storm, the area affected, the probable direction of travel and the anticipated strength of the wind. A complete list of such warnings and information contained therein is given in the A.L.W.S., Vols. I and II.

Warnings of Ice Accretion.†—Reference to Chapter XI will show that the meteorological conditions leading to ice accretion are now fairly well known ; but before a meteorological officer can issue an ice warning he must have good evidence that these conditions exist. From his charts and surface observations he is occasionally fairly certain that there is danger of ice accretion ; but there are many more occasions when some of the conditions are known to exist but there is no information as to whether the others are present or

† From a paper " Ice accretion on Aircraft " by G. C. Simpson, K.C.B.,

not. For example, he may know that a certain cloud layer has a suitable temperature and may be fairly sure that the cloud is sufficiently dense, but he has no knowledge whether the cloud particles are in the solid state and therefore harmless or in the liquid state and therefore highly dangerous.

If he issued a warning on all occasions when ice accretion appeared possible, he would have so many failures that pilots would lose confidence in him. On the other hand, if he issued warnings occasionally (on those occasions when he is fairly certain of success) the pilot comes to rely on them, and when no warning is received concludes that the meteorological officer sees no reason for anticipating danger—in that way disaster lies.

For this reason the Meteorological Office has been very reluctant to undertake the regular issue of ice warnings ; but the great increase of cloud flying has made it imperative that the meteorologist should give what information he can, and so the following system of ice warning has been introduced :—

(*a*) When information has been received that ice accretion has been encountered, warnings in the following form are issued :—

" Ice formation in clouds from 2,000 feet to 5,500 feet ; little chance of ice below 1,000 feet and above 8,500 feet (area or route affected specified)."

(*b*) When from the general synoptic situation it is concluded that there is a risk of ice accretion, messages in the following form are issued :—

" Conditions favourable for ice formation in clouds between 2,500 feet and 6,000 feet ; little chance of icing above 9,000 feet or below 1,500 feet (area or route affected specified)."

In these messages the heights specified are determined by the temperatures of the atmosphere (observed or computed) on the assumptions that the greatest ice accretion occurs between 20° F. and 32° F. and that dangerous icing does not occur at temperatures below 10° F. nor at temperatures above 35° F.

In the British Isles when reports are received from pilots that they have encountered ice accretion the information is immediately made available to other pilots, by broadcasting it from the Borough Hill Meteorological Wireless Station and adding it to other weather reports. There are some disadvantages in this dissemination of actual observations of ice accretion, for it is the tendency for pilots to report all ice accretion whether or not it is of a dangerous nature. It is not easy to formulate rules as to the intensity of the ice formation which should be reported, for ice accretion occurs in many forms and the time factor is so important. Then again ice accretion varies so greatly in intensity and is of such a local nature that pilots cannot assume that they will meet with similar accretion in other parts of the same cloud layer nor even at the same place at some subsequent time.

CHAPTER XVIII.

Types of Pressure Distribution and their Associated Weather

Isobaric Forms.

At an early stage in the history of weather forecasting with the aid of synoptic charts it became apparent that the isobars in a given region usually assumed one of a number of characteristic forms. With each of these forms there was found to be associated a characteristic type of weather. In its details the weather varied upon occasions when the same isobaric form was present but its general characteristics were always similar. For example, whenever the isobars grouped themselves in closed curves about an area of high pressure the weather was always of a quiet, settled type. There are seven main isobaric forms which are enumerated below. A detailed description of each form and its associated weather is given later in this chapter.

1. Depression.
2. Anticyclone.
3. Secondary Depression.
4. Trough of Low Pressure or V-shaped Depression.
5. Wedge of High Pressure.
6. Straight Isobars.
7. Col.

Of these seven types the depression and the anticyclone are of fundamental importance. The remaining types are either modifications of these or arise as a result of their special arrangement.

Methods of Forecasting.

Weather forecasting with the aid of synoptic charts was first carried out in this country by Admiral Fitzroy in 1861, a systematic study of the subject having been prompted by the occurrence of a storm which wrecked the *Royal Charter* in 1859. The method developed gradually but up to 1918 its basis remained unchanged, namely, a knowledge of the existing pressure distribution and anticipation of its probable changes in the immediate future. By 1885, as a result of experience with synoptic charts, certain rules for forecasting had become established and they were, in that year, formulated and published by the Hon. Ralph Abercromby. The method of forecasting based upon the pressure distribution has accordingly come to be known as the Abercromby method. It rests upon two basic principles. The first is that each of the isobaric forms (see above) is associated with its characteristic type of weather. The second is that the isobaric forms travel, and, since they are accompanied by their characteristic weather,

the latter may be said to travel also. In practice the forecaster first identifies the areas of high and low pressure and other isobaric groupings on the synoptic chart. He is then in a position to say, roughly, what type of weather is occurring in any part of the region considered. The synoptic reports from ships and land stations supply the actual details. His next step is to anticipate the probable movement of the pressure systems. Often, however, the problem is complicated by the fact that depressions and anticyclones may appear within a given region not having travelled into it from outside. That is, they may form within the region; likewise they may disappear from the region. Weather forecasting by the Abercromby method therefore involves recognition of the isobaric forms with their associated weather, and anticipation of their formation, travel and disappearance.

The modern method of forecasting, introduced in Norway during the Great War of 1914-1918, and since developed, is known as the Polar Front method. The weather is regarded as characteristic of broad currents of air having either a polar or tropical origin, and especially of the boundaries or fronts between different air currents. The pressure distribution has not the prominence accorded to it by Abercromby, yet, since it may be regarded as controlling the flow of air streams through the relation expressed by Buys Ballot's law, its importance has scarcely diminished. Accordingly, it still affords the best approach to the subject of weather forecasting, so that the first part of the present chapter is devoted to consideration of the isobaric forms based upon the ideas of Abercromby.

The Depression and its Associated Pressure Types.

The Depression—General Description.—The depression is often connoted by the alternative names, Cyclone, Disturbance, Centre of Low Pressure or simply a Low. It appears on the synoptic chart as a series of isobars usually oval or circular in shape, surrounding an area of low pressure. Following Buys Ballot's law the wind circulates round the depression in an anti-clockwise sense in the northern and a clockwise sense in the southern hemisphere. A typical depression is shown in Fig. 104.*

* Figs. 104, 108, 109, 121, 122, 123 and 124 are plotted according to the following scheme:—Barometer: isobars are drawn for intervals of two millibars. Barometric change from 4h. to 7h. in tenths of millibars is entered beneath the temperature. Wind: arrows fly with wind. A full length feather indicates two steps on the Beaufort scale, and a short feather one step. Calm is indicated by circle outside weather symbol. Temperature is given in degrees F. Weather symbols: clear sky. sky less than 3/10 clouded. sky 4/10 to 6/10 clouded. sky 7/10 to 9/10 clouded. overcast sky. rain falling. fog. mist. haze.

Depressions fall into two main classes, those of temperate and high latitudes, and those of the tropical seas. Only the former class is dealt with in the present chapter, consideration being reserved until later for the tropical storm.

The sizes of individual depressions vary enormously, the smallest having diameters of the order of 50 miles and the largest 3,000 miles. The two terms, depth and intensity, are used to describe a depression, the former referring to the pressure at the centre, the latter to the steepness of the pressure gradients within it. No precise figures have been laid down, but a depression in which the pressure at the centre is less than 990 mb. may be called deep ; one in which the central pressure is above 1,010 mb. may be called shallow. An intense depression is one in any part of which there are steep pressure gradients ; one in which the gradients are slack is called a feeble depression. Deep depressions are usually intense and shallow depressions feeble, but not always so, since the pressure gradient between the centre and outer boundary of a depression depends not only upon the pressure at the centre but also upon the radius of the depression and the pressure beyond its outer boundary. A depression in which the central pressure is falling is said to be deepening ; one in which it is rising is said to be filling up. The life of a depression, that is, the time for which it appears as a distinct feature on the synoptic chart, is generally about 4 days but it may be as much as 20 days. A depression usually deepens during the earlier stages of its existence and fills up during the later stages, then often breaking up into a number of separate feeble centres. Sometimes, however, after a depression has begun to fill up it will deepen again. This process is called regeneration.

The Travel of Depressions.—Depressions move very irregularly and there is no path so tortuous that it cannot be regarded as a possible one. Nevertheless, the tracks have certain features in common, the most noteworthy of which is the general tendency for movement from a westerly to an easterly point. East to west movement is rare. It is possible in a vague way to specify certain tracks as being favoured by the depressions of a particular locality. Fig. 105 shows the tracks followed by the main depressions of the year 1935 in the neighbourhood of the British Isles. The two most common tracks are that skirting our north-west seaboard with its extension to Scandinavia, and the eastward track between Iceland and Norway. There is a tendency for depressions to move over the sea rather than land ; thus depressions sometimes move south-eastwards into the North Sea from the vicinity of Iceland, or eastwards along the English Channel and thence into the Baltic. Coastlines and mountain ranges appear to exert a steering influence. The general trend of the coast and mountains of Europe is from south-west to north-east and so is also that of the main tracks of depressions. In the North Atlantic the general direction of travel is east-north-east. In the North Pacific it is north-east

To face page 340

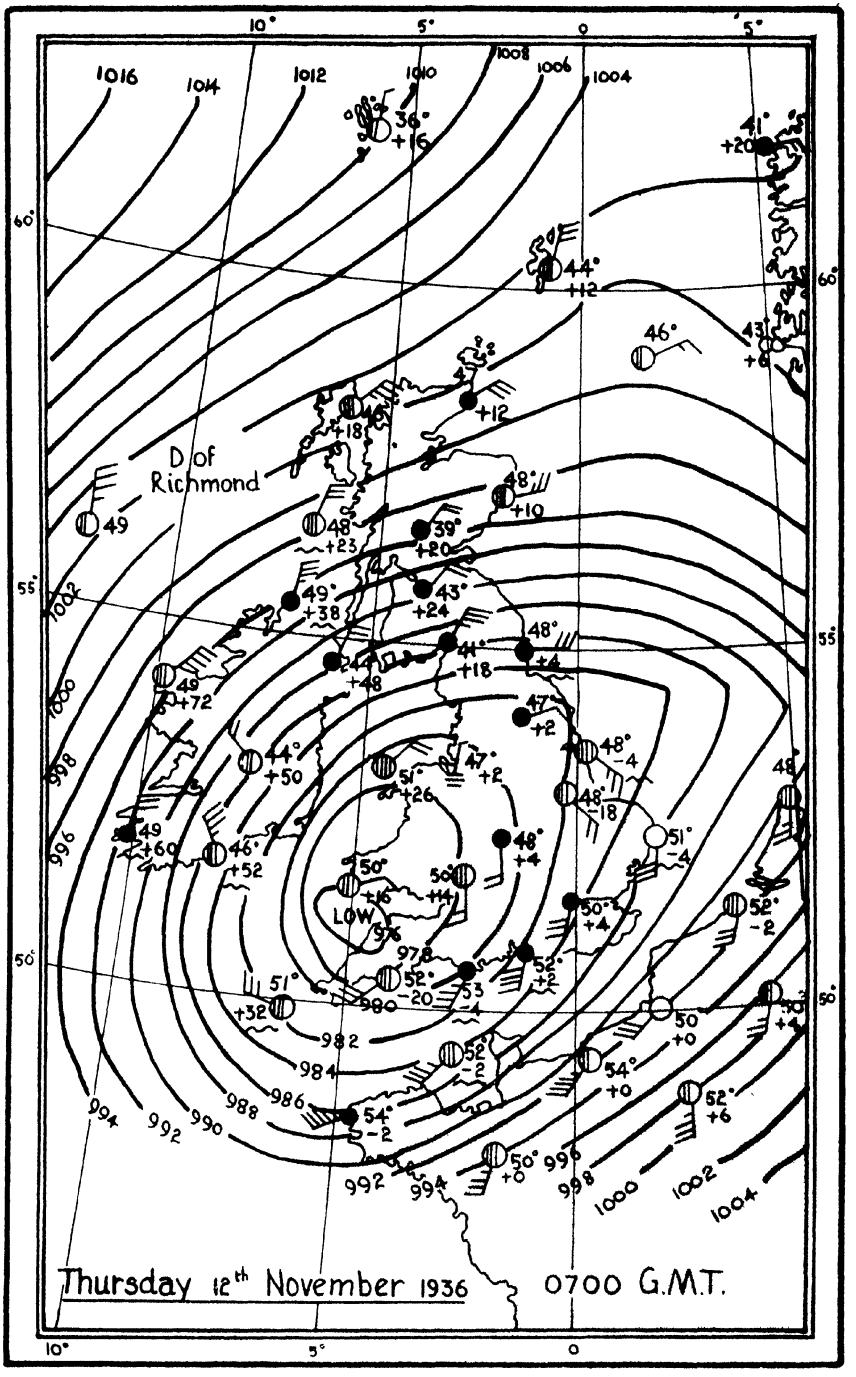


FIG. 104. DEPRESSION.

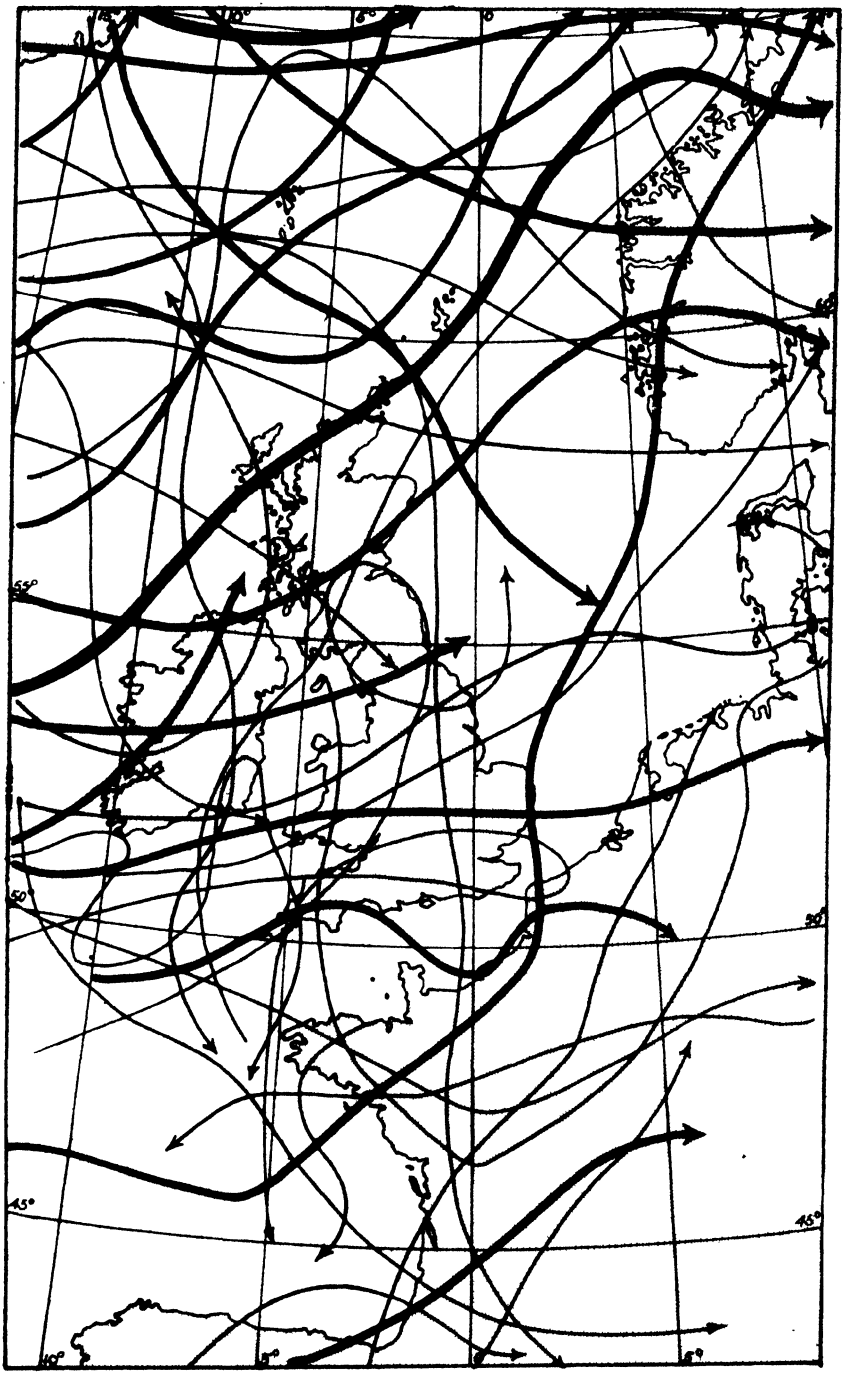


FIG. 105. TRACKS OF DEPRESSIONS. 1935.

The relative number of depressions following each track is indicated by the thickness of the line.

although on meeting the American coast with its high mountain ranges depressions sometimes turn south-eastwards. Over North America depressions enter the continent from Alaska, or form near the SE. coasts of the United States and move south-east and north-east respectively to the Gulf of St. Lawrence where the main tracks unite and pass into the Atlantic. In the southern hemisphere depressions move in a fairly regular sequence from west to east owing to the almost uninterrupted belt of ocean encircling the globe in the southern temperate latitudes.

The speed at which depressions move varies up to about 80 knots. In western Europe and the North Atlantic the average speed is 15 knots, in North America 25 knots, and in the North Pacific 20 knots. There is a tendency, especially in North America and the North Pacific, for speeds to be greater in winter than in summer. Sometimes depressions remain stationary for short or long periods. Small depressions tend to move rapidly ; large depressions, especially after they have begun to fill up, move slowly and often become stationary. Small depressions are often carried along bodily in the general air current in which they are situated. Large depressions themselves so modify the air currents around them that a general current independent of the depression cannot be said to exist.

The Distribution of Weather and Cloud in a Depression.—The straight line *AB*, Fig. 106, representing the instantaneous path

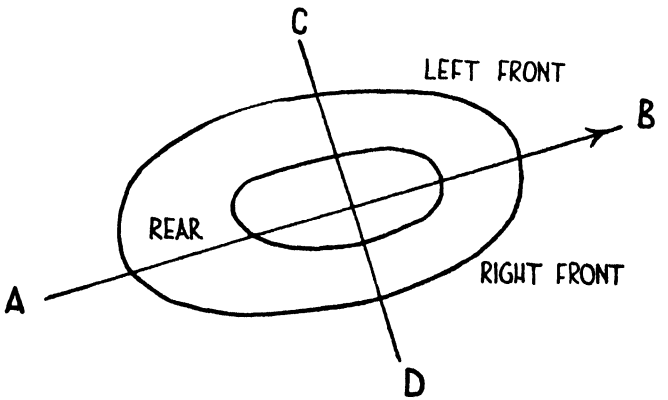


FIG. 106.—PARTS OF A DEPRESSION.

of a depression and the line *CD* drawn perpendicular to *AB* through the centre, called the trough line, divide the depression into four quadrants or parts. This division corresponds roughly to the distribution of cloud and weather in the depression. The two quadrants in advance of the trough line are called the right and left fronts respectively; those on the opposite side of the trough line are together known as the rear. In the northern hemisphere

the distribution of cloud and weather is roughly as follows. The right front is mild and cloudy with rain, the left front cold and dull with more persistent rain, and the rear showery, cool and invigorating, alternating between blue sky and cloud. In the southern hemisphere the distribution of weather as regards the right and left parts is reversed from that in the northern hemisphere.

The weather and cloud associated with the parts of a depression are shown in Fig. 107. In the front, some 500 miles

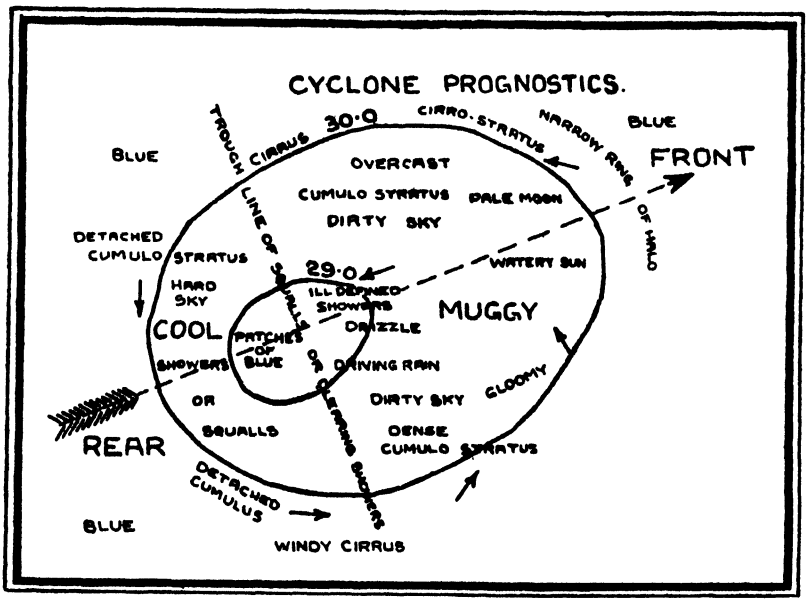


FIG. 107.—DISTRIBUTION OF WEATHER IN DEPRESSIONS (ABERCROMBY).

in advance of the centre, there is an outer ring of cirrus which thickens to cirrostratus on its inner side. Halos are often observed in the cirrostratus and are therefore regarded as prognostic of coming depressions. Within the cirrostratus there is a region of altostratus thickening towards the centre and becoming lower and merging into low nimbostratus in the central area. Rain usually falls from the thicker altostratus and the nimbostratus. In the northern hemisphere the wind is generally from south to south-east in the right front and the air is often muggy and oppressive. In north-west Europe the air has often moved over the ocean from the neighbourhood of the Azores, and is therefore mild and damp, particularly in winter. The oppressive nature of the air is often apparent even when the temperature is not high. The air in the left front is usually cold, especially in north-west Europe in winter, when it has been drawn from the cold continent. The cold easterly wind and driving persistent rain make the left front the most unpleasant quarter of the depression and in winter the precipitation

often takes the form of snow or sleet. In the rear of the depression the winds are generally from the north or north-west. The air is unstable to a greater or less degree, owing to heating at the surface as it moves to warmer latitudes. The cloud therefore is mostly cumulus or cumulonimbus with showers. Visibility is usually good. The cool invigorating winds are in striking contrast to the oppressive conditions of the right front.

There is a marked seasonal variation in the character of depressions. They are more frequent, deeper and more intense in winter than in summer, hence the greater frequency of gales in the former season. In winter, the clouds are usually indefinite and soft in outline and the precipitation, although persistent often occurs in the form of light rain or drizzle. In summer, skies are harder, cumulonimbus being frequent with copious rainfall in frequent and heavy showers. Shallow depressions tend to become stationary or slow moving in summer and then often develop thunderstorms over land areas. The weather and its association with the parts of these shallow depressions is ill-defined as are also the cloud types, which at a distance from the centre often assume the degenerate forms cirrocumulus and altocumulus instead of the more usual cirrostratus and altostratus.

Owing to friction, the wind at the surface flows not along the isobars, but across them towards the lower pressure at an angle which varies up to about 40°. This is known as incurvature. It follows that there is an inflow of air into a depression at all points along any one of its closed isobars. The air thus flowing into the depression can escape only by ascensional movement. There is thus good reason for anticipating the cloudy and rainy nature of the weather of a depression.

The Sequence of Weather Associated with the Passage of a Depression.—The sequence of weather experienced at a fixed station during the passage of a cyclonic depression differs according to whether the depression passes to the north or south of the station. In the northern hemisphere, depressions occur most frequently in about latitude 60°, and consequently pass to the north of the majority of stations. At a station south of the path, the first signs of the approaching disturbance are the backing of the wind to south or south-east, the barometer commencing to fall and the appearance of cirrus in the west and north-west. This gradually spreads over the whole sky, thickening to cirrostratus with possibly a halo round the sun or moon, and later to altostratus, giving the appearance known as "watery sky." Rain commences to fall from the altostratus, fitfully at first, but soon becoming continuous, and usually persists for several hours until the passage of the trough line. Near to the centre the altostratus merges into nimbostratus, which is frequently below 1,000 feet in altitude and sometimes down to sea level. It almost always envelops the higher ground. Meanwhile, the barometer continues to fall and the wind freshens.

perhaps increasing to a gale with little change of direction. At the trough line the barometer falls to its minimum and then begins to recover, sometimes abruptly. The amount of incurvature is often greater in the right front of the depression than in the rear, so that the change of wind at the trough line to west or north-west may be abrupt. Sometimes it is accompanied by a squall and a short period of heavy rain. The thick pall of nimbostratus soon breaks and the continuous rain ceases, giving place to showers from detached cumulus and cumulonimbus, while the medium and upper clouds rapidly disappear. Near to the centre the wind is often of gale force with violent gusts or squalls, but gradually decreases as the depression passes away eastwards, while the barometer continues to rise. There is often a sudden fall of temperature with the setting in of the cool winds on the passage of the trough. At a station to the north of the path there is a gradual transition from the dull rainy conditions of the left front to the cool, clear and showery conditions of the rear, the wind backing through east and north to north-west or west. The changes are nowhere abrupt. The same cloud sequence is observed as on the south side, but the rain is usually more persistent. If the centre of the depression passes over the station, the sequence of cloud and weather is the same as to the north of the track, but the wind falls light and variable as the centre passes, later setting in from the north-west.

The foregoing description applies to a fully developed depression of moderate or great intensity in the northern hemisphere. The sequence of cloud and weather is similar in the southern hemisphere. It should, however, be noted that what has been said above regarding the weather of the south side (right front) in the northern hemisphere applies to the north side (left front) in the southern, and *vice versa*, and also that the circulation of the wind is reversed.

The Secondary Depression.—The secondary is a small depression on the periphery of a larger primary or parent depression. If well developed it has its own system of closed isobars, but frequently is represented only as a bulge in the isobars of the primary. A typical secondary of the former type is shown centred off south-west Ireland in Fig. 108, (*see* footnote on page 339), the primary being situated in the neighbourhood of Iceland. The secondary is associated with cloud and rain, the weather being similar to that of an ordinary depression, but frequently more intense, with stronger winds, heavier rainfall and a more pronounced fall and rise of the barometer during its passage. Secondaries usually develop on the equatorial side of the primary and circulate around it in the general air stream at the same time participating in the general movement of the system as a whole. It is unusual, however, for the secondary, while it appears as such on the synoptic chart, to complete more than half a revolution round the primary, owing to the fact that the primary is often filling up, while the secondary

To face page 344.

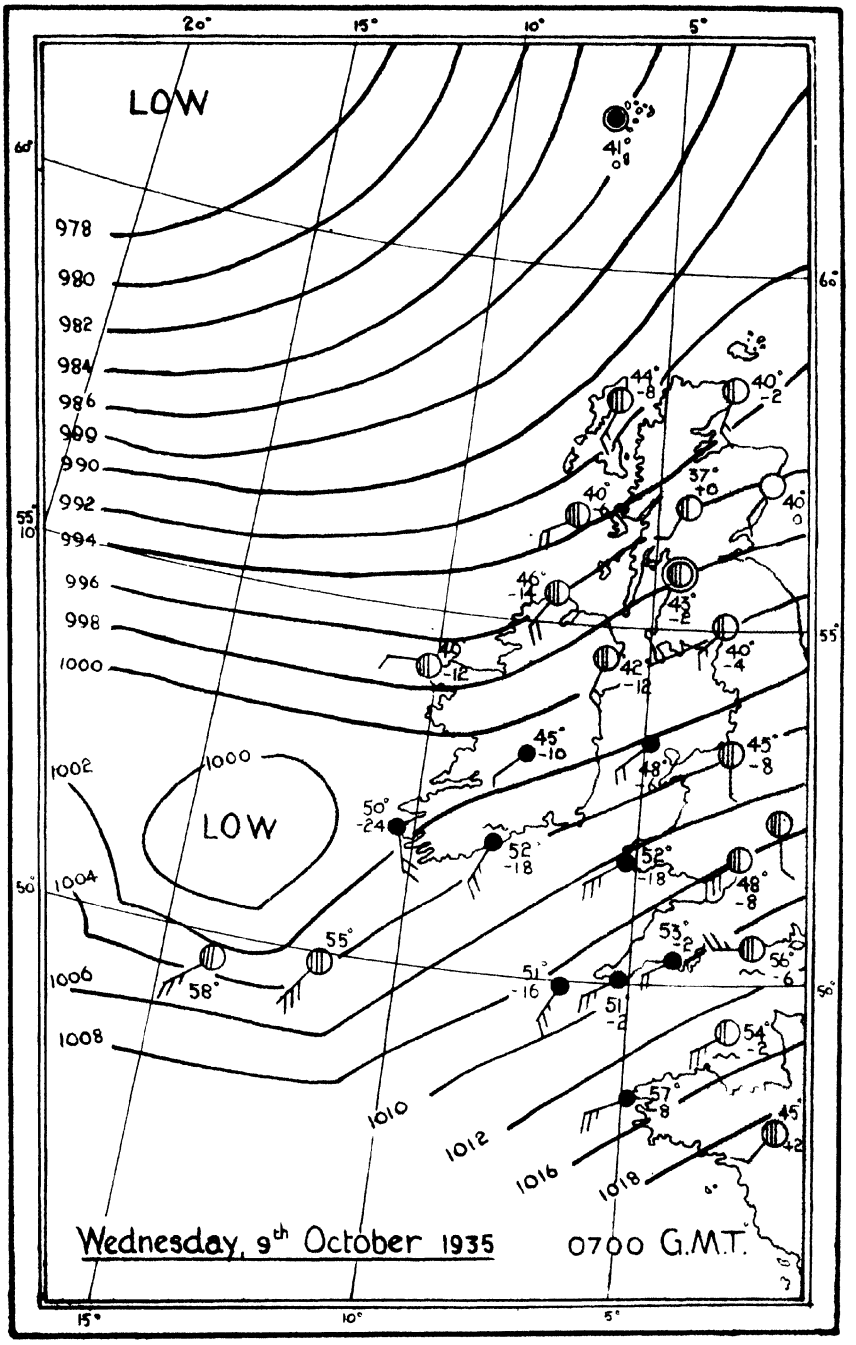


FIG. 108. SECONDARY DEPRESSION

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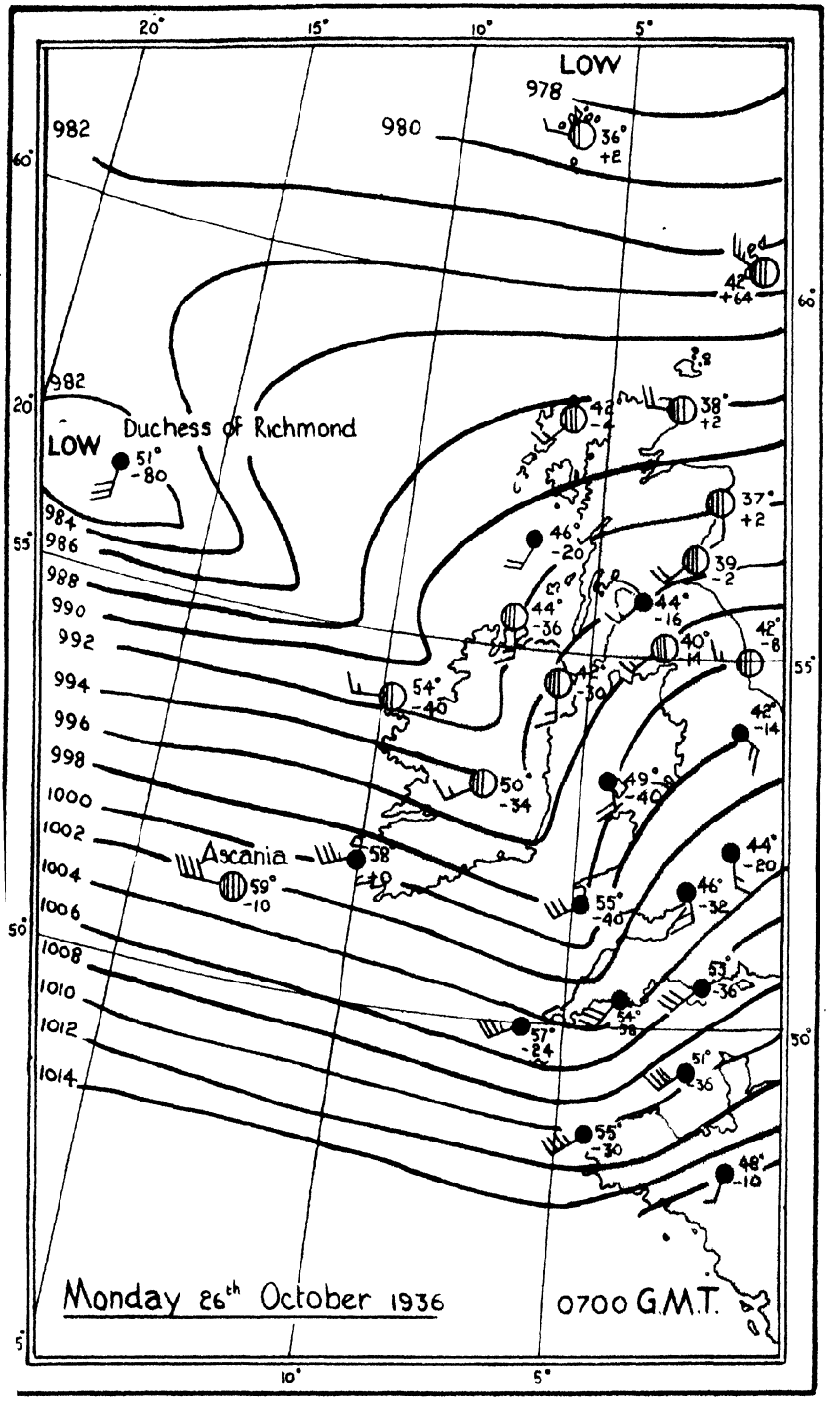


FIG. 109. TROUGH OF LOW PRESSURE.

is deepening, so that the latter becomes the dominating member of the system. Under these circumstances it may absorb the primary, or the latter may remain as a shallow area of low pressure indicated by a large bulge in the isobars of the former secondary. For this reason, it is often difficult from a single synoptic chart to distinguish between the primary and secondary depressions of a system, and it is therefore necessary to consult a sequence of charts. When a secondary attains approximately equal size, depth and intensity as its primary, the two together form a " dumb-bell-shaped low," and the centres then rotate anti-clockwise (in the northern hemisphere) about the point midway between them, at the same time both advancing with the speed and direction of motion of the system as a whole.

The region between a secondary and its primary is one of very slack barometric gradients, and consequently of light and variable winds, while on the side of the secondary remote from the primary the gradient is often steepened. One effect of this is that the appearance of a secondary may cause gales to occur at a greater distance from the primary centre than was anticipated, while light winds only may occur where gales were expected. Thus, secondaries moving eastwards on the south side of a depression to the north of Scotland sometimes cause gales in the English Channel and only light winds in the north of England where gales would normally have occurred.

Even those secondaries which are represented only by a bulge in the isobars may bring considerable rainfall and cloud to the area they traverse. Great care should therefore always be exercised in the drawing of isobars in order that secondaries of this type may be readily detected.

The Trough of Low Pressure or V-Shaped Depression.—This isobaric form consists of a series of isobars more or less V-shaped with low pressure within the " V." An example is shown in Fig. 109 (*see* footnote on page 339), extending from near the ship *Duchess of Richmond*, to the north-west of France. The trough may be regarded as an attenuated secondary, since it usually projects from a large and distant depression. The point of the " V " is often sharp, so that an abrupt veer of wind occurs at the trough. It is usually directed towards a southerly point in the northern hemisphere, and *vice versa*, and moves from a westerly to an easterly point. The winds and weather on its eastern side have the same characteristics as those in the right part of a depression, and the same cloud sequence, Ci., Cist., Ast., Nbst., occurs with rain as the trough approaches, while the barometer falls to a minimum at the passage of the trough, rising subsequently. In the rear of the trough, cool north-westerly winds occur with Cu. or Cunb. cloud, showers and good visibility. If the trough is sharp it moves with the speed and direction of the gradient wind in its rear.

THE POLAR-FRONT THEORY OF DEPRESSIONS.

The origin and development of the polar-front theory of depressions and the method of weather forecasting associated with this theory are largely due to the investigations of the Norwegian meteorologists V. and J. Bjerknes, father and son. The method has now been generally recognized for some years. With its adoption there began a new era in the science and practice of weather forecasting.

Origin and Development of a Depression.—The theory of the origin and development of depressions is based on the assumption that they are formed in regions where masses of air of widely different characteristics are brought into juxtaposition; the subsequent interaction between these masses is the cause of the development of the depression. In the following description of the development of an idealized depression of the northern hemisphere, the atmosphere is treated as being broadly divided into two types of air mass, one cold and the other warm. These two masses are termed "polar" and "tropical," as they normally originate in the polar and tropical regions respectively. Polar air is usually cold and dry, and is characterized by a large lapse-rate of temperature; tropical air, on the other hand, is warm and moist, while the lapse-rate of temperature, in the lower levels at least, is small. The two masses of air come into juxtaposition in the temperate regions where, before a depression develops, they are flowing in parallel streams but in opposite directions, in accordance with the general circulation of air in those regions (*see* Fig. 68); the polar air flows from an easterly or north-easterly direction on the northern side of the line of separation, while the tropical air is flowing from the opposite direction on the southern side (*see* Fig. 110a).

The boundary between these two air masses is a line of discontinuity with widely different conditions on its two sides. It is known as the "polar front." The surface of separation between the warm air and the cold air is not simply a vertical surface, but rises with a gentle slope towards the cold air. The slope is of the order of 1 in 50 to 1 in 200. The stability of this system is easily disturbed so that a wave may develop on the surface of separation, which is shown on the earth as a tongue of warm air projecting northwards into the polar air. This is the first stage in the life-cycle of a depression and is shown in Fig. 110b.

The next stage of development will be understood by reference to Fig. 110c.

The fall of pressure, which is in part due to the invasion of the cold air by the tongue of warm air, causes the cold air in front of the bulge at *A* to move from a southerly direction, *i.e.*, it is deflected to the right, and it circulates round the tip of the tongue of warm air, finally coming in from the north-west in the rear of the bulge at *B*. The flow of cold air is indicated in the figure by a single

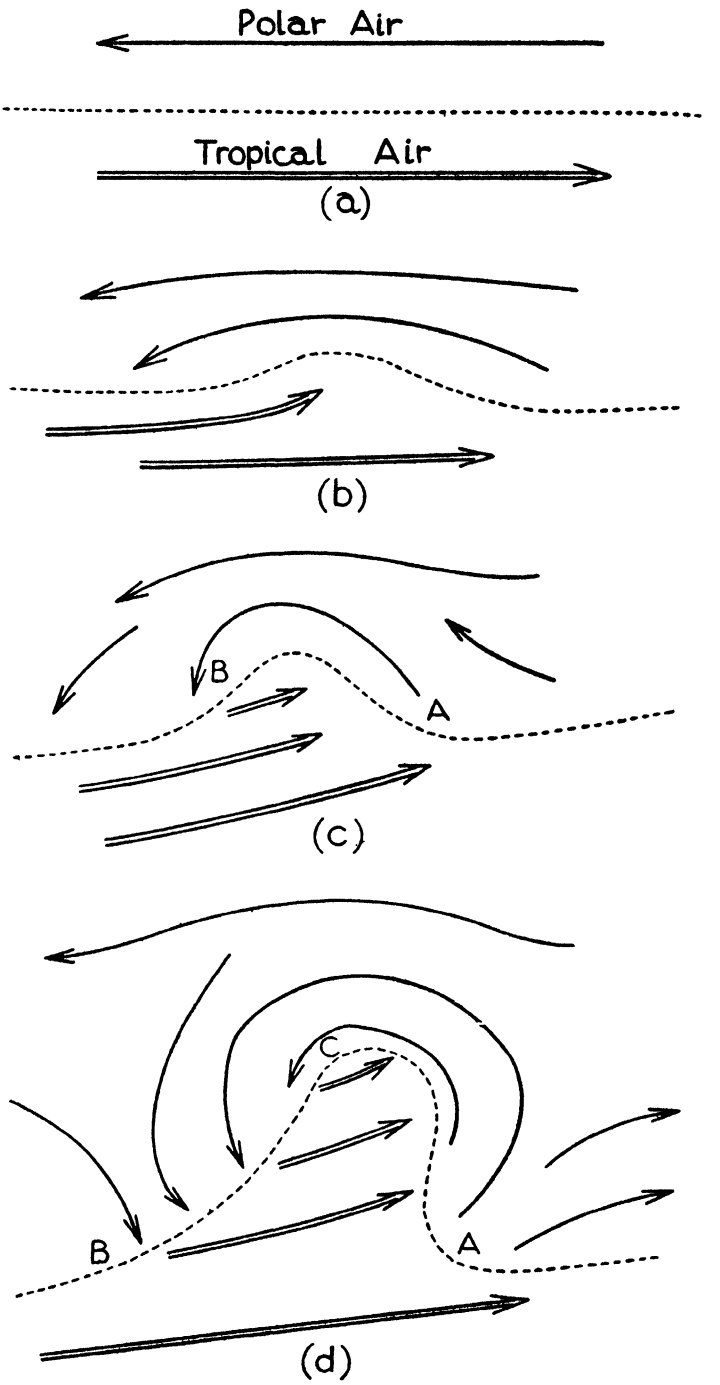


FIG. 110 (*a*) (*b*) (*c*) (*d*).—STAGES IN DEVELOPMENT OF A DEPRESSION ON THE POLAR FRONT.

line ; that of the warm air by double lines. The isobars begin to assume the form associated with depressions and the development of the disturbance continues until there is a complete cyclonic circulation of air.

In Fig. 110d, which illustrates an idealized depression in a more advanced stage of development, that part of the polar front marked *CA* represents the forward boundary between warm and cold air, while the part *CB* represents the rear boundary between cold and warm air at the earth's surface. *CA* is called the " warm front," *CB* the " cold front," and the area enclosed by the two is known as the " warm sector " of the depression, whose centre lies at the point *C*. An account of the movement of depressions will be given later in this discussion, but it may here be stated that a disturbance formed in this manner is not normally stationary, but travels along the polar front as a wave, in an easterly or north-easterly direction, carrying its weather conditions with it. Consequently, a stationary observer on the earth will experience a more or less definite sequence of weather conditions associated with those parts of the depression enumerated above as they pass over.

Fig. 111 shows, in ground plan and vertical section, the structure of an idealized depression of the northern hemisphere moving eastwards, the bottom diagram illustrating the weather conditions prevailing above the line *XY* to the south of the centre, the top diagram the conditions above the line *PQ* to the north of the centre. In the plan diagram the rain areas in advance of the warm front and behind the cold front are shaded. The weather conditions experienced along the line *XY* and appropriate to the different parts of the depression will now be discussed.

The Warm Front.—It will be seen that the air in the warm sector blows against the side of the cold air at the warm front and on account of its relative lightness will rise over the cold air ; the slope of the surface of discontinuity between the cold and warm air is about 1 in 100. The moisture in the warm air, as it ascends over the cold, will be subject to condensation, so that cloud and precipitation will form as the ascensional process goes on. In the advanced upper portion of the frontal zone, about 500 miles ahead of the warm front and at a height of approximately 5 miles, the originally warm air will manifest itself in the form of cirrus cloud, which in turn is followed by cirrostratus. Further down the surface of discontinuity, altostratus cloud appears and rain commences to fall. Behind this again comes nimbostratus or rain-cloud, which gives continuous rain, for in this region there is more moisture available for condensation ; this frontal rain, which covers an area from one hundred to two hundred miles wide, continues until the passage of the warm front. The barometer falls steadily as the lower surface of the warm air nears the ground, because cold air is being replaced by lighter warm air in the upper levels. The warm

front passes the observing station with a gradual transition from cold to warm air and a shift of wind. In certain regions thunderstorms may accompany a warm front, but such thunderstorms are rare in the vicinity of the British Isles.

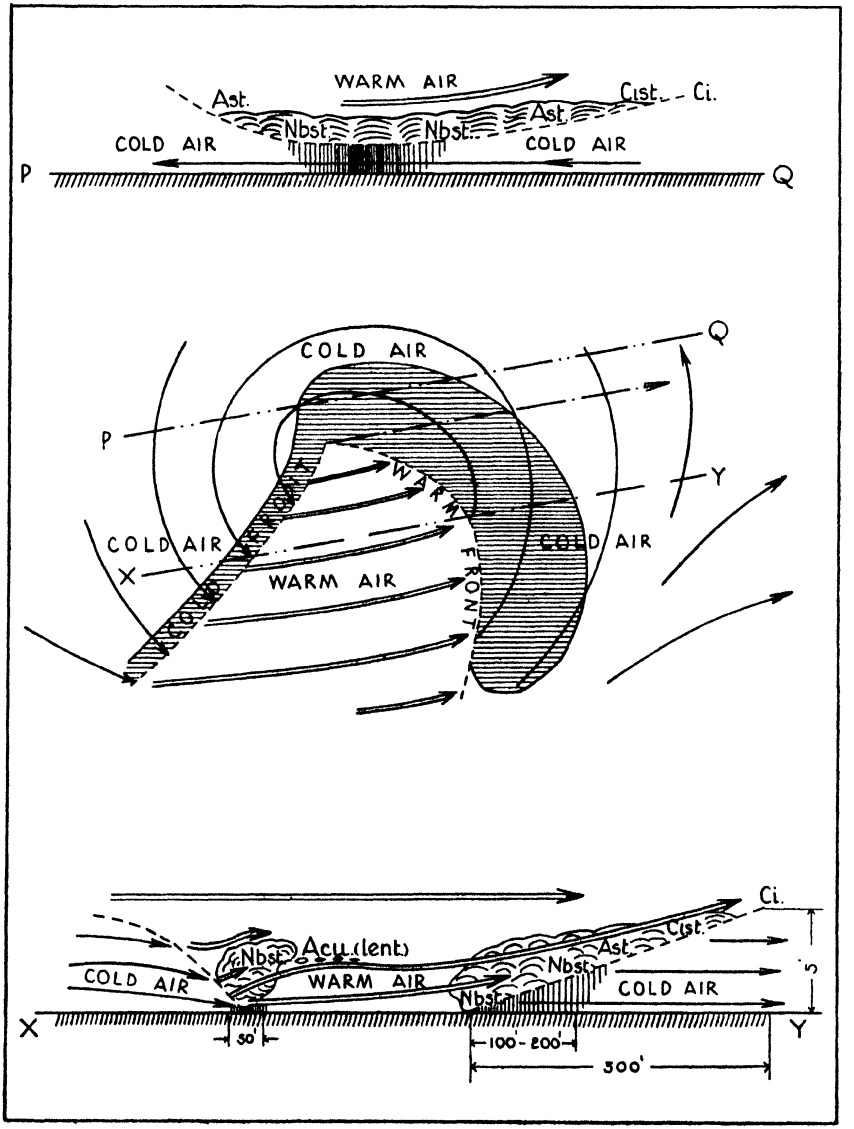


FIG. 111.—PLAN AND CROSS-SECTION OF A TYPICAL DEPRESSION (AFTER BJERKNES).

The Warm Sector.—In the central region, known as the warm sector, tropical air will be found both at the surface and in the upper layers. The air is warm and moist and the sky cloudy;

rain, if any, partakes of the character of drizzle, and will be due to orographic influences or to the turbulent uplift of the warm moist air as it encounters surface irregularities. The wind in the warm sector is steady in direction, but has a tendency to gustiness ; the barometer falls slowly or remains steady, depending on whether or not the depression is deepening or the centre of low pressure is approaching the observing station.

The Cold Front.—At the cold front, the cold dense air is undercutting the warm air, forcing it to ascend, the slope of the surface of discontinuity between the two air masses being of the order of 1 in 50. As this slope is much steeper than that experienced at the warm front, the ascensional movement of the warm air is more violent than in the case of the warm front. As a result, precipitation at the cold front is more in the nature of heavy rain lasting for a short time ; frequently, however, a period of rain is experienced in advance of the cold front. Under suitable conditions of temperature, humidity and wind discontinuity the violence of the upward current at a cold front may be such as to produce thunderstorms and line-squalls. As the cold front passes, a sharp fall in temperature and humidity is experienced at the observing station, together with a shift of wind and a rapid rise in the barometer, as the warm tropical air is replaced by the colder and drier air in the rear. Surface air at a cold front is frequently retarded by surface friction, causing it to be overrun by the cold air at a higher level (*see* detailed structure of a line-squall, Chapter XII). This would explain the sudden rise in the barometer and as the cold air probably tends to subside, the precipitation is of short duration.

The following is a summary of the sequence of weather conditions experienced in a depression along any line such as XY (Fig. 111).

(*a*) *Before the warm front :*

Wind backs to SE. or S. and increases gradually (veers to NE. or N. in southern hemisphere).
Barometer falls steadily.
Temperature increases slightly.
Clouds appear in the order :—cirrus, cirrostratus, altostratus, with rain commencing.
Altostratus merges into nimbostratus with continuous rain.
Relative humidity not high, but increases due to rain.
Visibility fairly good except in so far as it is affected by the rain.

(*b*) *At the warm front :*

Wind continues to increase and veers to SW. or W. (backs to NW. or W. in southern hemisphere).
Fall of the barometer is arrested or slackened.
Temperature increases.
Cloud is low but sky becomes lighter.
Rain ceases or falls as a light drizzle.
Relative humidity increases.
Visibility deteriorates.

(*c*) *In the warm sector :*
Wind from SW. or W. Gusty (NW. or W. in southern hemisphere).
Barometer stationary or falling slightly.
Temperature continues high. Air usually warmer than the sea.
Broken skies with drizzle. Overcast near the centre of the depression.
Relative humidity continues high.
Visibility moderate.

(*d*) *Before the cold front :*
Bank of thick cumulonimbus cloud sometimes appears to the westward with some altocumulus (lenticularis). Short period of pre-frontal rain.

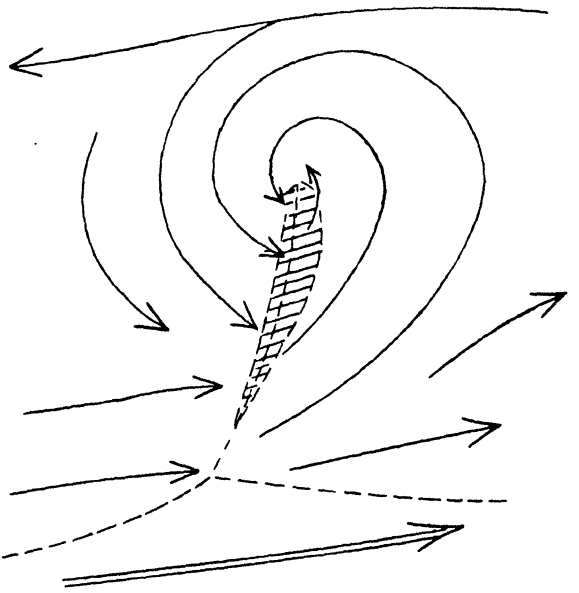
(*e*) *At the cold front :*
Wind veers suddenly to NW. Squally (backs to SW. in southern hemisphere).
Barometer begins to rise, possibly suddenly at first.
Temperature falls.
Narrow belt of heavy rain, mostly behind the front ; nimbo-stratus cloud ; possibly thunder and hail with cumulonimbus cloud followed by stratocumulus.
Relative humidity may increase temporarily due to rain : thereafter it decreases.
Visibility improves.

(*f*) *Behind the cold front :*
Wind between W. and N. Squally (between W. and S. in southern hemisphere).
Barometer continues to rise.
Temperature low : air usually colder than the sea.
Sky clears apart from detached cumulus and cumulonimbus clouds.
Showers ; sometimes thunderstorms.
Relative humidity low.
Visibility becomes very good after the passage of the front, but it may be very bad at the front itself.

To the north of the depression, *i.e.*, along the line *PQ* (Fig. 111), the tongue of warm air protrudes into the upper levels and the sequence of weather conditions will be somewhat similar as regards cloud and rain ; the wind will back gradually from south-east to north, while the barometer will fall slowly in advance of the depression and rise slowly in the rear. The fronts, however, do not appear at the observing station and consequently there is no warm sector apparent there.

The Occlusion.—It has already been stated that a depression moves along the polar front, generally in an easterly or north-easterly direction. The warm front and the cold front which are

integral parts of the depression will therefore move in this general direction. The rate of advance of both these fronts depends on the speed of the cold air. The cold air in advance of the warm front must be moving more slowly than the warm air behind it, otherwise the warm air would not move up the frontal surface and there would be no continuous production of cloud and rain. The cold air behind the cold front must, on the other hand, be moving more rapidly than the warm air in advance of it; otherwise the cold air would not undercut the warm air and produce the cloud and heavy rain of the cold front. Thus the cold front travels forward more quickly than the warm front, for there is no substantial change of wind speed in the warm sector. As a result, the cold front catches up on the warm front until the two coincide, while the warm air is gradually lifted off the ground. The line of coincidence thus established is called an occlusion (*see* Fig. 112).



The cold front will usually catch up the warm front, first, near the centre of the depression, so that the occlusion generally starts near the centre and extends outwards. It usually assumes a curled shape round the north and east sides of the depression and divides into a warm and a cold front at a point which gradually increases in distance from the centre. Many of the depressions which reach the British Isles from the Atlantic are occluded. If the cold air were homogeneous, there would be no difference at the earth's surface on the two sides of a line of occlusion. Generally, however, the air in front of an occlusion differs considerably from that in the rear, owing to the difference in the life-history of the two masses.

If the air mass behind the occlusion is the colder, the occlusion is of the cold-front type; if warmer it is of the warm-front type. Consequently, the weather in the vicinity of an occlusion partakes of the character of the weather associated with either a cold front or a warm front according to the type of occlusion.

Fig. 113 shows the weather conditions prevailing above an

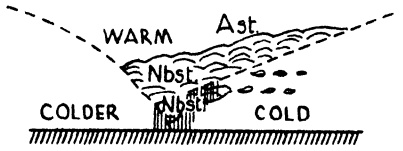


FIG. 113.—OCCLUSION (cold type).

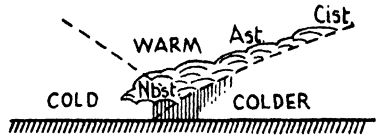


FIG. 114.—OCCLUSION (warm type).

occlusion of the cold-front type; Fig. 114 the conditions above an occlusion of the warm-front type. Of the two, the former is the more common. The first type is preceded by the cloud of the lifted warm front, while at the front itself there is a short period of heavier rain analogous to that of a cold front. In the other type the rain is of the warm front character, following on a similar sequence of cloud. Intermediate between the two types of occlusion there is the case where no marked contrast in temperature exists between the two air masses.

To summarize the conditions prevailing at an occlusion :—

The phenomena to a certain extent are a combination of those associated with cold and warm fronts.

A cold front and an occlusion are distinguished from one another by the associated cloud types and width of the rain belt.

Barometric tendencies do not serve to distinguish very clearly between a cold front and an occlusion, but the barometer is said to fall more rapidly before an occlusion than before a cold front.

There should be no difficulty in distinguishing between a warm front and an occlusion as the former has warm air behind it, while the latter has polar air behind it.

Relative humidities on both sides of an occlusion are lower than in warm air (humidity should be measured well away from the line of occlusion, to avoid effects of precipitation).

Visibility on both sides of a line of occlusion is higher than in warm air.

While the weather phenomena discussed above both in connection with fronts and occlusions are those associated with the idealized depression, it must be borne in mind that the ideal type is not always encountered, in practice. Modifications in the physical processes involved, and therefore in the weather conditions themselves, occur, according as the co-existent air masses vary in their

internal characteristics, while other factors such as the effect of land and sea, instability, etc., largely control the ascensional activity of the warm air. Consequently all the phenomena enumerated in the foregoing discussion do not always manifest themselves, and in the prediction of frontal phenomena the forecaster must be guided by careful analysis of the air masses involved, from the synoptic data available, and by such information as his experience has taught him.

Source of Energy in a Depression.—The energy of a moving depression is derived from the potential energy of the two adjacent masses of air. Owing to their difference in temperature they cannot remain indifferent to each other. The colder air will penetrate beneath the warmer air, causing it to rise, and the general tendency will be for the warm air everywhere to overlie the cold air. In the development of a depression, this tendency results in a decrease of potential energy which reappears as the kinetic energy of the winds in the cyclonic circulation associated with the depression. So long as there is a warm sector there is present a store of potential energy and the depression will deepen. After occlusion, however, the source of potential energy disappears, the kinetic energy imparted to the air is dissipated by surface friction and turbulence, and the depression weakens or "fills up." An incipient disturbance increases its speed as it develops, finally becoming slow-moving again as its life-cycle draws to a close.

Back-bent Occlusion.—It is not infrequent for the cold front of a depression to overtake the warm front at a comparatively early stage in its life-cycle. The occlusion formed in this way starts originally from the centre of the depression, but as the depression progresses, the centre frequently moves along the line of occlusion towards the tip of the remaining warm sector; the occlusion, as it were, bends back behind the centre of the depression and forms a secondary front, usually a secondary cold front. Such a formation may be understood by reference to Figs. 115a, b and c, which show a back-bent occlusion in the various stages of its development.

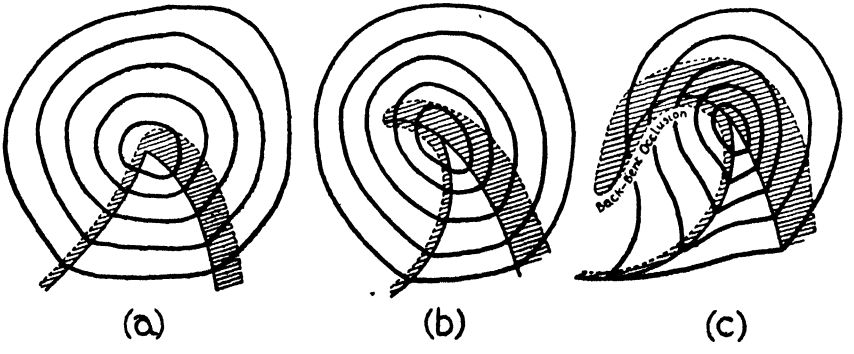


FIG. 115.—STAGES IN DEVELOPMENT OF BACK-BENT OCCLUSION.

