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The Geological Society of America
Memoir 80

STILLWATER IGNEOUS COMPLEX, MONTANA

A QUANTITATIVE MINERALOGICAL STUDY

By
H. H. HESS
Princeton University, Princeton, New Jersey

With an Appendix by J. R. Smith on
OPTICAL PROPERTIES OF LOW-TEMPERATURE PLAGIOCLASE

Presidential Address before
The Mineralogical Society of America

1960

Made in the United States of America

PRINTED BY WAVERLY PRESS, INC.
BALTIMORE, MD.

PUBLISHED BY THE GEOLOGICAL SOCIETY OF AMERICA
Address all communications to the Geological Society of America
419 West 117 Street, New York 27, N. Y.

The Memoir Series
of
The Geological Society of America
is made possible
through the bequest of
Richard Alexander Fullerton Penrose, Jr.

PREFACE

J. W. Peoples, A. L. Howland, and Edward Sampson began the study of the Stillwater complex, Montana, in 1930. The writer commenced a detailed study of the mineralogy of the complex in 1935 after several periods in the field in preceding years with the three other men. Originally it was planned to prepare a joint memoir in which Peoples would write the general geology of the complex, Howland the floor rocks and sulfide deposits, Sampson the chromite deposits, and the writer the mineralogy. Twenty years has elapsed, and a war has intervened. Peoples, Howland, Sampson, and the writer have published a number of short papers on the complex. (Peoples, 1933; 1936; Howland, Peoples, and Sampson, 1936; Hess (1938a; 1939); Peoples, 1939; Peoples and Howland, 1940; Howland, 1955). Since there seemed little likelihood of preparing a joint memoir within a reasonable number of years, the writer has gone ahead on his part independently.*

This manuscript was written in 1939 and largely rewritten while the author was at the University of Cape Town in 1949–1950. It was presented as the Presidential Address before the Mineralogical Society of America in New Orleans in November 1955. Stress of administrative duties has prevented the writer from modifying the whole manuscript in light of all the recent literature, and some conspicuous deficiencies in this respect may be noted. Rather than delay it any longer, however, the writer prefers to present it at this time with those shortcomings.

The work of J. R. Smith on plagioclase in Appendix 3 supersedes a long but relatively unsuccessful investigation by the writer. While Smith certainly profited by the writer's mistakes, the excellence of his achievement with this group of minerals adds much to the value of this report.

In the writing of this report the author wishes to acknowledge his great dependence on the work of his colleagues—Peoples, Howland, and Sampson—without which it would not have been possible to accomplish the results herein contained.

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July 1956

* *Igneous and tectonic structures of the Stillwater Complex, Montana*, by W. R. Jones, J. W. Peoples, and A. L. Howland is in press for issue as Bulletin 1071-H of the U. S. Geological Survey.

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ABSTRACT

The Stillwater igneous complex of Montana is an intrusion of lopolithic form. The floor and about 16,000 feet of layered rocks are exposed on the East Boulder plateau. It is estimated that another 40 per cent of the complex and its roof are hidden beneath a cover of Paleozoic rocks. North of the complex some Precambrian roof rocks are exposed by faulting and erosion.

The order in which the essential minerals separated from the Stillwater magma is as follows: olivine, orthopyroxene, plagioclase, and clinopyroxene. Near the uppermost outcrops hypersthene inverted from pigeonite appears. The only other mineral present in rock-forming amount is the accessory, chromite, which is concentrated in a few layers in the lower part of the complex. The complex represents a natural example of fractional crystallization on a vast scale. Crystals separate from the magma and accumulate on the floor only to be covered by more precipitate which thereby prevents reaction with the remaining liquid.

The chemistry of the essential minerals—plagioclase, orthopyroxene, and clinopyroxene—is thoroughly investigated. New optical-property curves are presented, and the accuracy and suitability of various determinative procedures are discussed. Smith's curves for low-temperature plagioclase represent a marked improvement over existing curves, and his determinative procedure is an advance both in precision and ease of measurement. The exsolution phenomena in the pyroxenes are used to interpret the subsolid relationships and to draw petrologic conclusions particularly as regards temperature during crystallization.

Curves are drawn to show the order in which the essential minerals separated from magma and their composition changes as crystallization proceeded. The general trend of the curves from more magnesian to more iron-rich for the ferromagnesian minerals and from more calcic to more sodic for the plagioclases is evident. Of particular interest are the minor oscillations of these curves. These can be correlated with the effect of pore-space liquid on the accumulated crystals on the floor. The pore space in the crystal mush is at least 25 per cent by volume. The effect produced by this pore-space liquid is very much less than what would be expected; consequently it is postulated that diffusion from the pore-space liquid to the overlying main mass of magma and vice versa has tended to diminish the effect. When accumulation of crystals on the floor is slow diffusion may be so effective that the phases formed will not be appreciably different in composition from the accumulated crystals; when accumulation is rapid the pore-space liquid can have a large effect in modifying the composition of the crystals formed. Phases present in small amount among the accumulated crystals can be modified to a greater extent by pore-space liquid, and phases present in large amount are modified to a lesser extent. At any rate minor fluctuations in the mineral-variation curves can be correlated with the local rate of accumulation of crystals on the floor. The process also accounts for the formation of certain monomineralic rocks. Without diffusion the trapped pore-space liquid would have contributed small amounts of other phases to the rocks.

With respect to layering in the Stillwater complex both the observations and conclusions closely parallel those of Wager and Deer for the Skaergaard intrusion. A theory to account for rhythmic layering is advanced. If crystals with different settling velocities are present and if the magma is in motion, rhythmic layering will result provided the vertical component of the motion in the magma is not constant. Velocities such as $100m/\text{yr}$ are of the right order to produce the desired result if the viscosity is 3000 poises. Well-developed rhythmic layering, however, is almost always associated with very large differences in settling velocities between phases. Thus it is conspicuously developed in the Stillwater complex, Norite Zone, where the orthopyroxene crystals are comparatively large and plagioclase comparatively small.

The settling velocity difference is 25 to 1. This suggests that the process operates in a layer of magma very close to the floor, perhaps in the lower 100 meters.

The composition of the magma is estimated from the composition of the border facies. It is very similar to the Bushveld complex. It is a basaltic magma, just saturated, and is comparatively low in TiO_2, Fe_2O_3, and K_2O. Otherwise it is much like the Karroo dolerites or Palisades diabase. An estimate of the composition of the hidden uppermost rocks is made by subtracting the composition of the exposed complex from the original magma in suitable proportions. If 40 per cent is hidden the resulting composition by subtraction is such that it corresponds to rocks of the uppermost portion of the Bushveld complex or Great Dyke—ferrogabbro.

An estimate of the rate of heat loss during crystallization suggests that about 10 cm of crystals are deposited on the floor a year. If the roof is much thicker than 1 km this value may have to be reduced. Data on viscosity of basalts, densities of phases at high temperatures, and settling rates of crystals are analyzed. It is shown that two-phase convection in the magma is probable.

In appendices certain general propositions are discussed. The calc-alkaline magma series so differs in its trends from those produced by fractional crystallization that some other origin seems required. A partial melting sequence is suggested. This leads to a consideration of the generation of magmas and the vertical composition gradient in continental crust. Differentiation of dolerite sills parallels the trend observed in the Stillwater complex but with a lower degree of fractionation. Absence of igneous lamination suggests that the process is unrelated to crystal settling. A process of inward crystallization from cooling surfaces analogous to the crystallization of the pore-space liquid in the Stillwater crystal mush is suggested; a diffusion mechanism removes the film of residual liquid.

Chapter I

Introduction

A. SUMMARY OF MAJOR RELATIONS

The Stillwater igneous complex is a lopolith—a huge concordant sheet of which the base and lower portion, some 16,000 feet thick, are exposed. The roof and upper portion are hidden beneath overlapping Paleozoic sedimentary rocks. It is estimated that this covered upper portion constitutes about 40 per cent of the total thickness of the complex.

The magma cooled slowly after intrusion so that extreme fractional crystallization was possible. Apparently crystals accumulated in layers on the floor until the lopolith became solidified.

The layering in the igneous rocks is approximately parallel to the floor, and it is reasonably certain that this layering developed in horizontal or nearly horizontal position. At present both the floor and the layering are tilted northward with an average dip of about 55° on the East Boulder Plateau. The northward dip is a result of Precambrian deformation. Peoples has shown that there is an angular unconformity between the layering and basal Cambrian. Furthermore, the entire plateau surface seems to be very near the former Precambrian–Cambrian contact since remnants of Cambrian are found downfaulted into it in a number of places. Laramide deformation has steepened the dips along the mountain front so that at some places the layers are vertical or even overturned (northern margin of the exposed complex).

B. SCOPE AND PURPOSE OF INVESTIGATION

The well-founded opinion that the crystals accumulated layer upon layer on the floor is based on careful field work by the writer's colleagues, in particular Peoples, and checked by the writer's field observations. Assuming this opinion to be correct the writer believed that a detailed study of the composition of the crystalline phases in the rocks from the floor to the uppermost exposed outcrops would shed much light on the course of crystallization of a basaltic magma under conditions of extreme fractionation.

The problem as to whether the plagioclase rose or settled in the magma or followed some more complex course such as rising first and settling later might also be solved by this type of investigation.

C. RESULTS OBTAINED

The investigation has now been completed. The compositions of thousands of grains have been determined. The opinion stated above concerning accumulation of crystals on the floor has been completely substantiated. The results obtained have not been as simple as it was originally thought they would be. In addition to the accumulation of crystals on the floor, certain other subordinate processes at work have modified the expected simple relations. The most important of these is the effect of (1) the interstitial liquid between the accumulated crystals (amounting to about ap-

proximately 28 per cent of the rock by volume) and (2) probably a process of diffusion of material from the magma above, into, and out of the interstitial liquid. This process is thought to be operative through several feet of crystal mush in contact with overlying magma during crystallization.

Minor changes in the compositions of the essential minerals—plagioclase, clinopyroxene, orthopyroxene, and olivine—are significant in the interpretation of the results. This necessitated much work to devise practicable techniques for the determination of optical properties of grains with a degree of accuracy as much as one decimal place beyond that ordinarily attained in standard practice. The attainment of a slightly greater degree of accuracy involved a disproportionately greater degree of difficulty and labor. Furthermore, the chemical and optical data available on three of the four essential minerals were so inadequate that they were practically useless for the purpose of this investigation. Only olivine, the least important of the four, was sufficiently well known to require no further investigation. Calcic plagioclase, orthopyroxene, and clinopyroxene had to be completely reinvestigated. This involved a long detour, almost, but not quite, finished at this time. Instead of 1 year's work, as originally planned, more than 20 years has been necessary to complete this project. The writer, however, has no regrets at this turn of events, since the work proved most interesting, and the results will no doubt be of more value to other investigators than the original project.

D. FIELD WORK AND COLLECTION OF SPECIMENS

Several weeks in each of the field seasons 1931, 1932, and 1933 were spent on the Stillwater complex. In 1931 and 1932 the writer visited Peoples while he was mapping the complex. During the 1933 field season the writer assisted Sampson at the Mountain View Lake chromite deposits and made several trips across the complex independently. Thus the writer had a fairly good background before starting work on the problems presented in this report.

After consulting Peoples and Sampson, the writer decided that a traverse across the East Boulder Plateau would be the best available cross section of the complex for the purposes of this investigation. Along this line the greatest thickness of the complex is exposed.

In 1935 the writer, assisted by R. K. Doten and G. L. Davis, mapped (Pl. 1) the East Boulder Plateau and collected about 100 specimens along a line extending from the base of the complex to the highest exposed horizon. Study of these specimens indicated that additional specimens were needed to explain the minor changes indicated in the mineral-variation curves. In 1936 Peoples and Howland collected an additional 100 specimens from the East Boulder Plateau section, as well as 100 specimens from parallel traverses at other places in the complex. A large number of specimens collected by Peoples, Howland, Sampson, and the writer in previous years were also available for study.

The East Boulder Plateau map (Pl. 1), made during the 1935 field season, was constructed essentially to locate the specimens and tie in to the line of traverse of the cross section such other features as seemed desirable. The U. S. Geological Survey topographic sheet of the area was at that time incomplete. After completion of this

sheet, the data from the 1935 field work was transferred to it (Pl. 1). Some names appearing on Plate 1 are local topographic features or station designations which have been added for convenience in describing various features or relationships within the mapped area.

E. PLAN OF PRESENTATION

The writer attempts to separate factual and descriptive data from hypotheses and conclusions. A complete separation is not possible without a great deal of undue repetition making for tedious reading and unwieldy length of manuscript. The general conclusions are given in the later chapters following the descriptive portion of the report. Minor conclusions must necessarily be coupled closely with the factual detail on which they are based.

F. ACKNOWLEDGMENTS

Grants from the A. H. Phillips Fund administered by the Department of Geology of Princeton University covered the cost of this research including the field work, chemical analyses, laboratory equipment, and thin sections. An allotment from the Higgins Trust of Princeton University made possible the X-ray work on the pyroxene.

The writer wishes first to express his indebtedness to his three colleagues in this investigation—Peoples, Howland, and Sampson—and for their aid and advice over the period of years during which this work was done. In particular, the writer gratefully acknowledges his dependence on the firm foundation of Stillwater geology established by Peoples. Discussion of mutual problems with Peoples as the work progressed was an important factor in the final definition of ideas embodied in this report.

Originally the chemical and optical study of the minerals was undertaken jointly by Professor A. H. Phillips and the writer. Before his death Professor Phillips had completed many of the chemical analyses of minerals appearing in this report. As all who have known Professor Phillips can well understand, the absence of his hearty laugh ringing from the chemical laboratory is a source of much regret to all of us.

Since Professor Phillips' death a number of chemical analyses were made for this report by R. B. Ellestad, L. C. Peck, and E. K. Oslund at the Rock Analysis Laboratory at the University of Minnesota. In working on a small group of minerals and rocks over a period of years, it is not difficult to judge the quality of chemical analyses of given specimens. The writer wishes to congratulate the analysts for the high quality and consistency of their results.

Throughout the course of this investigation, the writer had relied much on the critical judgment and advice of his colleague, Professor A. F. Buddington. Almost daily debates with Buddington on the petrological problems have been both a source of stimulation to solve the problems of the Stillwater complex and a clearing house where many ill-conceived ideas were efficiently eliminated. The writer also wishes to thank Buddington for critically reading and correcting the manuscript.

The field work in 1935 was carried out with the aid of R. K. Doten and G. L. Davis, to both of whom the writer wishes to express his thanks for generous and capable assistance.

For assistance in the laboratory in 1939 and 1940, and in particular for the making of Rosiwal analyses of many of the rocks, the writer wishes to thank Frederick Betz for his careful and painstaking work.

During 1953–1955, L. C. Coleman determined the unit-cell dimensions of the clinopyroxene.

A short trip across the Stillwater complex with Drs. A. L. Hall and Alexander DuToit, of South Africa, in 1933 proved most helpful because of their intimate knowledge of the analogous Bushveld complex. Similarly trips across the Bushveld with C. A. Strauss and across the Great Dyke with F. L. Amm, J. W. Wiles, and J. G. Stagman were very fortunate experiences for the writer.

Certain published works were made use of extensively during the investigation; among these are particularly the following: Bowen—*The evolution of igneous rocks*; Hall—*The Bushveld complex*; Bowen and Schairer—*The system MgO-FeO-SiO$_2$*; Lombaard—Several papers on the Bushveld complex; Wager and Deer—*The Skaergaard intrusion*; Cooper—*The Bay of Islands complex, Newfoundland*; Walker's papers on the Palisades diabase and Karroo dolerites, the latter in collaboration with Poldervaart; Benson—on the basic igneous rocks of Eastern Otago; Burri and Niggli—*Eruptivgestein des mediterranen Orogens.*

The writer is much indebted to Arie Poldervaart and H. D. Holland for critical reading of the manuscript.

Chapter II

Determinative Methods

A. INTRODUCTION

No comprehensive treatment is attempted here of all the methods applied in the determination of properties of crystals examined. Comments will be made on the methods found to be more precise or convenient than those commonly employed. None of these methods is new, but some are seldom used in practical petrography. Some are convenient combinations of well-established methods. Many methods were tested and found cumbersome or lacking in the precision usually attributed to them. Inasmuch as the observations were made from 1937 to 1953, few of the techniques favored at the completion of this report were in use during the earlier years, and most of the progress in improving methods was made at the end of the period.

It is no exaggeration to say that hundreds of thousands of observations were made of various optical properties on various mineral grains during this investigation. It would have been more helpful if the data had been so collected that exact numerical values for precision could have been given for all the observations, and errors systematically estimated. Although this was not done in many cases, one can from such a large number of observations set down a considered opinion on precision and accuracy which is almost as good, although perhaps less convincing.

B. INDEX OF REFRACTION MEASUREMENTS

1. GENERAL

This investigation has indicated that the accuracy of index of refraction measurements by routine methods is probably much lower than generally supposed. There are manifold reasons for this. Estimates of the errors for the minerals plagioclase, orthopyroxene, and clinopyroxene are as follows:

(1) White Light vs. Sodium-Vapor Lamp

Error for white light to be expected is $\pm.001$, although it varies with the relative dispersions of the mineral and immersion liquid.

(2) Calibration of the Liquid with a Good Abbe-Type Refractometer with Water-Temperature Control

Error about $\pm.0002$ in the feldspar range but increasing as the index of the glass prism in the refractometer is approached.

(3) Temperature of the Immersion Liquid at Time of Observation

If a temperature-control cell is used this can be held down to about $\pm.0002$. On an ordinary glass slide with no precautions to filter out the heat from the light source, the error can be $\pm.0015$. With a heat filter most of this error can be eliminated.

(4) Change of Index of Liquid by Evaporation or Decomposition

This depends on the length of time for the observation, the vapor pressures of the immersion liquids (if they are not single-component liquids), and whether or not a photo-sensitive liquid such as an iodide is being used. In the feldspar range this error is usually negligible. For pyroxene it is probably negligible for observations lasting 10 to 15 minutes at room temperature, but errors mount rapidly above 30°C.

(5) Error in Orientation of Grain

The error from this source can be stated only in terms of a particular mineral. Wherever use of a convenient parting or cleavage containing one of the three principal optical directions is possible it should be taken advantage of. It is preferable to expend the time to determine this one direction accurately rather than three principal directions with less accuracy. Determination of minimum and maximum indices of refraction of plagioclase in powders commonly involves an error of at least ±.001 and can easily be ±.002.

(6) Matching Error

Ordinarily for rock-forming minerals there is an interval of about ±.0005 on either side of the true match between the liquid and an index of the grain, in which the liquid and grain will not differ perceptibly in index for the Becke line or half-shadow test. The nature of grain boundaries, minute foreign specks in the mineral, and absorption of the mineral, if colored, all contribute to this. On glasses, ideal for observations, this error may be as low as ±.0001. If observations are made by changing temperature on a temperature cell in small steps or by rotating on the microscope stage (as described below) limits may be set between a reading of the index of the grain definitely higher and definitely lower than the liquid. Splitting this interval in half should normally cut this error to about ±.00025.

(7) Inhomogeneity of the Mineral

Although the individual grains of minerals from rocks of layered complexes such as the Stillwater are extraordinarily homogeneous and unzoned nevertheless composition variation from grain to grain may generally be expected in the range of 2–4 per cent in An content of plagioclase and a similar amount in Mg:Fe ratio of pyroxene. A sufficient number of grains must be measured to insure a suitable average. Normally 5–15 observations are necessary; the number is based on the variation limits found in the first 5 observations. In some cases it is simpler to take care of this statistically by observing a large number of grains at a fixed index for the liquid and recording the number above, equal to, and below the liquid. Then by changing the temperature in small steps and repeating the observations a good statistical curve for index distribution, and hence average composition, can be obtained.

(8) Time vs. Accuracy

In a laboratory set up for index determinations, one index of refraction can be measured with an accuracy of ±.002 in 10 minutes. Added accuracy necessitates an exponential rise in time consumed. To increase accuracy to ±.001 probably increases the time factor by 5, and to ±.0005 by 25 using standard methods. Smith has succeeded in improving determinative technique to a point where ±.0003 accuracy can be obtained on plagioclase in 1 hour (Appendix 3).

2. PLAGIOCLASE

All the common methods of determining indices of refraction of plagioclase were used at one time or another in this investigation. Maximum and minimum index of powders gave fairly good results so far as speed was concerned, but this method was comparatively inaccurate.

For several years the Emmons double-variation method with the universal stage and temperature control was tried but proved slow and cumbersome. Results were generally poor because insufficient light was obtained from the monochromator or too broad a band of wave lengths had to be used. Early difficulties with temperature control were effectively dealt with by building equipment such as described by Hurlbut (1947). Powders cannot be used satisfactorily on the universal stage because all but the largest grain slides from view when the stage is tilted. Gelatin coating of the glass slide to prevent this makes it difficult to see the Becke line because of a little ridge of gelatin around the grain. In later years measurements on the universal stage were made by dissolving a small portion of a thin section off a slide and using the resulting plates of the mineral for mounting on the stage or by using solid mounting media. If a good monochromator were available the Emmons double-variation

method would perhaps be practical. It is advisable, however, to remove a refractometer from the system because it introduces a long period of waiting, after each temperature change, for the system to settle down to thermal equilibrium. The index liquids may instead be calibrated in advance for temperature and wave length.

The same result may be obtained more effectively by a method devised by Merwin at the Geophysical Laboratory in which an appropriate glass of known index and dispersion is mixed with the powder to be determined. If the dispersion curves for the immersion liquids and mineral are known, the temperature control and the refractometer may be dispensed with. If the dispersion of the mineral is unknown, temperature changes are necessary to determine the curve. Matching the glass and liquid by changing wave length gives the proper dispersion curve to be used for the liquid at the temperature obtaining. Matching the mineral and liquid gives the index of the mineral for the particular wavelength of the match. Because we lacked a good source of monochromatic light of variable wave length this method was not used.

Dissolving thin plates of a mineral off a thin section for use on the universal stage is not entirely satisfactory because it requires making a section and partially destroying it, and also because such plates when tilted commonly have poor edges for observing the Becke line and relatively few edges to observe compared to a powdered mineral. Smith and the writer developed methods employing solid mounting media which proved much more satisfactory. Ideally one should have a set of such media to cover the plagioclase range, perhaps six or eight ranging in index by 0.015. Such a set has not yet been developed.

Glycol phthalate (N = 1.5708 at 25°C.) was used for plagioclase in the range An_{70} to An_{85} with promising results. The difficulty encountered was in making a precise determination of the index and temperature coefficient of the solid medium. This could best be done by the minimum-deviation method described by Fairbairn (1952), but this has not yet been tried by the writer. In lieu of determining the index of the solid medium, glycol phthalate was used to coat glass slides. The powdered mineral was sprinkled on the surface of the slide after warming. The grains stick firmly to the surface when cooled and are thus held in place for universal-stage work. The grains are then immersed in an appropriate index liquid, and a cover glass placed over them. For holding the plagioclase grains this proved a much more suitable means than gelatin because it is clear and hard enough so the grains are not pushed into it. It cannot be used with index liquids above 1.60 because it reacts with them. Tests showed no appreciable reaction with the Shillaber index liquids from 1.52 to 1.58. By scraping off the mineral grains and some of the glycol phthalate with a razor blade, the same slide may be used indefinitely.

The standard universal-stage procedure for index determination of crystals such as plagioclase is well described by Emmons (1943) and further refined by Smith (Appendix 3). Briefly, the principal optical planes are set up so that one is N.–S. and vertical, one E.–W. and vertical, and the third horizontal. If N_z is vertical and N_x horizontal and N.–S., and the grain is mounted in a liquid of index slightly higher than N_x, then rotation on the control axis from N_x toward N_z will pass through a point where $N_{x'}$ equals the liquid. In many cases one can find $N_{x'}$ for rotation both northward and southward on the control axis. This tends to eliminate the error

resulting from slight misorientation of the principal optical directions in the N.–S. plane. Knowing the birefringence either by measurement or by reading it from the plagioclase optical-property curves after a first estimate for N_x, and the angle of rotation on the control axis, one can find the partial birefringence $N_{x'} - N_x$ on Emmons' (1943) Plate 10a. The $N_{x'}$ is the index of the mounting medium so that, with $N_{x'} - N_x$ known, N_x can be obtained by difference. If the rotation had been from N_z toward N_y or N_y toward N_x the same procedure would follow except that the partial birefringence $N_z - N_y$ or $N_y - N_x$ would have to be determined or read from the optical-property curves. If $N_z - N_x$ and $2V$ are known the Wright (1911) curves can be used to obtain either of the partial birefringences (also Emmons, 1943, Pl. 11).

This will henceforth be referred to as the "orthodox" method. An unorthodox method which is theoretically unsound but practically very effective was developed and used by A. P. Subramaniam, J. R. Smith, and the writer in 1953. Instead of rotating the grain on the control axis the microscope stage was rotated. Thus if N_z and N_x for example were horizontal the angle from $N_{x'}$ to N_x was read on the stage. This permits a rotation of 90° if necessary. Such an observation can be made in 2–3 minutes. One observes a Becke line going into the mineral and one going out into the medium. When these are balanced in intensity $N_{x'}$ is assumed equal to the medium. This setting can be made to about ±1°. Using the angle of rotation found and the birefringence, Plate 10a of Emmons solves for partial birefringence and hence N_z. A rather extensive test of the method was made since it has no clear basis in optical theory. Fifty crystals of plagioclase Eb 38 gave N_x by the unorthodox method as 1.5672, and 20 crystals by the orthodox method gave 1.5674. Five crystals measured by rotation on the control axis followed by rotation of the stage gave N_x by the orthodox method also .0002 higher than by the unorthodox method. While the method is suitable for plagioclase or other minerals of low birefringence, large errors result if it is used for minerals whose birefringence is higher than plagioclase. It can be applied to orthopyroxene cleavage fragments without the use of a universal stage.

Care must be taken to control the temperature of the slide. This can be done by using a water-temperature controlled stage or by allowing the mount to come to equilibrium with room temperature (20–30 minutes) and filtering the heat from the sodium-vapor lamp.

Introduction of a quartz standard for the 1.544–1.553 index range makes temperature control unnecessary. Rotation from N_ω to N_ϵ of quartz determines $N_{\epsilon'}$ where the quartz index equals the medium *at the temperature obtaining*. Immediately after this determination, $N_{x'}$ for the plagioclase is compared with the mounting medium. The quartz reading may be made again if any temperature change is suspected. If a series of suitable synthetic crystals could be made up to take the place of quartz for other indices of refraction, this would be the ideal means of accomplishing accurate and rapid determinations of index of refraction. It eliminates temperature control and the refractometer error.

The determination of composition of plagioclase by using indices of refraction on

cleavage fragments representing {010} or {001} has been favored by many authors (Tsuboi, 1923). The writer had little success with this method for Stillwater plagioclase. Good recognizable cleavage fragments are not common in the powders. About one in five such fragments observed by the writer were prism parting plates {110} or {1$\bar{1}$0}. Thus with four possible types of plates present it becomes necessary to identify which type one is observing, and this is rather difficult. Furthermore, adequate index curves for cleavage plates are not available and were not made up for the present investigation.

3. ORTHOPYROXENE

All index observations on orthopyroxene were made by using the prismatic cleavage and determining N_z (Walls, 1935). For many of the routine determinations this value was observed on ordinary slides in sodium light with a heat filter using a set of liquids spaced .002 apart and making observations on different days for different room temperatures so that the effective spacing was .001. In recent years observations were made on a temperature cell; the temperature was increased by steps of 1°C., and the number of grains above, equal to, and below the liquid was recorded for each temperature. Commonly 10–15 grains were observed. After completion of the observations for this report a better method was devised, the "unorthodox" method described above for plagioclase.

The orthopyroxene grains are placed in a liquid of index slightly less than N_z. Using {210} prism cleavage fragments, the angle of rotation is found from N_z to the point where the liquid and $N_{z'}$ are equal. This is done on an ordinary glass slide using sodium light and measuring the temperature carefully on the microscope stage before and after the observations. Several grains can be so measured in a few minutes. To solve for N_z the optic angle and birefringence must be known. For routine work these can be estimated from optical-property curves and an estimate of N_z. If great accuracy is desired $2V$ should be measured, but birefringence can still be taken from the curves inasmuch as it changes very slowly with change of composition. The necessary calculations can be made rapidly using Wright's (1911) curves relating $2V$ and the total and partial birefringences in the principal optical planes and solving for the partial birefringence on {210} and partial birefringence $N_z - N_{z'}$ using Emmons' (1943) Plate 10a. To make the procedure clear, an example is worked out in Table 1 and illustrated in Figure 1.

4. CLINOPYROXENE

The N_y index of clinopyroxene was determined on either {100} or {001} parting plates. It is difficult to determine which of the two is being used, and immaterial so far as the N_y determination is concerned. Both have low birefringence and show an off-center optic-axis interference figure. The grains are tested by observing the interference figure and orienting the isogyre so that it is E.–W. If the optic plane is vertical the E.–W. isogyre exactly bisects the field, and N_y is N.–S. (Fig. 2) as described by Slawson and Peck (1936). Index N_y is then determined in sodium light on a temperature cell.

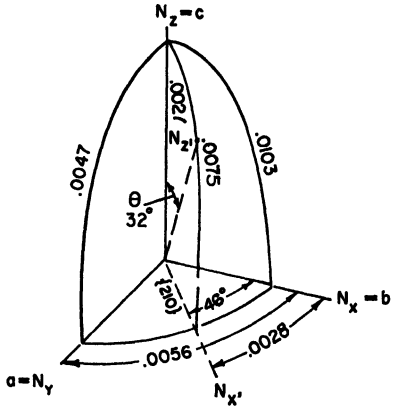


FIGURE 1.—*Diagram illustrating calculation of partial birefringences and N_z for an orthopyroxene*

TABLE 1.—*Calculation of three indices of refraction from observed N_z', birefringence, and optic angle*

$2V$ (−) = 85.0° measured
Birefringence, $N_z - N_x$ = .0103 (from curve, Fig. 9)
Partial birefringence $N_z - N_y$ = .0047 from Wright's curve
.0103 − .0047 = $N_y - N_x$ = .0056
$N_{x'} - N_x$ = .0028 from Emmons' Plate 10a
$N_{x'}$ 46° from N_x in N_y, N_x plane {001}
$(N_z - N_z) - (N_{x'} - N_z)$ = .0075 partial birefringence on {210} cleavage
$N_z - N_{z'} = \theta$ = 32° Observed angle of rotation on microscope stage to make $N_{z'}$ equal to liquid
$N_z - N_{z'}$ = .0021 32° from N_z in {210} plane for partial birefringence .0075 from Emmons' Plate 10a

Index of oil =	1.6819	25°C.
Temperature—		24°C.
Temp. Correction =	+.0006	
	1.6825	
$N_z - N_{z'}$ =	.0021	
N_z	1.6846	
Correction for method	−.0002	
N_z =	1.6844	
$N_z - N_y$ =	−.0047	
N_y	1.6797	
$N_y - N_x$	−.0056	
N_x	1.6741	

C. BIREFRINGENCE MEASUREMENTS

Path difference for thin plates of a mineral can be measured with a Berek compensator with an accuracy of about .0002 in terms of birefringence. To obtain birefringence from path difference, thickness of the plate must also be measured, and this presents the greatest difficulty. Various focusing methods were tried and gave large errors on thin plates. One may focus with a high-power objective on the top of the

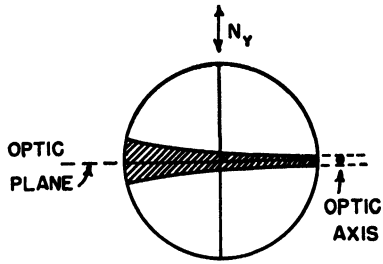


FIGURE 2.—*Use of single optic-axis interference figure on pinacoidal parting fragment to locate optic plane and hence N_y direction*
After Slawson and Peck (1936)

plate and then through the plate on its bottom surface correcting for index of refraction. Or one may focus on the top and then to one side on the glass plate on which it is resting. Focusing errors were of the order of .00n mm. Using a plate 1 mm thick or thicker such errors would be acceptable, but on plates .01 to .04 mm thick errors of 10–40 per cent result.

Mixing the mineral powder to be determined with quartz and using the quartz to solve for thickness seemed to be the only means of getting the desired accuracy. Mixtures of quartz and the mineral cemented by balsam were ground to thin sections, but the quartz was more (or rarely less) resistant to grinding than was the mineral powder. Thus the two were not exactly the same thickness. This could be overcome only by using a hard bakelite cementing medium and grinding and polishing both sides of the mount on a hard, very flat, steel lap.

The preparation procedure is as follows. A well-sorted, clean, crushed quartz with average grain diameter 0.2 mm was mixed with finely ground bakelite powder and the mineral in the ratios by volume of 1:1:1. The mixture is spread in a thin layer in the center of the plug (Fig. 3). The sleeve of the pressure-mounting apparatus is lowered over the plug, and bakelite powder is poured onto the sample until the sample is covered by 2 or 3 cm of bakelite. The piston is inserted above this in the sleeve, and a pressure of 25 psi placed on it. The cylinder is then heated slowly to 90°C., the melting temperature of bakelite, at which point pressure is increased to 4000 psi, and heating is continued to 150°C. The sample is then removed and cooled under tap water.

About .01 mm is ground off the bottom of the bakelite plug containing the sample, using MgO 302½ abrasive on a flat steel lap and an oil lubricant. Then it is further

polished for about 1 minute with MgO 305 abrasive. The sample is then cemented to a glass slide polished face down, excess bakelite sawed off, and the sample ground down to about .05 mm. Grinding and polishing are repeated as above on the upper surface of the slide, and the sample is ready for use.

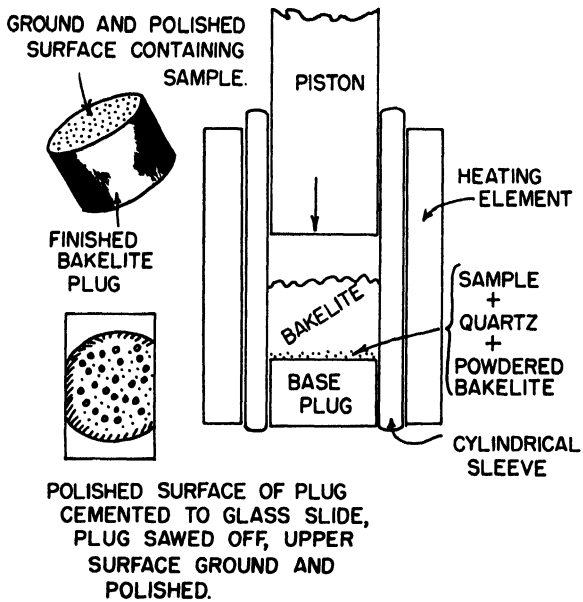


FIGURE 3.—*Preparation of slides for birefringence determination*

The slide is mounted on the universal stage, and a grain of the mineral with the optic plane nearly horizontal is found. After orienting this grain its path difference is measured. Five or six convenient quartz grains around it are measured to determine thickness. A sketch of the grain measured and surrounding quartz grains is made using a grid ocular and drawing the field on graph paper (fig. 41). Thickness can be determined to ±.0002 mm, and birefringence of the mineral can be determined with comparable accuracy. One such measurement of birefringence takes about half an hour. To minimize random errors and to average out slight differences resulting from variation in composition of the grains making up the analyzed sample, 10–15 measurements were made. A curve was constructed to convert Berek-compensator readings on quartz directly to thickness and thereby save time in the computations. The birefringence of quartz was taken as .00910 at 5500 Å. Most of the measurements were made with an ordinary lamp with maximum intensity at about this wave length. For materials with high dispersion of the birefringence a sodium-vapor lamp was used instead. This was not necessary in the minerals examined in the present investigation.

D. OPTIC-ANGLE MEASUREMENTS

The precision and accuracy of optic-angle measurements has recently been discussed fully by Fairbairn and Podolsky (1951). In the present investigation all optic-angle measurements were made by direct rotation on the five-circle universal stage from one optic axis to the other and application of the usual corrections. The precision obtained was slightly lower than that given by Fairbairn, probably for two reasons: (1) the minerals used by Fairbairn were more nearly ideal materials for measurement, and (2) the thickness of slides used varied somewhat so that distance between the upper and lower hemispheres was not exactly that which made the outer surface of the hemispheres a perfect sphere.

For the plagioclase and orthopyroxene in which the optic angles are large either the obtuse or acute bisectrix was measured. No consistent error was noted between the larger and smaller angle measured. For the plagioclase better results can be obtained on slides 2 or 3 times normal thickness. Conoscope observations such as Hallimond (1950) suggests were not used. This method, however, has been found to be excellent on minerals such as sphene or thick plates of micas. The path difference must be such that the isogyres are sharp and narrow. It is also convenient for rapid measurement of dispersion of the optic axes. It has advantages if the optic angle is very small and difficult to measure by orthoscopic observation.[1]

E. EXTINCTION-ANGLE MEASUREMENTS

1. PLAGIOCLASE

Extinction-angle measurements on plagioclase have been far more popular among petrologists than the information so derived warrants. To begin with existing curves for calcic plagioclase are in error by 5–8 per cent in An because they were originally based on inadequate chemical analyses. Secondly the measurements seem to have certain inherent errors which make their usefulness doubtful except for rough estimates of composition. These errors can be reduced by making a very large number of observations, but this number is so large that other methods for composition determinations seem preferable.

Offhand it would appear that by setting {010} vertical and N.–S. on the universal stage and observing extinction angles on Carlsbad-albite twins for 10-degree intervals of rotation on the control axis very good results might be obtained. In 1939–1940 more than 25,000 such observations were made by the writer with rather discouraging results considering the effort expended. The observations consisted of measuring the extinction angle on each of a pair of albite twins first on one side of the Carlsbad composition plane and then on the other. These four angles were each read four times. The observations were repeated at nine positions of rotation of the control axis from 40°N. to 40°S., and the whole procedure repeated on about 10 crystals to get an adequate average consistent with the chemical composition of the sample. On the basis of such measurements improved extinction-angle curves for calcic labradorite and bytownite were constructed (Fig. 14). In Figure 4 the distribution of points for

[1] *See* Smith, Appendix 3, section on Error from strain in measuring $2V$ in powders on universal stage.

observations on one sample is given. The scatter of the points shows clearly the unreliability of measurements at a single setting on a single grain.

Although the albite composition plane can be set vertical within ±1°, the extinction angles on the albite twins were usually not symmetrical. A 2-degree difference

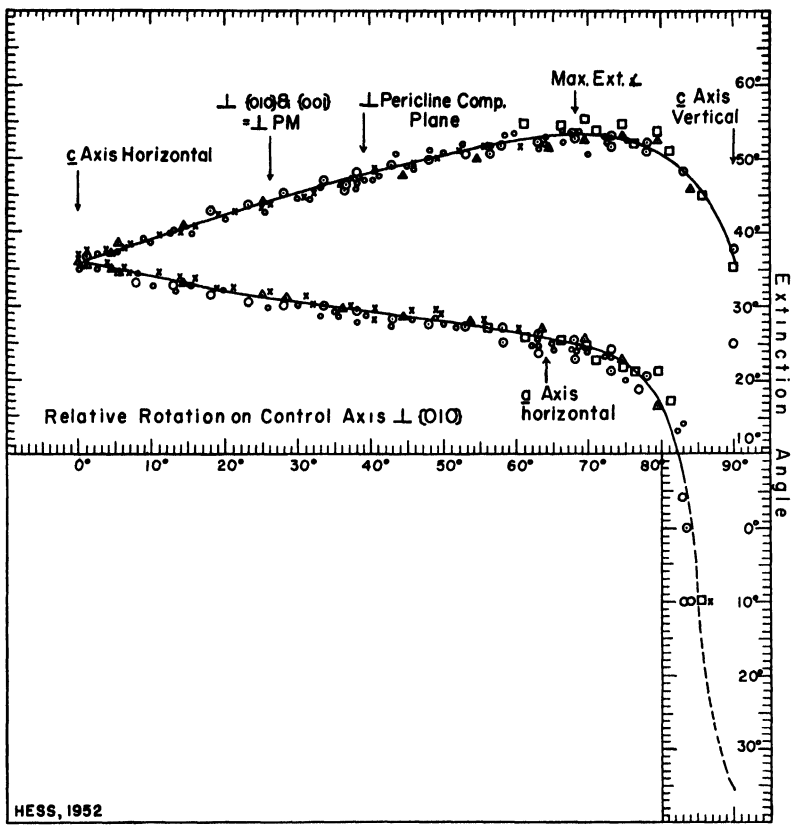


FIGURE 4.—*Extinction angles measured on plagioclase EB 18, An 85.9, in zone* ⊥ {010}, *one of the samples used in construction of Figure 14*

was common, and a 4-degree difference not unusual. It may be, as R. C. Emmons once suggested, that the twin lamellae differ slightly in composition. But it was also noted that, if extinction angles were read on the same lamellae by rotation both to the right and to the left of the crosshair, the positions were not 90° apart. A departure from 90° in this case, of 2° ,was also common. The explanation for this phenomenon is not apparent. Could it be that, in passing through the numerous glass and mounting-medium interfaces on a tilted universal stage, the plane of polarization is rotated slightly? The fact that this difficulty exists is clear whatever its cause. One

can average out this error and make a good composition determination with 1500 observations, but this is hardly practical.

In plagioclase such as that from the Stillwater complex one set of albite lamellae is commonly very narrow so that it is difficult to observe the extinction position on them accurately.

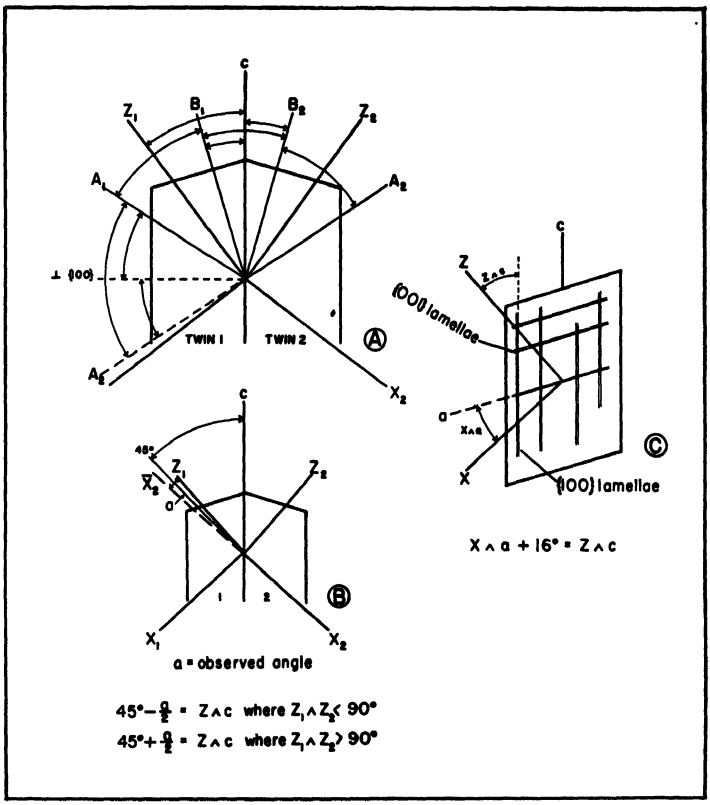


FIGURE 5.—*Diagram to illustrate angles measured on twinned augite in plane* ⊥ {010}

2. CLINOPYROXENE

Difficulties were also encountered in trying to measure directly $Z \wedge c$ of clinopyroxene. The change of $Z \wedge c$ with composition of the Stillwater clinopyroxene is small enough to be neglected. Nevertheless the observations show 3° or 4° variation from grain to grain in attempts to measure this angle on the {010} face. Part of this error may be due to the difficulty of locating the *c*-axis accurately, but probably the same unexplained error as found in the case of the plagioclase is the reason.

Fortunately a satisfactory indirect method to measure $Z \wedge c$ on Stillwater clinopyroxene is available. This stems from the fact that accumulated augite crystals in

the Stillwater are almost invariably pair twins. Thus it is possible on most crystals to measure the angle between an optic axis in each twin or in many cases two optic axes in one and one in the other. With the optic plane vertical and N.-S., rotation on the control axis not only brings these several optic axes in the twins to vertical but

TABLE 2.—*Angles measured on twinned augite EB 175 in {010} plane*

Grain	Observed								Calculated	
	OA_A^1	OA_B^1	OA_B^2	*c*-axis vert.	$B_{x_{a_s}}$	$2V$ or $1V$	$OA_B \wedge^1 OA_B^2$	$OA \wedge$ *c*-axis	$Z \wedge c$	$2V$
1	31½ ↑	¾ ↑	—	15¾ ↑	25½ ↓	26¼	30¾	15⅜	41¼	52½
2	34¼ ↑	4¾ ↑	—	22 ↑	—	—	29½	14¾	—	—
3	36½ ↓	16½ ↑	43¼ ↑	29¼ ↑	—	53	26¾	13⅜	40	53
4	38¾ ↓	12 ↑	43 ↑	25¼ ↑	—	50¾	31	15½	41	50¾
5	—	—	—	—	—	—	—	—	41¾	—
6	18 ↓	—	—	2 ↓	—	—	32	16	—	—
7	—	—	—	—	—	—	—	—	40¾	—
8	—	10¼ ↑	18¾ ↓	3 ↓	—	—	29	14½	—	—
9	—	—	—	—	—	—	—	—	41½	—
10	18 ↑	34½ ↓	47¾ ↑	32 ↑	—	52½	29¾	14⅞	41⅛	52½

Weighted average $Z \wedge c = 41°$; $2V = 52°$.