As will be seen, this secondary front is usually of limited length, but it is of considerable importance for it advances behind the cold front and brings, after the precipitation of the cold front itself has passed, a new region of low cloud and precipitation, instead of the fair, bright weather which would otherwise be expected. The precipitation, indeed, may be quite abundant and of several hours' duration, especially in the rear of the occlusion and in that part of it nearest the centre, but the width of the rain band decreases towards the outer end of the formation. A back-bent occlusion is frequently characterized by marked deformation, so that it is not necessarily of the same shape as the cold front which precedes it. Furthermore, there is in such a case, a tendency for the associated depression to be strengthened or " regenerated " by reason of its re-acquiring a warm sector due to the introduction of a fresh supply of warm air into the area between the cold front and the occlusion, or by a fresh supply of polar air in the rear of the occlusion. By this means, new energy is supplied to the depression, so that it tends to increase in intensity until it again becomes occluded.

Secondary Cold Front Due to Subsidence.—There is another way in which a secondary cold front may be produced in the rear of the primary cold front of a depression, viz., by the descent of subsidence of the cold air in the neighbourhood of the surface or separation between the cold and the warm air. This cold air, as it subsides, is warmed by compression, and although colder near the surface than the air in the warm sector, is nevertheless warmer than the horizontal moving cold air in the rear of the cold front. Thus there sometimes appears at the earth's surface, the primary cold front dividing this descending air from the warm sector, followed by a secondary cold front dividing the cold surface air from this descending air. The distance between the primary and the secondary cold front may be anything up to 100 miles, but is generally much less. A phenomenon of this nature will result in a diffusion of the characteristics of the cold front over a greater horizontal distance, and, naturally also, in some modification of the frontal phenomena. This development appears most likely to arise with fast moving cold fronts. Sometimes it results not in two distinct and separate cold fronts, but in a diffuse front, which takes the form of a zone of gradual change instead of a surface of discontinuity. The formation of a secondary cold front in this manner is shown in Fig. 116.

Skagerrak Depression.—There are some regions in which a depression tends to form at the tip of the residual warm sector of an existing occluded, or partially occluded, depression. The phenomenon is in part orographic, due to the warm front being held up by mountain ranges, and is particularly liable to happen over the Skagerrak when a depression centre passes to the north

of Scandinavia. Such a formation is known as a Skagerrak depression.

Closely associated with this phenomenon is the process known as "seclusion." The delay in the progress of the warm front occurs well to the south of the original depression centre, so that the cold

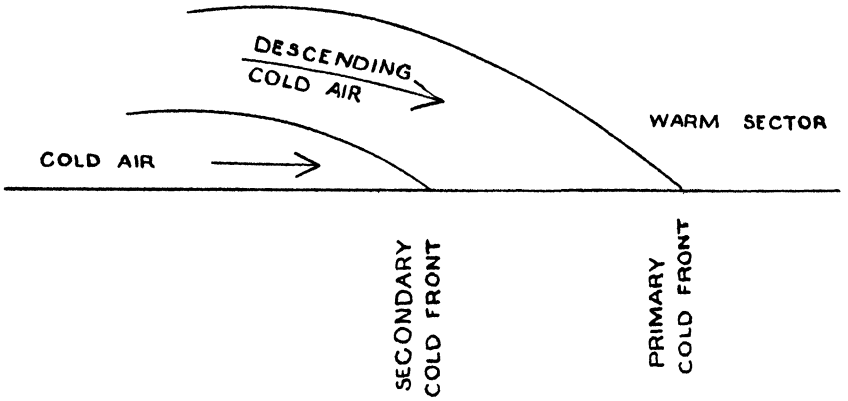


FIG. 116.—SECONDARY COLD FRONT DUE TO SUBSISTENCE.

and warm fronts coalesce to form an occlusion in that region, while there may still remain a warm sector to the northward, in which the warm air is "secluded," and which may subsequently be divorced from the secondary depression which has formed to the south.

Fig. 117 shows the formation of a Skagerrak depression simultaneously with the seclusion of the parent depression to the north.

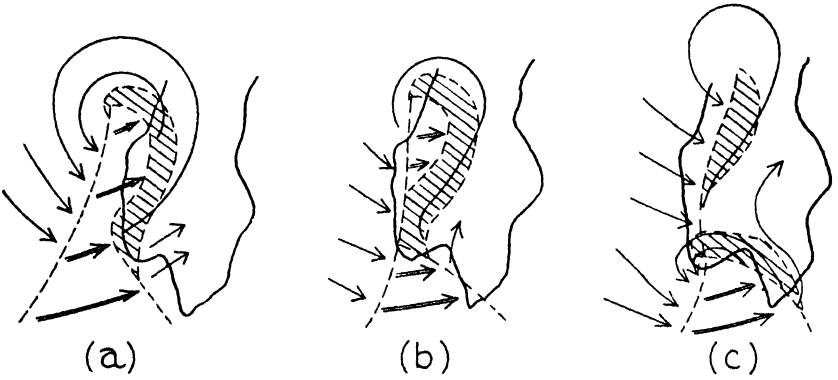


FIG. 117.—SUCCESSIVE STAGES IN THE FORMATION OF A SKAGERRAK DEPRESSION.

Depression Families.—In the foregoing description of the life-cycle of an ideal depression of the temperate regions, the disturbance has been treated as an independent entity which develops and travels along the polar front, while the phenomena of the

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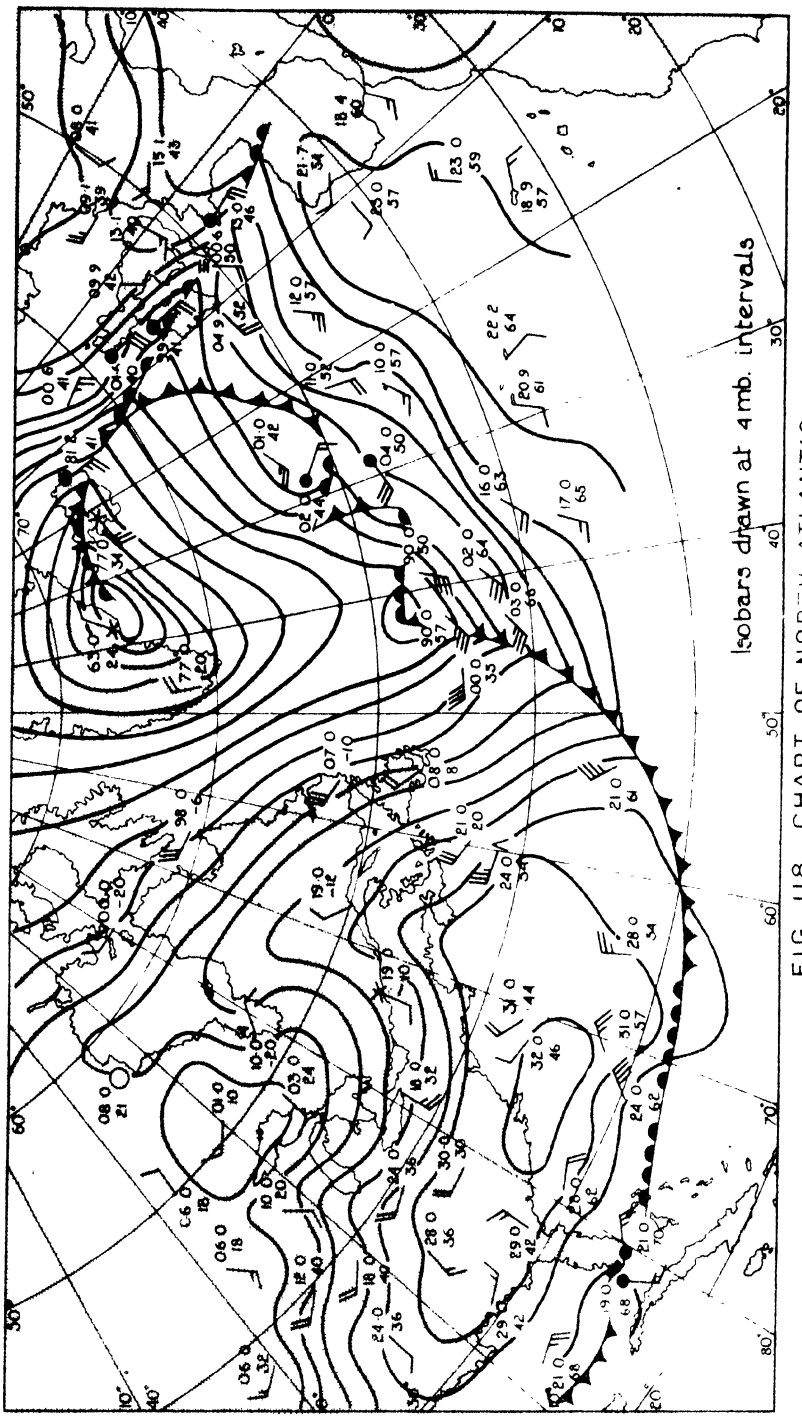

FIG 118 CHART OF NORTH ATLANTIC
SHOWING SERIES OF DEPRESSIONS ALONG THE POLAR FRONT

" back-bent " occlusion and the secondary cold front which frequently accompany the depression (together with the particular case of the Skagerrak depression) have also been discussed. The development of a depression, as we have seen, is due to the interaction of polar and tropical air masses, so that it is reasonable to suppose that such a development may take place wherever these air masses are brought into juxtaposition, *i.e.*, anywhere along the polar front. According to this theory, then, the polar front may be pictured as a line of discontinuity with depressions lying at intervals along it, like beads upon a string.

Fig. 118, which is drawn for the particular case of the northern hemisphere, shows the polar front through a series of depressions of different ages, and on the original theory this could be traced right round the hemisphere ; but investigations show that breaks occur from time to time in the polar front, where polar air penetrates into the trade-wind belts, and this has led to the conception of " depression families." Such families, consisting of 4 to 6 individual depressions, appear as a procession of depressions along the polar front, each fresh depression forming further to the rear, *i.e.*, to the westward along the trailing cold front of its predecessor. It will be seen from Fig. 118 that the polar front does not in general lie along a parallel of latitude, but tends to follow a south-west to north-east orientation ; as a result, in the general west to east movement of the polar front as a whole, each successive depression passes somewhat further to the south than the preceding one. Furthermore, in summer, the polar front has a tendency to recede polewards, so that the tracks of depression centres lie in lower latitudes in winter than they do in summer. As the procession advances along the polar front, depressions die out at one end and new ones appear at the other ; finally, an anticyclone forms in the rear of the last member as the polar air current bursts through. The most easterly members of the family as they pass the observer are of the dying type, while subsequent members are usually at an earlier stage of development and show increasing intensity as they advance. Further, the number of depressions which crosses any given meridian may exceed the number in the family as represented on a synoptic chart at a given instant, for new members may be born during the passage of the older ones across the meridian. The average life of an individual depression is about a week, while that of a family would last theoretically for ever, its existence being maintained by the birth of new members as the older ones die. Actually, investigations have shown that the average time of passage of each family over a given meridian is about 5½ days.

Fig. 119, which is a representation of the general circulation of wind in extra-tropical regions of the northern hemisphere, shows a series of four families of depressions, each family being separated by a gap where the polar current outburst above referred to breaks through to reinforce the trade winds, with the subsequent develop-

ment of an anticyclone (*H* in Fig. 119). It will be seen from the figure that when this transport of cold air from the polar to the tropical regions has been effected, the polar front, which has had a tendency to trail into lower latitudes, reforms far to the northward, owing to warm air moving to higher latitudes on the western side of the anticyclone. Consequently, the first members of the next

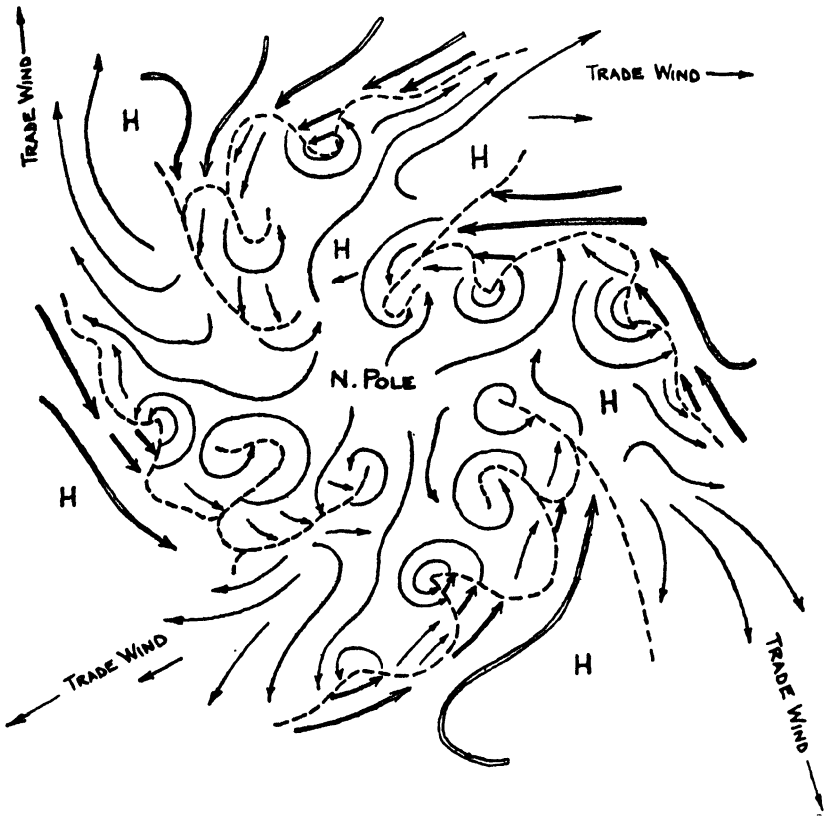

FIG. 119.—DEPRESSION FAMILIES IN GENERAL CIRCULATION OF NORTHERN HEMISPHERE.

family of depressions will usually appear on a more northerly track than that taken by the last members of the previous family, and the entire family will follow a new polar front which is not connected with its predecessor. The whole system is supposed to have a counter-clockwise rotation round the pole, corresponding to the eastward motion of the families themselves and the anticyclones separating them. Individual members of a family of depressions will move faster than the general system, for they have a two-fold movement—that of the family as a whole, and, in addition, a movement of their own along the polar front on which the family is formed.

Practical Aspect of the Polar Front.—In the foregoing account of the polar-front theory of depressions, an attempt has been made to explain the origin and development of these phenomena, together with the physical processes involved in the formation of fronts and their characteristics. It has been stated that the ideal type of depression is presumed to develop at the boundary between a west-bound air mass of polar origin and an east-bound air mass of tropical origin. Such conditions, however, would imply the existence of large areas covered by straight isobars, a state of affairs seldom if ever experienced in those regions of the globe where low-pressure systems are encountered. Actually, investigations have shown that a disturbance is more likely to develop between two air masses of different physical characteristics moving in the same direction, viz., towards the east in both hemispheres, so that initially there is no east wind and east winds only develop as the depression develops, as part of the cyclonic circulation of air round the centre of low pressure. Such a development is illustrated in Fig. 120.

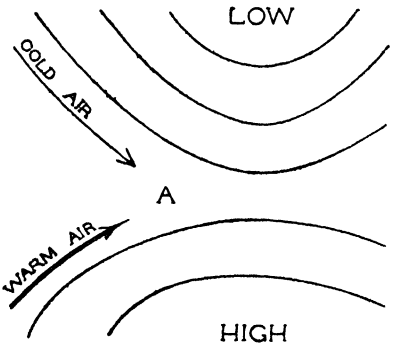

FIG. 120.—DEVELOPMENT OF A DEPRESSION BETWEEN TWO WESTERLY AIR CURRENTS OF DIFFERENT ORIGINS.

The synoptic situation shown, presupposes the existence of an anticyclone to the south of and adjacent to an already existent depression. In the north-west quadrant of the anticyclone the direction of circulation is such as to bring warm air towards the area marked *A*, while that in the south-west quadrant of the "low" brings cold air towards the same area. The co-existence of these two air masses of different characteristics may result in the formation of a new depression at *A*, which subsequently travels along in the westerly current between the original "high" and "low." Any such synoptic situation which reveals the existence of a polar westerly current adjacent to and to poleward of a sub-tropical westerly current, is recognized as an almost unmistakable sign of the development of a young deepening depression; such a develop-

ment is usually characterized by a marked slackening of the pressure gradient in the area concerned.

In the ideal depression referred to in the treatment of the polar-front theory, the classification of air has been based on the distinction between air originating from the polar regions and air originating from the equatorial zone. The nature of the life history of any air mass, however, may have a marked influence on its physical characteristics, and in many cases it is extremely difficult to trace back air to either of these two areas, so that a further classification of air must be introduced. The subject of air masses will be dealt with in Chapter XX, but it should be noted here that depressions may form at the boundary between any two air masses whose properties, notably temperature and humidity, differ, although the origin of such masses may be somewhat obscure. Furthermore, discontinuities may and do occur within air masses which at first sight appear homogeneous; this is particularly the case with air of polar origin, in which fronts may be detected, many of them of a minor or local character, but none the less associated with cloud and weather phenomena of a typical and persistent nature, while some of them are substantial enough to originate relatively large disturbances. It is the modification in the original characteristics of the separate air masses and the effect of such changes on the associated weather conditions, that make frontal analysis from synoptic data the complex problem that it undoubtedly is, for in many cases the discontinuities may be smoothed out to such a degree as to be almost unrecognizable, and only by the equally complicated process of tracing back the source of the air on earlier charts can their position be determined.

In practice, the problem which confronts the forecaster, resolves itself into the determination of the outline of the polar front (or part of it) on a synoptic chart, and the prediction of the subsequent movement of depressions and their associated fronts.

In spite of their irregularity it has been found possible to formulate certain practical working rules, the judicious application of some or all of which will enable the forecaster to predict the movement of depressions with some measure of success, once their position has been established on a synoptic chart. These rules are given in the chapter on general forecasting. A similar set of rules governing the movement of fronts and occlusions as separate entities, is also given in the same chapter. From these considerations the forecaster is able to predict the weather conditions likely to be experienced over an area traversed by fronts and occlusions, for, as we have seen, such weather conditions are due to the physical processes that go on in frontal zones, and they in turn are carried along as the fronts develop and advance. In common with the rules governing the movements of depressions, these rules are by no means infallible, and the forecaster must be guided largely by his own experience when applying them.

The Anticyclone and its Associated Pressure Types.

The Anticyclone.—The anticyclone appears on the synoptic chart as a series of closed curves surrounding an area of high pressure. Following Buys Ballot's law, the wind circulates round the anticyclone in a clockwise sense in the northern and an anticlockwise sense in the southern hemisphere. At the surface, owing to friction, the wind has a component outwards across the isobars, and there is thus a leakage of air from the anticyclone. If the anticyclone is to persist this loss must be made good, which can be accomplished only by air flowing into the anticyclone at high levels and descending. The descent takes the form of a slow settling down of the air over the anticyclone—the process known as subsidence. Sir Napier Shaw has estimated that in a large anticyclone the velocity of the descending air currents is of the order of 300 feet per day. In most developing anticyclones it may be 3,000 feet per day. The descending air is warmed dynamically, as a result of which precipitation very rarely occurs in anticyclones.

In the centre of a well-developed anticyclone the barometric pressure reduced to M.S.L. usually exceeds 1,030 mb. In the great anticyclone which forms over Siberia in winter it sometimes rises to 1,050 mb. An anticyclone in which the pressure is very high at the centre is said to be intense ; this term does not, as in the case of the depression, refer to the barometric gradient, but to the actual pressure at the centre. It has been seen on page 175 that near the centre of an anticyclone the pressure gradient is slight. It follows that an intense anticyclone must be large. On the average, anticyclones cover larger areas than depressions. Small anticyclones, owing to their slack pressure gradients and light winds, are comparatively insignificant features of the synoptic chart, and in this respect differ from depressions, the smallest of which are often the most active. An anticyclone in which the pressure is rising at the centre is said to be building up or increasing in intensity ; one in which it is falling is said to be giving way, or decreasing in intensity.

Anticyclones are of two main types. The first type is the semi-permanent or seasonal stationary anticyclone such as forms over the oceans within the sub-tropical high-pressure belts, and over the continents in winter, as for example the great Siberian winter anticyclone. The second is the migratory type, which at intervals appears in temperate latitudes interrupting the normal cyclonic activity. The movement of an anticyclone of this type is usually slower than that of depressions, and is an irregular drift rather than definite movement. Sometimes an anticyclone remains stationary for many days.

It is sometimes erroneously stated that anticyclones are always accompanied by fine weather. It is true that the weather is usually dry and quiet, but otherwise almost any conditions may be experienced. From the point of view of weather, anticyclones may be

divided into two classes, those accompanied by clear skies and those accompanied by overcast skies. The former are more frequent in summer, the latter in winter. An example of a summer anticyclone with centre situated over southern England is shown in Fig. 121, (*see* footnote, page 339). At most stations within the isobar of 1,026 mb. there is little or no cloud. The weather of this type of anticyclone is known as radiation weather, since it is largely controlled by the solar and terrestrial radiation to which the light winds and clear skies give full play. Subsidence leads to high upper-air temperature and, in so far as it warms the air, to the disappearance of cloud. It frequently results in the formation of an inversion of temperature at some 2,000–3,000 feet above the earth's surface. In summer, day temperatures rise to high levels over land owing to solar heating, but since the upper air is stable, no convection cloud, or only a few small cumuli are formed. The high temperatures are not in evidence in Fig. 121 because that is an early morning (0700 G.M.T.) chart. The nights are cool, since the earth's surface loses heat freely to the clear sky, and ground mists or fogs frequently occur but are dissipated shortly after sunrise. During the day visibility is often only moderate, owing to the accumulation of dust haze below the temperature inversion—the turbid atmosphere of dry weather.

If a clear sky anticyclone occurs in winter there is sharp frost at night owing to the prolonged and intense radiational cooling. Fog frequently forms over land with a pronounced inversion of temperature at the surface. Often the solar heating disperses the fog during the forenoon, but occasionally the fog persists throughout the day, and then the day temperature rises little above its low night level.

Often a layer of cloud forms below the subsidence inversion and then the anticyclone becomes characteristic of the second type. The weather may remain dull, cold and foggy or misty for days on end, and in the interiors of the continents even for weeks. Above the cloud layer, which is usually thin, there is brilliant sunshine and the temperature is several degrees higher than at the surface. This condition is known as anticyclonic gloom. It does not become established in summer because solar heating disperses the cloud layer and destroys the temperature inversion by the heating of the surface.

Diurnal effects are the main features of the radiational weather of anticyclones. There is a large diurnal range of temperature. Land and sea breezes are prominent on the coasts since the differential heating of land and sea is marked and there is no vigorous wind circulation to mask them.

The Wedge of High Pressure.—The wedge of high pressure is a wedge-shaped extension of an anticyclone and is thus the antithesis of the trough. It usually lies between two depressions as in the

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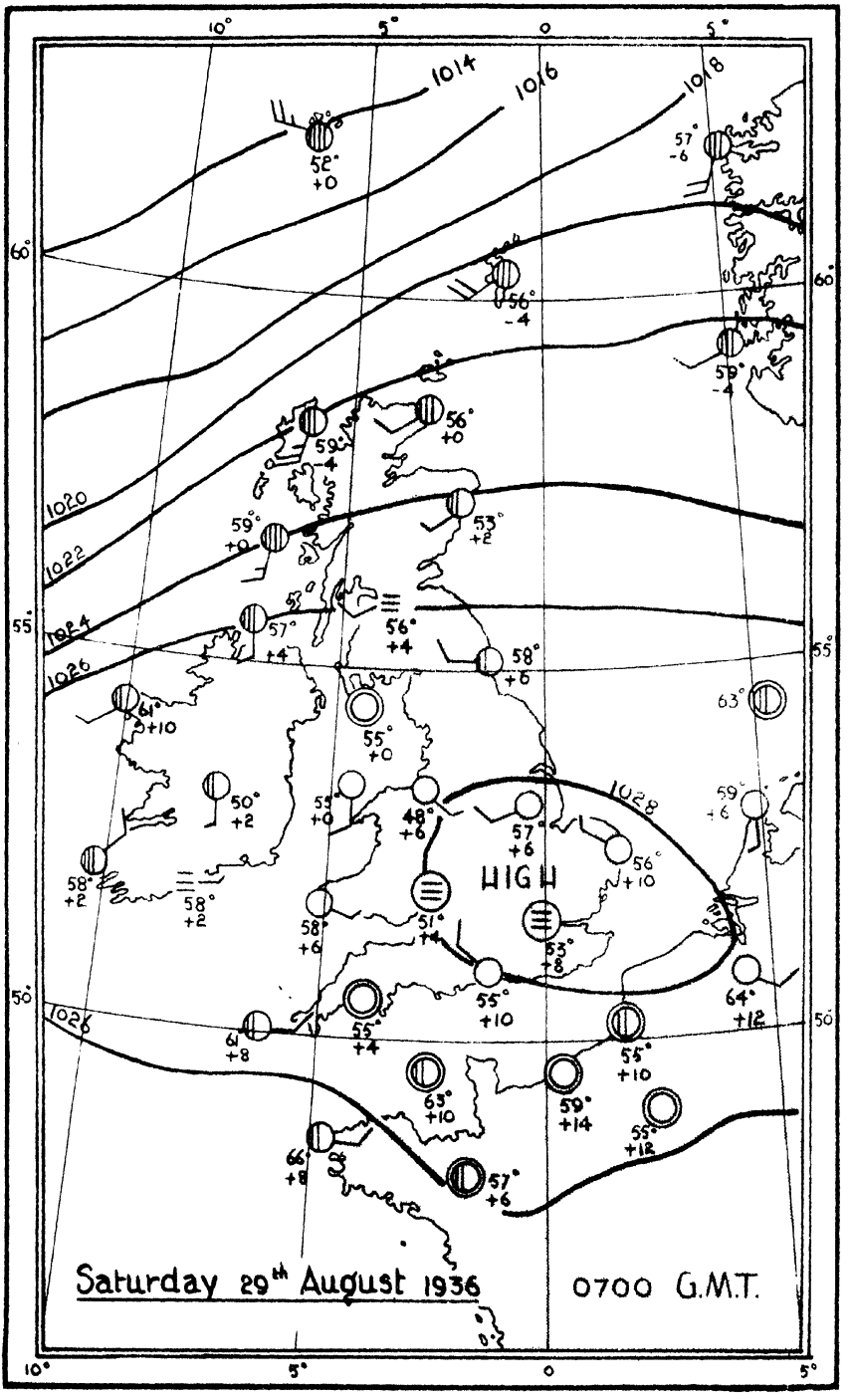

FIG. 121 ANTICYCLONE

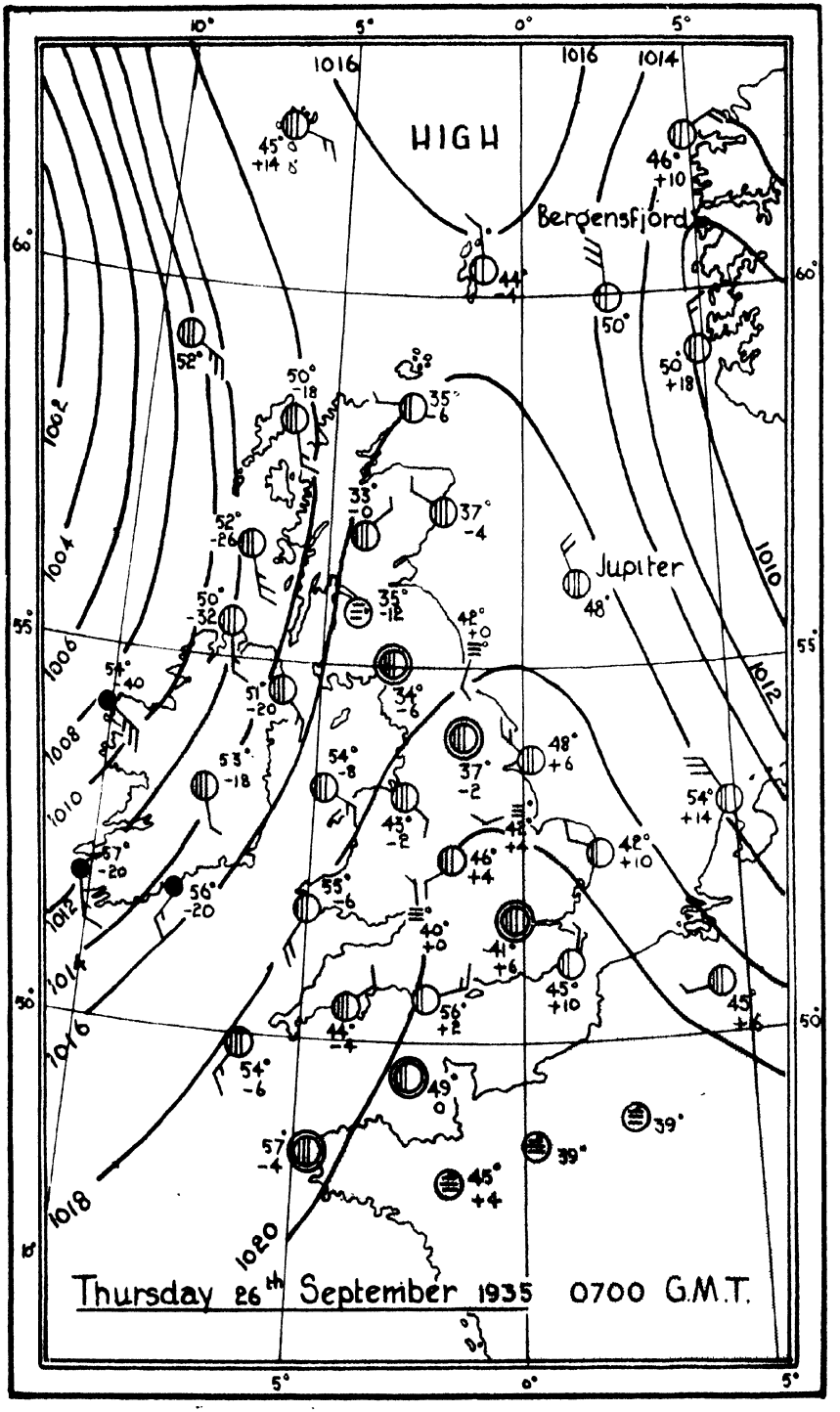

FIG 122 WEDGE OF HIGH PRESSURE

example of Fig. 122, (*see* footnote page 339). There the wedge extends across England and Scotland from an anticyclone over France and lies between depressions centred over the Baltic and the north-east Atlantic. In the northern hemisphere the point of the wedge is usually directed towards the north and in the southern hemisphere towards the south. The movement of the wedge is determined by that of the two depressions between which it lies and is therefore usually from west to east. In the northern hemisphere the winds are north-westerly on the eastern side, south-westerly on the western side and light variable in the centre. The weather of the front and central zones is often of the most brilliant type but deteriorates rather rapidly in the rear zone owing to the following depression. It is more consistently fine in a wedge than in an anticyclone but since the wedge moves between two depressions the fine weather is short-lived. Visibility within the wedge is usually good, but if the wedge is sufficiently broad the central area of light winds and clear skies may be associated with fog and in winter with frost. Fog is shown at a number of British and French stations in the wedge of Fig. 122. The change of wind at the axis of the wedge is always gradual, contrasting with the abrupt change which sometimes occurs at a trough of low pressure.

Other Pressure Types.

The Col.—A col is the distribution of pressure between two anticyclones and two depressions arranged so that the two highs (or lows) are diametrically opposite to one another. In Fig. 123, (*see* footnote page 339), a col over Ireland, Wales and western England separates anticyclones over Scandinavia and off south-west Ireland and depressions over France and to the north-west of Ireland. The positions of the anticyclones and depressions may be orientated in any way so long as they retain the same relative configuration with respect to one another. The col is a region of indefinite pressure gradients and light winds. It is also the meeting place of winds from sources which are widely different. In the example chosen, air drawn from the continent circulating round the Scandinavian high is brought into contact with Atlantic air circulating round the low off north-west Ireland. Such currents of air often differ widely in respect of their temperature and humidity and when they meet condensation by mixing occurs giving rise to cloud and local fog and rain. The latter is occurring at the widely separated stations, Valentia, Renfrew and Brest in the example chosen. In summer the stagnating air is subject to prolonged surface heating over land as a result of which thunderstorms are frequent. Later in the day for which Fig. 123 was chosen, thunderstorms, severe in some cases, developed over southern England. The col forms an easy path for depressions between the two anticyclones and it is therefore not a lasting feature of a series of synoptic charts.

If the col is attenuated along its high-pressure axis it is called a ridge or belt of high pressure. The weather associated with such a system is similar to that of an anticyclone. Similarly, if the col is attenuated in the direction of its low-pressure axis it forms a belt of low pressure, in which the weather is usually of the unsettled cyclonic type.

Straight Isobars.—Occasionally the isobars run in more or less parallel straight lines across a large area. The configuration is then known as straight isobars. They are usually the outlying portion of a large and distant depression or anticyclone, being straight owing to their large radii of curvature. The associated weather is changeable, small disturbances being liable to form as small bulges on the isobars, which may be almost imperceptible on the synoptic chart. These move along the isobars in the direction of the general wind circulation and bring with them the cloud and rain associated with secondaries. In the example shown, Fig. 124, (*see* footnote page 339), the rain at Brest, Cranwell (Lincs.), and in central Scotland appears to be associated with small travelling secondaries, which impression is confirmed by the backing of the wind from the general direction at these and nearby stations. The weather with straight isobars depends very much upon the direction in which the isobars run. Thus over the British Isles straight isobars from north to south with low pressure to the east bring cold weather with frequent showers of snow or sleet in winter and spring. Another common type is that with isobars running from south-west to north-east and low pressure in the neighbourhood of Iceland, associated with mild weather and alternating fair and rainy periods.

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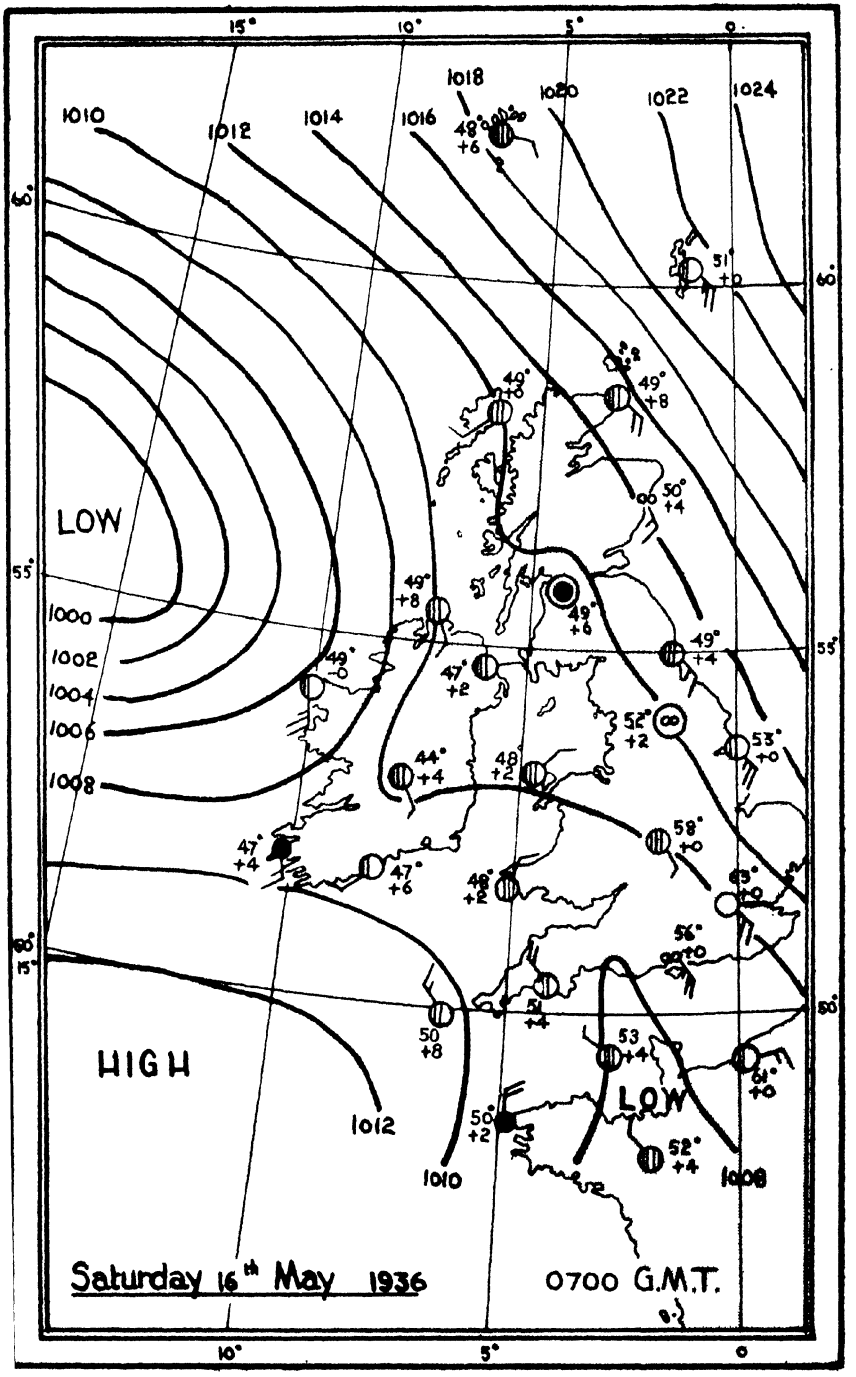

FIG. 123 COL.

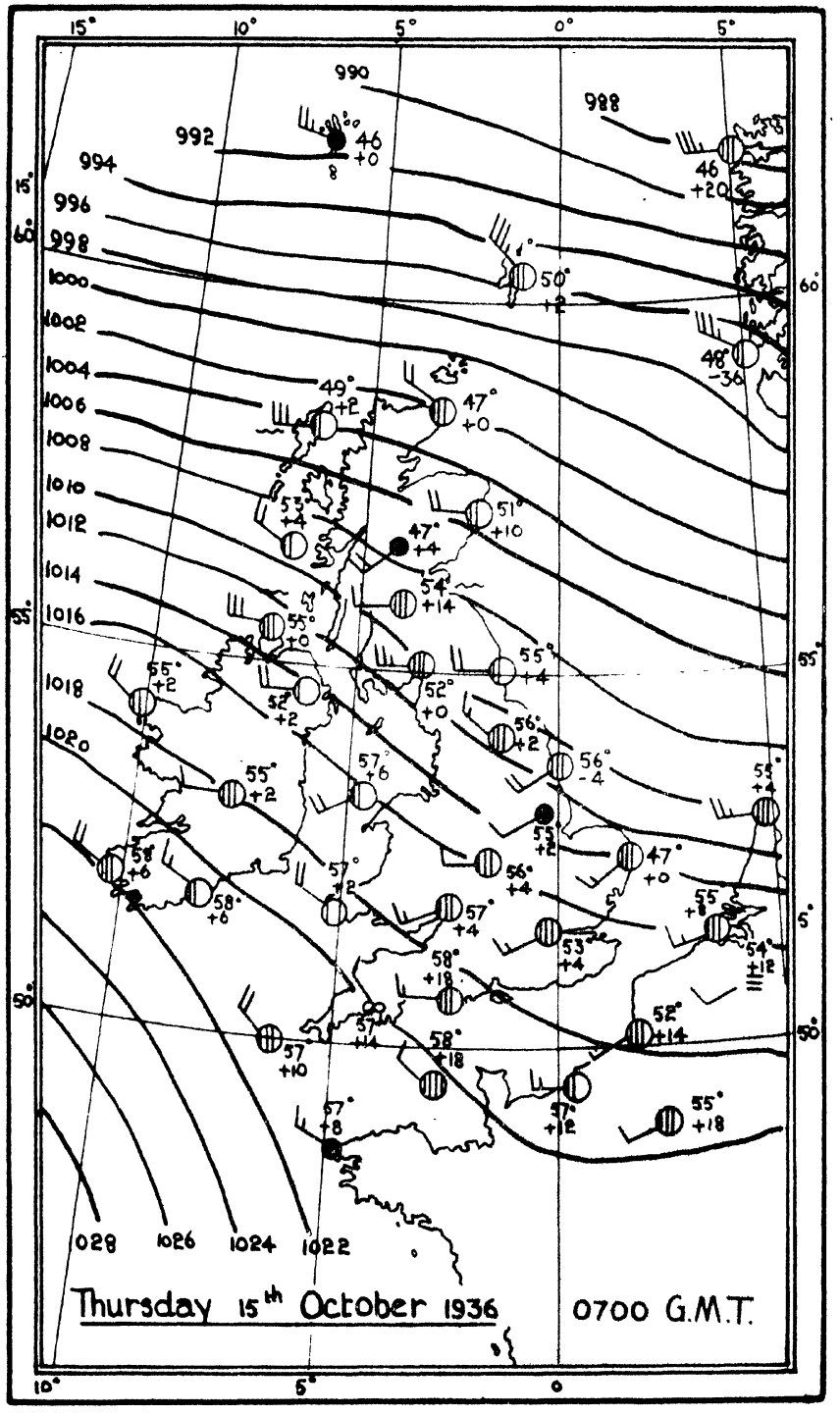

FIG. 124 STRAIGHT ISOBARS.

CHAPTER XIX.

Tropical Storms

General Characteristics.

Tropical storms are confined to very definite regions of the earth and occur for the most part in the tropical portions of the western sides of the great oceans. The term cyclone is applied to all tropical storms (the extra-tropical cyclone is generally known as a depression), but they are designated by various names according to the locality in which they are experienced. The following are the regions in which tropical cyclones occur and the names by which they are known :— (*See* Fig. 125.)

(1) North Atlantic Ocean. The West Indies, the Gulf of Mexico, the Caribbean Sea and the Florida coast—**hurricanes.**

(2) North Indian Ocean. The Arabian Sea—**cyclones.**

(3) North Indian Ocean. The Bay of Bengal—**cyclones.**

(4) North Pacific Ocean. The China Sea, the Philippine Archipelago and Japan—**typhoons** (in the Philippines they are called **baguios**).

(5) North Pacific Ocean. Off the coast of Central America and southern California—**hurricanes** or **cordonazos.**

(6) South Indian Ocean. To the east of Madagascar, near the islands of Mauritius and Réunion ; also between the north-west coast of Australia and the Cocos or Keeling group of islands—**cyclones** (on the north-west coast of Australia they are called **willy-willies**).

(7) South Pacific Ocean. From the Queensland coast of Australia to the Paumotu Islands—**hurricanes.**

It will be seen from the above that the only ocean bordering on the equator in which tropical storms never occur is the South Atlantic.

These storms are of a seasonal nature and have their greatest frequency towards the end of the hot season in each hemisphere. The following table shows the number of occurrences of tropical cyclones reported over various periods for different parts of the world.

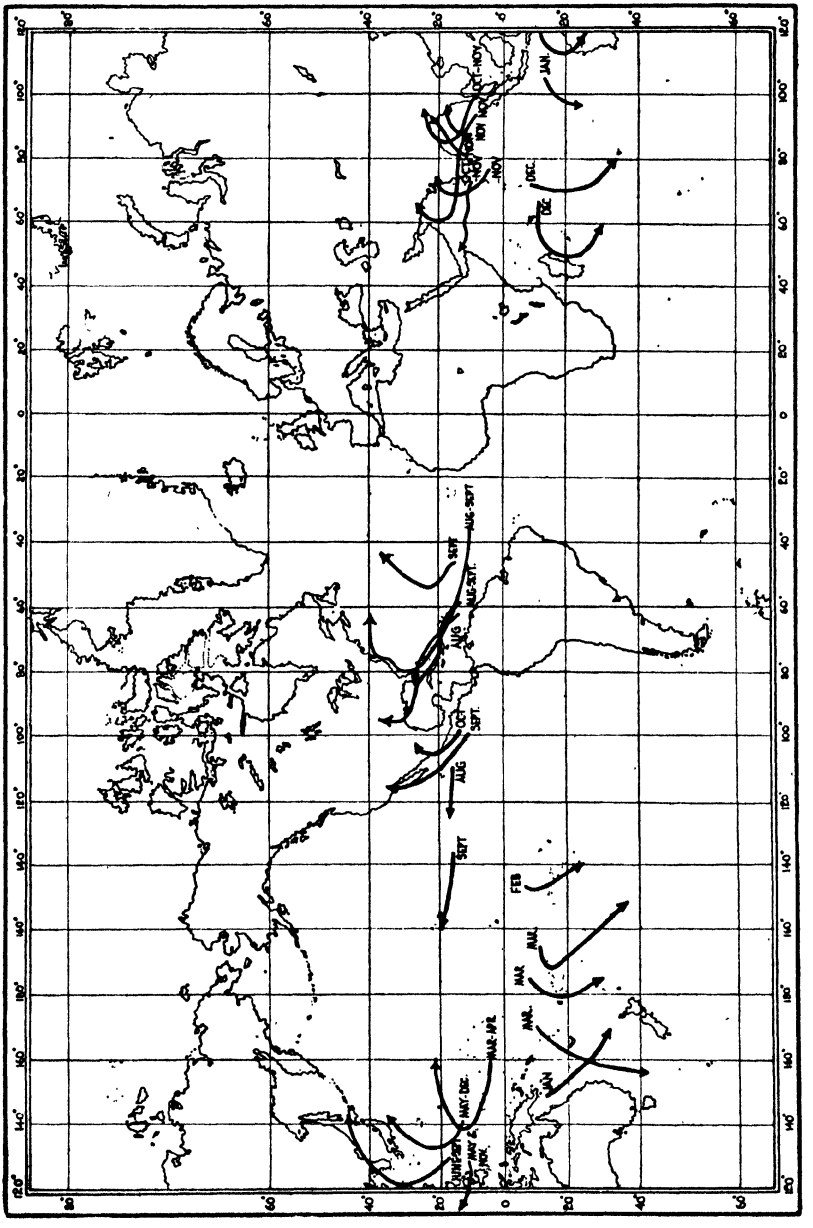

FIG. 125.—CHART OF THE WORLD, SHOWING LOCALITIES AND TRACKS OF TROPICAL STORMS.

Region and Period.	January	February	March	April	May	June	July	August	September	October	November	December	Total
West Indies, 1887–1923	—	—	—	—	—	11	13	35	61	49	8	2	179
Arabian Sea, 1842–1925	2	—	—	4	10	12	3	—	1	12	13	4	61
Bay of Bengal, 1891–1923	—	—	—	4	8	2	3	1	5	9	10	4	46
China Sea and western N. Pacific, 1893–1918 ..	31	17	18	14	33	34	90	92	110	95	52	34	620
Eastern N. Pacific, 1910–1928	—	—	—	—	1	11	14	15	34	17	2	1	95
Eastern and western South Indian Ocean, 1886–1917	42	54	39	18	6	—	—	—	—	2	8	25	194
South Pacific, 1789–1924	66	48	62	17	3	2	1	1	3	4	7	33	247

There are certain terms used to describe the various parts of a tropical cyclone, based on its forward movement and on the characteristics which are peculiar to these parts (Fig. 126). They are :—

Vortex or Eye of the Storm.—The central calm area within the ring of hurricane winds, where the barometer is lowest.

Centre.—That part of any cyclonic wind system where the barometer is lowest.

Angle of Indraft.—The angle which the direction of the wind makes with an isobar.

Track.—The track along which the centre has travelled.

Vertex.—The most westerly point reached by the centre when recurvature takes place. Also called the "cod" of the track.

Right Semi-circle.—That half of the storm which lies to the right when looking along the track in the direction of movement of the storm.

Left Semi-circle.—That half of the storm which lies to the left when looking along the track in the direction of movement of the storm.

Trough.—The line of lowest barometric pressure athwart the track, extending right across the storm-field.

Storm-field.—The region covered for the time being by the winds forming the storm system.

Dangerous Semi-circle.—The semi-circle which lies on the side of the track towards the usual direction of recurvature, *i.e.*, the right semi-circle in the northern hemisphere and the left semi-circle in the southern hemisphere. The dangerous semi-circle is so called because a ship caught in it may be blown towards the path over which the vortex will pass, or the storm may recurve and the vortex pass over her.

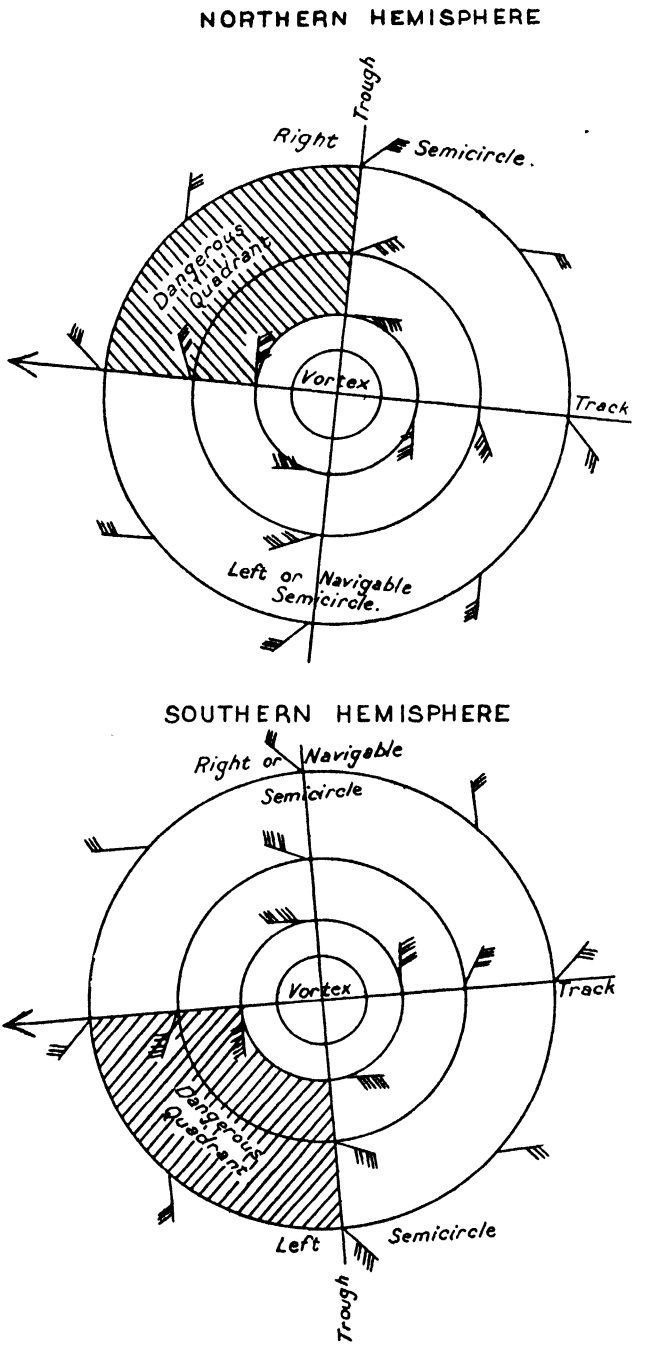

FIG. 126.—DIAGRAMMATIC REPRESENTATION OF TROPICAL STORMS OF THE NORTHERN AND SOUTHERN HEMISPHERES.

Dangerous quadrant.—The advance quadrant of the dangerous semi-circle.

Navigable Semi-circle.—That semi-circle which lies on the side of the path away from the usual direction of recurvature, *i.e.*, the left semi-circle in the northern hemisphere and the right semi-circle in the southern hemisphere.

The winds associated with tropical cyclones are amongst the most violent on the earth, the rainfall is torrential, tides are abnormal and the combined effect works extraordinary havoc both on land and sea. The storms are essentially maritime and seldom penetrate far inland, but the islands in the seas frequented by them, and sometimes the continental coasts, have to bear all their fury and they often suffer severely.

Comparison between Tropical Cyclones and Extra-Tropical Depressions.

The tropical storm and the extra-tropical depression are in essentials alike, for they are both areas of low pressure. They have the same wind circulation, counter-clockwise round the centre of low pressure in the northern hemisphere, clockwise in the southern hemisphere ; but while winds of hurricane force are the rule rather than the exception in the case of tropical storms, winds of this strength are rare in their extra-tropical prototypes. This difference in the strength of the wind is largely due to the difference in the pressure gradient. For example, the pressure at the centre of a tropical cyclone is frequently about 960 mb., and at its outer edge about 1,020 mb., while the diameter of the storm area, although it may reach 800 miles, is generally much less. In a depression, similar differences of pressure occur between the centre and outer edge, but the diameter of the depression may be 1,000, or even 2,000 miles. In tropical storms of exceptional severity the barometric gradient may be as much as 0·5 mb. per mile. Although barometric pressures of less than 900 mb. have been recorded, the barometric minimum is not necessarily a good criterion of storm intensity, assuming intensity to mean the strength of the wind. Indeed, many of the most destructive storms have had barometric minima of over 980 mb. and wind speeds of over 130 knots have been experienced in such cases. In a typhoon which struck the island of Guam in the Pacific Ocean, in March 1923, the wind speed reached 134 knots with a barometric reading of 987 mb. While this reading does not indicate the minimum pressure at the centre of the typhoon, since the centre did not pass over the station mentioned, it does show that it is not necessary for the bottom to fall out of the barometer before winds of hurricane force may be expected.

There are several other important aspects in which the tropical cyclone differs from the extra-tropical depression. These are :—

(i) The isobars of the tropical cyclone are, in general, more symmetrical and more nearly circular than those of the depression.

(ii) The temperature distribution around the centre of the cyclone is more uniform than it is around the depression.

(iii) In tropical cyclones rains are torrential and more or less equally distributed around the centre ; this is not the case in a depression.

(iv) Tropical cyclones are most frequent during the late summer of the hemisphere in which they occur, while extra-tropical depressions have their greatest frequency in the winter.

(v) Tropical cyclones frequently move to higher latitudes, where they often acquire the characteristics of extra-tropical depressions ; the latter, on the other hand, never occur in tropical regions, nor do they acquire the characteristics of tropical storms.

Origin.

There are two main theories to account for the origin and persistence of tropical storms—the convectional, and the mechanical. The convectional theory presupposes the fulfilment of the following conditions :—

(i) The air must be hot and of high humidity.

(ii) The earth's rotation must be effective. On the equator itself there is no force tending to convert an inward flow of air into a rotational movement, and, unless there is a rotational circulation, an area of low pressure cannot persist.

(iii) The lapse-rate must be such that saturated air which commences to rise as a result of convection at the surface will not cease to rise until it has attained a great height (something approaching 8 miles).

(iv) A mechanism must exist which is capable of carrying away the vast amount of air which is forced up to the top of the vortex.

(v) The general air current in which the vortex is formed must be homogeneous, within certain limits, throughout the troposphere, otherwise shearing of the column will take place.

When these conditions are present, a column of revolving air will be set up which will eventually become extremely stable, and which, having reached this stage, will be protected from outside

influences which might destroy it, by the ring of violent winds surrounding its axis.

The mechanical hypothesis is that tropical storms result from a disturbance of the normal balance between two opposing currents of air, on or near the boundary separating them. A reduction in air pressure takes place and denser air flows inwards towards the region of lower pressure. Except on the equator, the deflective effect of the earth's rotation will be to set up a cyclonic circulation, and the energy released as a result of the condensation of water vapour will stimulate the incipient cyclonic formation.

Whatever the causes may be, and it is probable that both theories play their part, the conditions enumerated above are only fulfilled over the tropical parts of the ocean where the trade winds are dying out and merging into the equatorial calms of the doldrums. It should be noted in support of the mechanical theory that the only ocean in which tropical cyclones never occur is the South Atlantic, and in this ocean the SE. trade wind extends right up to the equator in all seasons, and blows with remarkable persistency. In this ocean, therefore, there are no surface converging currents near the equator at any time of the year, such as exist on either side of the doldrums in the other oceans. The large seasonal differences in the frequency of tropical cyclones also suggest that convergent winds play an important part in the formation of these storms, since the seasonal range of both air and sea temperature is very small in the tropics, and hence convection should be scarcely less active in the winter than in the summer. Possibly, however, upper-air conditions form the controlling influence over both seasonal and annual tropical cyclone frequency.

It is significant that the western sides of the great oceans are most frequented by these storms, although there are exceptions in the case of the eastern North Pacific, the Bay of Bengal, and the eastern South Indian Ocean. A possible explanation of this lies in the fact that the flow of air to the equatorial side of the oceanic high-pressure areas takes place on the eastern side of these areas, and thus the air over the eastern portions of these oceans has but recently arrived from higher latitudes and is relatively cold. Before it can attain the requisite heat and water-vapour content to start a tropical cyclone it must travel westward over a hot sea surface for a considerable distance. Furthermore, tropical cyclones move from east to west during the early part of their lives, and a large number are not detected until they approach the coast bordering the western limit of the ocean. In the case of cyclones that occur in the Bay of Bengal and Arabian Sea, it is interesting to note that they have their greatest frequency during the transitional periods at the changing of the monsoons, while the origin of the storms that occur off the west coast of central America and southern California is probably associated with that of the hurricanes of the Atlantic.

The oceanic regions within which tropical storms form, are all studded with many small islands and it is possible that the trigger action which may be necessary to start them is provided by the thermal inequalities over adjacent land and sea masses, such as exist in the vicinity of these islands. It is again significant that in the South Atlantic Ocean there are no such chains of islands, and, as we have seen, no tropical cyclones. Once the rotational movement in such a storm has been well established, the storm itself is carried along by the general air current within which it happens to be at any particular stage of its existence.

In general, the season of maximum tropical cyclone frequency corresponds with the period during which the doldrums are at their maximum distance from the equator, in both hemispheres. Seldom do they originate in latitudes lower than 5° and they have their greatest frequency of development between the parallels of 10° and 20°, the zone of origin progressing polewards in conformity with the swing of the doldrums during the summer. As the season of greatest frequency advances, an annual longitudinal movement of the zone of origin also takes place, eastwards, in all regions, which is followed by a westerly retreat after the month of maximum incidence is past. But there are many exceptions to these precepts which do but indicate the general tendency.

Size and Movement.

Tropical storms vary greatly in size. In the vicinity of their place of origin they may be no more than 50 miles in diameter, while on becoming fully developed and passing into higher latitudes they spread out and their storm-field may have a diameter up to 800 miles. The average in such cases is about 500 miles. The clouded area associated with these storms, however, is usually more extensive than that of the surface storm-field itself. Tropical cyclones of small diameter are generally more severe than those covering a large area, but there are exceptions to this rule, for storms of large area can be very violent, and, conversely, small storms do not always have winds of hurricane force. The shape of the storm is generally circular, but the isobars are frequently crowded together on the poleward side in the direction of the anticyclone which the storm as a whole is skirting. As a result one would expect to encounter the strongest winds in that portion of the storm.

The usual track of the tropical cyclone resembles a parabola, *see* Figs. 127 and 128. In the northern hemisphere these storms almost invariably move westwards from their place of origin at a speed of about 10 knots and seldom more than 12 knots, gradually turning to the north-west as they proceed. On advancing into higher latitudes they tend to incline more to the northward and on reaching the vicinity of latitudes 20° or 25° begin to swing round through north, finally setting off in a north-easterly direction. This

is what is known as " recurving." During the recurvature process, the progressive motion is somewhat arrested, but having once established its course to the north-east, the storm increases its speed rapidly until it is travelling at 20 knots or more. The mean point of recurvature moves progressively north and east as the season of maximum storm incidence advances, corresponding to the direc-

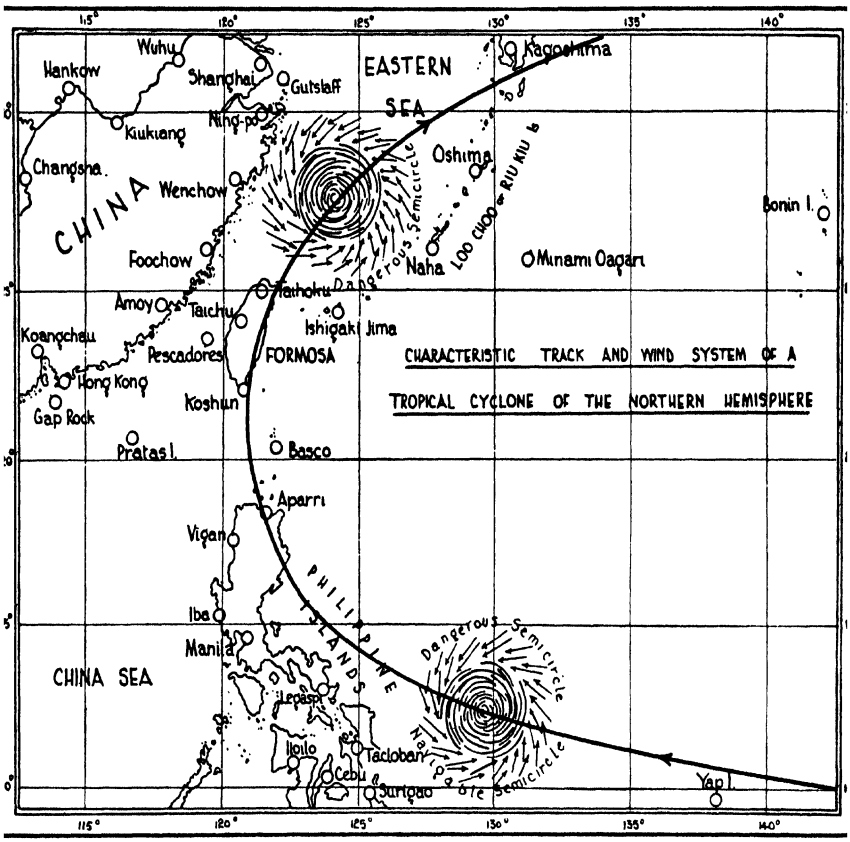

FIG. 127.—TRACK OF TROPICAL CYCLONE (NORTHERN HEMISPHERE).

tional swing in the place of origin. It would appear then that the track of tropical cyclones in the northern hemisphere is such as to keep an area of high pressure on the right-hand side of the storm as it advances. This statement is intended to apply only to the general movement of these storms in a clockwise direction around the permanent oceanic " highs," but it is by no means an infallible rule, as tropical storms often recurve on the south-east side of a continental anticyclone, thereby placing such a high-pressure area on their left-hand side. Moreover, individual storms occasionally describe a complete circle, thus passing over the same place twice. Frequently, too, they turn suddenly and move south-west even

in a comparatively high latitude, and near the point where one would expect them to recurve on to a north-easterly course. Others again move north-east in a region in which these storms normally travel towards the west. Sometimes a storm may remain practically stationary for a considerable time or conversely it may advance at a speed greatly in excess of the normal.

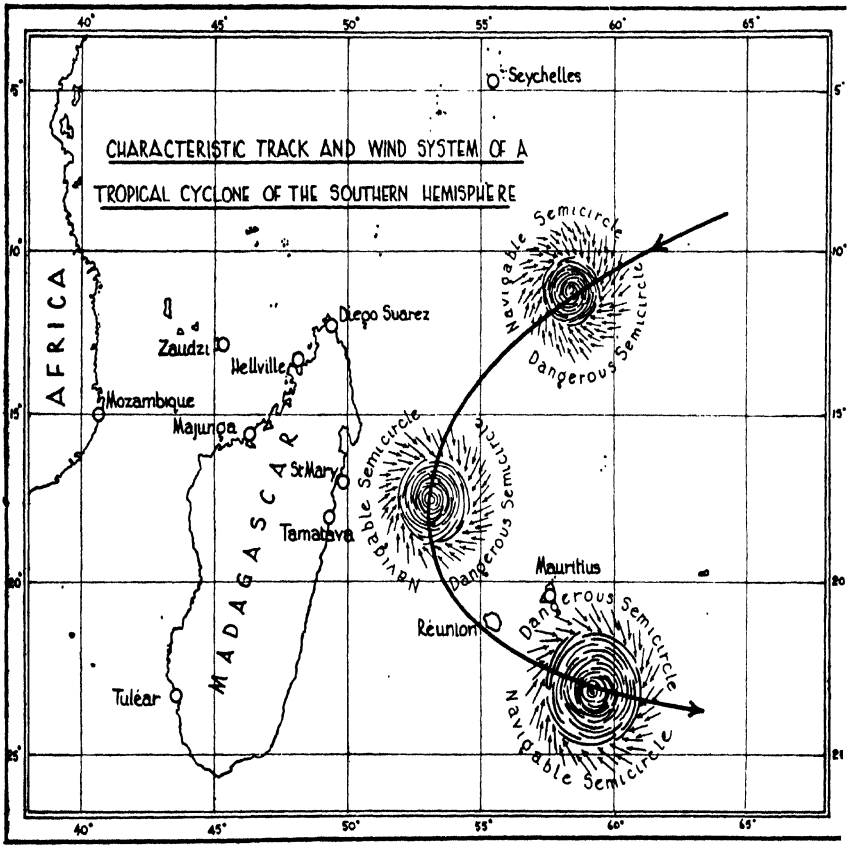

FIG. 128.—TRACK OF TROPICAL CYCLONE (SOUTHERN HEMISPHERE).

In the southern hemisphere the general movement of tropical cyclones is of a similar nature, the corresponding directions being west, south-west, south and finally south-east, so that in general they travel counter-clockwise around the permanent oceanic "highs" of the southern hemisphere. The speed of advance in the different stages is similar to that of cyclones of the northern hemisphere, while the mean point of recurvature progresses south and east as the season advances, but is nearer the equator than that of storms of the northern hemisphere. These rules, however, only apply to the general movement. They are by no means infallible and

should be used guardedly, for these storms are subject to the same vagaries of movement as those of the northern hemisphere.

It has been found possible, however, to formulate certain more definite rules with regard to the movement of tropical cyclones of both hemispheres, but again they are not infallible. These rules are :—

(i) Tropical cyclones tend to move parallel to the isobars which lie outside the storm-field.
(ii) They will avoid a stationary anticyclone by passing round it one way or the other.
(iii) If an anticyclone advances towards a tropical cyclone the latter will generally either turn away from the advancing high-pressure area or will cease to move.
(iv) Tropical cyclones tend to follow in the wake of a retreating anticyclone, and under these circumstances they may move across the isobars.
(v) A tropical cyclone tends to recurve into a trough of relatively low pressure that may exist when the storm arrives in the region, irrespective of the longitude or the time of year.

It would be useless to pretend that these rules give results which are completely satisfactory; the best that can be said for them is that they may be of some assistance to the forecaster, especially when used in conjunction with certain precursory signs. (*See* page 377.) They have been known to fail on many occasions but have given conspicuously successful results on others.

In general, tropical cyclones may be expected to decrease gradually in intensity as they advance into higher latitudes and more especially after recurving. They fill up appreciably when they are over the land for any length of time, particularly if it is high land, for the sea is the source of the warm moist air necessary to maintain the energy of the whirl; but if they once more reach the sea they often deepen again. More frequently, however, they develop the characteristics of extra-tropical depressions, persisting for a long time and travelling enormous distances. For instance, West Indian hurricanes and typhoons of the China Sea which have become depressions, have been traced as far as Europe and the coast of British Columbia respectively.

Wind and Weather.

The surface winds in a tropical cyclone blow round the low-pressure area, counter-clockwise in the northern hemisphere, clockwise in the southern hemisphere, with a convergence or angle of indraft towards the centre of about 30°. In other words, if an observer faces the wind the centre of a cyclone will bear approximately 120°, *i.e.*, 10 or 11 points, on his right-hand side in the northern hemisphere, and on his left-hand side in the southern

hemisphere. The angle of indraft, however, appears to vary in different quadrants of the storm-field, being greatest in the right rear quadrant and least in the left front quadrant, in the case of West Indian hurricanes that have been studied. It also increases with the distance from the centre, and it is less over the sea than over the land owing to the greater friction of a land surface. From the rough bearing of the centre an observer should be able to judge in which quadrant he is situated. The angle of indraft diminishes with height until at an elevation of about 3,000 feet the winds are parallel to the isobars; above this level it is believed that the circulation is outwards. There is no evidence to show to what height tropical cyclones extend but it is known that they reach 15,000 feet and it is probable that during a large proportion of their existence they extend to the tropopause.

The speed of the wind varies greatly for different storms, and also for different parts of the same storm. Near the centre or eye of the storm, an area of fine weather in which the cloud is often broken, the winds are light and variable. The diameter of this area varies from about 4 miles in low latitudes to about 60 miles in higher latitudes. In this central zone, mountainous confused seas and swells run which are a greater danger to shipping than the hurricane winds blowing around the area. Away from the centre, winds of great violence are experienced, accompanied by even more violent squalls with sudden shifts of wind direction. In the case of a typhoon, the average wind experienced is force 12 at 35 miles from the centre, force 11 at 50 miles, and it gradually decreases outwards, force 6 being met with on the average at 150 to 200 miles; these figures may be assumed to hold good, on the average, in the case of tropical storms in other regions. It would appear that in the case of storms of the northern hemisphere the strongest winds are encountered in the right semi-circle, while in the southern hemisphere they are strongest in the left semi-circle. This is due to the fact that in these semi-circles the cyclonic wind circulation is reinforced by the general air current in which the storm is moving. The maximum wind force occurs immediately before the central calm is reached, and it comes as a terrific squall, the peak value of which is generally greatly in excess of any previously recorded. As the central zone passes, another terrific squall, from a direction diametrically opposed to the previous one, blows with hurricane force, its arrival heralded by an ominous roar which bursts upon the comparative stillness of the calm centre. If the centre of the storm does not pass over the observer, no calm period will be experienced. The wind direction will, of course, shift as the bearing of the centre of the storm changes, but such shifts will not be as sudden as that which takes place when the central zone passes over the observer.

Heavy rainfall, sometimes accompanied by thunder and lightning is associated with all tropical cyclones. On the average, the

rain sets in at about 100 or 150 miles from the centre, but this distance varies with different storms. The extreme record of rainfall in a typhoon is 46 inches in 24 hours at Baguio, in the Philippines, but falls of as much as 20 inches in a day or two are not uncommon. During the period of rainfall the sky is covered with a dense pall of nimbostratus cloud, the visibility decreases, and the temperature of the air drops appreciably. In the centre, the cloud diminishes in amount, or even disappears completely, the temperature rises, and the rain generally ceases; but as the central zone passes, the sky once more becomes overcast, the temperature falls again, the visibility drops, and the rain comes down with renewed violence. In the rear of the storm it is not generally of such long duration, and as the storm recedes the layer of dense cloud lifts, giving way to fractocumulus and cirrus; the rain ceases, the temperature rises, the barometer rises to its normal level, and the wind falls to a moderate or gentle breeze. Except for the wreckage and the ominous heaving of the ocean, one would not know that a storm had passed.

Such then is the sequence of events in the life of a tropical storm. It cannot be over-emphasized, however, that such a description is at best a general one, as the inherent characteristics of such storms differ not only from region to region of their occurrence, but also between storms experienced in the same locality.

Warnings of approach of tropical storms are now broadcast by the meteorological services in the various countries adjacent to the regions of their occurrence. These warnings generally indicate the position of the storm at certain times, and its intensity and general movement. For details *see* A.L.W.S., Vols. I and II.

There are certain precursory meteorological signs associated with tropical cyclones which, used in conjunction with this information and having regard to the general characteristics of such storms, will enable an observer to predict with some measure of accuracy their subsequent movement, and the probability or otherwise of their passing near the ship or station concerned. Moreover, in the open sea and when no such information is available from official sources, careful observation and judicious application of these portents may reveal to the single observer the existence or development of a storm, and so enable him to take the necessary action to avoid imperilling the ship.

Precursory Signs of Tropical Storms.

Sea.—The first signs of the approach of a tropical storm are to be found in both sea and sky. In the open sea, provided there is no intervening land between the ship and the storm centre, the swell will probably give the first indication, and under these conditions it probably takes pride of place as a warning. The swell may extend to a distance of over 1,000 miles from the seat of the disturbance, and can usually be relied upon to make itself felt 400 miles

from the centre. As its movement is outwards from the centre of the storm, its direction gives a good idea of the bearing of the storm. Moreover, a change in the direction of the swell may often be used as a means of determining the storm track. The swell increases as the centre of the storm approaches, but the state of the swell for any given distance from the centre depends on the intensity of the storm, and so cannot be relied upon to give any reliable indication of distance.

Sky.—Cirriform clouds generally, but not always precede tropical storms, their appearance being a more reliable indication in the tropics than in higher latitudes. On the other hand, the appearance of such cloud does not always indicate the existence of such a storm. In many cases, when the storm is yet far distant, and before the barometer commences to fall or the least sign of bad weather is observed, a thin cirrus haze covers the sky, which may cause lurid red sunsets, and halos or rings about the sun by day and the moon by night. The wind falls light or drops completely, the air is sultry and oppressive, the visibility abnormally high, and there is an indefinable feeling of impending danger. The cirrus haze becomes true cirrus which usually stretches in bands across the sky and begins to thicken into cirrostratus or sometimes cirrocumulus. There is little known about the distribution of cirrus cloud in the different quadrants of a tropical cyclone, but in the case of West Indian hurricanes it is more frequently observed in the two front quadrants, and the left rear quadrant is particularly free. In the tropics, if cirrus cloud stretches in convergent bands, and the convergent formation remains visible in a fixed direction, the centre of the storm probably lies in the direction of the point of convergence. Such a formation may appear as much as 500 miles ahead of the storm centre. Following on the cirriform cloud comes altocumulus, after which, banks of black cloud appear low down on the horizon. As time passes, these banks of cloud coalesce into a single sheet of nimbostratus ; this is the true storm cloud, and is termed the " Bar of the Storm." At a distance the cloud bank retains its shape and position for hours, and as it forms about the storm centre, its direction will give an approximate bearing of the centre, while any change in bearing will give an indication of the direction of the storm's movement. From the main body of this cloud portions of fractonimbus or " scud " are detached from time to time and drift across the sky, their progress marked by squalls of rain and wind of increasing force. It has already been stated that the angle of indraft decreases with height and that beyond a certain level (about 3,000 feet) the wind circulation is probably outwards from the centre. The clouds at various heights, therefore, borne along as they are by the air currents at these levels, will travel in different directions in conformity with the wind currents. Thus, if the centre of the storm lies, say, due south of the observer, in the northern hemi-

sphere, the surface wind will blow more or less from east-north-east, and, on the above assumption, the lowest clouds will move from east, the altocumulus from east-south-east, the cirrostratus from south-east, the cirrocumulus from south-south-east, and the light cirrus from south. The last stage in the development of the advancing storm is reached as the layer of nimbostratus gradually envelops the sky. The wind gradually increases to gale force, the rain falls in torrents, and the sea is lashed into fury.

Barometer.—By carefully watching the barometer, comparing it with the normal pressure and using the information in conjunction with other evidence, timely warning of the existence or approach of a tropical storm may be obtained, especially in regions where the day-to-day variation is not very marked. In low latitudes the diurnal variation, (*see* page 137), shows daily maxima, at 1000 and 2200 local time, respectively, so that any departure from this regular oscillatory movement should be regarded with suspicion. In the outer regions of the tropical storm-field, the barometer is often slightly above normal, due allowance being made for the diurnal variation. Later it may be followed by a restless oscillating or pumping of the mercury caused by the disturbed conditions in the atmosphere. Under ideal conditions, when a tropical cyclone is going to pass reasonably close to the observer, there are usually three fairly definite phases in the fall of the barometer:—

(i) A slow fall during which the diurnal variation is still apparent on the barograph trace. This usually occurs at from 500—120 miles from the centre of the storm.
(ii) A distinct fall, during which the diurnal variation is almost completely masked. This usually occurs at from 120—60 miles from the centre. Throughout this phase the barometer is sometimes very unsteady.
(iii) A rapid fall. This usually occurs at from 60-10 miles from the centre.

It is not uncommon for the barometer at the centre to stand 60 or 70 mb. lower than in the region just outside the storm-field. In the rear of a tropical cyclone the barometer rises as rapidly as it fell in advance of the storm.

The three phases in the fall of pressure may be seen in Fig. 129, which shows a typical barograph trace associated with the passage of a typhoon.

To allow for the diurnal variation the following table has been prepared. In comparing successive readings of the barometer corrected in this manner any change due to a tropical revolving storm in the vicinity will be apparent. In addition, if any particular barometric reading corrected by this table is found to be 3 mb. or more below the average for that part of the ocean as found from the " Meteorological charts for the World," it is likely that such a storm is forming or has formed at no great distance.

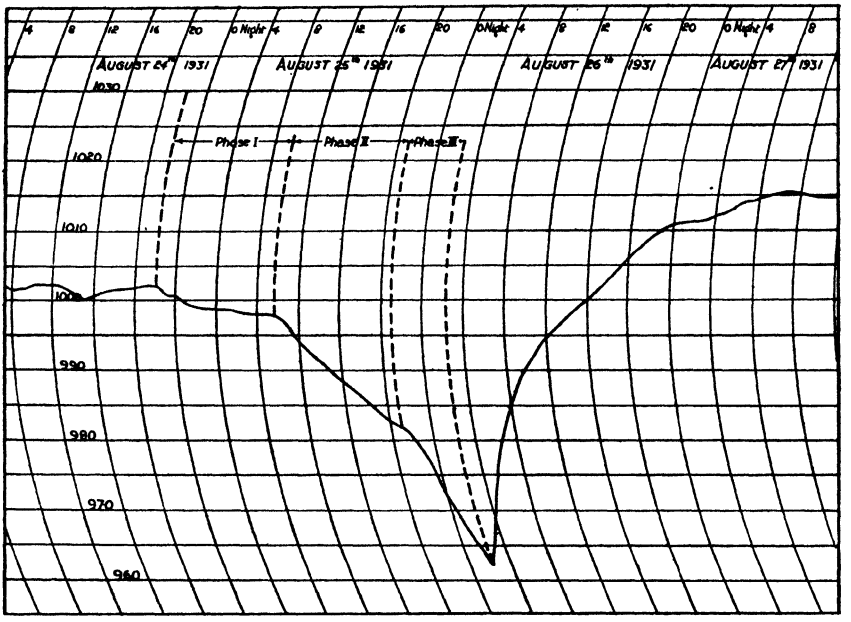

FIG. 129.—TYPICAL BAROGRAPH TRACE DURING PASSAGE OF A TYPHOON

TABLE TO CORRECT BAROMETRIC PRESSURE FOR DIURNAL RANGE AT SEA WITHIN THE TROPICS.

NORTHERN HEMISPHERE.

Zone Time.	Spring.	Summer.	Autumn.	Winter.
	mb.	mb.	mb.	mb.
0400	+0·8	+0·7	+0·8	+0·3
0800	−1·1	−0·9	−0·9	−0·9
1200	−0·9	−0·6	−0·7	−0·6
1600	+1·3	+1·2	+1·3	+1·4
2000	+0·1	+0·1	−0·1	0·0
2400	−0·4	−0·3	−0·3	−0·2

SOUTHERN HEMISPHERE.

Zone Time.	Spring.	Summer.	Autumn.	Winter.
	mb.	mb.	mb.	mb.
0400	+0·6	+0·7	+0·7	+0·5
0800	−1·0	−1·0	−0·8	−0·9
1200	−0·5	−0·4	−0·4	−0·4
1600	+1·4	+1·3	+1·1	+1·2
2000	0·0	−0·1	−0·2	−0·2
2400	−0·5	−0·4	−0·4	−0·5

Although the individual precursory signs of the approach of a cyclone have been dealt with independently, it is important to realize that *all* the factors of sea, wind, cloud, weather and barometer must be taken into consideration together. Thus, when the ominous swell is first observed, the barometer may be pursuing its normal course, while the air is sultry and oppressive ; but on the appearance of the storm cloud over the horizon the barometer will have begun to show unmistakable evidence of a fall. Again, when a departure from the normal diurnal variation is discernible on the barograph trace, the observer should watch for the appearance of the tell-tale cirrus cloud. It is also helpful to study the recorded paths of storms in different months (*see* weather handbooks for different stations, etc.), since, as we have seen, the place of origin and area traversed, vary with the seasons. By so doing, the probable course of the storm may be determined.

It is only by utilizing all the information available that the fullest measure of success in prediction can be obtained.

Handling a Ship Within the Storm-Field.

If, from the indications given above, and such others as experience has taught him, the seaman is led to believe in the approach of a tropical cyclone he should at once :—

(1) Determine the bearing of the centre.
(2) Estimate its distance.
(3) Plot its apparent path.

The following is a rough rule for determining the bearing of the centre :—

Face the wind, and the centre of the storm will be on the right-hand side in the northern hemisphere, and on the left-hand side in the southern hemisphere. The number of points to right or left may vary during the same storm, and as the wind usually shifts in squalls, its direction should be taken just after a squall. If the observer is in front of the storm, a good average allowance to make is ten points, but a larger allowance should be made in the rear. The bearing of the centre, from the direction of origin of the lower clouds, is approximately eight points. If very near the centre, the bearing of the centre from that of the surface wind direction should be reduced to eight or nine points in the front quadrants.

In the absence of synoptic information or official warnings, it is extremely difficult to estimate the distance of the storm centre. Constant observation of weather conditions and the behaviour of the barometer may give approximate results, and this information may be supplemented by assuming a rate of advance and plotting the apparent path. Two bearings of the centre with an interval between, of from two to three hours will, in general, be sufficient

to determine the course of the storm, provided that there has been a wind shift during the interval and that the ship's course and distance made good, are known. Should the wind not shift, but continue to blow steadily with increasing force, and with a falling barometer, it may be assumed that the ship is on or near the storm-path.

The relation between the position of the ship and the position and anticipated path of the storm will now indicate the proper course to pursue :—

(*a*) To enable the ship to keep out of or escape from the dangerous semi-circle and to avoid the centre of the storm.

(*b*) To enable the ship to ride out the storm if unable to escape from it.

Having plotted the apparent path of the storm it should be obvious in which semi-circle the ship is situated. Furthermore, the particular quadrant can be ascertained, for, if the barometer is falling the ship is before the trough and if rising, behind the trough (allowance having been made for the ship's course and speed). If a ship finds herself on or near the path of the storm and in advance of the centre, the best procedure, given plenty of sea room, is to run with the wind well on the starboard quarter if north of the equator, and on the port quarter if south. The ship will thus be in the navigable semi-circle and be constantly increasing her distance from the centre. The wind will draw more forward as she recedes from the centre, but the compass course first set should be adhered to until well clear. The procedure is the same, if the observations place the ship anywhere within the navigable semi-circle.

The most critical situation is that of a ship finding herself in the dangerous quadrant. In this case she should shape a course so as to bring the wind on the starboard bow, and proceed as fast as possible.

If there is any doubt whatsoever as to the position of the ship relative to the centre, it may be advisable to stop, until such time as the path of the centre has been located by observing the shifts of the wind and the behaviour of the barometer. In such cases, as indeed in all cases of the imminence of a storm, the general object of putting as much distance as possible between oneself and the storm centre should be constantly borne in mind.

The rules for handling a ship in a tropical storm so far as they may be generalized are :—

If in the path of a storm, run with the wind on the starboard quarter in the northern hemisphere, and on the port quarter in the southern hemisphere, away from and at right angles to the assumed path of the storm, until the barometer begins to rise.

If in the dangerous semi-circle (*i.e.*, the right-hand semi-circle in the northern hemisphere and left-hand in the southern hemi-

sphere), a steam vessel should steam to windward away from and at right angles to the assumed path of the storm. If in a sailing vessel, heave-to on the starboard tack in the northern hemisphere, and on the port tack in the southern hemisphere, as the wind will then always be drawing aft. Be careful to note what land is in the vicinity, as it may be possible to run into harbour, or under the lee of the land, for shelter, but in some conditions it is better to remain at sea.

If in the navigable semi-circle (*i.e.*, left-hand semi-circle in the northern hemisphere, and right-hand in the southern hemisphere) run with the wind on the starboard quarter in the northern hemisphere, and on the port quarter in the southern hemisphere, away from and at right angles to the assumed path of the storm, until the barometer begins to rise.

It is better to proceed at full speed ahead of the storm into some recognized shelter rather than be caught in narrow waters, as the visibility becomes very low due to driving rain and spray, and frequently quite large sets are experienced. When in harbour, a knowledge of the position of the storm-centre may enable the ship to be pointed in the direction of the violent squall which follows the passage of the central zone.

STORM TIDES.

Tropical storms are in all cases preceded by storm tides, the water commencing to rise on the coast in front of the cyclonic disturbance, one or two days before the storm is experienced. Over the sea, the winds in the storm-field induce waves of a size and length which bear a relation to the sustained wind speed and the length of time the waves are under the influence of the same wind direction. Waves of different sizes are developed in proportion to the winds which prevail in the different parts of the storm-field. Investigation of hurricanes of the West Indies has led to the conclusion that the largest and longest waves are created in the right rear quadrant ; these waves pass on through the storm-field and move towards the shore, where they cause a rise in the water in front of and to the right of the line over which the centre of the hurricane is advancing. This rise begins when the centre of the hurricane is 300—500 miles distant, and continues until the storm moves inland. The heights of the water reached at the shore near the storm centre, vary from 8—15 feet above the predicted tide level ; changes in the position of the maximum rise of the storm tide indicate changes in the course of the storm.

Information is lacking as to the effect on the production of storm tides by cyclonic winds in other regions, but it is to be assumed that the results are of a similar nature. In the southern hemisphere it is reasonable to suppose that the largest and longest waves are developed in the left semi-circle, probably in the rear quadrant,

and that the maximum storm tide effect will be experienced in front of and to the left of the line of advance of the storm. In general, slow-moving storms create the highest tides, so that the effect will normally be greatest in low latitudes. It must be remembered, however, that tropical storms differ greatly in intensity as they do in other respects, and the magnitude of the storm tides created by such storms must essentially be influenced by the configuration of the coast-line and the distribution of islands in adjacent waters.

CHAPTER XX.

Air Masses

In Chapter XVIII attention has been concentrated on the fronts or boundaries between extensive portions of the atmosphere differing widely from one another in their physical properties, but in themselves more or less homogeneous. Such portions of the atmosphere are termed air masses. Although the most striking weather phenomena take place in association with fronts (so that the forecaster's first aim is to locate and identify fronts on his synoptic chart), it is important also to remember that characteristic types of weather occur within air masses. A particular ship or station will be situated more often within the body of an air mass than near to a front, so that weather characteristic of an air mass is likely to be experienced more frequently than frontal weather. Moreover, it frequently happens that on a synoptic chart fronts cannot be classified directly, but only after consideration has been given to the air masses between which the fronts form the boundaries.

An air mass is an extensive portion of the atmosphere which is approximately homogeneous as regards its horizontal distribution of temperature, humidity and lapse-rate. It acquires its uniform properties by remaining stationary for a considerable period over a region of the earth's surface which is itself uniform. Such a region is known as a Source Region. The source region must be large and the physical nature and temperature of the earth's surface must be approximately the same over its whole extent. These conditions are satisfied by (for example) snow-covered land areas, tropical seas and desert areas. The polar regions, whether of land or sea, are also effective source regions, for the snow coverings of land and frozen water are identical as regards their effects on the overlying air. The conditions of middle latitudes, however, are not sufficiently uniform, owing both to the distribution of land and sea areas and the rapid variations of temperature with latitude, for them to be regarded as effective source regions.

The air must stay over the source region for a sufficiently long time for it to acquire uniform characteristics. When it finally moves away from the source region it carries with it the properties acquired at the source. These properties, however, are subject to gradual modifications as the air moves over surfaces, the conditions of which differ from those at the source. The modifications are not uniform within the moving air masses, but the differences thus introduced are usually slight compared with the differences at fronts, so that the air mass may continue to be regarded as

homogeneous. The properties of an air mass therefore depend upon :—

(*a*) The properties acquired at the source, and
(*b*) Modifications of these properties after leaving the source.

Classification of Air Masses.

The simplest classification is based upon the source. In middle latitudes, for which air-mass analysis was first carried out, two main air-mass types were originally recognized as being associated with depressions and their frontal systems. These, originating in low and high latitudes were called equatorial and polar air, respectively. It was found later, however, that it was often not possible to trace air back to the polar or equatorial regions. The warm air was often found to have come from the sub-tropics, that reaching the British Isles, for example, from the vicinity of the Azores. Similarly, the cold air was often more accurately to be described as sub-polar air originating on the North Atlantic or in the interior of Russia. It was therefore necessary to extend the source classification of air masses to four types. These are denoted by letters as follows :—

A. Arctic or true polar air, originating in the polar regions.
P. Sub-polar air, originating in high latitudes, but not the polar regions themselves. For brevity this is often called polar air, a misleading term, but confusion may be avoided by using the term arctic air exclusively for type *A*.
T. Sub-tropical air, often referred to as tropical air.
E. Equatorial air.

Important modifications of the properties of air masses arise according as to whether the air originates or performs a long journey over sea or land areas ; in the former case the air is called maritime, in the latter case, continental. These sub-types are distinguished by the prefixes *m*. and *c*., respectively, thus :—

mA. Maritime arctic air.
cA. Continental arctic air.
mP. Maritime polar air.
cP. Continental polar air.
mT. Maritime tropical air.
cT. Continental tropical air.

The General Characteristics of Air Masses.

The distinguishing properties of an air mass are its absolute and relative humidities, and its temperature at all levels. The latter is conveniently expressed in terms of the surface temperature

and lapse-rate. The properties of the air mass are for the most part acquired from and modified by the surface over which the air is situated or moves. Changes in the air mass therefore take place from below. Consequently the lowest layers are affected before the upper by modifying influences. In a given time the lowest layers are generally affected to a greater extent than the upper layers. The temperature of the air is affected by contact with the surface and by radiation to or from the surface. The lowest layers of air are heated by a warm surface and cooled by a cold one. These effects are gradually transmitted to the upper levels by turbulence, and, in the case of heating, also by convection. The humidity of the air is increased by evaporation from a warm sea-surface and to a less extent from land, cold sea, ice and snow-covered surfaces. The water vapour is transported to the upper levels by turbulence and convection.

It follows that the properties of the lowest layers of air are subject to rather rapid modification and are therefore dependent to a large extent upon the immediate past history of the air. For example, polar air, blowing over a warm sea-surface, is rapidly warmed near the surface ; tropical air moving for a long distance over snow may have its surface temperature reduced nearly to freezing point. Different air masses with similar immediate past histories may, in fact, be indistinguishable from one another if surface observations alone are considered. An example is furnished by the maritime polar air, which sometimes reaches the British Isles when a large depression is centred to the north-west of Ireland. This air, known as returning maritime polar air, originating on the western North Atlantic, often moves on a track which carries it south to latitude 45°, and it finally reaches the British Isles as a west-south-west wind. Its long passage over the sea renders it very little different as regards surface temperature and humidity from tropical air reaching the British Isles from the Azores region. In the upper layers, however, the air tends to conserve its original characteristics. Temperature differences of 20° F. are sometimes found between tropical and polar air at the same level over the British Isles. A good example is furnished by the graphs of upper air temperature (*see* page 432) for October 28th and 29th, 1936, shown in Fig. 130. Both curves are plotted from observations taken at about 0700 G.M.T. at Duxford (Cambs.). On the 28th October, Duxford was situated in a current of polar air flowing southwards round an occluded depression centred over the Skagerrak, and temperatures in the upper air were low. On the 29th October a warm front extending from the Hebrides to Land's End was moving eastwards and the cold air had been replaced by warm at all levels above 900 mb. At the 500 and 450 mb. levels, the temperature difference between the tropical and the polar air amounted to over 20° F., while smaller but considerable differences occurred down to the 850 mb. level. The air mass between the surface

and 900 mb. did not change in the interval between the two observations, as is shown by the approximate equality of temperature on the two days. The surface readings, however, show in a very striking manner the effect of the earth's surface upon the air in immediate contact with it. The early morning of 28th October was not favourable to radiation cooling in south-east England, because, although the sky was almost cloudless, the wind was moderate to fresh, so that the surface temperature was maintained above 40° F. throughout the night. During the night of 28th—29th October, however, the sky cleared and the wind fell light in the wedge of high pressure which then covered the country, and the surface temperature fell to 28° F., 15° F. lower than on the previous morning, although in the same air mass. This case illustrates how surface observations alone may be misleading for the purpose of identifying air masses. The distinguishing features must be sought for in the upper air.

On moving to temperate latitudes, tropical air is far less subject to modification after leaving the source region, than is polar air. Tropical air, moreover, tends to remain homogeneous ; polar air readily develops minor discontinuities with which are associated cloud and rain. Occasionally these discontinuities develop further into well-marked fronts on which vigorous depressions may form—entirely in polar air. The reason for this difference between polar and tropical air appears to be that the former, moving towards warmer regions, has its lower layers warmed and becomes unstable to an increasing degree ; the latter, moving towards colder regions, becomes stabilized owing to the cooling of its lower layers. The effects of changes of temperature and humidity at the surface are, therefore, in tropical air confined to the lowest layers, and the main body of the air mass conserves its original characteristics. In polar air the surface changes are transmitted rapidly upwards by convection. The effects of differing conditions experienced by different parts of the polar current, such as contact with land and sea surfaces of widely varying temperatures, penetrate the whole depth of the air mass forming active discontinuities.

Arctic Air(*A.*).—Arctic air is characterized by very low temperature at all levels, especially in the lowest layers, where contact with the cold surface causes a pronounced inversion, which may be further intensified by subsidence. If there is no subsidence the lapse-rate is nearly dry adiabatic in the upper levels. The moisture content is low on account of the low temperature, but the relative humidity is often high owing to surface cooling, and fog, consisting of ice crystals, is not uncommon. Moving away from the source region over the sea, the air mass becomes very rapidly modified (*mA*). The lowest layers acquire heat and moisture from the sea surface, increasing the lapse-rate. Cu. and Cunb. clouds are formed and showers occur, in winter taking the form of snow or soft hail. If the air leaves the source region on a land

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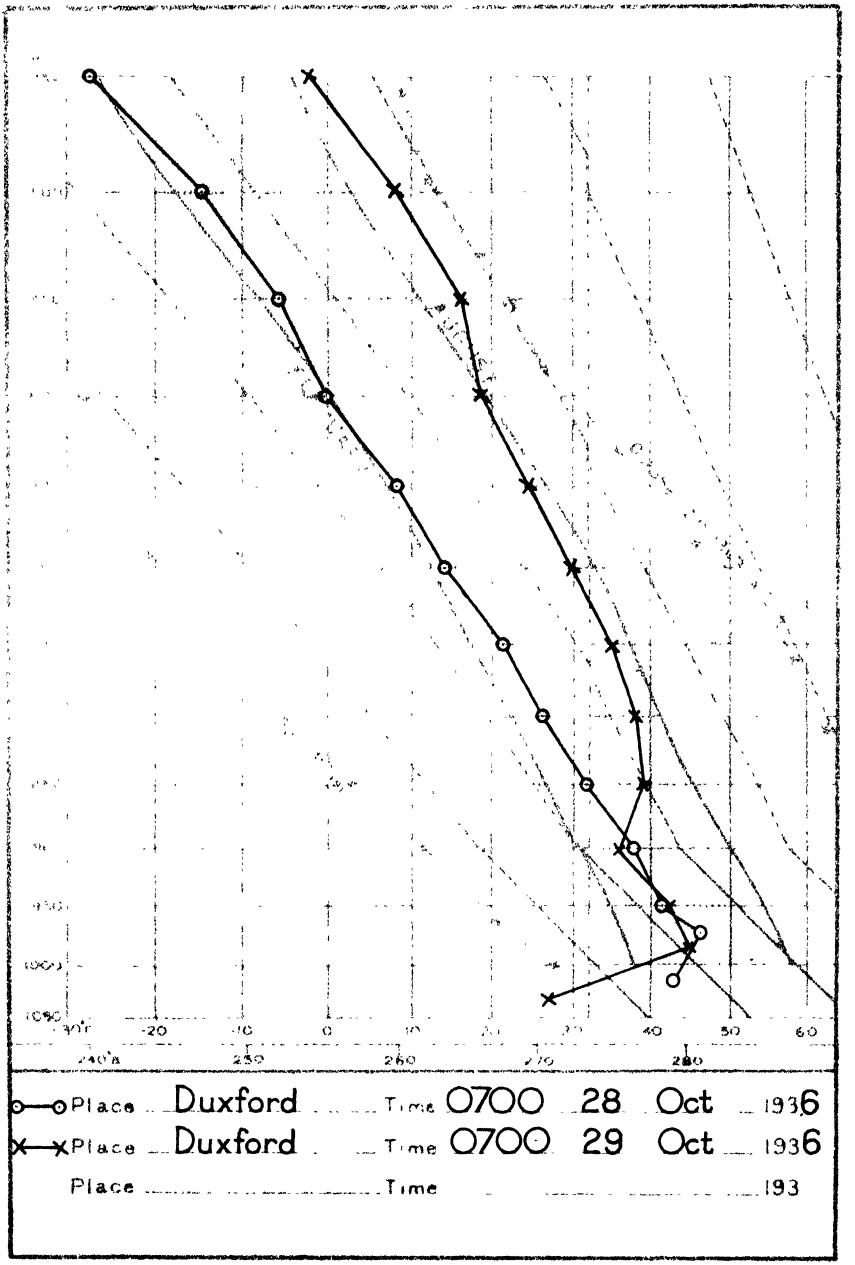

FIG. 130. GRAPHS OF UPPER-AIR TEMPERATURE TROPICAL AND POLAR AIR.

track (*cA*), it is subject to very little modification in winter, since the snow-covered continent is equally as cold as the source region. In summer the land surface becomes heated and the *cA* air is rendered unstable, but owing to the low moisture content, cloud formation and precipitation are only slight.

Continental Polar Air (*cP.*).—The source properties of continental polar air differ considerably between summer and winter, owing to the extremes of temperature to which the source region is subject. In winter, *cP* air has similar characteristics to *cA* air. In summer *cP* is warm with a large lapse-rate, but there is little cloud and precipitation owing to the low moisture content.

Maritime Polar Air (*mP.*).—The temperature of the air at the surface tends to approach that of the sea over which the air moves, so that it is relatively cool in summer and warm in winter. The lapse-rate approximates to or slightly exceeds the S.A.L.R. Subsidence is liable to occur in *mP* air behind a cold front, and the temperature at the surface may then be relatively high, even in summer. In such a case identification of a cold front by surface temperature is difficult, as there is often no change across the front. The moisture content is usually fairly high at all levels in *mP* air, and increases with the time spent by the air over the sea. It is, however, generally less than that of *mT* air, having the same surface temperature and *mP* air may often be distinguished from *mT* air by its lower dew-point. Nearly all varieties of weather may be experienced in *mP* air. It may act as cold air at a cold front in juxtaposition with *mT* or *cT* air, and as warm air at a warm front in juxtaposition with *A* or *cP* air.

Maritime Tropical Air (*mT.*).—At the source *mT* air is warm at all levels with a high moisture content. If it moves to higher latitudes the surface temperature and lapse-rate in the lowest layers are decreased by contact with the cool sea surface. If the cooling proceeds below the dew-point of the air, sea fog is formed. Orographic rain occurs at high coast-lines and in hilly districts.

Continental Tropical Air (*cT.*).—*cT* air is warm and dry, often unstable in the lowest layers. Owing to its low moisture content at the source the instability is rarely manifest as cloud or precipitation.

Equatorial Air (*E.*).—Owing to the vast quantities of water vapour carried upwards by convection in the equatorial regions, equatorial air is warm and damp at all levels. At the surface it rarely occurs outside the source regions. Instability is readily induced within it by surface heating over land.

AIR MASS TYPES ON NAVAL STATIONS.

The Home Station.

The source regions and the usual tracks followed by the air streams reaching the vicinity of the British Isles are shown in Fig. 131.

Maritime Arctic Air.—Maritime arctic air moves southwards from the Arctic between Greenland and Spitzbergen in the rear of a deep depression over Scandinavia. There is frequently an anticyclone over east Greenland which subsequently moves south-east to Europe. The cold air is usually of great depth with average lapse-rate about 4° F. per 1,000 feet. The clouds are chiefly Cu. and Cunb. with anvil Ci. Showers of snow or sleet are frequent in winter and early spring. The visibility is generally good. The wind is usually from a northerly point, often fresh and squally. Maritime arctic air frequently forms a secondary cold front, or arctic front, with the polar air in the rear of the Scandinavian depression.

Continental Arctic Air.—Continental arctic air usually moves from the region of Nova Zemlya and Spitzbergen south-eastwards and finally turns westwards round an anticyclone centred over north Russia. In winter the original low temperature of the surface layers of air is maintained by their passage over the snow-covered continent. Reaching England the arctic winds bring frost by day and night in the east and south-east. In other parts of the country the air usually arrives after a passage over the North Sea or English Channel sufficiently long to raise its temperature above the freezing-point, but conditions are still cold. There is often a pronounced inversion of temperature at about 6,000 feet, owing to subsidence. Visibility is often only moderate owing to haze. The characteristic clouds are Stcu. or fair weather Cu. and owing to the dryness of air precipitation is infrequent.

Arctic air-mass characteristics are not experienced in the British Isles in July and only rarely in the other summer months owing to the rapid modifications which the air undergoes in moving over warm sea or land.

Maritime Polar Air.—Maritime polar air originates over Canada or the western North Atlantic and may reach the British Isles by a variety of tracks the more southerly of which carry it into comparatively low latitudes. It is the most commonly occurring air mass over the British Isles. It occurs in the rear of cold or occluded fronts associated with depressions to the north and north-west and high pressure in the Azores region. The wind is mainly westerly or north-westerly, fresh and squally. The average lapse-rate is 3·5°—3·8° F. per 1,000 feet. The visibility is good. Air following the most northerly of the *mP* tracks of Fig. 131 is often called maritime polar cold (*mPC*). If it displaces tropical air at a cold front there is a decided fall of temperature at all levels. In

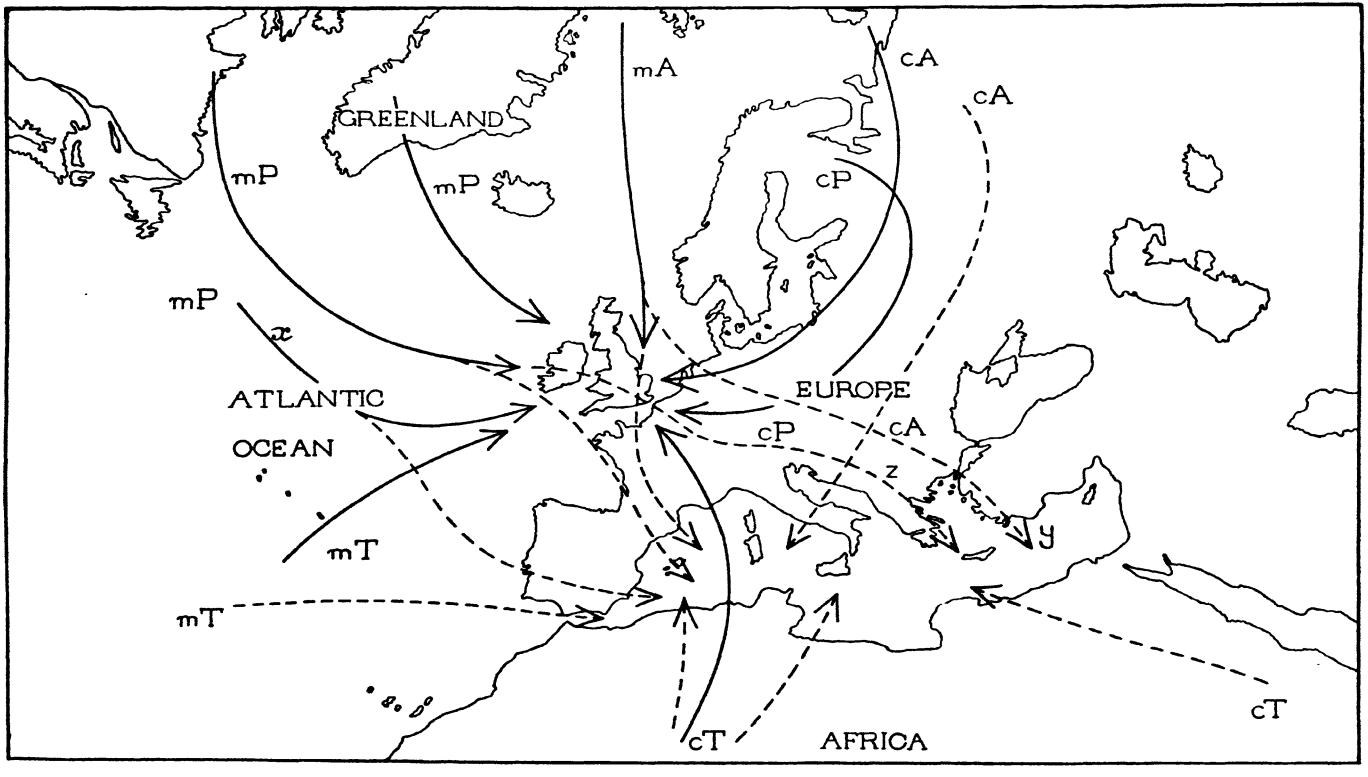

FIG. 131.—AIR MASSES OF HOME AND MEDITERRANEAN STATIONS.

winter, however, it may displace air whose lowest layers have been cooled by radiation over land ; it is then accompanied by a rise of temperature at the surface. In spring and autumn showers are frequent with chiefly Cu. and Cunb. clouds but sometimes also Cu. and Stcu. together with various forms of Acu. In summer thunderstorms or thundery showers are likely owing to surface heating over land.

Maritime polar air following the most southerly *mP* track denoted by *x* in Fig. 131 is, in its lowest layers, often indistinguishable from *mT* air. On approaching the British Isles it is moving over a sea surface which becomes steadily cooler and consequently becomes stabilized in the lowest layers. It may give orographic rain or drizzle on the coasts and hills since it becomes practically saturated, firstly on account of the water vapour acquired over the sea and secondly on account of the cooling at the surface. This air is the " returning " maritime polar air of page 387. The clouds are mostly Stcu. but large Cu. occur over heated land in summer. In winter it may act as warm air at a warm front if the air in advance of the front is either *A* or *cP*.

Continental Polar Air.—The continental polar air reaching the British Isles originates over Finland, Lapland or the interior of Russia. It occurs in the British Isles with high pressure over Scandinavia and north Russia. In winter it is generally similar in its properties to *cA* air. In summer it acquires a high temperature over the heated continent with an average lapse-rate, 3·2°—3·5° F. per 1,000 feet. Owing to its dryness there is no precipitation and little cloud formation except for fine weather Cu. in the afternoon. Conditions are often hazy owing to pollution by smoke from the industrial areas of the continent.

Maritime Tropical Air.—Maritime tropical air originates in the Azores high-pressure area. It is relatively warm and moist at all seasons and, being cooled at the surface as it flows over the ocean which becomes colder northwards, it is stable. It occurs in the British Isles either in the warm sector of a depression or with the straight isobars associated with low pressure to the north-west. The wind is usually fresh or strong from the south-west. Conditions in the British Isles are mild in autumn, winter and spring, but in summer, when the sea is relatively cold, the temperature in maritime tropical air is only moderate except over land in sunshine when conditions are unpleasantly warm and close. Over the continent in winter a surface inversion frequently forms making it difficult to trace the tropical air and the warm front from surface temperature alone. Sea fog occurs frequently in *mT* air in spring and early summer, particularly in the Channel approaches. Orographic rain and drizzle occur at all seasons. The cloud forms are mostly stratiform, Stcu., Nbst., Acu. and Ast.

Continental Tropical Air.—Continental tropical air reaches the British Isles from north Africa and (in summer) south-east Europe by way of France, with an anticyclone centred over east and south-east Europe. It is generally of low relative humidity, for even if it acquires moisture from the Mediterranean most of this is deposited on the mountains of southern Europe. In summer its temperature is high with the lapse-rate about the average, 3·5° F. per 1,000 feet. Owing to its low moisture content there is little tendency to convectional cloud formation and insolational heating over land is therefore pronounced, giving " heat-wave " conditions in the British Isles. In winter *cT* air rarely reaches this country. If it does it is so modified by its passage over the relatively cold continent that it is accompanied by rather cold dry weather with overcast skies mostly of Stcu. formed below the temperature inversion produced by the surface cooling.

The Mediterranean.

The properties of Mediterranean air masses are described briefly in this paragraph. A more detailed account will be found in the Meteorological handbook for the Mediterranean station (Weather in the Mediterranean, Vol. I).

The properties of polar air masses in the Mediterranean are, in general, similar to those in the British Isles, apart from modifications introduced by the passage of the air across the continent of Europe. Tropical air, on the other hand, reaches the Mediterranean without much modification. The principal source regions and tracks are shown in Fig. 131.

Maritime Arctic Air.—Maritime arctic air only penetrates to the Mediterranean when, in addition to the depression over Scandinavia required for its presence over the British Isles, there is another depression over north France and the Netherlands. The arctic air stream then sweeps across France, giving rise to the mistral (Chapter VIII) of the Rhone Valley and the Gulf of Lions. In the Mediterranean area, as in the British Isles, maritime arctic air is unstable and its instability is increased over the warm sea. In autumn it is accompanied by heavy showers and thunderstorms and in winter and spring by hail showers and snow on the hills of the northern Mediterranean lands, while gales are frequent. Arctic air mass characteristics are not experienced in summer.

Continental Arctic Air.—Continental arctic air generally moves from North Russia, south-westwards round an anticyclone lying from south-west to north-east across Europe.

Occasionally in the eastern Mediterranean it is an originally maritime air mass which has become dried by travelling across central Europe and the Balkan Peninsula (track *y* of Fig. 131). In winter it reaches the shores of the Mediterranean as a very cold air stream with low relative humidity, the effect of which is to

intensify its apparent coldness. Flowing over the sea it acquires moisture and soon becomes unstable through being heated at the surface, giving rise to Cu. and Cunb. cloud with showers. Sometimes it is associated with the formation of depressions over the central Mediterranean, on the north side of which heavy rain occurs. The bora of the Adriatic (Chapter VIII) usually accompanies an outflow of *cA* air in winter. In summer its properties are little different from those of *cP* air.

Continental Polar Air.—In the winter months the *cP* air reaching the Mediterranean is similar in its properties to *cA* air but since its instability is less pronounced the ensuing weather phenomena are not so intense. Like *mA* air, *mP* air may, by following a track over central Europe and the Balkans, become indistinguishable from *cP* air on arrival in the Mediterreanan (track *z* of Fig. 131). In summer *cP* air arrives from the heated continent. Passing over the cooler sea it becomes stabilized and being relatively dry it is not likely to give rise to sea fog. It gives the fine weather characteristic of the eastern Mediterranean in summer.

Maritime Polar Air.—In its general characteristics *mP* is similar to *mA* air and in winter brings cool and showery weather to the Mediterranean. The phenomena are, however, less intense than in *mA* air. In summer *mP* air becomes warm and relatively dry during its passage over the heated land to the north and gives fine settled weather especially in the central and eastern Mediterranean.

Maritime Tropical Air.—The source and source properties of *mT* air reaching the Mediterranean are the same as for the British Isles. From the vicinity of the Azores and Canary Islands the *mT* air travels by way of Spain and Morocco. On reaching the Mediterranean it sometimes reacts with *mP* air already present to form depressions which subsequently pass eastwards. In summer the Mediterranean is warmer than the neighbouring parts of the Atlantic so that the *mT* air arrives cooler than the sea surface, the heating due to which causes feeble convection, changing the form of cloud from stratiform to cumuliform but not introducing a sufficient degree of instability to cause precipitation. In contrast to conditions in the British Isles visibility is generally good with *mT* air in the Mediterranean.

Continental Tropical Air.—In winter the chief source region of *cT* air is north Africa, whence it often moves to the Mediterranean as a southerly current in advance of a depression over the western part of that sea. Originally it is warm with a very low water-vapour content but on moving over the sea it rapidly picks up moisture and arrives at Malta and the north shores of the Mediterranean as the humid and oppressive sirocco (Chapter VIII). In summer c*T* air is drawn from north Africa, Syria, Palestine and the Balkans. At the source it is very warm with an extremely low

water-vapour content and the cooling consequent upon its movement over the sea gives rise to a large inversion of temperature in the lowest layers. If the winds are strong, mixing takes place and the water vapour acquired over the sea is diffused rapidly upwards so that the air retains its dry and dusty characteristics for long distances over the sea. If the winds are light little mixing takes place and the moisture accumulating in the lowest layers, the air very soon becomes oppressive in character.

North America—Atlantic.

The continent of North America consists of a broad land mass on its polar side, narrowing rapidly towards the south. It follows then that polar air is mainly continental in origin and tropical air maritime. The absence of an east to west mountain barrier allows the polar and tropical air to penetrate freely to low and high latitudes respectively, forming well defined fronts. Most of the frontal systems reaching western Europe originate over the eastern U.S.A. or the adjacent ocean. Owing to the contrasting properties of the polar and tropical air masses the eastern part of the U.S.A. is subject to very large fluctuations of temperature. The barrier of the Rockies largely prevents Pacific air from penetrating to the middle and eastern parts of the continent, but even if it does so penetrate it is so modified by the continental influences as to have lost all its original maritime properties. In North America no distinction is made between arctic and polar air, for originating over the snow covered regions of the Arctic and north Canada respectively, both are effectively continental in origin and are designated continental polar. The principal air masses with their usual tracks are shown in Fig. 132.

Continental Polar Air.—The main sources of continental polar air are northern Canada, Alaska and the Arctic. The cold air moves southwards under the influence of high pressure over Alaska with low pressure over Greenland and Labrador, and advances across the U.S.A. as a well-defined cold front, in the rear of which there is often precipitation. The front sometimes sweeps across the Gulf of Mexico ; the winds that follow it are the well-known Northers of that sea. Very often depressions form on the advancing cold front and move along it on to the Atlantic, whence they often travel to western Europe. Many of the depressions reaching the British Isles have originated on the cold front of an outbreak of polar air in North America. The southward moving cold air performs a long journey over land which, although to a great extent covered by snow in its northern part in winter, becomes increasingly warmer as the Atlantic coast is approached, rendering the air unstable. Owing, however, to its low humidity there is little convection, and over land and near the coast the air mass is characterized by clear skies or only a few Cu. clouds. Over the sea the air becomes very

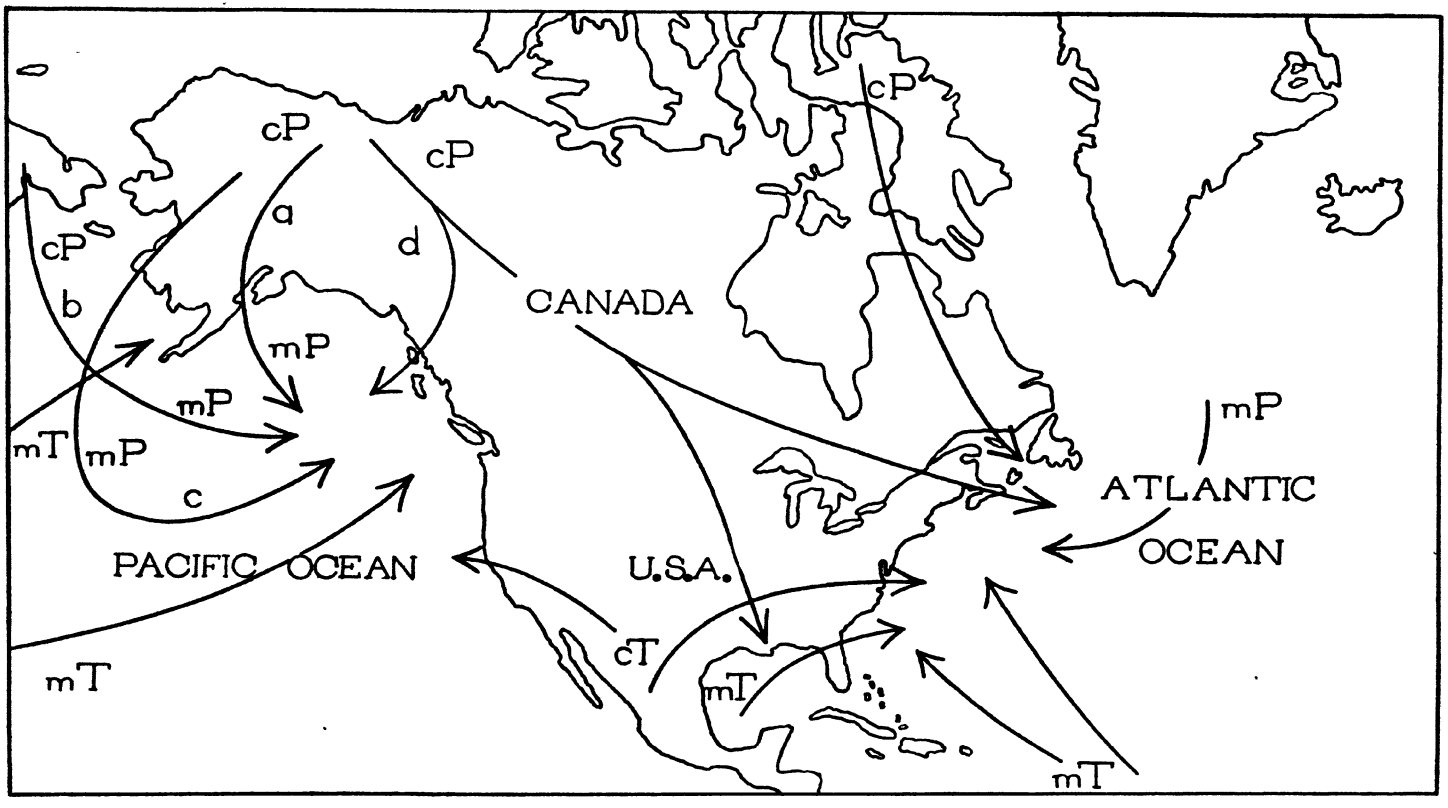

FIG. 132.—AIR MASSES OF NORTH AMERICA.

rapidly transformed into a moist unstable cloud- and rain-bearing mass with much Cu. development and showers. Depressions forming on the cold front usually move north-east along the Atlantic coast, the winds on their rear sides helping to strengthen the polar outflow.