Arrows indicate directions of rotation on control axis.

These are routine measurements with no attempt to maintain better than average precision. The angles $OA_A^1 \wedge OA_A^2$ and $OA_A \wedge \perp \{100\}$ might also have been measured, but no grains favorably oriented for such observations were noted in the first 10 crystals measured in this particular case.

also to positions at which the interference colors of the twins become identical, and thus where the *c*-axis is either vertical or horizontal. This is a very sensitive method of finding the exact position of *c*. Figure 5 shows the various angles which may be measured, and Table 2 gives a series of observations on one thin section showing how they may be combined to obtain both $Z \wedge c$ and $2V$.

F. DENSITY DETERMINATIONS

Difficulties in making density determinations are nicely summed up by Brian Mason (1944). Most of the density values in the present report were determined at the Rock Analysis Laboratory, University of Minnesota. The method used there is that described by Goldich and Parmelee (1947). For several months the writer experimented with density determinations in hopes of developing a better method for use on small quantities of powders.

For pyroxene the following method was used. The powder (60 mesh) was suspended in 20 cc of clerici solution. The solution was diluted by adding drops of water until the grains just sank. A small glass bulb filled with mercury (wt. in air 7 gms) was then weighed in the liquid. Drops of clerici were added until the grains floated, and the bulb was weighed again. The volume of the bulb had been determined previously by weighing in air and in toluene of known density. The main difficulty involved with this method is the long delay to determine whether the grains which are suspended after stirring are going to float or sink. During this interval tempera-

ture changes or evaporation of water from the clerici complicate the measurements. An improvement was developed by N. M. Bass using a centrifuge to speed up the procedure, and a density balance was substituted for the mercury-filled glass bulb.

The procedure with the centrifuge is to fill three tubes with the heavy liquid. One is used as a dummy to bring the bob of the density balance to the same temperature as the liquid. The mineral is placed in one of the two tubes, and both are placed in the centrifuge. If the mineral floats a drop of water is added to the clerici solution containing the mineral, and centrifuging is repeated. Henceforth water is added to each tube drop by drop until the mineral sinks. Both tubes are then corked to prevent evaporation. The bob of the density balance is immersed for a minute in the dummy solution to bring it to the temperature of the solution. The densities of both the liquids in the other tubes are then determined. The tube containing the mineral has one more drop of water than the second tube, and the density of the mineral lies between the values for the two liquids. The measurement can be made to about $\pm.003$ g/cc.

G. X-RAY MEASUREMENTS

The measurements of cell dimensions were made on pyroxene powders with a Norelco X-ray spectrometer, goniometric radius 17 cm. $FeK\alpha_1$ or $CuK\alpha_1$ radiation was used with wave lengths assumed as 1.93597 Å and 1.54050 Å respectively. The peaks obtained from the pyroxene powders were compared to silicon assuming $a_0 = 5.43062$ Å or, earlier in the investigations, to quartz, $a_0 = 4.9131$ Å and $c_0 = 5.4046$ Å. The precision of the cell dimensions given is about $\pm.004$ Å.

Early in the investigation elaborate precautions were taken to reduce preferred orientation and get a smooth flat upper surface on the sample (Hess, 1952). This was done by mixing the sample with liquid bakelite on a glass slide and baking it until hard. The upper surface was then ground down to a smooth plane on a flat steel lap. These precautions do not seem to be necessary and actually have adverse effects on the results. Dilution by bakelite decreases the peak heights, resulting in considerable loss of accuracy on the weaker peaks. The surface was not as flat as an ordinary glass slide. The bakelite after cooking had considerable stress in it so that after a few days some of the glass slides burst. An internal standard was used at first, but this was abandoned too in the later work. Again the effect of dilution of the sample was disadvantageous, as was interference between peaks from the standard and sample.

The method finally found to be best was to make a smear of the sample diluted with acetone directly on the glass slide. The acetone evaporates in a few seconds. With a little practice a smooth layer about .01 mm thick could be spread on the slide. Preferred orientation of the pyroxene powders does seem to be serious when so prepared. Before and after measurement of a given line, a line of the standard (on a slide prepared in the same manner) of similar 2θ was measured, and the correction to be applied to 2θ was obtained. It is important to determine the correction by means of the standard at the time the sample is measured. Changes in room temperature or warming of the goniometer by heat from the X-ray tube underneath make appreciable changes in the necessary correction.

In the case of the orthopyroxene, the cell dimensions of an enstatite were first

measured approximately. On the basis of these values all possible d_{hkl}. values for an orthopyroxene were calculated. The entire record from $2\theta = 150°$ to $2\theta = 20°$ was then indexed, and the most favorable peaks for precise measurement were chosen. The indexing was done by the method cited by Henry, Lipson, and Wooster (1951, p. 178). A series of records on samples with increasing iron content was then indexed by comparison.

In practice it was found convenient to determine a_0 and b_0 on 12, 0, 0 and 060 first, then substitute the values in 250, 420, 610, 650, and 14, 5, 0. A re-estimate of a_0 and b_0 was then made to give the best fit for those peaks in the above list which were of good quality. Using the a_0 value already determined, c_0 was first estimated in 004 and then in 202 and 502. Similarly 521, 531, or 11, 3, 1 were used to get a value for c_0 substituting predetermined a_0 and b_0 vaules.

Some Stillwater and Bushveld clinopyroxene was investigated by L. C. Coleman and the writer in a general survey of variation of dimensions in this group with composition. These results are still to be published.[2] Diopside was investigated with a Weissenberg camera, and all spots were indexed. Next the X-ray powder pattern from the Norelco instrument record was indexed on the basis of the Weissenberg results. The a_0 and b_0 were normally measured first on 600 and 060, then substituted in 150, 350, 750, 330, and 510, and re-estimated to make a better fit as before. Quantity c_1 was determined first on 041 and 062. Subtracting the pairs 531-53$\bar{1}$ and 311-31$\bar{1}$ gave values for β and hence a_0 and c_0. Using the a_0, b_0, c_0, and β as determined, these could be checked and adjusted if necessary by comparing calculated and observed d_{hkl} values for 531, 53$\bar{1}$, 311, etc.

Results from a few peaks in each sample were anomalous. Normally this could be attributed to overlap of the peak measured with some other peak or peaks.

The reconnaissance traverses for indexing were usually run at 4 minutes per degree 2θ and a chart paper speed equivalent to 1 inch per degree 2θ. Precise measurement of individual peaks was made at either 8 minutes per degree or 4 minutes per degree. Each peak was measured several times. The mean of these measurements is good to at least $\pm$ .02° and in most cases to $\pm$.01°.

H. ROSIWAL ANALYSES

Rosiwal analyses to determine volume per cent of the various minerals in a rock were made with an ordinary mechanical stage. The data were recorded in vertical columns on a piece of ruled paper, and the length of traverse across *each* grain was indicated separately. With practice this is not much more tedious than using an integrating stage or a point-count system, but it has the advantage of giving average grain intercept values as well as volume per cent. Average grain intercepts are a function of grain size and the only means available to express grain size in reasonably quantitative terms.

Grain-size deductions from average grain intercepts are significant only for the accumulated crystals and are relatively meaningless for the interstitial minerals which have crystallized from the pore-space liquid of the mush.

[2] A preliminary survey of the clinopyroxene group was given by Kuno and Hess (1953).

For chemically analyzed rocks three or four slides were traversed for a total traverse length of 30–40 cm. To get averages for the compositions of the various zones of the complex and averages comparable to the bulk chemical composition of the composite samples I and II (Table 33), one thin section of each of about 100 rocks was traversed. Although the error for a given rock sample is rather large in this case, the large number of samples used averages out random errors.

I. MINERAL SEPARATIONS

Nearly all the separations were made magnetically using a Frantz Isodynamic Magnetic Separator. For chemically analyzed minerals the sample was chrushed initially to pass on 80-mesh screen, and the −120 fraction was removed. The dust and fines were eliminated by washing and decanting. Removal of dust is important for good separations. Magnetite and iron filings from the crushing process must be taken out before placing the sample in the separator. This is accomplished with a bar magnet wrapped in thin paper. A trial run was made to determine the current at which the mineral in question just changed from the less magnetic to the more magnetic side of the split trough. Runs are made at settings 0.1 to 0.2 amperes below and above the cross-over point of the mineral to be separated, thus removing all the very much more magnetic and very much less magnetic fractions. The process may be repeated several times, each time narrowing the interval between the cut off above and below. If a clean concentrate is not obtained at this point, the mineral is crushed to −150 and +200 mesh, washed, and the process repeated. If impurities are still present further purification with heavy liquids may be necessary.

Clinopyroxene is less magnetic than orthopyroxene so that separations of the two ordinarily present no difficulties. The separation of pigeonite and augite is generally very difficult. The mass magnetic susceptibility curves for the two minerals have different slopes but intersect. No separation can be effected if field strengths near the values where the curves intersect are used. Change in the fore–aft tile of the separator is necessary so that the separation may be made at a higher or lower field strength.

Biotite commonly sticks to other mineral grains and may be difficult to remove particularly in cold, dry weather. This difficulty is caused by electrostatic charge on the flakes. The most effective means of eliminating biotite in this case is to pour the sample on and off sheets of paper repeatedly. The biotite tends to stick to the paper.

The magnetic separator was also used to make rapid separations of small quantities of minerals for index of refraction work. In this case a pure separate is not necessary. It is a great saving in time during index of refraction work to have a good concentrate of the mineral to be determined. If about a gram of the rock is crushed and poured through the separator, splitting it into four or five magnetic fractions, the concentrate usually is satisfactory. Such a rough separation may be made in 2–3 minutes.

Chapter III
Mineralogy

A. INTRODUCTION

The mineralogy of the Stillwater rocks is relatively simple. There are only five primary essential minerals: olivine, two types of orthopyroxene, a clinopyroxene of the augite species, and calcic plagioclase. The only abundant secondary mineral is serpentine, and it is limited to rocks originally olivine bearing. The near absence of secondary alterations and the moderately coarse grain size make the rocks ideal for mineral separations for chemical analysis and optical study.

Minerals from the Bushveld complex and Great Dyke were also analyzed and studied. The earlier formed minerals of the Bushveld and Great Dyke are virtually identical to their counterparts in the Stillwater. The former two intrusions supply the continuation of the pyroxene trends into the more iron-rich field, which cannot be seen in the Stillwater.

The treatment of mineralogy in this chapter is neither exhaustive nor balanced. It represents the status of an uncompleted investigation. The olivine was not investigated inasmuch as it is simple in composition, and its optical properties *vs.* chemical composition are relatively well known, although it would be interesting to determine the content of such minor constituents as Ca, Mn, Ni, etc., with respect to variation in the Mg:Fe ratio. A disproportionate amount of work was done on the pyroxene, much of which has already been published. Much time was spent investigating the properties of the calcic plagioclase, but compared to the effort expended the results were meager and not in satisfactory form for publication until Smith's work was completed in 1954.

Throughout this report such conventions as $An_{60}Ab_{40}$, $Fo_{90}Fa_{10}$, and $Wo_{40}En_{40}Fs_{20}$ refer to *molecular* and not to *weight* per cent. Wo, En, and Fs where used to describe pyroxene as a rule refer to norms of chemical analyses of rocks. In all other cases Ca, Mg, Fe ratios are used to indicate the composition of pyroxene either from chemical analyses of minerals or as deduced from optical properties. The Fe in the above ratio includes both Fe^{+2} and Fe^{+3} ions. This is done because the optical properties of the pyroxene more closely reflect the *total* iron content, and thus the composition estimates can be made more satisfactorily on this basis than on one neglecting the Fe^{+3}. In the Stillwater complex the Fe^{+3} content of the pyroxene is quite small so that Ca-Mg-Fe values are almost the same as Wo-En-Fs values would be. The orthopyroxene is often described in terms of Mg and Fe, as for example $Mg_{80}Fe_{20}$ or simply Mg_{80}, neglecting the small but almost constant Ca content. Where Ca may be of significance in the discussion, however, it may be given as $Ca_3Mg_{78}Fe_{19}$.[3]

B. ORTHOPYROXENE OF THE BUSHVELD TYPE

This type of orthopyroxene is characterized by fine exsolution lamellae parallel to {100} as described by Hess and Phillips (1940). The exsolved lamellae are composed

[3] In previous papers (Hess and Phillips, 1938; 1940) the Fe^{+3} content of the orthopyroxenes was neglected.

of diopsidic clinopyroxene and are 0.004–0.0001 mm thick. Problems related to this exsolution phenomenon are discussed below following a general treatment of the orthopyroxene. The salient properties of the Bushveld type orthopyroxene may be summarized as follows: cleavage prismatic on {210},[4] partings common on {100}

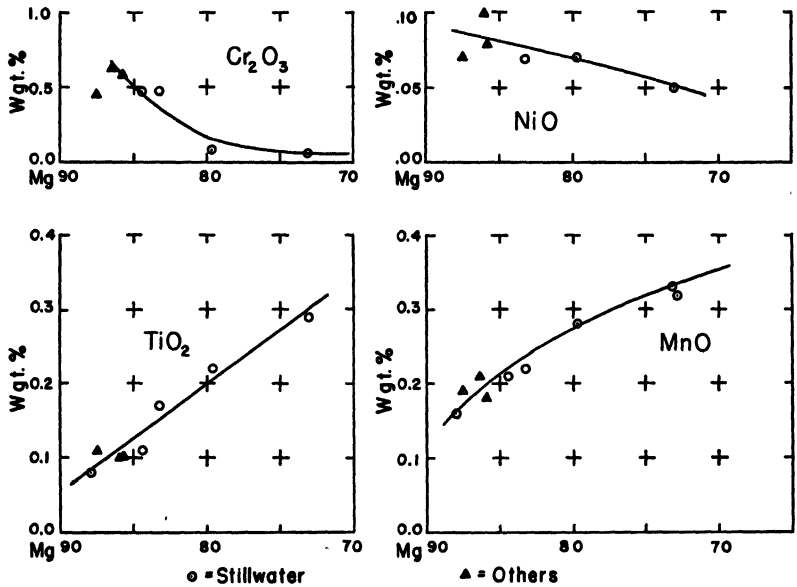


FIGURE 6.—*Variation in minor constituents in orthopyroxenes of the Bushveld type*

and {010} and less so on {001}; color in reflected light pale brown; pleochroism as seen in thick fragments Z pale green, X pink, and Y intermediate between X and Z. In thin sections of normal thickness the enstatite and bronzite are colorless, but the hypersthene is faintly colored. The optic plane is parallel to {100} if the crystallographic orientation of orthopyroxene is set up so that $a > b$. This orientation is required to be consistent with the fixed orientation of the clinopyroxene. Optical properties and chemical compositions of typical Stillwater, Bushveld, and Great Dyke orthopyroxene of this type are given in Table 3 and Figures 7, 8, and 9.

The major variable in chemical composition of the orthopyroxene is the decrease of the Mg:Fe ratio as fractional crystallization proceeds. The orthopyroxene of the Bushveld type ranges in composition from Mg_{88} to Mg_{70} in the Stillwater and Bushveld complexes. The Al_2O_3 content is fairly constant, near 1.7%, which is much lower than in the associated clinopyroxene (2.4%). Cr_2O_3 enters the most magnesian orthopyroxene up to about ½% as compared to 1% in the clinopyroxene. Similarly TiO_2 enters the orthopyroxene to about half the extent that it does the clinopyroxene.

[4] The cleavage of orthopyroxenes is commonly given as {110} but should be modified to be consistent with the unit cell as determined by X rays (Warren and Modell, 1930).

(Kuno suggests (oral communication) that intensity of pleochroism of the orthopyroxene is dependent on the TiO_2 content, not the iron content. This checks with the writer's observations. Orthopyroxene very rich in Fe but poor in Ti is nonpleo-

TABLE 3.—*Superior analyses of orthopyroxenes of the Bushveld type*
(Analyst, L. C. Peck)

Spec. No.	1 7666	2 7704	3 EB 13	4 EB 38	5 EB 130
SiO_2	55.94	55.95	55.36	55.20	53.67
Al_2O_3	1.61	1.53	1.72	1.50	1.65
Fe_2O_3	.97	.50	.54	.84	1.14
FeO	7.15	8.63	9.91	11.86	15.30
MgO	32.12	31.35	29.79	28.14	25.37
CaO	1.48	1.35	1.63	1.93	1.81
Na_2O	.00	.00	—	—	—
K_2O	.00	.00	—	—	—
H_2O+	.09	.07	.18	.30	—
H_2O-	.00	.01	.10	.06	.09
TiO_2	.11	.10	.17	.22	.29
Cr_2O_3	.45	.57	.47	.07	.06
MnO	.19	.18	.22	.28	.33
NiO	.07	.08	.07	.07	.05
	100.11	100.02	100.28	100.47	99.76
$Mg = \frac{100\ Mg^{+2}}{Mg^{+2} + Fe^{+2} + Fe^{+3}}$	87.5	85.8	83.3	79.7	73.1
$Ca = \frac{100\ Ca^{+2}}{Ca^{+2} + Mg^{+2} + Fe^{+2} + Fe^{+3}}$	2.8	2.5	3.2	3.7	3.6
Wt% R_2O_3	3.03	2.63	2.73	2.41	2.85
N_z	1.6802	1.6818	1.6856	1.6893	1.6979
$2V_z$	92	87¾	84	77	68¼
a in Å	18.251			18.273	18.284
b in Å	8.834			8.856	8.870
c in Å	5.193			5.199	5.207

1. Bushveld complex, Transvaal; density in g/cc 3.302/22°C.
2. Great Dyke of Southern Rhodesia
3, 4, 5. East Boulder Plateau, Stillwater complex

chroic.) The variation of the minor constituents with Mg:Fe ratio is shown in Figure 6. With decreasing Mg, Mn and Ti increase progressively. Cr decreases rapidly with increasing Fe beyond $Mg_{80}Fe_{20}$. The Ni content is very small and appears to decrease slowly with increase of Fe. Alkalies enter the orthopyroxene in such small amounts that they are not ordinarily detectable by the usual wet methods of analysis.

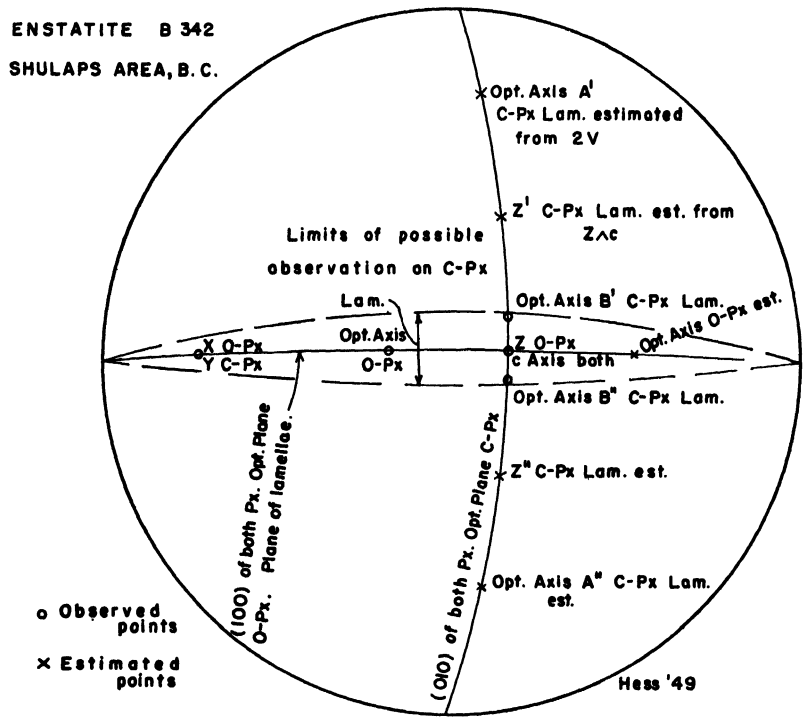


FIGURE 7.—*Stereographic projection of optical directions in enstatite and its included diopside lamellae*

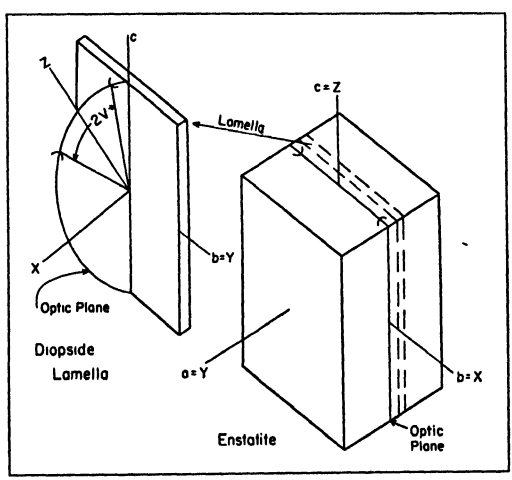


FIGURE 8.—*Block diagram of arrangement of optical and crystallographic directions for enstatite and its included diopside lamellae*

Orthopyroxene of the Bushveld type appears to be identical in bulk composition to that without exsolution lamellae found in most volcanic rocks and hypabyssal intrusives.

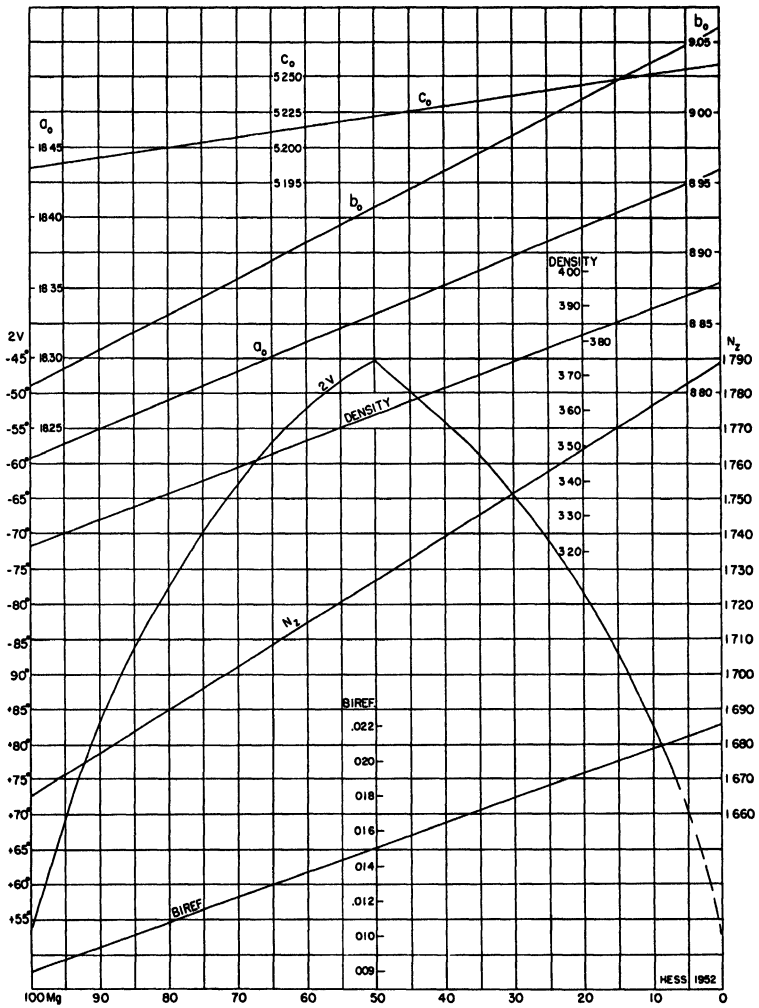


FIGURE 9.—*Properties of orthopyroxenes*

C. ORTHOPYROXENE OF THE STILLWATER TYPE

The orthopyroxene ranges from about Mg_{70} to Mg_{56} in the Stillwater complex and to about Mg_{45} in the Bushveld. The characteristic feature is the development of coarse plates of augite parallel to a plane which represents {001} of a clinopyroxene

predecessor. The plates are 0.1–0.01 mm thick. This pyroxene originally crystallized as pigeonite (clinopyroxene) and later inverted to the orthorhombic form (Hess, 1941; Edwards, 1942). The augite plates exsolved from the pigeonite before inversion, on the monoclinic {001} plane. The presence of such plates in pigeonite which has not inverted was described by Hess and Henderson (1949).

TABLE 4.—*Analyses of pigeonite and inverted pigeonite, orthopyroxene of the Stillwater type*

	1 Hakone	2 Stillwater	3 Bushveld	4 Stillwater	5 New Jersey dolerite	6 Moore County, North Carolina	7 New Jersey dolerite	8 Mull
		EB41	7493	EB89	TR-1-P	Meteorite	LAM-12-P	
SiO_2	52.84	52.81	51.40	50.24	51.53	49.37	51.24	49.72
Al_2O_3	.44	1.65	1.78	4.52	1.64	1.55	1.07	.90
Fe_2O_3	1.06	1.13	1.40	.41	.18	1.83	.05	1.72
FeO	16.89	18.13	20.68	21.81	23.35	26.44	26.85	27.77
MgO	23.51	19.81	19.37	15.81	17.27	15.54	14.85	12.69
CaO	4.06	5.13*	4.07	4.39	4.47	4.60	4.31	3.80
Na_2O	.19	.05	.11	.18	.10	—	.02	.23
K_2O	.00	.00	—	.04	.00	—	.02	.12
H_2O^+	—	.54	.37	1.42	.32	—	.36	1.27
H_2O^-	.22	.16	.04	.04	.14	—	.18	.08
TiO_2	.22	.37	.30	.69	.51	.62	.50	.85
MnO	.56	.38	.48	.43	.49	.37	.55	.98
	99.99	100.16	100.00‡	99.98	100.00‡	100.32	100.00‡	100.13
Ca	8	10†	8.5	10	9.5	10	9.3	8.5
Mg	64	58	57	50	50.8	46	44.4	39
Fe	28	32	34.5	40	39.7	44	46.3	52.5
Analyst	Nagashima	Ellestad	Ellestad	Ellestad	Peck	Henderson	Peck	Radley

* CaO high because of augite impurity
† Estimated augite impurity subtracted
‡ Known amount of augite of known composition subtracted and analysis recalculated to 100%
References: 1. Kuno and Nagashima (1952); 5 and 7. Hess (1949); 6. Hess and Henderson (1949); 8. Hallimond (1914)

In examining a slide of the norite from the Bon Accord quarries north of Pretoria in the Bushveld, the writer discovered a single small grain of uninverted pigeonite enclosed in augite. The other orthopyroxene in this rock was of the usual Stillwater type.

The general properties of the orthopyroxene host are those normally found in orthopyroxene and will not be further elaborated on here.

Orthopyroxene of this type is not uncommon. It is found in the Bushveld, Great Dyke, Duluth gabbro, Skaergaard, Adironack syenites, and most large plutonic intrusions of basaltic composition. It is not ordinarily found in dolerites such as the Karroo, Palisades, etc., where its place is taken either by uninverted pigeonite or a third variety of orthopyroxene which might be called orthopyroxene of the Palisades type.

In some cases orthopyroxene of the Stillwater type has irregular distributed coarse

rounded blebs of augite rather than regular plates parallel to the monoclinic {001} plane. This is particularly true of the Bon Accord gabbro orthopyroxene (north of Pretoria in the Bushveld) and of many of the gabbros of the Main Zone of the Bushveld. In some of the above cases the blebs show a definite tendency toward elongation on the monoclinic {001} plane, but some also show elongation parallel to {100}.

The bulk chemical composition of this pyroxene is identical with that of pigeonite with which it is compared in Table 4. The main chemical difference from ordinary orthopyroxene is in the Ca content which is about three times as high. Ca amounts to approximately 9½% of the total Ca plus Mg plus Fe in contrast to about 3% in orthopyroxene of the Bushveld type. The content of minor constituents and their variation with increasing Fe is consistent with a continuation of the values found in orthopyroxene of the Bushveld type.

D. EXSOLUTION LAMELLAE AND CALCIUM CONTENT OF ORTHOPYROXENE

The literature of the past 30 years is replete with descriptions of "lamellae", "striae", and twinning in orthopyroxene (Bushveld type). All these refer to the same phenomenon with different interpretations of its nature. Henry (1942) summarizes most of the pertinent literature to that date and takes exception to the proposal by Hess and Phillips that the "lamellae" are exsolved diopsidic clinopyroxene. Scholtz (1936) suggested that the laminated appearance of the orthopyroxene found in the thick dolerite sills of Pondoland resulted from a disorientation of thin plates of the orthopyroxene parallel to {100} by a slight rotation about the *b* axis. This was based solely on optical observations. Hietanen (1947) came to a similar conclusion on similar evidence for hypersthene from the Turku district of Finland. Henry (1942) took X-ray oscillation photographs of this type of orthopyroxene and concludes as follows:

> "In the photographs taken about the *c*-axis, ... there are additional spots forming a complete oscillation photograph of a similar crystal with its *c*-axis inclined at about 7° to that of the main crystal in the (010) plane. The spots of this tilted oscillation photograph match those on the layer-lines of the main photograph and are connected to them by Debye-Scherrer curves on which are superposed, in some cases, the spots from other preferred orientations. These photographs show that most of the subsidiary lamellae, which produce the tilted oscillation photographs, are inclined to one side of the main crystal and that one or two orientations are strongly preferred. But the elongation visible in the strong spots shows that a small proportion of the material is inclined on both sides of the main crystal at angles up to 15° from it. From this evidence it is clear that:
>
> (a) The lamellae are orthopyroxenes of the same composition as the main crystal.
>
> (b) The lamellae are inclined to the main crystal in the (010) plane of the latter. (In some cases a slight inclination of the *b*-axes of the two sets of lamellae can be detected, so that the displacement is not always strictly in the (010) plane).
>
> (c) The structure is not a regular twin structure because there is a range of orientations."

The writer took about 20 X-ray oscillation photographs which corroborate Henry's observations. Henry, however, is incorrect in ascribing the observed disorientation to the fine lamellae. Actually the disorientation is entirely in the orthopyroxene host material. Apparently the material of the lamellae is present in too small amount to give X-ray spots in either Henry's or the writer's X-ray photographs. One can easily observe the slight disorientation in the orthopyroxene host under the microscope which is essentially what Scholtz and Heitanen had already done. The lamellae differ

in chemical composition from the host and have the optical properties of diopside.

Leech (1953) made the following significant observation in his study of the ultramafic rocks in the Shulaps Range of British Columbia. The fresh orthopyroxene from these rocks contains the fine lamellae characteristic of orthopyroxene of the Bushveld type. Where the rocks are serpentinized the host enstatite is much more susceptible to replacement than are the lamellae. Cores of enstatite with lamellae are surrounded by serpentine, into which the lamellae extend. Serpentine pseudomorphs after enstatite contain the fine lamellae intact and unreplaced, although the enstatite has disappeared. The writer has since found the same relationship in enstatite from serpentinized ultramafic rocks from a number of other localities. There can thus be no question that the lamellae have a different composition from that of the orthopyroxene host.

The indices of refraction of the host and lamellae overlap in range, but the birefringence of the lamellae is about 2½ times that of the host. Birefringence measurements made on the {010} plane of an orthopyroxene by Berek compensator are given in Table 6. Inasmuch as (4) in the table was taken from optical-property curves for orthopyroxene, it is subject to a probable error estimated to be 0.0005 and certainly not more than 0.001. The true birefringence of the lamellae would, therefore, be subject to about twice the error for the orthopyroxene birefringence, plus approximately 0.0005 observational error in the Berek-compensator measurement. It is quite evident that the birefringence of the lamellae must be close to 2½ times that of the orthopyroxene host, all errors taken into consideration. Similar measurements were made on several other specimens with comparable results. The birefringence obtained for the lamellae is consistent with observed birefringences of common clinopyroxene.

The Z direction of the lamellae makes an angle of about 40° with the c-axis of the orthopyroxene host in sections of the orthopyroxene cut parallel to {010}. Observed extinction angles in the lamellae for samples from seven localities are given in Table 7. For a given crystal some lamellae have the Z direction on one side of c in {010}, and some have Z on the other side. This 40-degree angle is much larger than ±7° which might have been expected from Henry's interpretation of the X-ray data.

By grinding a basal section of the orthopyroxene to about one-third normal thickness (0.01 mm) universal-stage observations can be made on the lamellae so long as the plane of the lamellae is not tilted more than approximately 10° from the vertical. Fortunately with this much tilt it is possible to locate sufficient of the optical elements within the lamellae to determine the orientation of the indicatrix; with this plus the extinction angle $Z \wedge c$ the optic angle can be estimated to be approximately 60°. The optical directions in the lamellae and host as observed are given in stereographic projection in Figure 7 and also are reconstructed in block diagrams in Figure 8.

These facts leave little room for doubt that the lamellae are diopsidic clinopyroxene. The exsolution takes place in such a way that the lamellae and host have their b and c axes, {100} plane, in common. This is the only plane on which the clinopyroxene and orthopyroxene have the same structure, and it is the plane which might be expected to give the lowest energy interface between the two structures.

In orthopyroxene of the Stillwater type exsolution takes place on the monoclinic

{001} plane before the original phase, pigeonite, inverts to the orthorhombic form (Pl. 3, fig. 1). Why {001} should have been preferred over {100}, {010}, or some other plane in the exsolution of one clinopyroxene from another is not yet clear. The exsolution plane may be related to the ease of migration of the ions in the structure rather than to the differences between the two structures. This is also suggested by the fact that the lamellae in the Stillwater type orthopyroxene are 25 to 100 times as thick as those of the Bushveld type in which the ions must migrate across the SiO_4 chains rather than parallel to them. After inversion of the pigeonite to orthopyroxene the augite lamellae and orthopyroxene host are so oriented as a rule that they have their *b* and *c* crystallographic axes in common just as was the case in the Bushveld type. Where, however, the exsolution has been more irregular, as rounded blebs rather than well-formed lamellae, the crystallographic orientation of the blebs may be as described or may be seemingly random. After inversion the orthopyroxene host may exsolve more diopsidic clinopyroxene parallel to {100} as in the case of the Bushveld type.

The calcium content of the orthopyroxene is of interest petrologically for it may throw some light on conditions of crystallization. In Table 5 the calcium content of many orthopyroxene crystals from different localities and formed under different conditions is given. These are analyses of material of exceptional purity and in which special care was taken in determination of CaO. In chemical analyses in the literature the small CaO content, particularly in orthopyroxene rich in MgO, seems to be subject to rather large errors. The majority of the first nine analyses in Table 5 are of orthopyroxene of the Bushveld type, but analyses of one volcanic hypersthene and one high-temperature metamorphic hypersthene are given for comparison. CaO content varies from the median of 1.65% by only $\pm 0.3\%$, and part of this variation no doubt represents error in analysis plus a minute amount contributed by impurities. Probably 1.65% CaO is near the maximum Ca^{+2} which the orthopyroxene structure can hold at the temperature of ordinary magmatic crystallization. If the environment in which the orthopyroxene crystallizes is deficient in Ca then orthopyroxene with a lower Ca content will result. Analysis MVL 3a from the Stillwater which has only 1.11% CaO is an example of this. In MVL 3a at least half of the orthopyroxene was derived from pre-existing olivine by reaction with surrounding basaltic magma. The olivine contains almost no CaO. In those peridotites which the writer suggested came from an ultramafic magma the CaO content of the orthopyroxene is as a rule exceptionally low. Such a magma if it exists would be very low in CaO and thus could not ordinarily supply the Ca^{+2} ions in sufficient amount to reach the maximum Ca content for orthopyroxene. In the group of orthopyroxene crystals from such ultramafic rocks the average CaO content is only 0.9%. Some orthopyroxene from this type of magma does have the normal amount of Ca. This can be attributed to two causes—fractional crystallization which involves the separation of Ca-deficient olivine and thus enrichment of the residual magma in CaO; and the crystallization of a very small amount of orthopyroxene per unit volume of magma which permits extraction of a large part of the relatively small amount of CaO present. The fact that orthopyroxene in these ultramfic rocks is generally deficient in CaO compared to that crystallized from basaltic or andesitic magmas could support the postulate of

TABLE 5.—*Calcium content of orthopyroxene*

Spec. No.	Locality	Wt. % CaO	Ca	Mg	Fe	Rock type	Plutonic, Volcanic, Metamorphic	Magma type
7666	Bushveld	1.48	2.8	85.6	11.6	Bronzitite	P	Basaltic
7704A	Great Dyke	1.35	2.5	83.8	13.7	Bronzitite	P	Basaltic
J	Bushveld	1.48	2.8	83.5	13.7	Bronzitite	P	Basaltic
EB13	Stillwater	1.63	3.2	80.7	16.1	Feldspathic bronzitite	P	Basaltic
EB38	Stillwater	1.93	3.7	76.7	19.6	Norite	P	Basaltic
R	Bushveld	1.58	3.1	76.4	20.5	Norite	P	Basaltic
EB130	Stillwater	1.81	3.6	70.5	25.9	Gabbro	P	Basaltic
5718	Adirondacks	1.59	3.3	51.3	45.4	Syenite	P; M	Syenitic
DOM I	Dominica Is.	1.56	3.3	51.2	45.5	Andesite	V	Andesitic
Average CaO = 1.60; Range 1.35–1.93; Average Ca^{+2} 3.2 (of total Ca + Mg + Fe)								
From Calcium-Poor Ultramafic Magma(?)								
Web. 26 Gn	Webster, N. C.	0.23	0.4	93.7	5.9	Enstatolite	P	Ultramafic
B323	Shulaps, B. C.	0.79	1.5	91.4	7.5	Enstatolite	P	Ultramafic
Web. 1	Webster, N. C.	0.74	1.4	85.8	12.8	Websterite	P	Ultramafic
Web. 26 Bn	Webster, N. C.	0.57	1.0	87.4	11.6	Bronzitite	P	Ultramafic
B295	Shulaps, B. C.	1.95	—	—	—	Harzburgite	P	Ultramafic
374	Bay of Is., Nfld.	1.06	2.1	74.8	23.1	Hypersthenite	P	Ultramafic?
Average CaO = 0.89; Range 0.23–1.95; Average Ca^{+2} 1.8								
Orthopyroxene Derived Partly from Olivine Reaction with Basaltic Magma								
MVL 3a	Stillwater	1.11	—	—	—	Harzburgite	P	Basaltic
Inverted Pigeonites and Pigeonites								
M.C.	Moore County, N. C.	4.60	10	46	44	Meteorite	P	Basaltic (pigeonite)
EB41	Stillwater	5.13	10.5	57.5	32	Gabbro	P	Basaltic (inverted pigeonite)
EB89	Stillwater	4.39	10	50.5	39.5	Hornfels	M	Basaltic (inverted pigeonite)
M	Mull	3.80	9	40	51	Basalt?	V	Basaltic (pigeonite)
7493	Bushveld	4.45	8.5	57	34.5	Norite	P	Basaltic (inverted pigeonite)
Average CaO = 4.31; Ca^{+2} 9.6								

an ultramafic magma deficient in CaO. Calcium deficiency may also be explained in some cases on other grounds. In the Great Dyke, for example, cooling has been extremely slow. The dike extends downward at least 10 miles below the present outcrops. If the orthopyroxene were held for a long time in the temperature range some-

TABLE 6.—*Birefringence of lamellae in an orthopyroxene of the Bushveld type*
Specimen B 342, Shulaps harzburgite, British Columbia

(1) Path difference in enstatite {010} plane ($N_z - N_y$)	112 mu
(2) Path difference in lamella same grain, same orientation, ($N_z - N_z$ of lamella)	490 mu
(3) Optic angle of enstatite	80° (+)
(4) Birefringence of enstatite consistent with (3) (taken from curves for orthopyroxenes)	0.0095
(5) Partial birefringence of enstatite in {010} plane ($N_z - N_y$) calculated from (1), (3), and (4)	0.0057
(6) Birefringence of lamella calculated from (1), (2), and (5)	0.025

$$\frac{X}{0.0057} = \frac{490}{112} \qquad X = 0.025$$

TABLE 7.—*Extinction angles of the lamellae on {010} in orthopyroxene of the Bushveld type*

Specimen No.	Locality	Rock type	$Z \wedge c$ in lamellae in {010}
7666	Bushveld complex, Transvaal	Bronzitite	40° ± 5°
EB38	Stillwater complex, Montana	Norite	39°–43°
B342	Shulaps area, British Columbia	Harzburgite	40°–41°
WEB	Webster, North Carolina	Websterite	40° ± 5°
59	St. Pauls Is., Labrador	Anorthosite	40°–43°
PG13	Oriente, Cuba	Harzburgite	42° ± 5°
Q324	Thetford, Quebec	Harzburgite	42° ± 5°

what below its crystallization temperature the Ca^{+2} ions would have a much longer time to diffuse through the crystal. Many might reach the boundary of the crystal and escape. This apparently has happened in the case of the Great Dyke. The lowest CaO value for an orthopyroxene from a basaltic magma is that shown in Table 5 for Great Dyke sample 7704A = 1.35%. The Ca^{+2} and other ions lost in the above process add to the interstitial chrome augite present in these rocks.

The amount of CaO in orthopyroxene is also probably a function of temperature of crystallization. Metamorphic orthopyroxene which has crystallized below magmatic temperatures presumably would have less CaO than that from basaltic igneous rocks. Reliable data on this point are lacking. Howie (personal communication) finds very low CaO in orthopyroxene from charnockites. In some basaltic lavas in Japan Kuno and Nagashima (1952) find orthopyroxene with abnormally high CaO. The lavas may well have crystallized at a somewhat higher temperature than equivalent intrusives as a result of loss of part of their volatiles or perhaps of oxidation reactions.

Table 8 gives the number of exsolution lamellae per mm and range of thickness of

the lamellae for several orthopyroxene crystals of the Bushveld type. Both quantities show considerable variation. The finest lamellae are below the resolving power of the ordinary petrographic microscope. Through the courtesy of Professor A. Parpart, equipment which he was using on biological specimens was adapted to make the measurement. A television camera was substituted for the microscope eyepiece, and the lamellae were measured on the television projection at a magnification of 1000.

TABLE 8.—*Thickness, number, and volume of exsolution lamellae in orthopyroxene, Bushveld type*

Specimen No.	Mineral	Locality	Rock type	Magma type	Thickness of lamellae in mm	No. lam. per mm	Estimated vol. diopside	Vol. from chem. anal., per cent
Web. 1	Enstatite	Webster NC	Websterite	Ultramafic	.004–.0005	40	2.0	3
Web. 2	Enstatite	Webster NC	Websterite	Ultramafic	.0015–.00015	100	1.5	3
B323	Enstatite	Shulaps BC	Enstatolite	Ultramafic	.001–.0001	60	1.3	3
7666	Bronzite	Bushveld	Bronzitite	Basaltic	.0008–.0001	250	3.1	5.6
EB13	Bronzite	Stillwater	Feldspathic bronzitite	Basaltic	.003–.0003	112	3.4	6.4

Measurement of the number and thickness of the lamellae gives a fairly good approximation of the amount of diopsidic pyroxene exsolved.

Pigeonite contains about 9½% CaO. With slow cooling about two-thirds of this CaO is removed by exsolution of augite plates parallel to {001}. Thus upon inversion of the pigeonite to hypersthene the CaO content is not above that which the orthopyroxene can hold in its structure. If, however, the cooling is faster, inversion may take place before exsolution. The excess Ca^{+2} cannot enter the hypersthene structure and minute specks of diopside are rapidly expelled all through the orthopyroxene. This is fairly common in hypabyssal dolerites and might be referred to as orthopyroxene of the Palisades type. If cooling is more rapid, particularly in lavas, pigeonite is preserved intact without either exsolution or inversion.

E. AUGITE

The augite crystals have nothing particularly distinctive about them. They have good prismatic cleavage, and partings may be observed on any of the three pinacoids. In reflected light they are greenish brown but they are almost colorless in thin section. No pleochroism is observed in thin section, but thick fragments may show pale green parallel to the *c*-axis and faint pink parallel to the *b*-axis. This effect is due to exsolved hypersthene lamellae parallel to {100} and not to the clinopyroxene host. The most magnesian varieties are also chromiferous and show this in the hand specimen by an emerald-green color.

The optic plane is {010}, $Z \wedge c$ is near 41°, and $2V$ is near 50°. Exsolution lamellae of orthopyroxene are parallel to either {100} or {001}, or both. The optical properties and chemical composition are summarized in Table 9 and Figure 11. Most of the settled crystals are simple pair twins with equal development of the two individuals. The interstitial augite which has crystallized from pore-space liquid is not commonly

twinned. There is a strong suggestion that the pair twins develop when the first few unit cells are forming into the crystal, but that after this initial favorable stage the tendency for additional unit cells to add to the crystal in the twinned position is strongly diminished.

As fractional crystallization proceeds the major variable in chemical composition is a decrease in the Mg:Fe ratio. Ca^{+2} is about 40 per cent of the total $Ca^{+2} + Mg^{+2} + Fe^{+2} + Fe^{+3}$ in the samples analyzed, but it begins to decrease in the most ferriferous samples. No analyses are available of the clinopyroxenes richest in iron from the Bushveld, but in these it might be expected that Ca^{+2} would decrease to a minimum near 30 per cent. The enrichment in Fe does not extend over nearly so great a range as in the Skaergaard; the minimum Mg:Fe ratio is close to 1:1.