In summer the ground over which the *cP* moves becomes warmed by insolation, and the lowest layers of the air are warmed by contact. It is impossible to tell from surface observations alone whether the air is continental or maritime in origin. The surface warming leads to instability, but the air is usually cloudless except for a few small Cu. in the afternoon. Near the south coast the prolonged heating is sometimes sufficient to cause an outbreak of thunderstorms.

Maritime Polar Air.—The source region of maritime polar air is that part of the North Atlantic extending north-east from the Gulf of Maine, over which the sea surface temperature is abnormally low throughout the year. The air mass is often modified *cP* air which reaches the Atlantic coast from north Canada and moving out over the ocean becomes subject to maritime influences. Occasionally it turns westwards round the anticyclone which, following the polar outbreak, often extends over the Atlantic from eastern Canada ; or it may return to American coastal waters under the influence of a depression moving north-east along the Atlantic coast of the U.S.A. In winter, owing to the coldness of the sea surface, the temperature of the air is little above the freezing-point, but upper-air observations show that there is often an inversion of temperature aloft, probably due to subsidence. The weather is generally overcast with low Stcu. clouds and poor visibility, while light snow flurries or rain and drizzle frequently occur. Maritime polar air, however, is most prominent in spring and early summer, when the sea, in its source region, is relatively coldest and contrasts with the warming land areas to the west. Whenever the normal west to east circulation is interrupted, the air stagnates over the cold ocean, forming a thermal anticyclone, and cold air may flow outwards along the Atlantic coast, penetrating occasionally as far as Florida. Since this air was originally *cP* air, which after being warmed over the continent is subsequently cooled over the sea, it is stable. As a rule, however, the cooling is not generally sufficient to produce fog ; the visibility remains good.

Continental Tropical Air.—Originating in the restricted areas of the southern U.S.A. and Mexico, continental tropical air plays only an insignificant part in the weather of American coastal waters. It is warm, relatively dry and cloudless.

Maritime Tropical Air.—The sources of maritime tropical air are the Gulf of Mexico, the Caribbean Sea and the North Atlantic to the south of Bermuda. The air is warm at all levels, stable with high water-vapour content in the lowest layers. Because of its stability it is associated with clear skies or cloud of Stou. type.

Only when it is lifted at a front by polar air does it give appreciable precipitation, which is then likely to be heavy. Normally, the low-lying coasts of the south-eastern states do not lift it sufficiently to cause rain. Moving north-eastwards along the Atlantic coast, the *mT* air is rapidly cooled at the surface by contact with the cold water. At first an inversion of temperature is produced with a layer of St. or Stcu., but as the air moves into the vicinity of Newfoundland, dense fog of great depth is formed which later develops into a fine drizzle. Fogs occur on the Newfoundland Banks on about 50 per cent. of days in spring and early summer, when the contrast between the air and sea temperatures is greatest. *mT* air occurs frequently in American coastal waters in summer, being drawn in by the low pressure which then develops over the heated continent. It is conditionally unstable, and on reaching land it often gives rise to an outbreak of thunderstorms.

North America—Pacific.

Continental Polar Air.—Practically all the polar-air masses of the north-east Pacific are continental in origin. The principal source regions are the Arctic Basin, north-east Siberia, Alaska and northern Canada. The air reaches the Pacific coast either by a long track over the ocean, which broadens rapidly towards the south, or by south-westward movement across Canada. In the former case the air is so modified by its ocean passage that it must be designated *mP* air. In the latter case it must cross the almost continuous barriers of the Rockies, and on descent to the coast be subject to föhn heating. Practically all the *cP* air reaching the Pacific coast consists of föhn-heated currents following tracks similar to *d* of Fig. 132. It is therefore warm, relatively dry and markedly stable, for descent tends to increase the stability of an originally stable air mass. A relative humidity of less than 20 per cent. is not unusual in this air stream, even several miles out to sea. Normally, the air retains its stability during its subsequent motion over the sea, which in this area is warmer than the average for the latitude. But if the *cP* air is originally very cold, even the mechanical heating on descent of the mountains is insufficient to raise the air to the temperature of the adjacent sea surface and movement over the sea very rapidly brings about instability. Then, showers of snow or sleet occur, sometimes even as far south as California. Also, if the *cP* air reaches the coast by way of the gaps caused by the Fraser and Columbia rivers, it undergoes only gentle descent with little dynamical heating. It then arrives as a very cold polar outbreak over the ocean, the warmth of which soon induces instability. If the air mass moves south off the coasts of Oregon and California the instability becomes pronounced. Secondary depressions then often form along the front between it and the *mP* air which it displaces.

Maritime Polar Air.—Maritime polar air is continental in origin and acquires its maritime characteristics during its long sea passage over the North Pacific. Three typical paths *a*, *b* and *c*, are shown in Fig. 132. Near to the source region the paths are labelled *cP* to denote the continental origin of the air. The same tracks are labelled *mP* after the air has travelled sufficiently far over the ocean to acquire maritime characteristics. As in western Europe, the extent to which the maritime influences modify the air depends upon the time spent over the ocean and also upon how far the air travels southward. Air traversing paths like *a* and *b* of Fig. 132 is unstable to a greater or less degree, and, having acquired moisture, is accompanied by Cu. and Cunb. clouds and showery weather. Air traversing a path such as *c* first becomes warmed and acquires moisture on its southward journey, then, turning north becomes cooled from below, perhaps almost to its dew-point, and stabilized in the lowest layers. Surface observations alone may not distinguish it from *mT* air. If *mP* air penetrates to the Hawaiian Islands, the heating from below penetrates to a very great height during its long passage over the warm sea, and the air becomes unstable to a marked degree.

Maritime Tropical Air.—The main sources of the tropical air reaching the north-east Pacific are the Java, Celebes and China Seas. The air subsequently moves north-eastwards round the sub-tropical anticyclone and interacts with polar air to form the depressions of the North Pacific, which occur with the greatest frequency in the region of the Aleutian Islands. Upper-air observations show that *mT* air is initially unstable, but as a result of its northward movement it becomes stabilized by cooling at the surface. At Honolulu the air is often unstable at all levels and produces heavy showers. At San Diego, as a result of cooling it is stable below about 3,000 feet and unstable above, and as at other sea-level stations in California it is characterized either by clear skies, or by low St. cloud with fog. Only on the mountains does it produce precipitation.

When *mT* air is carried to high latitudes between Japan and the Aleutian Islands, the cooling by the cold sea is pronounced. Widespread sea fog then occurs and occasionally spreads to the north-east Pacific.

Frontal Systems.—The majority of the frontal systems which originate in the Pacific do so in the neighbourhood of Japan, the China Sea or the Yangtze Valley. Depressions with well-defined fronts and open warm sectors are then formed and advance north-east towards the Aleutian Islands. By the time Alaska is reached the depressions are usually occluded and the majority of them subsequently die out. Sometimes, however, they are strongly re-generated over the Gulf of Alaska by a westward flow of cold air down the Yukon valley. This air reacts readily with that over

the gulf, which is warmed by the abnormally warm sea surface. As a rule, the depressions, with their occluded fronts, subsequently pass into Canada, but they may become stationary and die out slowly over the ocean.

Depressions known as Kona Storms often develop as waves on quasi-stationary fronts in the vicinity of the Hawaiian Islands. They are usually slow in developing, and often die out without attaining much vigour. Occasionally, however, when they form as secondaries on the fronts of systems situated further north they develop considerable energy.

Depressions are sometimes also generated off the Pacific coast of the U.S.A. and Canada by the interaction of maritime and continental air masses. Warm fronts are formed by occluded systems in the vicinity of Alaska drawing in *mT* air off the coast of California. Secondary cold fronts are formed by the outbreaks of cold *cP* air from Canada by way of the Fraser and Columbia rivers, which, interacting with the maritime air forms secondary cyclonic systems.

The China Seas.

As in North America, no distinction is made between polar and arctic air, for the source regions, North Siberia and the Arctic Basin, are similar, being largely covered by snow and ice. The cold air masses are therefore designated polar. They are effectively continental in origin, but acquire maritime properties if they move southwards over the ocean.

The flow of air in the western North Pacific is governed by the high- and low-pressure systems which form in summer and winter respectively over the land mass of Asia. The predominating air mass at any season is therefore closely associated with the monsoon, to which the seasonal distribution of pressure gives rise. In winter there is an almost uninterrupted outflow of cold air from north China and Siberia on the north-east side of the great Siberian anticyclone, and the principle air mass of the western North Pacific is therefore *cP*. In summer the flow of air is mainly from the south-east, circulating round the continental low pressure whose centre lies over north India, and the predominating air mass is therefore *mT*. Air masses with other characteristics than the two mentioned do, however, occasionally appear within the region. The tracks of the principal air masses of the China seas are shown in Fig. 133.

Continental Polar Air.—Continental polar air originates over Siberia and the arctic regions to the north thereof. In winter it is cold at all levels, especially at the surface, owing to the coldness of the land surface itself. Moving south-eastwards over the continent, it is subject to little modification apart from a slight surface warming, and arrives at the coast as a cold relatively dry current, characterized by clear skies. The flow is practically

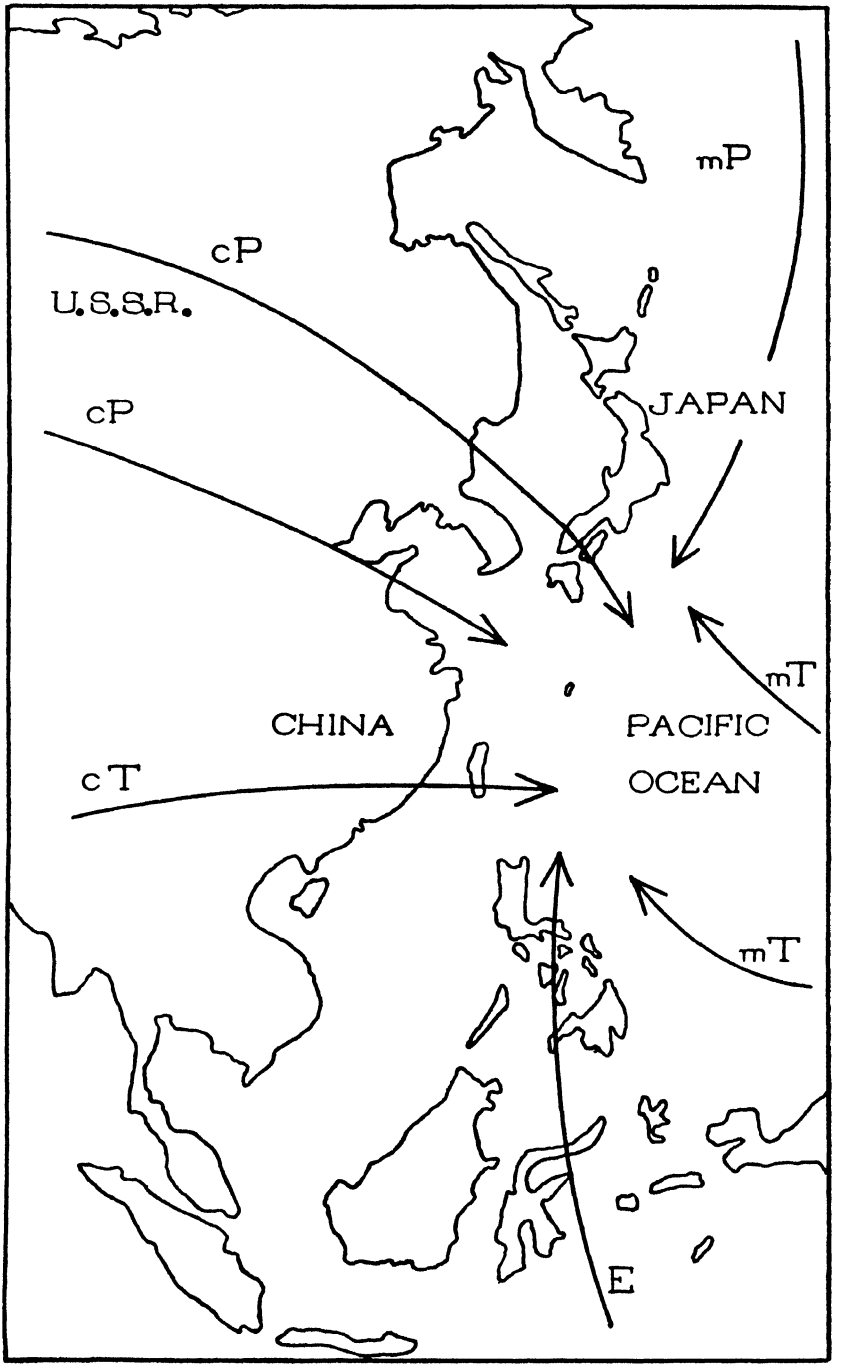

FIG. 133.—AIR MASSES OF THE CHINA SEAS.

continuous throughout the winter months. On passing out to sea, which off China and Japan is warm, it is rapidly heated from below, and acquires large quantities of moisture, causing it to become at first cloud-laden and later rain-producing owing to the instability generated; in fact it very rapidly becomes *mP* air. In summer, high temperatures occur in the source region and the *cP* air is probably unstable, although, owing to lack of moisture, little convection takes place. Since, however, in summer the general air flow is on shore, the China Seas are little affected by *cP* air.

Maritime Tropical Air.—Maritime tropical air originates in the sub-tropical anticyclone of the North Pacific, and during the summer constitutes the south-east monsoon of east Asia. The moisture content is high near the surface, but decreases rapidly above 5,000 feet. Moving north, the *mT* air is cooled from below and therefore stabilized in the lowest layers. Near the coast fog is produced, especially in spring and early summer, when the sea is relatively coldest. Off the coasts of Honshu and Kamchatka, washed by the cold Kamchatka current, fog occurs on half the days in summer. Over the open sea the replacement of *mP* air by *mT* air often leads to a general improvement of weather, the stability of the *mT* air being unfavourable both to cloud and precipitation. In the south, *mT* air is scarcely distinguishable from *mP* air, except when the latter has travelled very rapidly over the ocean, when its arrival may be accompanied by frontal rain and a fall of temperature.

Equatorial Air.—Equatorial air is met with in the south of the region during the summer monsoon. It has a high temperature and a high moisture content at all levels. Moving northwards it is cooled from below and stabilized. It does not therefore yield large quantities of precipitation over the open sea, nor in general over flat country. When, however, the land is subject to insolational heating or is mountainous in character, so that the air ascends either owing to thermal or orographical convection, the resulting trigger action often causes copious rainfall. If the equatorial air penetrates above 35° N. it becomes rapidly cooled by contact with the sea surface, whose temperature above that latitude falls off very quickly. The most dense and persistent fogs of the north-west Pacific then occur.

Frontal Systems.—It has already been mentioned that the majority of the frontal systems of the Pacific originate near to its western coast. This is essentially on account of the great contrasts of temperature between the continental and maritime air masses which meet in this region. On the continent itself depressions occur from time to time, but they are usually occluded and of no great intensity. They are of most frequent occurrence along the Yangtze valley, but also occur in north China. Moving off the land

to the warm ocean, they often intensify rapidly, developing new frontal systems and generally then move north-eastwards across the Pacific.

The Tropics.

The part played by fronts between interacting air masses in the weather of the tropical regions is relatively small. Much of the area is covered by tropical ocean of a uniformly high temperature. The proportion of land area within the tropics is small, so that the surface conditions are more uniform than in temperate latitudes, and the tendency is towards the formation of one extensive air mass with more or less uniform properties. These properties, which are most clearly defined in the belt of the doldrums which coincides approximately with the thermal equator, are high temperature and high humidity at all levels. Owing to the high water-vapour content, convection takes place readily over heated land and the rainfall is correspondingly heavy. The weather is subject not to the irregular variations such as are caused by the frontal systems of the temperate zones, but to regular diurnal and seasonal changes produced by the corresponding changes in insolational heating. The weather is therefore to be regarded as characteristic of the time of day or the season of the year. An example of the diurnal variation is furnished by the incidence of thunderstorms over tropical islands. These occur almost daily. After clear skies at dawn, Cu. and Cunb. clouds are gradually built up as a result of insolational heating and culminate in thunderstorms during the afternoon. The cloud clears in the evening. A similar sequence is repeated day after day. The seasonal variation is associated with the annual north-south migration of the pressure and wind belts. It is only within the doldrums that the conditions just described are fully developed. On either side of this belt are the trade-wind belts within which generally dry weather prevails. A particular place may at one time of the year be under the influence of the trade winds, and at another of the doldrums, and consequently experience a dry and a rainy season annually. It must, however, be emphasized that the rain is convectional and not frontal.

In the absence of orographic influences, as over the open ocean, the weather of the trade belts is fine and clear, since the wind usually blows from cooler to warmer seas, and consequently has a relatively low moisture content. Moreover, since the air of the trade-winds is usually air which has subsided in the sub-tropical anticyclone, it is usually stable. But when the trades blow over a warm ocean they acquire sufficient moisture so that if forced to ascend at a mountain barrier they may yield heavy rainfall. Such orographic rain occurs in south-east Africa and in east Australia.

Near to the western sides of the tropical oceans the uniformity of the weather is disturbed from time to time by tropical storms (Chapter XIX). The cause of their formation is unknown, but

instability in a slowly-moving mass of warm damp air over a warm sea surface appears to be an essential condition. This condition is fulfilled by the air masses of the areas affected. Tropical storms are not as a rule encountered on the eastern sides of the ocean, for these are under the influence of the cool, stable and relatively dry trade-winds.

The Southern Hemisphere.

No complete air-mass analysis for stations in the southern hemisphere appears yet to have been carried out. It is therefore only possible to indicate the probable general characteristics of the air masses and what follows must be largely conjectural.

Australia.—A considerable proportion of the area consists of tropical ocean, and is therefore appropriate to the preceding paragraph. The continent lies within the sub-tropical high pressure and the south-east trade belts, and is sufficiently large and uniform to act as a source region for *cT* air. The other two important air masses are *mP* and *mT*, originating over the southern ocean and tropical seas, respectively. The polar and tropical air masses interact to form depressions with associated frontal systems which pass eastwards in regular succession. The majority of tracks lie to the south of Australia in the belt of the "roaring forties." Only the southern part of Australia is in general affected by these frontal systems, and then only in the winter half-year. In summer the seasonal swing of the pressure belts causes the depressions to pass too far to the south to affect any portion of Australia. Within the region of influence of the depressions the weather is similar to that experienced in the British Isles. The clockwise circulation of the winds round the centres of low pressure should, however, be noted, and also the fact that the warm sector lies to the north of the centre, with a corresponding orientation of the fronts.

Throughout most of Australia and the neighbouring seas there is little interaction between air masses. Over the ocean in the latitude of the continent the weather is typical of the trade-wind and sub-tropical high-pressure belts, fine and clear. But the seas to the east of Australia are so warm that the winds acquire a high moisture content in passing over them, and, despite the prevailing anticyclonic conditions, the moisture is deposited as orographic rain on the mountain range parallel to the coast. In summer, the monsoon brings equatorial air, warm and humid, to northern Australia, which experiences rain at this season along the coast; but the rainfall rapidly diminishes with increasing distance from the sea. North Australia is subject to tropical cyclones, known as "willy-willies" in the north-west and as "hurricanes" off the Queensland coast. It is unusual for tropical cyclones to occur on the western side of a continent, since these are, in the main, washed by cold currents. They do so in Australia because the sea there is relatively warm. Sea fog does not occur off Western Australia,

whereas it is frequent off the corresponding cold-water coasts of Africa and America.

The most noteworthy frontal phenomenon in Australia is the "southerly buster," a violent line squall which sometimes accompanies the passage of a cold front over south-east Australia. Fig. 134

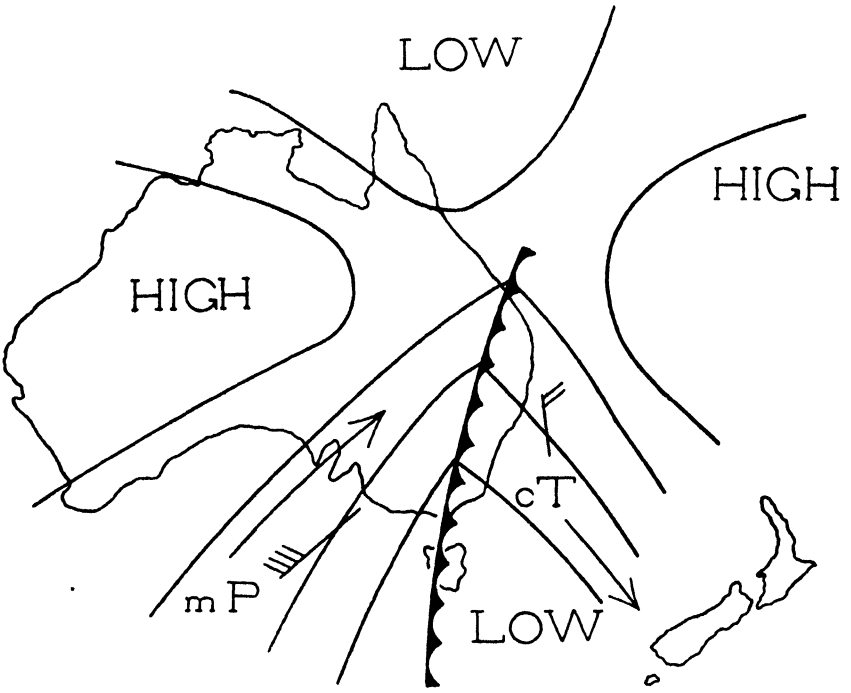

FIG. 134.—SOUTHERLY BUSTER.

shows the usual configuration of the isobars with the cold front approaching south-east Australia from the interior. The north-westerly winds in advance of the front consist of tropical air, which, since it is drawn from the interior of the continent, is exceedingly hot, dry and often dust-laden. In Sydney this wind is known as the "brickfielder." The air behind the front is *mP* and to reach Sydney, where the "buster" is most pronounced, it has to cross only a small part of the continent and therefore retains its cool maritime characteristics. The south-west wind sets in suddenly, often with a violent squall and a large fall of temperature amounting to 20° F. or more. Sometimes heavy rain occurs, but not always, since the warm air is usually dry.

New Zealand.—The southern part of New Zealand lies in the paths which are most frequently followed by the depressions of southern temperature latitudes and the weather experienced is very similar to that of the British Isles. Surrounded on all sides

by an ocean of vast extent New Zealand experiences only maritime air masses. Even the *cT* air from Australia acquires so much moisture in crossing the Tasman Sea that it gives heavy orographic rain on the mountains near the western seaboard. Frontal rain is likely in all parts of the Dominion but especially in the south.

South Africa.—Except for the immediate vicinity of the Cape Province, in winter (southern hemisphere) when the wind and pressure belts lie farthest north, South Africa does not come under the influence of the depressions of the southern temperate zones. The ocean to the south of the Cape is, however, subject to extremely stormy conditions and owing to the fact that the interacting air masses have pronounced maritime properties there is considerable rainfall at fronts with, also, frequent thunder and lightning. Violent squalls, similar to the southerly buster, are experienced in South Africa, especially near Durban whose geographical situation is similar to that of Sydney. Elsewhere on the Africa station frontal systems are not generally encountered since a large part of the region lies within the equatorial and trade-wind belts.

CHAPTER XXI.

General Forecasting

The Issue of Forecasts by H.M. Ships.

Forecasts issued by meteorological officers in H.M. ships and other naval establishments should, in general, be on the lines indicated in Chapter XVII. When naval and air operations are being carried out they should be issued with special reference to the meteorological conditions which may affect these operations.

A complete weather forecast consists of three parts, namely, the General Inference, the Forecast proper and the Further Outlook.

The general inference is a brief statement of the existing pressure distribution and the anticipated changes due to the movement of depressions, anticyclones, etc. The position and movement of well marked fronts are also included. The date and time of the synoptic chart on which the general inference is based should be stated.

The forecast should consist of a concise and definite statement regarding the meteorological conditions for a specified area and period. These should always be stated explicitly.

The further outlook is a statement in general terms of the meteorological conditions in the period immediately following that to which the forecast applies. When the future course of events is reasonably certain the forecaster may issue a further outlook with some confidence but on the frequent occasions when the meteorological situation is indefinite the forecaster will not feel justified in issuing a further outlook. In such a case the term "none issued" should be used.

There are two classes of forecasts which meteorological officers will be most frequently called upon to issue. The first is the general (sea-area) forecast for ordinary purposes. The second, which has become increasingly important since ships have been carrying aircraft, is the aviation forecast. These two may be regarded as basic forecasts, and the lines on which they should be drawn up will now be discussed.

General Forecast.—The elements included are wind, weather, visibility, sea, swell and sometimes also temperature.

Wind.—The probable direction and speed of the wind at the surface should be given. The direction should be stated, preferably in degrees from true north, but may also be given in compass points. The speed should be given as the force on the

Beaufort Scale, or in descriptive terms which are related to the Beaufort force as described in page 332.

If the direction is given in degrees and the speed in knots, round numbers should be used, to avoid giving the impression that it is possible to forecast the wind accurately to one degree and one knot. It will usually be found necessary to state a range of 10° or 20° as regards direction and 5 to 10 knots as regards wind speed. Thus a typical forecast might be "Wind 220° to 240°, 10 to 15 knots." If gusty or squally winds are anticipated this should be mentioned.

Weather.—Under the term weather are included the state of the sky, precipitation, fog and miscellaneous phenomena such as thunderstorms, dust-storms, line-squalls, etc.

To describe the state of the sky or general character of the weather in the absence of precipitation, the following terms are used:—fine, fair, cloudy or overcast.

The terms "fine" and "fair" must never be used in conjunction with precipitation. When precipitation occurs the sky is usually cloudy or overcast, but it may occur occasionally with less than 8-tenths of cloud. The term "variable cloud" may then be used, for the precipitation is usually of the shower type and accompanied by rapid variations in the amount of cloud. As regards precipitation, the nature, locality, intensity, time of occurrence and continuity should be stated. The following terms are typical of those used:—

Intensity.—Heavy, slight, thick (drizzle, fog), severe (thunderstorm, line-squall).

Locality.—Local, general, widespread, scattered, in many places, in most places, name of district or sea area in which phenomenon is expected.

Time of occurrence.—Occasional, frequent (showers), intermittent, at times, continuous.

Visibility.—A more precise statement of visibility than is included as regards fog under the heading "weather" should be given. The distance of visibility may be stated, or, alternatively, descriptive terms may be used related to the code figures for visibility (page 47) as follows:—

Code Figure.	Descriptive Term.
0, 1	Thick fog
2, 3	Fog
4	Mist
5	Haze or poor visibility
6	Moderate visibility
7, 8	Good visibility
9	Excellent visibility

State of Sea.—In the open sea the state of the sea may be related to the wind as follows :—

Wind (Beaufort Force).	Sea.	
	Description.	Mean height of waves (feet).
0	Calm—glassy	—
1	Calm—rippled	½
2—3	Smooth—wavelets	1—3
4	Slight	5
5	Moderate	9
6	Rough	14
7	Very rough	19
8	High	25
9—10	Very high	28—40
11—12	*Phenomenal	45 or over

* As might exist at the centre of a hurricane.

The descriptive terms in the second column should be used in association with the winds as specified. In harbour and in coastal waters account must be taken of the wind direction and the sheltering effect of the land. Only experience will guide the forecaster at any particular place. (See also page 458).

Temperature.—The probable average temperature may be stated in degrees Fahrenheit, a range usually being given. Often, however, it is more informative to relate the anticipated temperature to the present temperature by the use of such terms as " becoming colder," " temperature rising decidedly," " little change of temperature," etc.

Aviation Forecast.—The forecast for aviation may be required for reconnaissance, flying in a specified area, or for a flight over a specified route. It should include the following elements :— Wind at the surface, upper wind, weather, cloud amount and height, visibility.

Wind at the surface.—The specification should be similar to that in the sea forecast, except that the speed should always be given in knots.

Upper wind.—Unless otherwise specified the wind at a height of 2,000 feet should be given. This is a convenient height for flying, and, moreover, in the absence of upper-air observations the direction and speed of the wind at 2,000 feet may be obtained from the isobars on the current synoptic chart. Winds at other heights may be required. If possible the most favourable wind for a particular

flight should be given, *e.g.*, for a flight from Plymouth to Portsmouth, if the wind direction is 135° at 2,000 feet and 220° at 6,000 feet, obviously the latter is the better flying height and a statement of the wind at that height should be included in the forecast.

The direction of the upper wind should be given in degrees from true north and the speed in knots.

Weather.—This should be on the same lines as in the general forecast. A note on ice accretion should be added, *see* p. 337.

Cloud.—The forecast should state the amount and height above sea level of low cloud, that is, cloud below 8,000 feet. The amount should be specified in tenths of sky covered, usually as a range, such as, 0 to 3-tenths, 3- to 7-tenths, etc. If 10-tenths of cloud are expected at any place and time during the flight, special mention should be made of the fact. The height of the cloud should be given in feet above M.S.L., using round numbers. The occurrence of very low cloud at any time and place, and also whether high ground is likely to be enveloped by cloud, should be noted.

Visibility.—The horizontal visibility should be stated in yards or miles. Localities in which the visibility is likely to be poor or bad should be specially mentioned.

The remarks regarding the use of round numbers and a range of numbers under the heading "Wind" in the sea forecast, apply also to the upper wind, height of cloud and visibility.

THE PRACTICE OF FORECASTING

In the present chapter the practice of forecasting with the aid of synoptic charts is discussed from four aspects as follows :—

(A) The recognition of the isobaric forms described in Chapter XVIII, with their associated weather and anticipation of their formation, travel and disappearance.

(B) The recognition of fronts and occlusions, with their associated weather and anticipation of their development, travel and dissipation.

(C) The properties of air masses. Many important weather phenomena, for example, convectional rain, thunderstorms, fog and cloud, arise as a result of physical changes within an air mass remote from fronts. The occurrence of these phenomena depends upon the properties of the air mass, particularly as regards its lapse-rate and water-vapour content, so that in this connexion it is necessary to consider observations of upper-air temperature.

(D) Diurnal, seasonal, orographic and other local effects. The forecasts arrived at on the basis of (A), (B) and (C) above, are modified by considerations such as the time of day, the season of the year and the locality. Some typical effects of these kinds are :—

Diurnal effects.—The tendency of fog to form over land at night ; convection over land on summer afternoons with the outbreak of showers or thunderstorms.

Seasonal effects.—The occurrence of sea fog in late spring and early summer; the contrast between summer and winter anticyclones. (Chapter XVIII.)

Local effects.—The prevalence of sea fog in various parts of the world (Chapter IX); local and regional winds (Chapter VIII); orographic rain.

These effects should always be borne in mind in preparing a forecast of weather. It must be emphasized that the four aspects of forecasting listed above are mutually dependent. In preparing a forecast, due consideration must be given to each of them. For convenience of treatment they are, however, to some extent, discussed separately in the present chapter.

(A.) The Method of Forecasting Based on Isobaric Forms.

General.—The aim of the forecaster is to derive a picture of the pressure distribution at any time during the period of the forecast. It is often useful to make a rough sketch on a blank chart, of the anticipated pressure distribution at the end of the period. Not only the current chart, but the immediately preceding charts should be considered, so that the past history of the anticyclone or depression may be known. The previous path of a depression often gives information regarding its future movement. It may not be possible to tell, without reference to past charts, whether a depression is deepening or filling up.

If forecasting regularly for the same area, the synoptic charts should be retained for reference when similar situations arise in the future. It may reasonably be expected that the development of a synoptic situation will follow that of a similar predecessor, provided that the time of year is the same for both. To some extent this is a question of memory and experience, but it would be useful to make a list of charts classified according to the positions of the highs and lows and to compile an index of the dates on which each type occurred. Previous examples of a given type of pressure distribution could then be rapidly looked up. A classification of pressure types for the British Isles and an extensive index of their dates of occurrence has been prepared in the Meteorological Office by Lt.-Col. E. Gold, F.R.S. (*Meteorological Office, Geophysical Memoirs No.* 16, *Aids to Forecasting*).

It must, however, be strongly emphasized that the prevailing weather is not dependent upon the pressure distribution alone. For example, two depressions, similar as regards their shape and number of isobars, may be associated with different distributions of temperature, visibility, cloud and rainfall. Therefore, in forecasting, the characteristics of the individual depression or anticyclone, as shown by the station and ship reports must be taken into account. There is, moreover, a peculiar tendency to "moods" in the weather. In this connexion, Sir Napier Shaw (*Forecasting*

Weather, p. 395) remarks upon " the peculiarity which the weather seems to display of taking on 'moods' . . . If a thunderstorm occurs on one day in any part of the country, it is quite common for other parts of the country to be visited by similar phenomena on the next day or the day after, although the meteorological conditions do not necessarily point specially to the occurrence. . . . Why the conditions of instability should be set up in one locality because there was instability in another locality yesterday or the day before, it is difficult to see, but no forecaster can fail to draw the conclusion from experience. It is peculiarly true of thunderstorms and showers that they are conspicuously local, yet at the end of a spell few places seem to have been exempt. Moreover, the same habit is almost equally characteristic of rainfall. After a dry period, the weather sometimes seems unable to rain even under barometric conditions which are apparently most favourable for it, and on the other hand, rain sometimes falls without any recognisable meteorological reason, even as we have seen, in the central region of an anticyclone. The peculiar 'mood' or 'fit' of the weather is a great difficulty for the forecaster."

The forecast should be based on the general characteristics of the types of pressure distribution present as given in Chapter XVIII, together with any peculiar characteristics of the particular situation. In addition, with depressions and anticyclones, certain rules as set out in the two succeeding paragraphs may be applied.

Practical Rules for the Movement and Development of Depressions.—

The following rules are applicable to both hemispheres.

1. The centre of a depression which is not occluded will move in a direction parallel to the isobars in the warm sector, with the logical consequence that an occluded depression becomes almost stationary, except in those cases where the depression is a " secondary " in the general current with which it moves.

This rule is derived from the hypothesis that the depression is formed by the intrusion of warm air above cold air and hence the movement of the depression will be governed by that of the warm air, which, in the lower layers at least, is determined by the isobars in the warm sector. This rule was formulated by the Norwegian meteorologists and is regarded as one of the most important precepts of modern forecasting.

2. A depression will probably move towards the area of greatest falling barometric tendency, *i.e.*, towards an isallobaric " low."

This assumes that the barometric tendencies at a given instant are likely to persist. An isallobar is a line drawn through places at which the barometer has risen or fallen by the same amount during some specified period, *i.e.*, through places having the same barometric tendency. Regions in which the isallobars have

the largest positive values are known as "isallobaric highs," in the same way that regions where isobars have their highest values are termed "isobaric highs." Similarly, a region where the barometer is falling most rapidly will be termed an "isallobaric low."

3. A depression with winds above normal on all sides will fill up, while one with winds below normal will deepen.

4. Those parts of a depression in which winds are below normal indicate directions in which the depression may advance.

5. Divergent winds indicate the approach of a low-pressure system.

6. Convergent winds indicate the departure of a low-pressure system, or a rise of pressure.

Rules 3, 4, 5 and 6 were formulated by the French meteorologist Guilbert, in 1891.

Divergent winds are those which blow away from a low-pressure area, while convergent winds are those which blow more directly towards a low-pressure system than is normal. The wind circulation round a depression is counter-clockwise in the northern hemisphere, clockwise in the southern hemisphere with an incurvature or angle of indraft towards the centre of 30 to 40 degrees over the land and 10 to 20 degrees over the sea. An incurvature in excess of this value is abnormal.

The following table shows the distance apart, in nautical miles, of 2 mb. isobars for normal winds, corresponding to values of wind force on the Beaufort scale up to force 8. Provided the force of the gradient wind in any synoptic situation can be correctly determined, comparison with the values given in this table will show whether such winds are above or below normal.

Normal Wind, Beaufort scale.	Distance apart of 2 mb. isobars in nautical miles.
1	180
2	90
3	60
4	45
5	36
6	30
7	25
8	22

The fundamental difficulty in the application of Guilbert's rules, however, lies in the observation and recognition of normal, divergent and convergent winds. If all observing stations had perfect exposures to the wind, with no mountains and few trees or houses in the vicinity, the difficulty would not arise, but this ideal is rarely attainable, and in practice all observing stations, including ships at sea, have obstructions to the free flow of air in one or more directions. This means that a wind may appear

abnormal when compared with the pressure gradient because of local peculiarities; the reason may not lie in any abnormality in the atmospheric conditions, such as might reasonably be expected to influence the movement of the depression.

7. Small depressions tend to move with the general current in which they are borne.

Large depressions, however, so modify the air currents that this rule cannot be applied. In the case of two centres forming a "dumb-bell-shaped" depression, the system tends to rotate counter-clockwise in the northern hemisphere (clockwise in the southern hemisphere) about some point intermediate between the two centres.

8. Each successive depression of the same family crossing a given meridian, follows a track which is nearer the equator than that of its predecessor.

9. Depressions that are small and active tend to move quickly, while large and dying depressions tend to become stationary.

10. A depression deepens with increase in the instability of the surrounding air masses; hence, other considerations being equal, in winter a depression may deepen when passing from land to sea and in summer when passing from sea to land.

11. A pronounced polar current behind a depression greatly favours its deepening. Marked development is therefore more likely to take place at a straight front having a south to north orientation than at a straight front having a west to east orientation.

12. Up to the stage where the depression becomes occluded, the rate of advance increases, while after occlusion it diminishes.

13. After occlusion the path of a depression tends to curve to the left in the northern hemisphere and to the right in the southern hemisphere.

14. An old occlusion is often divorced from the depression with which it is associated while the latter lags behind or remains stationary.

15. In the case of the regeneration of a depression (due to the intrusion of cold air from outside) the rate of its movement increases.

16. If on the synoptic chart there appears in the region of a depression a zone of noticeably higher temperature in contrast to one of lower temperature, then the depression moves in a direction perpendicular to the line connecting the belts of high and low temperature, *i.e.*, to the horizontal temperature gradient. The higher temperature lies to the right of the path of the depression in the northern, and to the left in the southern hemisphere.

17. If the negative barometric tendencies in advance of a depression are of the same order of magnitude as the positive tendencies in its rear, the depression is not changing in intensity. If the negative tendencies are the greater, the depression is deepening; if the positive tendencies are the greater the depression is filling up.

The signs of development of a depression are as follows :—

(i) Falling barometer.
(ii) Cyclonic circulation of wind, *e.g.*, the appearance of an easterly or southerly wind at a station in a hitherto general westerly current.
(iii) Slackening of the pressure gradient shown by the opening out of the isobars in a broad unidirectional airstream.
(iv) The appearance of the frontal clouds, Ci., Cist., Ast., Nbst.
(v) The appearance of precipitation other than showers.

These should be regarded as danger signals whenever they appear, especially if in conjunction with one another. Since most depressions develop where two masses of air of widely different temperatures and humidities are in juxtaposition, *i.e.*, at a front, the above signs are particularly significant in such circumstances. They should be specially watched for along the line of a back-bent occlusion or of an old dying front, even if its presence is only suspected from previous charts and no evidence of its existence remains on the current chart. In the absence of fronts the signs may indicate (for example) the formation of a depression in polar air, or the appearance of a small secondary in straight isobars.

Rapid movement of high (cirriform) cloud from the NW. (in the northern hemisphere), as shown by nephoscope observations, often indicates the existence of a deepening depression to the west. Fig. 135

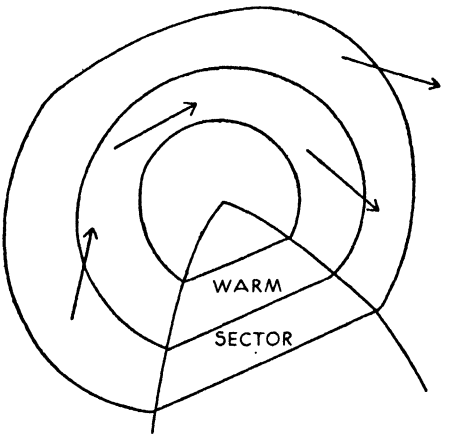

FIG. 135.—MOVEMENT OF HIGH CLOUD IN A DEPRESSION.

illustrates the movement of high cloud in advance of a depression with a well-defined warm sector ; the wind at high level circulates clockwise round the region of high temperature, *i.e.*, the warm sector, and therefore on its eastern side moves from the N. or NW. whereas the wind at the surface, conforming to the isobars, is

from S. or SW. In countries such as the British Isles and in ships where observations to the west may not be available, such rapid movement of high cloud is sometimes the only indication of an approaching depression. In the southern hemisphere the movement of high cloud in advance of a depression is from the SW.

The Movement of Anticyclones.—The movement of anticyclones is neither so regular as that of depressions nor are the signs of their formation so definite. The probable immediate direction of movement is given by the barometric tendencies if these are consistently greater on one side of the high than on the other; the high is then likely to move towards the region of greatest positive tendency. If the positive tendencies in advance of the high are generally greater in magnitude than the tendencies (positive or negative) in its rear, the high is increasing in intensity; if the negative tendencies are the greater, it is decreasing in intensity. A general rise of pressure over the area covered by an anticyclone shows that it is building up and likely to persist; a general fall of pressure that it is likely to collapse. A warm anticyclone, *i.e.*, one in which the temperature is high in the upper levels, is likely to persist longer and to move more slowly than a cold anticyclone. An approaching front may sometimes cause the complete disappearance of an anticyclone, but often only a limited area of the anticyclone nearest the front is affected.

In an intensifying anticyclone, the subsidence inversion which retards convection, intensifies and settles lower day by day.

The centre of an anticyclone moves in the direction in which the temperature is falling the fastest.

The approach of a new cold front (cold air outbreak) into an old anticyclone is accompanied by a weakening of the latter.

(B) Recognition of Fronts on a Synoptic Chart.

The synoptic charts referred to in this chapter are plotted according to the method described in Chapter XVI, either in full or abbreviated, except Figs. 121 and 122, for which see footnote on page 339.

The practical application of the polar-front theory of depressions to the elucidation of the weather phenomena associated therewith depends primarily upon the recognition of fronts on a synoptic chart. A front (and the term "front" where the context permits is taken also to include an occlusion) is a line across which some or all of the elements of pressure, wind, weather, cloud, temperature and dew-point are discontinuous. The sequence of such elements experienced during the passage of fronts has been described in Chapter XVIII, so that from these considerations it is possible to recognise and draw fronts on a synoptic chart by a careful examination of the meteorological observations made by a network of stations. The general method of identification of these lines of

discontinuity will first be discussed and will be illustrated subsequently by examples taken from synoptic working charts.

The most important element in the definition of a front is temperature, and, in fact, the difference of temperature across a front may be regarded as the cause of the discontinuities in the other elements. The temperature of the surface air is, however, modified to a large extent by the land or sea surface over which the air moves, by insolational heating or nocturnal cooling, so that the original difference of temperature on the two sides of a front is often obliterated. For this reason, surface temperature is often the least reliable of the above elements in the identification of a front. Temperatures at, say, 2,000 feet would afford a much more accurate criterion, but such temperatures are not usually available.

Of the other elements, pressure and wind usually exhibit the most prominent discontinuity, and it is with reference to these that a front may be most easily located. The pressure distribution usually shows a more or less well-defined trough or sharp refraction of the isobars, and this is necessarily accompanied by a discontinuity of wind direction and speed. If the front is diffuse, as is frequently the case, the sharp refraction of the isobars is smoothed out, and if the zone of transition is wide, the isobars at the zone must be rounded, so that the change of wind is gradual. A rounded trough, however, is not always a front so that additional evidence is required in such cases.

Barometric tendencies are also of great value in identifying fronts; normally the tendencies are negative in advance of an occlusion or warm front and positive in the rear of an occlusion or cold front, but they are sometimes modified by general pressure changes, due to the deepening or filling up of the associated depression or the approach of new systems. In general, it may be said that tendencies are " more positive " in the rear of a front than in advance of it. Thus, if the tendencies are negative on both sides of a front the greater tendencies occur in advance of the front; if the tendencies are positive on both sides of the front, the greater tendencies occur in the rear. The negative tendencies reported immediately in the rear of a front behind which there is a general rise of pressure, are sometimes misleading, owing to the fact that the amount of the barometric tendency is the net change of pressure in the 3 hours preceding the time of observation, and although at that time the barometer may be rising, the pressure may not have recovered its value of 3 hours previously. For example, suppose that a front passes the observing station one hour before the time of observation, and that before its passage the pressure was falling at the rate of one mb. per hour and has since been rising at the same rate. The 3-hourly tendency would then be—1 mb.; the fact that the barometer has commenced to rise would be shown by the characteristic " falling then rising."

Other useful factors in the recognition of fronts are the types and heights of clouds, and the presence and nature of precipitation. The evidence in favour of the association of rain and fronts is so overwhelming that if it is known from other evidence that warm and cold air are in juxtaposition in a certain region, it is reasonable to connect with this fact any rain which falls in that region. Considerable caution is necessary, however, in making such use of precipitation to define a front, since the rain often appears ahead of the front, and also because a few casual showers may appear in line on the chart without being due to the existence of a front. The pre-frontal cloud sequence and wide belt of precipitation serve to identify a warm front or occlusion; similarly the narrower belt of cloud and precipitation help to determine a cold front. It should be remembered, however, that much stratiform cloud and widespread precipitation often occur in a warm sector: the large cumulus and cumulonimbus clouds and showers which occur up to considerable distances in the rear of the cold front or occlusion serve to identify the nature of the post-frontal air mass, and hence, the type of front.

Dew-points are also of use in recognizing fronts, though they are apt to mislead if deduced from observations when rain is falling, since they are then raised by evaporation. Dew-points, indeed, are a more conservative element of an air mass, than temperature; hence a discontinuity in the dew-point may often be detected at a front when the temperature discontinuity has become obliterated. When, however, a warm front advances, most of the rise of dew-point may take place before the warm front arrives, owing to the rain. There is, too, another effect to guard against. Both wet-bulb temperature and dew-point rise quickly in polar air over the sea, especially when the layer of cold air is shallow. There may then be large differences between inland and coastal stations, especially in the morning in winter. Sometimes there are degenerate cold fronts behind which the dew-point does not immediately fall. These are most frequent in summer when the air over the sea tends to be more stable than in winter, but they occasionally occur in winter if the air has a very long sea track.

Recognition of a Warm Front.—Fig. 137 shows the synoptic situation over the British Isles and the western European continent at 0700 G.M.T. on the 18th June, 1935. A warm front associated with a depression over the Irish Sea is shown running east-south-east across England and thence southward through Belgium and France. The individual weather conditions leading to the identification of the warm front will now be discussed:—

Pressure and wind.—A glance at the wind observations shows a discontinuity, in the form of a shift of wind from south-east to south or south-west, running approximately east-south-east across England from north Wales, and continuing over Belgium and France, the line of discontinuity passing between stations 133 and

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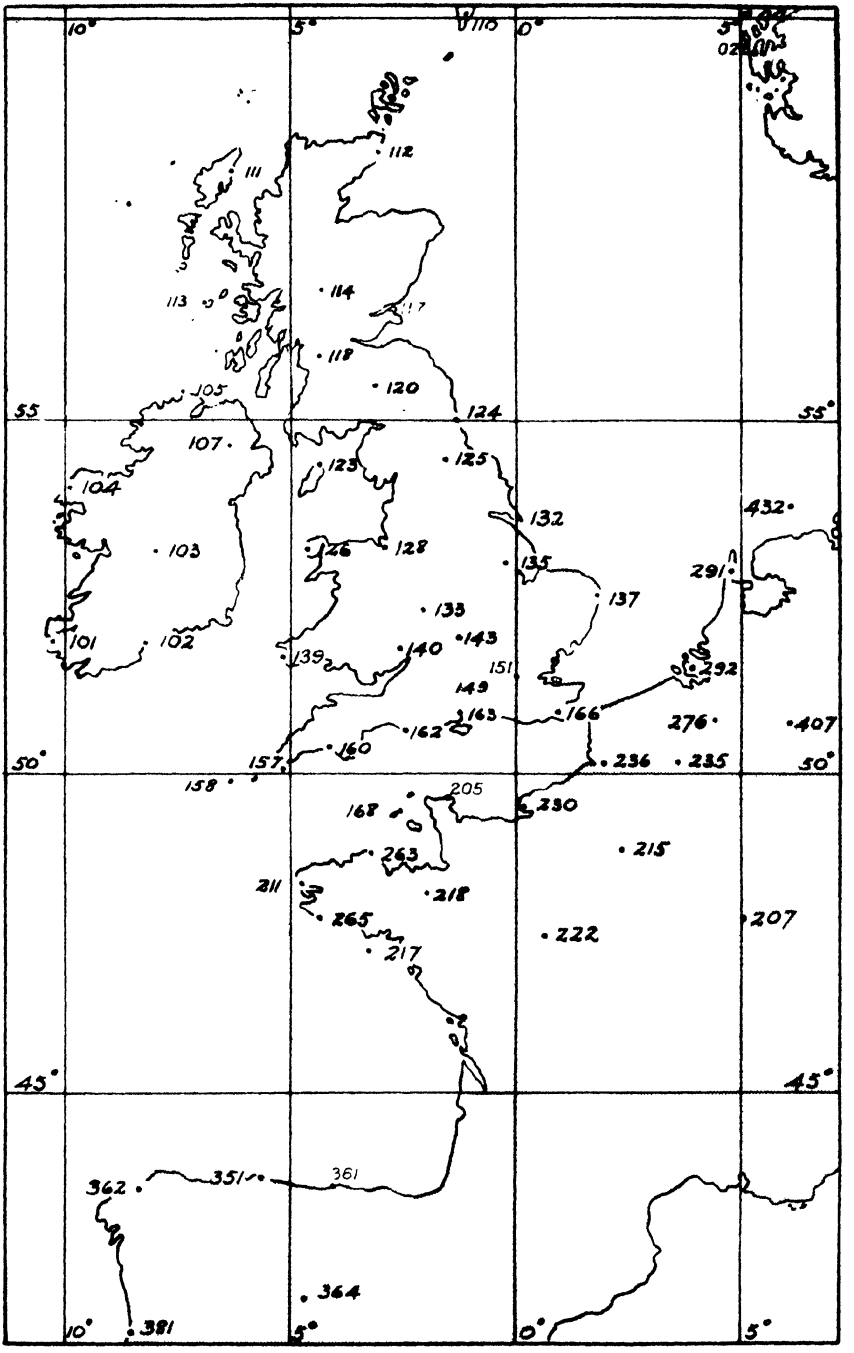

FIG. 136. KEY MAP OF STATIONS REFERRED TO IN CONNEXION WITH TYPICAL CHARTS (FIGS. 137-154)

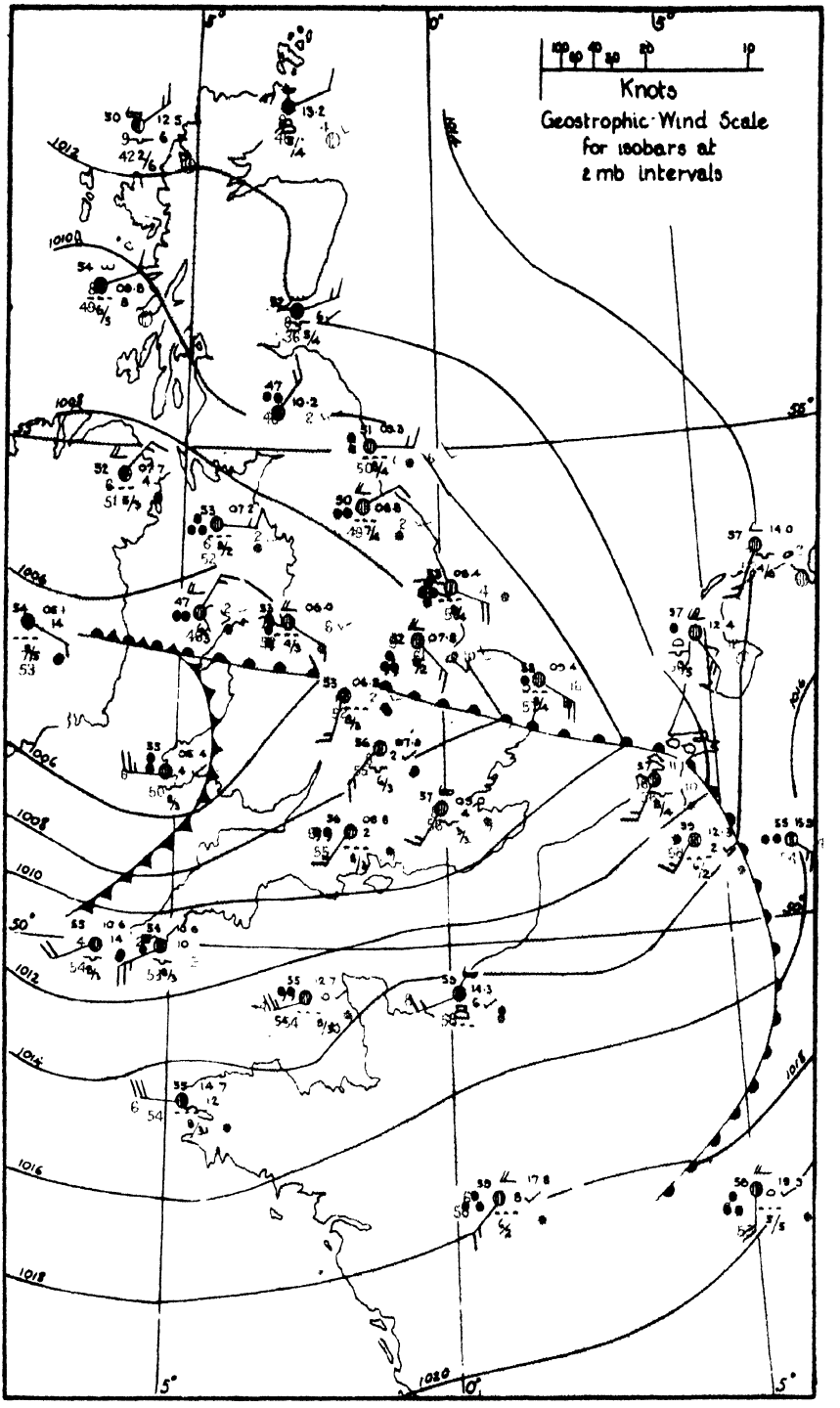

FIG. 137. WARM FRONT 0700 G M T 18 JUNE 1935

135, between 276 and 407, and thence between 222 and 207. Such discontinuities alone indicate the existence of a front and so the isobars can be drawn roughly as shown and the approximate position of the front determined, passing through the points where the isobars show a sharp refraction.

Temperature.—It will be seen that the temperatures at the inland stations 143 and 135 differ from one another by as much as 4° F., while between the continental stations 276 and 407 there is a similar difference; in each case the higher temperature prevails to the south-west or west of the front, *i.e.*, in the warm sector.

Barometric tendency.—On the north-east side of the front the barometer shows falling tendencies, *e.g.*, stations 132, 291, 432 and 407, while to the south-west, observations generally indicate a change from falling to steady or rising, *e.g.*, stations 133, 292 and 276.

Precipitation.—Over a wide area in advance of the front the rain is continuous, the rain-belt being bounded roughly by the front and a line through stations 103, 107, 120, 124 and 291. At a number of stations the precipitation is moderate or even heavy. Behind the front, precipitation at the majority of observing stations is slight and intermittent and characteristic of the warm sector of a depression. It will be noticed that near the south and south-west coasts of England rain and drizzle are reported, while fog is observed at station 157.

Dew-point.—Across the front from NE. to SW., dew-points have risen corresponding with the rise of temperature. Thus, at stations 125, 291 and 407 the dew-points are 49°, 50° and 54° F., respectively, while to the SW. of the front at stations 133, 292 and 276 they are respectively 55°, 56°, and 58° F. Near to the front where rain has been falling for some time, the dew-point has been raised by evaporation from the rain, *e.g.*, at stations 137, 135 and 128 the dew-point readings are 57°, 51° and 52° F. respectively.