The Stillwater clinopyroxene has a consistently higher Al_2O_3 content than the Bushveld and Great Dyke clinopyroxenes. This is peculiar in that the Stillwater and Bushveld magmas seem to be almost identical; the phases separating are the same, and the temperatures of crystallization indicated by pyroxene inversions are the same. The only condition which might have been different in the two cases is pressure, the depth at which crystallization occurred. The writer (1952) suggested that, for orthopyroxene, Al_2O_3 increased with pressure. The effect of substitution of Al^{+3} in the Y position in clinopyroxene results in a marked contraction of the unit cell so that high pressure would presumably favor increased Al_2O_3 in the clinopyroxene. The change in Al_2O_3 in clinopyroxene with stratigraphic height in the Stillwater (or in the Bushveld) is small. The variations in Al_2O_3 in both cases are not much greater that the expectable analytical errors so no firm conclusions may be drawn from them. Inasmuch as a good part of the crystallization is thought to take place near the roof of the magma chamber, the difference in Al_2O_3 between the Stillwater samples and Bushveld samples may represent only thickness of the cover over the complexes. This suggests that the Stillwater complex had a thicker cover. This effect is worthy of further investigation.

TiO_2 is comparatively low in the Stillwater and Bushveld augite as it is in the magma from which they were derived. It increases from ¼ per cent to about ½ per cent with increasing Fe in the samples analyzed. The Cr_2O_3 content is slightly more than 1 per cent in the most magnesian augite but drops off to negligible after Fe^{+2} reaches 10 per cent. Alkalies, largely Na_2O, generally amount to about 0.3 per cent. The NiO content in the earliest augite is about 0.1 per cent and decreases slowly as Fe increases. Mn increases slowly with increasing Fe from 0.1 to 0.3 per cent in the analyzed specimens.

F. EXSOLUTION IN AUGITE

With slow cooling the augite exsolves a lime-poor pyroxene and changes in composition from $Ca^{+2} = 40$ per cent toward $Ca^{+2} = 50$ per cent or in other words toward a diopside-hedenbergite clinopyroxene. In the more magnesian augite the exsolved lime-poor pyroxene is enstatite or bronzite and the plane of exsolution of the lamellae is {100}. This is exactly the same as in orthopyroxene of the Bushveld type but wth host and lamellae interchanged. As the magma becomes enriched in Fe there is a

TABLE 9.—*Chemical analyses of clinopyroxene*

(Analyst, R. B. Ellestad)

	Stillwater clinopyroxene					Bushveld clinopyroxene			
	MVL 101/2	I52	EB175	EB43	EB41	7666	7503	7648	7493
SiO_2	51.43	52.55	52.61	51.83	51.86	54.07	52.92	52.47	51.39
Al_2O_3	3.00	2.98	2.72	3.07	2.33	2.08	2.80	2.28	2.45
Fe_2O_3	1.37	1.17	1.36	1.38	1.60	.56	.85	1.49	1.26
FeO	4.23	3.98	5.85	7.21	9.45	2.53	5.57	6.61	11.63
MgO	19.44	18.21	15.97	16.00	14.50	17.39	16.40	15.71	14.21
CaO	17.86	19.23	20.50	19.21	18.92	22.12	19.97	20.28	18.12
Na_2O	.28	.32	.28	.27	.23	.41	.35	.31	.27
K_2O	.02	.00	.02	.02	.00	.00	.01	.02	.02
H_2O+	.70	.14	.19	.47	.37	.04	.10	.20	.03
H_2O-	.10	.08	.04	.11	.09	.06	.07	.04	.07
TiO_2	.27	.24	.44	.49	.55	.21	.50	.46	.41
Cr_2O_3	1.09	1.18	—	—	.01	.98	.88	.08	—
MnO	.14	.12	.18	.17	.24	.09	.15	.20	.32
NiO	—	.05	—	—	.02	.04	.10	.04	—
	99.93	100.25	100.16	100.23	100.17	100.58	100.67	100.19	100.18
Ca	37	40	42.3	41	39.5	45	41.5	41.5	37.6
Mg	56	52	46	45	42.5	50	48	45.5	41.1
Fe	7	8	11.7	14	18	5	10.5	13	21.3
% Al in Z	4.9	4.6	3.9	4.4	3.9	3.0	4.2	3.7	3.7
Nx	1.6810	1.6805	1.6823	1.6832	1.6870	1.6742	1.6818	1.6820	1.6905
Ny	1.6853	1.6852	1.6874	1.6870	1.6921	1.6800	1.6865	1.6875	1.6951
Nz	1.7057	1.7075	1.7093	1.7100	1.7149	1.7019	1.7085	1.7101	1.7190
Biref. × 10^4	247	270	270	268	279	277	267	281	285
$2V_z$	48¾°	49°	51½°	52°	50¾°	55½°	49°	51¾°	47½°

Disp. B Opt. Ax.	mod.	mod.	weak	weak	mod.	weak	mod.	mod.	strong
Disp. A Opt. Ax.	mod.	mod.	weak	none	weak	weak	weak	mod.	mod.
$Z \wedge c$	40.5°	40.5°	41°	42.1°	42°	40.25°	40.75°	42.5°	44°
Lamellae	{100}	{100}	{100}	{100}	{001} {100}	rare	{100}	{001} {100}	{100} {001}
Twins	—	—	pairs on {100}	pairs on {100}	pair twins on {100}	—	—	pairs {100}	pairs on {100}
Remarks			$2V$ lam. 66° ± 2 $Mg_{75}Fe_{25}$						
a_0 in Å			9.749		9.755				9.756
b_0 in Å			8.926		8.932				8.935
c_0 in Å			5.251		5.248				5.252
β			74°2′		74°5′				74°4′

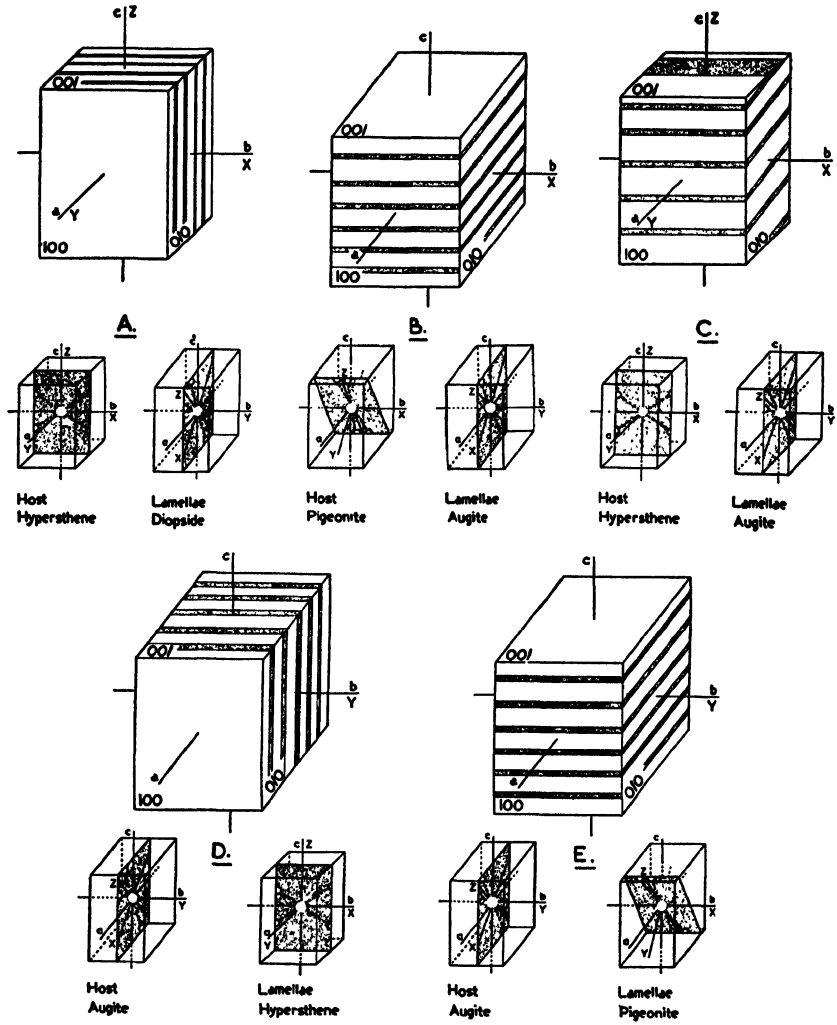


FIGURE 10.—*Exsolution lamellae in clinopyroxenes*
After Poldervaart and Hess (1951)

stage at which the inversion curve orthopyroxene → pigeonite is crossed (Fig. 12). Upon crossing this curve pigeonite instead of orthopyroxene exsolves as lamellae, but the lamellae are parallel to {001}; in other words the situation is the same as in orthopyroxene of the Stillwater type but again with host and lamellae interchanged. With cooling the temperature drops below the inversion curve, and further exsolution takes place, but the exsolved material is hypersthene, and a second set of lamel-

lae form parallel to {100}. The various possibilities for exsolution are summarized in Figure 10 after Poldervaart and Hess (1951). Even though the lamellae parallel to {001} may now be orthopyroxene there can be little doubt that originally they were pigeonite. In the chilled roof-border facies of the Great Dyke augites with the pigeonite lamellae preserved have been found (Hess, 1950).

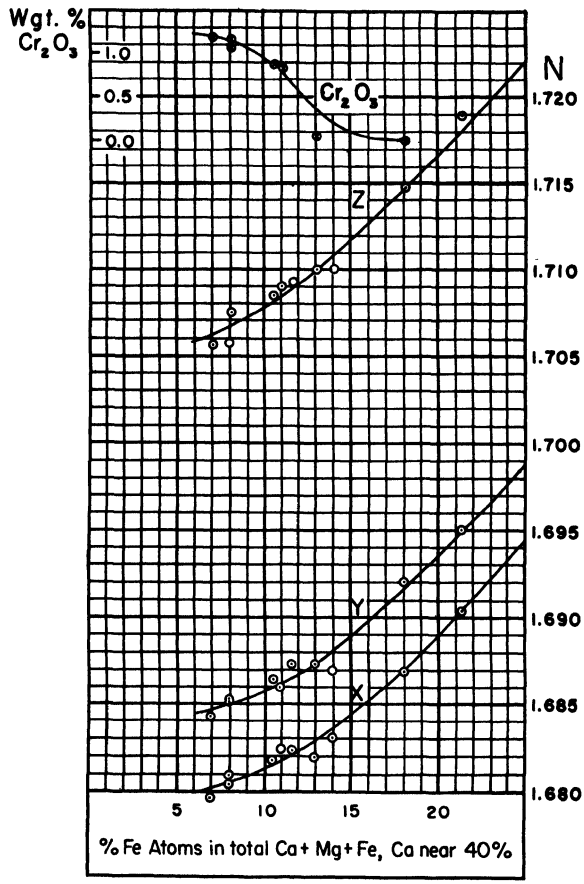


FIGURE 11.—*Properties of augite*

G. PYROXENE AS GEOLOGIC THERMOMETER

Figure 12, taken from Hess (1941), shows how with enrichment in Fe the course of crystallization moves from the field of orthopyroxene, across the inversion curve, into the field of clinopyroxene (pigeonite). The inverstion curve is taken from Bowen and Schairer (1935). On cooling, the pigeonite thus formed inverts to orthopyroxene of the Stillwater type, whereas the orthopyroxene which crystallized as such from the magma becomes orthopyroxene of the Bushveld type. In both the Stillwater and

Bushveld complexes the inversion curve was crossed when the crystals had an Mg:Fe ratio of 70:30. This would indicate a temperature near 1100°C. Inasmuch as all the crystallization up to this point was below the inversion curve, the initial crystallization from the Stillwater and Bushveld magmas must have taken place below

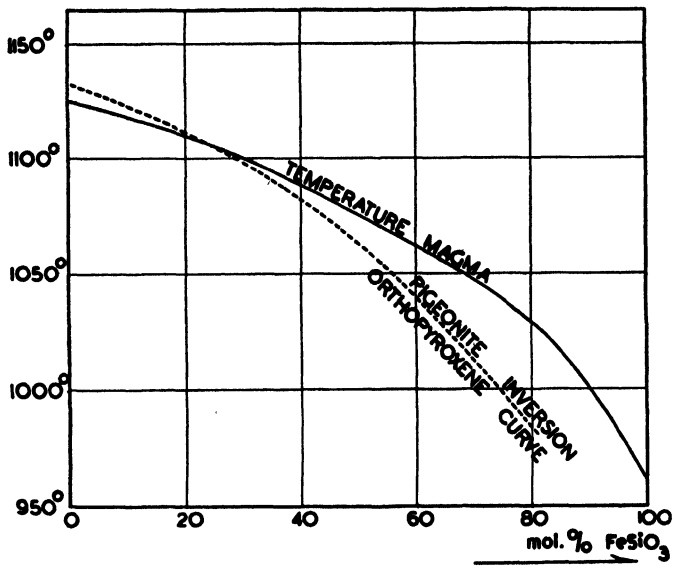


FIGURE 12.—*Pigeonite-orthopyroxene inversion curve* vs. *temperature of magma with crystallization*

approximately 1140°C., the inversion temperature for enstatite to clinoenstatite. In the Stillwater the pigeonite began to crystallize at a horizon represented by the highest outcrop stratigraphically or, as will be seen from later discussion, after about 60 per cent of the magma had crystallized. The temperature change during the first 60 per cent of crystallization is very small.

Figure 13 shows the composition relations between augite and orthopyroxene or pigeonite which crystallized contemporaneously. The augite also furnishes a thermometer. Where exsolution of lamellae from the augite takes place at a temperature below the orthopyroxene → pigeonite inversion curve the plane o exsolution is {100}, and the exsolved material is orthopyroxene. Where exsolution takes place above the inversion curve the material exsolved is pigeonite, and the plane of exsolution of the lamellae is {001}. Exsolution on {001} in the augite appears soon after orthopyroxene of the Stillwater type makes its appearance in the stratigraphic succession. From this information can be obtained the temperature at which exsolution began in those augites in which the transition takes place from lamellae on {100} to lamellae on {001}. Evidently the exsolution began immediately below, and very close to, the crystallization temperature as determined by the appearance of orthopyroxene of the Stillwater type slightly lower in the stratigraphic succession.

H. PLAGIOCLASE

A study of the calcic plagioclase was begun by A. H. Phillips and the writer in 1933. Most of the analyses (11) were completed by Phillips before his death except for a correction to be applied to the alkali determinations. R. B. Ellestad redetermined the Na_2O and K_2O on these samples. Not as much attention was paid to the

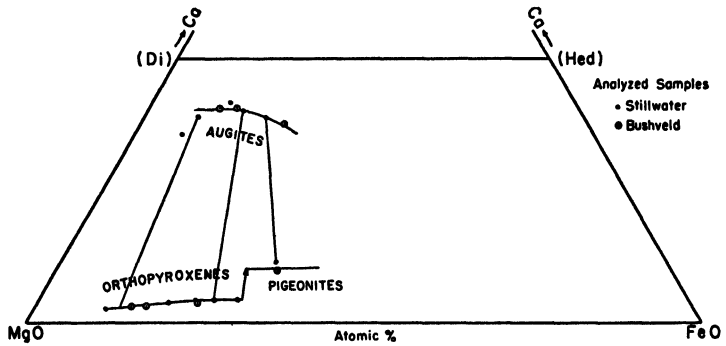


FIGURE 13.—*Course of crystallization of ortho- and clinopyroxene*

minor constituents of the plagioclase as there would have been if the investigation had been undertaken in the last few years. It is hoped that these may be investigated at a later date. J. R. Smith has recently studied the optical properties of the plagioclase from the Stillwater, Bushveld, Skaergaard, and Great Dyke; he has improved the curves and made a general survey of determinative methods. Inasmuch as his results supersede those of the writer the plagioclase will be dealt with briefly here and more extensively by Smith in Appendix 3.

The original study was started in 1933 because there were few superior chemical analyses of plagioclase between An_{30} and An_{90} and because the optical properties of these had not been adequately determined. The situation did not improve materially (Chayes, 1952) until 1954. Current optical-property curves may be in error by as much as 8 per cent in An content.

The fresh plagioclase of the Stillwater, Bushveld, and Great Dyke are generally white in the hand specimen or powder, but those in the ferrogabbros may be yellowish.

The common well-developed cleavages on {001} and {010} are present in all. Partings on {110} are also common in the powders. Thus the powders have tablets representing four planes with respect to the optical indicatrix so that the use of "cleavage" fragments as an aid to determination of optical properties is often not practical.

In the lower part of the Stillwater complex zoning of the plagioclase is so slight that it might be overlooked. High in the complex, plagioclase generally shows a narrow rim of more sodic composition about an unzoned interior. Some of these grains have on their margins a minute amount of a vermicular intergrowth or an exsolution structure, the host being plagioclase and the vermiculae a mineral of much

lower index of refraction, presumably K feldspar. They are too minute for specific determination.

Although zoning is slight, individual settled crystals seem to vary in An content from ±1 to ±4 per cent from the mean. Some of this may be an initial difference between settled crystals accumulating on the floor, but most of it is related to trapped pore-space liquid. In particular in layers where accumulation of crystals has been abnormally rapid, isolated liquid in the pores may react with and add to adjacent plagioclase to make it slightly more albitic. In general such samples were avoided in choosing material for analysis. Even though this was done, every property of the plagioclase must be measured on a number of grains in order to obtain an average comparable to the chemical analysis.

The plagioclase was prepared for chemical analysis by magnetic separation. Inasmuch as this is as a rule the only diamagnetic mineral in the rock a very clean separation could be made. The only impurities present are some minute unidentified specks which appear in little clusters in some grains. All samples had some of this, and some had much more than others. This does not seem to be reflected in the chemical analyses. The manner in which the aggregates of specks are arranged suggests an alteration product, and this in turn suggests that they may be a clay mineral. They are much too small to be identified in thin section or powders and could not be concentrated to examine with X rays. The amount present, estimated to be a fraction of 1 per cent, is also too small to find in X rays of the bulk sample.

Twinning on all the common twin laws for plagioclase has been observed. In twinning on the albite law commonly two or three very fine lamellae make up one component of the albite twins, and the host, a rather broad expanse, makes up the other. Strictly, therefore, one cannot say whether the albite twin lamellae are broad or narrow.

Without making a statistical analysis of the percentages of the various twin laws represented, the plagioclase from the Stillwater and that from the Bushveld do not appear to differ with respect to twinning. But one abnormal case was found in the eastern Bushveld in the region north of Magnet Heights. Here fine, oft-repeated albite twin lamellae are characteristic. Many of the feldspar crystals are slightly bent, and some have been fractured. This appears to be a twinning induced by strain. Samples taken over several square miles all exhibit this peculiarity.

If there is a difference between a high-temperature and a low-temperature calcic plagioclase, then all the samples studied in this investigation should be in the low-temperature group (Köhler, 1949). An examination of a synthetic (high-temperature) anorthite supplied by J. F. Schairer and a natural (metamorphic, low-temperature?) Anorthite described by Subramaniam (1952) show no appreciable difference in the indices of refraction and birefringence of the two, but, in the higher-temperature synthetic, $2V$ is 1.7° lower than in the metamorphic anorthite.

Analyses of 11 Stillwater, 2 Bushveld, 1 Great Dyke, and 1 Skaergaard plagioclase are given in Table 10. All but the last four were made more than 15 years ago. They are all quite satisfactory so far as the An:Ab:Or ratio is concerned, but minor constituents have not been adequately analyzed. There are some small excesses and deficiencies in Al and Si, probably the result of analytical error. SiO_2 is high in BV63

TABLE 10.—*Chemical analyses of plagioclase*

First 11 analyses by A. H. Phillips with alkalies checked by R. B. Ellestad; last four by E. H. Oslund

	S EB18	S C12	S EB21	S EB38	S EB72	S EB32a	B 7510	S EB68	S EB43	S 1b6	S EB42	S EB41	B BV63	GD GD29	SK 1963	
SiO_2	46.34	46.69	46.93	47.67	47.77	47.36	49.06	49.09	49.19	48.74	51.34	52.95	53.58	55.24	56.91	
Al_2O_3	33.36	33.71	33.06	33.46	32.28	33.13	32.14	31.62	31.82	31.49	29.98	29.48	29.07	27.63	26.60	
Fe_2O_3	.54	.46	.57	—	.49	.17	.27	.48	.55	.17	.55	.24	.08	.09	.65	
FeO	—	—	—	—	—	—	—	—	—	—	—	.32	.28	.35	—	
MgO	.38	.43	.53	.35	.54	.42	.20	.39	.41	.91*	.48	.10	.12	.18	.10	
CaO	17.31	16.81	16.53	16.23	16.04	15.89	15.38	15.25	14.95	15.05	13.39	12.70	12.04	10.55	8.94	
Na_2O	1.55	1.61	1.70	2.19	2.28	2.25	2.57	2.71	2.78	2.94	3.70	4.12	4.33	5.22	6.02	
K_2O	.05	.19	.08	.07	.05	.11	.17	.07	.09	.11	.15	.13	.36	.32	.36	
H_2O+	.23	—	.31	—	.59	.14	.13	.18	.28	.53	.35	.10	.24	.21	.11	
H_2O-	—	—	.05	—	.04	—	.03	.05	—	.03	.02	.00	.04	.04	.01	
TiO_2	—	—	—	—	—	—	—	—	—	—	—	.05	.05	.06	.11	
MnO	tr	—	tr	—	—	—	—	—	—	—	—	—	—	—	.10	
SrO	—	—	—	—	—	—	—	—	—	—	—	—	—	—	.05	
	99.76	99.90	99.76	99.97	100.08	99.47	99.95	99.84	100.07	99.97	99.96	100.19	100.19	99.89	99.96	
An	86.0	84.3	83.9	80.2	79.5	79.1	76.0	75.4	74.4	73.5	66.1	62.9	59.4	51.9	44.2	Mol %
Ab	13.7	14.6	15.6	19.4	20.2	20.2	23.0	24.2	25.0	25.9	33.0	36.3	38.6	46.2	53.7	
Or	0.3	1.1	0.5	0.4	0.5	0.7	1.0	0.4	0.6	0.6	0.9	0.8	2.0	1.9	2.1	

* Serpentine veinlets in sample.

S = Stillwater; B = Bushveld; GD = Great Dyke; SK = Skaergaard.

and GD29 probably as a result of the presence of a little quartz in the analyzed material. MgO is high in 1b6 because of minute veinlets of serpentine. The whole group of analyses otherwise shows a rather consistent MgO content near 0.4% and total iron oxides content near 0.5%. MnO is virtually absent, and TiO_2 in the few

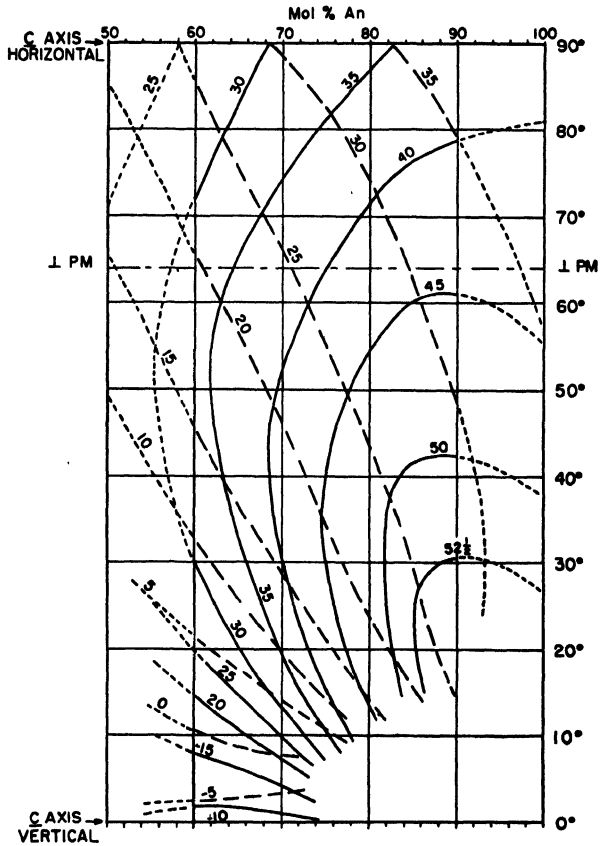


FIGURE 14.—*Extinction angles for Carlsbad-albite twins in calcic plagioclase*

cases determined is near .05%. The water content averaging near 0.3% does not seem to correlate with the slight alteration noted in a few of the specimens. Even though much of this is H_2O+ it seems possible that it is water adsorbed on the surface of the grains in the fine powders. The Or content increases gradually with increasing Ab. Or in analysis C-12 is probably a little high as a result of analytical error. There is a suggestion of somewhat lower Or in the Stillwater samples than in the Bushveld and Great Dyke.

Those optical properties which were measured precisely are given in Smith's Table 54 and Plate 12. Only the more recent determinations are given. These were

made after much experimentation to improve methods and precision of observation. Properties of most of the analyzed samples were measured earlier in the investigation, but the accuracy of these was so much less satisfactory that they have not been listed. Some of the samples show too large a range of variation in An content to make precise optical determinations worth while. It would take so many observations to get a satisfactory average that the task was not undertaken. Extinction-angle curves for Carlsbad-albite complex twins are given in Figure 14. The writer is not entirely satisfied with these curves although they are an improvement over those given in the literature. *See* comments in Chapter 2 on extinction-angle measurements. While the curves are constructed for Carlsbad-albite twins, one can get from them the extinction angle perpendicular to the two cleavages or the maximum extinction angle in the plane perpendicular to {010}.

I. OTHER MINERALS

The quantities of minerals other than the essential ones are exceedingly small except for serpentine as an alteration product of olivine in the Ultramafic Zone. The only accessory minerals observable in the field are chromite and sulfides.

Chromite occurs as an accessory mineral in the lower and middle portions of the Ultramafic Zone and is concentrated in a few layers near the middle of this zone. (*See* various U.S.G.S. bulletins by Peoples and Howland.) The details of the chemistry and mineralogy of the chromites is being studied by E. Sampson.

Sulfides—chalcopyrite, pyrrhotite, and pyrite—occur in appreciable concentrations at some localities near the floor of the complex. They also may appear in traces in rocks throughout the complex. At several horizons near the base of the Upper Gabbro Zone and to a lesser extent near the base of the Anorthosite Zone the concentration is high enough to permit tracing a rusty oxidized layer in the field, but the sulfide content is generally less than 1 per cent. The sulfides at the base of the complex as well as the smaller amounts higher in the complex probably represent precipitation of an immiscible sulfide liquid. The sulfides of the complex were studied by Howland (1932, Ph.D. thesis, Princeton Univ.).

The minute traces of sulfides and other accessory minerals in the gabbros, norites, and anorthosites are mainly found in those rocks in which the essential crystals accumulated more rapidly than usual, and accessories precipitated from the trapped residual liquid in the pore space. The accessory sulfides commonly show a black opaque rim which appears to consist of iron oxide, probably hematite.

Quartz is the most abundant accessory mineral above the Ultramafic Zone. It is present commonly only as a trace but may reach concentrations of 1%. It is everywhere anhedral and in thin sections as a rule fills small triangular areas between plagioclase and pyroxene crystals. It is limited to those layers in which accumulation of crystals was rapid and it was precipitated from pore-space liquid. Although it may have crystallized originally as tridymite, no confirmatory evidence was observed.

The presence of titaniferous magnetite or ilmenite in the Stillwater has not been confirmed. Minute black opaque specks in some rocks might be one of these minerals or other iron oxides. No doubt in the covered zone above the highest observable exposure there is a titaniferous magnetite zone as in the Bushveld, Great Dyke,

Sierra Leone gabbro, Duluth gabbro, and Skaergaard. These minerals apparently had not reached their saturation point in the lower 60 per cent of the complex now exposed.

Apatite, rutile, and in one case fluorite were noted in heavy-mineral conecntrates

TABLE 11.—*Chemical analysis and X-ray data on ferriferous serpentine mineral*
(Analyst, A. H. Phillips)

		Å	I
SiO_2	42.30	7.37	20
Al_2O_3	.59	4.60	7
Fe_2O_3	4.54	3.67	25
FeO	1.97	2.55	6
MnO	.03	2.49	20
MgO	35.88	2.097	4
CaO	.05	1.749	4
H_2O+	13.12	1.536	12
H_2O-	1.30	1.315	3
TiO_2	.00		
	99.78		

from a very few rocks and were seen in a few thin sections. Coral-pink zoisite is more common. It may be an accessory mineral or a late magmatic alteration product. No zircon or spinel was seen. Traces of biotite and fibrous amphibole were noted in a number of places and probably represent late magmatic alteration products.

As in the Bushveld, rarity of accessory minerals and alteration products is the rule rather than the exception. Partridge (1934) crushed several pounds of the Pyramids norite and found no heavy accessory minerals. The writer has had the same experience with many Stillwater gabbros and norites. This is not true of undifferentiated basalts and gabbros but is an expectable characteristic of crystal accumulates.

The peridotitic rocks of the lower portion of the Ultramafic Zone are partially or wholly replaced by serpentine. Olivine, originally the predominant mineral, goes to chrysotile serpentine, and the orthopyroxene to bastite-antigorite. Plagioclase, not very abundant in these rocks, also seems to be replaced by a white-weathering serpentinous material not specifically identified. It has an index of refraction near 1.56 and is almost isotropic. These appear to be late magmatic alterations formed during the cooling of the lopolith. The H_2O may have been derived by extraction from the floor rocks. All the minerals to be described below replace the serpentine minerals in such a way that a second post-consolidation epoch of mineralization must be postulated.

Large volumes of partially serpentinized harzburgite in the lower portion of the Ultramafic Zone have been converted to "secondary dunites". The olivine of these

rocks is slightly more ferriferous than that of the primary harzburgite. A serpentine-like mineral is found in cracks in, and replacing, the olivine, and also as veins extending outward into fresh bronzitites from adjacent "secondary dunites". It is rather variable in optical properties and color. A chemical analysis is given in Table 11. No magnetite is present in the sample so the iron is a constituent of the mineral. In plane light it is honey yellow to yellowish brown or green and in the hand specimen dark olive green. The outer surface where weathered is stained by limonite, and a weathered zone about an inch thick may be found on material from the coarser veins which is pink to flesh-colored. On polished surfaces spherical and wormlike structures can be seen; both the matrix and "worms" are serpentine, but they differ slightly in color. Under the microscope much of the material is isotropic and appears to be amorphous. Incipient crystallization in it develops radiate fibrous material and asbestiform veins. These crystallized portions have moderate birefringence, perhaps 0.02; interference colors range to yellow orange of the first order in sections of normal thickness. The indices of refraction are variable between 1.53 and 1.55. The crystalline material is noticeably pleochroic with X', parallel to the length of the fibers, yellow brown, and Z' olive green. The X-ray powder diffraction pattern resembles chrysolite serpentine. The main X-ray lines are given in Table 11. The mineral was examined by Dr. George Faust of the U.S.G.S. who considered it to be a variety of chrysotile.

Chapter IV

East Boulder Plateau Section

This chapter presents a petrographic description of the rocks collected on a traverse from the floor of the Stillwater complex to the highest exposed horizon. In part this duplicates descriptions by Peoples of the general relations in the complex, but, since the data are to be considered with details of the mineralogy of the rocks and in stratigraphic order from the bottom upward, the arrangement of the data in this order constitutes a somewhat different point of view. Furthermore it is perhaps valuable to record separately the observations of two investigators because the facts of occurrence are after all the most important contribution which this memoir can make. Where there is such an abundance of factual data to collect it is inevitable that each observer will see things overlooked by another.

A. STRUCTURAL UNITS, LAYERING, AND TEXTURAL RELATIONS

1. SUBDIVISIONS OF THE COMPLEX

In the field Peoples and Howland have divided the complex into conveniently mappable units. After laboratory study the present writer has redivided the complex on a mineralogical basis. For the lower two zones the mineralogical divisions and field units are identical. Peoples' and Howland's Banded and Upper zones, however, are split into four units. In the discussion of the mineral-variation curves and petrographic data, it would be confusing for the reader if the divisions were not based on the data. If further field work were to be done, it probably would be possible to map the upper part of the complex on the basis of mineralogical divisions. It might even be possible to map subdivisions of the major units.

Figure 15 shows the relation between Peoples' and Howland's zones and the mineralogical zones. Each mineralogical zone is named on the basis of its characteristic mineral assemblage.

2. LAYERING

Use of the term *banding* as commonly applied to "banded" norites and gabbros is open to criticism in that a band should be considered two dimensional. Strictly speaking the structures referred to are three dimensional and better designated as *layers* following Wager and Deer. The use of the term banding, however, to describe the features seen in the field may be reconciled with the fact that the actual outcrop presents a plane on two-dimensional surface, on which *bands* appear.

The writer has described the common variety of layering in the Stillwater complex and suggested an explanation for it (Hess, 1938b). It had previously been discussed by Peoples (1936). Wager and Deer (1939) studied the layering of the Skaergaard intrusion in considerable detail and suggested a terminology to describe its various features. This terminology is accepted in the present report and is as follows:

LAYERING: The arrangement of the crystals in layers analogous to stratification in sedimentary rocks. It is equivalent to Hall's (1932) "pseudostratification".

RHYTHMIC LAYERING: Conspicuous banding, rhythmically repeated and constant over wide areas, which produces layers forming almost plane surfaces. Layering of

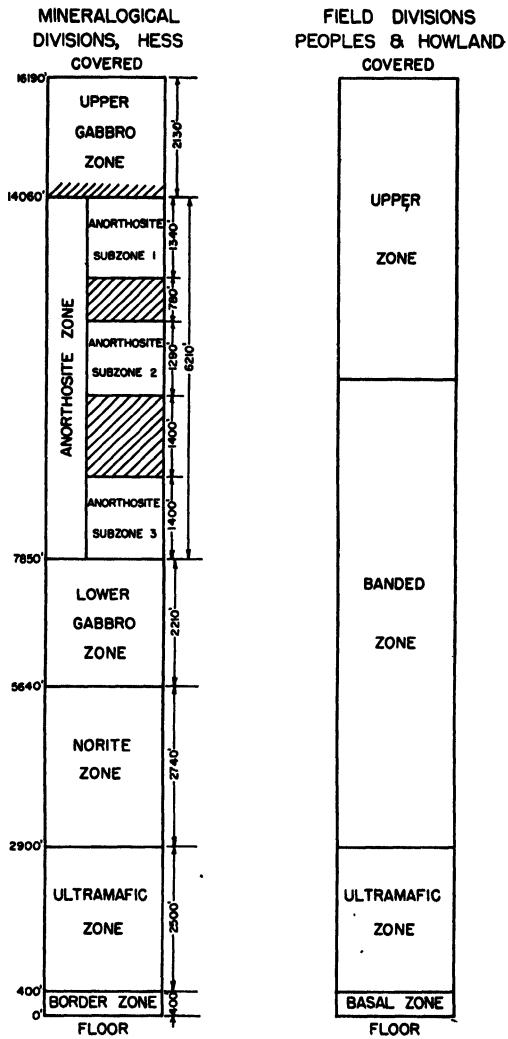


FIGURE 15.—*Subdivisions of the Stillwater complex*

this kind is due to changes in the relative proportions of the minerals. It is commonly associated with the phenomenon of "gravity stratification" (Buddington, 1936; Peoples, 1936) which may be described as a gradation upward in the individual layers from minerals of higher density to minerals of lower density: for example, pyroxene-rich at the base to plagioclase-rich at the top or chromite-rich at the base to olivine-rich at the top. This type of layering does not involve appreciable changes

in composition of the constituent minerals within a layer or from one layer to the next.

CRYPTIC LAYERING: The inconspicuous, and for the most part gradual, change in composition of the mineral phases upward in the intrusive. Wager and Deer also include under cryptic layering occasional abrupt changes resulting from the appearance or disappearance of a particular crystalline phase during the course of crystallization. The present writer calls this "phase layering."

PHASE LAYERING: The abrupt appearance or disappearance of a precipitated crystalline phase or phases during the course of crystallization of the magma. For example, the first appearance of precipitated augite crystals marking the base of the Lower Gabbro Zone could be said to divide the complex into two layers, one below without precipitated augite and one above with this phase. Chromite appears as a precipitated phase at the chromite horizon, but not above or below, and thus is an excellent example of phase layering.

IGNEOUS LAMINATION OR LAMINATION: The arrangement of platy crystals with their longer axes approximately in the plane of layering. This feature is common for the great majority of the Stillwater rocks whether they show pronounced rhythmic layering or not. Peoples (1936) has pointed out that, as a rule, no linear arrangement of the crystals is brought about by parallelism of their longest axes in the plane of lamination. In the Skaergaard intrusion, however, locally some linear structure has been noted. A small degree of orientation in the plane of lamination was found by Van der Berg (1946) in the Bushveld complex.

INCH-SCALE LAYERING: At several places in the Stillwater Complex in the Lower Gabbro Zone a type of layering has been noted which has not been described in the Skaergaard intrusion. Because its origin is uncertain the writer has called it *inch-scale layering*. It consists of layers slightly less than an inch thick composed of alternations of pyroxene and plagioclase (Pl. 8).

3. TEXTURAL RELATIONS

The changes in composition and grain size of the essential minerals and the changes in mineralogy of the rocks are readily presented graphically by plotting the variations against height above the floor. The one remaining variable, the textural relation of the minerals to each other, is difficult to handle in this manner. It has been tested only qualitatively. Inasmuch as it is quite important to correlate textural changes with composition changes if a full appreciation of the process of crystallization is to be obtained, it is necessary to reduce textural relations to a series of symbols. This is rather difficult, and it must involve much simplification of the actual relations and consequently some loss of detail. Unlike the quantitative relations it also involves the personal judgment of the observer. With due recognition of the difficulties, the writer uses a series of symbols and terms to describe textural relations in the discussion of the crystallization of the rocks of the complex.

The accumulation of early formed crystals on the floor of the magma chamber and the building up of the solid rocks layer upon layer in this manner until the whole, or practically the whole, of the complex crystallized are reflected in the textural relations of the rock specimens. Rocks developed by crystal accumulation contain a

large proportion of nearly euhedral crystals cemented by or modified by material crystallized from the interstitial liquid trapped in the pore space between the accumulated crystals. The term "peri-euhedral" is coined to describe the almost euhedral form of these crystals. In a peri-euhedral crystal the original euhedral outline may be recognized, but it has been slightly modified by addition of material from the pore-space liquid to form a thin "envelope" around the crystal, rounding off or producing minor irregularities on the originally smooth crystal faces. A rock composed largely of peri-euhedral crystals might be said to have a "peri-idiomorphic" texture.

The accumulated crystals in a specimen may be plagioclase, orthopyroxene, clinopyroxene, or olivine, or any combination thereof. Certain accessory minerals may also exhibit the same features, as for example chromite and presumably titaniferous magnetite, judging from the Bushveld complex, but these layers are quantitatively a minute fraction of the complex. Similarly the material which has crystallized from the interstitial liquid may consist of one or any of the essential minerals. To understand the textural relations one must study certain extreme types—for example, specimens in which accumulated crystals are represented by a different mineral phase or mineral phases from that which crystallized from the interstitial fluid. From such specimens it has been estimated that the pore-space liquid originally amounted to 26–29 per cent of the whole by volume.

B. BORDER ZONE AND FLOOR ROCKS

Stratigraphically below the floor contact of the complex along the line of the section are found (1) the eulysites (to be described elsewhere by Howland); (2) a bluish, sugary-textured quartzite; (3) hornfels that is difficult or impossible to distinguish in many places from the next overlying rock; (4) the doleritic "chilled" border facies of the complex. The writer is concerned only with the border facies and its distinction from the hornfels. The only igneous rocks noted below the floor contact are doleritic intrusive rocks with plagioclase phenocrysts. The phenocrysts are corroded and may reach several centimeters in cross section. It is not known whether these intrusives are related to the complex or are younger. Younger nonporphyritic dolerite dikes cut rocks of the complex.

A specimen (EB 89) collected at the base of the complex was thought to be a fine-grained facies of the "chilled" dolerite, although even when it was collected some doubt was entertained as to whether it was dolerite or hornfels. Although it has now been studied in thin section and chemically analyzed this doubt has not yet been removed. The specimen was examined by Howland, who has had much experience in distinguishing the hornfels from the rocks of the complex, and also by Peoples. It does not contain cordierite, phlogopite, or quartz, which are characteristic of the hornfels, but it does have a granulitic texture, a feature common in the hornfels but not in the dolerite. The pyroxene resembles that of the dolerite. It is concluded that it is probably hornfels but may be contaminated dolerite. In either case its composition is not representative of the border facies.

The chemical analyses of EB 89 and of Peoples' chilled border facies dolerite 463 G 70 are given in Table 12. EB 89 contains much more normative magnetite than Peoples' specimen, and this partially accounts for its relatively lower percentage

TABLE 12.—*Chemical composition of border facies*

	Analyses		Norms				
	1	2	1			2	
SiO_2	50.68	47.34	Q	.00		1.20	
Al_2O_3	17.64	17.12					
Fe_2O_3	.26	2.51	Pc Or 1.39			1.22	
			Pc Ab 15.72	56.17		6.45	50.34
			Pc An 39.06	($An_{68.2}$)		42.67	(An_{84})
FeO	9.88	12.05	Di Wo 5.04			2.49	
			Di En 2.60	9.95		1.30	4.91
MgO	7.71	8.23	Di Fs 2.31			1.12	
CaO	10.47	9.79	Hy En 16.70	31.82		19.30	37.65
			Hy Fs 15.12	($En_{59.4}$)		18.35	(En_{58})
Na_2O	1.87	.76	Mt	.46			3.71
			Il	.91			.99
K_2O	.24	.21	Ap	.20			
			H_2O	.48			.77
H_2O+	.42	.73					
H_2O-	.06	.04		99.99			99.57
			Modes (wgt. per cent)				
Cr_2O_3	.04	.04	1. (Peoples) Labradorite 53.5%, hypersthene 37.3%, augite 7.9%, opaque 1.3%				
TiO_2	.45	.52					
P_2O_5	.09	tr	2. Bytownite 41.6%, hypersthene 52.7%, augite 5.0%, opaque 0.7%				
MnO	.15	.17	(Hypersthene partly altered to colorless amphibole)				
	99.96	99.51					

1. Peoples' specimen 463 G 70. Hypersthene dolerite border facies (Analyst: R. B. Ellestad. Measured density 3.03; calculated norm density 3.02).

2. Specimen EB 89. Contaminated hypersthene dolerite or hornfels (Analyst: A. H. Phillips (Alkalies by Ellestad))

of SiO_2. The Mg/Fe ratios of the normative pyroxene are practically identical. The less abundant normative plagioclase in EB 89 is much more calcic—far too calcic for a normative plagioclase of the liquid of the basaltic magma. If EB 89 is contaminated dolerite it has presumably lost alkalies and gained magnetite, although in other respects the two rocks are chemically very similar.

The textures of EB 89 and 463 G 70 are shown in Plate 4. Specimen EB 89 is much finer-grained and in plane light has a panidiomorphic or granulitic texture, whereas 463 G 70 is considerably coarser-grained and has an ophitic texture. Under crossed nicols the rounded orthopyroxene grains in EB 89 extinguish simultaneously over 2 or 3 sq cm areas and are poikilitic. A portion of the orthopyroxene has been

altered to a fibrous, colorless amphibole and magnetite, but the plagioclase is fresh. The orthopyroxene was analyzed (Table 13); the sample analyzed was small and contained some alteration products after magnetic separation from the rock. It should not be considered a superior analysis.

TABLE 13.—*Chemical composition of hypersthene from specimen EB 89 (contaminated hypersthene dolerite or hornfels)*

(Analyst, R. B. Ellestad)

SiO_2	50.24	
Al_2O_3	4.52	
Fe_2O_3	.41	
FeO	21.81	
MgO	15.81	
CaO	4.39	
Na_2O	.18	$Ca_{10}Mg_{50}Fe_{40}$ or $Mg_{56}Fe_{44}$
K_2O	.04	
H_2O+	1.42	
H_2O-	.04	
TiO_2	.69	
MnO	.43	
	99.98	

TABLE 14.—*Mineralogical changes, Border Zone to Ultramafic Zone*

	Border facies dolerite 463 G 70	Mafic norite EB 37	Basal feldspathic websterite EB 92X	Bronzitites and harzburgites
Orthopyroxene	$Mg_{58}Fe_{42}$*	$Mg_{77}Fe_{23}$	$Mg_{81}Fe_{19}$	$Mg_{83}Fe_{17}$-$Mg_{87}Fe_{13}$
Clinopyroxene	$Ca_{41}Mg_{41}Fe_{18}$	$Ca_{41}Mg_{46}Fe_{13}$	$Ca_{40}Mg_{52}Fe_8$	$Ca_{40}Mg_{53}Fe_7$
Plagioclase	$An_{68}Ab_{32}$	$An_{74}Ab_{26}$	$An_{77}Ab_{23}$	$An_{66}Ab_{34}$-$An_{80}Ab_{20}$
Mode in wt.%	Hypersthene 37¼	Hypersthene 66¼	Bronzite 72	
	Augite 8	Augite 9¾	Chrome augite 21	
	Plagioclase 53½	Plagioclase 24	Plagioclase 7	
	Opaques 1¼			

* This is an inverted pigeonite

This is a Stillwater type hypersthene, an inversion product of pigeonite. It has the characteristic augite plates parallel to what originally was {001} of the pigeonite; chemically it exhibits the higher CaO content, as would be expected. Howland states that he has not observed this type of orthopyroxene in other hornfelses. Nevertheless weighing all the evidence the writer is inclined to regard EB 89 as most probably a hornfels.

The hypersthene of the border facies specimen 463 G 70 is similar to that in EB 89.

It also has the augite plates indicating inversion from pigeonite and has approximately the same Mg:Fe ratio.

Stratigraphically above EB 89 is a porphyritic rock containing plagioclase phenocrysts 2–3 cm in diameter (EB 90). The total thickness of the hornfels (?) and the "chilled" dolerite of the Border Zone appears to be of the order of magnitude of 400 feet, but inasmuch as the dip of the contact is not definitely known in this section, and the outcrops are small and scattered, the estimate is at best only a rough guess.

Above the fine-grained facies is a coarser-grained facies of the Border Zone. This rock has a fairly well developed layering approximately parallel to the floor of the intrusive. Streaks and segregations of coarse, norite pegmatite parallel the layering. Approaching the contact with the Ultramafic Zone above, the Border Zone becomes enriched in hypersthene so that it changes from hypersthene dolerite below to mafic norite above.

In the mafic norite (EB 37) most of the orthopyroxene is peri-euhedral, whereas the plagioclase and clinopyroxene are anhedral. Both the texture and the much more magnesian character of the orthopyroxene (compared to the orthopyroxene of the "chilled" facies below) suggest enrichment in orthopyroxene by settling of crystals. Such crystals of bronzite or olivine would have a composition near $Mg_{87}Fe_{13}$ when entering the Border Zone but might then be modified to an intermediate composition by reaction with the surrounding liquid. The somewhat more calcic character of the plagioclase and slightly more magnesian character of the clinopyroxene cannot, however, be accounted for similarly by settling of crystals. Plagioclase and clinopyroxene are interstitial to the hypersthene and have no doubt crystallized *in situ*; they only appear as accumulated settled crystals thousands of feet higher in the complex after a vast amount of bronzite and olivine has separated. The change in composition of the plagioclase and clinopyroxene may be ascribed to another process—namely, the loss upward of some of the less refractory constituents of the magma by diffusion of the more refractory constituents from the overlying magma to the site of crystallization. Although diffusion of ions is an exceedingly slow process so also is the cooling of an intrusion of this size. Later in this report the writer considers the diffusion process and the time available for it to take place.

In the basal portion of the Ultramafic Zone, the feldspathic websterite (EB 92 X) indicates greater enrichment in magnesian orthopyroxene than in EB 37 described above. This may again be ascribed to settling of orthopyroxene or to the settling of magnesian olivine which reacted with the magma to form orthopyroxene en route. The plagioclase has diminished in amount compared to EB 37 and has become more calcic, suggesting that the postulated process of diffusion has also become more effective. The orthopyroxene is peri-euhedral, and other constituents are interstitial to it.

After review of the petrographic relations in the Border Zone, the following deductions may be made. Immediately after intrusion the zone within a few hundred feet of the contact lost heat to the cold wall rocks. The heat loss at this stage was not sufficient to cause immediate crystallization but merely increased the viscosity of the magma. From this point onward the rate of cooling decelerated so that crystallization took place slowly and resulted in the moderately coarse grain size and the even

texture of the "chilled" dolerite. It also permitted local segregations and accumulations of volatile-rich patches which resulted in the formation of the norite pegmatite schlieren here and there throughout the Border Zone. Possibly some of the volatiles represent water extracted from the sediments of the floor.

After emplacement and before appreciable crystallization, the viscosity of the liquid decreased from the floor upward. Almost immediately after emplacement some immiscible sulfide droplets of relatively high density as compared to the magma settled through the Border Zone and accumulated in the vicinity of the contact; Howland will describe this elsewhere. The viscosity of the lower portion of the Border Zone was too high to permit settling of the less dense silicate crystals through it, and thus it crystallized before it could be modified in composition by such a process. With the viscosity decreasing upward, however, an increasing number of ferromagnesian crystals penetrated the liquid. The trapped settled crystals increased in amount upward through the Border Zone up to the contact of the Ultramafic Zone. At this horizon the settled crystals formed a crystal mush upon the floor; they were as tightly packed as their shapes would permit and they contained only enough interstitial liquid to fill the pore space. From this point, the accumulation of crystals on the floor of the magma chamber foot by foot and solidificaton of the resulting mush is the dominant process in the development of the complex.

The peri-euhedral hypersthene crystals which represent the crystals that settled into the Border Zone probably started settling as olivines and reacted with the magma to form pyroxene during the final stages of settling or even after settling. In the lower half of the overlying Ultramafic Zone crystals of olivine are the predominant settled crystals; bronzite forms poikilitic crystals around them. The bronzite is of later crystallization and no doubt formed in the mush after accumulation of the olivine. In general bronzitite, developed by accumulation of bronzite crystals on the floor, does not occur below the middle of the Ultramafic Zone, and it forms the dominant rock of the upper half of the Ultramafic Zone. There can be no doubt that olivine crystallized first from the magma and was succeeded and finally replaced in the crystallization process by orthopyroxene.

C. ULTRAMAFIC ZONE

1. GENERAL RELATIONS

The lower portion of the Ultramafic Zone is predominantly harzburgite, and the upper portion bronzitite. A medial chromite horizon (or horizons) approximately marks the boundary between them. The harzburgites are partially serpentinized, but the bronzitites are not. Also present within the lower portion of the zone are irregular areas of dunitic rocks clearly derived from the primary harzburgites. For purposes of discussion the dunitic and related secondary rocks are considered under the heading Secondary Group of Rocks, and the primary harzburgites, bronzitites, and chromitites under the heading Primary Group of Rocks.

The East Boulder Plateau section is not particularly favorable for study of the mineral changes across the Ultramafic Zone because: (1) the lower portion is highly serpentinized; (2) this portion is to a large extent made up of the secondary group of rocks which is in many cases difficult to distinguish from the primary group; (3)

the dip of the layering is variable and can be observed only at a few horizons so that it is difficult to determine the exact stratigraphic position of the samples; (4) although outcrops are excellent on the walls of many small cirques, intervening gently rolling areas are grass-covered. It is, therefore, necessary to depend somewhat on the use of Peoples' Little Rocky–Nye Basin divide samples for the study of mineralogical variations within the Ultramafic Zone. The East Boulder Plateau section, however, is well adapted to a study of the secondary group of rocks and the relations between the Ultramafic Zone and overlying Norite Zone.

The Ultramafic Zone on the East Boulder Plateau is estimated to be 2500 feet thick. To the west across the Boulder River this zone thins and finally pinches out. Eastward in the region of the Stillwater River and Peoples' described sections it thickens considerably.

Since Peoples has already described the Little Rocky–Nye Basin divide section in considerable detail the writer merely refers to Peoples' described section without further discussion, and the rock types normally encountered in the Ultramafic Zone will be briefly described here.

2. PRIMARY GROUP OF ROCKS

a. Rock types.—This group consists entirely of various kinds of harzburgites and bronzitites. During the formation of the Ultramafic Zone olivine, bronzite, and chromite were the only minerals accumulating on the floor of the magma chamber, and the chromite was distinctly subordinate. Plagioclase and chrome augite are also present in the rocks of this zone, but in small amounts, and they are entirely interstitial to the olivine and bronzite. Their presence, however, results in such rock types as feldspathic, chrome augite-bearing harzburgites and bronzitites. The total amount of plagioclase plus augite in these rocks does not exceed 30 per cent by volume (except in certain mafic pegmatites) and in most is less than 10 per cent. The mafic pegmatites are composed of coarse plagioclase, chrome augite, bronzite, and more rarely olivine. On the basis of both mineralogy and relation to surrounding rocks it appears likely that these mafic pegmatites represent squeezed-out segregations of the liquid which were trapped in the mush of accumulated early crystals. The bulk composition of these pegmatites, however, does not represent the composition of the residual magma at this instant in the crystallization but is rather far from it. The reason for this will be given when crystallization from the interstitial liquid is considered in Chapter VI.

b. Chromite horizon.—The main chromite horizon occurs approximately at the middle of the Ultramafic Zone. It ranges in thickness from a few feet to about 35 feet. It has been considered by Peoples and Howland (1940), and Sampson will describe it in detail, so no further description of it is given here. The horizon represents that episode in the history of the crystallization of the Stillwater magma when chromite reached its saturation point and precipitated. Chromite in large amounts was never precipitated again. The original content of Cr_2O_3 in the magma, if the Border Zone chilled dolerite is an adequate sample, is .04 per cent. A considerable fraction of this amount must have been removed by the precipitation at the chromite horizon. Chromite continues as a minor accessory mineral in the ultramafic rocks above the chromite horizon for a short distance but is absent from the rocks of the uppermost

part of this zone. For the remainder of the history of crystallization of the magma Cr_2O_3 enters the pyroxene in minute amounts until it is almost exhausted. The clinopyroxene of the Ultramafic Zone may contain as much as 1 per cent Cr_2O_3. The amount of chromium in the augite decreases upward in the complex, so that in the highest horizons it has diminished to a negligible amount.

c. Rock types above Chromite Horizon.—Below the chromite horizon about 80 per cent of the original rocks present are harzburgites, many of which contain a few per cent of plagioclase and a little chrome augite. The remaining 20 per cent are bronzitites, some of which also contain minor amounts of plagioclase, chrome augite, and olivine. Chromite is, of course, an accessory mineral in both rock types. From above the chromite horizon to the top of the Ultramafic Zone 80 per cent of the rocks are bronzitites, and the remainder are harzburgite; all the harzburgite is in the lower portion of this group of rocks.

d. Serpentinization.—Throughout this zone serpentinization of the olivine is the rule rather than the exception. Perhaps 85 per cent or more of the olivine in the harzburgites of the primary group of rocks has been serpentinized. Small relics of original olivine remain in some of the serpentines, and in other specimens the serpentinization is complete. The outline of original euhedral olivine crystals averaging 2–5 mm in diameter is seen in practically all the harzburgites. Bronzite shows much greater resistance to serpentinization. In most of the bronzitites little serpentinization is apparent. In most harzburgites the poikilitic bronzite is unserpentinized except along cracks, whereas the olivine is almost completely or completely serpentinized. Less commonly the bronzite is also serpentinized.[5] The interstitial chrome augite also resists serpentinization in most cases although the amount of augite is small and its distinction from serpentinized bronzite is usually not possible. Plagioclase in the partially serpentinized harzburgites is fresh as a rule, but in highly serpentinized harzburgites it also alters and presumably to a mineral of the serpentine group. The alteration product of the plagioclase appears as dull, cloudy, brownish to colorless areas under the microscope in plane light. Under crossed nicols these areas may be almost isotropic or may consist of fine flaky aggregates resembling the sericitic alteration of feldspars in felsic rocks. The flakes have moderate birefringence since they appear to be white or yellow of the first order under crossed nicols even though the individual flakes are much thinner than the thin section itself. Although the flaky substance looks like sericite it may well be a mineral of the serpentine group. The gradation of the flaky areas into isotropic areas supports this possibility.

The writer believes the original serpentinization of these rocks took place during cooling of the complex. It preceded certain younger dikes that cut the complex and apparently preceded the formation of the Secondary Group of Rocks to be discussed below. A second period of serpentinization with a different type of serpentine accompanied the formation of the Secondary Group.

e. Olivine-bronzite relations.—In the lower portion of the Ultramafic Zone harzburgite is the predominant rock type, and olivine is the predominant mineral that accumulated on the floor. Almost all the bronzite of the harzburgites formed poikilitic

[5] Each bronzite is replaced by a bastite single crystal. Such crystals are commonly considered to be antigorite. The writer has not determined this.

crystals either enclosing olivine or interstitial to it (Fig. 17B). Much less common are rocks containing both well-formed olivine and bronzite crystals. In most of the harzburgites the bronzite either crystallized directly from the interstitial liquid or by reaction of the liquid with the accumulated olivine crystals. The composition of the magma was such that it was always near the boundary between the olivine field and the orthopyroxene field. This accounts in part for the intercalated layers of bronzitite which appear in the harzburgite and the oscillation from harzburgite to bronzitite in the central portion of the Ultramafic Zone. In the upper portion of this zone olivine ceased to crystallize, and only bronzite accumulated on the floor. The reappearance of olivine much higher in the complex will be dealt with in the discussion of the Anorthosite Zone.

The bronzitite layers in harzburgite have sharp boundaries. The absence of rhythmic layering in the olivine-bronzite rocks may be noted. The measured densities of olivine and bronzite from these rocks are virtually identical so that rhythmic layering would not be likely if crystals of both minerals were available at the same time. Some digression here seems warranted to account for the precipitation alternately of olivine and orthopyroxene. Evidence discussed in later chapters suggests that crystallization takes place in a very small portion of the total volume of the magma in a layer just below the roof. Such a small volume of magma might, after precipitating olivine, move over in composition into the pyroxene field (although the great bulk of magma in the chamber would remain in the olivine field). If crystallization continues in that volume of magma which has reached the pyroxene field, a bronzitite layer may result. Such inhomogeneities in the magma must be quite temporary inasmuch as they will probably be swept away by periodic convection within the magma chamber. Mixing by convection currents will bring the composition and crystallization back into the field of olivine until finally the magma as a whole passes into the pyroxene field.

Most harzburgites have olivine crystals (or serpentine pseudomorphs after olivine) which are peri-euhedral. The interstices are filled by poikilitic bronzite. In such cases the bronzite has crystallized from the pore-space liquid. More commonly the olivine is present as rounded rather widely spaced relics in the poikilitic bronzite. In these a considerable part of the bronzite is derived by reaction of pore-space liquid with the olivine crystals. A harzburgite specimen from the chromite horizon shows quite clearly that part of the poikilitic bronzite is attributable to reaction and part to direct crystallization; an original mush of euhedral olivine crystals had in its pore space a liquid crowded with minute chromite octahedra arranged in delicate branching chain structures. The pore liquid crystallized as poikilitic bronzite including the chromite octahedra. The liquid also reacted with the olivine crystals forming bronzite, but this bronzite is free of chromite octahedra. In this case about a third of the original olivine has been transformed to bronzite.

The bronzite and olivine show very little change in composition throughout the Ultramafic Zone. Only in the uppermost few hundred feet is there any definite trend from more magnesia to more iron. Throughout the major portion of the zone an oscillation of the Mg:Fe ratios of the olivine and orthopyroxene is noted; the two phases vary sympathetically in Mg:Fe ratio. These minor oscillations in composi-

tion are probably a function of the variable rate of accumulation of crystals on the floor and the consequently greater or lesser modifying effect of the trapped interstitial liquid on the settled crystals (Table 15).

One objective of this study is to compare critically the observable relations among common natural silicates with silicate systems investigated in the laboratory. The application of Bowen and Schairer's (1935) work on MgO–FeO–SiO_2 to the ultramafic rocks is evident. The reaction relationship of olivine to pyroxene is well shown in the Ultramafic Zone. The MgO:FeO molecular ratio of approximately 60:40 in the norm of the border facies, which should be representative of the original magma, as compared to the MgO:FeO ratio of the earliest crystals of olivine ($Fo_{88}Fa_{12}$), is in good accord with the liquidus–solidus relations in Bowen and Schairer's results. One might expect from the experimental work that the olivine associated with a given orthopyroxene would be slightly more ferriferous than the pyroxene; in these natural occurrences the Mg:Fe ratios of the olivine and pyroxene are about the same. This may be accounted for by the fact that the pyroxene continued to crystallize in contact with pore-space liquid for a considerable period after the olivine ceased to form. Many of the effects of fractional crystallization also become evident higher in the complex. Early separation of highly magnesian olivine and pyroxene result in a rapid decrease in the Mg:Fe ratio of the residual liquid. Early crystallization of olivine from a magma which was just saturated with respect to SiO_2 (has neither normative olivine nor normative quartz) results soon afterward in the appearance of interstitial quartz.

f. Plagioclase, augite, and mafic pegmatite.—Plagioclase and augite crystallize directly from the interstitial liquid and are anhedral in the Ultramafic Zone. The plagioclase ranges from An_{60} to An_{80}, but the augite seems to be nearly constant in composition, $Ca_{41}Mg_{52}Fe_7$, with a maximum range of about ±3% Ca, Mg, or Fe. Augite contains about 1% Cr_2O_3 and in the hand specimen can be distinguished from the brown bronzite by its emerald green color.

The bronzitites indicate that the orthopyroxene crystals will pack together in a mush so that they contain about 28% by volume of pore-space liquid. The composition of the liquid at this stage in the crystallization of the complex is not far from that of the original magma as represented by the border facies. If this magma were trapped in the pore space, two thirds of it, or 18% of the whole rock by volume, should crystallize as plagioclase and augite. Nevertheless the great majority of bronzitites contain much less interstitial plagioclase and augite than this, and a great many contain virtually none. The bronzitite near the chromite horizon on the east side of the Boulder valley is a good example. An analysis quoted from Peoples (1932 Ph.D. thesis, Princeton University) is given in Table 16. How such monomineralic bronzitites form, or how any of the monomineralic rocks of the complex form, is an interesting problem. In the writer's opinion the explanation is as follows. Starting for example with a bronzite mush on the floor, if the distances are not too great and if the channels to the overlying magma are kept open, two-way diffusion of ions between the pore-space liquid and overlying magma may take place. This would permit continued growth of the bronzite crystals of the mush at the expense of the pore-space liquid, a process which could continue until the liquid was exhausted.

If, however, accumulation of bronzite on the floor were more rapid, the mush might become too thick for ionic diffusion to be effective, in which case some or all of the pore-space liquid might be trapped. Thus the pure bronzitites may represent very

TABLE 15.—*Composition variation in orthopyroxene and olivine in the Ultramafic Zone*

Orthopyroxene		Olivine Fo†	Rock types	
N_g	Mg*			
1.6875	81		Norite	Norite zone
1.6860	82		Bronzitites	Ultramafic zone
1.6840	83			
1.6825	84			
1.6815	85			
1.6825	84	85		
1.6820	86			
1.6790	87			
1.6810	86	88	Chromite horizon	
1.6770	88	87		
1.6815	85		Harzburgites	
1.6815	85	86		
1.6840	83			
1.6830	84	86		
1.6815	85	87		
1.6810	86	88		
1.6825	84	86		
1.6840	83	85		
1.6825	84	84		
1.6825	84	80		
1.6855	82			
1.6940	77		Websterite	
				Border zone

* Mg = Number of Mg^{+2} ions × 100 divided by $Mg^{+2} + Fe^{+2} + Fe^{+3}$ ions in the orthopyroxene. In as much as Fe^{+3} is included in this value, it is about 1 per cent lower than the standard En which is commonly used.

† Fo in molecular per cent

slow accumulation of crystals on the floor, and increasing amounts of interstitial plagioclase plus augite may indicate increased rates of accumulation. The same reasoning may be applied to anorthosites and also to harzburgites if modified slightly to take into consideration the reaction relationship of olivine to bronzite.

Layers of mafic pegmatite are found at many places in the Ultramafic Zone. They are characterized by crystals several centimeters long and consist of plagioclase, augite, bronzite, and rarely olivine. The layers range from a few inches to a couple

of feet in thickness. Unlike most other layers the pegmatites have wavy or scalloped boundaries. It seems likely that these pegmatites also represent pore-space liquid (perhaps enriched in volatiles) drained off from the mush by some form of filter pressing. A layer of pegmatite a few inches thick in many cases underlies layers of

TABLE 16.—*Chemical analysis of bronzitite*
465-E-3h, Boulder River valley; analyst, T. Kameda

SiO_2	54.68
Al_2O_3	1.80
Fe_2O_3	0.50
FeO	9.19
MgO	30.19
CaO	2.22
Na_2O	0.04
K_2O	0.03
H_2O+	0.46
H_2O-	0.05
TiO_2	0.11
P_2O_5	0.02
Cr_2O_3	0.47
MnO	0.21
	99.97

chromite. One might suppose that the comparatively high thermal conductivity of the chromite might have something to do with the development of pegmatite beneath it, but the writer can present no really adequate explanation of its origin.

3. SECONDARY GROUP OF ROCKS

a. Distribution.—The Secondary Group of Rocks consists mostly of dunites and serpentinites. They occur only within the Ultramafic Zone where in many places they cut sharply across the layering of the Primary Group. Their absence from the floor rocks, the Border Zone, and the rocks of the complex above the Ultramafic Zone indicates a close genetic relationship between the Secondary Group of Rocks and the Primary Group of Rocks of the Ultramafic Zone. They are fairly abundant near the base of the Ultramafic Zone, fairly common as high as the chromite horizon, but absent from the upper portion. They are more abundant on the East Boulder Plateau and along the Boulder River section than elsewhere.

b. Relations to wall rocks.—The dunitic secondary rocks are almost invariably associated with partially serpentinized primary harzburgites. The contacts between the two are exceedingly irregular. The relations between them can be observed only on favorably weathered surfaces where the bronzite in the harzburgite stands out in relief. The weathered dunite is light grayish brown, and the weathered harzburgite is more yellowish or reddish brown. Passing from the normal harzburgite to dunite one finds first blebs and irregular veins of dunite appearing in the harzburgite. These increase in number and volume until the predominant material is dunite with ir-

regular relics of harzburgite (Hess, 1938a, Fig. 2). The strike and dip of the layering in these relics where observable is constant and consistent with the strike and dip of the layering of the surrounding primary harzburgites, indicating that the relics have not been rotated in the dunite. This is the strongest point in favor of the development of the dunites by transformation *in situ* from solid harzburgite to solid dunite

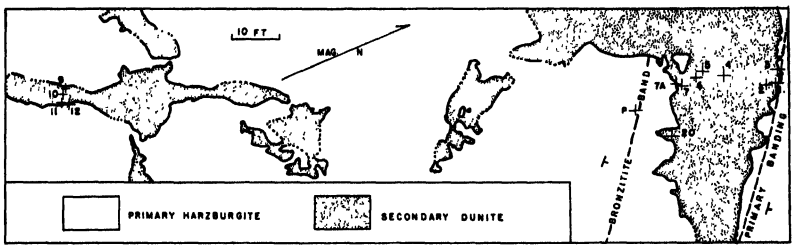


FIGURE 16.—*Map showing irregular contacts between secondary dunite and primary harzburgite*
After E. Sampson

without going through a liquid state. Occasional corroded relics of large bronzite crystals are found in otherwise pure dunites.

Dunite contacts with bronzitite, in contrast to those with harzburgite, are almost invariably sharp. No gradations from bronzitite to dunite have been observed. Sampson's unpublished map (Fig. 16) shows a large mass of dunite in contact with bronzitite. The irregularity of the contact here is partly a result of the low dip of the plane of the bronzitite layer with respect to the surface mapped. A bronzitite overlying a large mass of dunite on "Dunite Ridge" is sharply transected by small pipes and dikes of dunite (Hess, 1938a, Figs. 3, 4). For a fraction of an inch from the contacts the bronzitite is replaced by a peculiar, ferric iron-bearing variety of serpentine, and a system of radial veins containing the same serpentine extends outward from the small pipes. For a few inches from the contact with larger masses of dunite, the bronzitite shows a slight alteration to a colorless amphibole, probably tremolite, and magnetite. This alteration is very mild and commonly occurs only along grain boundaries or cracks in the bronzite. The alteration is interpreted as hydrothermal and is presumably produced by solutions derived from or related to the formation of, the dunite.[6] At one locality a bronzitite overlying dunite was altered largely to talc for several feet from the contact; both the type of alteration and the volume altered are quite exceptional.

c. Petrography.—The mineralogy of the Secondary Group of Rocks is exceedingly simple. Olivine is the only essential mineral in the fresh rocks, and chromite and magnetite the only accessory minerals. All these rocks, however, are somewhat serpentinized (35%–100%). Magnetite may accompany the development of serpentine. Most of the serpentine is the peculiar ferric iron-bearing variety, although some ordinary serpentine (chrysotile) is also present.

[6] The term dunite has been used throughout this discussion to include partially or completely serpentinized as well as fresh dunite.

OLIVINE: The olivine is very uniform in composition (Mg_{86} to Mg_{89}) and approximately of the same composition as the olivine in the primary harzburgites, although the bulk composition of the dunite is considerably richer in iron than is the harzburgite. The olivine is coarser-grained than in the ordinary harzburgites (1–3 cm in diameter). The coarse olivine crystals are anhedral with interlocking grains having sutured boundaries; in many cases they are divided into bands having slightly different extinction positions. The boundaries between these bands are straight lines, and, although they resemble twinning at first glance, they are an effect of strain in the crystals; adjacent segments of these crystals have been rotated through a few degrees with respect to one another. This is not unlike the wavy extinction bands seen in strained quartz crystals.

SERPENTINE: The Secondary Group of Rocks are in all cases partly serpentinized, and in some cases serpentinization is complete. The two analyzed dunites indicate approximately 45 per cent serpentine by weight. Thin sections suggest far less serpentinization, but this illusion results probably from the fact that the eye is attracted by the high relief of the olivine in plane light and the bright colors under crossed nicols.

A yellow-brown ferric iron-bearing serpentine is characteristic of these rocks. Normal serpentine plus much magnetite is, however, common in the highly serpentinized specimens.

A chemical analysis of the yellow-brown serpentine is given in Table 11. No magnetite is present in this sample which was taken from a vein associated with one of the dunite pipes, so the ferric iron is definitely a constituent of the mineral.

This variety of serpentine is common in the less serpentinized dunites, along contacts of the secondary dunites with the wall rocks, and as veins and replacements extending a short distance into the wall rocks. It is only a minor constituent of the serpentinites, in which normal serpentine (chrysotile) plus abundant magnetite take its place.

The dunites very commonly exhibit two sets of rather wavy and irregular serpentine-filled cracks which divide the rock into small roughly quadrilateral areas. These quadrilateral areas between veins are occupied by olivine or a rounded central relic of olivine in serpentine. The serpentinites show a similar texture; the vein pattern is accentuated by abundant central fillings of magnetite. In many cases one set of veins (the one at right angles to the wall rocks in contact specimens) is more conspicuous than the other giving the rock a foliated appearance. Some specimens show three or four sets of veins and complicated or irregular patterns. In many specimens one set of veins is isotropic, and another is anisotropic. The isotropic veins generally appear to be older. The veins commonly exhibit banding with variations in the properties of the serpentine from the wall to the center. At the center of veins a very thin band of a colorless mineral with slightly higher indices than the serpentine and birefringence near .015 has been observed in many specimens. It may be amesite, but the band is too thin to obtain exact optical properties.

OTHER MINERALS: Chromite is an accessory mineral in all the Secondary Group of Rocks. It commonly forms dark-brown, euhedral octahedra. Many of the crystals appear ragged and anhedral apparently as a result of later deposition of magnetite

along their margins. In a strong light, a ragged black opaque margin can be distinguished from the brown translucent chromite of the core. Fairly coarse anhedral chromite was also noted in some specimens. Many of the chromite grains contain rounded inclusions of serpentine and olivine.

A colorless monoclinic amphibole is *very common* along the contacts of the dunites with bronzitite. It also replaces relics of bronzite within the secondary dunites. Rarely stubby prisms of a clear, colorless amphibole, possibly tremolite, are found between olivine grains. The development of this amphibole precedes serpentinization of the secondary dunite.

A chlorite, probably clinochlore, is found sparingly in some of the dunites. It has the following properties: colorless, birefringence approximately .004, indices of refraction higher than the serpentine but much lower than the tremolite, optic angle about 10°, and sign positive. Some grains show anomalous greenish-gray interference colors characteristic of certain chlorites. In some cases it appears to fill miarolitic cavities. Olivine shows euhedral outlines against it. These are the only cases in which crystal faces of the olivine are observed in the dunites.

Very rarely a minute amount of carbonate is found at the center of serpentine veins.

Pyroxene is found in many of the dunite specimens, but with one exception it is believed that these pyroxenes (as a rule bronzites) are relics derived from the primary harzburgites. Many of them are large (1 to 2 cm), as are the poikilitic bronzite crystals of the harzburgites. They are corroded by and eaten into by fingers of olivine. Acicular amphibole replaces the borders of the pyroxene and some entire crystals. Some magnetite is also developed with the amphibole. The amphibole does not replace olivine. It appears that bronzite is altered by solutions contemporaneous with the crystallization of the dunitic material. The same type of alteration occurs at contacts where dunite cuts bronzitite.

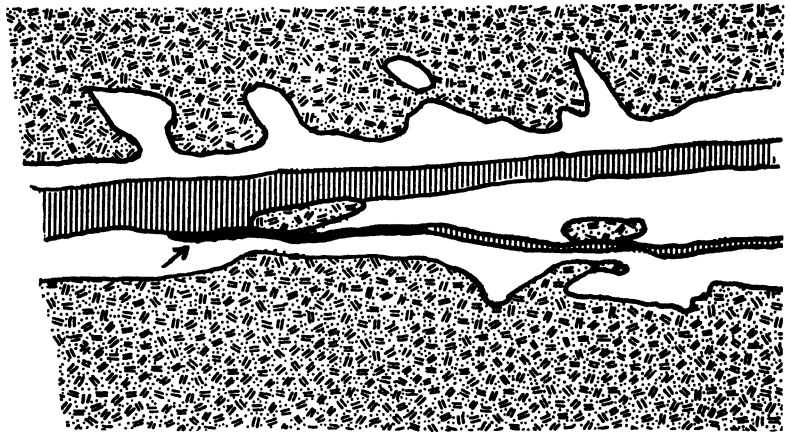
In one case noted in the field pyroxene had developed along a central vein in a small dunite "dike" (Fig. 17a). Unfortunately none of this material was collected so that information is not available as to the nature of the pyroxene.

d. Chemical composition.—The variation in chemical composition of the Secondary Group of Rocks reflects the simplicity of their mineralogy. Table 17 gives the chemical composition of four analyzed specimens; SiO_2, Fe_2O_3, FeO, MgO, and H_2O are the major constituents. Also in Table 17 the molecular ratios of SiO_2, Fe_2O_3, FeO, and MgO have been calculated, converted to percentage of the four constituents, and compared to the total water content which is proportional to the degree of serpentinization. From this recalculation it can be seen that MgO and SiO_2 vary little in the four analyses, total iron increases somewhat with increase in serpentinization, and ferrous iron shows a reciprocal relationship to ferric—the iron becomes more oxidized as serpentinization increases. An analysis of the pure yellowish-brown serpentine is included with the four analyses for comparison.

4. ORIGIN OF THE SECONDARY GROUP OF ROCKS

A summary of relations of the Secondary Group is as follows:

The dunites do not occur in the floor rocks and are most abundant near the base of the Ultramafic Zone.



a

b

FIGURE 17.—*Dunite "dike" with central pyroxene vein* (a), *and textural relations in typical harzburgite EB 75*(b)

a. Wall rocks primary harzburgite, dunite, white; pyroxene, vertical lines; arrow indicates chromite or magnetite vein.

b. Typical harzburgite EB 75; single poikilitic orthopyroxene crystal with rounded serpentine pseudomorphs after olivine in it and a few olivine relics (dotted). Chromite (black).

TABLE 17.—*Chemical analyses of secondary dunites and serpentinites*

Analyst, A. H. Phillips

	1	2	3	4	A
SiO_2	40.06	40.64	39.29	35.48	42.30
Al_2O_3	1.67	1.25	.71	1.14	.59
Fe_2O_3	6.05	7.10	8.66	13.75	4.54
FeO	7.52	6.03	6.25	2.79	1.97
MgO	35.55	36.03	36.21	32.01	35.88
CaO	2.34	1.49	.00	.29	.05
Na_2O	.32	.48	—	—	—
K_2O	.12	.05	—	—	—
H_2O+	6.01	6.53	9.44	12.24	13.12
H_2O-	.28	.31		1.18	1.30
TiO_2	.06	.03	—	—	.00
Cr_2O_3	.21	.21	—	.72	—
MnO	.19	.18	—	.11	.03
P_2O_5	.00	tr	—	—	—
	100.38	100.33	100.56	99.78	99.78
Molecular per cent					
SiO_2	39.3	39.5	38.5	39.0	42.7
Fe_2O_3	2.3	2.5	3.1	5.5	1.5
FeO	6.0	5.0	5.1	2.5	1.5
MgO	52.4	53.0	53.3	53.0	54.3
$\frac{100(FeO+Fe_2O_3)}{MgO}$	100 8.3	100 7.5	100 8.1	100 8.0	100 3.0
Weight per cent					
H_2O (total)	6.3	6.8	9.4	13.4	14.4
Calculated per cent serpentine	43	47	65	93	100

1. MB 20 Dunite. East side Boulder Valley (Location shown on Figure 18a)
2. MB 4 Dunite. East side Boulder Valley (Location shown on Figure 18a)
3. Olivine nodule in bronzitite. Quoted from Peoples (1932, Ph.D. thesis, Princeton Univ.)
4. EB 92 Serpentinite. East Boulder Plateau, "Dunite Ridge" near base of complex

A. Serpentine. Pure serpentine from vein related to the dunite. East Boulder Plateau, "Dunite Ridge"

They crosscut and are younger than the primary layered harzburgites and bronzitites.

They appear to be developing directly from the harzburgite bands on "Dunite Ridge" but are found in the bronzitites only as clearly crosscutting dikes, pipes, and veins.

The lamination in relics of harzburgite surrounded by dunite maintains the strike and dip characteristic of the nearby outcrops of primary harzburgite. Evidently the relics have not been rotated.

They consist essentially of olivine and serpentine. Magnetite is fairly abundant in the more serpentinized varieties. Accessory chromite is present in minute amounts.

The fresh dunites contain coarse interlocking grains of olivine suggesting a pegmatitic type of crystallization.

They are richer in iron than are the primary ultramafic rocks. The Mg/Fe ratio is relatively low for an ultramafic rock.

They are analogous to the hortonolite dunites of the Bushveld, although considerably poorer in iron and to the black pipes associated with the Taguil (Ural Mtns.) platinum deposits.

The investigation of the system MgO-SiO_2-H_2O by Bowen and Tuttle (1949) supplies a basis for hypothesis as to origin of the dunites of the Secondary Group of Rocks. Hess (1938a) suggested that this group of rocks originated by partial melting of the primary serpentinized harzburgite forming a secondary magma capable of intruding the overlying bronzitites as dikes and pipes. Bowen and Tuttle's data offer a better explanation of both the chemistry and the field relations; therefore the writer discards his previous hypothesis.[7]

The regeneration of olivine by thermal metamorphism of serpentine was first noted by Durrell (1940) and later by Macdonald (1941), both in the Sierra Nevada. Leech (1953) has described much more extensive prophyroblastic development of olivine from serpentine in the Shulaps area of British Columbia. The regeneration of olivine from serpentine in the presence of water vapor above approximately 500°C. is amply demonstrated in Bowen and Tuttle's laboratory investigation. Regeneration of olivine from the primary serpentinized harzburgite seems to be the major element in the development of these *secondary dunites*. That the change has been accomplished largely in the solid state is indicated by the constancy of the strike and dip of the primary lamination in isolated relics of harzburgite surrounded by dunite and its consistency with the primary lamination of nearby bronzitites and harzburgites. The coarse interlocking texture of the olivine in the dunites is compatible with porphyroblastic crystallization.

The erratic distribution of areas of secondary dunite suggests that the heat necessary for the reaction may have been carried in by pneumatolytic solutions (presumably water vapor) rather than by general rise of the temperature of the surrounding terrane. The pipes and small dikes extending upward from large areas of dunite and sharply crosscutting overlying bronzitite correspond to features for which Bowen and Tuttle have suggested a pneumatolytic origin.

In a broad way the system MgO-SiO_2-H_2O seems to offer a general basis for understanding the secondary dunites, but many of the details remain unsolved, probably because we are dealing not with the closed system of a laboratory bomb, but with an

[7] The writer discards his previous hypothesis with regard to the origin of a so-called secondary peridotite magma, but he nevertheless continues to believe that the serpentinized peridotites of alpine-type mountain systems are derivatives of an ultramafic magma. The field evidence seems to him to necessitate such a hypothesis even though it is at present contrary to what is known of the chemistry.

open system—in which constituents in unknown amounts may both enter and leave. If the reactions were taking place in the region immediately above the stability field of serpentine (somewhat above 500°C.) we should expect to find the association of phases talc-enstatite-forsterite, talc-forsterite, forsterite-brucite, forsterite-brucite-periclase, or talc-enstatite-quartz. Actually the phases brucite-periclase-quartz have not been found. Enstatite (bronzite) is present apparently only as relics of the primary harzburgite, and the relations indicate that it is being resorbed and that where the reactions have gone to completion it has disappeared from the system. Talc is present in significant quantity in only one site examined. Two varieties of serpentine are present, but these are probably a younger generation of phases that developed after the temperature had again dropped. A trace of colorless amphibole thought to be tremolite is common and may represent a metastable phase corresponding to the anthophyllite mentioned by Bowen and Tuttle, but inasmuch as it is probably a calciferous amphibole it is outside the chemical field investigated by them.

The differences in chemical composition between the original harzburgite and present dunite are comparatively small. These differences no doubt were somewhat greater before the second stage of serpentinization which must have tended to move the composition back toward the original state. The Fe:Mg ratio of the dunite is higher than in the original rock. Residual concentration of Fe by extraction of Mg is contrary to the laboratory evidence of extreme lack of mobility of MgO under pneumatolytic conditions in this system. Introduction of the Fe seems more probable. This must occur during the high-temperature stage along with the crystallization of olivine inasmuch as no total increase of Fe is noted with increasing degree of serpentinization but only an oxidation of the Fe. Loss of SiO_2 and H_2O should have accompanied the crystallization of the olivine, but the same constituents are returned in the subsequent younger stage of serpentinization.

5. CONTACT OF ULTRAMAFIC ZONE AND NORITE ZONE

In the field the upper contact of the Ultramafic Zone is clear cut and sharp wherever observed. It is the most pronounced line of demarcation within the complex. Below the contact feldspathic augite-bearing bronzitite is found, and above feldspar-rich norites or anorthosites. It marks that episode in the crystallization of the Stillwater magma when plagioclase begins to accumulate on the floor in large amounts. In the Ultramafic Zone the plagioclase is interstitial and has crystallized from the liquid trapped between the accumulated pyroxene and olivine crystals. Across the upper boundary of this zone, peri-euhedral crystals of plagioclase are abundant. Only a few such crystals may be noted in the ultramafic rocks immediately below the boundary.

Table 18 gives the composition and textural changes observed in the vicinity of the boundary. The Mg content of the orthopyroxene decreases in the norite but is fairly constant through the 2500 feet of the Ultramafic Zone. The change in composition of the clinopyroxene is slight. The increase in An content of the plagioclase from 78–80 in the Ultramafic Zone to 85–86 in the Norite Zone is to be expected in view of the different conditions of crystallization. In the Ultramafic Zone the plagioclase crystallizes from the pore-space liquid, and in the Norite Zone it is predominantly settled accumulated crystals.

A few feet below the top of the Ultramafic Zone a peculiar anorthosite xenolith was found embedded in bronzitite. It is a slab about 6 inches thick by 2 feet long. The plagioclase in it is too badly altered to cloudy white material to permit determination of its composition. The origin of this xenolith is most puzzling, and no explana-

TABLE 18.—*Mineral variations at contact of Ultramafic and Norite zones*

	Ultramafic Zone			Norite Zone				
Spec. nos.....	EB 13	EB 14	EB 15	EB 16	EB 17†	EB 18	EB 19	
Feet from floor	2560	2785	2825	2900	3000	3160	3235	
Orthopyroxene	$Mg_{83.3}$*	Mg_{84}	Mg_{83}	Mg_{82}	$Mg_{80.5}$	Mg_{80}	Mg_{80}	Composition
Plagioclase	An_{78}	An_{79}	$An_{80.5}$	An_{85}	An_{83}	An_{86}*	An_{86}	
Orthopyroxene	78.2	86	73.5	43	trace	13.3	15	Mode in weight %
Clinopyroxene	10.3	5	2	3	trace	10.7	3.5	
Plagioclase	11.5	9	24.5	54	100	76	81.5	
Orthopyroxene	.7	1.8	1.8	1.5	—	1.0	1.3	Average grain intercept in mm.
Clinopyroxene	.6	1.7	.6	.5	—	.4	.5	
Plagioclase	.3	.8	.9	1.0	1.3	.8	.9	
All grains	.6	1.6	1.3	1.1	1.3	.75	.95	
Orthopyroxene	P to E	P to E	P to E	P to A	—	P	P to A	Textural Relations
Clinopyroxene	A	A	A	A	—	A	A	
Plagioclase	A	A	A	P to E	A	P to E	P to E	

* Chemically analyzed

† Pegmatitic anorthosite; some grains several cm in diameter; great variation in grain size

E = Euhedral

P = Peri-euhedral

A = Anhedral or crystallization interstitial to accumulated crystals

tion is offered. Cognate xenoliths of bronzitite are found at several horizons in the lower portion of the Norite Zone. These are identical to bronzitites found lower in the complex and do not present so much of a problem.

D. NORITE ZONE

1. GENERAL STATEMENT

No marked differences are found in the broader features of the rocks of the four upper zones (Norite, Lower Gabbro, Anorthosite, Upper Gabbro, Fig. 16). The essential minerals—calcic plagioclase, orthopyroxene, and clinopyroxene—are common to all. The only other mineral present in more than minute amounts is olivine. The amount of olivine in the upper zones is quantitatively very small. It occurs only at certain horizons in the Anorthosite Zone. The subdivision of the upper part of the complex is based chiefly on the relative proportions of the three essential minerals and secondarily on certain textural relations.

The Norite Zone is estimated to be 2740 feet thick. The basal 200 feet is composed

of normal norites interlayered with anorthosites. On the average the amount of pyroxene decreases upward so that above the basal part of the zone anorthosites and anorthositic norites predominate for the next 2000 feet, except for a horizon represented by EB 21, 2390 feet above the base of the zone. EB 21 is a norite containing 60 per cent pyroxene. Toward the top of the zone pyroxene again increases so that the upper 550 feet averages more than 40 per cent pyroxene compared to 20 per cent for the main part of the zone below. One hundred and fifty feet from the top of the zone pyroxene increases to such an extent that the rocks may be classified as pyroxenites (EB 199, EB 200). Above the pyroxenite the uppermost rocks of this zone are norites containing 40–50 per cent pyroxene.

The Norite Zone exhibits the finest examples of rhythmic layering found in the complex. The thickness of the layers is highly variable. One may find a layer of pyroxene crystals only one or two crystals thick in a larger layer of anorthosite. Layers of anorthosite or norite of fairly uniform composition from top to bottom ranging from 50 to 100 feet thick may also be observed. The rhythmic layers commonly range from a few inches to a few feet thick. About two-thirds of them show gravity stratification. From a sharply defined lower contact of pyroxene-rich rock they grade upward into more feldspathic rock or anorthosite.

A few small bronzitite xenoliths were noted in the lower part of the Norite Zone on the East Boulder Plateau. The xenoliths evidently are derived from a bronzitite horizon in the Ultramafic Zone. The bronzite correspond in composition (approximately Mg_{86}) with that of the Ultramafic Zone bronzitite and is considerably more magnesian than the bronzite of the surrounding norites. Xenoliths of the same type are present in the lower part of the Norite Zone of the Stillwater Valley area, but there they are larger and more abundant. No harzburgite xenoliths have been found.

2. TEXTURAL RELATIONS

In the Norite Zone both plagioclase and orthopyroxene occur predominantly as peri-euhedral crystals representing the crystal mush of these phases from which the rocks were formed. Interstitial anhedral augite has crystallized entirely from the pore-space liquid of the original mush. A certain amount of late plagioclase and orthopyroxene also crystallized from the interstitial liquid. In the norites the late plagioclase and orthopyroxene normally have rimmed the already present crystals of the same species, but in anorthosites the late orthopyroxene may be anhedral and interstitial to the plagioclase; in the rare pyroxenites the plagioclase may be similarly interstitial to the orthopyroxene. Quartz in minute amounts makes its first appearance in this zone. It is the last mineral to crystallize and fills the last remaining pores before complete solidification of the rock. Augite may show crystal faces where it borders on quartz. An ophitic type of crystallization of minerals from the pore-space liquid is fairly common. In a few exceptional samples large crystals of plagioclase, which have grown from the pore-space liquid, poikilitically include not only orthopyroxene but also many smaller accumulated plagioclase crystals of the mush.

3. MINERALOGY

The plagioclase and orthopyroxene representing the accumulated crystals range in composition from An_{86} to An_{78} and Mg_{83} to Mg_{72} respectively. In one specimen,

EB 199, the plagioclase is An_{73}. From the interstitial liquid the same two minerals plus clinopyroxene have crystallized. The clinopyroxene of all samples examined (with one exception, EB 21) is anhedral and commonly forms skeleton poikilitic crystals so that small interstitial patches scattered over the area of a whole thin section extinguish simultaneously. The clinopyroxene ranges in composition from approximately $Ca_{41}Mg_{49}Fe_{10}$ to $Ca_{41}Mg_{47}Fe_{12}$.