Cloud.—The clouds in advance of the front are of the characteristic warm-front type, some or all of altostratus, nimbostratus and fractonimbus being reported at stations 128, 125, 135 and 432. In the rear of the front there is much low and medium cloud of the stratiform type. It should be noted that almost all types of stratiform cloud can occur within a warm sector, so that the existence of altostratus and nimbostratus is only significant for weather in advance of a warm front when accompanied by the other signs discussed above.

From an examination of earlier charts it is known that the front has a general movement from SW. to NE., as would be expected, so that the considerations enumerated above lead to the conclusion that the line of discontinuity is a warm front.

Recognition of the Approach of a Warm Front.—The synoptic information supplied by observing stations in certain regions may

indicate the approach of a warm front, although it may not be possible to locate the position of the front owing to the absence of observations taken at the front itself and to the rear of it.

Fig. 138 shows the synoptic chart of the British Isles and France drawn for 1300 G.M.T. on the 17th of June, 1935. Consider certain weather conditions prevailing over the south-western area of the chart :—

Wind.—In the south-west of Ireland, south-west of England and west of France, the wind is observed to have backed from south-west to south or south-east.

Temperature.—There is little or no evidence from surface temperatures to indicate the approach of a warm front. (The upper air temperatures observed at Henlow, Beds., however, show a rise of temperature at a height corresponding to 700 mb., from 20° F. at 1200 G.M.T. on the 17th of June, to 30° F. at 0600 G.M.T. on the 18th of June. Such a rise in temperature indicates the arrival of warm air in the upper layers and suggests the approach of a warm front or occlusion, *see* pages 348 and 350.)

Barometric tendency.—Over the same areas as mentioned under "wind" above, the barometer is observed to have a falling tendency, considerable falls being reported at stations 101, 158, and 211, while the falling tendency diminishes in a north-easterly direction.

Precipitation.—Rain is reported at all stations to the south-west of a line running approximately through stations 101, 102, 160 and 217.

Cloud.—Cirrus cloud appears in Northern Ireland, north-west England and the Midlands, while altostratus or altocumulus is observed at stations 149, 166, 143, 162, 160, 139 and 107. At stations 160, 157 and 158, fractonimbus cloud is reported. (It should be noted that in summer, over the land, clouds have a tendency to assume cumulus form owing to the increase in the lapse-rate. In particular, frontal altostratus cloud tends to be accompanied by altocumulus, so that this type of cloud, in such circumstances, may be regarded as signifying the approach of a front. Summer warm fronts, moreover, are often feeble and degenerate, and the altocumulus observed with them is really degenerate altostratus).

The above considerations indicate the approach of a warm front or occlusion towards the south-west coasts of the British Isles. It is impossible, however, to distinguish between these two types of discontinuity from observations made in advance of the front alone. Similarly, it is impossible from such observations to trace the exact location of the front ; this can only be accomplished by utilizing similar evidence at the front, and in its rear. The ship in lat. 51° N., long. 12° W. reports a south-south-westerly wind, temperature 57° F., and no precipitation in present weather. This would lead to the supposition that the front lies to the eastward

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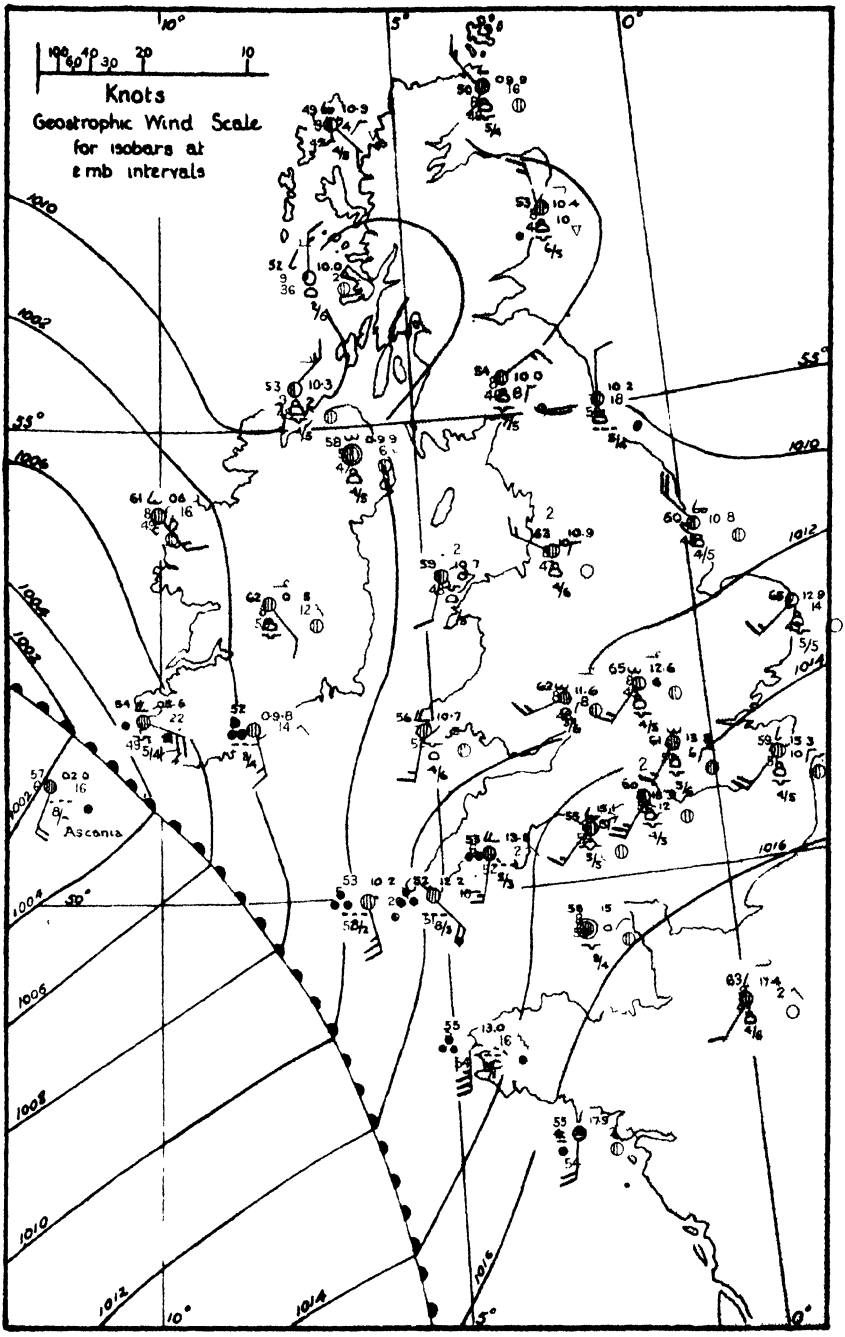

FIG. 138. APPROACH OF WARM FRONT. 1300 G.M.T.
17 JUNE 1935.

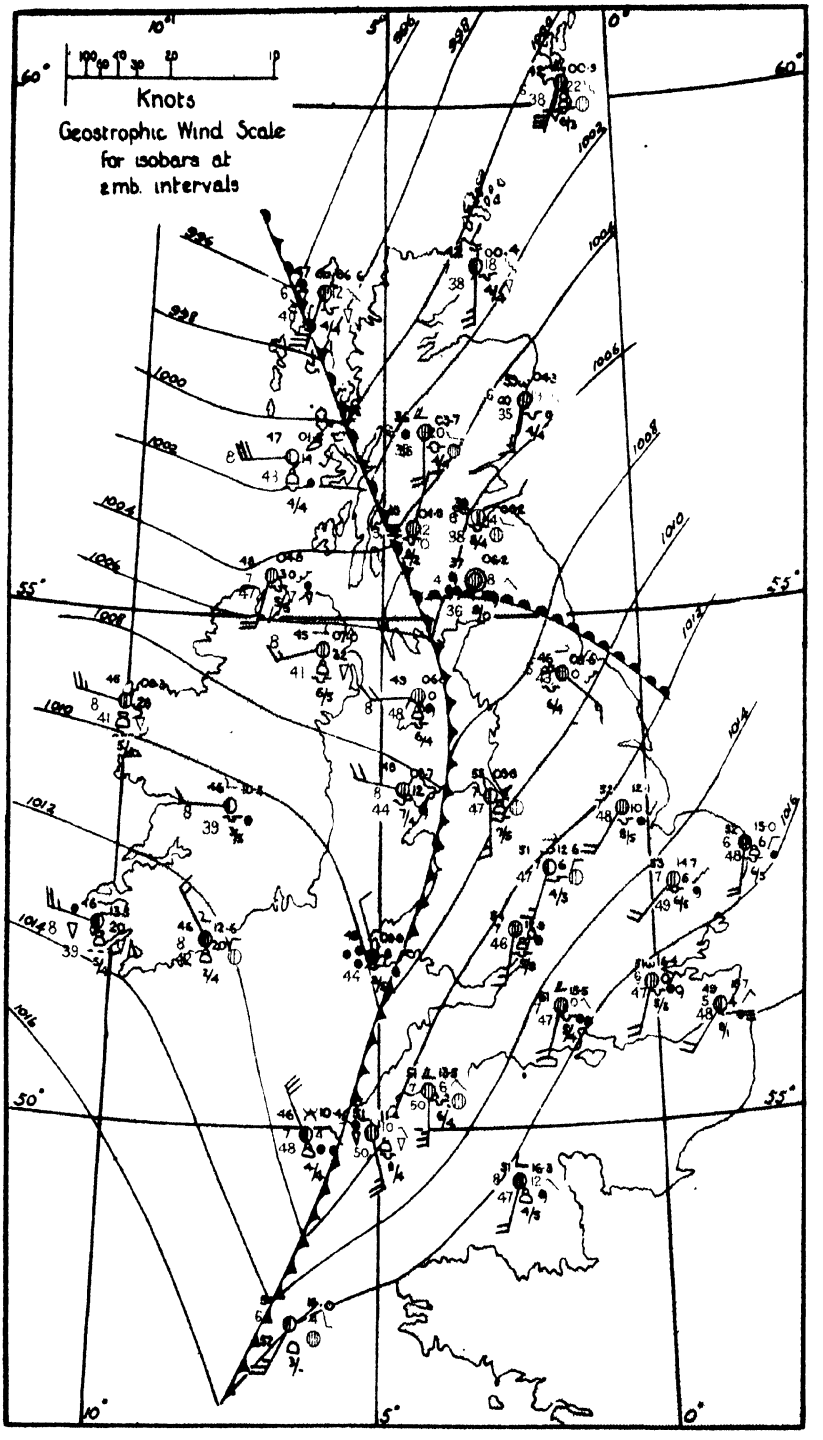

FIG. 139. COLD FRONT. 1300. G.M.T 29. DECEMBER 1936.

of the ship, but beyond this information, and the fact that the isobars appear to bend sharply in that area, the evidence of the existence of a warm front is inconclusive.

Recognition of a Cold Front.—Fig. 139 shows the synoptic situation over the British Isles at 1300 G.M.T. on the 29th December, 1936. A front extends from Stornoway to Land's End and exhibits properties of a cold front south of the Solway Firth. As before, consider the weather conditions in the vicinity of the front :—

Pressure and wind.—A trough of low pressure is shown along the line of the front with wind discontinuities from a southerly to a westerly or north-westerly direction. The trough is particularly sharp, and the wind changes pronounced over Wales and south-west England.

Temperature and dew-point.—There are considerable differences of temperature and dew-point across the southern part of the front, *e.g.*, at station 140, 54° and 46° F., at station 139, 48° and 44° F., at station 157, 51° and 50° F., and at station 158, 46° and 45° F. Over Scotland, however, the air in advance of the front has, at many stations, a lower dew-point and temperature than the air in its rear. The polar air over Scotland does, in fact, appear to be of a different type from the milder air over England, and the juxtaposition of these two air masses over southern Scotland is probably the cause of the drizzle at stations 118 and 120. Accordingly, a feeble warm front has been indicated over southern Scotland, and if this diagnosis is correct, the air over England assumes the role of a warm sector, while the portion of the front over western Scotland must be regarded as an occlusion. This would account for the negative barometric tendencies over Scotland, and the altostratus or nimbostratus cloud at stations 110, 114 and 111. The absence of general precipitation in advance of the occlusion is to be attributed to the dryness of the warm air.

Barometric tendency.—Over Scotland, the pressure tendencies are negative to the east of the front, while they are positive to the west. Over England, tendencies are mainly positive, although there are small negative tendencies in the south-west. In the rear of this portion of the front the positive tendencies are larger than the prefrontal tendencies, *e.g.*, + 2 at station 128, + 12 at 126, + 10 at 135 and + 28 at 104.

Precipitation and cloud.—The precipitation which occurs immediately in the rear of an ideal cold front is not well marked in this example, the only reports of rain being moderate rain at station 139, and rain in the previous hour at 158. Further to the rear of the front, showers are reported, *e.g.*, at stations 101, 104, 107 and 105, with large cumulus or cumulonimbus cloud at many stations. In advance of the front there are stratiform and cumuliform clouds, the latter indicating that the warm air is to some extent unstable. Now, having regard to the time of year, the instability cannot have

been brought about by insolational heating over the land ; therefore the air is not true tropical air, which in winter is always stable in temperate latitudes, but probably returning maritime polar air. This conclusion is supported by the difference between the temperature and dew-point at most stations ; if the air were true tropical air it would almost certainly be saturated. The comparatively low water-vapour content of the warm air may also account for the absence of general rain immediately in rear of the front.

From a study of preceding charts, the front is known to be moving eastwards, so that the above considerations lead to the recognition of the front as a cold front.

Recognition of the Approach of a Cold Front.—It is not possible as a rule to recognize an approaching cold front from observations in advance of the front. There is no well-defined pre-frontal cloud system, although cumulonimbus occurs at the front itself, and altocumulus (lenticularis) a short distance ahead of the front ; if the sky is clear of lower cloud, therefore, these clouds provide some indication of the approaching front. Often, however, they are obscured by the lower warm-sector cloud. The barometer often falls slightly with a backing of the wind before a cold front, and there may be a period of pre-frontal rain.

Recognition of an Occlusion.—Fig. 140 shows the synoptic situation over the British Isles at 1300 G.M.T. on the 2nd October, 1935. The frontal system is shown as an occlusion extending from south of station 113 to west of station 126, south of which it divides into a warm front passing between stations 160 and 157, and a cold front passing between stations 157 and 158. As before, consider the weather conditions at the fronts :—

Pressure and wind.—The frontal system coincides with a trough of low pressure on the east side of which winds are mainly from south or south-east, and on the west side, mainly westerly. The change of wind direction is abrupt, *e.g.*, between stations 126 and 139 between 160 and 157, and between 265 and 211.

Temperature and dew-point.—Although the temperatures and dew-points vary considerably from station to station, there is no systematic variation across the front as would be the case with a definite warm or cold front. For example, the values 53° F. and 46° F. for temperature and dew-point, respectively, at station 133, may be compared with 55° F. and 44° F. at station 103 ; 55° F. and 51° F. at 162, with 56° F. and 49° F. at 102.

Barometric tendency.—The barometric tendencies are negative over the whole area, and, while the greatest tendencies occur immediately in advance of the front (—50 at stations 160 and 263, —44 at 126), there are also large negative tendencies well in the rear of the front (—34 at station 101, —20 at 104). Only the small negative tendencies (—08 in each case) at 105 and 158 in comparison with the larger tendencies in advance of the front, tend to show

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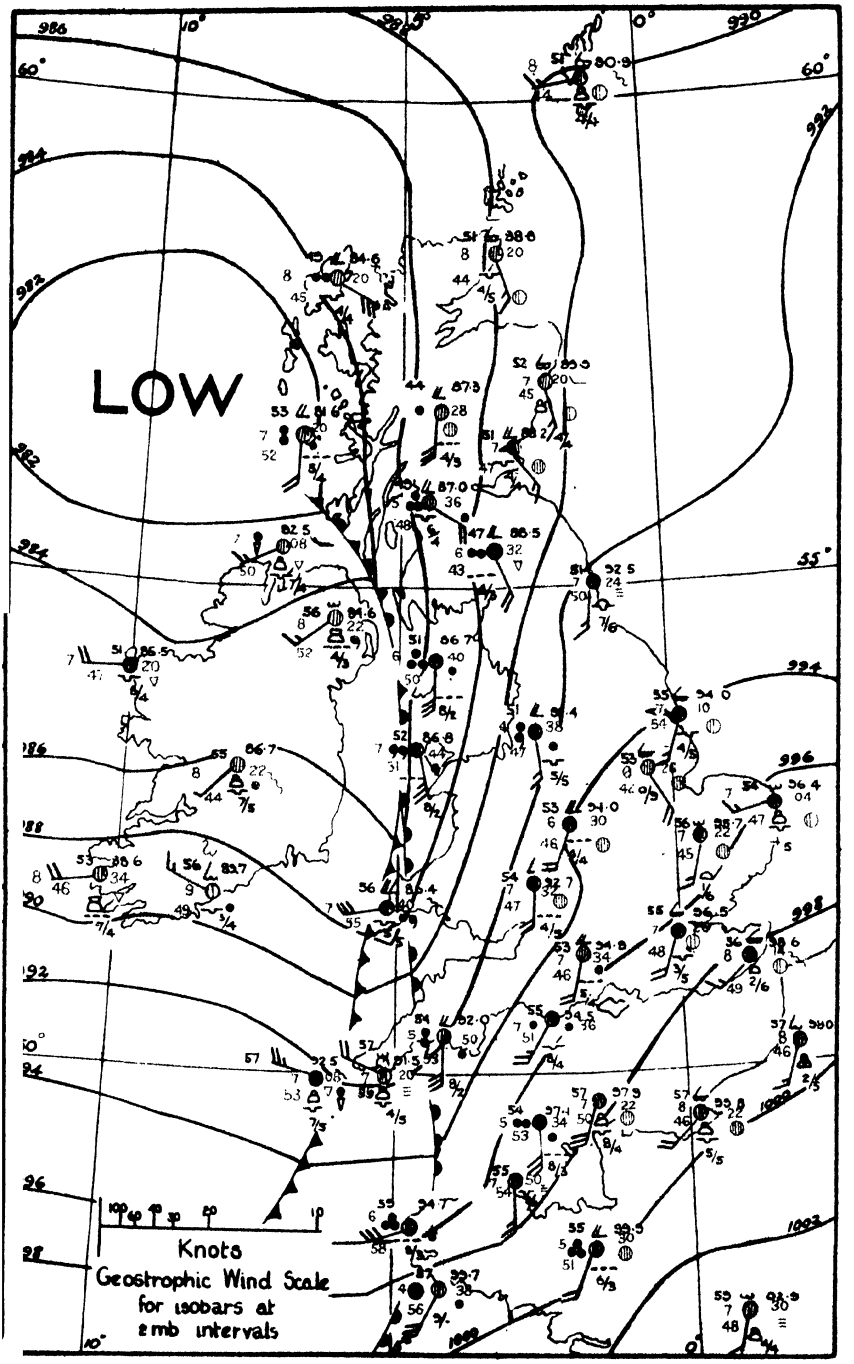

FIG. 140. OCCLUSION 1300 G.M.T. 2. OCTOBER 1935

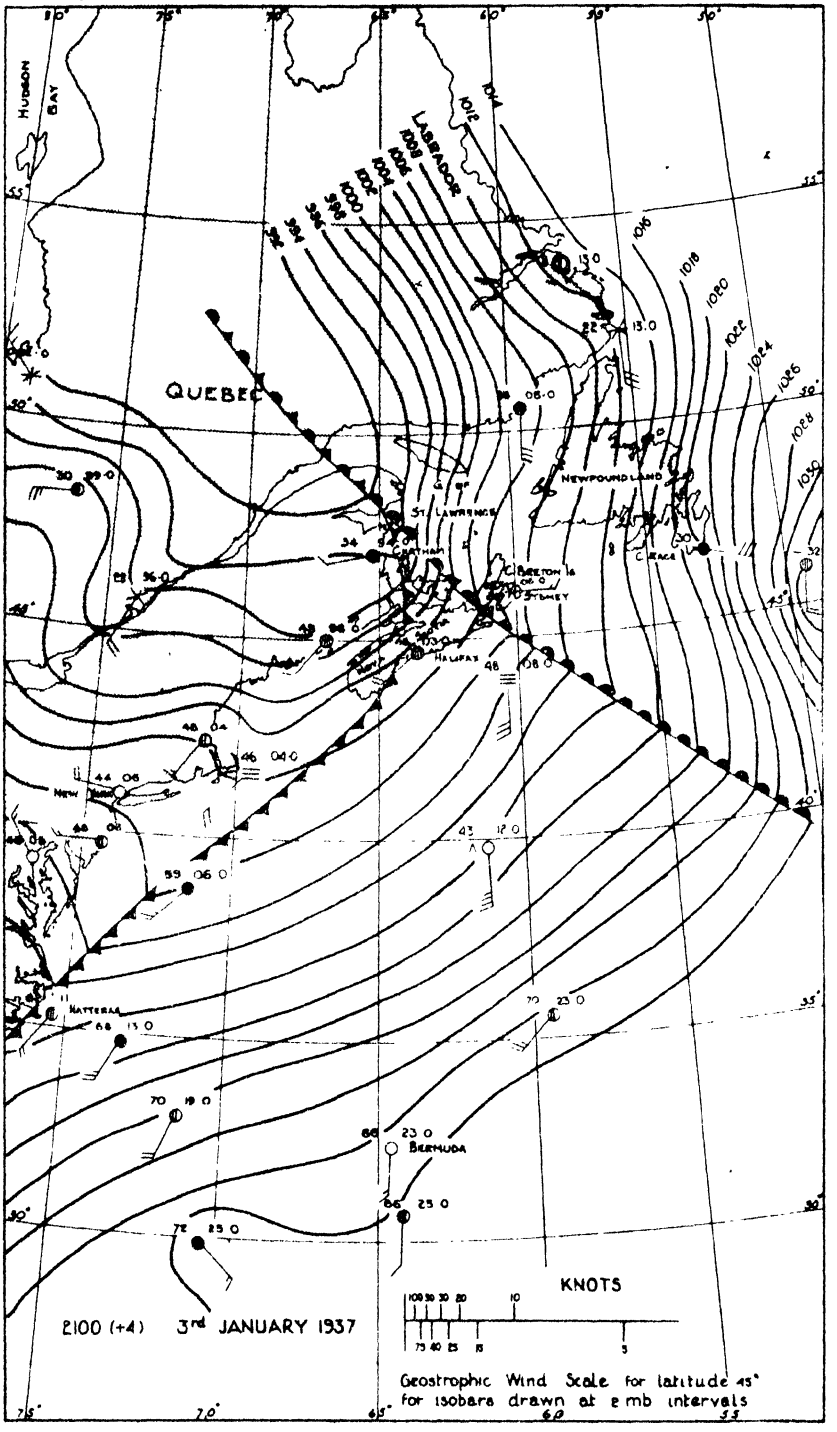

FIG. 141. NORTH AMERICA 0100 G.M.T. 4 JANUARY 1937.

that the front has passed these stations ; but the evidence supplied by the other factors is so definite in locating the front that the large negative tendencies in its rear must be ascribed to causes other than frontal. In this case they are due to the approach of a new depression from the Atlantic.

Cloud and precipitation.—There is a belt of precipitation about 100 miles wide in advance of the front, indicating either a warm front or occlusion. This is confirmed by the cloud, cirrostratus being reported at station 112, and altostratus or nimbostratus at many stations in England and Scotland ; near to the front there is much fractonimbus. In the rear of the front, however, over Ireland, large cumulus and cumulonimbus clouds are reported, indicating that the air in that region is polar in origin and therefore that the front is occluded.

A narrow warm sector is indicated extending southwards from the Irish Sea, the principal evidence for its existence being furnished by the high temperatures and dew-points reported from stations 139, 157 and 211. Moreover, at each of these stations the wind has veered to a westerly direction, showing that the warm front has passed them. The fall of dew-point between stations 157 and 158, together with the showers reported at 158, indicate that the cold front has passed that station.

Example of Frontal Analysis in Eastern North America.—Fig. 141 shows a specimen synoptic chart of this area for 0100 G.M.T. on the 4th January, 1937. Only the observations of wind, weather, pressure and temperature are available. A large deep depression is centred to the east of Hudson Bay, while an extensive anticyclone covers the western North Atlantic. A warm front extends in a south-easterly direction from the southern part of the Gulf of St. Lawrence while a cold front crosses Nova Scotia and runs approximately parallel to the eastern coast of the U.S.A. An occlusion is shown extending from the centre of the depression to the Gulf of St. Lawrence. Depressions with open warm sectors, and fronts with marked temperature discontinuities are characteristic of this region. These features are well shown in the example selected. In advance of the warm front, *i.e.*, in Labrador, Newfoundland and Cape Breton Island temperatures range from 8° to 38° F. Within the warm sector, at Halifax and the ship in lat. 44° N., long. 60° W., the temperature is 48° F., while farther south, temperatures of 60° F. or 70° F. occur. There is widespread precipitation in advance of the warm front, at some stations in the form of snow owing to the low temperature. The approach of the front is marked by a backing of the wind to east, *e.g.*, at Cape Race and Cape Breton Island ; within the warm sector the wind is mainly south or south-west. A further indication of tropical air is furnished by the fog reported by the ship in lat. 44° N., long. 60° W. The cold front is located chiefly by temperature discontinuities. At American coastal stations temperatures are generally below 50° F. ;

at Chatham the temperature has fallen to 34° F. In the neighbourhood of New York, the wind has veered to west ; further evidence in support of the cold front is furnished by the rain at Halifax and at the ships in the vicinity of New York and Cape Hatteras. There is insufficient evidence in the vicinity of the occlusion to locate its position ; it has therefore been drawn from the centre of the depression to the meeting point of the warm and cold fronts in the Gulf of St. Lawrence.

Example of Frontal Analysis in the Southern Hemisphere.— Fig. 142 shows the synoptic situation in the Australia-New Zealand area at 2300 G.M.T. on the 24th July, 1936. A depression is situated over the eastern part of the Great Australian Bight, and from the centre a warm front extends eastwards and a cold front north-north-east. Only the observations of wind, weather, pressure and temperature are available in this example. In advance of the warm front, which is moving clockwise with respect to the centre, winds are mainly easterly, and temperatures range between 47° and 53° F. Precipitation in advance of the warm front is scanty ; this, however, is usual in Australia since the warm air is generally of continental origin and, therefore, dry. At the warm front the wind backs sharply to a northerly direction, and in the warm sector temperatures range between 51° and 66° F. The continuous drizzle at station 298 is typical of warm sector conditions, while the rain showers at stations 271 and 283 are to be attributed to convection over the heated land. At the cold front there is a further sharp backing of the wind to west or west-north-west, with a general fall of temperature to about 51° F., while a rain shower is reported from station 140. Frontal analysis in Australia is difficult on account of the scantiness of the information supplied, owing to the dryness of the warm air and consequent general absence of precipitation at a front, and also owing to the situation of the area on the edge of the sub-tropical high-pressure belt, with consequent modification of the original properties of the air masses by subsidence.

Practical Rules for the Movement of Fronts.

(Applicable to both hemispheres.)

1. A cold front advances at about the instantaneous speed of the gradient wind in its rear resolved at right angles to the front.

It has been shown that normally the cold front of a depression moves faster than the warm front, so that it eventually overtakes the warm front. The advance of the cold front is not seriously retarded by friction at the earth's surface, although the lower layers of cold air in immediate contact with it may be retarded more than the cold air at higher levels, so that the cold front overhangs in the lowest layers. At a height of about 2,000 feet, however, the cold air is not affected by friction, and moves forward so as to

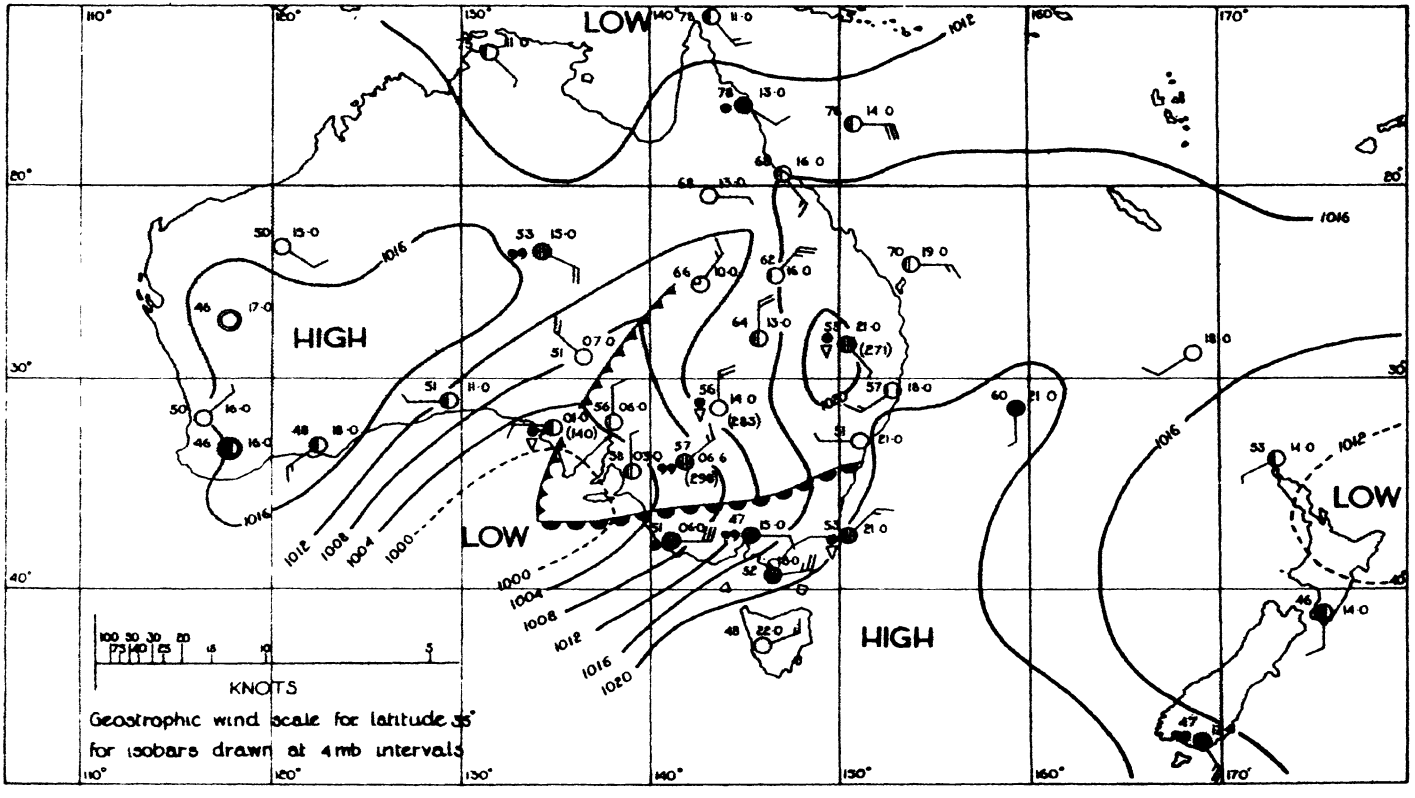

FIG 142 AUSTRALIA 2300 G.M.T. 24 JULY 1936.
(Note, The index numbers referred to in text are inserted in brackets e.g.(298).

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overhang the warm air in the warm sector, thus producing an unstable arrangement which breaks down from time to time and gives rise to squally winds (see Fig. 96, page 286). After such a breakdown the overhanging front begins to build up again, and the process may be repeated several times without appreciably retarding the advance of the cold front. When a cold front encounters a mountain barrier the warm air which is trapped between the mountain and the oncoming cold wedge must be forced upwards, so that large quantities of rain may be added in this way to the pure orographic and cold front rain.

2. A warm front advances at a speed which is less than the instantaneous speed of the gradient wind in its rear resolved at right angles to the front. The amount of this difference varies; on the average it is 10 knots, but it may be as much as 20 knots depending on the inclination of the isobars to the front, and hence the direction of the wind in advance of the front.

In the case of a warm front, the surface layers of air both in front of and behind the warm front may be retarded by surface friction, but at higher levels surface friction can have no appreciable effect and so the front at 2,000 feet should advance freely as far as surface friction is concerned. The only way in which the retardation of the warm front can be explained is by the asumption that the cold air has an acceleration down the slope of the front, sufficient to produce the difference between the rate of advance observed and the gradient wind.

3. A front may move forwards as a warm front, or backwards as a cold front, depending on whether the observing station experiences a transition from cold air to warm air, or *vice versa*. This is particularly applicable when a front has a westerly to easterly orientation, and in the case of small waves passing along the surface of discontinuity, the alternation between the roles of warm and cold front may be quite rapid.

4. If a front lies along the isobars, it will remain more or less stationary, for there is then little or no gradient in the rear from which it can derive a progressive movement.

5. When fronts form along the isobars, the stronger winds are on the side of higher pressure, *i.e.*, where the isobars are closest together.

6. In summer a warm front may dissolve on passing from sea to land, while a cold front has a tendency to accentuate under the same conditions.

7. In anticyclonic regions, fronts tend to become stationary.

8. Cold fronts tend to become diffuse and to dissipate when they are accelerating, and sharp when they are retarding, while the reverse is the case with warm fronts.

9. A warm front moves the faster the greater the negative barometric tendencies ahead of it; a cold front moves the faster, the greater the positive tendencies in its rear.

10. A cold front approaching a mountain range intensifies. In the rear of a cold front, and immediately to leeward of a mountain range, intense squalls, frequently associated with heavy hail showers, may occur. The wind speed reached in the squalls bears no relation to the pressure gradient.

11. Mountain ranges, in general, slow down all fronts; the retarding effect is greater the higher the barrier and the lower the height of the frontal surface.

In general, the less the impulse giving rise to a frontal disturbance, *i.e.*, the shorter the length of the wave formation and the less the speed of the gradient wind perpendicular to the front, the less is the possibility of occlusion. It is the short waves and those of small amplitude that remain as waves on the polar front, but their passage, although sometimes indicated only by a sinuosity in the isobars, and not by a closed net of isobars, may be attended by considerable rainfall and strong shifting winds. Consequently, their detection and the consideration of their travel is of some importance on that account. The rapidity of movement of such small frontal waves, however, is extremely variable.

The same practical rules given above may be applied to the movement of occlusions, depending on whether the occlusion belongs to the cold-front or the warm-front type, although in considering the speed of advance of an occlusion of the former type, it is probably more accurate to allow for a value somewhat less than that of the gradient wind behind it.

In the absence of more convincing evidence to the contrary, it is well to assume that the immediate future movement of fronts and occlusions, will be a direct continuation of that in the recent past, both as regards direction and speed.

Rules 1 and 2 above may be applied to locate the future position of a front at the end of a period of a few hours' duration. The front having been located on the current chart, the geostrophic wind scale is applied along the front (not necessarily perpendicular to the isobars); the reading will then give the value of the component of the geostrophic wind perpendicular to the front. The distance through which the front will move, perpendicular to itself, in a given time can therefore be calculated and the future position of the front estimated. It should be noted that this assumes that the geostrophic wind in the rear of the front does not change during the period. The rule cannot, therefore, be applied to the portion of a front near to the centre of a depression since, there, the pressure distribution is usually changing rapidly; it is best to assume that that portion of the front moves with the centre of the depression. It is sometimes possible by this method to locate a front on the current chart, when, from the observations available, its position

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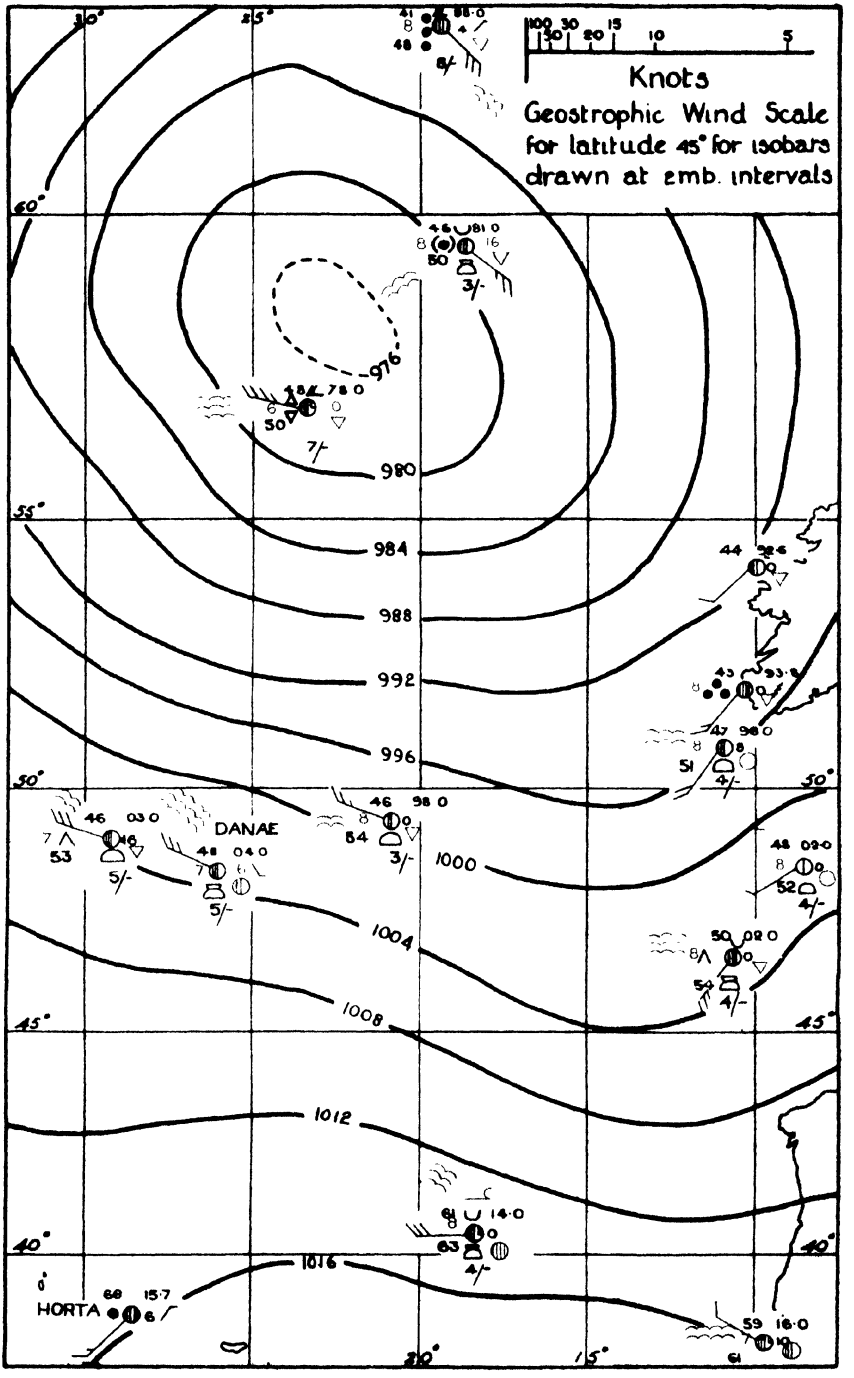

FIG. 143. 0100. G.M.T. 16. NOVEMBER 1935

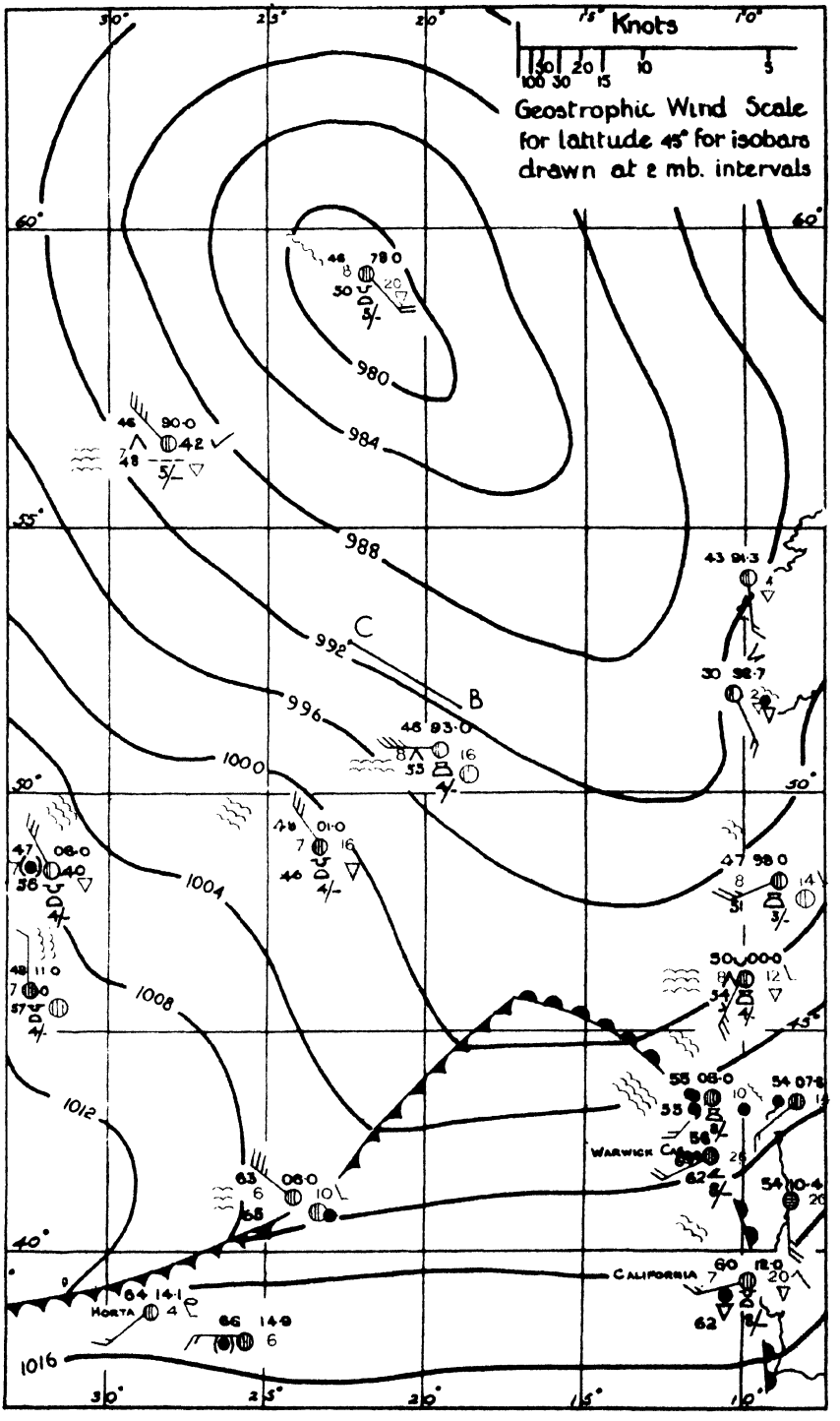

FIG. 144 0700. G.M.T. 16 NOVEMBER 1935.

is ill-defined; it is located by laying off its movement from its position as determined from preceding charts.

Development of a Depression between Two Westerly Air Currents.—Fig. 143 shows the synoptic situation in the eastern North Atlantic Ocean at 0100 G.M.T. on the 16th November, 1935; a depression is centred some 300 miles south of Iceland.

It is not possible, from the information available, to trace the outline of the polar front, but a comparison of temperatures and wind directions between the ships in mid-Atlantic and Horta suggests the co-existence of two eastbound air masses of widely different characteristics. This is confirmed by tracing back the source of these air masses on earlier charts. To the north of the boundary between them the air is of polar origin; to the south the air is of tropical or sub-tropical origin. Furthermore, it will be observed that there is a decided slackening of the pressure gradient between H.M.S. *Danae* and Horta, while altocumulus and cirrostratus cloud are reported by the ship in lat. 40½° N., long. 18°W. These factors indicate the possibility of a new development in the area between H.M.S. *Danae* and Horta.

The synoptic chart for 0700 G.M.T. on the same day (Fig. 144), shows a marked bulge in the isobars in the area west of the Bay of Biscay. Falling pressure with wind backing towards the south is reported by the ship in lat. 46° N., long. 10° W. and station 362, together with rain at a number of ships and shore stations in that area. *California* and *Warwick Castle* report a veer of wind to west-south-west, while the temperature at *California* has risen to 60° F., compared with 54° F. at stations 381 and 362. This information suggests a warm front in the position indicated in the figure, and this assumption is confirmed by *Warwick Castle*, which reports temperature 58° F., ten-tenths of nimbostratus cloud, and continuous slight drizzle, *i.e.*, warm-sector weather. Ships in mid-Atlantic, generally, report westerly to north-westerly winds, cumulus and cumulonimbus cloud, and temperatures between 46° and 48° F., which are, in the main, several degrees lower than the sea-temperature—indications that these ships are all in polar air and, therefore, that somewhere to the south of them there is a cold front separating that polar air from the air in the warm sector. There is not sufficient evidence to locate the position of the cold front exactly, but the ship in lat. 42° N., long. 24° W., reports a veer of wind to north-west with rain in the previous hour; these considerations, compared with the south-westerly wind and absence of precipitation at Horta, suggest that the front has passed the ship, but not Horta. The comparatively high temperature of 63° F. reported by the ship is, however, difficult to explain; possibly it is due to subsidence at the cold front.

Fig. 145 shows the synoptic situation at 1300 G.M.T. on the 16th November. The configuration of the isobars to the west of

the Bay of Biscay and off western Spain, indicates at least, the presence of an elongated trough of low pressure extending southwards from the west of Ireland. Furthermore, the west-south-westerly wind reported by *Rangitane* suggests that there is a closed centre to the south of that ship, off north-west Spain, although definite evidence on this point is lacking. The large tendencies reported by *Adda* and *California* (in the warm sector), indicate a rapidly deepening disturbance, a fact which is confirmed by the tendencies at stations 364 and 211. The rain reported by *Warwick Castle* and station 364, and the rise of temperature between stations in Spain and the three ships off the coast, indicate a warm front in the position shown. The veer of wind to north, fall of temperature to 57° F., and the rising pressure tendency reported by the ship in lat. 42° N., long. 21° W., indicate that the cold front has passed that ship ; the position of the cold front is not definitely established by the information available, but may be approximately located by laying off its advance from the position on the 0700 chart, calculated on the basis of the measured geostrophic wind of 40 knots.

The synoptic chart drawn for 1800 G.M.T. on the 16th November (Fig. 146) shows the secondary now situated over the Bay of Biscay. There are still, however, no observations off the south-west coast of England to indicate a separate wind circulation. Southerly winds, large negative tendencies and widespread rain reported by stations in western France and coastwise shipping in the Bay, indicate the approach of a warm front or occlusion. The veer of wind and rise of temperature between stations 361 and 351, indicate to some extent that station 351 is in the warm sector, and the warm front in the position shown. The cold front has advanced to western Spain, as is revealed by the generally rising tendencies at stations 362 and 381, and a number of ships off western and north-western Spain ; moreover, most of these ships report north-westerly winds and temperatures some 4 degrees lower than that of the sea, while H.M.S. *Danae* and the ship in lat. 42° N., long. 14° W. report cumulonimbus cloud. Since 1300 the centre of the secondary depression has moved east-north-east, corresponding to the direction of the isobars in the warm sector.

The synoptic situation at 0100 G.M.T. on the 17th November (Fig. 147) shows the secondary depression now situated over the Channel Islands. It will be seen that the disturbance has now developed a complete wind circulation ; pressure at the centre has fallen to 981·7 mb. indicating considerable deepening since 1800. Information is scanty but there is no evidence of a warm sector in the central part of the depression ; the winds, barometric tendencies, temperatures and rainfall over southern England and northern France indicate an occlusion in the position shown. The warm sector over southern France has been drawn entirely on the basis of estimated frontal movement (see rules) ; the wide belt of persistent rain to the north-east of the centre should be noted.

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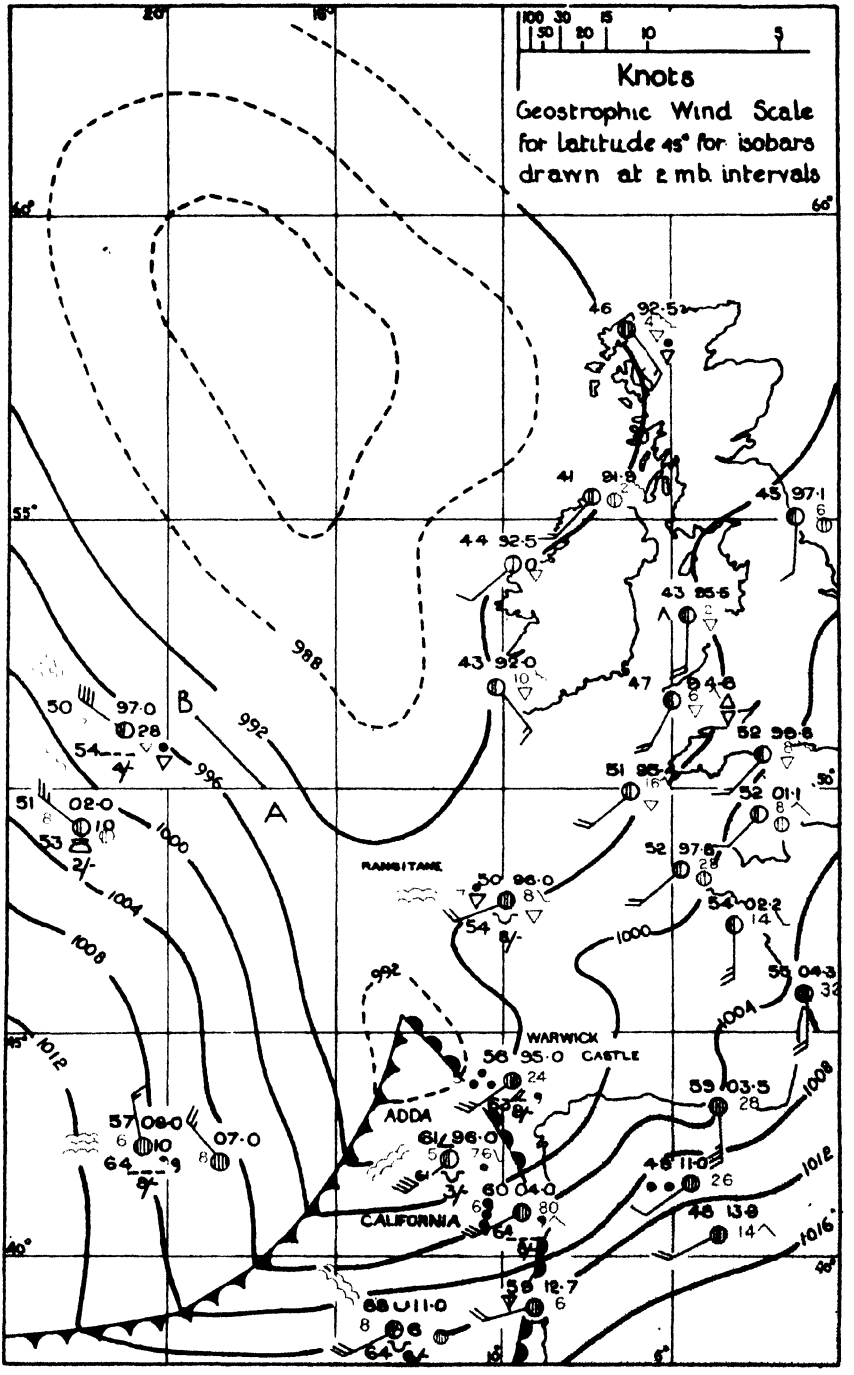

FIG. 145. 1300. G.M.T. 16. NOVEMBER 1935.

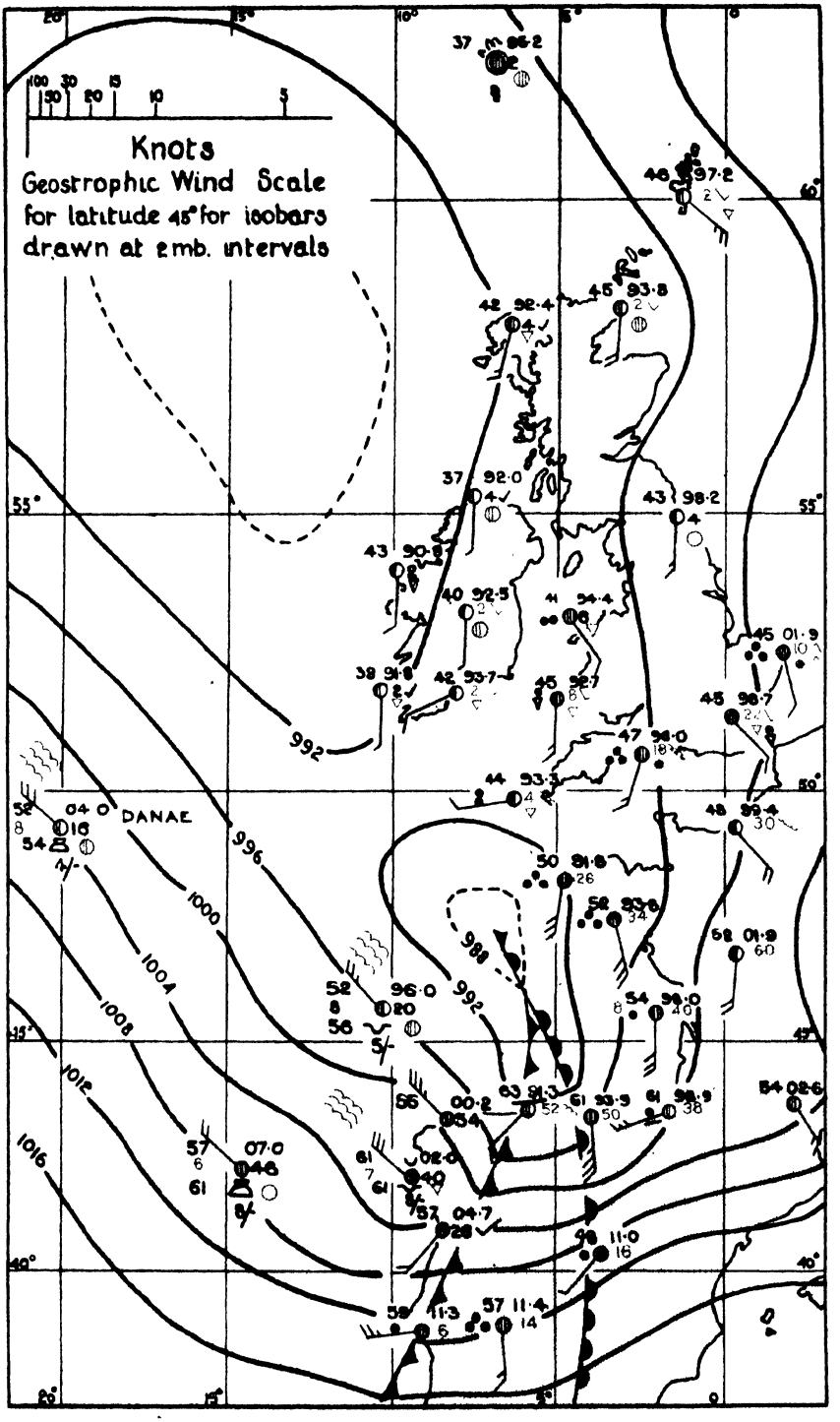

FIG. 146. 1800. G.M.T. 16. NOVEMBER 1935.

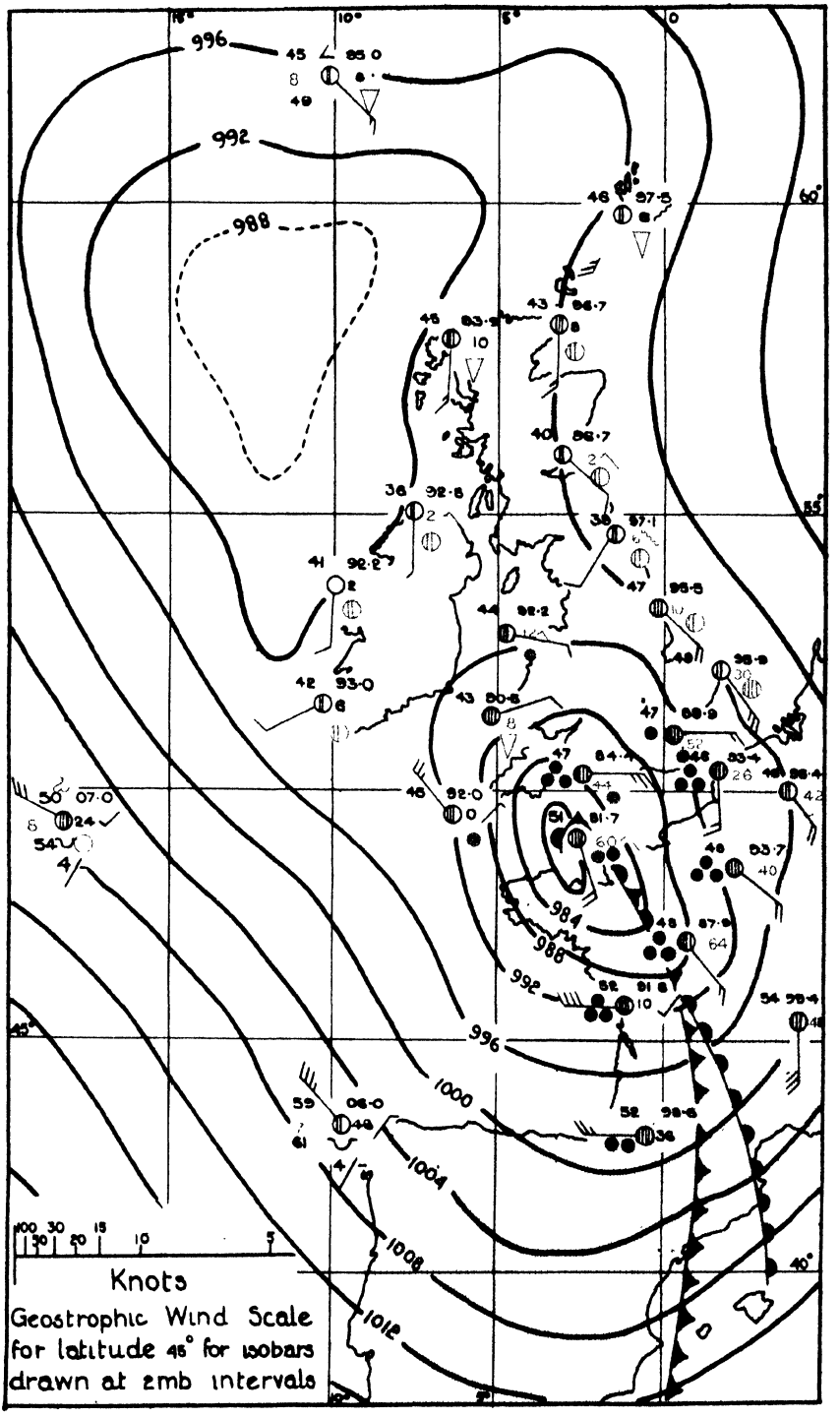

FIG. 147. 0100. G.M.T. 17 NOVEMBER 1935.