It is characteristic of rocks formed by crystal accumulation that the common accessory minerals are exceedingly rare or, not uncommonly absent. Partridge (1934) ground up several pounds of the Bon Accord norite of the Bushveld complex for the purpose of separating and identifying the accessory minerals. He found no magnetite, ilmenite, apatite, or zircon and only a trace of iron oxide (hematite) and sulfide.

Except for the sulfides, all the minor primary constituents of the rocks, here grouped under the heading Accessory Minerals, have crystallized late. The sulfides, probably chalcopyrite and pyrrhotite chiefly, are not uncommonly inclusions in crystals of the essential minerals. They appear as rounded grains and may represent droplets of an immiscible sulfide liquid which accumulated on the bottom along with the accumulated silicates. Traces of sulfide are present in almost every rock examined, whereas the distribution of the other accessory minerals is much more irregular. Excluding the sulfides, the accessory minerals appear to have crystallized after the late, essential-mineral, silicates. The total volume of accessory minerals present is in all cases less than 1 per cent of the rock, and as a rule it amounts to a minute fraction of 1 per cent.

In 30 specimens of rocks from the Norite Zone traces of sulfides were noted in all but 2. Quartz was found in 17, and black opaque minerals (ilmenite-magnetite or hematite) in 4. Biotite was observed in 6, rutile in 3, amphibole in 2, and apatite in 1. Grains of sulfide are rimmed by a black opaque mineral. It is uncertain whether this rim is an alteration product or represents late crystallization. Secondary minerals produced by post-consolidation alterations have not been included above. A few specimens show partial alteration of plagioclase to clinozoisite-epidote minerals and of pyroxene to chlorite-amphibole-magnetite.

4. BULK COMPOSITION

Based on 33 Rosiwal analyses of rocks from the Norite Zone the bulk mineralogical composition of the zone is as shown in Table 19.

The analyses were not merely averaged directly but were multiplied by the thickness of the layers which they represent and then all of them added and recalculated to 100 per cent. This was necessary since relatively more samples were collected from some parts of the zone than from others. The pyroxene ratio {100 orthopyroxene/orthopyroxene + clinopyroxene} is characteristically high in the Norite Zone. Excluding the anorthositic rocks (with 75 per cent plagioclase or greater) the average pyroxene ratio is 85 based on 25 specimens. Only two of these specimens have pyroxene ratios below 72 (that of EB 21 is 64, that of EB 182 is 53). After clinopyroxene appears among the early accumulated crystals (in the overlying zone) the ratio drops abruptly to 50 or less. This feature is used to delineate the boundary between the Norite Zone and Lower Gabbro Zone above it.

5. DESCRIPTION OF SPECIMENS

Anorthositic norite specimen EB 18, 3160 feet from the base of the complex or 260 feet from the base of the Norite Zone, is fairly typical of the lower half of this zone. It consists of 76 weight per cent plagioclase, An_{86}, 13⅓ per cent orthopyroxene, Mg_{81}, and 10⅔ per cent clinopyroxene as estimated from Rosiwal analysis. The

TABLE 19.—*Bulk mineralogical composition of Norite Zone and average composition of its minerals*

	Weight per cent	Average composition of minerals in molecular per cent
Plagioclase	63½	An_{81}
Orthopyroxene	28	Mg_{78}
Clinopyroxene	8½	$Ca_{41}Mg_{48}Fe_{11}$
	100.0	

grain size is normal for the Norite Zone; the average grain intercept is about 0.75 mm Average grain intercept is about half the average dimensions of the grain.

All the clinopyroxene has crystallized at a late stage as poikilitic crystals including peri-euhedral plagioclase and bronzite. Both plagioclase and orthopyroxene form peri-euhedral to subhedral crystals. It is evident from the texture that much of the plagioclase and bronzite represents accumulated crystals and that clinopyroxene, as well as some additional plagioclase and orthopyroxene, has crystallized from the interstitial liquid. The amount of accumulated bronzite present is rather small so that the effect of addition of a little more orthopyroxene from the interstitial liquid is more pronounced in changing orthopyroxene composition than where the crystalline phase under consideration is a more abundant constituent of the rock.

Figure 18a is a tracing made from a projected thin section showing one of the poikilitic areas of clinopyroxene and the general relations between the three essential minerals. Accessory minerals are negligible. A few minute grains of sulfide were noted as well as a smaller amount of black opaque minerals, probably hematite or ilmenite. An analysis of the plagioclase of this specimen appears in analysis 1 of Table 10. The rock as a whole is similar to Peoples' (1932, Ph.D. thesis, Princeton University) analyzed anorthositic norite 463 C 2, collected from an equivalent stratigraphic horizon in the eastern portion of the complex.

Specimen EB 21 is a mafic norite from the 3890-foot horizon. It contains 36 per cent plagioclase, An_{84} (Table 10, analysis 3), 41 per cent orthopyroxene, Mg_{80}, and 23 per cent clinopyroxene. It is abnormal in that some of the clinopyroxene occurs as rounded to peri-euhedral grains of about the same size as the peri-euhedral bronzite. This is the only observed specimen, within the Norite Zone, in which clinopyroxene appears to be among the accumulated crystals. Most of the bronzite is peri-euhedral, and the plagioclase is peri-euhedral to subhedral. In this particular rock it is also evident that some of the plagioclase has crystallized from the interstitial liquid. A

few rather large plagioclase crystals include several small nearly euhedral crystals of plagioclase. There is little or no difference in composition between the early and the late plagioclase. A little accessory sulfide is present, and a few small patches of alteration may be seen in which epidote, clinzoisite, chlorite, amphibole, and magnetite occur.

Hypersthene-bearing anorthosite EB 30 from the 5060 horizon near the middle of the Norite Zone contains 90 per cent plagioclase, An_{80}, 9½ per cent orthopyroxene, and ½ per cent clinopyroxene. The pyroxene is altered, but the plagioclase is fresh. Most of the pyroxene has crystallized late and is interstitial to the peri-euhedral to subhedral, predominantly early plagioclase. A few crystals of peri-euhedral hypersthene are present. Late poikilitic hypersthene includes euhedral plagioclases and has narrow rims of still later clinopyroxene. In some cases, outward from poikilitic hypersthene is a wide border zone of "skeleton poikilitic" clinopyroxene. These "skeleton poikilitic" crystals of clinopyroxene are seen in the hand specimen as large spots, single crystals, perhaps 3 cm in diameter, but within the spot only 10–20 per cent of the volume consists of clinopyroxene, and the remainder is included crystals, mostly plagioclase.

This specimen shows the first quartz found above the base of the complex in the East Boulder Plateau section. About 0.1 per cent of quartz is present commonly as triangular areas interstitial to the plagioclase laths. It is the last mineral in the rock to crystallize.

At 5500 feet in the complex, 140 feet from the top of the Norite Zone, a conspicuous pyroxenitic horizon was noted. This is one of two pyroxenite occurrences above the Ultramafic Zone. EB 199 is a mafic augite norite to feldspathic augite-bearing hypersthenite. The specimen studied contains 74¾ per cent by weight of hypersthene, Mg_{78}, 9 per cent clinopyroxene, and 16¼ per cent plagioclase, An_{73}. The hypersthene is nearly euhedral, and both the plagioclase and clinopyroxene are interstitial forming poikilitic crystals. The textural relations are shown in Figure 18b, a tracing from a projection of the thin section. It is one of the rare rocks from which an estimate of the total volume of pore space may be made since the orthopyroxene alone represents the accumulated crystals and clinopyroxene plus plagioclase represent all the material which crystallized from the interstitial liquid. In this case the volume of plagioclase + clinopyroxene amounts to 26.4 per cent of the rock (calculated from Rosiwal analysis of four thin sections 2 cm by 3 cm). Results obtained from other similar rocks range from 26 per cent to 29 per cent for the volume of the interstitial liquid. The clinopyroxene forms rather larger poikilitic crystals (±1.5 cm across), and the plagioclase smaller poikilitic crystals (±.5 mm across). A minute amount of sulfide and biotite was noted as well as a few very small areas of green amphibole which may be an alteration product.

The pyroxenite grades both upward and downward into norite. A norite below the pyroxenite, EB 32a, furnished material for both hypersthene and plagioclase chemical analyses.

6. DEAD TREE DETAILED SECTION

It was thought advisable to study a small part of the East Boulder Plateau section in great detail in order to correlate minor changes in composition and texture in the

rocks and to determine exactly how rapidly these changes take place. For this purpose 42 specimens were collected in the 1220 feet from 4760 to 5980 feet in the section. Previously a suite of specimens collected from a much shorter section and spaced a few inches to a few feet apart was studied, but changes in the composition of the minerals were barely appreciable so it was apparent that a longer section with wider spacing of samples was necessary. Doten, Davis, and the writer collected specimens from the Dead Tree section in 1935. After study of these a large number of additional samples were collected by Peoples and Howland in 1936 at specified points along the section. The results of this detailed investigation are given in Table 20 and also presented in the form of curves with discussion in the chapter on Mineral-Variation Curves (Pl. 9). The Dead Tree section includes the upper part of the Norite Zone and lower part of the overlying Lower Gabbro Zone.

The bulk mineralogical composition of the detailed section is 62% plagioclase, 28% orthopyroxene, 9% clinopyroxene, and 1% accessory minerals.

E. LOWER GABBRO ZONE

1. GENERAL STATEMENT

The Lower Gabbro Zone is estimated to be 2210 feet thick extending from the 5640-foot horizon to the 7850-foot horizon. It consists predominantly of hypersthene gabbro and anorthositic hypersthene gabbro. Lesser amounts of anorthosite and augite-rich norite are also present. The ratio of clinopyroxene to orthopyroxene is close to 1:1, but clinopyroxene tends to be slightly in excess. The base of the zone represents the horizon at which clinopyroxene joins plagioclase and orthopyroxene as one of the phases settling and accumulating on the floor. As in the Norite Zone the lower and middle portions of the Lower Gabbro Zone are anorthositic, averaging about 20 per cent of total pyroxene, and the upper portion becomes progressively more mafic until a pyroxenitic facies results (Eb 175) about 400 feet from the top, after which the pyroxene content drops off rapidly to the upper contact of the zone.

The exposures in the central portion of the zone are poor along the East Boulder line of section, and the relations are complicated by a normal fault striking parallel to the layering. The fault itself is found in a saddle in the lower part of the zone. It appears to be vertical and has an estimated displacement of 800 feet with the north side up. The fault-line scarp has receded for a considerable distance northward, and its slope toward the fault line is talus-covered, accounting for the lack of outcrops in this portion of the zone. A fault of this displacement would cause on its north side a repetition of some layers found on the south side. The area in which this repetition would take place is covered. South of the fault a small patch of limestone—probably Cambrian—is found on the surface lying unconformably on the Precambrian.

2. TEXTURAL RELATIONS

The only significant change in textural relations in the rocks of this zone as compared to the Norite Zone is in the form of the clinopyroxene. Instead of crystallizing late and commonly forming poikilitic crystals including the other two essential minerals, it crystallizes in part early, so that distinct but rounded or irregular crystals are formed which are comparable in size to the average early orthopyroxene.

The orthopyroxenes are more nearly euhedral than the clinopyroxene crystals; it would seem that more material was added to the clinopyroxene crystals from the interstitial liquid, after they accumulated on the bottom, than was added to the orthopyroxene. Since orthopyroxene begins to crystallize considerably before clinopyroxene in the crystallization sequence of the accumulated crystals it may well be that more clinopyroxene than orthopyroxene will crystallize from the interstitial liquid. It is, however, difficult to make any generalization on this point since a number of specimens contain a small amount of very late orthopyroxene definitely younger than late clinopyroxene.

3. MINERALOGY

The three essential minerals—plagioclase, hypersthene, and augite—range in composition between the following limits:

Plagioclase	An_{86} to An_{76}
Orthopyroxene	Mg_{81} to Mg_{73}
Clinopyroxene	$Ca_{41}Mg_{49}Fe_{10}$ to $Ca_{41}Mg_{45}Fe_{14}$

Explanation of Table 20

In column 5 figures are given for the average grain intercept determined while making the Rosiwal analyses of the rocks. It is arrived at by dividing the total number of grains of a particular mineral found in the Rosiwal analysis into the total length of the traverse across these grains. It is the most satisfactory method of arriving at an absolute figure for grain size. In the case of poikilitic crystals the traverse across a given crystal may be broken into several parts by its included crystals of other minerals. The disconnected portions of one poikilitic crystal may either be added together and considered to be one grain or they may be considered as separate grains. For this investigation the latter method was used.

The composition of the plagioclase is given in molecular per cent of anorthite. The composition of the orthopyroxene is given in per cent of Mg ions of the total $Mg^{+2} + Fe^{+2} + Fe^{+3}$ assuming that these are pure Mg and Fe pyroxenes. Actually about $3\frac{1}{2}$ per cent of Ca^{+2} ions is present in these crystals along with the other two components. The clinopyroxene composition is less accurately determined than the compositions of the plagioclase and orthopyroxene. It is based on the assumption that the crystals of clinopyroxene are all of the normal type similar to the analyzed specimens. Only Nx was determined, and this was converted to per cent Ca^{+2}, Mg^{+2}, $(Fe^{+2} + Fe^{+3})$ from the curve based on the analyzed specimens.

Under Time of Crystallization the minerals are listed as having crystallized early (E) or late (L). Where most of the crystallization was early the notation given is "E + L"; where most was late, "L + E". These determinations were based on complex textural relations of the rocks and are necessarily subject to uncertainties. Nevertheless it gives a fairly good approximation of the relations in a manner which may be easily used for comparison with composition and grain-size variations.

Under accessory minerals the term *black opaque* was used so that both ilmenite-magnetite and hematite could be included under a single heading because it is as a rule difficult to distinguish between the two in thin section. The accessory minerals make up less than 1 per cent of any of the rocks in this table and commonly are present only in a minute fraction of 1 per cent. Minerals which are definitely secondary alteration products of the rock are not included in the list.

In column 10 pyroxene ratio refers to the quantity

$$\frac{\text{orthopyroxene}}{\text{orthopyroxene} + \text{clinopyroxene}} \times 100$$

Compositions in the table are in many cases given in fractional parts of a per cent such as $Mg_{70\frac{1}{2}}$ or $An_{80\frac{1}{2}}$ although the accuracy of the measurements does not warrant such fractions. The compositions given are those read from optical-property curves, and the fractions merely place the samples in the correct positions with respect to one another.

	1	2	3			4			5				6			7			8		9		10	11
			VOL. %		MODE	WGT. %		MODE	AV. GRAIN		INT'	CP'T	PLAGIOCLASE			ORTHOPYROXENE			CLINOPYROXENE		ACCESSORY MINERALS		Px RATIO	REMARKS
	Specimen Number EB	Height From Floor (feet)	Pc	Opx	Cpx	Pc	Opx	Cpx	Pc	Opx	Cpx	Rk	Comp % An	Range	Time Cryst.	Comp % Mg	Range	Time Cryst.	Comp Ca Mg Fe	Time Cryst.	Sulfides Black Opaque Quartz Rutile Apatite Biotite Amphibole			
Lower Gabbro Zone	218	6140	63.5	18	18.5	60.5	19.5	20	.48	1.30	1.10	.75	78	Small	E+L	77	Small	L?	41-47.5-11.5		X		49	Badly altered
	217	6110	78	7	14.25	75.25	8	16	.33	.66	.70	.41	78	76-78.5	E+L	77.1	Small	L+E	41-46.5-12.5	E+L	X		-	Small and large Pc
	216	6052	65.5	13.5	21	62.5	14.5	23	.36	1.00	.89	.56	77	Small	E+L	76.8	Small	L+E	41-46 -13	L+E	X		39	Pc as above
	214	5935	54	26	19	51	28	20	.30	1.20	.80	.64	-	-	E+L	77.1	76.5-77.6	L+E	41-49 -10	E+L	X		56.5	Pc as above. Rk altered
	213	5920	66.5	15.25	18.25	63.5	16	19	.26	.74	.70	.35	78	Small	E+L	77.6	77.1-78.2	L+E	41-48 -11	E+L	X		40	Pc as above
	212	5900	76	16.25	7	74	18	8	.33	1.40	.93	.54	-	-	E+L	-	-	E	-	E+L			69	Pc as above. Px altered
	211a	5880	72	11.25	16	69	12.5	18.5	.37	1.10	.70	.50	-	-	E+L	-	-	L+E	-	L+E	X		42	Pc as above
	211	5865	70	13	17	67	14.25	18	.28	.96	.75	.54	78.5	Small	E+L	77.3	77-78	E+L	-	E+L	X		43	Early Pc crystals very small
	210	5830	77	0	23	74	0	25.25	.63	-	.68	.64	78	Small	E+L	70.7	Small	-	-	L+E		X	-	Badly altered
	209	5790	56	22	22	53	23.5	23.5	.49	.97	.69	.59	79	77-79.5	E+L	71.3	70.5-72.5	L+E	40-43 -17	E+L	X X X		50	Altered Px
	208	5758	76	16	8	73	17.5	8	.66	1.00	.59	.70	79.5	Small	E+L	78	Small	L+E	41-47.5-11.5	E+L	X		66.5	
	207	5712	94	6	0	94	6	0	.88	.49	-	.84	80.25	79.25-80.5	E+L	72.2	71.5-72.5	L+E	40-46 -14	-	X	X	-	Altered Px
	206	5679	61.5	14.5	24	58.5	15.5	26	.43	.70	.60	.49	78.5	Small	E+L	77.7	Small	L+E	41-49 -10	E+L	X		37.5	
	205	5652	76	4	20	73	4	22.25	.68	.73	.76	.70	78	Small	E+L	73.7	Small	L+E	40-44 -16	E+L	X X X		17	
Norite Zone	204	5614	55	33	12	52	35	13	.55	1.30	.82	.71	78.5	Small	E+L	77.2	Small	E	40-45.5-14.5	L	X X X	X	73	
	203	5583	50	46	4	47	49	4	.57	1.40	.36	.76	-	-	E+L	77.5	Small	E	41-49 -10	L	X X		92	
	202	5545	63	36	1	60	39	1	.96	1.60	.30	1.10	78.25	Small	E+L	77	Small	E	41-46.5-12.5	L	X X X	X	98	Determination not possible
	201	5522	39	47	14	36	49	15	.85	1.50	1.10	1.10	-	-	E+L	77.2	Small	E	41-49.5- 9.5	L	X X X X	X	77	EB 200 Pyroxenite badly altered
	199	5498	18.6	73.6	7.8	16.25	74	9	.82	1.80	1.70	1.50	73.25	Small	L	77.9	Small	E	41-48 -11	L	X	X X	83	
	198	5480	43	55	2	40	58	2	.55	1.00	.21	.70	78.25	Small	E+L	78.2	Small	E	-	L	X X		96.5	
	197	5468	51	43	6	48	45.5	6.5	.59	1.10	.61	.74	-	-	E+L	78.6	Small	E	41-47.5-11.5	L	X X		88	
	196	5444	62	31	7	59	34	7	.53	.84	.56	.60	-	-	E+L	78.4	Small	E	41-49.5- 9.5	L	X		82	
	195	5428	45	49	6	42	51.5	6.5	.51	.91	.33	.62	-	-	E+L	78.5	Small	E	-	L	X X		89	
	194	5417	57	32	10	54	34.5	11	.69	.60	.70	.66	-	-	E+L	78.3	Small	E	-	L	X X		75	More sulfide than commonly
	193	5410	77	21	2	74.5	23.25	2.25	.56	1.30	.30	.62	-	-	E+L	77.8	Small	E+L	41-46 -13	L	X X X		91	
	192	5400	62.5	31.25	6.5	59.5	33	7	.52	1.50	.25	.60	79	Small	E+L	77.5	77-78	E+L	-	L	X X X		83	
	191	5367	73	22	5	70	24	6	.46	.92	.38	.52	79.25	Small	E+L	77.2	Small	E+L	-	L	X X		81.5	
	190	5333	69	24	7	66	26	8	.53	1.00	.38	.58	79.25	Small	E+L	76.5	Small	E+L	-	L	X X X X		74	
	189	5300	60.5	32	6	57.5	35.5	7	.62	1.50	.49	.73	79.25	Small	E+L	76.4	Small	E+L	41-46.5-12.5	L	X X X	X X	83	
	188	5260	70	27	3	67	29	3.25	.56	.91	.42	.61	79	76.25-79.5	E+L	76.5	Small	E+L	-	L	X X X X	X	90	
	187	5220	73	23	4	70	25	5	.55	1.00	.44	.61	80	78 - 80	E+L	76.7	76.4-78.0	E+L	40-46 -14	L	X X X		83	
	186	5205	62	37	1	59	40	1	.72	2.10	.20	.91	81.25	79 - 82	E+L	71.4	70.0-72.0	L	40-46 -14	L	X X		97.5	
	185	5180	51	35	13	48	37	14	.43	1.00	.80	.63	81	79 - 81.25	E+L	78.2	Small	E+L	-	L	X		72	
	184	5155	51	43	6	48	45.5	6.5	.38	.86	.29	.49	-	-	E+L	78.6	Small	E+L	41-46.5-12.5	L	X		87.5	
	183	5125	67	33	tr	64	36	tr	.90	2.30	-	.90	80.5	79 - 81	E+L	77.5	Small	E+L	41-46.5-12.5	L	X		99	
	182	5045	51	26	23	47	27	24.5	.70	1.30	.80	.86	80.25	79 - 81	E+L	72.5	Small	E+L	40-44 -16	L	X		53	More Ilm-Mt than commonly
	181	4985	91	3	6	90	3.25	6	.60	.40	.40	.60	80	79 - 81.5	E+L	71.3	-	L	40-43.5-16.5	L	X X X		-	
	180	4920	69	29	2	66	32	2	.60	2.00	.20	.70	81	80 - 82.5	E+L	75.9	Small	E+L	41-49.5- 9.5	L	X X X		93	

TABLE 20.—*Mineral variations in the Dead Tree detailed section*

Compositions of minerals are given to a fraction of a per cent although absolute accuracy of determination was much less. This was done merely to list them in proper relative positions. The Rosiwal analyses also have much larger errors than suggested by the 25.

Chemical analyses of the plagioclase and hypersthene from specimen EB 38, of plagioclase from EB 68 and EB 72, and of the augite from EB 175 are given in the chapter on Mineralogy.

In 20 specimens examined all but one contained some accessory minerals; 16 had a little sulfide, 8 ilmenite-magnetite, 7 quartz, 2 green amphibole, and 1 biotite. In each specimen the total amount of accessory minerals was exceedingly small.

TABLE 21.—*Bulk mineralogical composition of Lower Gabbro Zone and average composition of its minerals*

	Weight per cent	Average composition of minerals in molecular per cent
Plagioclase	60½	An_{79}
Orthopyroxene	17	Mg_{77}
Clinopyroxene	22½	$Ca_{41}Mg_{48}Fe_{11}$
	100.0	

4. BULK COMPOSITION

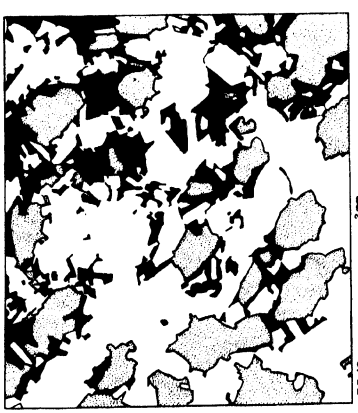
On the basis of 20 Rosiwal analyses the composition of the Lower Gabbro Zone is as shown in Table 21, neglecting the small amount of accessory minerals.

The pyroxene ratio is 45 or nearly 1:1 in contrast to 85 for the Norite Zone (excluding the anorthositic rocks).

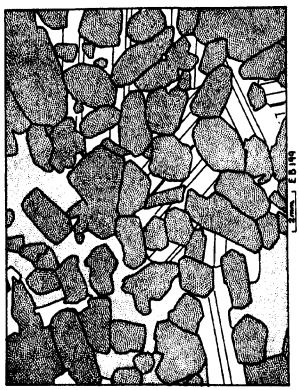
5. INCH-SCALE LAYERING

As indicated earlier in this chapter, a peculiar variety of layering, designated as *inch-scale layering*, was observed in the Lower Gabbro Zone. Outcrops presenting a plane perpendicular to the layering show a regular repetition of plagioclase and pyroxene bands. These bands, from the center of one pyroxene layer to the center of the next pyroxene layer, have a very uniform thickness, five-sixths of an inch. In polished hand specimens the contact between the pyroxene layer and the plagioclase layer presents a wavy appearance without any suggestion of early euhedral pyroxene projecting into the plagioclase portion of this layer. In this respect the layers differ markedly from the normal rhythmic layering. Rhythmically layered rocks consisting of more than 30 per cent pyroxene contain a fair proportion of peri-euhedral pyroxene crystals. This suggests the possibility that the pyroxene crystals in the inch-layered rocks have crystallized wholly or to a large extent from the interstitial liquid. The problem is discussed further in Chapter VII.

Inch-scale layering is well developed in the cliff outcrops west of "Cliff Station" (Pl. 8) near the East Boulder River. The inch-scale layering occurs near the top of the Lower Gabbro Zone. It was not observed on the line of traverse across the East Boulder Plateau. It may be covered there or may die out before reaching the line of the cross section. Inch-scale layering was also observed at a similar horizon on the east side of the Stillwater River Valley.



a



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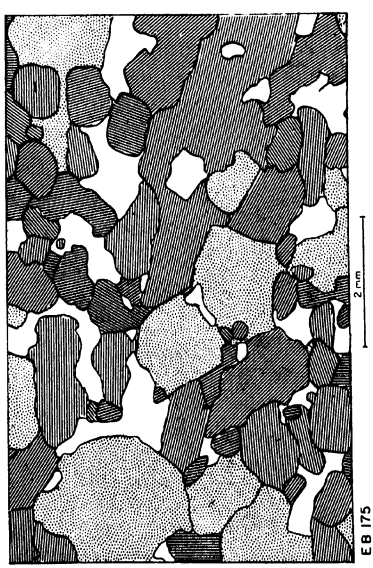

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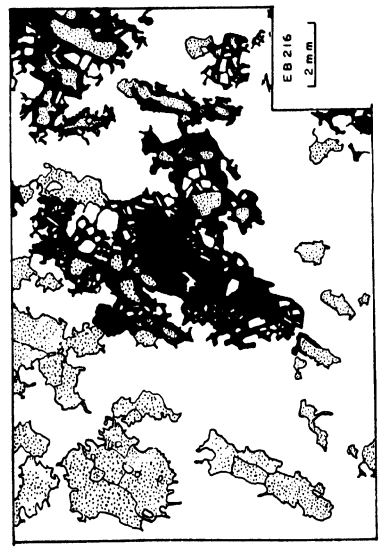


d

FIGURE 18

6. DESCRIPTION OF SPECIMENS

EB 175 is a mafic hypersthene gabbro or very nearly a feldspathic websterite having the following composition:

	Weight per cent	*Composition*
Orthopyroxene	40	Mg_{77}
Clinopyroxene	45	$Ca_{42}Mg_{46}Fe_{12}$
Plagioclase	15	$An_{76.5}$

It occurs 400 feet from the top of the Lower Gabbro Zone at 7410 feet from the base of the complex. EB 175 (Fig. 18) represents one of the two ultramafic horizons above the Ultramafic Zone. (The other is the EB 199–EB 200 horizon in the Norite Zone previously described.) In textural relations EB 175 is markedly different from EB 199. All three of the essential minerals—plagioclase, orthopyroxene, and clinopyroxene—are represented in the accumulated crystals in contrast to only orthopyroxene in EB 199. All three essential minerals show peri-euhedral to subhedral forms. The clinopyroxene seems to have continued to crystallize slightly longer than the other two, so that its boundaries are somewhat more irregular, and it may engulf small crystals of plagioclase and rarely orthopyroxene. The orthopyroxene may also include small crystals of clinopyroxene or plagioclase but only near borders of the grains, indicating a central accumulated crystal and later marginal addition. Plagioclase crystals rarely include pyroxene, but a few include clinopyroxene. The lack of included orthopyroxene may be ascribed to the uniformly larger size of the accumulated orthopyroxene crystals (1 to several millimeters in maximum diameter). Many of the early clinopyroxene and plagioclase crystals apparently were less than half a mm in diameter. The fact that inclusions in orthopyroxene are limited to the margins of crystals whereas some clinopyroxene grains show inclusions well toward the center may be ascribed to the fact that some of the larger clinopyroxene crystals have crystallized to a considerable extent from the interstitial liquid.

Figure 18c shows the general distribution of minerals and their textural relations. A chemical analysis of the clinopyroxene of this rock is given in the chapter on Mineralogy.

A small amount of sulfide is present in the rock either along grain boundaries or included in the margins of clinopyroxene. A little black opaque oxide, probably ilmenite-magnetite, is associated with the sulfide. It commonly surrounds sulfide grains. Many of the sulfide grains are embedded in opaque material which appears brown in reflected light and probably is an alteration product, limonite.

A little dark-brown biotite was noted. Some appears to be a primary accessory

---

FIGURE 18.—*Textural relations of minerals in EB 18, EB 199, EB 175, and EB 216*

a. EB 18. White plagioclase and dotted orthopyroxene which is peri-euhedral; settled crystals with considerable addition from pore-space liquid. Black is interstitial augite.

b. EB 199. Euhedral orthopyroxene (coarse dots), interstitial augite (fine dots), and interstitial plagioclase (white).

c. EB 175. Peri-euhedral clinopyroxene (lined), peri-euhedral orthopyroxene (dotted), and interstitial plagioclase (white).

d. EB 216. Black, orthopyroxene; dotted, clinopyroxene; white, plagioclase

mineral and some an alteration product of pyroxene. Some magnetite-chlorite-limonite alteration is also noted. Alteration is very slight.

Except for the high ratio of pyroxene to plagioclase the rock is fairly typical of specimens from the Lower Gabbro Zone.

Specimens EB 208 and EB 209 from near the top of the Dead Tree detailed section illustrate an apparent lack of correlation of textural features with the composition

TABLE 22.—*Mineral composition of EB 208 and EB 209*

	Weight per cent in rock			Composition of crystals		
	Plagioclase	Ortho-pyroxene	Clino-pyroxene	Plagioclase	Orthopyroxene	Clinopyroxene
EB 208	73¾	17½	8¾	$An_{79\frac{1}{2}}$	Mg_{78}	$Ca_{41}Mg_{49}Fe_{10}$
EB 209	53	23½	23½	An_{79}	Mg_{72}	$Ca_{41}Mg_{45}Fe_{14}$

of the minerals. EB 208 is from 5758 feet and EB 209 from 5790 feet, which is in the lower portion of the Lower Gabbro Zone. Both specimens have rather small plagioclase crystals which appear to be predominantly of early crystallization (accumulated crystals). A great many very small almost euhedral plagioclase crystals are included in the pyroxenes. Figure 18d shows a rock with similar textural relations (EB 216).

In EB 208 both the orthopyroxene and clinopyroxene are in large part late (have crystallized from the interstitial liquid). Cores in some of the clinopyroxene crystals appear to represent early accumulated crystals, but the very ragged outline of the crystals and abundant inclusions along the margins indicate that a considerable proportion crystallized late. The orthopyroxene may have some early cores but is also predominantly late and definitely continued to crystallize after the clinopyroxene had stopped so that in some specimens orthopyroxene rims clinopyroxene).

In EB 209 the pyroxenes are similar, but apparently more of the clinopyroxene crystallized early. The orthopyroxene of EB 208 continued to crystallize after the clinopyroxene ceased, and a larger proportion of the orthopyroxene is definitely later than in EB 209. Table 22 shows the composition of both rocks.

The plagioclase in both rocks is early, and the pyroxene in both rocks is predominantly late and interstitial. One might expect the pyroxene, because it is late, to be more ferriferous than the early pyroxenes in adjacent rocks; but in EB 208 the pyroxene is quite as magnesian as the early pyroxene crystals in the nearby rocks. In EB 209, however, the pyroxene is ferriferous even though more of it is of early crystallization than in EB 208. One might, therefore, object to the generalization that the minerals which crystallize from the late interstitial liquid have a greater percentage of the less refractory constituents inasmuch as these data indicate poor correlation between being late in the crystallization sequence and having more of the low-melting constituents. This generalization does actually hold for most cases, and there is another generalization which is valid in nearly all cases—namely, if a pyroxene is relatively richer in iron as compared to the same mineral in adjacent layers,

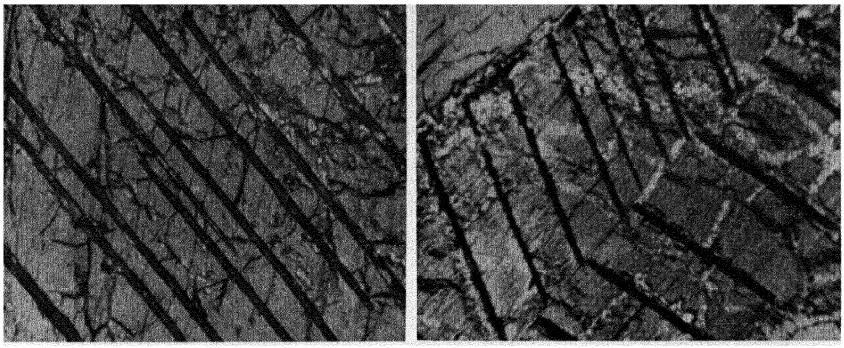
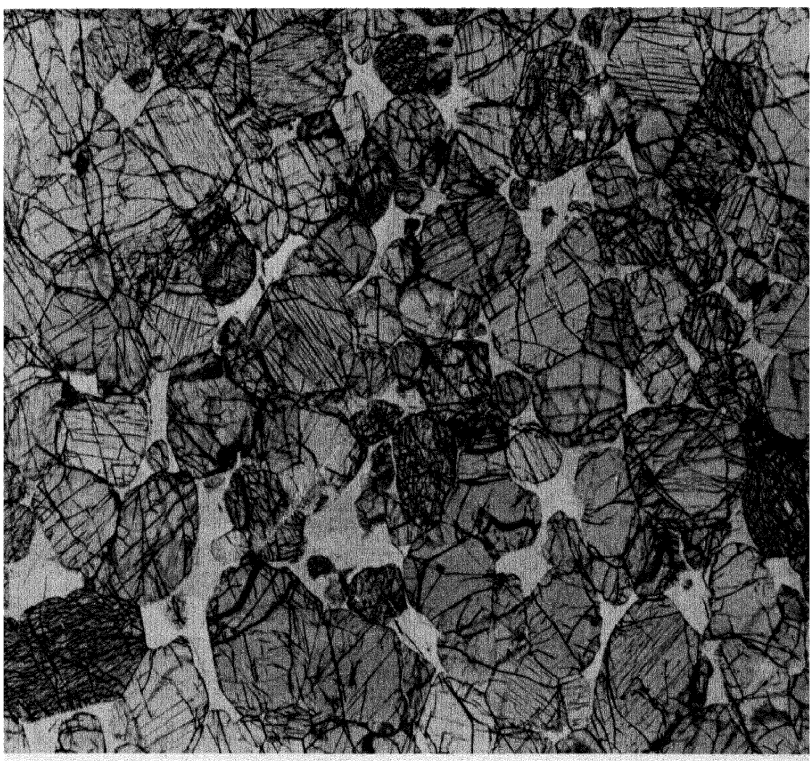


FIGURE 1. — Pigeonite inverted to hypersthene with coarse augite lamellae parallel to (001) of original pigeonite, EB 135



FIUGRE 2. — Feldspathic, augite-bearing bronzitite, EB 13

PHOTOMICROGRAPHS

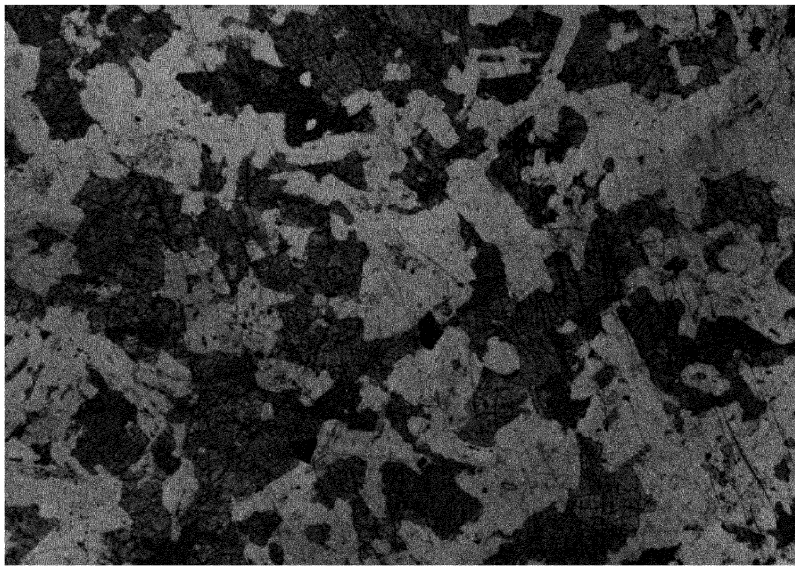


FIGURE 1

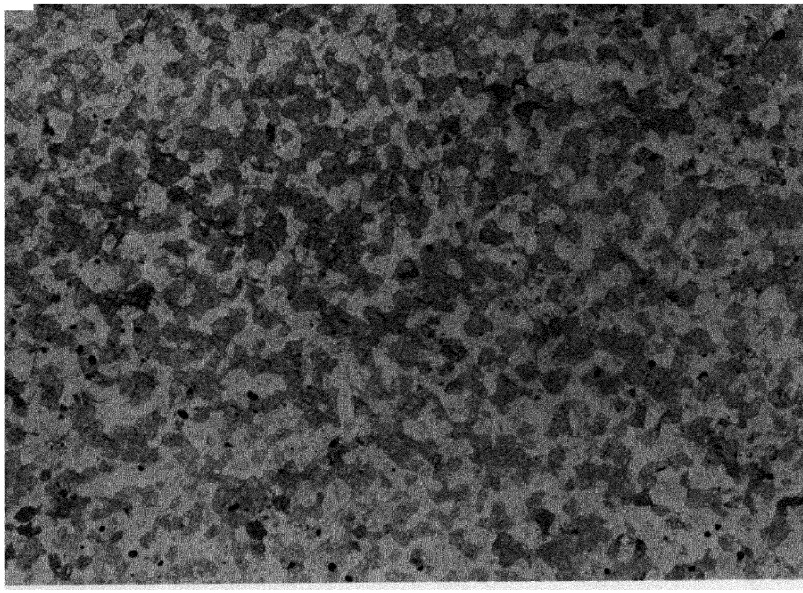


FIGURE 2

TEXTURAL RELATIONS IN BORDER FACIES (463 G 70) AND IN THE HORNFELS (?) BELOW IT (EB 89)

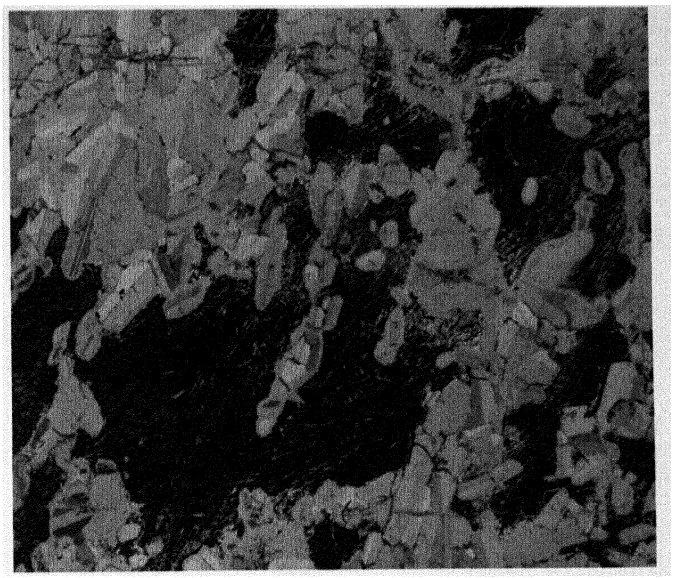


FIGURE 1. — Dark areas are poikilitic olivine with narrow rim of serpentine between olivine and plagioclase

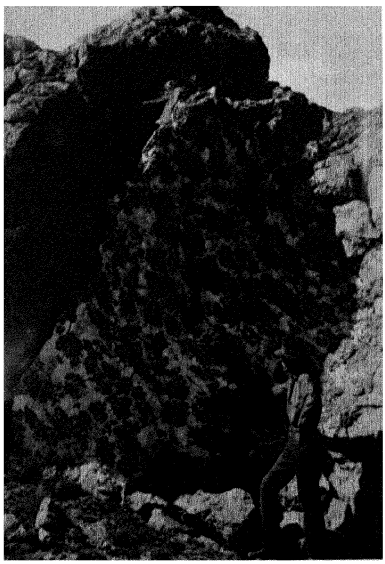


FIGURE 2. — "Pillow" troctolite layer

FIGURE 3. — Close-up view of same rock

OLIVINE-PLAGIOCLASE RELATIONS IN "PILLOW" TROCTOLITE, EB 70

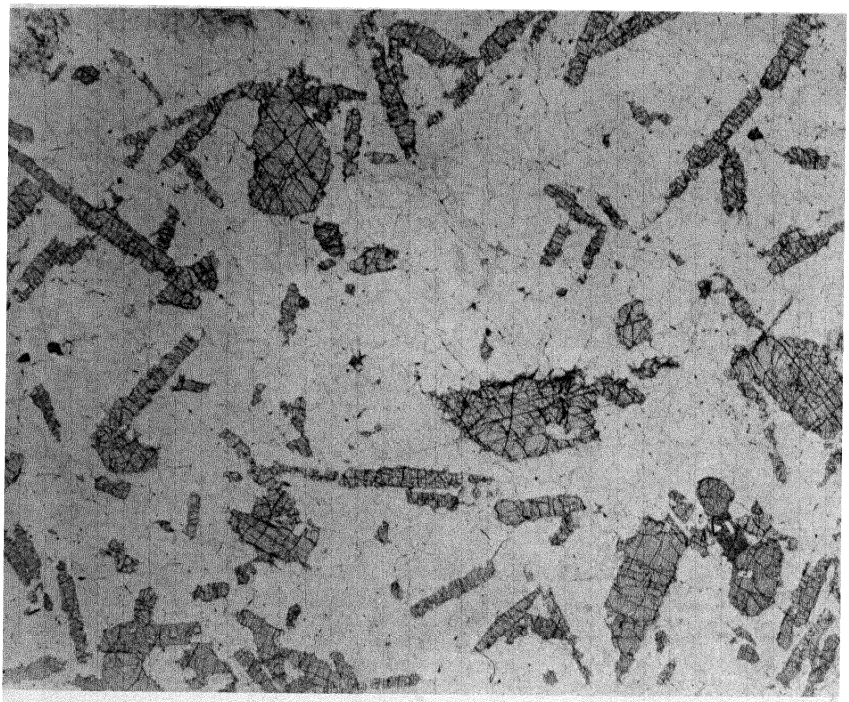


FIGURE 1. — Rock is cut parallel to lamination. Width of picture 1 inch

FIGURE 2. — Same rock cut at right angles to lamination; same scale

PHOTOMICROGRAPHS OF NORITE EB 38

the other pyroxene is also iron rich, and the plagioclase is more sodic than the plagioclase of nearby layers. All three minerals vary sympathetically in relative proportions of the less refractory constituents. The percentage change in the plagioclase is much smaller as a rule than in the pyroxene. These generalizations are of importance, the writer believes, in understanding the causes of mineral variations in the rocks and the process of crystallization and differentiation of the magma. This will be referred to again in Chapter VI.

F. ANORTHOSITE ZONE

1. GENERAL STATEMENT

The Anorthosite Zone is the thickest of the subdivisions of the complex described by the writer (6210 feet). It represents several difficult but fascinating petrological problems. It cannot be said that these are all solved satisfactorily at the present time. Tentative explanations for the observed relations are advanced with due recognition of the fact that additional investigations of the silicate chemistry as well as the petrological study of Stillwater and similar rocks in the field and laboratory will be necessary before final conclusions may safely be reached.

The Anorthosite Zone has three thick layers of anorthositic rocks separated by two layers containing normal amounts of ferromagnesian minerals (Fig. 19). Above the Ultramafic Zone, olivine is restricted to the Anorthosite Zone. The boundaries of the zone were arbitrarily placed below the thick anorthosite 1 and above the thick anorthosite 3.

2. TEXTURAL RELATIONS

The gabbros and norites of this zone show the same textural variations as do the rocks of the lower zones. The pure anorthosites are made up of peri-euhedral plagioclase grains. Approximately three quarters of the rock consists of accumulated crystals of plagioclase, and one quarter later plagioclase from the interstitial liquid. Most of the anorthosites contain some pyroxene, all of which crystallized from the interstitial liquid. Clinopyroxene, orthopyroxene, or both may be present in any specimen. The grain size of the plagioclase is noticeably larger in these anorthosites than in other feldspar-bearing rocks of the complex. The average grain intercept is 1–1.5 mm instead of .5–.75 mm commonly found in the other zones or in the norites, gabbros, and troctolites of this zone.