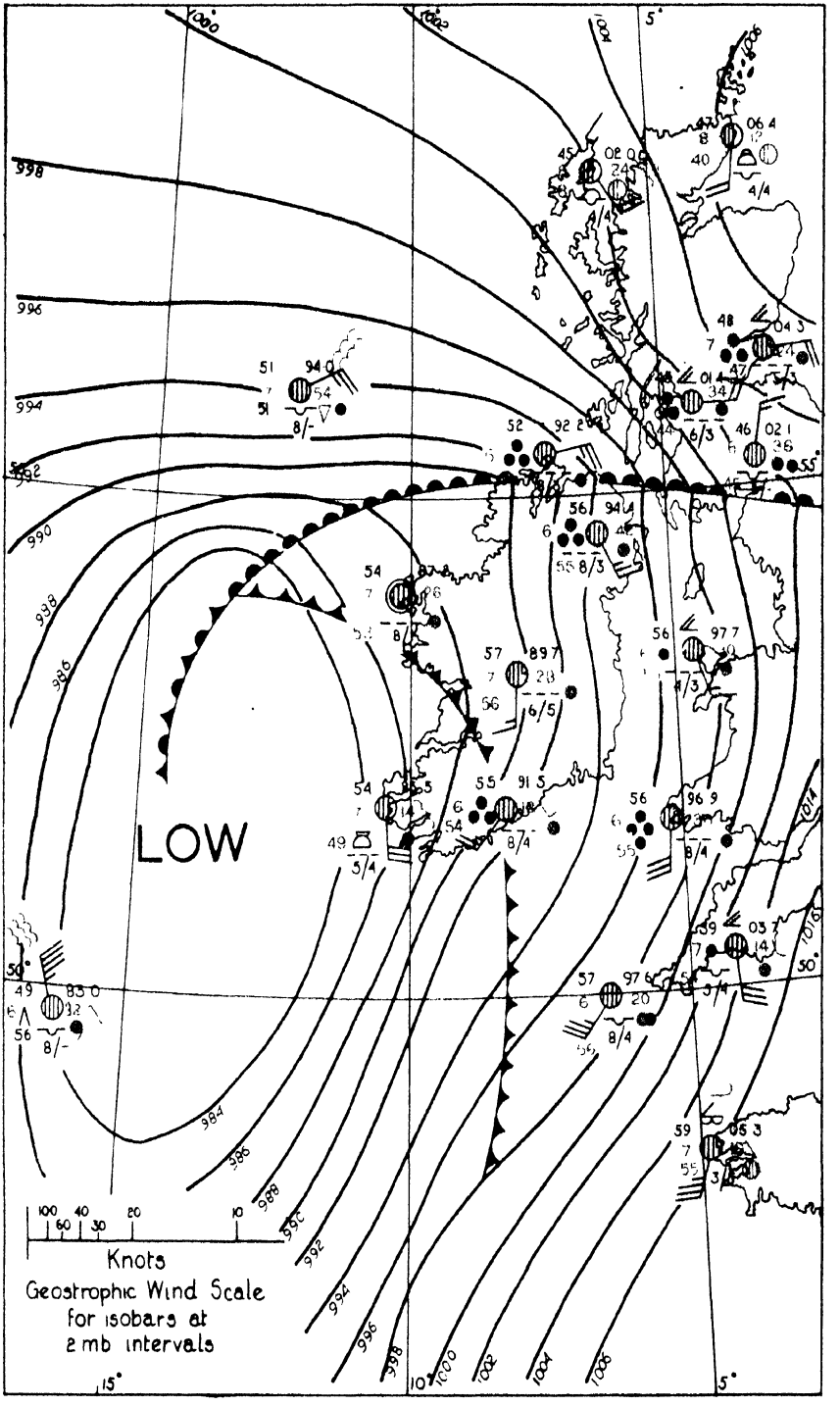

FIG. 148. 0700 G M T 25 OCTOBER 1934.

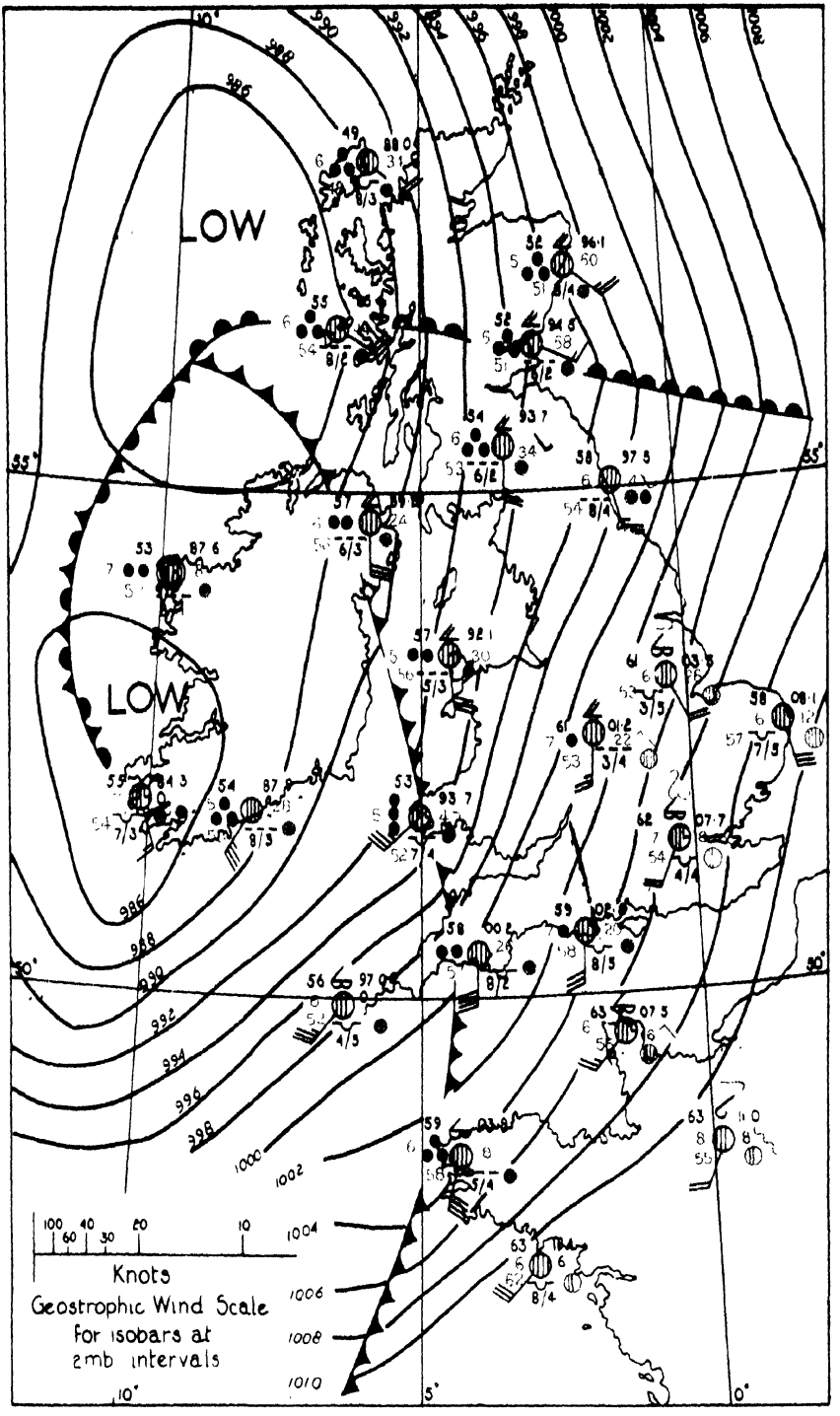

FIG 149 1300 G.M.T 25 OCTOBER 1934

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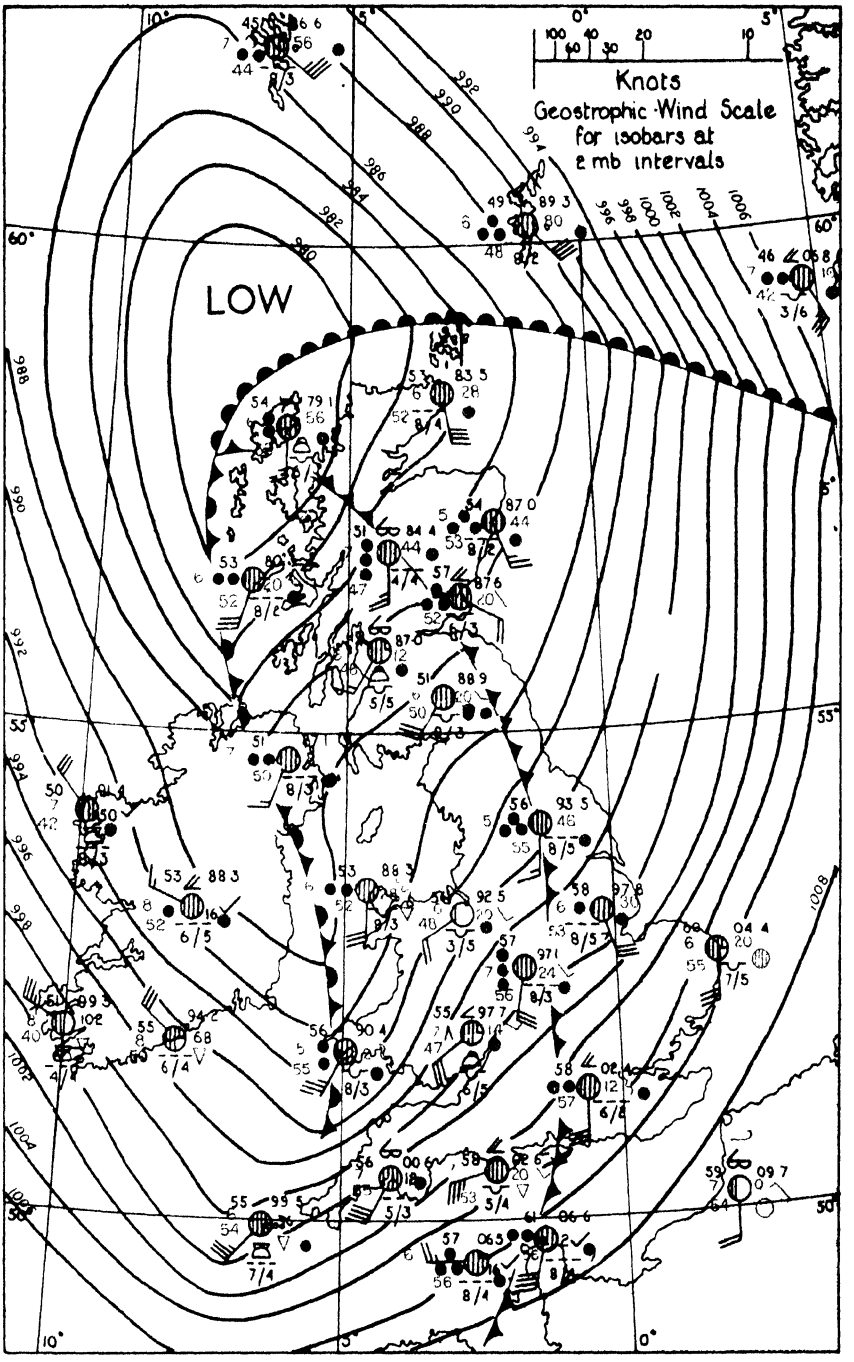

FIG. 150 1800 G M.T. 25. OCTOBER 1934.

Formation of a Back-Bent Occlusion.—Fig. 148 shows the synoptic chart for the British Isles at 0700 G.M.T. on 25th October, 1934. A deep depression is situated to the west of Ireland and from its centre a warm front extends across northern Ireland and south Scotland while a cold front extends south-eastwards across south-west Ireland. The position of the warm front is clearly defined by the winds, which in advance of it are mainly easterly and in its rear, south-east or south. In this case, too, the temperature and dew-point rise sharply on crossing the front from north to south. The cloud type and height, and the rainfall to the north of the front confirm the diagnosis, but the indications of the barometric tendencies are not definite, the barometer falling somewhat rapidly on both sides of the front. The warm sector, covering most of England is well defined by the high temperatures and dew-points. A number of stations report rain, which is to be expected in the warm sector of a depression which the barometric tendencies show to be deepening, while the clouds are also of the warm-sector type. At station 101, however, there is a partial clearance of the sky, the cloud is Cunb., the dew-point has fallen to 49° F. with relative humidity 85 per cent., indicating the arrival of polar air. The cold front has accordingly been placed east of this station, practically through 102, where continuous rain is falling. The continued though diminished fall of the barometer at 101 indicates either the approach of a secondary cold front or a back-bent occlusion, or it may be attributed solely to the deepening of the depression; at the moment it is impossible to say which.

The synoptic situation at 1300 G.M.T. is shown in Fig. 149. The warm front has moved slightly north, its position still being defined chiefly by the rise of temperature and dew-point and the veer of the wind across it. The rise of temperature across the front is less marked than at 0700, indicating that the front has become rather diffuse. This frequently happens, when, as in the present case, the front is nearly perpendicular to the isobars, and it is not possible then to draw the position of the front with certainty. The position of the cold front is immediately suggested by the rain along the line of the stations 107, 126, 139, 160 and 211, and is confirmed by the veer of wind to SW., the partial clearance of the sky and the fall of dew-point to 52° F. at station 158. While there is still no definite evidence of the bent-back occlusion, the precipitation, low stratiform cloud and falling barometer at the Irish stations indicate its possibility.

By 1800 G.M.T. (Fig. 150) the back-bent occlusion has appeared with well defined cold-front characteristics and extends from the centre of the depression, now near the Hebrides in a southerly direction as far as Pembroke (139). The features defining it are:—

The veer of wind from S or SW. at 113, 107, 126 and 139 to NW. at 101, 102, 103 and 104;
The rising barometric tendencies at the four latter stations;
The low dew-points and the Cunb. clouds at 101 and 104;
The rain along the line of stations 113, 107, 126 and 139.

The centre of the depression remains at the tip of the warm sector. The warm front obviously lies south of stations 110 and 028, and north of 111 and 112. It is difficult to locate the cold front exactly, the assumed position being based on the following considerations :—

The veer of wind between stations 117 and 120, 125 and 128, 149 and 162, 205 and 168 ;
The precipitation along the line 117, 125, 133, 149, 205 ;
The low dew-points at 128 and 140 ;
The showers at 126 and 162.

Development of a Depression as a Wave on a Cold Front.—Fig. 151 shows the synoptic situation over the British Isles at 1800 G.M.T. on the 30th October, 1936. A cold front extends across central England from east-north-east to west-south-west ; the front has been located by the observations of falling tendencies, high temperatures and mainly south-westerly winds on its southern side ; rising tendencies, northerly to north-westerly winds, lower temperatures and a belt of rain about 100 miles wide on its northern side. There is no reason to anticipate a wave development, except possibly the opening out of the isobars over central England and the North Sea.

The synoptic situation at 0100 on the 31st October is shown in Fig. 152. By this time the cold front has advanced in a south-south-easterly direction, and is located as before by barometric tendencies, winds and temperatures. The shape of the isobar for 1,016 mb. suggests that a secondary is in process of development over the Bristol Channel. This is confirmed by the fact that the negative tendencies over southern England are much greater than at 1800 on the 30th October, while the positive tendencies immediately in the rear of the front have almost disappeared.

The synoptic chart drawn for 0700 on the 31st October (Fig. 153) shows that a shallow secondary depression with a complete cyclonic circulation has developed over the Isle of Wight. The front is located chiefly by wind discontinuities, *e.g.*, a westerly wind at station 166, south-westerly at 230, south-westerly at 291, north-westerly at 137, northerly at 151 and east-north-easterly at 149 ; west-north-westerly at 168, northerly at 211 and south-westerly at 217 and 218. The temperature differences across the front noted on previous charts still persist, *e.g.*, 45° F. at station 140, 46° F. at 135, 47° F. at 160, 53° F. at 166, 55° F. at 230 and 57° F. at 217. The temperatures of 53° F. at station 151 and 52° F. at 149 and 168, all of which stations are immediately behind the front, are attributed to mixing at the front. The front has assumed the form of a wave at the centre of development. It would appear from the rain over the Midlands and from the run of the isobars to the south-west of London that the portion of the front between Portsmouth and the Thames estuary has become a warm front. The next subsequent chart (Fig. 154) shows, however, that the whole

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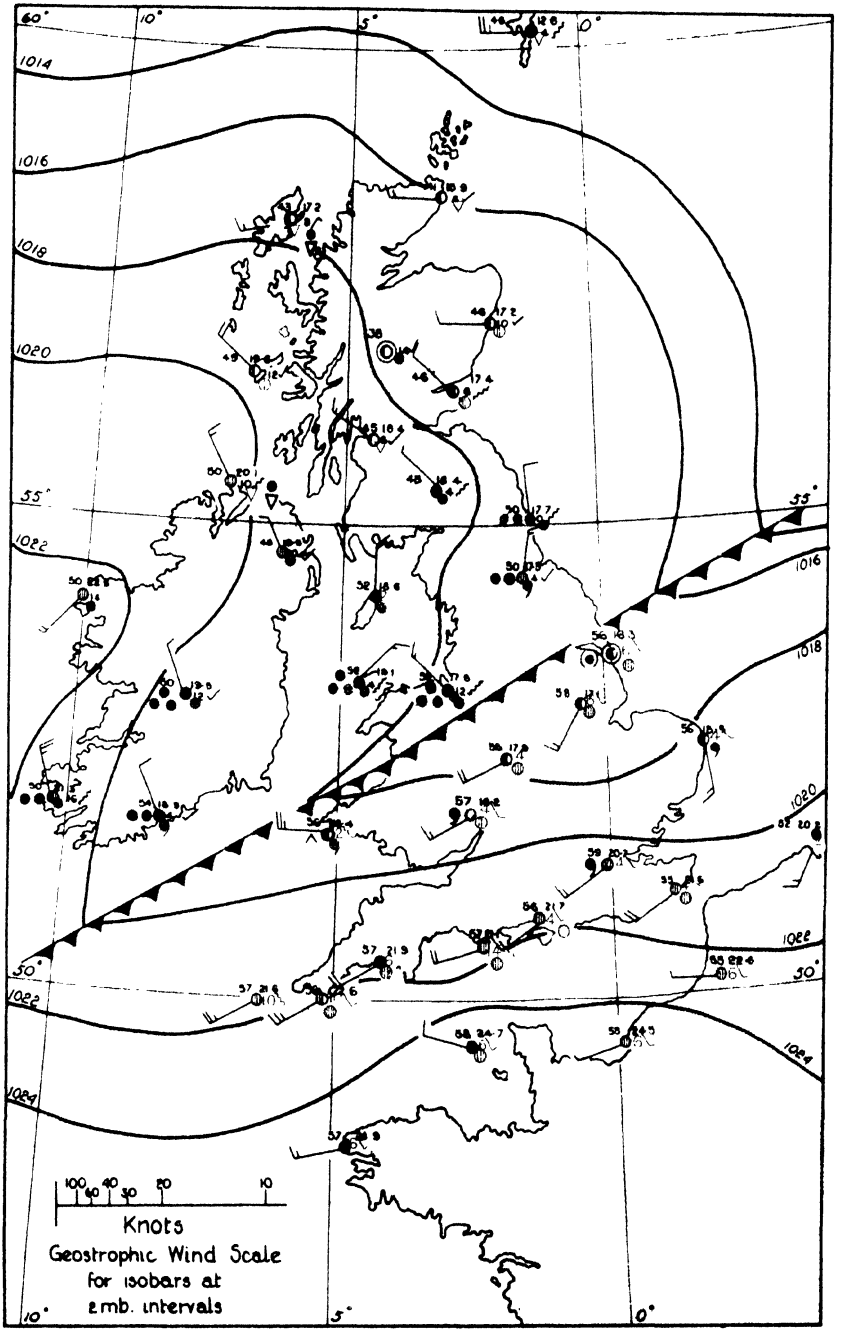

FIG. 151. 1800. G.M.T. 30 OCTOBER 1936

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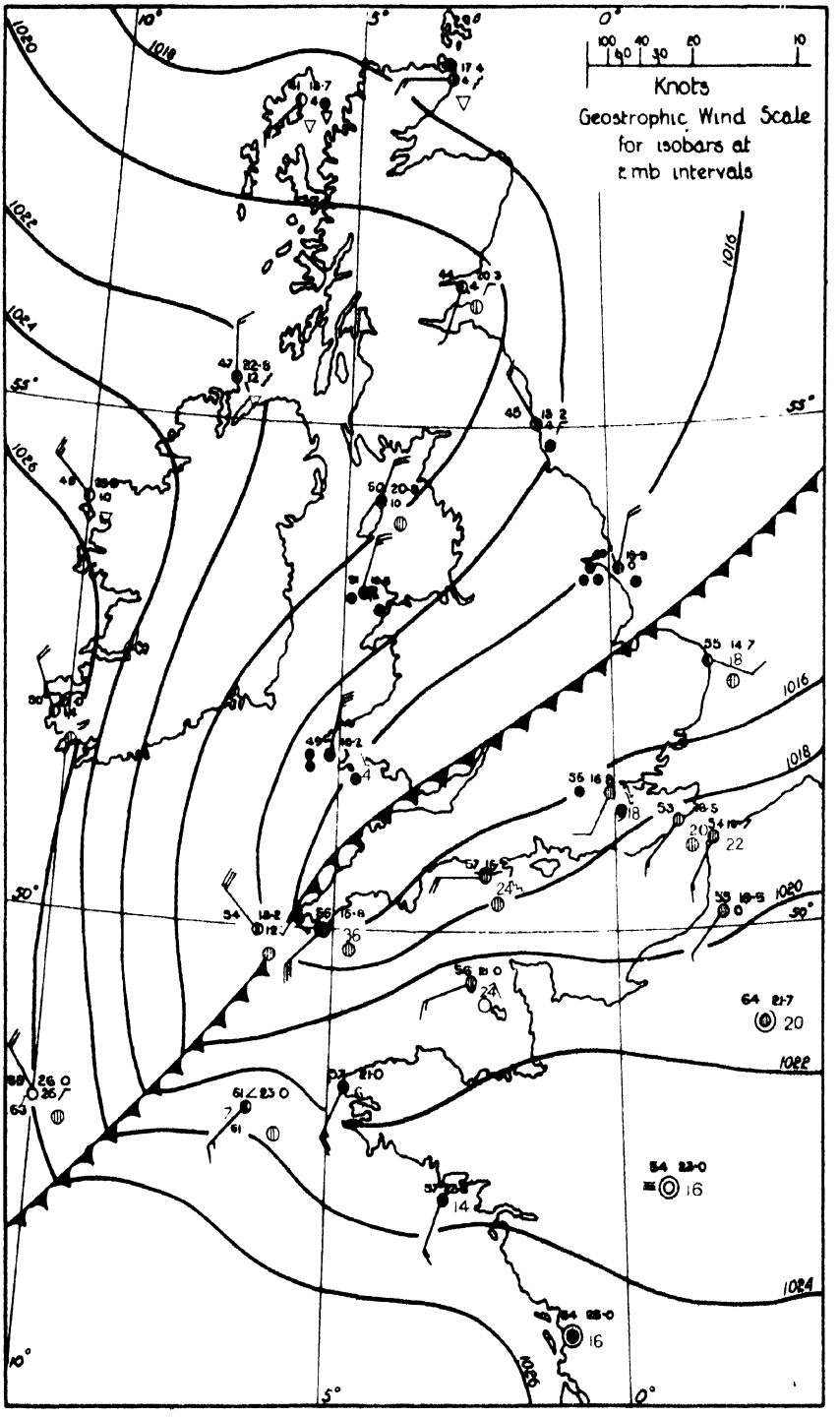

FIG 152 0100. G M T 31 OCTOBER 1936.

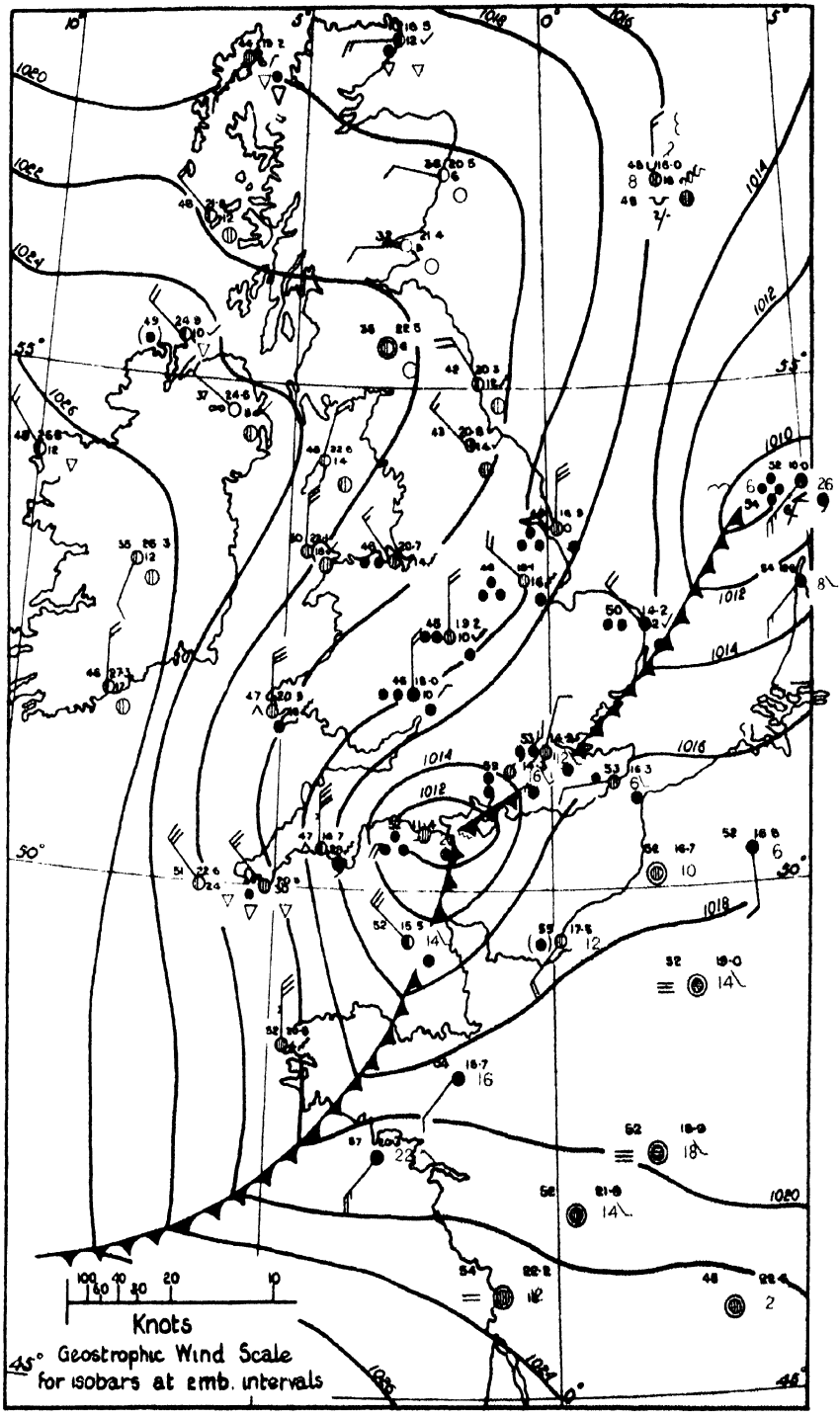

FIG. 153. 0700. G.M.T. 31 OCTOBER 1936.

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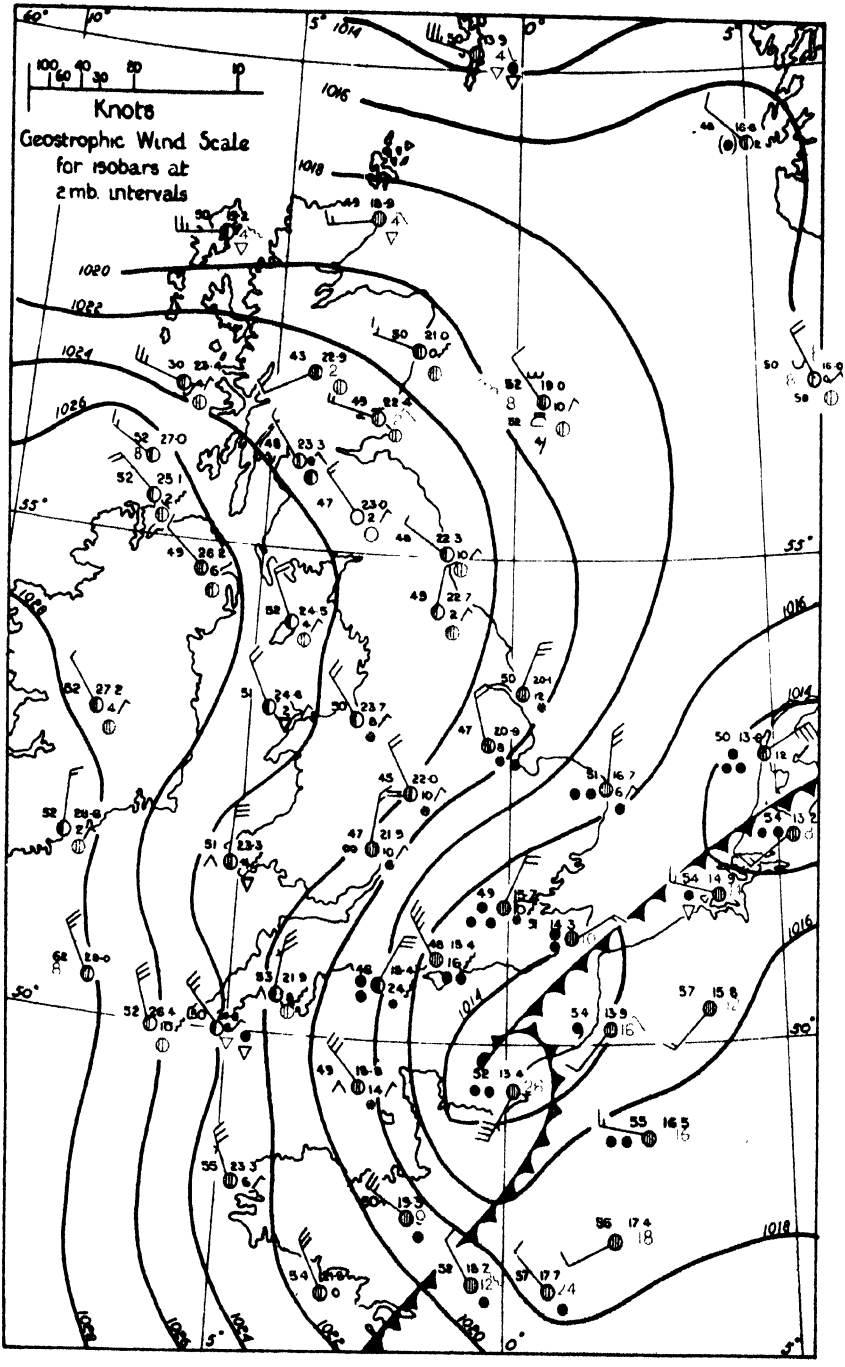

FIG. 154. 1300 G.M.T. 31 OCTOBER 1936.

front continued to move south-east and it has therefore been drawn as a cold front. The other portion of the front which extends from the centre of the depression to station 217 remains a cold front.

Fig. 154 shows the fully developed depression at 1300 on the 31st October, centred over the eastern channel. The portion of the front extending north-eastwards from the centre is again located by wind and temperature discontinuities, while the remaining portion is shown extending south-westwards near to station 230, separating negative tendencies on its south-easterly side from positive tendencies on the opposite side. The wind discontinuities are not well marked as regards direction, winds to the south-east of the front being light and somewhat variable in direction, those on the opposite side mainly strong northerly.

(C) Air Masses and the use of Upper-Air Temperature Observations in Forecasting.

The general properties of air masses are described in Chapter XX, and a consideration of these properties will, as a rule, enable an air mass to be identified. In many cases also it is possible to deduce the nature of an air mass from the wind direction, *e.g.*, in the British Isles tropical air with south-westerly winds, maritime polar air with west-north-westerly winds, continental polar air with easterly winds. Sometimes, however, the direction of the wind at a particular station may be misleading as regards the air mass; *e.g.*, maritime polar air circulating round a depression immediately to the south of Ireland may arrive over southern England as a southerly current. Similarly, with an anticyclone centred to the west of Ireland, tropical air moving northwards on its western boundary, circulates clockwise round the centre and arrives over the British Isles as a warm damp current from the north-west. In such cases the properties of the air mass may often be deduced by tracing it back to its source. For this purpose a series of synoptic charts, preferably at 6-hourly intervals with the isobars accurately drawn, should be used.

The procedure is as follows :—Suppose it is required to find the source of the air arriving at a position A (50° N., 17° W.) on synoptic chart of Fig. 145. Assume that during the 6-hour period centering at the time of the chart the movement of the air is given by the geostrophic wind determined from that chart. At A this is 40 knots. For this chart, therefore, it is assumed that during the preceding 3 hours that air has moved with the geostrophic speed at A. Project back the air at A for a period of 3 hours, by drawing a line AB parallel to the isobar at A, in the direction from which the wind has come, of length equal to the distance that the air has travelled in 3 hours, *i.e.*, 120 nautical miles, obtained by multiplying the geostrophic wind speed, in knots, by 3. Similarly, project back from B to C for a period of 6 hours using the geostrophic wind

determined by the isobars at *B*, for the chart drawn 6 hours previously, *i.e.*, Fig. 144. Repeat this process, using for each 6-hour period the geostrophic wind given by the synoptic chart for the mid-point of the period. In this way the approximate track of the air may be obtained.

Upper-Air Temperature Graphs.—In order that observations of upper-air temperature may be utilized in weather forecasting they are plotted in graphical form. A simple form is that known as the Upper-Air Temperature Graph plotted on the diagram shown in Fig. 155. The co-ordinates are pressure and temperature, the scales being logarithmic. For convenience in plotting, the scale of pressure is graduated for multiples of 50 mb. corresponding to the levels at which the temperature is read in the aircraft ascent. If required, the heights may be found from the pressure-height formula.

The shaded belt running across the diagram shows the limits of the mean values of the temperature over south-east England. The highest mean temperature occurs in August, the lowest in February, as indicated by the two bounding lines. The irregular dotted lines labelled respectively 5th April, 1911 and 2nd October, 1908, are examples in the nature of extreme values for south-east England.

The full sloping lines of which *AB* is an example, from the base of the diagram to the line of 900 mb., are dry adiabatics, drawn on the assumption that on the average, air ascending from the surface becomes saturated at 900 mb. Above this level the lines continue as saturation adiabatics, typified by the slightly curved, pecked line *BC*. Such of the saturation adiabatics as cross the line of 32° F., as at *D*, there show a discontinuity, due to the latent heat released on freezing of the condensed water.

The height is determined by means of the scale on the right of the diagram, as follows :—

> On the pressure scale on the left, mark off a point corresponding to the M.S.L. pressure and draw a horizontal line through it to cut the sloping line marked " surface " on the height scale. From the point of intersection draw a vertical line, which, by its intersections with the sloping lines marked 2,000, 4,000 feet, etc., forms the appropriate height scale.

Example.—Surface pressure, 1,030 mb. Mark off the point *P* on the pressure scale corresponding to 1,030 mb. and draw *PQ* horizontally to cut the line marked " surface " in *Q*. From *Q* draw the vertical *QR*, which then gives the height scale required.

To find the height corresponding to a given pressure, say 650 mb., produce the horizontal line marked 650 to cut *QR* in *S*. This is also the point where the sloping line marked 12,000 cuts *QR* and the required height is therefore 12,000 feet. If *S* does not fall on one of the sloping lines the height must be found by interpolation along *QR*.

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GRAPHS OF UPPER AIR TEMPERATURE

The pressure and temperature scales are logarithmic. All changes of temperature according to the dry adiabatic law are therefore represented by parallel straight lines. The broken lines show lapse rates for saturated air. These are for air masses which contain just sufficient moisture at 1015 mb pressure to be saturated after rising 1 kilometre in an environment having the dry adiabatic lapse rate.

The condensed moisture is assumed to be retained thereafter in the air as it rises. The full lines running from the bottom of these broken lines to the base of the diagram are dry adiabatics. The scale on the right-hand side shows where heights of 2,000, 4,000 ft etc, occur with various pressures and with the normal temperatures over S.E. England corresponding with these surface pressures (from Table VII of Geophysical Memoir No 2). The two dotted lines show extremes of temperature measured in the South of England.

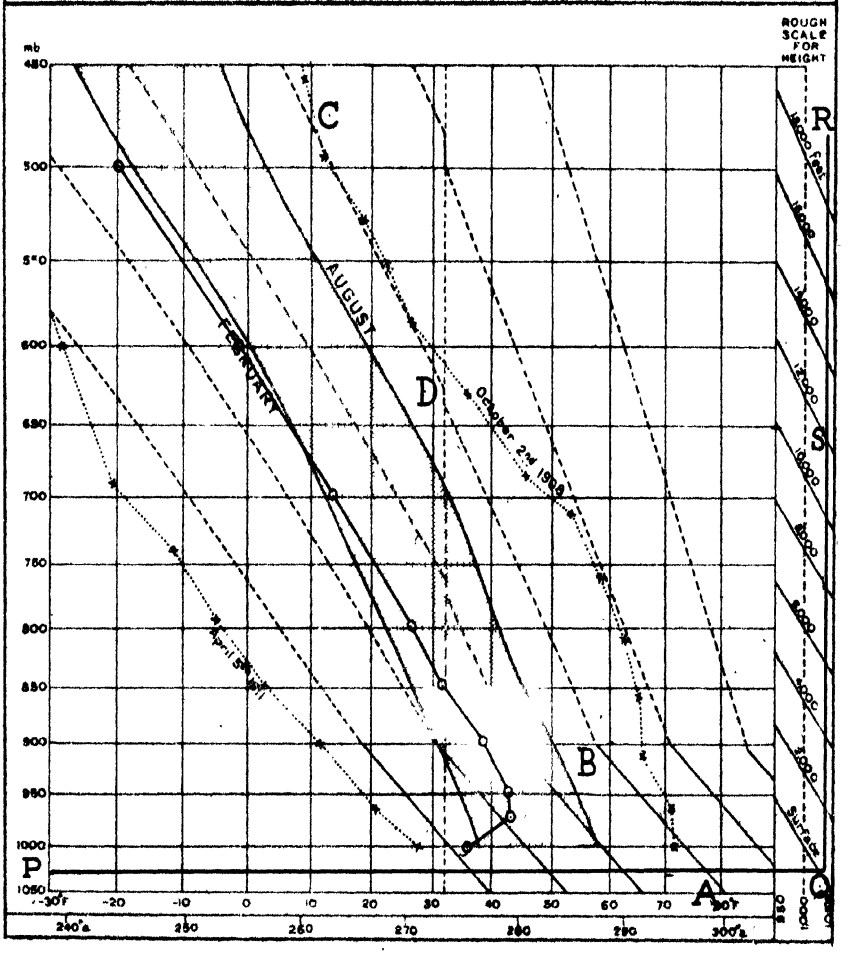
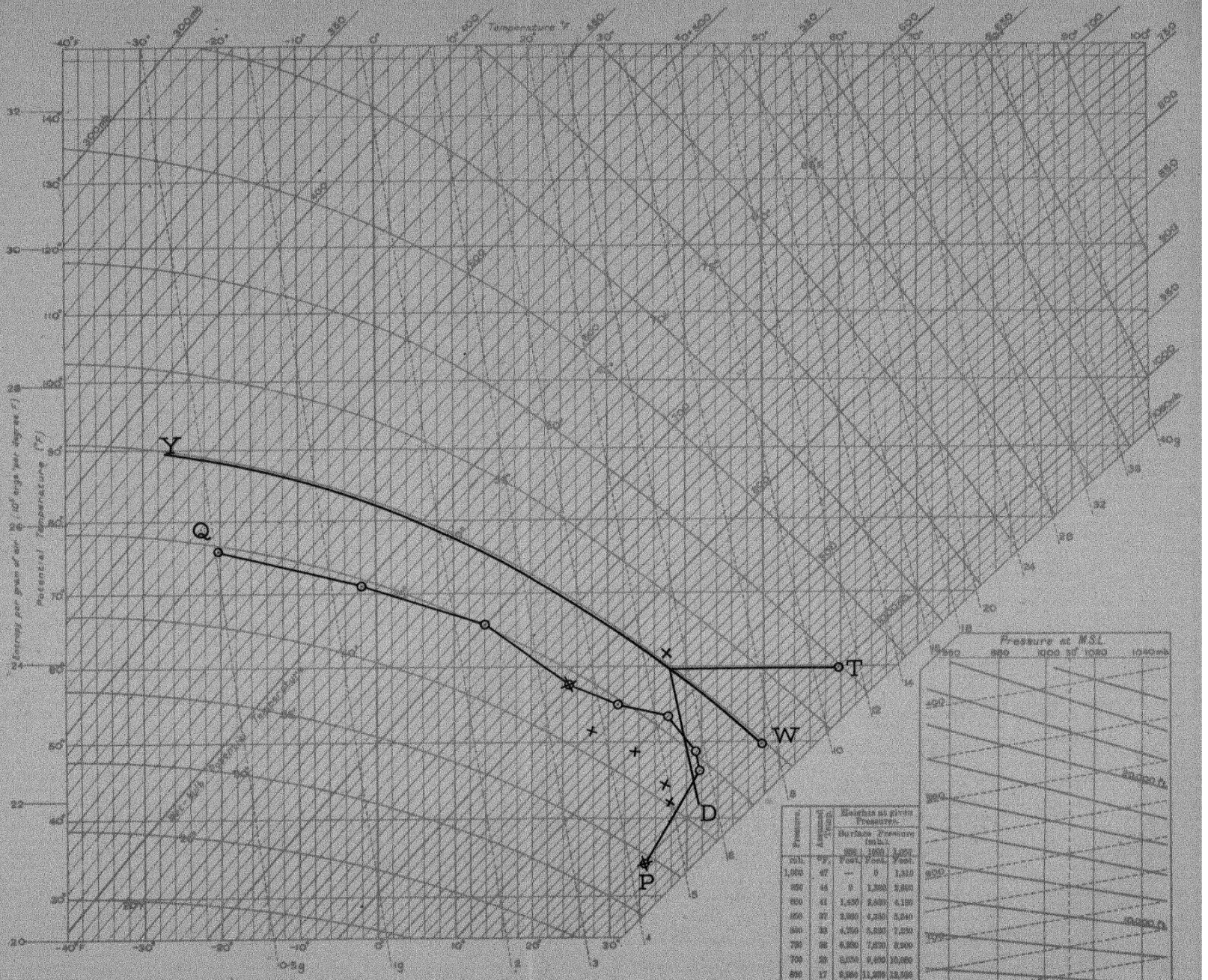


FIG. 155. GRAPHS OF UPPER-AIR TEMPERATURE

T-φ GRAM FOR USE WITH AEROPLANE RECORDS.



Pressure	Approximate Height	Heights at given Pressures.			
		Surface Pressure (m.b.s.)			
		950	1000	1050	1100
m.b.	°F.	Feet	Feet	Feet	Feet
1,000	47	—	0	1,310	2,620
950	45	0	1,360	2,650	3,940
900	41	1,430	2,820	4,170	5,500
850	37	2,960	4,230	5,540	6,850
800	33	4,750	5,930	7,150	8,370
750	24	6,820	7,870	8,940	10,000
700	20	9,070	9,950	10,860	11,770
650	17	9,980	11,370	12,770	14,170
600	13	11,980	13,390	14,800	16,210
550	1	14,100	15,500	16,910	18,320
500	-7	16,350	17,730	19,120	20,510
450	-17	18,770	20,150	21,530	22,910
400	-27	21,350	22,690	24,030	25,370

Let the "condition" of a gram of air be represented by a point on the diagram. Change the condition of the gram of air so that that this point describes a closed curve of one sq cm area. The work required to effect this change is 2·78 x 10⁷ ergs (if the curve is described in a counter-clockwise direction). It is independent of the shape of the closed curve or its position on the diagram and depends only on its area, (i.e. for 2 sq cm the energy is twice as much.)

FIG. 156.

The information regarding the results of upper-air observation is usually given in coded form. The various forms of message are described on page 317. That most commonly used gives the temperature at various pressure-levels. From it, a graph can be plotted on the diagram. In Fig. 155 the specimen message interpreted on page 318 has been plotted and the series of points obtained joined by a curve giving the graph of upper-air temperature for the ascent under consideration.

The upper-air temperature graph gives a general picture of the variation of temperature in the upper air, and may be used to estimate roughly the degree of stability or instability of the atmosphere. The lapse-rate is determined by the slope of the graph—the more nearly horizontal the graph the greater is the lapse-rate. The slope should in practice be compared with the slopes of the dry and saturation adiabatics, and the general conclusions drawn as follows :—

(i) A layer in which the lapse-rate is less than the S.A.L.R. is stable. If the lapse-rate is less than the S.A.L.R. at all levels, the atmosphere is stable throughout and no convection phenomena, viz., cumulus cloud, showers or thunderstorms are to be expected.

(ii) A layer in which the lapse-rate exceeds the D.A.L.R. is unstable. This rarely occurs except in the lowest layers as a result of powerful heating at the surface.

(iii) A layer in which the lapse-rate lies between the D.A.L.R. and the S.A.L.R. is unstable only if the air is saturated and is said to be conditionally unstable. If the lapse-rate exceeds the S.A.L.R. through a considerable range of height, convection phenomena are probable.

The Tephigram

The properties of, and changes which take place in moist air may be conveniently studied with the aid of the Tephigram, a diagram devised by Sir Napier Shaw. It has the advantage over the upper-air temperature graph of dealing quantitatively with the properties of damp air, namely, wet-bulb temperature, dew-point, relative humidity and water-vapour content. As with the graph, the path curve and environment curve are plotted on the diagram, and the stability or instability of the atmosphere investigated. The tephigram is shown in Fig. 156. Its framework consists of a number of systems of straight lines and curves, as follows :—

(*a*) **Lines of Constant Temperature or Isothermal Lines.**—The variation of temperature is represented horizontally on a uniform scale. The lines of constant temperature are therefore vertical lines spaced at uniform intervals. Lines are drawn for every 2° F. from — 40° to + 100° F., the lines for 10°, 20°, etc., being thickened.

(*b*) **Lines of Constant Potential Temperature.**—The vertical axis of the diagram is graduated primarily in units of entropy, and the horizontal lines on the diagram are therefore lines of constant entropy. Now, it may be shown that the entropy of a mass of air is proportional to the logarithm of its potential temperature, so that the lines of constant entropy are also lines of constant potential temperature. Moreover, the potential temperature of a mass of air is constant so long as it is subject only to dry-adiabatic changes ; a line of constant potential temperature is therefore also a dry adiabatic. Every horizontal line on the tephigram is therefore a dry adiabatic and adiabatic changes of temperature and pressure in a mass of unsaturated air correspond to the movement of its representative point along such a line. Thereby the difficult conception of entropy is avoided.

The horizontal lines are drawn at intervals of 10° F. from 30° F. to 140° F. If the potential temperature and ordinary temperature of a mass of air are known, its representative point may be plotted with respect to the horizontal and vertical lines.

(*c*) **Lines of Constant Pressure.**—The slightly curved lines sloping upwards from left to right are lines of equal pressure, drawn at intervals of 10 mb. For convenience in plotting, the lines for multiples of 50 mb. are thickened. The formula for potential temperature is used for the construction of these lines. This formula (*see* page 151) is :—

$$\log \theta = \log T - 0{\cdot}286 \log \frac{p}{1000}$$

where θ = potential temperature
T = actual temperature
p = pressure in mb.
} of the air.

By assigning a definite value to p a relation between θ and T is obtained from this formula. On plotting with respect to the network of horizontal lines on the tephigram the line of constant pressure p is obtained. It should be noted that in the above formula θ and T are expressed in degrees absolute and were converted to Fahrenheit before constructing the diagram.

(*d*) **Saturation Adiabatics.**—The curved lines sloping upwards from right to left of the diagram are "irreversible saturation adiabatics," computed as relations between pressure and temperature using the known saturation values of water-vapour content at varying pressures

(*d*) **Saturation Adiabatics.**—contd.
and temperatures. They are computed on the hypotheses (i) that all condensed water falls out of the air as soon as it is formed; and (ii) that water in the atmosphere freezes at 32° F. Neither assumption is always true. The diagram is constructed so that a saturation adiabatic line passes through every point of intersection of the 1,000 mb. line with each 10° F. isothermal line. As in the temperature-height diagram, the saturation adiabatics are path curves for rising saturated air. At low temperatures where there is little vapour in a saturated atmosphere the S.A.L.R. approaches the D.A.L.R. and the saturation adiabatics become nearly parallel to the dry adiabatics, *i.e.*, they tend to become horizontal. On the other hand, with high temperature where the S.A.L.R. becomes small the saturation adiabatics tend to become more nearly vertical.

(*e*) **Lines of Equal Water-Vapour Content for Saturation.**—The system of nearly vertical pecked lines are lines of equal saturation values of water-vapour content expressed in grams of water vapour per kilogram of dry air. They are labelled from 0·5 g. to 40 g. indicating saturation water-vapour contents from 0·5 grams to 40 grams of water vapour per kilogram of dry air. Air at the temperature and pressure represented by any point on the diagram will be saturated if it contains water vapour to the extent indicated by the water-vapour content line passing through the point. If no water-vapour content line passes through the point, the saturation value of the water-vapour content must be estimated by interpolation with respect to the pecked lines.

Example.—Temperature 50° F. Pressure 770 mb. The water-vapour content line labelled 10 g. passes through the point 50° F., 770 mb. Therefore 10 grams of water vapour per kilogram of dry air will saturate air at the given temperature and pressure.

RELATIONS BETWEEN THE TEMPERATURE, RELATIVE HUMIDITY, DEW-POINT AND WET-BULB TEMPERATURE GIVEN BY THE TEPHIGRAM.

(1) **Given the Pressure and Water-Vapour Content, to find the Dew-Point of a Mass of Air.**—Since the dew-point is the temperature at which the amount of water vapour present would be just sufficient to saturate the air, it is the temperature at which the given water-vapour content line cuts the line corresponding to the given pressure.

Example.—Water-vapour content 16 g./kg., pressure 1,000 mb. The pecked line 16 g. cuts the 1,000 mb. line at 70° F., which temperature is therefore the required dew-point.

(2) **Given the Pressure, Temperature and Relative Humidity, to find the Water-Vapour Content and the Dew-Point.**—From the diagram read off the water-vapour content for saturation at the given pressure and temperature. Take the percentage value of this corresponding to the relative humidity to obtain the actual water-vapour content. To obtain the dew-point proceed now as in (1).

Example.—Temperature 70° F., pressure 1,000 mb., relative humidity 60 per cent. From the diagram, the saturation value of the water-vapour content corresponding to 70° F., 1,000 mb. is 16 g./kg. The actual water-vapour content

$$= 60 \text{ per cent. of } 16 \text{ g./kg.}$$
$$= 9 \cdot 6 \text{ g./kg.}$$

The dew-point is found as in (1) to be 56° F.

(3) **Given the Dry- and Wet-Bulb Temperature and the Pressure, to find the Dew-Point, Water-Vapour Content and the Relative Humidity.**—It is more convenient in upper-air ascents to record the wet-bulb temperature than either the relative humidity or the dew-point. By a theorem due to Normand the wet-bulb temperature can be related to the other properties.

Normand's theorem is as follows :—

If simultaneous readings of the dry bulb, ventilated wet bulb and temperature of the dew-point are plotted on the isobar appropriate to the place of observation, then the horizontal (dry adiabatic) line through the dry bulb, the saturation adiabatic line drawn through the wet bulb, and the vapour-content line through the dew-point will meet in a point.

In Fig. 156 let T, W and D respectively represent the dry-bulb temperature, wet-bulb temperature and dew-point of a mass of air whose pressure is (for example) 1,000 mb. Normand's theorem then states that the dry adiabatic TX through T, the saturation adiabatic WX through W and the vapour-content line DX through D all meet in a point X. The lines TX, WX and DX must be interpolated with respect to the various systems of lines, if necessary.

Example.—Pressure 1,000 mb., dry-bulb temperature T, 60° F. wet-bulb temperature W, 50° F. Plot T and W at 60° F. and 50°F. respectively on the 1,000 mb. line. Draw the dry adiabatic TX through T and the saturation adiabatic WX through W to meet in X (pressure 863 mb., temperature 38·5° F.). By interpolation the water-vapour content corresponding to point X is 5·7 g./kg. Draw the line of constant vapour content 5·7 g./kg. from X to cut the 1,000 mb. line in D, which then gives the dew-point temperature (42° F.). By (*e*) above the water-vapour content of the air is 5·7 g./kg. The water-vapour content for saturation at 60° F., 1,000 mb. is 11·4 g./kg. The relative humidity is therefore $100 \times 5{\cdot}7/11{\cdot}4$, i.e. 50%.

The Environment Curve and Path Curve.—The network of isotherms and pressure lines is used as a frame of reference for plotting observations of temperature in the upper air. Such observations are usually available in the message forms described on page 317, which give corresponding values of pressure and temperature. By way of illustration the observations contained in the specimen message interpreted on page 318 are plotted on a tephigram in Fig. 156. The curve PQ thus obtained is called the environment curve. The dew-points are plotted by the method described on page 436, and are shown in Fig. 156 as a series of crosses.

The curve traced out on the tephigram by the representative point for air moving vertically is called the path curve. For rising unsaturated air and for descending air the path curve is a dry adiabatic, *i.e.*, a horizontal line. The path curve for rising saturated air is clearly a saturation adiabatic. The provision of saturation adiabatics and saturation water-vapour content lines on the tephigram enable the path curve for rising air to be accurately plotted if the initial pressure, dry-bulb and wet-bulb or dew-point temperatures are known.

(1) *Temperature, pressure and dew-point known.*—In Fig. 156 let the initial pressure of the air be 1,000 mb., dry-bulb temperature 60° F. and dew-point 42° F., the last two being represented by the points T and D respectively. The water-vapour content of the air is that corresponding to the vapour-content line DX. The air being unsaturated its path curve is at first the dry adiabatic TX. At the point X the temperature of the air has been sufficiently reduced that the vapour content, which has remained constant during the process of rising, becomes just sufficient to saturate the air and subsequent cooling is accompanied by condensation. X is therefore called the condensation point. It corresponds to the level of the base of the cloud formed by convection of air from the surface and may be used to forecast the height of clouds formed by this process, *i.e.*, Cu. and Cunb. It is the point of intersection of the dry adabatic through T and the vapour-content line through D. Beyond X the path curve is continued as the saturation adiabatic XY. Therefore, to construct the path curve for air whose pressure, temperature and dew-point are known, plot the temperature T and dew-point D on the appropriate pressure line. Draw the vapour-content line through D and the dry adiabatic through T to meet in a point X and draw the saturation adiabatic XY. Then TXY is the required path curve.

(2) *Temperature, pressure and wet bulb known.*—Plot the dry-bulb temperature T and wet bulb W on the appropriate isobar. Then by Normand's theorem (page 436) the saturation adiabatic through W meets the dry adiabatic in the condensation point X determined as in (1). Therefore draw the dry adiabatic through T and the saturation adiabatic through W to meet in a point X and continue the saturation adiabatic WX to Y. Then TXY is the path curve.

(3) *Temperature, pressure and relative humidity known.*—Find the dew-point as in (2) (page 436), and proceed as in (1) above.

Stability and Instability.—The conditions as regards lapse-rate for stability and instability in the atmosphere have been considered on page 433. They are here re-stated with special reference to the tephigram.

An atmospheric layer for which the environment curve slopes downwards from right to left is unstable, since the lapse-rate exceeds the D.A.L.R. An atmospheric layer for which the environment curve at any point is more nearly vertical than the saturation adiabatic passing through the point is stable, since the lapse-rate is less than the S.A.L.R.

An atmospheric layer for which at any point the slope of the environment curve lies between the horizontal and the slope of the saturation adiabatic through the point, is conditionally unstable, since the lapse-rate lies between the D.A.L.R. and the S.A.L.R.

Positive and Negative Areas on the Tephigram.—It has been seen on page 155 that convection will go on spontaneously only if the path curve on the temperature-height diagram lies to the right of the environment curve. The considerations on page 155 apply equally to the tephigram. Air will continue to rise if at any level (corresponding to a given pressure line) its path curve on the tephigram lies to the right of the environment curve.

The area between the path curve and the environment curve is proportional to the energy (per unit mass of rising air) involved in the process of rising. Two cases arise. In the first case the path curve lies to the right of the environment curve; the rising air, being warmer than its environment, ascends freely, and energy is released in the convective process. The released energy maintains the vigorous ascent of the upward current by supplying it with kinetic energy, and is also the source of the electrical energy should a thunderstorm develop. A large area between the path curve and the environment curve is therefore associated with vigorous convection phenomena. The area representing released energy is called positive area. It is usual to shade it in red on the working tephigram. In the second case, the path curve lies to the left of the environment curve, and the air if it were forced

to rise would be colder than its environment, and therefore lack buoyancy. Energy would have to be expended to lift the air. The area between the two curves in this case represents a deficit of energy which must be made good before the air can be made to rise. In general, therefore, positive areas indicate instability and negative areas stability.

It often happens that the path curve and the environment curve do not intersect in the upper levels, so that the area between them is not closed and its magnitude cannot be determined, as in the case of the negative area of Fig. 157. This is, however, immaterial in practice, since the relative magnitudes of the areas are considered only qualitatively as indicating the degree of stability or the vigour of convection.

Types of Tephigrams.—Tephigrams may be classified into types according as to whether they yield positive or negative areas. Examples of the various types are shown in Figs. 157–9. In these diagrams, which are not drawn to scale, only the pressure lines and the saturation adiabatics have been shown, to avoid confusion. Positive areas are shaded, negative unshaded.

(1) *Stable Type* (Fig. 157).—Temperature is high in the upper levels and the lapse-rate is less than the S.A.L.R.

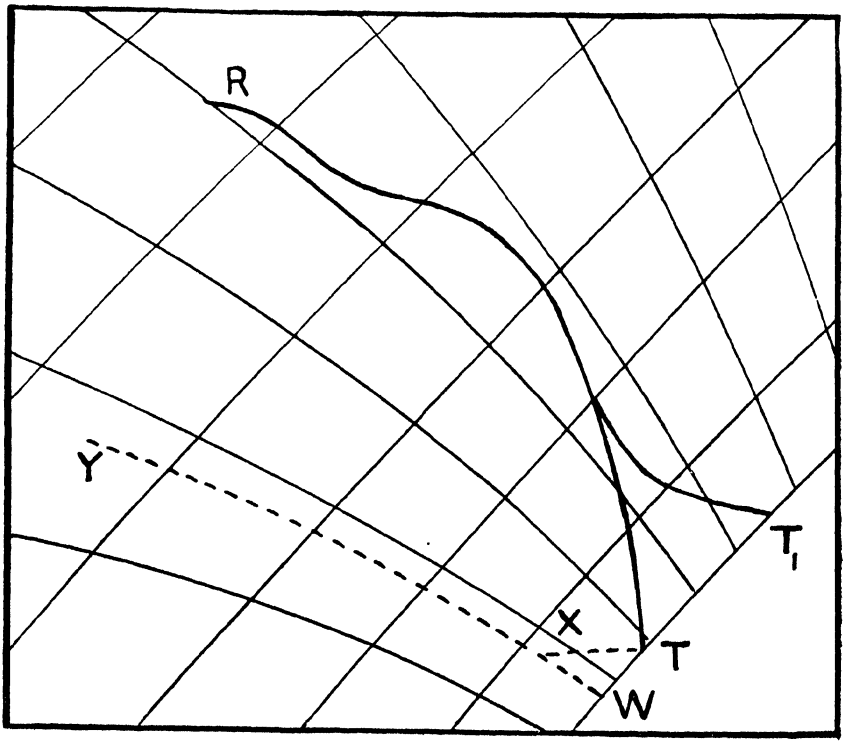


FIG. 157.—TEPHIGRAM—STABLE TYPE

(1) *Stable Type.*—contd.
through a large range of height. The environment curve takes the form TR and the path curve for surface air, TXY. There is a large negative area, indicating extreme stability. This type of tephigram occurs chiefly with anticyclones, the high upper-air temperature being due to subsidence. Over land in winter there is, however, frequently a pronounced inversion in the lowest layers, with consequently very low surface temperature. Over land on summer afternoons the lapse-rate in the lowest layers may approach the D.A.L.R. due to surface heating, the environment curve taking the form T_1R, in which case there may be sufficient convection to bring about the formation of small cumulus.

(2) *Convective Type.*—The lapse-rate is equal to or slightly exceeds the S.A.L.R. through a large range of height. In the lowest layers the temperature distribution is often a stable one, and the environment curve then takes the form $T'ZR$ of Fig. 158. In such a case a typical path curve is $T'X'Y'$, which yields a positive area RZY'

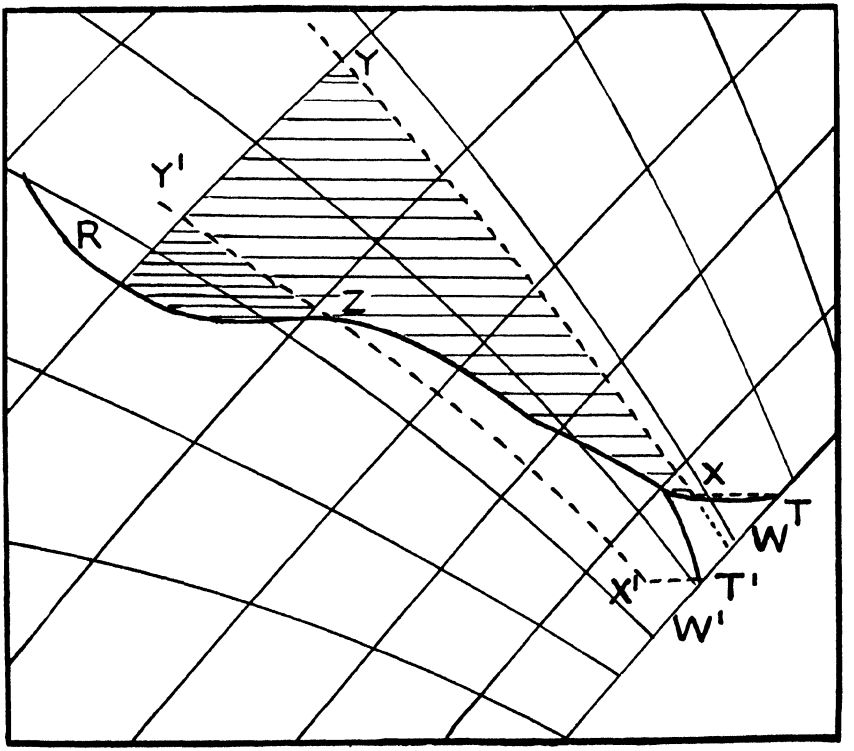


FIG. 158.—TEPHIGRAM—UNSTABLE TYPE.

only at very high levels. Convection would not take place owing to the negative area $T'X'Z$. Nevertheless, the fact that the lapse-rate exceeds the S.A.L.R. in all except the lowest layers, indicates the possibility of convection should the surface dry-bulb and wet-bulb temperatures rise. Assuming that these rise to T and W, respectively, while the temperature distribution in the upper levels remains unchanged, the environment and path curves take the forms TZR and TXY, and yield the large positive area $TZRYX$ with the likelihood of powerful convection.

The convective type of tephigram is obtained from ascents made in polar air which has become heated from below as a result of slow movement over heated land or a warm sea-surface. $T'ZR$ is typical of the curve obtained in the early morning when, owing to nocturnal cooling, the surface temperature has been reduced and the lower air stabilized. As a result of heating at the surface by insolation the dry-bulb and wet-bulb temperatures rise during the forenoon, and by midday the environment curve takes the form TZR. The heating at the surface is required to start off the convective process. It is for this reason that showers and thunderstorms are most frequent in the afternoon.

(3) Sometimes the tephigram takes the form shown in Fig. 159,

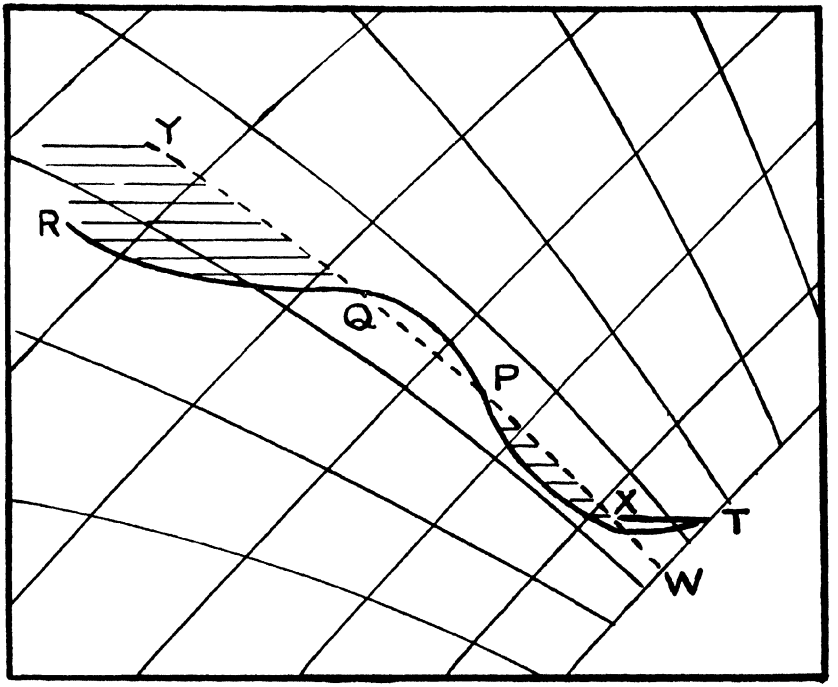


FIG. 159.—TEPHIGRAM WITH ALTERNATE POSITIVE AND NEGATIVE AREAS.

the principal characteristic being a negative area in the layer PQ between two positive areas RQY and TXP. Provided that the negative area is small, this type is effectively the same as type (2), for, convection having once started below P, the rising air, by virtue of its own momentum, may penetrate the layer corresponding to the negative area, and continue to rise spontaneously in the layer corresponding to the upper positive area.

Fronts and the Tephigram.

The tephigram sometimes gives useful information regarding fronts and occlusions. Consider first the case of an approaching warm front. The vertical structure is as shown in Fig. 160, where XY represents the ground and AB the sloping frontal surface. Suppose that an aircraft ascent indicated by the vertical line PQR is made from a point P far in advance of the front. Denote the points corresponding to P, Q and R by the same letters on the tephigram of Fig. 161. Within the cold air the lapse rate may be

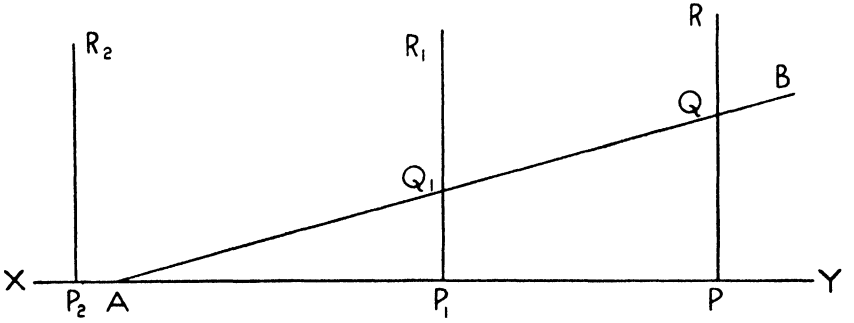


Fig. 160.

regarded as normal and the environment curve on the tephigram approximates to a saturation adiabatic PQ. At Q the ascent reaches the frontal surface. Owing to mixing, the boundary between the cold and the warm air is not a sharp surface, but a diffuse transition layer one or two thousand feet thick. Above this layer is the warm air with, it may be assumed, normal lapse-rate. On account of the general lapse of temperature with height the warm air is not necessarily at a higher temperature than the cold air beneath. There may be a continuous fall of temperature through the transition layer at Q. Sometimes there is an isothermal layer or even an inversion at Q, but more often the transition layer is one in which the lapse-rate is less than normal. On the tephigram this is represented by an upward bend in the environment curve at Q, above which the normal lapse-rate is resumed. The air at R is virtually warmer than that below Q since, if the normal lapse-rate were continued

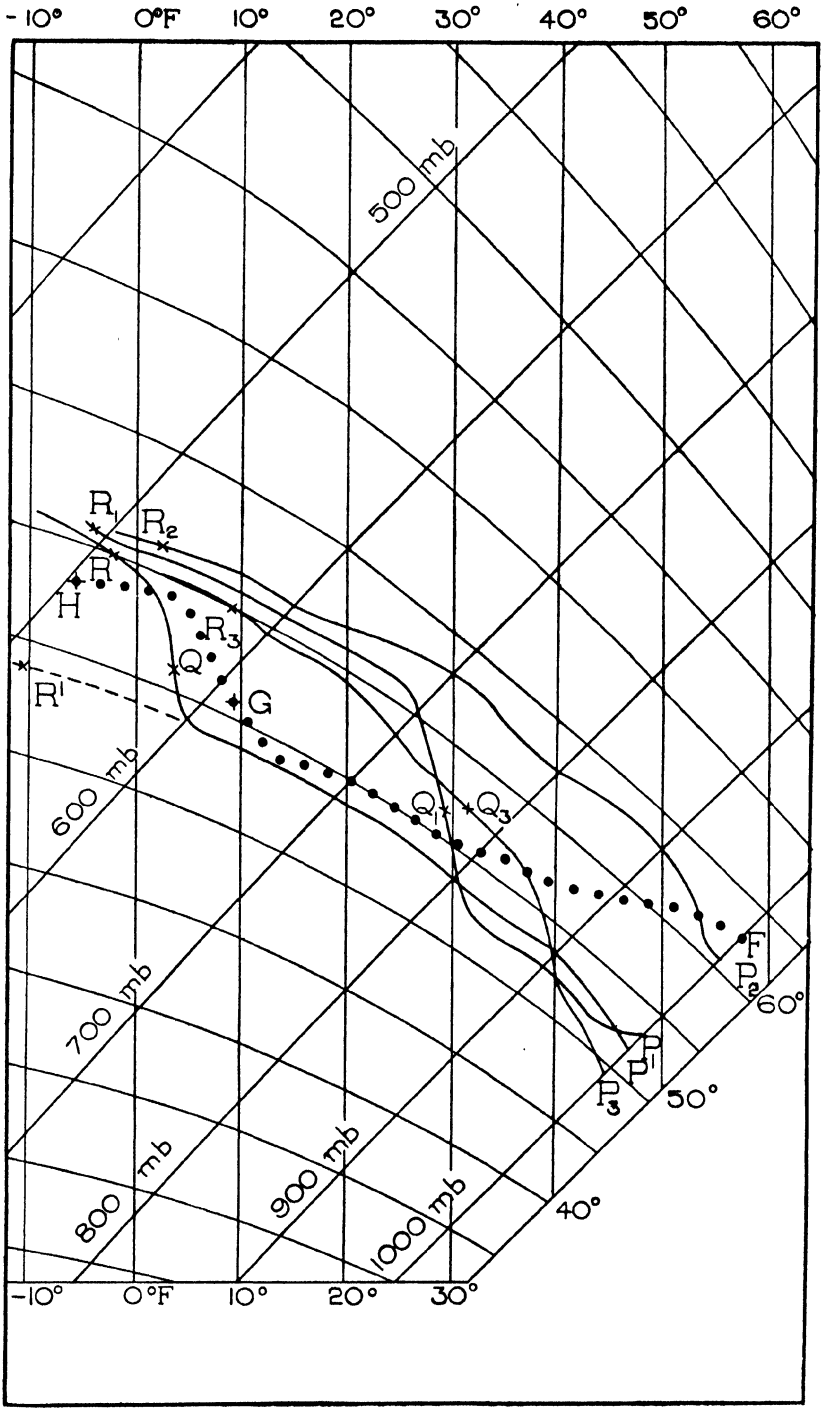


FIG. 161.—TEPHIGRAMS ASSOCIATED WITH FRONTS.

from P up to the level of R as indicated by the line PR' the temperature at the level R would be represented by R'.

The upward bend in the curve at Q and the relatively high temperature at R indicate the presence of warm air aloft and should be regarded as signs of an approaching warm front or occlusion.

A second ascent made from the same station as the first but some hours later may be represented by the vertical line $P_1Q_1R_1$, Fig. 160, where P_1 lies between P and A. As in the former case the transitional layer at Q_1 is indicated by an upward bend on the tephigram but at a lower level. The curve therefore takes the form $P_1Q_1R_1$, Fig. 161. Below the transition layer the second curve is approximately coincident with the first, since little change of temperature takes place in the cold air below the frontal surface.

A third ascent made after the passage of the warm front may be represented by the line P_2R_2, Fig. 160, to the left of A. The ascent takes place entirely in the warm air and the temperature is relatively high at all levels. The curve is similar to P_2R_2, Fig. 161, and there is no layer with diminished lapse-rate.

The foregoing may be summarized by saying that an approaching warm front or occlusion is indicated on the tephigram by a layer with lapse-rate less than normal, which appears at successively lower levels on the series of curves obtained from ascents at a particular station as the front approaches.

If a definite assumption is made regarding the slope of the frontal surface, the tephigram, in the case of a warm front, may be used to indicate the position of the front at the ground (point A of Fig. 160). It is usual to assume a slope of 1/100; therefore $PA = 100\ PQ$, and if PQ is known, PA can be determined. Q is taken as the mid-point of the transition layer. Its pressure can then be read off and its height approximately determined from the small table which appears on the tephigram form. For example, if the M.S.L. pressure is 1,000 mb. and the pressure at Q is 580 mb. the height of Q is approximately 14,000 feet, or $2\frac{1}{2}$ miles. PA is therefore 250 miles. The above calculation cannot be applied to an occlusion since the sloping frontal surface does not reach the ground.

Sometimes with a warm front or occlusion there is no well-defined transition layer, but the reduced lapse-rate is spread over a large range of height giving a curve similar to $P_3Q_3R_3$ of Fig. 161.

Above the frontal surface there is usually a thick layer of cloud extending often up to the cirrus level. Within the cloud layer the air is saturated and on the tephigram the dry-bulb and wet-bulb temperatures coincide.

Similar forms of tephigram are obtained in the rear of a cold front or occlusion. In Fig. 162 let KL represent the ground and MN the sloping frontal surface. Ascents after the passage of the cold front are represented by the vertical lines FGH, $F_1G_1H_1$.

At G and G_1 occurs the transition layer. A curve corresponding to the ascent at F is shown at FGH in Fig. 161. This is an actual record from an ascent at Duxford, Cambs., at 1300 G.M.T. on 30th September, 1935. The cold front then lay in a N.—S. direction

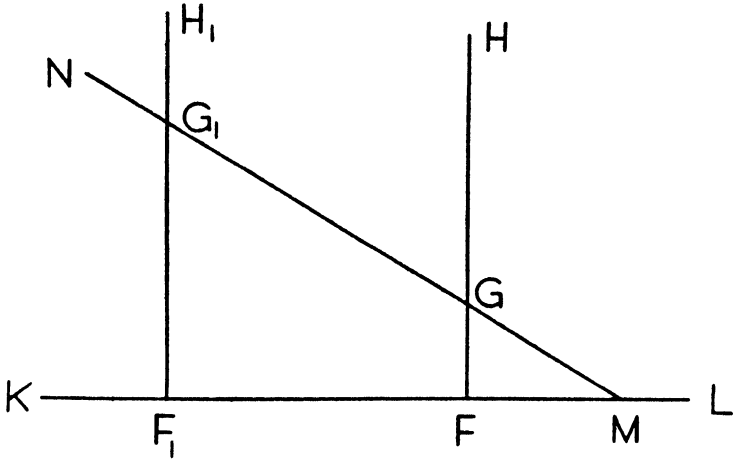


FIG. 162.

about 120 miles east of Duxford. A later ascent would show the transitional layer at a higher level.

If the ascent takes place in the rear of a cold front its distance at the ground may be estimated as in the case of the warm front except that the assumed slope is 1/50.

Trigger Action.—It has been stated in Chapter VII that convection in a conditionally unstable atmosphere may be started by a number of processes, known collectively as trigger action. These processes are here explained in relation to the tephigram.

(1) *Heating of the Surface Air.*—In Fig. 163 let TR represent the environment curve in which the surface values of the dry-bulb and wet-bulb temperatures are T and W respectively. By the usual construction the path curve $TXX'Y$ is obtained, X' being the point where it cuts the environment curve. Below the level of X' the path curve lies to the left of the environment curve and therefore convection will not start spontaneously from the surface. If now owing to heating at the surface the temperature there is raised to T', where T' is such that $X'T'$ is horizontal, and the wet-bulb temperature remains unaltered, the new path curve is $T'X'Y$. Since this lies entirely to the right of the environment curve convection is now spontaneous.

(2) *Increase of Water-Vapour Content of the Surface Air.*—Let the water-vapour content of the surface air increase so that its wet-bulb temperature rises to T, without change of dry-bulb temperature. The path curve is then TY' and lies entirely to the right of the environment curve TR and convection is spontaneous.

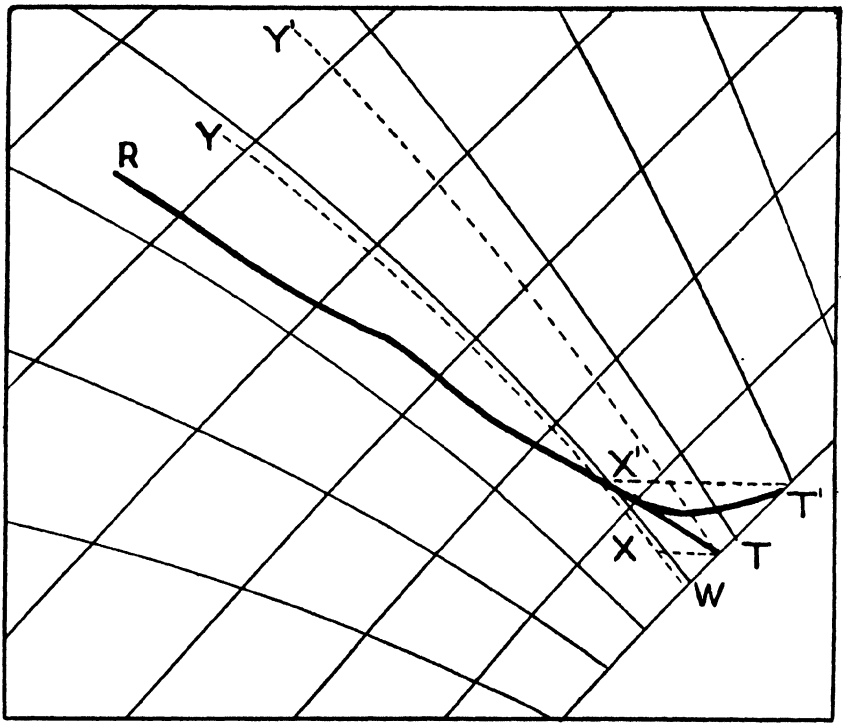


FIG. 163.—TEPHIGRAM ILLUSTRATING TRIGGER ACTION.

(3) *Mechanical Lifting of the Surface Air.*—If the surface air is raised mechanically, it follows the path curve $TXX'Y$. Below X' its temperature is lower than that of its environment so that a lifting process by some external agent is necessary to raise the air to the level X'. Above X' the air will ascend spontaneously.

In practice the lifting may be due to the forced ascent of an air-stream over a mountain barrier or at a front.

In the foregoing, only the path curves for air at the surface have been considered. The reasoning can clearly be applied without modification to the air at any level. In the majority of cases, however, it will be the surface air that is involved in the convective process, since that air is most likely to be heated or to acquire water vapour at the surface.

To face page 447

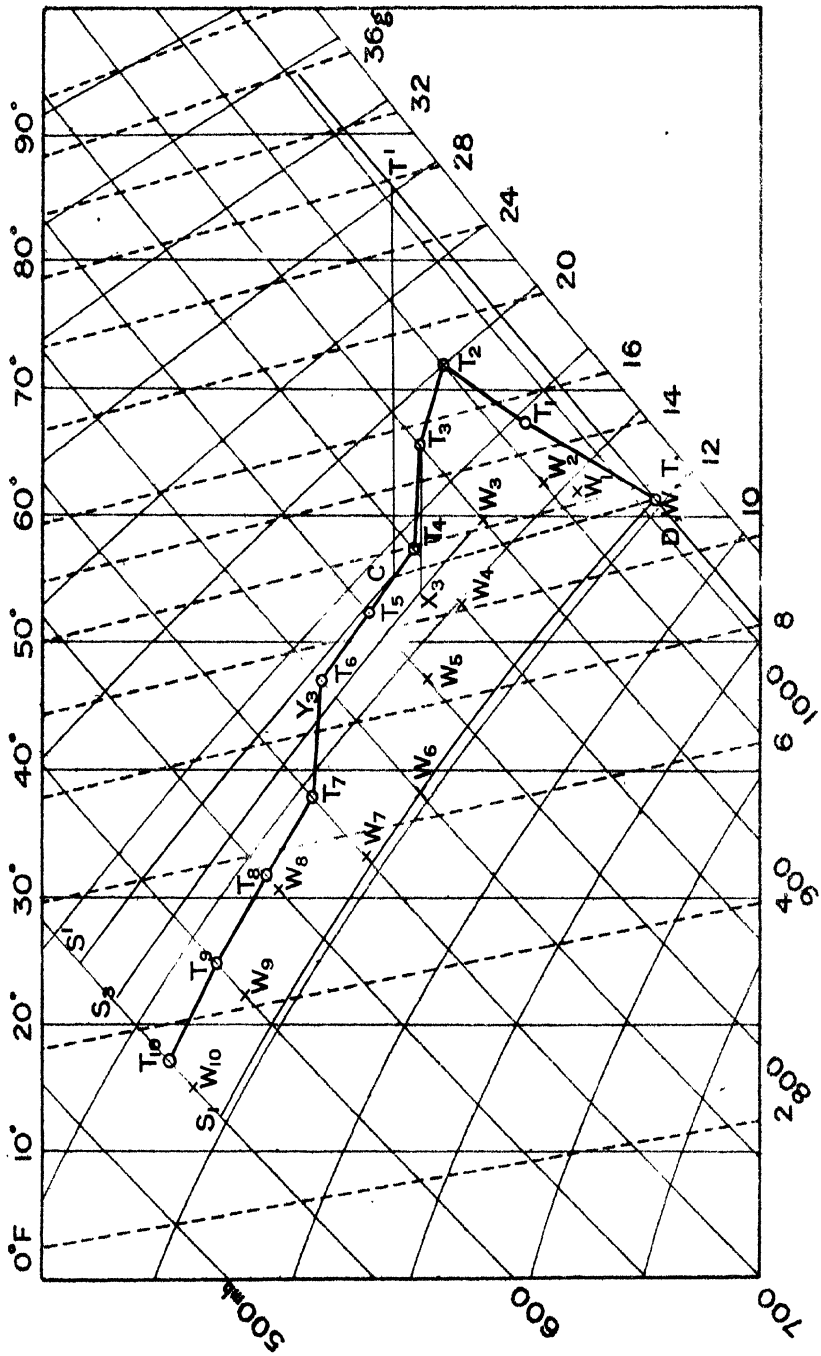


FIG. 164. TEPHIGRAM FOR HENLOW 0600 G.M.T. 25 JUNE 1935.

An Example of the Forecasting of Thunderstorms with the Aid of the Tephigram.

The following table gives the dry-bulb and wet-bulb temperatures observed in an ascent made at Henlow (Beds.) at 0600 G.M.T. on 25th June, 1935.

Pressure.	Dry bulb.	Wet bulb.
mb.	° F.	° F.
1,005	61·0	61·0
970	67·5	62·0
950	72·0	62·5
900	65·8	59·8
850	57·8	53·5
800	52·2	46·8
750	46·8	38·0
700	38·0	33·0
650	32·0	31·0
600	25·0	22·2
550	17·4	14·8

The dry-bulb and wet-bulb temperatures have been plotted on the tephigram of Fig. 164, giving the two series of points $TT_1T_2..T_{10}$ and $WW_1W_2..W_{10}$, respectively. The former have been joined to form the environment curve. Above T_2 the lapse-rate exceeds the S.A.L.R., indicating conditional instability, but the large inversion of temperature between T and T_2 prevents convection of the surface air. Since the air at the surface is saturated (W coincides with T) its path curve is the saturation adiabatic WS_1, which yields a large negative area, again showing that convection cannot take place freely from the surface.

Consider the path curve for the air at the 900-mb. level. Its dry-bulb temperature is T_3 and wet-bulb W_3. Draw the dry adiabatic T_3X_3 and the saturation adiabatic W_3X_3 to meet in the condensation point X_3, and produce the saturation adiabatic to S_3, cutting the environment curve in Y_3 (730 mb.). Then $T_3X_3Y_3S_3$ is the path curve. Below Y_3 it yields the negative area $Y_3X_3T_3T_6$, and above Y_3 the positive area $Y_3S_3T_{10}T_7$. Only if the air at 900 mb. were lifted mechanically to 730 mb. would it proceed to rise freely. Many tephigrams are of this type, showing alternate positive and negative areas. The probability of spontaneous convection must be estimated having regard to the disposition and magnitude of the positive and negative areas. Theoretically, path curves should be drawn for the air at all levels, but only those which yield positive areas are of interest; the rest can be ruled out as affording no possibility of convection. Generally it can be seen by inspection at which levels the air will yield a path curve of no interest. In Fig. 164, for example, the saturation adiabatics through all the wet-bulb readings except W_3 lie, for the most part, below the environment curve, and the path curves all yield large

negative areas, with, in some cases, only insignificant positive areas in the highest levels. It is concluded that the distribution of upper-air temperature is unfavourable for convection.

The forecaster using this tephigram cannot, however, assume that subsequent convection is unlikely, for allowance must be made for possible trigger action. In particular, since the observations are taken in the early morning, the temperature at the surface is likely to rise considerably. The upper air will be less subject to a rise of temperature, and it may be assumed that the temperature above the 900-mb. level will undergo little change. In the lowest layers the effect of heating at the surface will be to establish a lapse-rate approximating to the D.A.L.R.

The temperature to which the air at the surface must rise in order to bring about free convection may be estimated as follows. Assume that the temperature of the surface air rises, but that its dew-point does not change. This is not strictly true, since some water is certain to be evaporated into the air, thus raising its dew-point, but it is a convenient first step. Now, since the surface air is saturated, T, W and the dew-point D are the same. Through D draw the vapour content line to meet the environment curve in C. From C draw the dry adiabatic CT' to cut the surface isobar in T', the corresponding temperature being 86° F. If, now, the temperature of the air at the surface be raised to 86° F., the air will rise freely along the dry adiabatic $T'C$, and becoming saturated at C it will continue to rise freely along the saturation adiabatic CS', which lies entirely above the environment curve. The temperature T' is thus determined as the lowest, which will yield an entirely positive area.

There are, however, two other factors to be taken into account in estimating the probability of convection. The first is a possible rise in the dew-point of the surface air by the evaporation of moisture from vegetation and water surfaces. Under given conditions it is difficult to estimate the rise in the dew-point and a value must be assumed. In the case under consideration, assume that the dew-point rises to 64·3° F., corresponding to an increase of water-vapour content from 11·5 to 13 g./kg. To avoid confusion of the diagram, this and the succeeding discussions are illustrated in Fig. 165, in which the environment curve of Fig. 164 has been repeated. The new dew-point is represented by D' and the corresponding vapour-content line is $D'C'$, cutting the environment curve in C'. As in the previous case, drawing the dry adiabatic $C'T''$ to cut the surface pressure line in T'' gives 84° F., as the temperature to which the surface air must be raised to give free convection under the assumed conditions. The path curve yields the positive area, $C'S''T_{10}$, which is greater than the positive area $CS'T_{10}$ of the former case. Despite the smaller increase of surface temperature, the increase of water-vapour content provides more energy for release on convection, increasing the vigour of the convection and the intensity of the associated weather phenomena.

The other factor is the effect of momentum acquired by the rising

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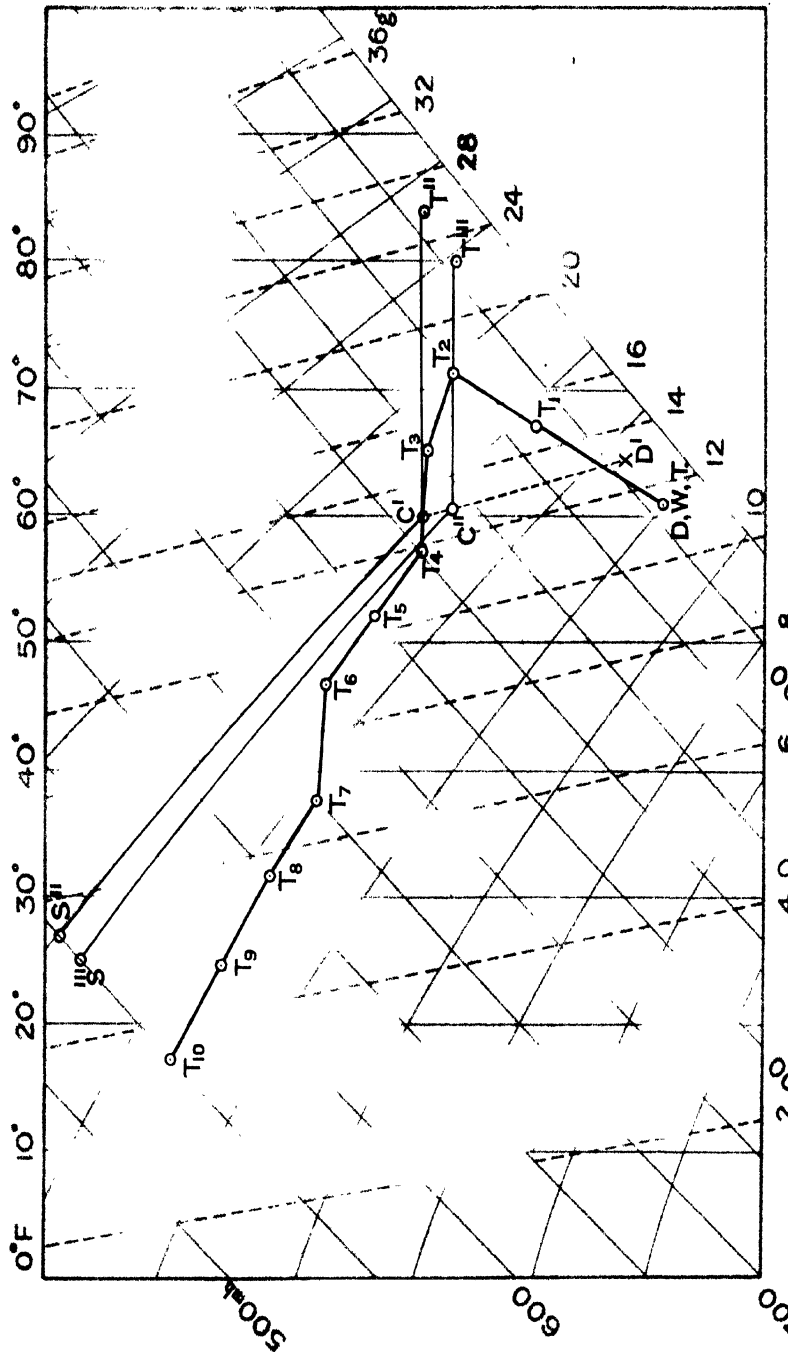

FIG. 165. TEPHIGRAM FOR HENLOW 0600 G.M.T. 25 JUNE 1935.

air in enabling it to penetrate stable layers associated with negative areas. To illustrate this, assume that the surface temperature is raised only to 80° F. (T'''), with the dew-point D' as in the preceding case. Construct the path curve $T'''C''S'''$ in the usual way to cut the environment curve in (by coincidence) T_2 and T_4, to give the negative area $C''T_4T_3T_2$ and the positive area $T_4S'''T_{10}T_9..T_4$. It would appear that the air could not rise through the layer corresponding to the negative area, since, while on the path curve $T_2C''T_4$, it would be colder than its surroundings. But in rising along $T'''T_2$ it would acquire a certain speed of ascent, the momentum due to which might carry it over that part of the ascent associated with the negative area, enabling it to reach the level of T_4, above which it would ascend freely. What allowance should be made for momentum in enabling air to penetrate the area associated with a negative area, cannot easily be estimated, but it is evident that some allowance must be made. Moreover, a small negative area will be more easily penetrated than a large one.

It appears then, that having regard to the possible rise of dew-point and the penetration of stable layers by virtue of acquired momentum, a rise of surface temperature to about 80° F. would bring about spontaneous convection of the surface air. In forecasting thunderstorms or other convection phenomena, therefore, the question to be considered is whether, having regard to the time of year and the state of the weather, the surface air temperature is likely to rise during the day to 80° F.

On the day in question the sky was cloudless in the early morning, and the time of year being June, a rise of temperature from 61° F. at 0600 to 80° F. by midday was considered probable, and thunderstorms were forecast. Actually thunderstorms of a most violent type broke out shortly after noon over a large area of southern England.

It is of interest to consider the tephigram constructed from the observations made at 1200 G.M.T. on the same day, when the thunderstorm outbreak was taking place. The observations are set out in the following table :—

Pressure.	Dry bulb.	Wet bulb.
mb.	° F.	° F.
1,005	80·8	72·0
970	73·5	68·0
950	71·5	66·5
900	64·0	62·5
850	57·0	56·2
800	50·5	48·5
750	46·0	40·6
700	38·8	34·0
650	32·0	31·0
600	23·7	21·2
550	15·0	12·5
500	9·0	7·5

Based on these observations the environment curve $T\,T_1..T_{11}$ has been plotted in Fig. 166 together with the wet-bulb reading at the surface, *W*. On comparison with the morning tephigram there is firstly to be noted a slight fall of temperature in the upper levels, but this is of no immediate interest. The surface temperature has risen to 80·8° F., which was estimated as sufficiently high to lead to convection. On applying Normand's construction to the points *T*, *W*, the value 67·5° F. is obtained for the dew-point *D*, which, as anticipated, has risen since the morning. Finally, the path curve for the surface air, *TCS*, yields a large positive area, and the condition for thunderstorms is therefore fulfilled.

Forecasting of Individual Elements.

Surface Winds.—The surface wind depends primarily upon the pressure distribution, and to forecast it, it is necessary to form a mental picture of the isobars during the period for which the forecast applies. The following rules may be used to relate the surface winds to the pressure distribution :—

(1) Over the sea the speed of the surface wind is approximately two-thirds that of the geostrophic wind, and its direction deviates by 10° to 20° from the direction of the isobars.

(2) Over the land the speed of the surface wind is approximately one-third that of the geostrophic wind, and its direction deviates by about 30° from the direction of the isobars.

In coastal waters, the sheltering effect of the land must be taken into consideration when applying these rules, an off-shore wind being taken as a land wind and an on-shore wind as a sea wind.

For forecasts covering very short periods the surface winds as actually reported, may be used.

On account of the effect of curvature of the isobars, the wind on the outer boundary of an anticyclone tends to be somewhat stronger than that estimated from the geostrophic wind. On the other hand, in a depression, the wind tends to be somewhat lighter than that estimated from the geostrophic wind. On an average, the difference amounts to about 5 knots.

When the lapse-rate is large, the wind is often squally and comparatively strong relative to the geostrophic wind. On the other hand with a small lapse-rate there is little gustiness, and the speed of the surface wind is light in comparison with the geostrophic wind.

In the early morning, over land areas, after a night during which radiation has been marked (*see* Chapter VI), the surface wind is often light and variable, bearing no relation to the pressure distribution. With the increase in turbulence during the day the surface winds gradually conform to the pressure gradient, in the manner expressed

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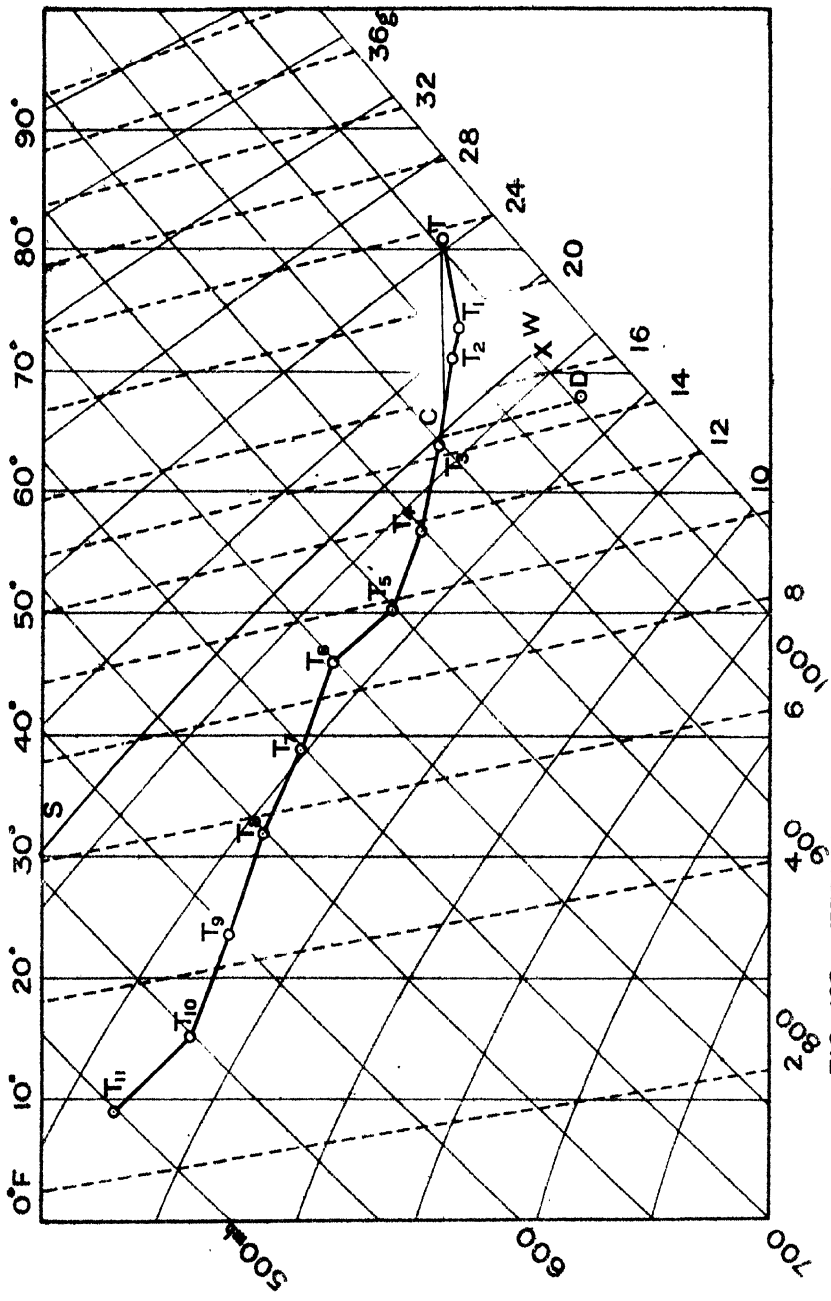


FIG. 166. TEPHIGRAM FOR HENLOW 1200 G.M.T. 25 JUNE 1935.

by the preceding rules, and allowance should be made for this effect when forecasts are issued in the early morning.

In low latitudes the geostrophic control is small, and the winds may show a marked departure from conformity with the isobars, tending rather to follow the slope of the country and the run of the coast, and to be generally affected by topographical features. Land and sea breezes and katabatic effects are particularly prominent.

The wind tends to accelerate past headlands and in straits.

In advance of a warm front or occlusion, the wind backs gradually, and sometimes also, but to a less extent, in advance of a cold front. It veers as the front passes and the change is comparatively sudden. The above applies to the northern hemisphere; in the southern hemisphere "backing" should be substituted for "veering" and *vice versa*.

Near the coast, the daytime sea breeze should be allowed for, if temperature is likely to become high over the land, and provided that the general wind does not mask its effect.

Winds of force 8 and above, *i.e.*, gales, should be forecast whenever it is anticipated that the pressure gradient will steepen so as to give a geostrophic wind in excess of 50 knots. The increase in the pressure gradient may occur either as the result of the deepening or the approach of a depression, or both in combination.

The deepening of a depression should be anticipated if the depression has a pronounced warm sector; if deepening is in progress, it is indicated by large negative barometric tendencies. Large negative tendencies also denote the approach of low-pressure systems.

In general, the rules given above for the forecasting of surface winds are applicable to the forecasting of gales, but consideration should be given primarily to the barometric tendencies. Gales do occur in certain localities when the general pressure gradient reinforces a local katabatic effect, as, for example, the mistral of the Gulf of Lyons and the bora of the Adriatic. Reference should be made to the appropriate handbooks for an account of the circumstances under which these local winds occur.

Gales subside gradually as the depression causing them fills up after occlusion, or moves away.

Upper Winds.—The wind at 2,000 feet is most conveniently determined from the M.S.L. isobars; its direction is parallel to that of the isobars, and its speed approximately equal to that of the geostrophic wind.

The effect of cyclonic or anticyclonic curvature is similar to that in the case of surface winds, and experience suggests that with cyclonic curvature, the geostrophic wind should be somewhat reduced to give the wind at 2,000 feet. The following table has been found suitable for use in the British Isles and may be taken as generally applicable to middle latitudes; it shows the amount

by which the geostrophic wind should be reduced in order to give the approximate wind at 2,000 feet :—

Geostrophic wind. Knots.	Reduction. Knots.
0–20	0
20–30	5
30–40	10
40–60	15
60–80	20

Geostrophic winds in excess of 80 knots occur as a rule only in central regions of a depression where the cyclostrophic component is large, and the wind at 2,000 feet is often much less than the geostrophic wind.

With anticyclonic curvature, the wind at 2,000 feet may be taken as equal to the geostrophic wind.

For very short period forecasts, recent pilot-balloon ascents give good approximations to the upper winds, except when a front has passed since the ascent was made.

At a front, the wind in the lowest few thousand feet changes in a manner similar to the surface wind.

For winds above 2,000 feet, the results of recent pilot-balloon ascents should be used ; if no ascents are available, the rules given in Chapter VIII regarding the variation of wind with height in the free atmosphere and its relation to the temperature distribution, may be applied.

Precipitation.—It has been seen in Chapter XI that there are three main types of rain, viz., convectional, frontal and orographic.

Convectional rainfall occurs with a large lapse-rate and high moisture content in the lowest layers of the atmosphere. In temperate regions it occurs mainly in polar air, *e.g.*, in the rear of a cold front, chiefly in the form of showers from detached cumulonimbus clouds, the individual showers being separated by fair or even bright intervals. Over the sea this type of rain is most prevalent in winter, when the relative warmth of the water increases the lapse-rate in the cold air moving over it. In summer, the sea is relatively cool and it is only over the land that the heating effect is sufficient to bring about convection. Owing to the diurnal variation of temperature, convectional rainfall over land in summer occurs chiefly in the afternoon ; on the other hand, over the sea, convectional rainfall may occur at any time of the day. In the tropics, the rainfall is mainly of the convectional type, and heavy rain, often accompanied by severe thunderstorms in the afternoon, is of almost daily occurrence. In this connexion it should be noted that the occurrence of rainfall at a particular place is largely dependent on wind direction, *e.g.*, an on-shore wind acquires abundant moisture during its sea passage and on reaching the land, deposits this moisture as orographic or convectional rain. In general, the

more vigorous the convection, the heavier is the rain associated with it, and in extreme cases, heavy hail and violent thunderstorms occur ; the probability of convectional precipitation and its intensity are estimated with the tephigram in the manner described elsewhere in this chapter. In summer, strong localized heating coupled with convergence of air due to incurvature across the isobars, often gives rise to heavy precipitation in shallow depressions moving slowly over land areas.

Much of the rainfall of temperate latitudes is frontal in origin ; in advance of a warm front or occlusion there is usually a belt of continuous rain up to 200 miles in width, but the rainfall is generally only light or moderate in intensity. All rain which precedes warm fronts and occlusions falls from altostratus or nimbostratus cloud. It commences when the base of the cloud is about 15,000 feet, but often evaporates before reaching the surface, until the cloud base has been lowered to 10,000 feet ; hence when the air ahead of the front or occlusion is very dry, the rainfall will be less. Similarly, if the upper-air temperature in advance of a warm front or occlusion is high, the rainfall will be slight, for there is then little interaction between the pre-frontal and post-frontal air masses, owing to the small difference of temperature. In this case the tephigram for the air in advance of the front or occlusion is of the stable type described on page 439.

Normally, in a warm sector, occasional slight rain or drizzle occurs, but it may become continuous near the centre of the depression and especially on hills and elevated coasts. If the depression is deepening rapidly, the rainfall in the warm sector is often continuous and widespread.

In advance of a cold front the rainfall is sometimes continuous for about an hour, while a shorter period of heavy rain usually occurs at the passage of the cold front itself. In the rear of the cold front, the rain generally clears rapidly, usually within about half an hour, but there may be renewal of rain for a short period if the cold front is followed by a secondary cold front (*see* Chapter XVIII). In the rear of a back-bent occlusion there is frequently a considerable period of continuous rain, similar to that which occurs in advance of a warm front. Rainfall is often heavy in deepening depressions and particularly in deepening secondaries.

Orographic rainfall should be forecast whenever a saturated or nearly saturated, air-stream meets a range of hills or elevated coast. In such a case the rainfall will continue so long as the supply of moist air lasts. Much of the rainfall of the monsoons of India and China is orographic. If the lapse-rate is sufficiently large, the orographic uplift often supplies the trigger action necessary to start convection, resulting in the formation of coastal thunderstorms (*see* Chapter XII). If, on the other hand, the lapse-rate is small, convection does not follow orographic uplift, which by itself is sufficient to cause only drizzle ; an example of this is furnished

by the drizzle which occurs on the coasts of south and south-west England in a stream of tropical air which has become stabilized in the course of its northward passage over the sea.

Orographic influence may be such as to delay the passage of a warm front, resulting in the prolongation of the frontal rain on the windward side of a range of hills.

The passage of air over a colder sea surface often leads to the formation of a continuous sheet of stratocumulus cloud, as a result of turbulent upward transport of water vapour, and occasionally the process is pressed so far as to lead to the occurrence of drizzle, even over the open sea or flat country. Such precipitation is common in eastern England with easterly winds in spring and early summer.

Drizzle often occurs on the outer boundary of an anticyclone; it is impossible to forecast precipitation of this type on physical principles alone, so that the synoptic chart should be carefully watched for its appearance and the forecast framed accordingly.

Rain often occurs also for no obvious reason in a col and again the chart should be carefully watched for its appearance.

The meteorological conditions associated with snow or sleet are, in general, similar to those associated with rain; the principal distinguishing factor is the temperature. As an approximate rule it may be assumed that if the surface temperature is below 37° F., snow or sleet will occur in any situation where rain would otherwise be expected. Snow, however, is possible with the temperature above 37° F., but in such cases the wet-bulb temperature is usually low. It is more likely to occur, at a given temperature, in depressions that have developed entirely in polar air than in normal depressions. It is comparatively frequent in the arctic air associated with secondary cold fronts moving southwards over the British Isles. If the air temperature is below 32° F. the precipitation usually takes the form of dry snow; if, however, there is a pronounced inversion of temperature aloft and the precipitation falls from above the inversion it will be in the form of rain provided that the temperature at the surface does not fall too much below the freezing point. Precipitation which begins as heavy rain, may, owing to evaporation, cool the air through which it falls to a temperature below the critical value and subsequently turn to snow.

Cloud.—A statement regarding the amount and height of cloud is required in all forecasts for aviation. The cloud type need not be specified, but should be considered in preparing the forecast, since it is often useful in estimating the amount and height.

The usual pre-frontal cloud sequence, viz., cirrus, cirrostratus, altostratus and nimbostratus should be associated with approaching warm fronts and occlusions. With the approach of the front an increase in amount to ten-tenths must be allowed for, and unless there is definite evidence to the contrary, a lowering to 300 feet of the cloud base at the front itself, with consequent envelopment

of high ground. In the warm sector of a depression there is often ten-tenths of low stratus or nimbostratus, especially in winter; in summer, however, the cloud in the warm sector often breaks over land. At a cold front, the sky is generally completely covered with cloud at about 1,000 feet, or slightly lower; the cloud sometimes takes the form of cumulonimbus, although this cannot be determined by viewing from below, since only the ragged base is visible. In the rear of a cold front, the cloud rapidly lifts and breaks, while the warm-front cloud sequence is often experienced in the reverse order within less than an hour. With feeble fronts, cloud sheets tend to break up, the normal cirrostratus, altostratus and nimbostratus assuming the forms of cirrocumulus, altocumulus and stratocumulus respectively. Cumulus and cumulonimbus clouds generally occur in the rear of a cold front, and are often persistent near to the centre of the depression, especially when the barometer fails to rise rapidly after the front has passed. At a greater distance from the cold front and the centre of the depression, cumulus and cumulonimbus diminish in amount, probably owing to subsidence, especially when the barometer is rising rapidly. The forecaster may therefore sometimes be misled when, as a result of the preponderance of cumulonimbus in the early morning due to the proximity of the front, he forecasts an increase of cloud and frequent showers during the day as a result of increased convection. On the other hand, when there is no front in close proximity, the early appearance of cumulus cloud should be regarded as an indication of the persistence of convection cloud during the day.

In anticyclonic conditions in summer, the sky is usually cloudless, apart from a few small cumulus clouds in the afternoon. Occasionally also, the sky may become overcast with stratus or stratocumulus during the night, especially if the wind has had a long passage over the sea. In winter, anticyclones are frequently accompanied by skies overcast with stratus or stratocumulus cloud, and it is not possible to forecast with any degree of certainty whether such a cloud sheet is likely to form or break.

In general, if the lapse-rate is large, clouds are mainly cumuliform, and, over land, increase in amount during the day, disappearing at night. If, however, the lapse-rate is small, clouds are of the stratus type, and there is a tendency for them to form during the night and to disappear by day. Over the sea, since there is no appreciable diurnal variation of temperature, clouds of the cumulus type may occur at all times of the day in a polar air current whose surface temperature is lower than that of the sea. If, on the other hand, the air is warmer than the sea, the tendency is for stratus clouds to form.

With light winds over the sea, turbulence does not force air high enough to form cloud, while with strong winds, the air is not long enough in contact with the sea to pick up sufficient moisture

to form cloud. The critical value for maximum cloud formation appears to be with a wind speed of 10 to 20 knots.

In summer, owing to the relatively large lapse-rate, most clouds assume a cumulus structure, while in winter, owing to the small lapse-rate, stratiform clouds predominate.

A stream of moist air striking a range of hills or elevated coast-line gives rise to orographic cloud on the windward side; with strong winds the cloud may be carried away to leeward, as, for example, in the case of the levanter experienced at Gibraltar.

The height of cloud should be forecast by using a table of average heights for the various types, such as the table on page 50. In the case of clouds formed by convection, *i.e.*, Cu. and Cunb., the base is approximately at the condensation level for air ascending from the surface, as determined by the tephigram.

In the absence of upper-air temperatures, the cloud height in feet is given by the following approximate formulæ :—

500 $(T - W)$	when the surface air temperature is about			40° F.
400 $(T - W)$	,,	,,	,,	50° F.
400 $(T - W)$	,,	,,	,,	60° F.

where T and W are the dry- and wet-bulb temperatures, respectively, of the surface air. During showers the cloud base becomes considerably lower.

The track of the wind before it arrives at a place must be studied before forecasting cloud amount. Thus, in south-east England, north-easterly winds passing over the North Sea are accompanied by a great deal of cloud, which will decrease if the wind veers, for it then has a shorter sea passage from the continent. In the northern Mediterranean, a southerly wind may be accompanied by much low cloud, while the same wind off the north African coast will give a cloudless sky.

Fog and Visibility.—(1) *Land fog.*—Land fog is important from the seaman's point of view because it may drift out and envelop coastal waters. Chapter IX should be consulted regarding the time of day, season and locality, favourable to the formation of land fog. It is formed in the central parts of anticyclones and wedges which are usually accompanied by clear skies and light winds. It may also form in a col, owing to the mixing of air streams of widely differing origins. Generally, in the winter half-year, there is a risk of fog formation whenever light winds occur in conjunction with clear skies at night. An empirical diagram for the forecasting of fog at land stations in the British Isles has been prepared by G. I. Taylor and is shown in Fig. 167. The probability of fog is estimated from the observations of dry- and wet-bulb temperatures taken during the evening, when the conditions regarding cloudiness and wind are favourable to radiation. Along the horizontal axis is plotted the dry-bulb temperature; along the vertical axis, the difference between the dry and wet bulb observed

at 2000. The dry bulb and depression of the wet bulb can then be plotted as a point on the diagram. It is found that the points thus plotted for nights on which fog subsequently developed are grouped in the lower right-hand half of the diagram; those plotted for nights on which no fog developed, in the upper left-hand half of the diagram. The two sets of points are separated by a straight line. The diagram may be used only when the sky is less than

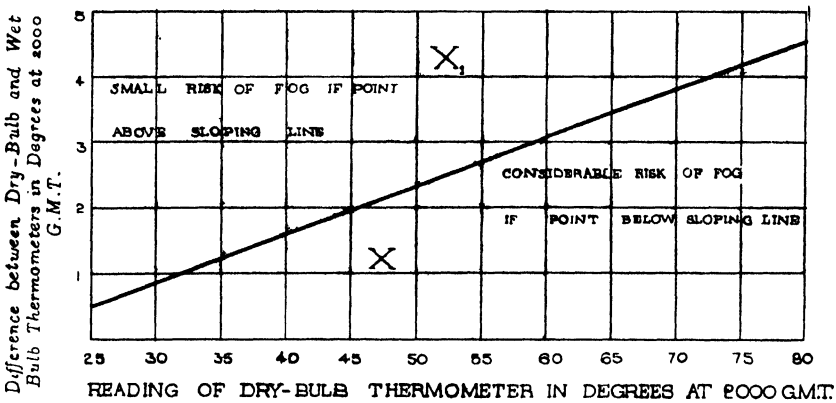


FIG. 167.—FOG FORECAST DIAGRAM

half clouded and the wind speed less than 7 knots at 2000. The temperature and depression of the wet bulb are plotted; if the point lies above the line, fog is improbable, while if it falls below the line, there is considerable risk of fog.

If conditions at 2000 are calm, fog or mist is likely whatever the thermometer readings are.

(2) *Sea fog.*—Reference should be made to Chapter IX for an account of the formation of sea fog, and the seasons and localities in which it is most frequent.

Sea fog should be forecast whenever the dew-point is likely to be above the temperature of the sea surface. In this respect the forecaster is largely guided by the latest synoptic chart, from which the origin of the air-stream may be determined and the dew-point estimated.

Fog of this type occurs chiefly in tropical air blowing towards higher latitudes, and, in particular, in the warm sector of a depression. As a rule it does not clear until the arrival of a cold front, although any increase in the sea temperature has a tendency to lift the fog.