3. MINERALOGY

The essential minerals of the Anorthosite Zone are plagioclase, clinopyroxene, orthopyroxene, and olivine (Table 23).

Accessory minerals are rare in the Anorthosite Zone except for yellow sulfides which are present in small amounts in most specimens. A trace of ilmenite-magnetite is also associated with the sulfide in most samples. Quartz was noted in 5 of the 34 specimens.

A short distance below the top of the zone, three layers 1–2 feet apart contain a larger concentration of sulfides than any of the rocks of the East Boulder section. It is estimated that sulfides in these layers may reach 1–2 per cent of the rock by

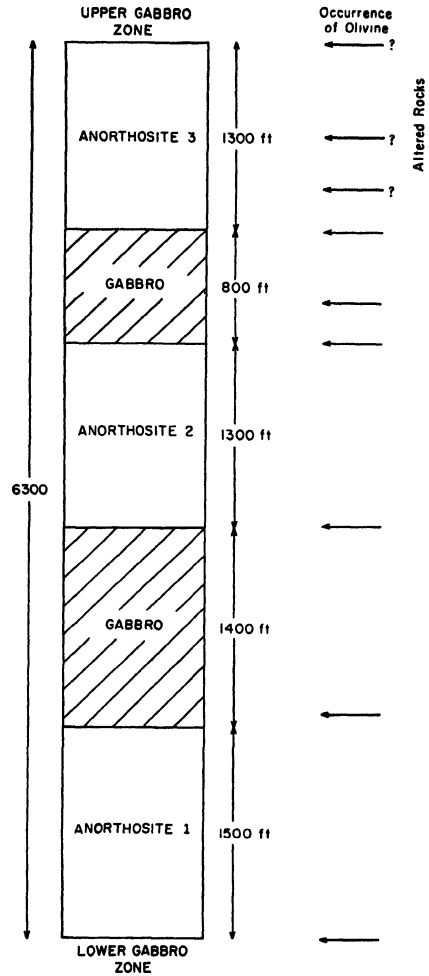


FIGURE 19.—*Columnar section of the Anorthosite Zone*

weight. These thin layers are probably equivalent stratigraphically to Howland's upper sulfide horizon in the Main Boulder Valley. They occur in gray anorthosite.

4. BULK COMPOSITION

The pyroxene ratio (100 orthopyroxene/orthopyroxene + clinopyroxene) is very low (23½). (*See* Table 23.)

5. DESCRIPTION OF SPECIMENS

The best developed troctolite of the complex is at the base of the Anorthosite Zone. This is the "Pillow troctolite" horizon (Pl. 5). A good example of this rock type is

EB 70, taken from 8000 feet above the base of the complex and about 500 yards west of the line of the main section, near Cairn Station. This consists of approximately one part olivine Fo_{80} to two parts plagioclase $An_{79\frac{1}{2}}$. The olivine crystals are generally 5–10 mm in diameter and include much plagioclase in their rims. Most of them have a plagioclase-free core suggesting settled crystals which have grown to a considerable extent from the interstitial liquid (Pl. 5, fig. 1). The olivine crystals are

TABLE 23.—*Composition of minerals in Anorthosite Zone*

	Range	Average
Plagioclase	An_{79}-$An_{75.5}$	$An_{77.5}$
Orthopyroxene	Mg_{79}-Mg_{60}	Mg_{75}
Clinopyroxene	$Ca_{41}Mg_{47}Fe_{12}$ $Ca_{40}Mg_{44}Fe_{16}$	$Ca_{41}Mg_{46}Fe_{13}$
Olivine*	Fo_{80}-Fo_{75}?	Fo_{77}?

* Many of the olivines are too altered for determination.

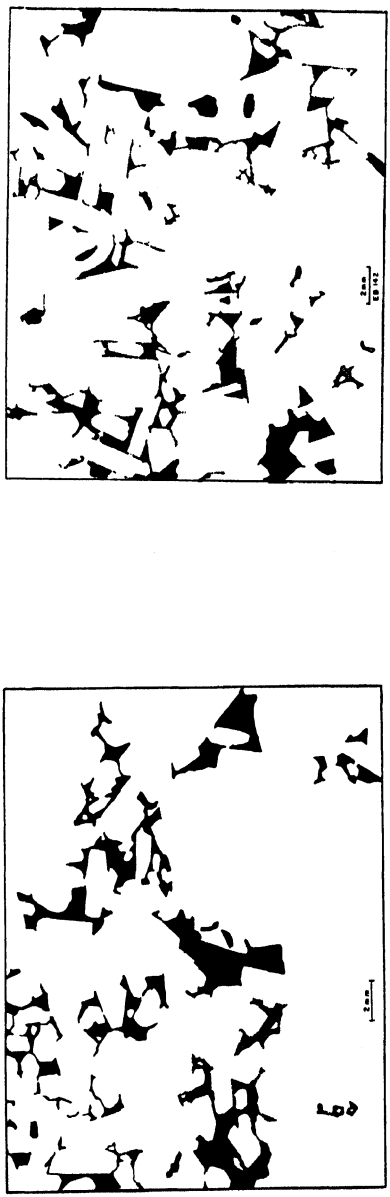
set in a matrix of much smaller (1 mm or less) peri-euhedral plagioclase crystals. This matrix is almost pure anorthosite, but a trace of both orthopyroxene and clinopyroxene may be found in it as skeleton poikilitic crystals. Rarely orthopyroxene narrowly rims the olivine. Normally there is a thin band of isotropic serpentine along the boundary between olivine and plagioclase. It appears to replace both olivine and plagioclase and is rather constant in width. The olivine crystals are fresh but are criss-crossed by thin veinlets containing magnetite.

Specimen EB 70 represents the "pillows" of the troctolite. The pillows are enclosed in a matrix of pure anorthosite made up of small peri-euhedral crystals of composition $An_{79.6}$.

The anorthosites of the three anorthosite subzones are all very similar. Pure anorthosites of coarse, interlocking texture are rather common. Many show incipient alteration to zoisite or clinozoisite. Others have large skeleton poikilitic pyroxenes. EB 142 found at 12,930 feet from the base of the complex is typical of these. It contains slightly more than 90 per cent plagioclase, $An_{77\frac{1}{2}}$, and skeleton poikilitic ortho- and clinopyroxene in about equal amounts. The orthopyroxene, Mg_{60}, is an inverted pigeonite with the usual coarse exsolution plates of augite. The clinopyroxene has an approximate composition $Ca_{39.5}Mg_{40}Fe_{20.5}$. A few minute ilmenite-magnetite grains are included in the pyroxene. The rock is fresh, with a trace of zoisitic alteration. A few plagioclase crystals are narrowly rimmed by antiperthite. The texture is illustrated in Figure 20b.

6. THE ANORTHOSITE PROBLEM

a. Facts.—Three subzones within the Anorthosite Zone are anorthositic in composition (85–100 per cent plagioclase). The remainder (0 to 15 per cent pyroxene) is all interstitial to the plagioclase and of late crystallization. The pyroxene forms large skeleton poikilitic crystals. Plagioclase was apparently the *only* mineral accumulating on the floor of the magma chamber during the building up of these layers that constitute about 4400 feet of anorthosite.



a b c

FIGURE 20

The plagioclase grains are somewhat larger in these anorthosites than in the normal Stillwater plagioclase-bearing rocks. The average grain intercept across the plagioclase crystals is two to three times greater.

Olivine is found near the upper and lower boundaries of these three thick anorthosite layers, but nowhere else above the Ultramafic Zone. Thus there is probably some direct relationship between the anorthosites and the crystallization of olivine.

Pyroxene-rich layers complementary to the anorthosite layers are not found, unless the Ultramafic Zone 5000 feet below be so considered.

The composition of the plagioclase in the anorthosites does not differ appreciably from that of the plagioclase in rocks adjacent to the anorthosites. There is a gradual decrease in An content of the plagioclase from the base of the zone to the top (An_{79} to $An_{75.5}$).

b. Hypothesis on origin of the thick anorthosites.—Evidence of crystal sorting in the Stillwater complex is abundant so that the most obvious hypothesis to explain the 4400 feet of anorthosite of the Anorthosite Zone is to consider it the complement by crystal sorting of the 2500 feet of pyroxene-rich and olivine-rich rocks of the Ultramafic Zone. This hypothesis must presuppose that, while the olivine and pyroxene were accumulating on the floor to form the Ultramafic Zone, plagioclase was either rising or remaining suspended in the magma.

Let us first inquire into the density relations. The measured density of the border-facies fine-grained norite at 20°C. is 3.02 g/cc. Bytownite An_{85} has a density of 2.74 g/cc. It is estimated that the magma temperature at the start of crystallization was 1125°C. Based on figures for the Vinal Haven diabase (Dane, *in* Birch *et al.*, 1942) and the Palisades diabase (Day, Sosman, and Hostetter, 1914) and interpolating for temperature, the density decrease from the crystalline state at 20°C. to the liquid state at 1125°C. should lie between 11 per cent and 13 per cent. This would give a density for the magma at 1125°C. of 2.65 $\pm$ 0.3 g/cc. The bytownite at 1125°C. would have a density of 2.69 g/cc. Thus the plagioclase would be 0.01 to 0.07 g/cc denser than the magma. With due regard for the various uncertainties and assumptions in these calculations, it is concluded that the plagioclase probably would not have floated upward, but might have settled or remained suspended if it were crystallizing at this time. If it remained suspended in the magma while the olivine and pyroxene settled, it would serve the needs of the above hypothesis. Thus the density data lead to no *positive* conclusion either favorable or unfavorable to the hypothesis. It is calculated that after separation of the olivine and pyroxene of the Ultramafic Zone amounting, it is estimated, to slightly more than 9 per cent of the volume of the complex, the magma density would be decreased by 0.08 g/cc. Plagioclase at this epoch in the history of the complex should settle if it were to crystallize, and there is clear evidence at the base of the Norite Zone that it did settle.

This hypothesis, when tested by petrographic detail, does not fare so well. One

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FIGURE 20.—*Textural relations in EB 156, EB 142, and EB 122*

a. EB 156. Black, orthopyroxene; dotted, clinopyroxene; both skeleton poikilitic with size of crystals larger than slide.

b. EB 142. Black, hypersthene inverted from pigeonite; dotted, interstitial clinopyroxene

c. EB 122. Black, peri-euhedral orthopyroxene; dotted, interstitial clinopyroxene

might expect to find some plagioclase crystals trapped in the mush of settled olivine or pyroxene crystals of the Ultramafic Zone if plagioclase had been separating at the same time, but no such peri-euhedral crystals are found. The contact of the Ultramafic Zone with the Norite Zone is sharp and abrupt. This condition is best accounted for by the postulate that it represents the sudden beginning of crystallization of plagioclase and its settling to the floor below with orthopyroxene. This early plagioclase of the Norite Zone has a composition of An_{86} which would be the plagioclase in equilibrium with the liquid An_{68}. This is the composition of the normative plagioclase of the border-facies analyzed rock, thought to be equivalent to the original magma. This strongly suggests that this bytownite was the first plagioclase to separate from the magma.

It seems unlikely that a hypothesis of mechanical sorting of crystals can by itself account for the thick layers of anorthosite in the Anorthosite Zone. It is significant to remember that pyroxene gives way to olivine as the thick anorthosites are approached from below, and finally the magma seems to pass into a field where no ferromagnesian mineral is crystallizing. Plagioclase alone is accumulating on the floor during the formation of the anorthosite layer. Passing out of the anorthosite at the top, olivine reappears, and then its place is taken by pyroxene, just the reverse of the sequecne found upon entering the layer from below. In a system governed by phase-rule equilibrium it would be quite possible to pass from the field of crystallization of pyroxene by reaction into a field of olivine crystallization, but it would be impossible to have the olivine phase then cease crystallizing without its place being taken by some new ferromagnesian phase. Perhaps we are dealing here with conditions which depart from equilibrium as envisioned by the general phase-rule requirements governing a closed system.

The observed phase relations could have been developed in the following manner. Some time after the magma was intruded it became thermally adjusted to the extent that the heat flow from the roof and floor reached a steady state. The flow would be predominantly from the roof upward. If the magma were emplaced with some superheat, it was first lost from the layer immediately beneath the roof. The outward heat flow from this layer is balanced by the heat of crystallization of the solid phases separating out, and crystallization proceeds with very little change of temperature. After this layer below the roof becomes sufficiently charged with crystals, the bulk density of the liquid–crystal mixture becomes high enough to result in convective overturn (as suggested by Grout in the Duluth gabbro and by Wager and Deer in the Skaergaard intrusion). The magma charged with crystals spreads out on the floor, and its crystalline phases slowly settle out of it. After plagioclase joins the minerals separating from such a layer, it also will settle out but at an extremely slow rate. We may suppose that all the pyroxene but not all the plagioclase has separated from such a layer before the next convective overturn arrives from the roof. The newly arrived relatively dense liquid–crystal mixture displaces the preceding layer now lighter because of loss of its pyroxene and much of its plagioclase. The displaced layer moves upward carrying part of its plagioclase with it. Upon reaching the central portion of the magma chamber it mixes with hotter undifferentiated magma still containing some of its original superheat. The suspended plagioclase crystals are resorbed, and the residual magma mixes with the undifferentiated magma. In this

manner a portion of the magma near the center of the magma chamber becomes relatively changed in composition and enriched in the constituents of plagioclase. Later, when crystallization starts from this "contaminated" portion of the magma, it is on the plagioclase side of the pyroxene–plagioclase cotectic line, and plagioclase alone separates until the composition moves back to the cotectic line where it will be joined

TABLE 24.—*Mineralogical composition of the Anorthosite Zone*
(Weight per cent)

	1 Zone as a whole	2 Three thick anorthosite layers	3 Norites and gabbros between the three anorthosites
Plagioclase	84 ($An_{77.5}$)	92	63
Orthopyroxene	3¾ (Mg_{75})	2	7½
Clinopyroxene	12¼ ($Ca_{41}Mg_{46}Fe_{13}$)	6	29½

by pyroxene (relations simplified for purposes of discussion; actually one or two pyroxene phases and olivine are involved). If the composition of the contaminated magma were appropriate, olivine might appear before pyroxene, and the troctolites found on the margins of the thick anorthosite layers might be so derived. Successive movements of masses of the plagioclase-contaminated magma to positions under the roof where they would lose heat, crystallize, and then move down to the floor could thus explain the successive thick anorthosites. By the time the upper part of the Anorthosite Zone had crystallized, no more magma enriched in plagioclase was left, and the production of such magma had ceased at some previous time when the last remnant of superheat for solution of the plagioclase had been exhausted.

Further information may be obtained from a consideration of the pyroxene: plagioclase ratios in the rocks and deduced contemporaneous liquids (Table 25).

TABLE 25.—*Pyroxene: plagioclase ratios in various units of the complex*

	$\frac{100\ Px}{Px + Pc}$ in wgt. %
1. Norite Zone plus Lower Gabbro Zone	38
2. Gabbros and norites intercalated with the anorthosites of the Anorthosite Zone	37
3. Upper Gabbro Zone	35
4. From composite analysis 2 including the above 1, 2, and 3 plus the anorthosites of the Anorthosite Zone	31.2
5. Ratio in original magma as indicated by border facies analysis	36.5
6. Above value less the material of the Ultramafic Zone (assumed to be 9.3 per cent of volume of complex)	31.3

The actual pyroxene:plagioclase ratio as obtained from the composite analysis of the rocks from the base of the Norite Zone to the highest exposure in the complex (4) is practically identical with the deduced ratio in the magma at the start of crystallization of plagioclase (6). The norites and gabbros below the Anorthosite Zone

should contain about 20 per cent more plagioclase than they would if pyroxene and plagioclase were separating at this time in proportion to their ratios in the magma. This 20 per cent extra plagioclase may represent the crystals carried upward and resorbed. The hypothesis would suggest that plagioclase separated from the contaminated magma in excess of the amount resorbed by the magma before the stage was reached at which it was again joined by pyroxene.

While the writer favors the above hypothesis of re-solution of plagioclase to account for the apparent cessation of crystallization of the ferromagnesian minerals from the magma, it has the obvious weakness of any hypothesis which assumes an appropriate amount and kind of contamination to produce the desired results.

UPPER GABBRO ZONE

1. GENERAL STATEMENT

The base of the Upper Gabbro Zone is approximately at the edge of the plateau. The rocks of this zone crop out on the northward slope from the plateau toward the east-west ridges of Paleozoic sedimentary rocks. Between the slope and the Paleozoic rocks is a saddle about a quarter of a mile wide which is thickly wooded and in which no outcrops of the complex are found. It would be interesting to trench or drill in this saddle and thereby obtain information on the upper, now hidden, portion of the complex. Streams crossing the Palezoic ridges have cut well into the Cambrian so that it would require relatively little drilling here to get to the Precambrian below. Only in this area on the northern rim of the East Boulder Plateau are rocks of the Upper Gabbro Zone found. Elsewhere Paleozoic rocks cover them.

In many respects it is the most satisfactory zone to describe. The relations are clear cut, and the mineral and textural relations are more straightforward than in any of the other zones.

Rhythmic layering is faint or absent in the rocks of the Upper Gabbro Zone although many of the rocks have a pronounced lamination. The conspicuousness of lamination may be a result of the more tabular form of the plagioclase crystals so that there is a greater tendency for them to lie with their longer axes in the plane of lamination. With short stubby plagioclase this tendency is not so marked.

The total thickness of this zone is 2130 feet, extending from 14,060 feet to the highest exposure of the complex at 16,190 feet.

2. TEXTURAL RELATIONS

In all the rocks of this zone examined the plagioclase is one of the accumulated minerals and has crystallized predominantly early with only a small contribution to this mineral from the interstitial liquid. Most of the crystals are peri-euhedral.

The orthopyroxene and clinopyroxene show all possible textural relations to the essential minerals present in any rock specimen. Either or both may be late or early. Either may therefore occur as peri-euhedral crystals, as completely late poikilitic crystals, or they may show some intermediate relationship. The textural relations will be further considered during the discussion of individual specimens and in connection with the mineral-variation curves.

3. MINERALOGY

The essential minerals are plagioclase, augite, and two varieties of hypersthene. One orthopyroxene is orthopyroxene of the Bushveld type with fine lamellae parallel to {100}; the other is orthopyroxene of the Stillwater type with coarse augite plates parallel to the monoclinic {001} plane. Stillwater-type orthopyroxene is commonly found only in the uppermost 800 feet of the exposed complex where it takes the place

TABLE 26.—*Bulk mineralogical composition and average composition of minerals in Upper Gabbro Zone*

	Weight per cent	Average composition of minerals in molecular per cent
Plagioclase	65	An_{69}
Orthopyroxene	17	$Mg_{71.5}$
Clinopyroxene	18	$Ca_{40}Mg_{46}Fe_{14}$
	100.0	

of the Bushveld type. Where the orthopyroxene is richer in iron than Mg_{70} the coarse plates are present in it. Chemically the bulk composition of this orthopyroxene shows about 9½ per cent of Ca atoms in the total Ca + Mg + Fe or nearly three times as much as in Bushveld-type orthopyroxene. It is considered that the Stillwater type represents an inverted pigeonite and the Bushveld type indicates direct primary crystallization of orthopyroxene from the magma as discussed in the chapter on Mineralogy.

The three essential minerals show the following ranges in composition:

Plagioclase	An_{75} to An_{61}
Orthopyroxene	
I. With fine lamellae only	Mg_{78} to Mg_{70}
II. With fine lamellae and coarse plates	Mg_{70} to Mg_{64}
Augite	$Ca_{41}Mg_{48}Fe_{11}$ to $Ca_{39}Mg_{43}Fe_{18}$

Accessory minerals seem to be even less abundant in the rocks of this zone than lower in the complex. Sulfides are the most common and are present in all but 2 of 19 specimens examined in detail. Quartz was found only in EB 122, a specimen from only a few feet above the base of the zone. Some isotropic pink specks in this quartz have a very low index of refraction and were considered to be fluorite; this is the only occurrence of this mineral noted in the complex. In the uppermost 6 specimens the ends of some of the plagioclase laths are replaced by a rim of dactylitic intergrowth of what appears to be a colorless mica (probably muscovite) and a more sodic plagioclase.

4. BULK COMPOSITION

The rocks of the Upper Gabbro Zone are augite norites or hypersthene gabbros. Anorthosites and also pyroxenites are lacking. There is thus a much greater uni-

a b c

FIGURE 21

formity in rock types than in any of the other zones and in this sense a much less pronounced local differentiation. The minerals present in the rocks, however, show a much more rapid change in composition with distance from the base than is found in the other zones so that in this sense the differentiation is more rapid.

The bulk composition of the zone is given in Table 26.

The pyroxene ratio ranges from 72½ to 20 and averages 51 or practically 1:1 orthopyroxene to clinopyroxene. Omitting three extreme specimens the range becomes 62½ to 40.

5. DESCRIPTION OF SPECIMENS

EB 122 occurs at 14,080 feet or 20 feet from the base of the zone. It has the following composition:

	Weight per cent	*Composition in molecular per cent*
Plagioclase	62½	An_{75}
Orthopyroxene	26	Mg_{73}
Clinopyroxene	11½	$Ca_{41}Mg_{47}Fe_{12}$

It also contains a minute amount of yellow sulfide, ilmenite-magnetite intergrowth, and quartz which occurs as triangular areas interstitial to plagioclase laths. In the quartz are small specks of pink isotropic material with a very low index of refraction, probably fluorite. This is the only observed occurrence of fluorite in the complex. It is also the only occurrence of quartz in the Upper Gabbro Zone.

Figure 20c shows the distribution of minerals and their textural relations. The orthopyroxene and plagioclase are evidently accumulated (early) crystals which have been added to slightly from the interstitial liquid. Relatively more has been added to the orthopyroxene than to the plagioclase. The clinopyroxene crystallized late from the interstitial liquid, and it includes euhedral or almost euhedral plagioclase and orthopyroxene crystals.

The plagioclase in the thin section is much elongated, and it and the orthopyroxene show a strong tendency to alignment. The thin section was cut approximately at right angles to the plane of layering so that the alignment indicates the lamination. In a plane parallel to the layering no alignment of the long axes of crystals is noted.

EB 43 outcrops at approximately 400 feet from the base of the zone, at the 14,460-foot horizon. It is a hypersthene gabbro. Table 27 gives the chemical composition of the rock and the chemical analyses of the three essential minerals—plagioclase, augite, and hypersthene.

Figure 21a shows the textural relations of the rock. The plagioclase is predominantly early but shows some late growth. The two pyroxenes appear to be about half early accumulated crystals and half later growth from the interstitial liquid. The orthopyroxene continued to crystallize for a short time after the clinopyroxene ceased so that it occasionally rims the clinopyroxene crystals and it includes a few small clinopyroxene crystals near the margins of its grains. A few grains of yellow sulfides are present in the thin section of the rock.

FIGURE 21.—*Textural relations in* (*a*) *EB 43,* (*b*) *EB 42, and* (*c*) *EB 40*
Black, orthopyroxene; dotted, clinopyroxene

TABLE 27.—*Chemical composition of rock EB 43 and its constituent minerals*

	Rock	Plagioclase	Orthopyroxene	Clinopyroxene
SiO_2	50.12	49.19	53.20	51.83
Al_2O_3	20.01	31.82	2.24	3.07
Fe_2O_3	.80	.55	.83	1.38
FeO	4.29	—	15.15	7.21
MgO	7.91	.41	24.58	16.00
CaO	13.97	14.95	2.61	19.21
Na_2O	1.74	2.78	.06	.27
K_2O	.05	.09	.01	.02
H_2O+	.69	.28	.48	.47
H_2O-	.04	—	.10	.11
Cr_2O_3	—	—	.10	—
TiO_2	.19	—	.26	.49
P_2O_5	trace	—	trace	—
MnO	.09	—	.32	.17
	99.90	100.07	99.94	100.23
Analyst	Ellestad	Phillips	Ellestad	Ellestad
Norm				
Pc (Or .28, Ab 14.67, An 46.56)	61.5†			
Hy (En 13.38, Fs 4.91)	18.29	$An_{74.4}Ab_{25.0}Or_{.6}$	$Mg_{73}Fe_{27}$	$Ca_{41}Mg_{45}Fe_{14}$
Di (En 6.52, Fs 2.15, Wo 9.45)	18.12		(a little augite and plagioclase impurity)	
Il	.38			
Mt	1.16			
H_2O	.73			
	100.19			

* Alkalies checked by Ellestad

† Calculated Pc, molecular per cent, $An_{74.6}Ab_{25.0}Or_4$. The normative plagioclase is slightly greater than the modal plagioclase since there is a little Al_2O_3 in the pyroxene. There is also slightly more clinopyroxene relative to orthopyroxene in the mode because of the relatively greater amount of Hy dissolved in the natural clinopyroxene than Di dissolved in the natural orthopyroxene.

EB 42 was collected approximately 1210 feet from the base of the zone at the 15,270-foot horizon. It is a hypersthene gabbro. Chemical analyses of the rock and its plagioclase are given in Table 28; an estimate of the composition of the two pyroxenes based on optical properties is also given.

TABLE 28.—*Chemical composition of rock EB 42 and its plagioclase*

	Rock	Plagioclase
SiO_2	51.11	51.34
Al_2O_3	15.36	29.98
Fe_2O_3	1.22	.55
FeO	6.63	—
MgO	10.04	.48
CaO	12.63	13.39
Na_2O	1.82	3.70
K_2O	.07	.15
H_2O+	.90	.35
H_2O-	.04	.02
TiO_2	.22	—
P_2O_5	trace	—
MnO	.08	—
	100.12	99.96
Analyst*	Phillips	Phillips

Composition of Minerals in Molecular Per Cent

Plagioclase, calculated from mineral analysis $An_{66.1}Ab_{33.0}Or_{.9}$

Clinopyroxene, calculated from optical properties $Ca_{40}Mg_{44}Fe_{16}$

Orthopyroxene, calculated from optical properties $Ca_{3\frac{1}{2}}Mg_{67\frac{1}{2}}Fe_{29}$ or $Mg_{70}Fe_{30}$

Norm		
Pc	An 33.64, Ab 15.46, Or .39	49.49†
Di	Wo 12.18, En 7.90, Fs 3.43	23.51
Hy	En 15.90, Fs 6.79	22.69
Ol	Fo .91, Fa .41	1.32
Il		.46
Mt		1.74
H_2O		.94
		100.15

* Alkalies checked by Ellestad

† Plagioclase calculated from rock analysis, $An_{66.7}$

The small amount of olivine indicated in the norm is probably the result of slight alteration of the rock forming some zoisite from plagioclase. No olivine is present in the rock. The water content also is a function of this alteration. In the case of the analyzed plagioclase the slightly altered material was removed by magnetic separation before analysis.

TABLE 29.—*Chemical composition of the rock EB 40 and the plagioclase and pyroxene of EB 41*

	Rock EB 40	Plagioclase EB 41	Orthopyroxene EB 41	Clinopyroxene EB 41
SiO_2	51.12	52.95	52.81	51.86
Al_2O_3	16.34	29.48	1.65	2.33
Fe_2O_3	1.09	.24	1.13	1.60
FeO	5.84	.32	18.13	9.45
MgO	8.49	.10	19.81	14.50
CaO	13.63	12.70	5.13	18.92
Na_2O	2.18	4.12	.05	.23
K_2O	.06	.13	.00	.00
H_2O+	.49	.10	.54	.37
H_2O-	.07	.00	.16	.09
Cr_2O_3	—	—	—	.01
TiO_2	.19	.05	—	.55
P_2O_5	.03	—	—	—
MnO	.15	—	.38	.24
NiO	—	—	—	.02
	99.68	100.19	100.16	100.17
Analyst	Phillips*	Oslund	Ellestad	Ellestad

Norm

Pc	An	34.57	53.24†
	Ab	18.34	
	Or	.33	
Di	Wo	13.92	26.91
	En	8.90	
	Fs	4.09	
Hy	En	9.80	14.35
	Fs	4.55	
Ol	Fo	.92	2.67
	Fa	1.75	
Mt			1.62
Il			.38
H_2O			.56
			99.73

Mode weight per cent

Pc	57
C-Px	26¾
O-Px	16¼
	100

Clinopyroxene $Ca_{39.5}Mg_{42.5}Fe_{18}$ (analysis)

Orthopyroxene $Ca_{10}Mg_{55}Fe_{32}$ or $Mg_{64.5}Fe_{35.5}$ (less small amount of augite impurity)

* Alkalies checked by Ellestad

† Plagioclase calculated from rock analysis $An_{63\frac{1}{2}}$ and An_{63} from mineral analysis EB 41

Figure 21b shows the textural relations. The plagioclase is predominantly early, peri-euhedral, accumulated crystals. Most of the clinopyroxene is also early but shows less tendency toward euhedral crystals, partly because of a slight replacement by younger orthopyroxene and partly because of relatively more material added from the interstitial liquid. Orthopyroxene is entirely late; it crystallized from the interstitial liquid. It encloses much clinopyroxene and plagioclase in poikilitic crystals.

As in specimen EB 122 (Fig. 20c) the plagioclase crystals are strongly aligned. The alignment represents the trace of the plane of lamination. The outcrops from which EB 42 was collected show a well-developed lamination but no rhythmic layering.

EB 40 was collected at the 16,190-foot horizon, the uppermost exposure of the complex. It is a hypersthene gabbro. Table 29 gives the chemical composition of the rock and of two pyroxenes from EB 41 collected a few feet below EB 40.

The olivine content of the norm is, as in EB 42, probably the result of slight alteration. No olivine is present in the rock.

The orthopyroxene is of the Stillwater type, with the coarse oriented plates of clinopyroxene. It is interpreted as an inversion product of a pigeonite.

Figure 21c shows the textural relations of the rock. All three of the essential minerals—plagioclase orthopyroxene, and clinopyroxene—appear to be accumulated crystals and predominantly of early crystallization. All three have had a considerable addition from the interstitial liquid (perhaps 25 per cent). There is a well-developed alignment of the elongate grains indicating the lamination of the rock. The pyroxene crystals tend to occur in clusters. The plagioclase crystals are quite variable in size. The three large crystals of plagioclase near the center of the figure are the largest crystals of plagioclase found in this specimen. Their margins and those of some of the small crystals are partly replaced by a dactylitic intergrowth of mica and more sodic plagioclase. Several small grains of yellow sulfide may be noted in the rock.

The four specimens described above illustrate the rock types found in the Upper Gabbro Zone as well as the variations observed in the minerals making up these rocks. EB 122 is from the base of the zone, EB 43 not far above the base, EB 42 approximately the middle, and EB 40 the top.

Chapter V

Bulk Mineralogical and Chemical Composition of the Stillwater Complex and its Subdivisions

A. DISTRIBUTION OF ROCK TYPES

More than 100 specimens of the Stillwater complex were examined quantitatively. All are composed essentially of the three minerals plagioclase, orthopyroxene, and clinopyroxene. (The olivine-bearing rocks of the Ultramafic Zone and the rare troctolitic rocks of the Anorthosite Zone are omitted from this discussion). These rock types can be plotted on a triangular diagram with the three essential minerals at the corners. The resulting diagrams show that certain portions of the field within the triangle are devoid of points representing rock types and that each zone has a well-defined field including all the rocks of that Zone with the exception of perhaps two or three abnormal specimens, and these lie but a short distance outside the field.

Figure 22 represents all the rocks examined. In the lower left-hand (orthopyroxene) corner a small field is separated quite distinctly from the bulk of the rock types on the diagram. This field contains entirely rocks of the Ultramafic Zone with one exception—the point representing the feldspathic pyroxenite EB 199 of the Norite Zone. Most of the other rocks examined lie within a field extending from 100 per cent plagioclase to 33 per cent plagioclase. Only six specimens, which may be considered abnormal, lie outside these two fields. The most interesting point is that there are no clinopyroxene-rich rocks; no specimen contains more than 50 per cent clinopyroxene, and only four contain more than 40 per cent.

Considering only the rocks above the Ultramafic Zone, eight contain less than 33 per cent of plagioclase, and three of the eight come from different points along the strike on the same horizon (EB 175, EB 23A, MB 223). The greatest concentration of points (not the average) seems to lie at about 60 per cent of plagioclase and 20 per cent of each of the two pyroxenes. The composition of the magma (border facies) lies about 10 per cent to the right of this; in other words it has relatively more orthopyroxene and less clinopyroxene.

The explanation of the distribution is fairly simple. Clinopyroxene does not accumulate as settled crystals below the Lower Gabbro Zone. From there on it is accompanied by orthopyroxene in a ratio not far from 1:1. This makes it impossible for a pure concentrate of clinopyroxene to be developed since the two pyroxenes have almost exactly the same density and cannot be separated by differential settling. Furthermore, above the horizon representing the top of the Ultramafic Zone, plagioclase is also accumulating as settled crystals. The plagioclase thus tends to dilute the concentration of clinopyroxene. Although the plagioclase may be, and is, separated from the pyroxenes by differential settling to a certain extent, the separation is not perfect, and therefore pyroxenitic horizons above the Ultramafic Zone are very rare—there are two in about 13,000 feet of layered rocks. Differential settling does seem to be more effective in producing nearly pure plagioclase rocks in the Norite and Anorthosite zones.

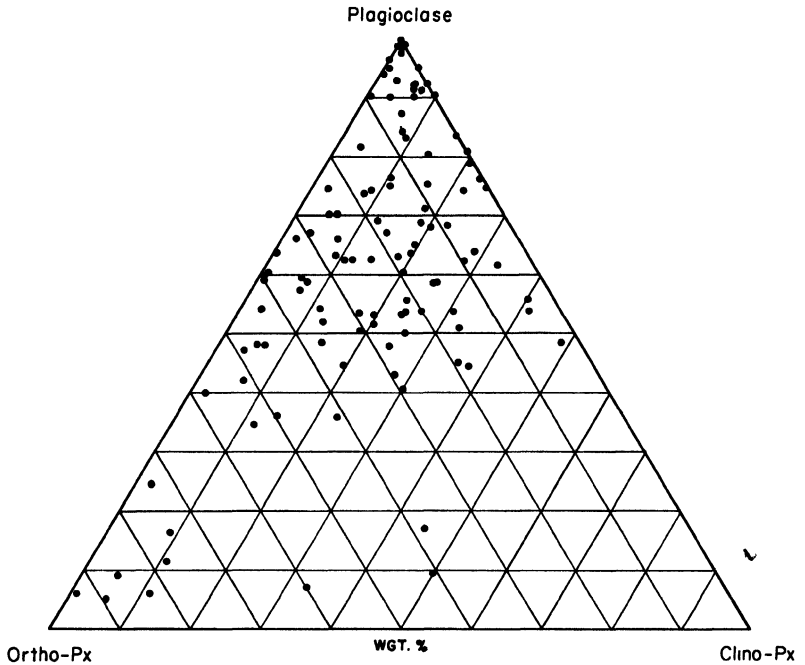


FIGURE 22.—*Relative percentages of plagioclase, orthopyroxene, and clinopyroxene in rocks of the Stillwater complex*

TABLE 30.—*Mineral proportions and compositions in each of the four upper zones of exposed complex*

	Weight per cent			Molecular per cent				
	Pc	O-Px	C-Px	Pc An	O-Px Mg	C-Px		
						Ca	Mg	Fe
Upper Gabbro Zone	65	17	18	69	71½	41	46	14
Anorthosite Zone	84	3¾	12¼	77½	75	41	47	13
Lower Gabbro Zone	60½	17	22½	79	77	41	48	11
Norite Zone	63½	28	8½	81	78	41	48	11
Average based on composite analysis (2)	69	13	18	76	77*	41	47	12

* Normative pyroxene slightly more magnesian than modal inasmuch as modal contains a little ferric iron.

The variability in the ratio of orthopyroxene to clinopyroxene in the three upper zones seems to be due largely to the contribution of material from the interstitial liquid. If only the early accumulated crystals could be examined then the ratio would probably be close to 1:1 in all cases. Where the percentage of pyroxene in the rock is

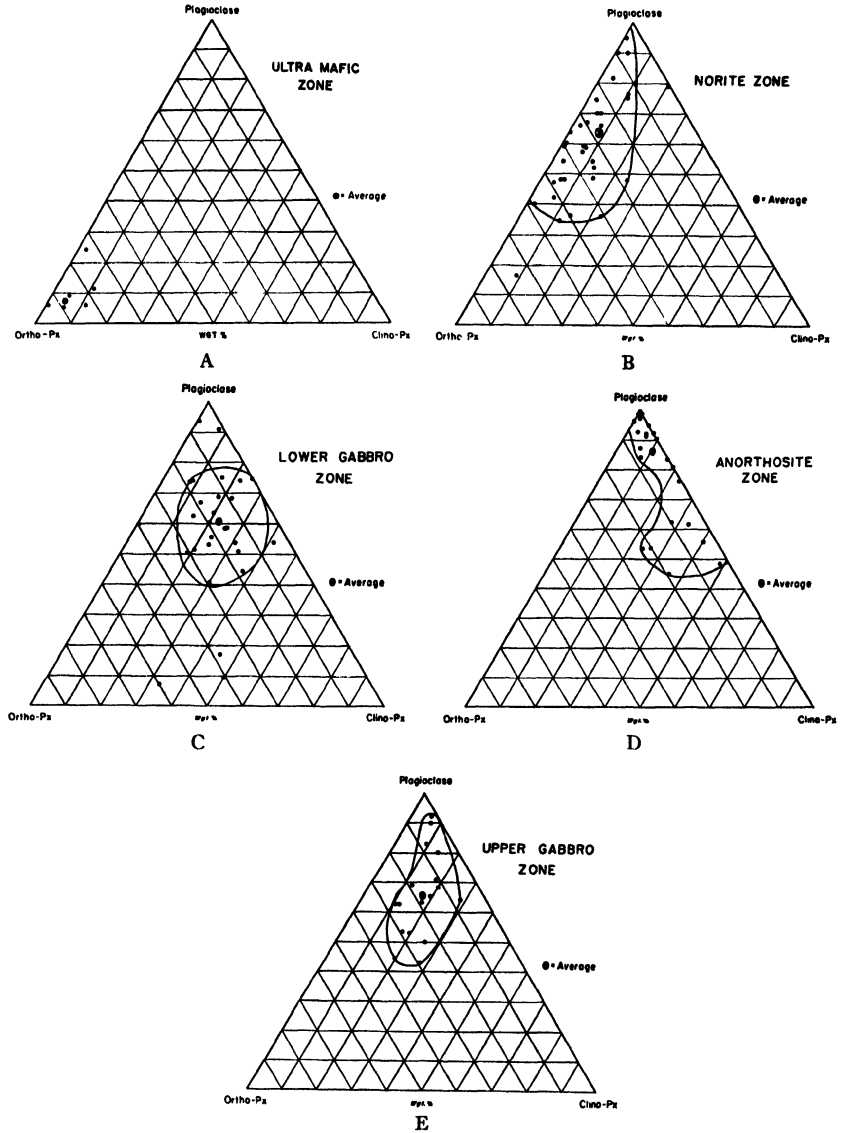


FIGURE 23.—*Plagioclase, orthopyroxene, and clinopyroxene percentages in each zone of the complex*

high the ratio is commonly close to 1:1 because the contribution of pyroxene from the interstitial liquid in these cases is relatively small. The contribution of pyroxene from the interstitial liquid may consist entirely of one pyroxene or the other, or more commonly it may be a mixture of both in no fixed ratio.

Figures 23 A to E show the distribution of rock types within each of the five zones again excluding the olivine-bearing rocks. The point representing the average composition of each zone is shown in the figures as an X in a circle. Figure 24F shows only the location of these average points.

TABLE 31.—*Chemical analysis, Composite 1, Ultramafic Zone*
Analyst, R. B. Ellestad

		Norm		
SiO_2	47.73	Cor	.66	
Al_2O_3	4.82	Pc An 10.42 / Ab 1.62 / Or .11	12.15	(An_{86})
Fe_2O_3	2.94			
FeO	6.54			
MgO	28.98			
CaO	2.44	Hy En 55.40 / Fs 7.00	62.40	
Na_2O	.19			
K_2O	.02			
H_2O+	4.91	Ol Fo 11.93 / Fa 1.78	13.71	
H_2O-	.49			
		Mt	4.18	
CO_2	.11	Il	.23	
TiO_2	.12	Cm	.70	
P_2O_5	.01	Py	.07	
		Cc	.31	
		H_2O	5.40	
S	.04			
Cr_2O_3	.48		99.81	
MnO	.17			
	99.99			

B. BULK CHEMICAL COMPOSITION

1. LOWER EXPOSED PORTION

An analysis was made of a composite sample of the Ultramafic Zone (Table 31). It consists of chips of 26 specimens collected by Peoples from his Little Rocky–Nye Basin Divide section. The plagioclase of the norm is about 5 per cent richer in An than in the mode inasmuch as some of the alumina is contained in the serpentine and the pyroxene. About half the pyroxene plus most of the olivine should be combined with the 5.40 per cent of water to form serpentine if the norm were to be modified to agree with the mode.

Chemical analysis of a second composite sample made up of 40 specimens of the East Boulder Plateau section (weighted with respect to the thickness of the section which each sample represents) is given in Table 32. In this case the norm and mode probably agree very closely.

By combining composite 1 and composite 2 in the correct proportions (the relative thicknesses of the Ultramafic Zone and the remainder of the exposed complex), the

analysis in Table 33 is obtained; it represents fairly accurately the composition of the whole exposed complex.

2. UPPER HIDDEN ZONE

The main objectives in determining the composition of the exposed complex are to estimate the amount and composition of the hidden remainder of the complex.

TABLE 32.—*Chemical analysis, Composite 2, exposed complex above Ultramafic Zone*
Analyst, R. B. Ellestad

		Norm			
SiO_2	49.70	Pc	An 51.78	67.31	($An_{76.9}$)
Al_2O_3	22.04		Ab 15.14		
Fe_2O_3	.66		Or .39		
FeO	4.02	Di	Wo 6.38	12.30	
MgO	7.03		En 4.20		
CaO	13.59		Fs 1.72		
Na_2O	1.79	Hy	En 12.60	17.35	($En_{77.8}$)
K_2O	.07		Fs 4.75		
H_2O+	.82				
H_2O-	.09	Ol	Fo .56	.76	
			Fa .20		
CO_2	.07	Mt		.93	
TiO_2	.16	Il		.30	
P_2O_5	.02	Ap		.05	
S	.04	Cc		.16	
Cr_2O_3	.03	H_2O		.91	
MnO	.09	Miscellaneous		.07	
	100.22			100.14	

Knowing the composition of the original liquid as preserved in the fine-grained doleritic border facies and the composition of the exposed complex, by subtracting the correct amount of the latter from the former, we can obtain the composition of the remainder. The unknown quantity in this equation is the ratio between the exposed and hidden portions of the complex.

Qualitatively some idea of the nature of the residue may be obtained by comparing the major constituents of the norms of the border-facies analysis and the composite analysis 1 + 2 (Table 34).

It is evident that the total amount of feldspar by weight in the hidden remainder must be lower than in the border-facies rock. But since a little olivine is present in the composite and none in the border facies, quartz should be present in the remainder. It would appear that feldspar + quartz, or the salic constituents, will be about the same in the remainder as in the border facies. The diopside content of both the border facies and composite is practically the same so there should be little change in the

TABLE 33.—*Bulk composition of exposed complex, Composite 1 + 2*

		Norm		
SiO_2	49.33			
Al_2O_3	19.28	Pc	An 45.59 Ab 12.94 Or .33	58.86 ($An_{76.5}$)
Fe_2O_3	1.02			
FeO	4.41			
MgO	10.48	Di	Wo 5.22 En 3.70 Fs 1.06	9.98
CaO	11.80			
Na_2O	1.53	Hy	En 17.40 Fs 4.49	21.89 (En_{83})
K_2O	.06			
H_2O+	1.45			
H_2O-	.15	Ol	Fo 4.27 Fa 1.12	5.39
CO_2	.08	Mt		1.49
TiO_2	.15	Il		.30
P_2O_5	.02	Cm		.21
S	.04	Cc		.20
Cr_2O_3	.10	H_2O		1.60
MnO	.10	Miscellaneous		.06
	100.00			99.98

TABLE 34.—*Comparison of norms of Border Facies and Composite 1 + 2*

Border facies			Composite 1 + 2	
Pc	An 39.06 Ab 15.72 Or 1.39	56.17 ($An_{68.4}$)	45.59 12.94 .33	58.86 ($An_{76.5}$)
Di	Wo 5.04 En 2.60 Fs 2.31	9.95	5.22 3.70 1.06	9.98
Hy	En 16.70 Fs 15.12	31.82 ($En_{59.4}$)	17.40 4.49	21.89 (En_{83})
Ol	Fo — Fa —		4.27 1.12	5.39

remainder. There is more hypersthene in the border facies than in the composite analysis. If, however, the olivine of the composite is converted to hypersthene and added to the original hypersthene then the composite contains almost as much as the border facies. Some uncertainty must exist as to the hypersthene content of the

remainder since an undetermined amount of the FeO probably has become oxidized, thus reducing the amount of FeO available for hypersthene. Except for this effect, it seems likely that the hypersthene content of the remainder would be fairly similar to that of the border facies.