It is of the utmost importance that the sea temperature should be accurately known, since a small change in its value is sufficient to cause the formation or dispersal of fog. The visibility will generally improve as the ship steams into warmer water, and deteriorate as it approaches colder water; hence, patchy visibility

is experienced in areas such as the North Sea, where numerous shoals force cold water to the surface as the tidal stream runs over them; a change in visibility may be experienced when the tidal stream turns. Rain, by lowering sea-surface temperature, may occasionally cause fog formation.

In general, the best visibilities both on land and sea are experienced when the lapse-rate is large; a small lapse-rate, and particularly an inversion, is often associated with haze, mist or fog. For this reason, visibility is generally good in polar air, but only moderate or poor in tropical air. The immediate past history of the air is, however, a factor equally as important as its origin; a marked deterioration of visibility occurs to leeward of a large town or industrial area, owing to pollution by smoke, *e.g.*, visibility in the Thames estuary is often seriously affected by London smoke with westerly winds. Easterly winds have a similar effect on visibility in the Mersey approaches. If the lapse-rate is small, the deterioration of visibility may be experienced at great distances from the source of pollution, *e.g.*, visibility on Salisbury Plain is frequently reduced to less than 2 miles by smoke from London, carried by an ENE. wind. The larger the lapse-rate, the more rapidly is the smoke dissipated and the smaller the distance at which the pollution is effective. Dry air is often associated with good visibility; on the other hand, since it is also associated with subsidence and a consequent inversion of temperature, it is often hazy. This is particularly the case with summer anticyclones, during the life of which, visibility gradually deteriorates.

Reference should be made to Chapter IX for a description of other types of fog, *e.g.*, warm-front fog and " mixing " fog; since fronts and occlusions are not always accompanied by fog they cannot be used as criteria for forecasting its occurrence. For example, a warm front may, or may not, be associated with fog at a particular place, and in general it is impossible to say which.

The forecasting of the state of sea is partly a matter of seamanship as opposed to meteorology (study of phenomena of the atmosphere) though it is largely dependent upon the wind forecast for the same period and that experienced during the preceding few hours. Other matters affecting the state of sea for a given force of wind are fetch, depth of water, length of time the wind has been blowing (in the case of the stronger winds), and, to a limited extent, tidal stream and current.

The correlative table given on p. 51 shows the state of sea to be expected for a given force of wind in deep water, far from the windward shore, and when the fetch is great; and the following notes are intended as a guide for modifying the forecast when fetch, depth, etc., are not as great as would be necessary for the full development of the waves.

(i) As the wave runs into shallower water, it becomes shorter, higher, and therefore steeper.

(ii) The fetch in miles required to raise waves of the order given in the correlative table varies with the force of the wind, and is probably in the nature of $\frac{W^6}{500}$ where W is the force of the wind, by Beaufort Scale.

(iii) The size of a wave is limited by the depth of water, and the least depth in fathoms in which a wind will cause a sea according to the table is about $\frac{W^3}{10}$.

(iv) The time taken for the wind to raise the sea given in the table depends upon the force of the wind, and the sea and swell already running ; for a wind of force 5 a little more than one hour is necessary from a state of calm, whereas probably 12 hours is necessary for a wind of force 8.

(v) Waves will be higher than expected if superimposed on swell, *i.e.*, temporarily in phase with the swell, or approaching shallow water, or due to over-riding.

If, then, the meteorological officer is required to make a forecast of the state of sea, he should allow for the effect of fetch and duration of wind, and leave the other factors to the recipient's judgment (since they are matters of seamanship and are dependent upon information not readily available to the meteorological officer.)

Swell.—It is impossible to lay down any reliable rules for the forecasting of swell, but the following considerations should be taken into account :—

Gales or strong winds reported or anticipated, will give rise to swell in the open ocean from the direction from which the wind blows. Such gales may occur at a considerable distance from the observing ship, and the swell will reach the ship only if the winds causing it are blowing approximately towards the ship. A heavy swell should be anticipated also when there is a deep depression or tropical storm at a great distance, which may be as much as 1,000 miles from the observing ship, provided that there is an open expanse of sea between. A fetch of 600–900 miles is required for the development of waves to their maximum height, which is about 70 feet. It is clear, therefore, that a synoptic chart covering a wide area, is necessary for the forecasting of swell. For this reason swell forecasts have been omitted from many of the examples at the end of this chapter. If there is no direct evidence of the existence of gales or any of the factors mentioned above, the forecasts for swell should be based on the actual reports of swell from the nearest ships.

Air Temperature.—(1) *Over the sea.*—The diurnal variation of temperature over the sea is small, amounting to not more than about 2° F. Consequently, the temperature depends upon the

origin of the air and its recent past history. After a long fetch over the sea, the air temperature at the surface approximates to that of the sea, except in the case of unstable polar-air currents where vigorous convection leads to the vertical interchange of air; for, in this case, the temperature of air which has recently descended to the surface, although affected by dynamical heating, is lower than that of the surface air which it has displaced. In coastal waters, with off-shore winds, the temperature of the air is largely determined by that of the land from which it has come. In summer it will be warmer, and, in winter, colder than the sea surface. In clear weather there may be a considerable difference of temperature between day and night. Changes of temperature in the open sea are caused by the arrival of new air masses, and therefore take place at the passage of fronts.

(2) *Over the land.*—The same considerations regarding the origin of the air stream and its recent past history, apply to a certain extent in forecasting temperatures over land. They are, however, subject to considerable modification by insolation and nocturnal radiation, the effects of which are to bring about a large diurnal variation of temperature. In the British Isles the average diurnal variation is about 10° F. in winter, and 20° F. in summer. Over the desert areas of the sub-tropical zones the diurnal variation is greater than this, and may be as much as 40° F. The effect of cloud is considerably to diminish the diurnal range.

In general, when the wind sets in from a cold or warm region, the temperature does not reach its minimum or maximum, respectively, for a day or two. The stronger the wind, the more rapidly does the cold spell or heat wave develop. Rain falling through air may cool it, owing to evaporation, to its wet-bulb temperature; hail or snow can cool it still further, possibly to freezing point, through melting. If available, tables of average temperature for particular localities and seasons are useful in forecasting. The forecasting of frost is sometimes important in confined waters; at an inland situation it is likely to occur at night in autumn, winter and spring, if the sky remains clear and the wind light; it occurs also in winter in air currents of arctic, polar or continental origin, except in low latitudes. The latter type of frost is more likely to do damage with frozen pipes, etc., owing to the penetrating effect of the cold wind.

Sea Temperature.—The forecasting of sea temperature should be made primarily by reference to charts of average sea-surface temperatures for different months of the year. In general, after a few days of flat calm, the sea surface temperature may rise a few degrees above the average; cold water from the depths may be forced to rise to the surface by the presence of shoals. In coastal waters the temperature of the sea surface is affected by the rivers flowing into them, while both sea and air temperature are considerably diminished in the neighbourhood of ice.

Situations in which Ice Accretion may be Encountered.*—*High Flights.*—If flights are made at heights where the temperature is below about 12° F. the clouds will be cirrus, altocumulus, altostratus or cumulonimbus. We know that cirrus clouds consist of ice crystals and therefore ice accretion cannot occur in them. Altocumulus and altostratus clouds are known, from the colours they show in the neighbourhood of the sun and moon, often to consist of supercooled water particles and there are records of ice accretion occurring in them. These clouds are, however, thin and cannot contain much water, therefore the ice accretion is not likely to be heavy. Prolonged stay in the clouds might produce appreciable accretions, but there are no records of heavy accumulation in either altostratus or altocumulus clouds.

The case of high flights in cumulonimbus clouds will be considered later.

Flight across a Warm Front.—A warm front (*see* Chapter XVIII) is associated with a very extensive almost horizontal layer of cloud. The base of the cloud is low near the front where there is usually steady rain in summer and snow or sleet in winter. Ahead of the front the base of the cloud becomes higher, and the whole cloud sheet thins out slowly into a layer of altostratus or cirrostratus cloud as one recedes from the front.

The main cloud sheet near the front is very thick, extending from a few hundred feet above the surface to a height of at least five or six thousand feet and often to a height well above ten thousand feet. The cloud sheet near the front, as seen from the ground, has apparently a level base and is dark in appearance owing to the great thickness of the cloud ; it is called nimbostratus. At all times of the year the top of this prefrontal cloud layer is in the region of the atmosphere where the temperature is below the freezing point. Hence there is a layer in the cloud in which ice accretion may, but does not always, occur.

The circumstances in which ice accretion occurs at fronts are not known ; but experience has shown that ice accretion is most likely where the front is " active," *i.e.*, where precipitation is falling.

As it is at warm fronts that warm air mounts over cold air the conditions are particularly favourable there for the formation of glazed frost. An aircraft flying through the precipitation at a warm front in the winter with surface temperatures at or near the freezing point may encounter very dangerous conditions.

High land near to a front is particularly dangerous for two reasons. In the first place, the base of the cloud may be lower than the hills, and, secondly, the air is forced to rise and so the accumulation of liquid precipitation is increased, thus making ice accretion, if it

* From a paper " Ice accretion on Aircraft " by G. C. Simpson, K.C.B., F.R.S.

occurs, much more rapid and heavy. Some of the worst disasters due to ice accretion have occurred in the cloud system accompanying a front where it passes over high ground.

In the summer there is little danger in crossing a warm front because the layer with temperatures below the freezing point is sufficiently high to leave plenty of room below it for navigation.

In the winter also there is usually room to navigate an aircraft under the cloud at a warm front, or to keep within the region where the temperature is higher than the freezing point if the cloud has to be entered. But if the temperature at the surface is below 40° F., the possibility of being forced to fly in a cloud in which ice accretion will occur and the especial danger of high ground must always be borne in mind.

The conditions at an occluded front are so similar to those at a warm front that it is not necessary to treat them separately. On the whole the main cloud layer at an occluded front is higher than near a warm front; but the weather there is frequently very "dirty" and rapid ice accretion may occur.

Flight across a Cold Front.—At a cold front (*see* Chapter XVIII) there is not usually an extensive unbroken layer of clouds. The clouds are generally of a broken nature and if they have to be entered they are soon left again. The clouds are, however, of the convection type with strong ascending currents which will carry cloud particles in the liquid state far into regions below the freezing point. Also the clouds originating in polar air behind the cold front are on the average at a lower temperature, so that the ice-accretion layer is nearer the surface than at the warm front.

Flight through Cumulus and Cumulonimbus Clouds.—Owing to the frequency with which cumulus clouds occur in this country, ice accretion is more frequently met with in these clouds than in the nimbostratus clouds associated with warm fronts and occlusions. When cumulus clouds develop into cumulonimbus clouds (the thunder cloud), the ice accretion becomes much more serious, extends through a greater thickness of cloud and occurs at the lowest temperatures.

The reason for this is not difficult to understand. A cumulus cloud makes visible an ascending current of warm air which is pushing its way upwards through the surrounding cooler air. The air enters the cloud at its base, then rises almost vertically with a speed which may be only small when the cloud first forms but becomes very large as the cumulus cloud develops into a cumulonimbus cloud. When the ascending currents become very violent, as in a thunderstorm, vast quantities of liquid cloud particles are carried far up in the atmosphere to very low temperatures, thus providing the most favourable conditions for heavy ice accretion.

The reports of ice accretion experienced by aircraft in these clouds bear out this conclusion. Out of 36 cases of ice accretion in cumulus

clouds six of them were " heavy " while of 24 cases in cumulonimbus clouds as many as 10 were " heavy " and four occurred at temperatures below 3° F.

Normally cumulus clouds are of limited extent, and there should be little difficulty for an aircraft flying through cumulus clouds and noting ice accumulation to get out of the cloud. Also ice accumulation seldom occurs in a cumulus cloud below 5,000 feet, so that there is plenty of room to get below the ice-forming layer. In cumulonimbus clouds ice accretion occurs still higher. A pilot does not readily enter a cumulonimbus cloud on account of the danger from lightning and violent ascending currents. Ice accretion adds one more to these dangers so that pilots are well advised to keep clear of cumulonimbus clouds if possible.

Flight through a layer of Stratocumulus Cloud.—Ice accretion in this country is met with more frequently in stratocumulus clouds than in all other types of cloud put together. Of the 217 cases reported during 1932–7, 124 of them occurred in stratocumulus clouds. One reason for this is that during the winter this type of cloud is very frequent and often extends in an unbroken layer over the greater part of the country. It is also a low cloud, and therefore in the winter it very frequently occupies that part of the atmosphere where the temperature is just below the freezing point.

Stratocumulus cloud is not formed in an ascending current of air, as is the case with cumulus clouds, nor in the forced continuous rise of warm air over cold air as at a warm front. The stratocumulus cloud layer owes its origin to the local raising of air by the turbulent motion of the wind. When closely examined stratocumulus clouds are found to have a cellular or wave-like pattern due to the air rising in some places and falling in others. There are frequent " holes " in a layer of stratocumulus clouds. As there is no continuous rise the cloud particles cannot be carried up and accumulate, the cloud particles are therefore of no great size and precipitation seldom falls from this type of cloud. Hence it is relatively seldom that aircraft experience heavy ice accretion within stratocumulus clouds unless they remain a long time within them. The base of the clouds is usually above 2,000 feet and the tops below 5,000 feet so that aircraft need seldom fly for any length of time within the cloud layer itself.

Practical Examples of Forecasts.

Forecast for Coastal Waters of the British Isles for the period 0900 to 2100 G.M.T. on the 29th August, 1936.—Fig. 121 (see footnote on page 339), shows the synoptic situation over the British Isles at 0700 G.M.T. on the 29th August, 1936. A large anticyclone covers the southern part of the British Isles, its centre being situated over south-east England. The barometric tendencies near and around the centre are positive generally, indicating that the high pressure is building up. The straight isobars running from south-west to north-east across Scotland, together with

negative barometric tendencies at stations 110, 111, and in western Norway, point to the possibility of small secondary disturbances affecting this region, particularly in the north and west of Scotland. For the purpose of the forecast the area is divided up into 3 parts :—

Area (1).—Off north and west Scotland and Northern Ireland.
Area (2).—Off eastern Scotland.
Area (3).—The remaining coastal waters of the British Isles.

In Area (1), the measured value of the geostrophic wind is approximately 30 knots, which gives 20 knots as the probable surface wind over the sea. Cloudy conditions, with occasional slight precipitation, are expected to be associated with the small secondaries crossing in a north-easterly direction. There are no observations of visibility available, but the probability is that the visibility will only be moderate in the warm south-westerly current. The forecast of the state of the sea should be based on the anticipated wind in the usual way.

In Area (2), the measured speed of the geostrophic wind is 15 knots, indicating a light surface wind, which is confirmed by observations at stations 112 and 124. It is not anticipated that the small secondaries will extend their influence sufficiently to cause precipitation in this area, cloud only being expected. With regard to visibility, the same conditions hold as for Area (1).

Near the centre of the anticyclone, the usual light winds and fine weather are anticipated. Over the open sea, visibility will be good, owing to the dryness of the air, but coastal waters may be affected by drifting radiation fog formed by night over the land ; for this reason, local coastal mist should be forecast.

The anticyclone shows no sign of breaking up, neither do the winds and tendencies indicate the approach of any well-marked depression to north-western districts ; it is therefore anticipated that similar conditions will continue.

The forecast should therefore be issued on the following lines :—

Inference.—A large anticyclone covering the southern part of the British Isles is increasing in intensity, while small disturbances are likely to move north-east between Scotland and Iceland.

Forecast.—Area (1). Moderate or fresh south-westerly wind ; cloudy ; slight rain or drizzle at times ; visibility moderate ; sea slight or moderate ; rather warm.
Area (2). Light or moderate south-westerly wind ; cloudy ; visibility moderate ; sea slight ; rather warm.
Area (3). Light variable wind ; weather fine ; visibility mainly good, local mist near coast ; sea calm ; rather warm.

Further outlook.—Similar conditions continuing.

Forecast for Area (1) the Irish Sea and Area (2) the Southern North Sea, for the period 0900 to 2100 G.M.T. on the 26th September, 1935.—Fig. 122 (see footnote on page 339), shows the synoptic situation over the British Isles at 0700 G.M.T. on the 26th September, 1935. A wedge of high pressure covers England and eastern Scotland.

The barometric tendencies are positive on the eastern side of the wedge and negative on the western side, indicating an easterly movement, while the large negative tendencies in western Scotland and Ireland indicate the approach of a new depression. In Area (1) the pressure gradient for southerly winds will probably steepen, and although no strong winds or gales are reported, experience shows that these are likely to develop in western districts of the British Isles with the advance of the depression. The rain, which has already commenced in southern and western Ireland, is expected to spread east and north. The high temperatures at stations 101 and 102 indicate the northerly movement of warm air, as a result of which visibility is expected to be moderate.

With the eastward movement of the wedge, the wind in Area (2) will back, becoming light in the centre of the wedge and finally south-westerly. The geostrophic wind measured between the isobars of 1,016 and 1,020 mb. on the west side of the wedge is 20 knots, giving a surface wind over the sea of about 12 knots, *i.e.*, a moderate wind, to be expected towards the end of the period. It is possible also that the cloud and rain associated with the depression will reach the area by evening, but throughout most of the period the weather will be that appropriate to the wedge, viz., fine with good visibility. Temperature will rise during the day on account of the passage of the air over the heated land.

On the information available, it is not possible to give a definite further outlook; the approaching depression does, however, indicate that unsettled weather, with westerly or south-westerly winds, is probable during the next 36 hours.

The forecast should therefore be issued on the following lines:—

Inference.—A wedge of high pressure covering England and eastern Scotland is moving east, and is being followed by a large and deep depression to the north-west of Ireland.

Forecast.—Area (1). Fresh southerly or south-easterly wind increasing to strong or gale in the afternoon; cloudy, rain later; visibility moderate; sea slight, becoming rather rough; mild.

Area (2). Light north-westerly wind, backing to south-west in the late afternoon and increasing to moderate; fine at first, cloud increasing with some rain this evening; visibility good; sea calm.

Further outlook.—Strong south-westerly to westerly winds, with rain at times.

Forecast for Sea Area within 50 miles radius of Portland for the period 0900 to 2100 G.M.T. on the 18th June, 1935.—Fig. 137 shows the synoptic situation over the British Isles and western European continent at 0700 G.M.T. on the 18th June, 1935. A warm front, associated with a depression over the Irish Sea, is shown running east-south-east across England and thence southward through Belgium and France.

Since 1300 G.M.T. on the previous day (*see* Fig. 138) the centre of the depression has moved slowly east-north-east, while the lowest pressure reported, has risen from 1,002 to 1,005 mb. Barometric tendencies are generally small and positive to the north, south and south-east of the centre; only immediately north of the front are there negative tendencies, and these are attributed to frontal movement. It is therefore assumed that the depression is filling up. There is no reason to anticipate any change in the weather in the western Channel, and so the forecast should be based on the general conditions in that area. The average geostrophic wind is about 30 knots; two-thirds of this value for the surface wind over the sea gives a wind-speed of 20 knots, *i.e.*, a fresh wind. The direction given by the normal angle of incurvature is about 250°, *i.e.*, west-south-west. Intermittent rain or drizzle and fog patches are anticipated in the warm sector, and are confirmed by the observations at stations 157 and 168.

The rising barometer and fall of dew-point at 139 shows that a cold front has passed that station, but it does not appear to have reached stations 157 and 158, so that it must trail back in a westerly direction nearly parallel to the isobars south of Ireland. It is therefore unlikely to reach Portland during the period of the forecast, but its passage should be allowed for in the further outlook. It will be accompanied by a short period of rain, a veer of wind to west, and a general improvement in visibility.

The forecast should therefore be issued on the following lines:—

Inference.—A shallow depression centred over the Irish Sea is moving slowly east-north-east and filling up; a warm front extending from North Wales to Belgium is moving very slowly north.

Forecast.—Moderate to fresh west-south-westerly wind; cloudy with occasional rain or drizzle; extensive fog patches; sea moderate; little change of temperature.

Further outlook.—Wind becoming westerly and fog clearing, after a short period of rain towards midnight.

Aviation Route Forecast from Firth of Forth to Spithead (direct route), commencing 0900 on the 18th June; duration of Flight 3½ hours.—Referring to Fig. 137, the movement of the front as measured by the geostrophic wind perpendicular to it between the isobars of 1,006 and 1,008 mb., is so slow that the front may be regarded as stationary. The aircraft will therefore pass through it in the position shown on the chart, *i.e.*, in the vicinity of Birming-

ham (station 133). The measured speed of the geostrophic wind is 25 to 30 knots over the whole route, and one-third of this value gives a surface wind of about 10 knots over the land, throughout the flight, and in accordance with the rule given on page 452, gives an upper wind of 20 to 25 knots. The wind directions are obtained from the isobars, that of the surface wind by assuming an incurvature of 30° (over land), that of the upper wind being assumed parallel to the isobars. The pre-frontal rain extends as far north as station 120, and therefore covers the whole of the route north of Birmingham. Near to the front, at stations 128, 132 and 135, the precipitation is heavy. South of the front, the usual slight rain or drizzle of the warm sector is anticipated. Cloud amounts and heights, and visibilities, should be based on the actual reports, allowing for the usual reduction of cloud height near to the front, and deterioration of visibility due to the smoke in industrial areas.

The forecast should therefore be issued on the following lines :—

Wind at surface.—060° in Scotland, veering to 100° to 120° over northern England, veering again to 210° from Birmingham onwards ; 10 knots throughout.

Wind at 2,000 *feet.*—090° in Scotland, 120° to 140° in northern England, 240° to 260° from Birmingham onwards ; 20 to 25 knots throughout.

Weather.—Continuous rain or drizzle in Scotland and northern England, heavy locally in the Midlands, occasional drizzle or slight rain south of Birmingham.

Cloud.—Mainly ten-tenths, 1,000 to 1,500 feet in Scotland and north of England, lowering to 300 feet over northern Midlands, lifting to 600 to 800 feet with slight breaks south of Birmingham.

Visibility.—2 to 6 miles at first, deteriorating to 1 mile in northern Midlands, and probably to 1,000 yards near large towns, improving to 2 to 6 miles south of Birmingham.

Forecast for the Bristol Channel for the period 1500 to midnight G.M.T. on the 17th June, 1935.—Fig. 138 shows the synoptic situation over the British Isles at 1300 G.M.T. on the 17th June, 1935. A depression is situated off south-western Ireland and an associated warm front extends from its centre to north-west Spain.

The movements of the depression and of the warm front are given by the direction of the isobars in the warm sector. The measured value of the geostrophic wind in the rear of that portion of the front intercepted between the isobars of 1,006 and 1,010 mb. and perpendicular to it, is approximately 25 knots. Reducing this value by the customary 10 knots gives 15 knots as the speed of movement of the warm front. At the end of the period, *i.e.*, 11 hours after the time of the chart, the front will therefore have advanced 165 miles, and measurement gives its position in the Bristol Channel as approximately along the meridian of 5° W. The

forecast is therefore divided into 2 portions, viz. :—Area (1), east of 5° W., which will be in the pre-frontal zone throughout the period, and Area (2), west of 5° W., which the front will cross towards the end of the period.

In Area (1), the wind will back gradually to the south-east and will freshen as the front approaches; the estimated wind force should be based on the observations at stations 101 and 158, which are close to the front. Rain, which is continuous in the pre-frontal zone, is expected to spread rapidly over the area; it has already been reported at station 160 which is 180 miles ahead of the front. The normal deterioration of visibility in the rain in advance of the warm front is anticipated; visibility 5 is reported from stations 158 and 211. The forecast for the state of the sea should be based on the rule given on page 409. Little change of temperature is expected since insolation will be cut off by the continuous sheet of cloud, and, moreover, temperatures reported from surrounding stations are more or less uniform.

In Area (2) conditions up to the time of arrival of the warm front will be similar to those for Area (1), except that owing to the closer proximity of the front, the rain is expected to be continuous throughout the period, and the wind will have backed and freshened by 1500. On the arrival of the front shortly before midnight, the wind will veer to the south-west and the rain should become intermittent or possibly cease altogether. Some sea fog is anticipated with the arrival of the warm air, the time of year being that most favourable for sea fog off the south-west coasts of the British Isles. The rise of temperature across the front amounting to only 3° F. between station 101 and *Ascania* is regarded as insignificant; the air temperature is therefore expected to approximate to that of the sea over which it has recently moved, as the wind is comparatively light.

There is no evidence of a cold front immediately following the warm front; it is therefore assumed that conditions appropriate to the warm sector will spread to all parts of the area concerned, and this should be stated in the further outlook. The forecast should therefore be issued on the following lines :—

Inference.—A depression off south-west Ireland and an associated warm front extending from its centre to north-west Spain are moving north-east.

Forecast.—Area (1): Light south-westerly winds backing slowly to south-east and becoming fresh this evening; rain spreading from the south-west during the late afternoon and continuing till midnight; visibility becoming poor with rain; sea increasing to moderate; little change of temperature.

Area (2): Moderate to fresh southerly or south-easterly wind veering to south-west shortly before midnight; overcast with rain, continuous at first but becoming

intermittent with veer of wind; visibility poor, fog patches towards midnight; sea moderate; little change of temperature.

Further Outlook.—Fresh south-westerly winds; intermittent rain; local fog.

Aviation Route Forecast for a flight from Strait of Dover to Plymouth, commencing at 1500 G.M.T. on the 17th June, 1935: duration of flight 2½ hours.—Referring to Fig. 138, consideration of the movement of the warm front shows that on the termination of the flight, *i.e.*, at 1730, the front will be in the vicinity of station 158. The anticipated wind for the first part of the flight should be based on the pressure gradient which is not expected to change during the period of the flight. The measured value of the geostrophic wind is approximately 20 knots; a reduction of this value by one-third gives a surface wind of about 13 knots over the sea, and the assumed angle of incurvature of 20° gives a direction of 220°. It should be noted that winds of force 4 and 5 are reported from stations 166 and 163 respectively; these values are not in conformity with the pressure gradient but the increase in speed is probably due to the sea-breeze effect which sets in as a result of the comparatively high temperature at stations in southern England. At station 168, where no sea breeze is expected, there is no wind and consequently little significance is attached to the reported increases in wind speed at stations 166 and 163. Towards Plymouth, the wind is expected to back and increase owing to the proximity of the front, the extent of the change being indicated by the observations at stations 157 and 158.

The forecast of upper wind should be based on the measured values of the geostrophic wind in the eastern Channel for the first part of the flight, and on that in the vicinity of station 158 for the approach to Plymouth. These measurements give 240°, 20 knots and about 170°, 40 knots respectively and should be reduced in accordance with the rule given on page 452 to obtain the forecast wind speeds.

The pre-frontal rain is spreading north-eastwards and it is expected that it will be encountered by the aircraft in the vicinity of Portland. The cloud amount and height for the first part of the flight should be based on the observations at stations 162, 163 and 166. An increase of cloud amount to ten-tenths and lowering of height to 600 feet are normal in advance of a warm front and should be forecast. These predictions are confirmed by the observations at stations 157 and 160. No change in visibility is expected over south-east England but a gradual deterioration to about 2 miles is probable in the pre-frontal rain, and this fact is confirmed by the reports from stations 158, 160 and 211.

The forecast should therefore be issued on the following lines:—

Wind at surface.—220°, 10 knots at first, backing gradually to 140° and increasing to 20 knots.

Wind at 2,000 feet.—230° to 240°, 15 to 20 knots at first, backing gradually to about 170° and increasing to 30 knots.
Weather.—Continuous rain west of Portland.
Cloud.—About five-tenths at 2,000 feet at first, increasing and lowering to ten-tenths at 600 feet beyond Portland.
Visibility.—Over 10 miles at first, deteriorating to 2 miles in rain beyond Portland.

Forecast for Portsmouth from 1800 to 0600 G.M.T., 29th/30th December, 1936.—Fig. 139 shows the synoptic situation over the British Isles at 1300 G.M.T. on the 29th December, 1936. A cold front extends across western Scotland, Wales and Cornwall. The speed of the cold front is taken as the perpendicular component of the geostrophic wind on that portion of the front intercepted between the isobars of 1,010 mb. and 1,014 mb. This measurement gives approximately 15 knots, so that the cold front is expected to reach Portsmouth, 10 hours after the time of the chart, *i.e.*, at 2300 G.M.T. The observed winds in advance of the front are mainly force 4 or 5, while the average geostrophic wind is about 35 knots. Normally, over the sea, the surface wind speed is about two-thirds of the geostrophic wind; over the land it is about one-third. For southerly winds, Portsmouth has a sea exposure, which would give a surface wind of about 24 knots, *i.e.*, force 6 at Portsmouth, although the wind will be to some extent modified by the sheltering effect of the Isle of Wight. In the forecast, therefore, a wind force of 4 to 5 should be allowed for. At the passage of the front the wind will veer sharply to between west and north-west, and will moderate later on account of the slackening of the gradient behind the front. Intermittent rain and possibly drizzle are expected in the warm sector, with the usual short period of heavier rain at the front itself. The probability of showers in the polar air behind the front is indicated by the showers and convection cloud actually reported by stations in Ireland and western England. It has already been mentioned on page 422 that the air in advance of the front is probably returning maritime polar air, and therefore the visibility within this air is expected to be good; this is confirmed by actual reports. Good visibility is also expected in the polar air behind the front.

The forecast should therefore be issued on the following lines:—

Inference.—A trough of low pressure, extending across western Scotland, Wales and Cornwall, is moving slowly east.
Forecast.—Moderate to fresh southerly winds veering sharply to north-west about midnight and moderating later; mainly cloudy with intermittent rain or drizzle, at first; short period of heavier rain about midnight, followed by showers; visibility mainly good; becoming cooler with veer of wind; sea moderate.
Further outlook.—Moderate westerly to north-westerly winds; some squally showers, with bright intervals.

Aviation Route Forecast from Liverpool to Belfast commencing at 1400 G.M.T., 29th December, 1936.—Referring to the same synoptic chart (Fig. 139), the cold front, in the vicinity of Liverpool, is found to be moving east at 20 knots, so that at the time of commencement of the flight, it will lie just off the Lancashire coast. The aircraft will therefore be within the frontal zone during the first 30 minutes, after which it will encounter the polar air in the rear of the cold front. The geostrophic wind immediately in advance of the front, near to Liverpool, is about 30 knots; taking one-third of this as the speed of the surface wind over the land, a speed of 10 knots is obtained for the surface wind at the time of commencement of the flight. This agrees with the force 3 reported by station 128. The direction of the surface wind is obtained by allowing for an incurvature of approximately 30° across the isobars. In the rear of the front the geostrophic wind speed is about 24 knots, and over the sea, the surface wind will be two-thirds of this value, *i.e.*, 15 knots approximately. The angle of incurvature over the sea has an average value of about 20°, and since the isobars there lie in the direction 290°, the surface wind direction is taken as about 270°. The wind at 2,000 feet is taken as the geostrophic wind reduced according to the rule given on page 452. The aircraft will be practically within the frontal zone at the commencement of the flight, and so rain should be forecast; this will probably give way to showers within the polar air towards Belfast. Near to the front, cloud will probably be ten-tenths, and the forecast height should be based on reported heights of clouds at stations 123 and 126, *i.e.*, 1,000 feet. Further from the front the cloud is expected to break and lift somewhat; the forecast amounts and heights should be based on the observations at stations 107 and 105. Within the frontal zone, visibility will probably be reduced to about 5 miles, but there should be a rapid improvement in the clear polar air following the front.

The forecast should therefore be issued on the following lines:—

Wind at surface.—180°, 10 knots, veering sharply after about half an hour to 270°—280°, 15 knots.

Wind at 2,000 *feet.*—210°, 20 knots, veering sharply after about half an hour to 280°—290°, 20 knots.

Weather.—Some rain at first, with showers towards Belfast.

Cloud.—10-tenths at 1,000 feet at first, decreasing to 7—9-tenths, and lifting to 2,000 feet over Irish Sea.

Visibility.—About 5 miles at first, improving to 10 miles or over.

Forecast for the Mersey Area for the period 1500 to midnight G.M.T. on the 2nd October, 1935.—Fig. 140 shows the synoptic situation over the British Isles at 1300 G.M.T. on the 2nd October, 1935. An occlusion, associated with a depression to the west of Scotland, extends southwards over the Irish Sea. The mean speed of movement of that portion of the occlusion likely to affect the

Mersey area is given by the geostrophic wind measured between the isobars of 984 and 988 mb., and its component perpendicular to the occlusion is found to be about 15 knots: the occlusion is therefore expected to reach the Mersey area at about 1800. Until that time, the weather will be appropriate to that of an approaching occlusion, viz. :—continuous rain and poor visibility. In advance of the occlusion, the speed of the geostrophic wind is 55 knots and the customary reduction of one-third, gives 36 knots as the average speed of the surface wind over the sea. It should be borne in mind, however, that the Mersey has not an open sea exposure to southerly winds, but is sheltered to some extent by the mountains of North Wales. The surface wind will therefore be somewhat less than that calculated from the above, and a better guide is furnished by the winds reported at stations 123 and 126, viz. :— force 6 and 5, respectively. At the passage of the occlusion the wind will veer sharply to a westerly direction, and, owing to the opening out of the isobars behind the occlusion, will decrease to moderate or light. The continuous rain will cease at the passage of the occlusion, and in its rear the reports show that there are only a few scattered showers. The normal improvement in visibility in the rear of the occlusion must be allowed for, and station reports indicate good visibility generally. There is little change of temperature across the line of occlusion. The abnormal negative barometric tendencies over Ireland are most probably due to the approach of a new depression to the west of Ireland ; the information available does not permit this system to be located, but it may be expected to cause a backing of the wind to south again, with a further period of rain.

The forecast should therefore be issued on the following lines :—

Inference.—A deep depression is centred off western Scotland, and an associated trough of low pressure extending southwards over the Irish Sea is moving slowly east.

Forecast.—Strong southerly winds veering to west and decreasing to moderate or light in the early evening ; continuous rain during the afternoon, occasional showers later ; visibility poor, improving to good in the early evening ; sea rather rough, moderating later ; little change of temperature.

Further outlook.—Renewal of strong southerly winds with rain spreading from the west.

Aviation Forecast for aircraft operating in sea area within 200 miles radius of the Firth of Forth for period 1500 to 2100 G.M.T. on the 2nd October, 1935. (Fig. 140)—It is necessary to divide the forecast into 3 parts, viz. :—

(1) That portion of the area within 100 miles of the coast.

(2) The remainder of the area north of lat. 56° N.

(3) The remainder of the area south of lat. 56° N.

The average speed of the occlusion, as determined by measurement with the geostrophic wind scale, of the portion intercepted between the isobars of 982 and 986 mb. is about 15 knots. At the end of the period of the forecast, the occlusion will just reach the Firth of Forth and will extend southwards from it, so that a belt extending about 70 miles from the coast is likely to be affected by the pre-frontal weather. On the chart under consideration, the rain has reached stations 114, 120 and 128, *i.e.*, about 100 miles ahead of the occlusion. This coastal belt should be further subdivided into northern and southern parts on account of the differing winds. For the first part of the period the winds should be forecast in the normal manner from the run of the isobars. On the approach of the occlusion the wind is expected to back and increase slightly; rain, low cloud and diminished visibility which are normal pre-frontal conditions are confirmed by the reports from central and southern Scotland. It is not expected that the pre-frontal rain and cloud will reach Areas (2) and (3), and, therefore, mainly fair conditions should be forecast for these areas with cloud and visibility based on the reports from stations on the east coast. The forecast of wind should be based on the isobars in the usual way.

The forecast should therefore be issued on the following lines :—

Area (1). *Wind at surface.*—160° to 180°, 10 to 15 knots, backing to 140° and increasing to 15–20 knots in the evening.

Wind at 2,000 *feet.*—240°, 15 knots, backing to 180°–200° in southern part and increasing to 30 knots; 180° to 200°, 20 to 25 knots in northern part, increasing to 30 knots in the evening.

Weather.—Rain spreading from the west during the evening.

Cloud.—Increasing to ten-tenths at 1,000 feet.

Visibility.—Deteriorating to 3–6 miles.

Area (2). *Wind at surface.*—160° to 180°, 5 to 10 knots.

Wind at 2,000 *feet.*—180°, 15 knots.

Weather.—Fair.

Cloud.—Mostly high or medium cloud, up to five-tenths low cloud above 1,000 feet later.

Visibility.—10 miles.

Area (3). *Wind at surface.*—210°, 5 to 10 knots, backing to 180° in the evening and increasing to 10–15 knots.

Wind at 2,000 *feet.*—220°–240°, 10 to 15 knots, backing to 200°–210° in the evening and increasing to 15–20 knots.

Weather, cloud and visibility.—As for Area (2).

Forecast for Ship in lat. 45° N., long. 55° W. from 0000 (+4) to 1200 (+4) 4th January, 1937.—Fig. 141, which is drawn for 2100 (+4) 3rd January, 1937, shows a depression east of Hudson Bay with associated warm and cold fronts. The depression is

expected to move north-eastwards, its movement being based on :— (i) the direction of the isobars in the warm sector, and (ii) because its previous movement has been towards the north-east. The trough of low pressure, lying along the occlusion and the warm front, is assumed to be moving in the direction of the geostrophic wind in its rear. The component of the geostrophic wind in the rear of the warm front and perpendicular to it, is measured by the geostrophic wind scale and found to be approximately 30 knots. Deducting (say) 15 knots from this value for the movement of the warm front, the speed of the latter is estimated as 15 knots, so that it should reach the ship's position about 0700 (+4) on the 4th January. Similarly that portion of the cold front in the vicinity of Halifax is estimated to be moving towards the ship at about 25 knots, so that it should reach the ship after an interval of about 16 hours, *i.e.*, shortly after noon on the 4th. Up to the time of passage of the warm front the wind is likely to remain easterly, probably increasing somewhat in force ; at the passage of the front it will veer to south. Continuous precipitation is expected in advance of the warm front, and on account of the low temperature will probably take the form of snow or sleet, which is already reported at one or two stations. With the advent of the warm air the continuous rain will cease, but there is a probability of fog, caused by the warm, damp air blowing over the cold sea. Visibility in advance of the warm front will probably only be moderate on account of the precipitation. The state of the sea should be forecast in accordance with the anticipated winds. The changes indicated in the " further outlook " should be based on those normally experienced at a cold front, which is expected to pass the position shortly after noon on the 4th January.

The forecast should therefore be issued on the following lines :—

Inference.—A deep depression centred to the east of Hudson Bay and an associated trough of low pressure extending south-east to the Gulf of St. Lawrence, are moving north-east.

Forecast.—Fresh to strong easterly winds, veering to south during the morning ; overcast with continuous rain, snow or sleet throughout the night ; visibility moderate or poor during the night, fog patches towards noon ; sea rough ; cold tonight, becoming decidedly milder during the forenoon.

Further Outlook.—Wind veering to south-west and moderating during the afternoon, with squally showers and fall of temperature ; fog patches clearing ; sea moderating.

Forecast for Bass Strait for the period 1200 (— 10) to 2400 (— 10), 25th July, 1936.—Referring to Fig. 142, which is drawn for 0900 (— 10), 25th July, 1936, it is assumed that the depression shown will move south-eastwards in accordance with the direction of the isobars in the warm sector, and will deepen on account of its having an open warm sector. The movement of the warm front is given

by the geostrophic wind diminished by 15 knots; the measured value of the geostrophic wind is approximately 30 knots, so that the speed of the front is taken to be 15 knots. This speed will probably increase somewhat as the depression deepens, so that an average speed of 20 knots may be assumed. On the chart in question, the warm front lies approximately along a line 250 miles north of the centre of Bass Strait, and is therefore expected to reach that point at about 2100 (−10) on the 25th. During the approach of the front the wind will continue easterly, but precipitation will probably only be slight on account of the dryness of the warm air. At the passage of the warm front the wind will back to north and will increase as the depression deepens. Occasional light rain or drizzle is expected in the warm sector, and visibility will only be moderate on account of the probable dust-laden nature of the warm air, and its stable temperature distribution consequent on the cooling of the lower layers over the sea. The state of the sea is forecast according to the rule given on page 458 while an increase in temperature is anticipated with the advent of the warm sector. With the deepening of the depression, the speed of travel of the cold front is expected to increase from 20 knots (obtained from the measured geostrophic wind) to 30 knots. The cold front will therefore pass over the centre of Bass Strait about noon on the 26th. Its passage will be accompanied by a backing of the wind to westerly, a fall in temperature, and possibly also showers of rain and thunderstorms; the rain again will be slight owing to the dryness of the warm air. The wind will decrease as the depression recedes, with a corresponding decrease in the sea.

The forecast should therefore be issued on the following lines:—

Inference.—A depression centred south-west of Adelaide will move south-east and deepen.

Forecast.—Fresh easterly winds backing to north tonight and increasing to strong or gale; occasional rain or drizzle; visibility moderate; sea increasing to rather rough; temperature rising during evening.

Further outlook.—Wind backing to west tomorrow and decreasing to fresh, with showers and possibly thunderstorms; becoming cooler; sea moderating.

Forecast for ship in lat. 45° N., long. 7° W. from 1500 to midnight G.M.T. on 16th November, 1935.—Fig. 145 shows the synoptic chart for the eastern North Atlantic at 1300 G.M.T. on the 16th November, 1935.

The pressure has fallen by about 10 mb. off NW. Spain between 0700 and 1300, indicating that the depression is deepening very rapidly. The positions of the fronts have been located as explained on page 427. At 0700 (*see* Fig. 144) the depression was apparently centred in lat. 45° N., long. 17° W. and on the 1300 chart has moved to a position lat. 45° N., long. 13° W., in accordance with the direction of the isobars in the warm sector at 0700. At 1300 the isobars in the warm sector run ENE. and give a geostrophic wind

of about 40 knots. It is therefore anticipated that during the next 5 hours the depression will move ENE. at this speed and at 1800 should therefore be centred in lat. 47° N., long. 8½° W. The component of the geostrophic wind perpendicular to the warm front measured between the isobars of 992 mb. and 996 mb. is 40 knots; a reduction by the customary 10 knots gives 30 knots as the speed of advance of the warm front, and, hence, at 1800 it will have just reached the ship. Up to 1800 the wind will therefore remain SW. to S. and owing to the further deepening of the depression should increase to gale force. There will be continuous rain, and visibility will be poor. At the cold front the perpendicular component of the geostrophic wind measured between the isobars of 992 mb. and 996 mb. is 40 knots. At 1800 therefore the front will lie approximately along the west coast of Spain, and if the line of the front is produced to the centre of the depression it passes very close to the ship. It is anticipated therefore that the passage of the warm front at 1800 will be very closely followed by the passage of the cold front. Moreover, the 1300 chart shows that there is little change of wind at the warm front so that the ship will experience only one shift of wind, *i.e.*, a sharp veer to NW. when the cold front passes. Since the ship is near to the anticipated position of the centre of the depression, the rain is not expected to clear immediately after the passage of the cold front, and therefore the showery and squally conditions with good visibility associated with the NW. wind, are not anticipated to set in until late in the evening. The forecast of swell should be based on the SW. swell reported by *Adda*, obviously associated with the SW. winds in advance of the depression. After the passage of the cold front it is expected that the swell will become NW. as reported by the ships in positions approximately lat. 50° N., long. 20° W.

There is no indication of further disturbances approaching from the west; ships out in the Atlantic report NW. winds and rising barometer. For the period of the further outlook, therefore, it is anticipated that NW. winds will persist, but will moderate slowly as the depression recedes.

The forecast should therefore be issued on the following lines:—

Inference.—A deepening depression centred in lat. 45° N., long. 13° W., and an associated warm front extending SE. from its centre are moving ENE. A cold front also extends SSW. from the centre of the depression.

Forecast.—Wind SW. to S. strong to gale, veering sharply to NW. at about 1800; continuous rain at first, occasional squally showers towards midnight; visibility poor, becoming good towards midnight; sea rough; moderate SW. swell becoming NW.; becoming somewhat cooler this evening.

Further outlook.—NW. winds slowly moderating; weather mainly fair with good visibility.

TABLE VI.

TABLE FOR OBTAINING RESULTANT WIND DIRECTION AND SPEED IN KNOTS FROM COMPONENTS IN HUNDREDS OF FEET PER MINUTE.

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	
1	*45* 1	*27* 2	*18* 3	*14* 4	*11* 5	*9* 6	*8* 7	*7* 8	*6* 9	*6* 10	*5* 11	*5* 12	*4* 13	*4* 14	*4* 15	*4* 16	*3* 17	*3* 18	*3* 19	*3* 20	*3* 21	*3* 22	*2* 23	*2* 24	*2* 25	*2* 26	*2* 27	*2* 28	*2* 29	*2* 30	1
2	*45* 3	*34* 4	*27* 4	*22* 5		*18* 6	*16* 7	*14* 8	*13* 9	*11* 10	*10* 11	*9* 12	*9* 13	*8* 14	*8* 15	*7* 16	*7* 17	*6* 18	*6* 19	*6* 20	*5* 21	*5* 22	*5* 23	*5* 24	*5* 25	*4* 26	*4* 27	*4* 28	*4* 29	*4* 30	2
3	*45* 4	*37* 5	*31* 6			*27* 7	*23* 8	*21* 8	*18* 9	*17* 10	*15* 11	*14* 12	*13* 13	*12* 14	*11* 15	*11* 16	*10* 17	*9* 18	*9* 19	*9* 20	*8* 21	*8* 22	*7* 23	*7* 24	*7* 25	*7* 26	*6* 27	*6* 28	*6* 29	*6* 30	3
4	*45* 6	*39* 6				*34* 7	*30* 8	*27* 9	*24* 10	*22* 11	*20* 12	*18* 13	*17* 13	*16* 14	*15* 15	*14* 16	*13* 17	*13* 18	*12* 19	*11* 20	*11* 21	*10* 22	*10* 23	*9* 24	*9* 25	*9* 26	*8* 27	*8* 28	*8* 29	*8* 30	4
5	*45* 7					*40* 8	*36* 9	*32* 9	*29* 10	*27* 11	*24* 12	*23* 13	*21* 14	*20* 14	*18* 15	*17* 16	*16* 17	*16* 18	*15* 19	*14* 20	*13* 21	*13* 22	*12* 23	*12* 24	*11* 25	*11* 26	*10* 27	*10* 28	*10* 29	*9* 30	5
6	*45* 8	*41* 9	*37* 10	*34* 11	*31* 11	*29* 12	*27* 13	*25* 14	*23* 15	*22* 16	*21* 17	*19* 18	*18* 19	*18* 20	*17* 21	*16* 22	*15* 23	*15* 23	*14* 24	*13* 25	*13* 26	*13* 27	*12* 28	*12* 29	*11* 30						6
7	*45* 10	*41* 11	*38* 11	*35* 12		*32* 13	*30* 14	*28* 15	*27* 15	*25* 16	*24* 17	*22* 18	*21* 19	*20* 20	*19* 21	*18* 22	*18* 23	*17* 24	*16* 25	*16* 26	*15* 27	*15* 28	*14* 29	*14* 29	*13* 30						7
8	*45* 11	*42* 12	*39* 13			*36* 13	*34* 14	*32* 15	*30* 16	*28* 17	*27* 18	*25* 19	*24* 19	*23* 20	*22* 21	*21* 22	*20* 23	*19* 24	*18* 25	*18* 25	*17* 27	*17* 28	*16* 29	*15* 30	*15* 31						8
9	*45* 13	*42* 13				*39* 14	*37* 15	*35* 16	*33* 16	*31* 17	*29* 18	*28* 19	*27* 20	*25* 21	*24* 22	*23* 23	*22* 23	*21* 24	*21* 25	*20* 26	*19* 27	*18* 28	*18* 29	*17* 30	*17* 31						9
10	*45* 14					*42* 15	*40* 15	*38* 16	*35* 17	*34* 18	*32* 19	*30* 19	*29* 20	*28* 21	*27* 22	*25* 23	*24* 24	*23* 25	*23* 26	*22* 27	*21* 28	*20* 28	*20* 29	*19* 30	*18* 31						10
11	*45* 15	*43* 16	*40* 17	*38* 18	*36* 18	*35* 19	*33* 20	*31* 21	*30* 22	*29* 23	*28* 23	*27* 24	*26* 25	*25* 26	*24* 27	*23* 28	*22* 29	*21* 30	*21* 31	*20* 32											11
12	*45* 17	*43* 17	*41* 18	*39* 19		*37* 20	*35* 21	*34* 21	*32* 22	*31* 23	*30* 24	*29* 25	*28* 26	*27* 27	*26* 27	*25* 28	*24* 29	*23* 30	*22* 31	*22* 32											12
13	*45* 18	*43* 19	*41* 20			*39* 20	*37* 21	*36* 22	*34* 23	*33* 24	*32* 24	*31* 25	*29* 26	*28* 27	*27* 28	*27* 29	*26* 30	*25* 31	*24* 31	*23* 32											13
14	*45* 20	*43* 20				*41* 21	*39* 22	*38* 23	*36* 23	*35* 24	*34* 25	*32* 26	*31* 26	*30* 27	*29* 28	*28* 29	*27* 30	*27* 31	*26* 32	*25* 33											14
15	*45* 21					*43* 22	*41* 22	*40* 23	*38* 24	*37* 25	*36* 26	*34* 26	*33* 26	*32* 27	*31* 29	*30* 30	*29* 31	*28* 31	*27* 32	*27* 33											15
16	*45* 22	*43* 23	*42* 24	*40* 25	*39* 25	*37* 26	*36* 27	*35* 28	*34* 29	*33* 29	*32* 30	*31* 31	*30* 32	*29* 33	*28* 34																16
17	*45* 24	*43* 24	*42* 25	*40* 26		*39* 27	*38* 27	*36* 28	*35* 29	*34* 30	*33* 31	*32* 32	*31* 32	*30* 33	*30* 34																17
18	*45* 25	*43* 26	*42* 27			*41* 27	*39* 28	*38* 29	*37* 29	*36* 30	*35* 31	*34* 32	*33* 33	*32* 34	*31* 35																18
19	*45* 27	*44* 27				*42* 28	*41* 29	*40* 30	*38* 30	*37* 31	*36* 32	*35* 33	*34* 33	*33* 34	*32* 35																19
20	*45* 28					*44* 29	*42* 29	*41* 30	*40* 31	*39* 32	*38* 32	*37* 33	*36* 34	*35* 35	*34* 36																20
21	*45* 29	*44* 30	*42* 31	*41* 32	*40* 32	*39* 33	*38* 34	*37* 35	*36* 35	*35* 36																					21
22	*45* 31	*44* 31	*43* 32	*41* 33		*40* 34	*39* 34	*38* 35	*37* 36	*36* 37																					22
23	*45* 32	*44* 33	*43* 34			*41* 34	*40* 35	*39* 36	*38* 37	*37* 37																					23
24	*45* 34	*44* 34				*43* 35	*42* 36	*41* 36	*40* 37	*39* 38																					24
25	*45* 35					*44* 36	*43* 36	*42* 37	*41* 38	*40* 39																					25
26	*45* 36	*44* 37	*43* 38	*42* 38	*41* 39																										26

Angles in Italics

SPEEDS IN CLARENDON.

For Components larger than those shown halve both Components: use these halved values in the table and double the Resultant speed (But not the Angle)

$V_E + V_N +$ $\begin{cases} V_E > V_N & \phi = 270 - \theta \\ V_E < V_N & \phi = 180 + \theta \end{cases}$

$V_E + V_N -$ $\begin{cases} V_E > V_N & \phi = 270 + \theta \\ V_E < V_N & \phi = 360 - \theta \end{cases}$

$V_E - V_N -$ $\begin{cases} V_E > V_N & \phi = 90 - \theta \\ V_E < V_N & \phi = \theta \end{cases}$

$V_E - V_N +$ $\begin{cases} V_E > V_N & \phi = 90 + \theta \\ V_E < V_N & \phi = 180^\circ - \theta \end{cases}$

TABLE X.

	N	C_L	C_M	C_H	a	W		Y	Q	T_d	K	V_S	C	A	S	H_1	
0									Lat N. 0°–90° W	-10		0		0	Calm		0
1								Sun.	Lat. N. 90°–180° W	-7		9		10	Smooth	1000	1
2								Mon.	Lat. N. 180°–90° E	-4		18		30	Slight	2000	2
3								Tues.	Lat. N. 90°–0° E	-2		27		50	Mod.	3000	3
4								Wed.		-1		36		70	Rough	5000	4
5								Thurs.	Lat. S 0°–90° W	+1		45		10	V.rough	7000	5
6								Fri.	Lat. S 90°–180° W	+2		54		30	High	10,000	6
7								Sat.	Lat S 180°–90° E	+4		63		50	V.high	13,000	7
8									Lat. S 90°–0° E	+7		72		70	Precipitous	16,000	8
9										+10		81			Confused	20,000	9

TABLE XI.

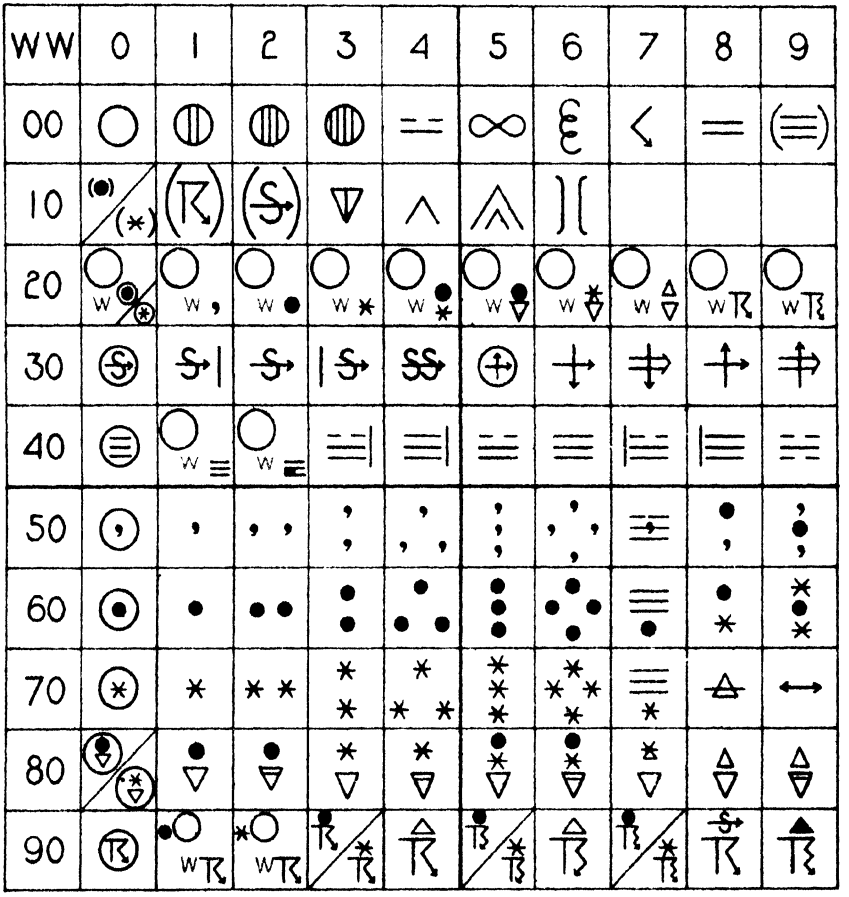
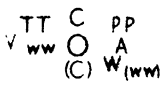


TABLE XIII.

First Figure	Present Weather ww Second Figure													Day of Week	Octant of Globe	Bar Tendency	Cloud	Past Weather	Swell	Dew Points
	0	1	2	3	4	5	6	7	8	9				Y	Q	A	C	W	K	U
0	b	bc	c	o	Low fog	m_o or z_o	Dust devil	l	m or z	jf	0				Lat N 0°-90°W	0	St	b	No swell	1
1	jp	t	jkz	u	q	Last 3 hours Heavy Squall	Last 3 hours Water Spout				1			Sun	Lat N 90°-180°W	1	Ci	bc	Low Short	44
2	/r d h, S or rs	/d	/r	s	/rs	/pr	/ps	/ph(r)	/tlr$_o$	/TLR	2			Mon	Lat N 180°-90°E	3	Cs	o	Low Long	35
3	kz	kz-	kz	kz+	kzz	ks	k/s$_o$	k/S	s/k	S/k	3			Tues.	Lat N 90°-0°E	5	Cc	kz	Moderate Short	27
4	f	/f	/F	f- sky vis.	f-	f sky vis.	f	F sky vis.	F	lf	4			Wed.		7	Ac	f	Moderate Average	21
5	d	ld$_o$	dd$_{o\,o}$	ld	dd	lD	DD	df	dr dr	Dr	5			Thurs.	Lat S 0°-90°W	1	AS	d	Moderate Long	16
6	r	lr$_o$	r$_o$r$_o$	ir	rr	lR	RR	rf	r s$_o$ rs	RS	6			Fri	Lat S 90°-180°W	3	Sc	r	Heavy Short	12
7	s, rs	ls$_o$	s$_o$s$_o$	ls	ss	lS	SS	sf	sh	Ice crystals	7			Sat	Lat S 180°-90°E	5	Ns	s	Heavy Average	8
8	pr	pr$_o$ pr	PR	ps$_o$ ps	PS	prs$_o$ prs	PRS	psh	ph(r) ph$_o$(r)	P H(R)	8				Lat S 90°-0°E	7	Cu	p	Heavy Long	4
9	tlr	r/tlr	S r s/tlr	tlr$_o$ (s)	tlh$_o$	tlr (s)	tlh	TLR(S)	tlkz	TLH	9						Cb	tlr	Confused	2

Note This table may be used with Fleet Synoptic Messages when it is considered undesirable to use the more complicated symbols of Decode Card H 354
Nos in column 'U' should be subtracted from TT to give the Dew Point.

Station Model



The circle denotes the position of the station or ship
A - Barometric tendency
C - Form of predominating cloud types of low cloud are plotted below the station circle
PP - Barometric pressure.
TT - Temperature
V - Visibility
ww - Present weather. When ww is 20-29, 41 or 42 it is plotted alongside W
W - Past weather

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NOTE.—*Where full indexing on any subject would involve merely a repetition of the Table of Contents reference has been made to that Table.*

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NOTE

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