Considering the individual minerals of the norms the following relations are evident:

Plagioclase

The An content in the remainder has decreased considerably, or in other words AB + Or has increased, with the Or increasing relatively more rapidly than the Ab, but still remaining very much subordinate to Ab.

Hypersthene and diopside

Iron has greatly increased with respect to magnesium in the femic constituents of the remainder.

The net result of all these observations may be summed up as follows:

The remainder is not a felsic rock and would not correspond to any normal rock type between ordinary quartz diorite and granite.

It would have a salic:femic ratio close to that of the original magma.

Specifically it must be a quartz-bearing ferro-norite. It seems most likely that the rocks of the hidden portion of the Stillwater are very similar to the rocks of the exposed portion of the Skaergaard intrusion minus olivine. As in the case of the Skaergaard only a very little granitic material should have been formed as a final differentiation product.

Quantitatively it is possible to define the limits within which the composition of the hidden portion of the complex must lie and to make a fairly accurate estimate of its exact composition.

The major oxide that shows the greatest percentage difference between the original magma (border facies) and composite 1 + 2 is MgO. The composite sample is much richer in MgO than is the border facies; therefore the remainder or hidden portion must contain very much less MgO than the original magma. A *minimum* value may be set on the amount of material in the hidden portion by assuming that all the MgO was used up by the crystallization of the exposed complex and none was left to go into the hidden portion. Assuming that no MgO is left, a simple graphical solution indicates that the hidden portion would amount to 27.7 per cent of the whole complex (or the exposed portion equals 72.3 per cent of the complex). This would mean that the hidden portion is at least 6200 feet thick, and the complex at least 22,390 feet thick.

The assumption of no MgO in the remainder, although useful in arriving at a minimum figure, is certainly far from correct. In the uppermost rocks of the exposed ocmplex the MgO:FeO molecular ratio is still about 68:32 in the normal pyroxenes, indicating a considerable amount of MgO in the magma when these rocks crystallized. In the experimental data on the pure system clinoenstatite–ferrosilite (Bowen and Schairer, 1935) a clinohypersthene with an MgO:FeO ratio of 68:32 would be in equilibrium with a liquid having the ratio 40:60. It is possible that in a natural complex silicate melt, such as the Stillwater magma, the MgO:FeO ratio of the crys-

tals as compared to the liquid will depart somewhat from the ratios of the liquidus and solidus of the pure melt because of the effects of other components in the magma.

A maximum value for the percentage of the whole complex represented by the

TABLE 35.—*Estimated thickness of Hidden Zone*

	Most unlikely (minimum)	Range of likely figures			Most unlikely (maximum)
			Best figure		
Per cent of complex represented by hidden portion.	27.7	35	40	45	50
Thickness of hidden portion (feet)....	6200	8710	10,640	13,240	16,190
Thickness of whole complex (feet).....	22,390	24,900	26,830	29,430	32,380

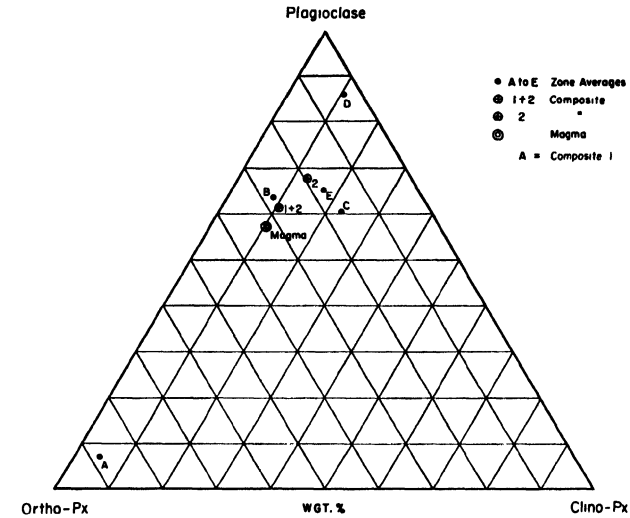


FIGURE 24.—*Orthopyroxene, clinopyroxene, and plagioclase proportions of 1, 2, and 3 of Table 36 compared with magma and average proportions of each zone*

hidden portion may be arrived at by a consideration of the plagioclase. If approximately 50 per cent of the complex were hidden, then the normative plagioclase of the liquid remaining after crystallization of the exposed complex would have the same An content as the plagioclase crystals found in the uppermost rocks of the complex. If the hidden portion exceeded 50 per cent, the magma remaining after crystallization of the exposed complex would have a normative plagioclase more calcic than the actual crystals which were separating and which now are found in the uppermost exposures. The great mass of information available on plagioclase in basaltic rocks indicates that it is most unlikely that the liquid could be more anorthitic than the crystals separating from it in an environment such as we must be dealing with in this case.

TABLE 36.—*Estimated composition of the Hidden Zone*

	1. 463G70* The magma	2. Composite 1 + 2* The exposed complex	3. Hidden portion equals 40% The residue
SiO_2	50.9	50.3	52.0
Al_2O_3	17.7	19.6	14.8
Fe_2O_3	.3	1.0	5.1†
FeO	9.9	4.5	17.1 / 12.0
MgO	7.7	10.7	3.4
CaO	10.5	12.0	8.3
Na_2O	1.88	1.55	2.39
K_2O	.24	.06	.51
TiO_2	.45	.15	.90
P_2O_5	.09	.02	.19
MnO	.15	.10	.23
	99.81	99.98	99.82

Norm of No. 3

Q		8.70	
Pc	Or 2.78	51.29	
	Ab 20.43		
	An 28.08		
Di	Wo 5.34	10.85	
	En 1.81		
	Fs 3.70		
Hy	En 6.63	19.70	
	Fs 13.07		
Mt		7.42	
Il		1.67	
Ap		.44	
		100.07	

$\{An_{53}\ En_{40}\}$ Molecular %

$\frac{100\ Px}{Px + Pc} = 37$ wgt %

This places the amount of the hidden portion as between the extremes of 27.7 per cent and 50 per cent of the whole complex. Furthermore it is most unlikely that the true value can be very close to either extreme. Several lines of evidence suggest that it may be closer to the upper value than to the lower. The writer considers the most likely figure to be 40 per cent with a reasonable limit of error of ±5 per cent.

Figure 24 shows the positions of the estimated composition of the magma, the exposed portion, and the hidden portion, and the points marking the average composition of each of the zones in terms of plagioclase, clinopyroxene, and orthopyroxene. Olivine in these is arbitrarily converted to orthopyroxene.

The validity of the conclusion concerning the composition of the hidden zone might be questioned on the grounds that: (1) the sample representing the exposed complex

TABLE 37.—*Alternate estimated of composition of Hidden Zone*

	1	2
SiO_2	52.0	51.5
Al_2O_3	14.8	13.4
Fe_2O_3	5.1	5.3
FeO	12.0	12.5
MgO	3.4	6.2
CaO	8.3	7.0
Na_2O	2.39	2.22
K_2O	.51	.50
TiO_2	.90	.90
P_2O_5	.19	.20
MnO	.23	.25
	99.82	99.95

		Norms 1		2	
Q		8.70		6.70	
Pc	Or 2.78	51.29		2.78	46.42
	Ab 20.43			18.34	
	An 28.08			25.30	
Di	Wo 5.34	10.85		3.60	7.31
	En 1.81			1.60	
	Fs 3.70			2.11	
Hy	En 6.63	19.70		13.90	29.61
	Fs 13.07			15.71	
Mt		7.42		7.66	
Il		1.67		1.67	
Ap		.44		.44	
		100.07		99.81	
		An_{53}		An_{53} molecular %	
		En_{40}		En_{54}	
		37		44 100 Px(Px+Pc) wgt %	

1. Original estimate with remainder equal to 40% of whole complex.
2. Estimate modified by assuming Ultramafic Zone to be half as thick as in first estimate thereby changing composition of residue.

(Table 36, No. 2) departs from the true composition; (2) the border-facies analysis departs from the composition of the original liquid; (3) the ratio of the exposed complex to the hidden remainder as chosen is incorrect; and (4) the composition of the remainder is such that the conclusion appears unlikely for one or more of the above reasons. It must also be remembered that estimating the composition of the hidden

TABLE 38.—*Composition of Hidden Zone of complex assuming that remainder represents 35, 40, and 45 per cent of total volume of the complex*

	1 (35%)		2 (40%)		3 (45%)	
SiO_2	52.2		52.0		51.8	
Al_2O_3	14.1		14.8		15.3	
Fe_2O_3*	6.3	18.9	5.1	17.1	4.8	13.5
FeO	12.6		12.0		11.1	
MgO	2.3		3.4		4.2	
CaO	7.8		8.6		8.7	
Na_2O	2.51		2.39		2.30	
K_2O	.58		.51		.46	
TiO_2	1.05		.90		.80	
P_2O_5	.21		.19		.17	
MnO	.25		.23		.21	
	99.90		99.82		99.84	

Norms

Q		10.20		8.70		7.70
Pc Or	3.34		2.78		2.78	
Pc Ab	20.96	49.88	20.43	51.29	19.39	52.41
Pc An	25.58		28.08		30.24	
Di Wo	4.99		5.34		5.22	
Di En	1.30	10.25	1.81	10.85	2.11	10.50
Di Fs	3.96		3.70		3.17	
Hy En	4.41	18.80	6.63	19.70	8.33	20.61
Hy Fs	14.39		13.07		12.28	
Mt*		8.12		7.42		6.96
Il		1.98		1.67		1.52
Ap		.59		.44		.39
		99.82		100.07		100.09
An		An_{50}		An_{53}		An_{56} molecular %
En		En_{29}		En_{40}		En_{47} molecular %
100 Px = Px + Pc ratio		37		37		36 weight %

* Arbitrarily assumed that 30% FeO was oxidized.

portion involves the assumption that the Stillwater complex is a huge single intrusion which has after emplacement differentiated *in situ*. Unquestionably points 1, 2, and 3 each involve a certain amount of error. In spite of all these possible objections to the estimate of the composition of the hidden portion, the writer believes that none of them individually nor several of them together would cause a large error in the conclusion reached.

The close similarity of the border facies of the Stillwater complex and of the Bush-

veld complex strongly supports the supposition that they represent the undifferentiated magma for intrusions of this type. The rock types resulting in each case are practically identical. The normative and modal plagioclase of the Stillwater border facies is An_{68}, and the first crystals of plagioclase to accumulate on the floor of the magma chamber are AN_{86}. This agrees very closely with the liquidus and solidus relations of Bowen's curves for synthetic plagioclase. Similarly the MgO:FeO ratio of the border facies (59:41) and the composition of the first olivine precipitated in the Ultramafic Zone (Fo_{88}) agree well with the solidus–liquidus relations as shown on the curves for the synthetic olivine (Fo_{85}) investigated by Bowen and Schairer.

The writer considers that the supposition that the composite analyses represent the exposed complex involves the most inportant single source of error in the calculation. In such a large sheetlike intrusion, there may have been some shifting or oscillation over the floor of the magma chamber of the areas on which the crystals were accumulative. If these are numerous enough it may be supposed that they would in the long run average out. The greatest possibility for error in the composite analysis might lie in having overestimated the average thickness of the Ultramafic Zone. Judging from the Bushveld complex the early formed olivine-bronzite rocks occur in the very bottom of the "saucer" so that, if the present erosion surface cuts across the upturned Stillwater complex in such a manner as to bisect the "saucer," relatively too large an amount of these rocks has been included. In Table 37 an estimate of the composition of the remainder is given based on a thickness for the Ultramafic Zone of 1250 instead of 2500 feet and assuming that the remainder equalled 40 per cent of the whole complex. The main difference in composition resulting from this recalculation is an increase in the amount of MgO and consequently a greater proportion of, and a more magnesian, Hy in the norm. Qualitatively it does not change the general conclusions.

The ratio of the exposed complex to the hidden portion has already been discussed. Variation of this ratio within the limits considered likely does not very greatly change the composition of the residue. Table 38 shows the composition of the hidden portion assuming that it represents 35, 40, and 45 per cent of the total volume of the complex. Variation of the ratios beyond these ranges has been shown to be unlikely.

About 15 other recalculations were made based on a variety of assumptions, but all of them resulted in very similar results. Therefore, analysis 2 of Table 38 will be taken as representative of the hidden remainder for purposes of discussion. While this is not a precise picture of the remainder it must be fairly close.

The estimated composition of the hidden remainder is compared to rocks of the uppermost parts of the Bushveld complex and Great Dyke in Chapter IX.

Chapter VI

Mineral-Variation Curves

A. GENERAL DISCUSSION

The reasoning in this chapter may not be entirely clear without a knowledge of the contents of Chapters VII and VIII. The same may be said of Chapter VII with respect to VIII. Nor would it help to reverse the order of these three chapters for, no matter which is discussed first, the succeeding chapters contain elements vital to the understanding of the preceding. For the reader who may wish to follow the reasoning closely, the only solution is to reread Chapters VI and VII after having read all three chapters in the order here presented.

Fractional crystallization accounts for the broad changes in mineral phases found and for changes in the compositions of these phases as one proceeds from the floor toward the roof. The causes of the minor fluctuations in the mineral-variation curves are extremely complex and could be worked out only in sections where the samples were closely spaced and only by study of all available data on proportions of the various phases present in the rocks and their textural relations to one another. The nature of the processes causing the fluctuations will be discussed first and will be followed by an interpretation of the actual mineral-variation curves.

B. INTERSTITIAL LIQUID AND ITS EFFECT

To estimate the amount of interstitial liquid in the crystal mush after accumulation of crystals on the floor, it is necessary to examine certain rare types of rocks—namely, those in which the accumulated crystals represent a phase or phases not developed by crystallization from the interstitial liquid. In such rocks euhedrons of the accumulated crystals are found in a matrix (commonly showing a poikilitic development) of anhedral crystals of the phases separating from the interstitial liquid of the mush. The proportion of euhedral crystals to the interstitial anhedral crystals then represents the original proportion of crystals to liquid in the mush. The best example found of such a rock is specimen EB 199 of the Norite Zone (previously described in detail). In this specimen euhedral orthopyroxene makes up 71 per cent of the rock by volume, and interstitial plagioclase and clinopyroxene make up 29 per cent. Presumably the mush originally contained about 29 per cent liquid. Similar specimens studied give values ranging between 26 and 29 per cent interstitial liquid.[8]

The interstitial liquid differs very markedly from the contained crystals. If the crystals and liquid were sealed off from the main mass of magma and allowed to crystallize completely, either the original crystals would be considerably changed in composition or the crystals separating from the interstitial liquid would be very different from the accumulated crystals. The point is illustrated by a specimen from the upper part of the Norite Zone in which the pyroxene:plagioclase ratio is near 1:1. The pyroxene:plagioclase ratio of the interstitial liquid would, after crystallization,

[8] Wager (personal communication) believes the interstitial liquid may be as much as 50 per cent if growth of the accumulated crystals after settling be considered. This possibility was neglected in the writer's estimate.

be approximately the same and would contain only a minute amount of other constituents. The Mg:Fe ratio of the pyroxene crystals would be 4:1 and of the liquid approximately 1:1. Similarly the plagioclase crystals would be An_{80}, and the liquid would be An_{60}. Assuming approximately 30 per cent interstitial liquid, the effect of the pyroxene fraction of the liquid on the accumulated pyroxene crystals would be as follows: 70 per cent pyroxene with Mg:Fe ratio of 4:1, plus 30 per cent pyroxene with Mg:Fe ratio of 1:1 equals a final pyroxene with an Mg:Fe ratio of 71:29 provided reaction occurred. If no reaction occurred (unlikely considering the extremely slow cooling) then two pyroxenes of ratios 4:1 and 1:1 would be found.

If another specimen were chosen from the same locality and the same relations obtained except that the pyroxene:plagioclase ratio of the accumulated crystals was 0:1 (an anorthosite), then the composition of the plagioclase of the resulting rock would change very slightly toward that of the liquid, but the pyroxene would have an Mg:Fe ratio of 1:1.

It should be expected, therefore, that, if a gravity-stratified layer be examined, a lower norite portion should have pyroxene with an Mg:Fe ratio of 71:29 and an anorthositic rock a few feet higher should have pyroxene with an Mg:Fe ratio of 1:1.

One of the first experiments in this investigation was a test of the above hypothetical relations. A number of specimens from a layer such as described above were collected, and the compositions of the pyroxene and plagioclase were determined in the field by immersion methods under the microscope. The results indicated no appreciable change in the composition of the plagioclase or pyroxene across a layer or series of layers ranging from pyroxene rich to plagioclase rich. More accurate determinations were later made in the laboratory on the same specimens. The more accurate work proved that, although in a few specimens there was no appreciable change, in most there was a slight change in the direction indicated by the hypothesis, a change of 1–2 per cent in most examples rather than the 20 per cent to be expected if the hypothesis were entirely correct. The maximum observed change was about 7 per cent.

The fact that the interstitial liquid does, in most cases, affect the composition of the crystals in the direction (but not to the extent) to be expected indicates that the process suggested by the hypothesis probably is operative, but, since the change in composition is only a small fraction of that to be expected, some other counterbalancing process must also be operative. The nature of this second process may now be considered.

Elsewhere an estimate has been made of the length of time it would take for a 26,000-foot sheet of basaltic magma to lose enough heat to crystallize completely. From the total time available, assuming crystallization proceeds at a uniform rate, an estimate may be made of the rate of accumulation of crystals on the floor. Evidence presented in the discussion of layering indicates that the crystal mush on the floor of the magma chamber at any given time is normally about 5 feet thick. This mush increases in crystallinity downward so that below it the rocks are entirely crystalline solids. Above the mush rests the liquid magma. Heat is being lost from the floor downward at a very slow rate so that a gentle temperature gradient, perhaps a fraction of a degree, exists between the magma above and the bottom of the

mush. The rates of diffusion of diopside and plagioclase in melts of these two minerals have been investigated by Bowen (1921). It would appear that these rates are just about sufficient, with respect to the rate of accumulation of crystals and the total thickness of crystal mush on the floor, to permit diffusion to be effective.

The manner in which diffusion might operate to counter-balance the expected large changes which might have been brought about by the interstitial liquid is as follows: Suppose the crystal accumulate consists entirely of plagioclase An_{80} making up 70 per cent of the total bulk of the mush. The interstitial liquid consists of 50 per cent potential plagioclase An_{60} plus 50 per cent potential pyroxene. So long as the temperature remains constant the liquid and crystals, being in equilibrium, will remain constant. An extremely small temperature difference between the magma above and the bottom of the mush will upset this equilibrium and will result in the crystallization of a little plagioclase near An_{80}. For simplicity let us consider only Ca^{+2} and Na^{+1} ions instead of the more complex An and Ab molecules. The crystallization of a little anorthitic plagioclase reduces the concentration of Ca^{+2} in the pore space and enriches it in Na^{+1}. Thus a composition gradient is set up between the pore space and overlying magma. Suppose the temperature is temporarily held constant. The liquid is in equilibrium with the plagioclase crystals, and no further crystallization will occur if nothing is changed. If a Ca^{+2} ion by chance enters the pore space from the overlying magma or if an Na^{+1} leaves, then the equilibrium is to this degree upset, and one more anorthitic molecule may be precipitated on the surface of the accumulated plagioclase crystals. With random movement of ions the plagioclase crystals will in a sense gobble up chance Ca^{+2} ions entering the pore space until they have grown to fill the entire space. We may look at this in a different way, disregarding also the effect of crystallization. If we have a pore space connected by a channel to the overlying magma, and the pore has a higher ratio of Na:Ca than the overlying magma, then the purely random result of diffusion will tend to iron out this composition gradient.

In the example described above no pyroxene was present among the accumulated crystals. The pyroxenic constituents of the interstitial liquid might begin to precipitate and then follow a course analogous to the plagioclase, or might diffuse upward as the space becomes smaller as a result of crystallization of plagioclase. In the latter case it must be remembered that the crystallization of plagioclase enriches the liquid relatively in pyroxene as well as in Ab with respect to An so that there would result a similar composition gradient and a similar tendency to remove pyroxene constituents by diffusion.

The effectiveness of the diffusion process is closely related to the rate of accumulation of crystals on the floor. Various lines of evidence suggest that the rate of accumulation fluctuated from time to time and place to place. The cause of the fluctuation is probably related to the location, direction of movement, and velocity of convection and other currents in the magma. Where accumulation was slow, diffusion overcame the tendency of the trapped interstitial liquid to change the composition from that of the accumulated crystals. In such a case a norite with 50 per cent pyroxene might be followed by an anorthositic rock with only 10 per cent pyroxene all of which was of late, interstitial crystallization; the pyroxene of the anorthosite,

nevertheless, would have an Mg:Fe ratio just as high as the accumulated crystals of pyroxene in the norite. With more rapid accumulation some interstitial liquid would be trapped too far below the top of the crystal mush for diffusion to be effective. In this case the final residuum of interstitial liquid could produce more marked effects on the final mineral products, thus explaining the more marked deviations noted at some places in the mineral-variation curves. These deviations are much more marked for the pyroxene than for the plagioclase, but this is to be expected since, throughout the course of crystallization of the exposed part of the complex, the Mg:Fe ratio of the liquid is much farther from the Mg:Fe of the crystals than is the An:Ab ratio of the liquid from the An:Ab ratio of the crystals. Also, plagioclase represents 60 per cent of the volume of the average rock so that the effect of the interstitial liquid will be smaller.

The effects on the pyroxene and plagioclase to be expected from the two processes may be summarized as follows:

(1) The effect of the interstitial liquid should be proportionately greater on accumulated crystals present in small amount and less on accumulated crystals present in large amount. Somewhat more iron-rich pyroxene should appear in pyroxene-poor rocks than in pyroxene-rich rocks at any given horizon in the complex. Plagioclase-poor rocks should have somewhat more sodic plagioclase than plagioclase-rich rocks at any given horizon.

(2) The effect of the interstitial liquid should be pronounced where accumulation of crystals is rapid and diffusion ineffective. The effect of the interstitial liquid will become less as rate of accumulation decreases and may even be eliminated if accumulation is exceedingly slow.

C. INDICATORS OF THE RATE OF ACCUMULATION OF CRYSTALS

If a current in the magma has a vertical-velocity component greater than the normal downward settling velocity of the pyroxene, accumulation of all crystals on the floor will be interrupted. Decrease of the vertical velocity of this current will ultimately bring it below the settling velocity of the pyroxene, but it will still exceed the settling velocity of plagioclase. More slowing down of the current will be necessary before plagioclase will again deposit on the floor. In this ideal case a series of pyroxene-rich layers followed by plagioclase-rich layers will result, and relatively slow accumulation may be assumed at least for the lower portion of this sequence. A similar result might be produced by some other process or processes. Where one finds this distribution of pyroxene and plagioclase and where the compositions of these minerals show little effect from the interstitial liquid, it is reasonable to suppose that the deduction as to rate of accumulation is probably correct.

The relative rates of accumulation may be estimated more accurately by another means. Study of areas in which the layering is disturbed by slumping or landsliding of crystal mush on the floor indicates that the mush ordinarily is approximately 5 feet thick. (*See* discussion of Origin of Layering for more details.) Inasmuch as the rate of loss of heat downward is practically constant, the amount of mush on the floor will be proportional to the rate of accumulation. If 5 feet of mush be considered normal, less than 5 feet would be slow accumulation, and more, fast accumulation.

The difficulty in applying this criterion is that areas of such disturbed layering are not sufficiently common that one can be found at a desired level in the complex. Well-developed examples of disturbed layering are almost limited to the Norite Zone. One conclusion may be drawn by application of this criterion: where crystals accumulated at the "normal" rate (5 feet of mush), the effect of the interstitial liquid is small. It can produce changes in Mg/Fe ratio of the pyroxene of about 1–2 per cent.

D. ORIGIN OF MONOMINERALIC ROCKS

Crystal accumulation from a basaltic magma can never produce a truly monomineralic rock. There is at least 26–29 per cent of trapped interstitial liquid of basaltic composition which should crystallize largely as both pyroxene and plagioclase. Nevertheless anorthosite layers are found with 100 per cent plagioclase, and similarly bronzitite layers with 100 per cent bronzite. The absence of plagioclase from most harzburgites in the Stillwater presents the same problem. Most of the explanation lies in a diffusion mechanism. For example, if only bronzite is accumulating on the floor, as in the upper part of the Ultramafic Zone, and, if accumulation was slow so that diffusion could be effective, then the following result might be expected. Slow loss of heat by conduction through the floor would permit a little more bronzite to crystallize, causing the crystals in the mush to grow. The pore-space liquid would be enriched slightly in Fe^{+2} with respect to Mg^{+2} and in plagioclase with respect to pyroxene. Diffusion from the pore space of the mush to the magma above and vice versa would remove the composition difference between the pore-space liquid and the magma above. More bronzite could then crystallize. If this went on slowly enough and for long enough, all the pore space would be filled with bronzite, and a monomineralic rock would result. Bronzitites and anorthosites formed in this manner are not uncommon among the Stillwater rocks, but it is more usual to find a not quite perfect development. Thus in the anorthosites of the Anorthosite Zone it is common to find skeleton poikilitic pyroxene crystals which have grown from the pore-space liquid. These are commonly an inch or more in diameter and may be separated from one another by several inches. In such a poikilitic crystal the pyroxene may make up only 20 per cent of the volume; the rest is included peri-euhedral plagioclase. Similarly in the harzburgites of the lower part of the Ultramafic Zone one finds poikilitic bronzite growing around (and at the expense of) the settled olivine of the mush. Again these poikilitic crystals are large—from an inch to more than a foot. It seems characteristic of crystallization from the pore-space liquid that very few crystal nuclei form but that once a single nucleus develops it will grow to comparatively large size. The plagioclase of feldspathic pyroxenites and feldspathic harzburgites also tends to form poikilitic crystals but generally smaller than the poikilitic pyroxene. It can thus be seen that there is no particular problem in deriving monomineralic rocks or almost monomineralic rocks from a basaltic magma in a Stillwater type of intrusion.

E. INTERPRETATION OF TEXTURAL RELATIONS

The interpretation of textural relations from thin sections alone is often misleading even if due consideration is given phase relationships known from laboratory

experiments. The field relations must also be known. Looking at a feldspathic pyroxenite with interstitial plagioclase one might conclude that the residual liquid was rich in plagioclase. An anorthositic rock perhaps only a few feet away from the pyroxenite in a rhythmically layered sequence would suggest a pyroxenic residual liquid judging from its interstitial pyroxene. Neither conclusion is valid. One may not draw conclusions as to paragenesis of phases separating from the magma from observations on such thin sections.

Harker (1908) followed by Tomkeieff (1945) fell into just this sort of error in interpreting the textural relations of the rocks from Rum. The mafic layers with interstitial plagioclase and the felsic layers with interstitial olivine were considered separate intrusions. Assuming a cotectic relationship between the plagioclase series and the olivine series they concluded that the composition of the magmas was successively on one side of the cotectic or the other as the textural relationships suggested. Wager and Brown (1951) recently reinterpreted the field relations at Rum as a rhythmically layered sequence rather than successive intrusions of differing composition.

Wells (1952) discusses textures of Bushveld norites as seen in thin section. The textures are described and illustrated most meticulously. Wells follows very much the same line of reasoning as Tomkeieff. This results in some highly complex alternative hypotheses to explain the textures and in different hypotheses for individual slides. If one keeps in mind that one is dealing with an accumulation of crystals and an interstitial liquid and then if one applies the diffusion mechanism to the liquid, all the relationships are explained by a single rather simple process.

The origin of ophitic texture in many basalts and dolerites has long been debated. It has been suggested that it indicates crystallization of plagioclase ahead of pyroxene or alternatively that plagioclase has a greater "crystallizing force" than pyroxene. Neither explanation is satisfying. A clue is provided by the Stillwater rocks. Here the poikilitic pyroxene crystals in anorthosites or harzburgites grew larger than the similar poikilitic plagioclase crystals in feldspathic pyroxenites and feldspathic harzburgites. This suggests that plagioclase in a basaltic liquid has a somewhat greater tendency to form nuclei than does pyroxene. If there were for example 20 plagioclase nuclei to 1 pyroxene, and equal volumes of both phases crystallized, ultimately a single pyroxene would enclose a large number of plagioclase crystals. This might be tested by comparing sizes of phenocrysts of pyroxene and plagioclase in basalts. The writer has made no such study. It certainly does not hold for all basalts, nor are all dolerites ophitic. Some special factor must be involved which tends to retard nucleation of the pyroxene relative to plagioclase.

F. INTERPRETATION OF MINERAL-VARIATION CURVES OF DEAD TREE SECTION

With the above theoretical discussion in mind the mineral-variation curves of the Dead Tree detailed section may now be considered (Pl. 9; Table 39). The minor oscillations of the curves can be successfully interpreted in terms of the effect of the interstitial liquid under varying rates of accumulation of crystals. The fact that this can be done gives the writer considerable confidence in the soundness of the theoretical discussion.

Among the extreme rock types, one good pyroxenitic horizon may be noted (specimen 199),[9] and two anorthositic horizons (181, and 207) (Pl. 9). In all these rocks, the minerals present in small amount (the plagioclase of the pyroxenite and pyroxene of the anorthosite) show marked effects produced by the interstitial liquid. In the two anorthosites the orthopyroxene has a composition ranging from $Mg_{71\frac{1}{2}}$ to $Mg_{72\frac{1}{2}}$, while the average composition of hypersthene, in adjacent rocks of normal Px:Pc ratio, is approximately $Mg_{77\frac{1}{2}}$. Thus it appears that the interstitial liquid has produced a 5 per cent decrease in the Mg:Fe ratio. The trapped interstitial liquid itself, however, if unmodified by diffusion, would have produced an orthopyroxene near Mg_{50} or 28½ per cent rather than 5 per cent lower in Mg:Fe ratio. Less than one-fifth of the effect to be expected without diffusion is found. In the pyroxenite, the late interstitial plagioclase present in small amount has a composition of $An_{73\frac{1}{2}}$, whereas the average composition of the plagioclase in adjacent rocks is $An_{78\text{-}79}$. A 5 per cent decrease in An content is found, but a 15 to 20 per cent change would have resulted if no diffusion had occurred.

The plagioclases of two of the anorthosites represent peaks on the composition curves. They show ½ to 1½ per cent higher An content than the plagioclase in adjacent rocks of normal composition. This suggests that in general the plagioclase of average norites and gabbros is changed in composition by the interstitial liquid to the extent of about 1 per cent enrichment in Ab. Similarly the pyroxene of the pyroxenitic rocks and mafic norites or gabbros shows small peaks in the composition curves amounting to from ½ to a little more than 1 per cent.

The peak in Mg:Fe ratio on the orthopyroxene curve comes not at the pyroxenite horizon (199) but about 65 feet lower in the section (195–197), and no peak in An content occurs at anorthosite 181. Both of these apparent discrepancies are thought to be related to the rate of accumulation of crystals; when the rate of accumulation and its limiting effect on diffusion are considered the deviations can be explained. Both pyroxenes of specimens 181, 182, and 30 are conspicuously low in Mg:Fe ratio even though 182 has a relatively high pyroxene content—nearly 40 per cent. There is also a marked sag in the plagioclase composition curve between specimens 180 and 186. Both observations indicate relatively rapid accumulation which has acted as a check on the effectiveness of diffusion.

From 191 to 196 the Mg content of the orthopyroxene is continually rising as is the total pyroxene content, but from 197 to 199 the Mg content of the orthopyroxene begins to drop, whereas the total pyroxene continues to increase. Presumably the rate of accumulation of crystals was decreasing from 191 to 196. Then followed a flood of pyroxene as an upward current subsided. This lessened the effect of diffusion and produced a downward trend of the Mg curve even though the total pyroxene, and hence the effectiveness of the interstitial liquid to change the pyroxene toward a more iron-rich composition, was diminishing. The marked downward trend of the plagioclase curve to 199, the greatest dip in this curve noted anywhere in the East Boulder Plateau section, is the result then of rapid accumulation, limiting diffusion, and the relatively low plagioclase content of the rock.

Between 199 and 207 are relatively pyroxene-rich norites, and then a deluge of

[9] The EB designation of the sample numbers has been dropped in this discussion and on Plate 9, and only the numerals are used

TABLE 39.—*Relations between composition and rate of accumulation of crystals*

	1 Deduced current in magma	2 Predictable effect of current on		3 Predictable effects complementary to (a) and (b) of		4 Combined effects of (a), (b), (c), and (d) to produce observed mineral-variation curves
		(a) rate of accumulation	(b) Px:Pc ratio	(c) interstitial liquid	(d) diffusion	
A	Magma quiet (190–191)	Normal	Normal (35:65±)	Effect small	Effect large	Pyroxene:plagioclase ratio equivalent to ratio of precipitation from magma and the composition of both pyroxene and plagioclase are average for this part of complex (±1% richer in Fs and Ab respectively than original accumulated crystals)
				Pyroxene and plagioclase present in more than small amount so effect of interstitial liquid on each will be small; accumulation normal so diffusion will be effective in eliminating all but about 1% of change which could have been produced by interstitial liquid on accumulated crystals		
B	Beginning of upward current velocity at first increasing (192–197)	Decreasing	Increasing	Small and decreasing	Large and increasing	Per cent pyroxene increasing with respect to plagioclase; pyroxene becomes increasingly magnesian to maximum at 197 as a result of (1) slow accumulation and high effectiveness of diffusion, and (2) high percentage of pyroxene so that interstitial liquid would in any case have small effect; plagioclase present in smaller amounts is more affected by interstitial liquid, but slow accumulation permits diffusion to operate effectively and thus prevent any downward trend in plagioclase curve
		But high in magma, as at A in Figure 29, a concentration of plagioclase is taking place; Failure of plagioclase to settle produces the increasing ratio in (b)		Slow rate of accumulation allows diffusion to be highly effective so that interstitial liquid has less effect on accumulated crystals, but plagioclase present in decreasing amount will be more affected by interstitial liquid than pyroxene		

C	Slackening off of current (198–200)	Increasing	Increasing	Increasing from small to moderate	Decreasing from large to moderate	Percentage of pyroxene reaches maximum; pyroxene curve bends downward as diffusion becomes less effective but high percentage of pyroxene limits effect of interstitial liquid to about 1% change, while plagioclase, present in small amount, drops five times as much
		Both pyroxene and plagioclase previously held back by upward current now begin accumulating at more rapid rate; pyroxene, which was held back less and which settles faster, reaches the floor in increasing amounts		Increasing rate of accumulation limits effectiveness of diffusion, but pyroxene present in great excess can be only slightly affected by interstitial liquid; plagioclase present in small amount may show extreme effect of interstitial liquid		
D	Magma quiet; but abnormal distribution of crystals in magma above still exists as a result of previous motion in magma (201–205)	Fairly rapid to normal	Decreasing to normal	Small to moderate	Large to moderate	Percentage of pyroxene somewhat higher than normal; effect of interstitial liquid small on each (1–2%); no marked lows on either pyroxene or plagioclase curves; grain size decreasing indicating some size sorting
		Fairly rapid precipitation of crystals previously retarded by rising current	Excess of pyroxene at first as result of previous current as in stage 3 (Fig. 29)	Pyroxene and plagioclase both present in more than small amounts; therefore interstitial will not have very pronounced effect on either of them; accumulation rate fairly rapid		
E	Magma quiet (206–214)	Fairly rapid	Low	Fairly large	Fairly small	Deluge of plagioclase complementary to pyroxenite (199–200) arrives at floor; high 1½% above average in An at plagioclase maximum since plagioclase cannot be much affected by interstitial liquid; accumulation fairly rapid, low pyroxene content in many of the rocks and practically all pyroxene of late interstitial crystallization; marked lows at 205, 207, 209,210 on pyroxene curves result
		Deluge of plagioclase previously held up by rising current now reaches floor	Plagioclase excess "D" of stage 3 greatly decreases pyroxene: plagioclase ratio	High plagioclase content prevents interstitial liquid from having much effect on plagioclase; accumulation rapid resulting in marked lows of pyroxene curve as result of insufficient time for diffusion		

TABLE 39.—(*Continued*)

	1 Deduced current in magma	2 Predictable effect of current on		3 Predictable effects complementary to (a) and (b) of		4 Combined effects of (a), (b), (c), and (d) to produce observed mineral-variation curves
		(a) rate of accumulation	(b) Px:Pc ratio	(c) interstitial liquid	(d) diffusion	
F	Magma quiet (215–218); effects of previous currents now passed	Normal	Normal	Small	Large	Pyroxene and plagioclase are now accumulating at rate of their precipitation from magma; accumulation is normal so effect of diffusion is sufficient to bring both plagioclase and pyroxene to about 1% below the An and En content of original accumulated crystals
				Same as A		

plagioclase forms the anorthosite at 207. This anorthosite may be the complement of the pyroxenite 225 feet below. A short distance above 207 a normal pyroxene:plagioclase ratio is again established, and normal rather uniform composition is indicated by the pyroxene and plagioclase curves from this point, 211, to the top of the detailed section at 218. No marked lows are found on the sensitive orthopyroxene curve in this region.

An analogous relationship exists with regard to the mafic norites 184–185 and a complementary anorthositic norite about 60 feet above.

If we turn ahead to the theoretical discussion of rhythmic layering (Chapter VII) we see that the hypothetical conditions set up and illustrated by Figure 29 with regard to changes to be expected in pyroxene:plagioclase ratio and relative rates of accumulation are exactly fulfilled in the detailed section from 190 to 218. At the same time, after the pyroxene:plagioclase ratio and rate of accumulation are taken into consideration, the effects on composition of the crystals are exactly as predicted in the discussion of the effect of the interstitial liquid and diffusion.

A summary of the correlation of currents, with their effects on rate of accumulation and pyroxene:plagioclase ratio, and the observed variations in mineral proportions and compositions found in the detailed section from 190 to 218 is given in Table 39. Inasmuch as the interrelations of all these factors and their effects on the mineral-variation curves are highly complex, the table may serve to integrate the relations. It is considered that the relations can be explained as a single cycle from quiet magma conditions to a rising current in the magma and return to quiet conditions.

A cycle of the type illustrated in Table 39 covers a stratigraphic range of nearly 1000 feet in the complex. With work as detailed as this particular section there are indications that half a dozen or more such cycles might be found in the complex. It must be remembered that much smaller cycles are superimposed on these large ones. In reality each gravity-stratified layer (rhythmic layering) represents such a microcycle. The gravity-stratified layers probably represent slight currents operating in a zone of the order of magnitude of perhaps 100 feet from the floor, whereas the large cycles probably represent something like a complete convective overturn of the whole magma mass. The microcycles are indicated by the gravity-stratified layers, in which one finds perhaps a dozen sharply defined layers within 10 feet. A process operating far above the floor would result in mixing of the pyroxene and plagioclase crystals and a blurring of these sharply defined layers.

G. INTERPRETATION OF MINERAL-VARIATION CURVES FROM FLOOR TO DEAD TREE SECTION

The East Boulder Plateau is not favorable for collecting a suite of samples of the Ultramafic Zone in stratigraphic sequence. The layering is gently folded and gently dipping. Many samples were examined, but their precise stratigraphic positions were not known. A suite of samples collected by Peoples from the Benbow property east of the Stillwater Valley was examined.

The Border Zone has an inverted pigeonite, orthopyroxene, composition Mg_{89}, and a plagioclase, composition An_{88}, agreeing with the norm of the chemical analysis.

The Border Zone rocks grade via feldspathic websterites into the typical harzburgites of the lower half of the Ultramafic Zone (Table 14). Above the harzburgites, bronzitites are the predominant rock type to the top of the zone (Table 15).

Throughout the Ultramafic Zone the bronzite is near Mg_{86}. Fluctuations in composition of $\pm 1-1\frac{1}{2}$ per cent are common, and the olivines of the harzburgites vary sympathetically with them, averaging near Fo_{87} in composition. In the upper few hundred feet the bronzite becomes more iron rich, reaching Mg_{82} at the top of the zone (Table 18) and further decreasing in the Norite Zone to Mg_{80} at the base of the Dead Tree section. The most magnesian bronzite (Mg_{89}) found in the complex occurs with the chromite of the chromite horizon. It would appear that the chromite takes the Fe out of the interstitial liquid thus permitting a slightly more magnesian orthopyroxene than normal to survive.

The interstitial plagioclase crystals scattered throughout the Ultramafic Zone generally are near An_{73} but vary considerably. At the base of the Norite Zone the plagioclase is An_{86}. These are the first settled plagioclase crystals and the most calcic in the complex. In the Norite Zone they gradually decrease to An_{81} at the base of the Dead Tree section.

Clinopyroxene is generally absent from the harzburgites of the lower part of the Ultramafic Zone but is found as large emerald-green poikilitic crystals in the bronzitites. The composition is fairly constant near $Ca_{40}Mg_{52}Fe_8$ with 1 per cent Cr_2O_3.

The bronzite crystals of the harzburgites are developed from the pore-space liquid and by reaction of the liquid with olivine and are all poikilitic. Nevertheless they show no significant difference in composition from the settled orthopyroxene crystals of the bronzitites. Only olivine was being precipitated from the magma during formation of the lower portion of the Ultramafic Zone, and only bronzite precipitated during formation of the upper portion. The middle portion consists of intercalated harzburgites and bronzitites. Evidently during development of this portion fluctuations between olivine and bronzite precipitation took place. Rarely are the two phases mixed in a single layer. This fluctuation probably is due to inhomogeneities in the overlying magma whereby certain portions pass through the point where crystallization changes from olivine to orthopyroxene only to be replaced in the cooling region under the roof by magma which has progressed less far in the course of crystallization resulting in a return to precipitation of olivine. The densities of olivine and bronzite of the compositions with which we are concerned are so close that there would be no difference in settling rates for crystals of the same size. The boundaries between harzburgite and bronzitite layers are ordinarily very sharp.

The variations in compositions of the phases present are shown in the general diagram for the complex as a whole (Fig. 25).

H. INTERPRETATION OF MINERAL-VARIATION CURVES FROM THE DEAD TREE SECTION TO THE HIGHEST EXPOSURE

Inasmuch as the data are sufficient to show fluctuations of the three main phases diagrammatically in Plate 10, no extended discussion of compositions of the minerals or rocks is necessary.

The orthopyroxene and clinopyroxene curves vary sympathetically, and most of

their fluctuations are also reflected in the plagioclase curve. The main difference is that the plagioclase curve shows gentle rises where the composition is anorthositic, while the pyroxene curves are depressed. Four distinct rises and falls of the curves suggest major convective overturns of the magma.

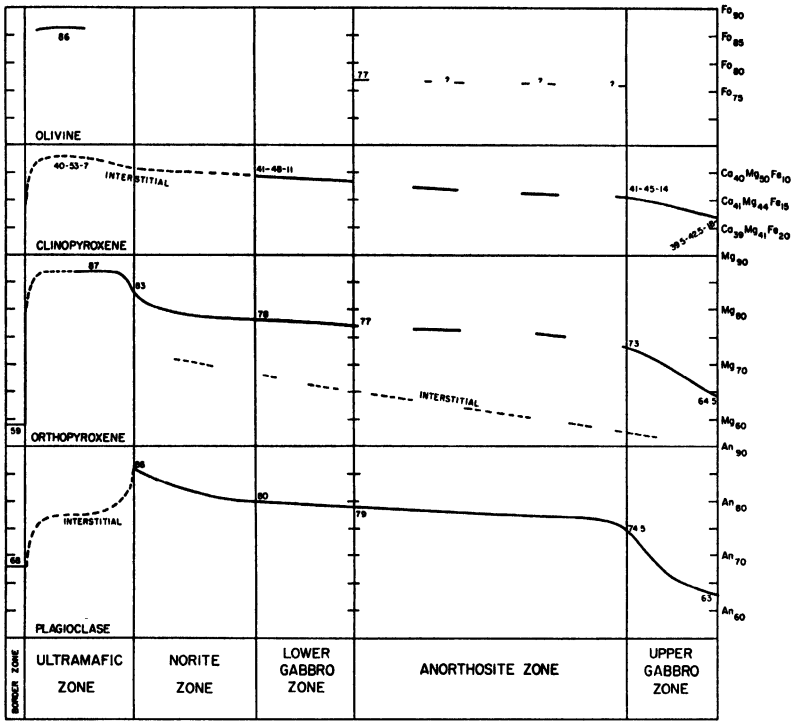


FIGURE 25.—*Mineralogical changes in complex as a whole*

I. INTERPRETATION OF MINERAL-VARIATION CURVES FOR THE COMPLEX AS A WHOLE

The general trend of fractional crystallization of the complex is well shown in Figure. 25 Plagioclase decreases in anorthite content from An_{86} to An_{63}, orthopyroxene decreases in Mg:Fe ratio from Mg_{87} to Mg_{66}, and clinopyroxene changes from $Ca_{40}Mg_{52}Fe_8$ to about $Ca_{39}Mg_{42}Fe_{19}$.

The magma has a normative plagioclase of An_{68}, and all but the last few hundred feet of the exposed complex is more anorthitic than the magma, leaving a large quantity of more sodic plagioclase in the hidden portion. All the pyroxene is much more magnesian than the Mg:Fe ratio of 60 of the normative pyroxene. This indicates great enrichment in Fe in the hidden portion and portends the crystallization of iron oxides.

At least six, and possibly more, major fluctuations of the curves are indicated.

Plagioclase and orthopyroxene curves steepen in the Upper Gabbro Zone, indicating more rapid changes in composition with decreasing bulk of residual magma.

J. OBSERVATIONS ON GRAIN SIZE

It would be profitable to make a much more comprehensive study of grain-size variation in the complex as a whole and also at certain critical horizons. Grain-size variation was not very evident in the field, and consequently in collecting this factor was ignored. It is unknown whether the settling units are single grains or glomeroporphyritic aggregates. Elsewhere it was suggested that the chromite crystals may have coalesced in chains during settling and that at least some of the plagioclase settled as glomeroporphyritic aggregates. Irregular spots of 10–20 grains of plagioclase are common in gabbros of the complex. On the other hand similar clusters of pyroxene were not noted.

In one of the olivine-bearing horizons of the Anorthosite Zone a most peculiar rock was encountered. It had oval-shaped plagioclase spots about the size of an egg, sprinkled rather regularly through the rock with a spacing of about 1 foot. Each "egg" had a single olivine crystal exactly at the bottom (stratigraphically) of the aggregate. There can be little question here that the olivine oriented the cluster as it settled. Whether glomeroporphyritic aggregates were common or rare during settling is a question. The writer's impression is that they were relatively rare and largely related to the plagioclase.

In Chapter VII settling rates are discussed at length. Considering that settling varies as the square of the radius and only as the first power of the density difference, the absence of conspicuous size sorting is puzzling. This might be accounted for by a rather uniform size of the crystals of each mineral at any given time during the settling. It is not uncommon for phenocrysts of one mineral in a porphyry to be all about the same size. There is no good method for determining grain-size variation in a given sample. The random section presented on the slides may give any size from the maximum diameter to vanishingly small size where the slice was made so that it cut through one small corner of a crystal. One does get a subjective impression that the larger crystals are two or three times greater in diameter than the smaller crystals.

Certainly one might expect size sorting in rhythmically layered sequences. Nevertheless no striking examples of this were noted. However, well-developed rhythmic layering was limited to the Norite Zone where both a size factor and a density factor worked in the same direction—large orthopyroxene and small plagioclase—so that the pyroxene had a settling velocity of 25 times that of the plagioclase.

The absence of evident size sorting and the fact that a very large differential in settling rates is necessary to produce rhythmic layering suggest that the distance settled was small—perhaps 300 feet or less. In Chapter VIII it is postulated that a dilute suspension of crystals forms in the magma beneath the roof, and this layer, as it becomes denser, periodically overturns and spreads out on the floor. Thus crystal settling would occur from a very thin layer.

On Plates 9 and 10 the most significant size-variation curve is that for the plagioclase. Plagioclase represents on the average almost 70 per cent of the crystals by

volume, and hence it is relatively little changed by the crystallization from pore-space liquid. The plagioclase crystals of the Anorthosite Zone are nearly twice as large as those in other zones. Particularly striking is the abrupt drop in plagioclase grain size at the contact of the Anorthosite Zone with the Upper Gabbro Zone. The larger size in the Anorthosite Zone might be a result of lack of competition for common ions if the writer's hypothesis on origin of these anorthosites were accepted. Only plagioclase would be separating as crystals from the magma. In other cases pyroxene crystals are also separating with resulting competition for Ca^{+2} ions, Si^{+4} ions, and to a lesser extent Al^{+3} ions.

In the average grain-intercept curves of Plates 9 and 10, most of the fluctuations are small, and commonly several points determine the slopes. One does not as a rule see erratic fluctuations which would require more points to determine the true position of the curves. This is rather clear evidence that size sorting on the same scale as rhythmic layering is not present. The actual fluctuations are commonly on a much broader scale suggesting gradual changes from sample to sample through a vertical range of 500 to 1000 feet. Inflection points at which the grain-size curves change from downward to upward slopes or *vice versa* may as a rule be correlated with petrographic changes in the rocks. In the lower part of the Anorthosite Zone plagioclase grain size decreases upward as far as a troctolite horizon, above which grain size increases. At the base of the Upper Gabbro Zone feldspar crystals are very small; they become increasingly larger up to the highest exposure. [Note the actual points measured are not shown on the average grain-intercept curves. For every rock for which a Rosiwal analysis was made a point for average-grain intercept was plotted. The positions of all these points can be seen on the curves showing percentage of minerals present. The grain-intercept curves were drawn by connecting adjacent points by straight lines with no smoothing.]

Chapter VII

Layering and Its Origin

A. DESCRIPTION OF LAYERING IN THE STILLWATER COMPLEX

1. GENERAL CONSIDERATIONS

The terminology used in connection with layering has already been described in Chapter IV; it is essentially the same as that used by Wager and Deer in their Skaergaard memoir.

Layering of several types and orders of magnitude are found in the Stillwater complex. Standing on an outcrop, one may observe layers of the order of magnitude of a few inches to a few feet in thickness. Looking at a valley wall several miles away, one may see conspicuous layering, but this is on a scale of 100 feet thick or more. Groups of the smallest-scale layers form apparent single large layers when viewed from a distance, and these are separated from one another by thick, essentially nonlayered norites or anorthosites. On a scale of thousands of feet, layers equivalent to the zones of this report may be postulated, but they are generally too large to be seen in the field as layers. They are recognized only by mapping and in some cases not until after laboratory study of the rocks.

2. RHYTHMIC LAYERING

a. Thickness of layers.—Looking at a series of layers in an outcrop, one would estimate that their thickness was commonly measurable in inches to a few feet. In most occurrences the layers are distinct and easily recognizable, but in some the definition of exactly what constitutes a layer becomes arbitrary. For example, one may find an anorthosite layer 1 foot thick in the middle of which is a thin sheet of pyroxene, two or three crystals thick. The pyroxene crystals might be disregarded and the whole considered a single anorthosite layer, or one might subdivide it into two 6-inch anorthosite layers with a thin pyroxene-rich layer between them. A series of layers, as defined by the writer, was measured in the field to determine thicknesses. In 217 feet, 75 layers were found. Thus the average thickness would be about 3 feet. If, however, 6-thick layers of uniform, nonlayered norite contained in this sequence are omitted, the remaining 69 layers average 16 inches in thickness.

b. Lateral extent and continuity of layers.—As a rule the layers form exceedingly flat lenses. It is rather difficult to determine the dimensions of these lenses. Standing on an outcrop one can rarely look along a 6-inch layer in both directions to the points where it lenses out. Many such layers were found and traced for hundreds of feet of continuous outcrop without reaching such a termination. A section of outcrop 500 feet long was mapped in detail to get some information on this point. A thin anorthosite layer extended the entire length of the outcrop, and the base of this layer was used as a horizon marker to map the sequence of layers above and below. Eleven traverses were made at right angles to the layering at approximately equal intervals along the strike. The result is shown in Figure 26. A vertical exaggeration of five times was used in constructing the diagram to emphasize any changes in the thickness of

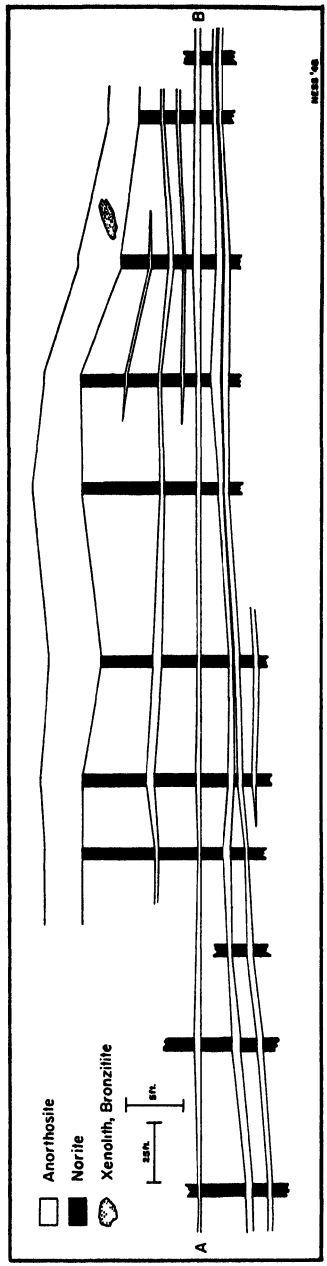


FIGURE 26.—*Section illustrating variation of rhythmic layers along the strike, west side of Stillwater valley, Norite Zone*
Vertical exaggeration × 5

the layers. Considerable changes in thickness do occur, and several layers pinch out within the limits of the diagram. Thus the diameters of the lenses are of the order of magnitude of several hundred to perhaps a thousand times their thicknesses.

c. Attitude of layers.—In general it appears that the layers originally formed in an essentially horizontal attitude, parallel or subparallel to the floor of the lopolith. In some cases they may have been deposited on a slight slope, perhaps as a result of sagging of the floor of the lopolith during crystallization, tectonic effects, or inequalities in the rates of sedimentation of crystals on the floor from point to point.

d. Contacts between layers.—Contacts are as a rule sharp, but may be gradational. Where they are sharp, the crystals interlock across boundaries in thin section.

e. Relations between layers.—Layers never crosscut one another as do dikes, but their relations are exactly like beds of clastic sediments.

f. Mineral variations across layers.—Some series of layers show practically no variation in the composition of constituent minerals across the whole sequence regardless of the proportions of the minerals present. Study of the mineral-variation curves shows that under certain conditions, presumably very slow accumulation of crystals on the floor, there is almost no change in the composition of the minerals. Under conditions of rapid accumulation of crystals, small changes of composition are observed, and these are correlated with effects produced by the trapped interstitial liquid. It is significant that rhythmic layering with conspicuous variation in mineral proportions does not *necessarily* entail variations in the composition of the minerals.

g. Orientation of crystals.—Elongate or tabular crystals tend to lie with their longer axes in the plane of layering. This produces what has been defined in Chapter IV as "lamination". It is conspicuously developed (apparent in field; recognizable in hand specimens) at a few horizons in the complex. As a rule the axes of the plagioclase and pyroxene crystals are not sufficiently elongated to make lamination conspicuous. No doubt petrofabric analyses would show lamination in almost any specimen (Pl. 6). Van der Berg (1946), in a study of dimensional orientation in Bushveld norites, defines the degree of orientation precisely and demonstrates a very slight degree of linear orientation in the plane of lamination.[10] Hand specimens cut and polished parallel to the plane of lamination as a rule show little or no linear orientation of the longest axes in this plane, which indicates that the crystals fell vertically and that currents moving horizontally were not present in most places, at least within a few feet above the floor. Channels with strongly developed linear orientation such as found locally in the Skaergaard intrusion by Wager and Deer were not found, nor were potholelike structures with tangential lineation as described by Schmidt in the Bushveld.

h. Gravity stratification.—Gravity stratification or density gradation within layers, described by Peoples (1936), consists of gradations from minerals of higher density at the base of a layer to minerals of lower density at the top of the layer. Layers may thus grade from pyroxene rich at the base to anorthositic at the top, or perhaps from chromite rich below to olivine rich above. This feature may be compared to graded bedding in sediments. In sediments graded bedding can be used to determine top

[10] The writer would attribute this slight lineation to deposition on a gentle slope rather than to orientation produced by deposition from flowing magma.

and bottom of a bed, and similarly gravity stratification can be used to determine top and bottom of a layer in tectonically disturbed areas (Buddington, 1936). Perhaps a third of the Stillwater layers show clearly developed gravity stratification; others show only traces of it or none. Some reversed gravity stratification is observed as a

TABLE 40.—*Comparisons of rhythmically layered sequences*

Section A	Section B
(Feet)	(Feet)
—	5.5
—	32†
20±*	1*
135†	47†
1*	21*
11†	22†
2*	16*
26†	28†
5*	28*
30†	6†
10*	5.5*
50†	30†
252	252
Totals 38 (15%)	77 (30%)*
214 (85%)	175 (70%)†

* Thin-layered

† Uniform norite

Section A. Ratio of thickness of uniform norite to thin-layered sequence above it ranges from 5:1 to 11:1.

Section B. Ratio of thickness of uniform norite to thin-layered sequence above it is highly variable.

result probably of either a downward current immediately above the floor as explained elsewhere or the telescoping of two layers.

i. Regularity of layers.—Most layers appear to be sharply bounded and regular. Within the limits visible from an outcrop, they appear to be bounded by plane-parallel surfaces. Only by mapping the layers and making sections with considerable vertical exaggeration do the lenticular character and variation in thickness along the strike become evident. Rare but highly significant disturbed layering can be found particularly within the Norite Zone of the complex and also at the chromite horizon of the Ultramafic Zone.

j. Distribution of layered sequences and thick norite layers.—In walking across the well-developed rhythmic layering of the Norite Zone one gets the impression that the entire mass is layered on a scale of about 1 foot, but detailed measurements indicate that actually sequences of these thin layers are separated by thick, as a rule uniform, norites. Two sets of measurements are given in Table 40. The thin-layered sequences make up only 15–30 per cent of the total section measured. In Section A there seems to be a fairly constant ratio between the thickness of the underlying

norite and the overlying thin-layered sequence, ranging from 5:1 to 11:1. In section B no such regular ratio is apparent. However, the well-developed thin-layered sequences are subordinate to the thick intervening, essentially nonrhythmically layered, norites. This suggests that the process forming rhythmic layering was operative only at occasional intervals during the deposition of crystals on the floor.

The suggestion of some sort of rhythmical deposition of thin-layered sequences followed by thick, nonlayered uniform norites and the seemingly nearly constant ratio of thickness between the two were mentioned by the writer in a previous paper (Hess, 1938b). In the Skaergaard memoir Wager and Deer (1939) describe and illustrate sequences which show a much more nearly perfect arrangement than any found in the Stillwater complex. In the Skaergaard (*see* Plate 8 of that memoir) considerable thicknesses of layered rocks show a sequence of uniform gabbro layers, all of about the same thickness, and each separated from the next above it by a single gravity-stratified layer about a tenth as thick. The uniform gabbro layers appear to be about 5 feet thick. The more nearly regular development of layers and groups of layers in the Skaergaard intrusion might well be related to its shape, in which conditions for regular convection-current cycles would be ideal. A sheetlike body such as the Stillwater might be expected to develop much more irregular convection currents. The ideal shape for a convection cell is one in which the horizontal diameter is slightly larger than the vertical dimension.

3. DISTURBED RHYTHMIC LAYERING

Disturbances of layering are quantitatively very rare in the Stillwater complex. They may be considered as exceptions to the rule of great regularity and represent exceptional conditions during their development. These disturbances occur at several horizons in the complex. It appears that they persist for long distances along such horizons. They seem to be much more common in the Norite Zone than elsewhere. Since, however, it would be much more difficult to see them where rhythmic layering was less well developed, it may be that they are present at many other horizons but their effects cannot be readily observed.

Two detailed sections were made of outcrops exhibiting disturbed layering (Pl. 7) on the west side of the Stillwater Valley in the Norite Zone. The disturbed sequences illustrated are 10 feet and 3 feet thick respectively. The floor below the disturbed area is smooth and straight, showing no indication of disturbance. The overlying layers are deposited on the disturbed sequence unconformably, filling depressions and thinning above upward projections just as would sediments deposited on an uneven floor. Above the unconformity the contacts between layers are smooth and straight showing no sign of disturbance. Some of the lowermost layers involved in the disturbance had evidently become partly solidified before the disturbance so that they break into slabs and blocks, while the upper layers of the disturbed sequence behave in a plastic manner (Fig. 27). It can be concluded that the disturbances took place in the crystal mush on the floor before it solidified. Below the mush the floor was almost completely crystalline and so was not deformed. Above, magma rested on the mush at the time of disturbance so that subsequently deposited layers lie upon the disturbed ones unconformably and are themselves not disturbed. The lowermost

disturbed layers have crystallized sufficiently so that they may bend to a certain extent but may also break in a brittle manner while the upper layers are still mushy and deform plastically.

In all the areas of disturbed layering in this vicinity, the overturning of folds and the thrust faulting of the more solid layers appear to be toward the east or northeast.

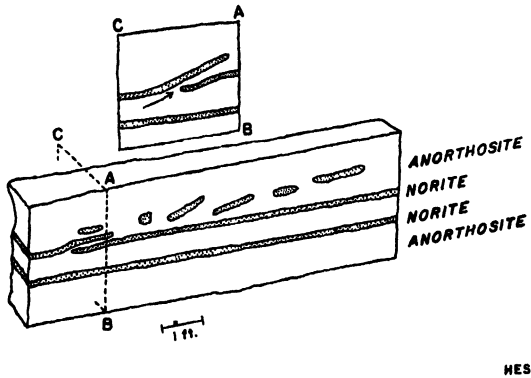


FIGURE 27.—*Disturbed layering, west side of Stillwater valley, Norite Zone*
One layer shows a brittle type of fracture and overthrust

It is suggested that the deformation of the layering is a result of slumping or "landsliding" of the crystal mush over the solid, crystallized floor. The movement may have been brought about by sagging of the floor of the lopolith during crystallization, or the crystals may have been deposited on a gentle slope and the sliding may have been started by an earthquake. The fact that the direction of motion seems to have been the same in all examples of disturbed layering in this particular area suggests a regional tilt of the floor.

Another type of structure formed in disturbed sequences is shown in Figure 28. This structure is apparently an upsqueeze produced by sliding of the crystal mush and is in many respects analogous to salt-dome structures in sediments.

A well-defined overthrust in the lower layers of a disturbed sequence is shown in Figure 27. Sufficient interstitial liquid seems to have remained below the thrust layer to permit sliding, but the thrust layer itself seems to have been quite solid and capable of breaking in a brittle manner.

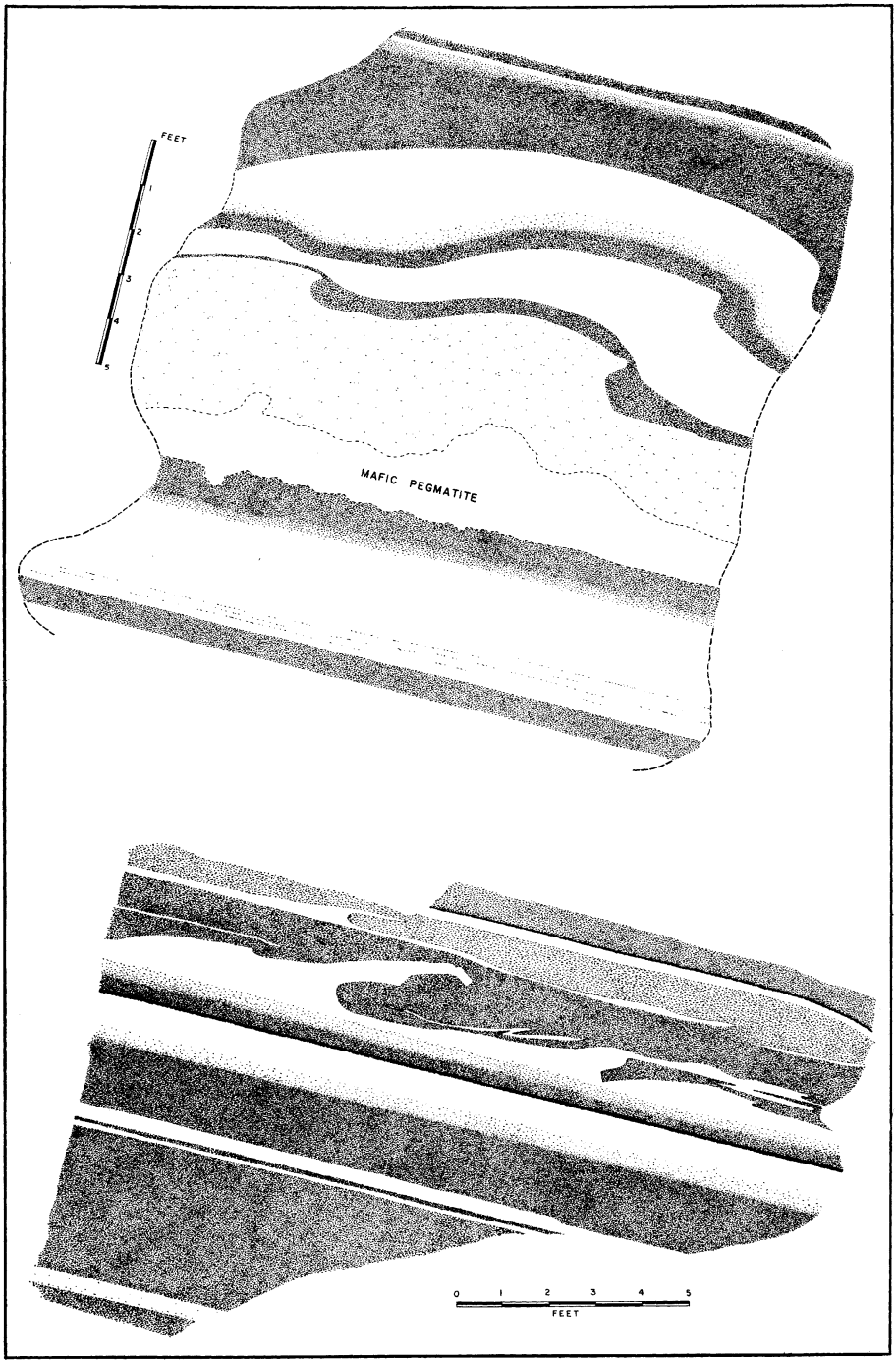
The following conclusions may be drawn from the data on disturbed layering:

The disturbances take place in the crystal mush on the floor of the magma chamber.

The total thickness of crystal mush normally on the floor during the period of crystallization is perhaps 5 feet, and the observed variation is from about 3 to 10 feet.

The mush increases in crystallinity downward so that the lower portion may bend and even break in blocks while the upper portion deforms plastically.

On the west side of the Stillwater Valley the overturning of folds and direction of overthrusting in disturbed sequences are toward the east or northeast, indicating a regional dip of the floor in those directions.



DISTURBED LAYERING, WEST SIDE OF STILLWATER VALLEY, NORITE ZONE

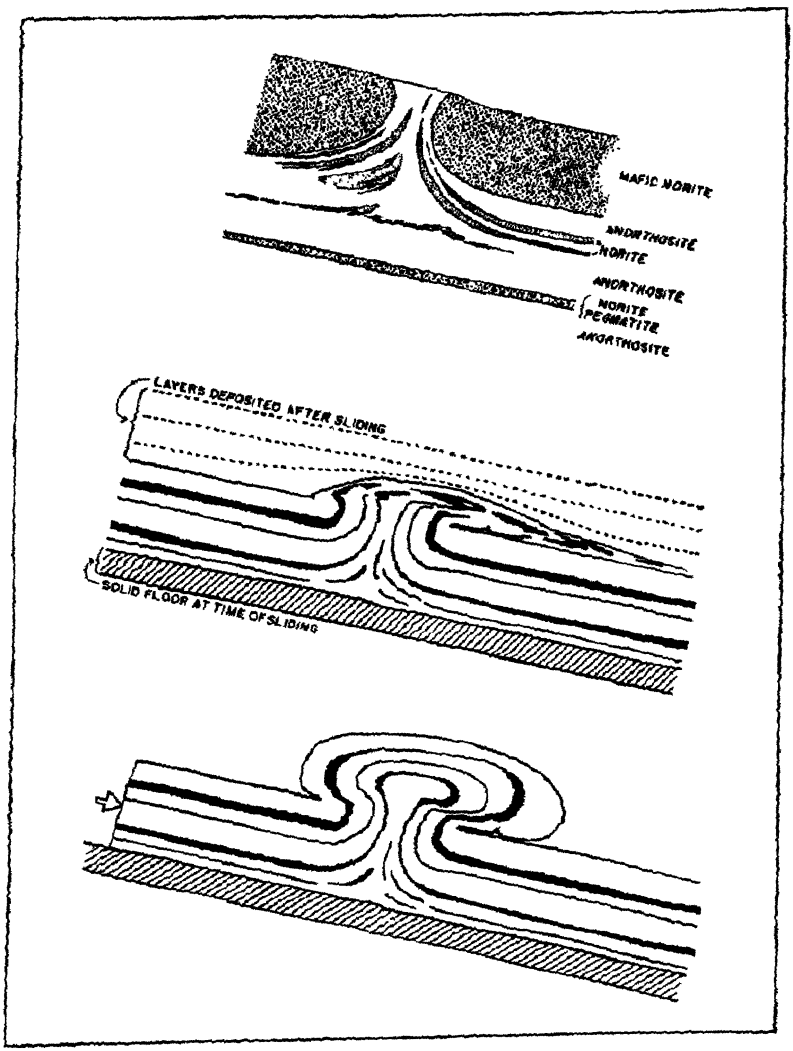


FIGURE 28.—*Upsqueezed structure formed in crystal mush.*
Lower drawings show hypothetical origin of the actual structure in the top drawing

The sliding may be due to increase of dip as a result of sagging of the floor of the lopolith or, starting with an initial dip, the sliding may have been set off by an earthquake or tectonic disturbance.

4. CRYPTIC LAYERING

Cryptic layering, a term coined by Wager and Deer, is layering resulting from changes in the solid phases separating from the magma as cooling and crystallization

proceed. By their definition it may include the gradual change in composition of a solid phase with continued fractional crystallization or it may be applied to the abrupt appearance or disappearance of a particular solid phase as crystallization progresses. The latter has been called "phase layering" in this report and is discussed separately. Cryptic layering in the limited sense of this report is shown by the gradual changes in composition of the plagioclase, pyroxene, and other minerals as one proceeds upward through the complex and is considered in Chapter VI dealing with mineral-variation curves.

5. PHASE LAYERING

The abrupt appearance of augite as a precipitated phase marking the base of the Lower Gabbro Zone is a good example of phase layering. So also is the abrupt appearance and disappearance of chromite as a precipitated phase at the chromite horizon. The term might also be used to designate a particular horizon at which an abrupt change in the compositions of all solid phases occurs as a result of the addition of undifferentiated magma to the main body of magma during crystallization. No such horizon has been recognized in the Stillwater complex, but it seems likely that such might explain the Merensky Reef of the Bushveld complex.

Table 41 shows the appearance and disappearance of various solid phases as crystals accumulating on the floor. In general this appearance and disappearance is one of the main bases for delimiting the zones of the complex as designated in this report.

The main chromite horizon in the Ultramafic Zone is an excellent example of phase layering. In this zone after a period of accumulation of olivine crystals on the floor, the magma reached its saturation point for a spinel, chromite. A flood of minute chromite octahedra was relatively suddenly precipitated from the magma and formed a chromite layer several feet thick on the floor. This layer is practically continuous for the whole length of the exposed complex (Peoples and Howland, 1940) and probably would be continuous and uniform were it not for vicissitudes of crystal sedimentation on the floor. For the same reason it is not everywhere a solid layer of chromite but may be disseminated through a much thicker layer admixed with olivine and later crystallizing bronzite. Although most of the visible chromite is contained in the main chromite horizon, sporadic occurrences of smaller chromite accumulations are found below and above. Not far above the main chromite horizon, chromite disappears as a solid phase separating from the magma and does not reappear.

The rhythmic layers are flat lenses whose lateral extent is equivalent to several hundred times their thickness. Phase layers are far more persistent; normally they extend through the whole lopolith. In the Bushveld, individual chromite layers or sequences of layers are persistent for very long distances—on the order of miles for a layer only a few inches thick. If the groups of chromite layers are considered as a unit they persist through the entire complex. Since great lopoliths have an areal extent measured in tens of thousands of square miles, it is unlikely that all parts of the vast magma chamber reach the stage of precipitation of chromite, for example, at exactly the same time. The time of beginning of precipitation in the complex is different from place to place. Thus the layer containing the chromite behaves exactly as would

a sedimentary rock "formation". The lithologic unit is persistent, but isochronous planes cut across it at a low angle.

In the case of phase layering resulting from addition of undifferentiated magma after partial crystallization of the original magma (*i.e.*, the Merensky Reef of the Bushveld), the layer formed, although very thin, may persist over the whole area of

TABLE 41.—*Phase layering resulting from appearance or disappearance of solid phases accumulating on the floor during crystallization*

	Ultramafic Zone	Norite Zone	Lower Gabbro Zone	Anorthosite Zone	Upper Gabbro Zone	Hidden Zone
Olivine	———			— — — —		
Chromite	—					
Orthopyroxene	——	———	———	— —	———	
Pigeonite					↓ ——	——→
Augite			———	— —	———	——→
Plagioclase		———	———	———	———	——→

the complex. Such a layer would be the analog of a bentonite bed deposited as the result of a single volcanic eruption. This type of phase layer would be everywhere the same age. Phase layers of the first type might well be expected to show a gradual transgressive relationship to the second type. Unless there are two Merensky reefs in the Bushveld, which seems unlikely, Fuller (1951, M.Sc. thesis, Univ. Capetown) finds such transgression in the Swartkopf district in the Northwestern Bushveld Here the Merensky Reef lies above titanomagnetite horizons, whereas along the southern and eastern margins of the complex the Merensky Reef is well below the magnetite layers. This probably means that the magma chamber was thinner toward the northwest and crystallization differentiation was more advanced there when the magma addition producing the Merensky Reef took place.

6. INCH-SCALE LAYERING

At several places in the complex in the Lower Gabbro Zone is a type of layering characterized by thin layers about five-sixths of an inch thick which may be repeated hundreds of times. It consists of alternations of pyroxene and plagioclase (one such pair is here defined as a layer). Figure 1 of Plate 8 shows the appearance of this layering in the field, and Figure 2 of Plate 8 is a closer view of a specimen of it. The origin of this layering is at present obscure, but it suggests growth of crystals on the floor in areas where crystals were temporarily not being contributed by settling from above. The regular repetition of layers remains unaccounted for.

B. ORIGIN OF RHYTHMIC LAYERING

The writer has previously advanced a theory concerning the origin of rhythmic layering (Hess, 1938b) in which most of the pertinent literature was briefly reviewed. The present discussion summarizes the theory and elaborates certain points.

To produce a rhythmic *repetition* of layers, such as observed in the Stillwater complex and in many other mafic intrusives, some physical condition must be modified

to interrupt the otherwise uniform rate of crystallization, settling, and accumulation of crystals. In a huge mass of magma such as is represented by the Stillwater, it is unlikely that composition, pressure, or temperature conditions could be changed rapidly and repeatedly to produce this effect. Furthermore if these conditions had been changed, the compositions of the crystals separating from the magma would have been affected, which, as shown in Chapter VI, is not the case.

There remains one variable capable of producing rhythmic layering—motion in the magma. If motion in the magma is not constant in velocity and direction, rhythmic layering will result, provided only that the two or more crystalline phases involved have considerably different settling velocities.

A number of factors can periodically produce motion in the magma. Loss of heat by the magma through the roof will result in crystallization, and, as Wager and Deer point out, a body of magma containing crystals will be heavier than the underlying magma without crystals. This will produce convective overturn. Also as Wager and Deer suggest the velocity of the sinking column with crystals will increase with its downward progress as a result of growth of the crystals favored by increasing pressure. Slight changes of shape of the magma chamber resulting from sagging of the floor or other tectonic forces will also result in motion in the magma. In fact it would be extremely unlikely that the magma could remain motionless or that the motion could be uniform (constant in direction and velocity).

In the consideration of the mineral-variation curves in Chapter VI and in particular in the discussion of the Dead Tree section, it is noted that megacycles in the mineral variations represented on the floor by thicknesses of crystal accumulates of the order of magnitude of 1000 feet possibly indicate one complete convective overturn of the magma. The smaller cycles, of which there are hundreds in this thickness, probably represent minor currents in the magma immediately over the floor perhaps within a few hundred feet of it. Most of the pronounced rhythmic layering is probably related to these minor currents. In the discussion of rhythmic layering it was shown that intervals of layered sequences are separated by much thicker layers of uniform nonlayered norite. This is interpreted as meaning that in the magma, immediately above the floor, currents are present for short intervals of time and for longer intervening intervals are quiescent.

Let us assume that crystals of two types are settling in a magma and that one type settles faster because of greater density or size. For convenience the faster settling crystals will be called "F" crystals, and the slower settling "S" crystals. We may now consider the general effect of any change in motion of the magma on such a system.

If the magma is moving so that the motion has an upward vertical component, it will slow the settling of the S crystals relatively more than the F crystals. For example if the F crystals have a settling velocity of 5 and the S crystals of 3 and the upward component of the velocity of the current is 2, the resulting combined velocities are $5 - 2 = 3$ for the F crystals and $3 - 2 = 1$ for the S crystals. Thus the current has changed the ratios of settling velocities from 5:3 to 3:1. The F crystals originally settled not quite twice as fast as the S crystals, but now they settle 3 times as fast. This change will at first increase the proportion of F crystals to S crystals arriving at the bottom, followed by a continuation of the arrival of F and S crystals in the

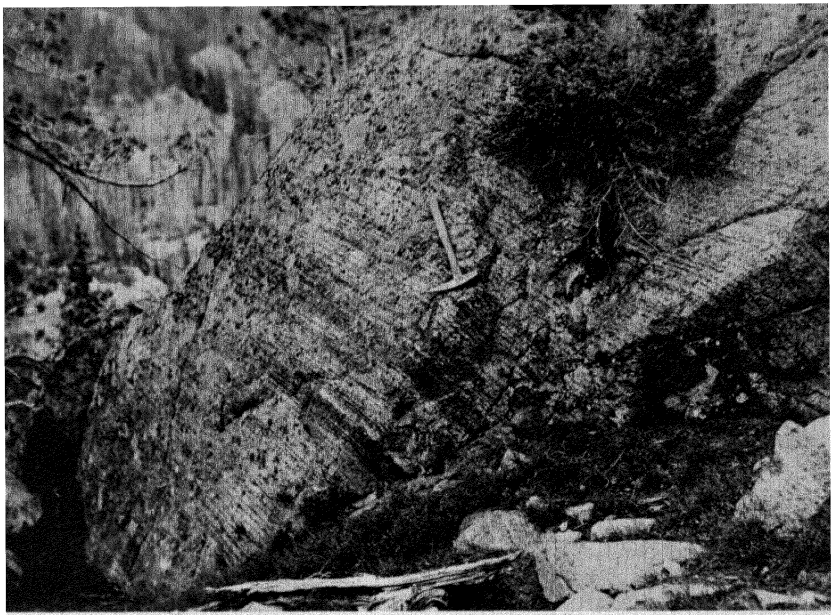


FIGURE 1. — Outcrops of inch-scale layering northwest of Cliff station, East Boulder Valley

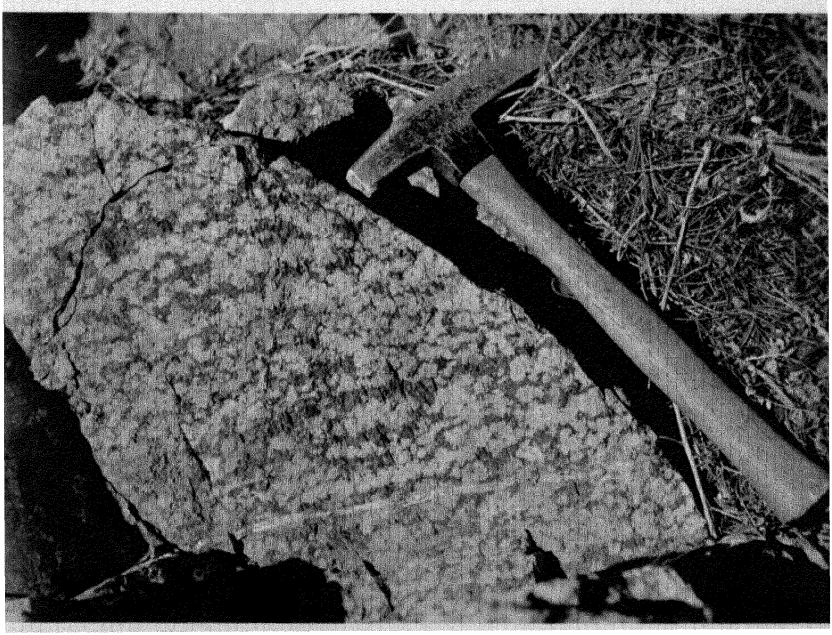


FIGURE 2. — Close view of a specimen showing inch-scale layering

same ratio as before the current started, provided the current persists with constant velocity. If the current slows or stops, the proportion of S crystals accumulating on the floor will increase, followed after a short interval by return to the normal ratio of arrival on the floor of F and S crystals. A current with a downward vertical component can cause very much less effect because it produces very much less change in the relative velocities of the two groups of crystals. For a downward velocity of 2 the F crystals have a net velocity of $5 + 2 = 7$ and the S crystals $3 + 2 = 5$ or a change in ratio of velocities from 5:3 to 7:5 which obviously can produce much less striking results than the current with an upward velocity of 2. The effect of a downward current on the arrival of crystals at the bottom would be to cause a slight increase of the S crystals at first followed by return to the normal ratio of arrival. Stopping or slowing of the downward current would result in an initial increase in the F crystals followed by return to normal. No doubt the observed cases of inverted gravity stratification are related to downward currents, but the fact that such inverted gravity stratified layers are less common and much less well developed than the normal gravity stratified ones is to be correlated with the comparative ineffectiveness of a downward current in changing the relative settling velocities of the F and S crystals. In most cases the results of a downward current are probably too slight to be readily observed in the field.

The mechanics of rhythmic layering are discussed in a series of examples assuming that pyroxene and plagioclase crystals of the same size are settling in the magma and that the relative settling rates are disturbed by motions in the magma. The reader is referred to Chapter VIII for a quantitative evaluation of settling rates with respect to densities, sizes of crystals, and viscosity of the magma. Assuming some motion in the magma—intermittent and perhaps varying in velocity—the following series of cases illustrate the results to be expected.

At *A* in Figure 29 (stage 1) plagioclase and pyroxene are entering in the ratio of their precipitation from magma. Plagioclase and pyroxene arrive at *B* in exactly the same ratio. Pyroxene settles much more rapidly so that pyroxene crystals are separated by greater intervals in the figure. Since they take less time to reach the floor there will be fewer suspended at any given instant.

In the special (limiting) case of an upward current of velocity Vc, equal to the settling velocity Vpc of plagioclase, the relations shown in stage 2 will result. Plagioclase will remain stationary, none will reach the floor, but newly crystallized plagioclase will pile up at *A*. Pyroxene will continue to pile up on the floor at first at a slightly decreased rate and then at the normal rate. The total number of crystals arriving on the floor will of course be halved (assuming plagioclase and pyroxene are separating at a 1:1 ratio) since no plagioclase will be arriving. Hence the net rate of accumulation will be slow.

With these conditions maintained a pyroxenite layer would be formed on the floor. If the upward current died out the relations in stage 3 would result. The relationship of pyroxene to plagioclase in the upper part of the figure is the same as in stage 1 and in the lower part the same as in the upper half of stage 2. At *B* on the floor the accumulation of pyroxene will be somewhat faster than normal, and the rate of accumulation of plagioclase will be normal until *D* reaches the bottom. The crystals

from *D* will arrive as a sudden deluge of plagioclase producing an anorthosite layer. The distance between the pyroxenite formed in stage 2 and the anorthosite of stage 3 will depend on the length of time the upward current is in motion. The layers be-

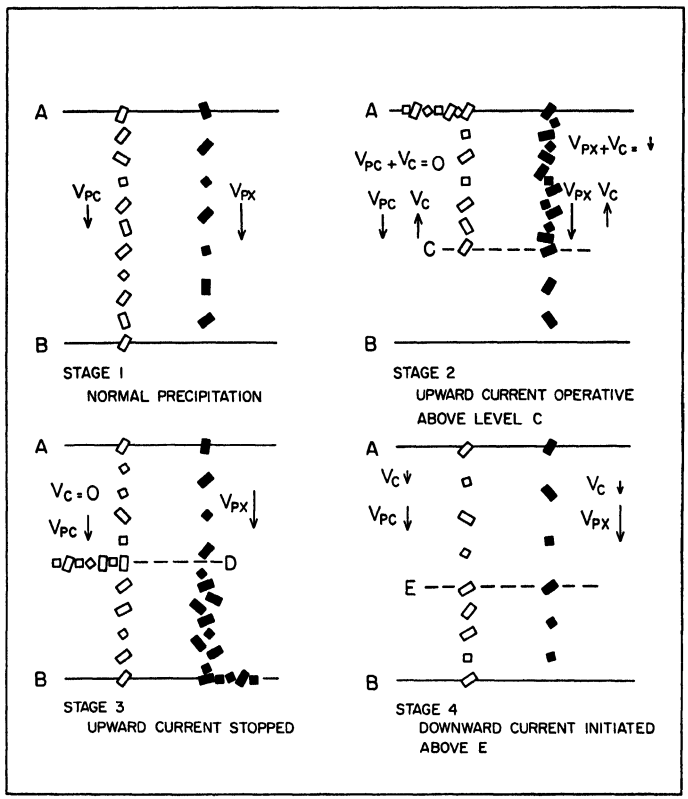


FIGURE 29.—*Effect of currents with a vertical component on crystals settling at different velocities*

tween the pyroxenite and anorthosite will accumulate somewhat more rapidly than normal and will be abnormally pyroxenic. The anorthosite layer probably would be deposited considerably more rapidly than normal. After its deposition the rate of accumulation and ratio of pyroxene to plagioclase arriving at the bottom would again be normal.

A downward current would produce much less spectacular results. It would at first cause an increase in the rate of accumulation on the floor and an increase in plagioclase with respect to pyroxene. Both these effects would cease when *E*, the arbitrary locus of input of crystals at the instant the current started, reaches the floor (stage 4). From this instant, the rate of accumulation of crystals and ratio of plagioclase to pyroxene would be normal. Stopping of this current would reverse the

process producing temporary slow accumulation and a relative increase in pyroxene until both again return to normal.

In the example of an upward current for simplicity a special case was chosen—namely, one in which the upward velocity of the current equalled the settling velocity of the plagioclase. It can readily be seen that any upward current regardless of its velocity would produce analogous results. A velocity less than the settling velocity of plagioclase would, however, produce no sharply defined pyroxenite and anorthosite. Any velocity greater than the settling velocity of plagioclase would produce qualitatively the same results as in the case in which the velocity was equal. The quantities of pyroxenite and anorthosite produced would be greater for greater lengths of time of operation.

Constant conditions such as no currents in the magma or currents in any direction of constant velocity after an initial stage will result in deposition on the floor of crystals in the proportion from which they separate from the melt, provided the upward vertical component of velocity of the current is less than the settling velocity of the plagioclase. Starting of a current or any change of the velocity of a current with a vertical component will result in disturbing the pyroxene:plagioclase ratio of the crystals accumulating.

Chapter VIII

Some Quantitative Aspects of Stillwater Petrology: Settling Rates, Thermal Relations, Diffusion, Convection

A. SETTLING VELOCITIES OF CRYSTALS IN MAGMA

Settling velocities of crystals in a magma may be estimated from Stokes' law for settling of spheres. The crystals in the Stillwater magma were not spheres, but neither were they conspicuously platy or rodlike. At the low velocities to be expected this probably does not introduce an appreciable error compared to the uncertainties in other quantities as, for example, viscosity.

$$V = \frac{2}{9} g r^2 \frac{\rho_c - \rho_m}{\eta}$$

where g = gravity constant, r = radius in cm, η = viscosity in poises, ρ_m = density of magma, ρ_c = density of crystals, and V = settling velocity in cm per sec. (Multiply by 31.5×10^4 to change to meters per year).

Rubey (1933) noted that settling of quartz in water obeyed Stokes' law provided the grain size was less than .02 mm. On theoretical grounds he set up an equation to cover settling of grains of all sizes up to coarse gravel, and confirmed his equation experimentally:

$$V = \frac{\sqrt{4/3 g \rho_m (\rho_c - \rho_m) r^3 + 9\eta^2} - 3\eta}{\rho_m r}.$$

In Chapter IV the density of the original magma was estimated at 2.65 g/cc at 1125°C., and the density after separation of the Ultramafic Zone at 2.58 g/cc at the same temperature. The latter value is used in the calculations inasmuch as it is particularly desired to apply the results to the discussion of rhythmic layering in the Norite Zone. Curves were constructed using a range of reasonable values for r, ρ_m, and ρ_c. For settling rates less than 1 cm/sec and viscosity greater than 100 poises Stokes' law gives exactly the same results as Rubey's equation.

Data available on the viscosity of basalt magma permit a considerable latitude in choice of a proper value. By Jeffrey's formula Nichols (1939) calculated viscosity of two Hawaiian lava flows from their velocity of flow, physical properties, dimensions, and floor gradients. For the Alika flow he obtained a viscosity of 43,000 poises and for the 1887 flow 47,700 poises. While these may be valid estimates for a lava flow, Minakami's (1951) results on the basaltic lava of Oo-sima indicate a much lower figure before eruption (Table 42). The exponential increase in viscosity with decreasing temperature and increasing distance from the crater is evident.

Kani (1934) and Volarovich, Tolstoy, and Korčemkin (1936) measured viscosities of fused basalts in the laboratory (Table 43). In each case the viscosity in the vicinity of 1200°C. is comparable. Minakami's value at 1125°C, is about comparable to fused basalts at 50°C. higher temperature. The difference no doubt has to do with contained volatiles in the natural lava as compared to the fused basalt samples.

If one extrapolates Minakami's values back to some point within the crater a value less than 5600 poises seems probable. Lava flows themselves lose volatiles and begin to solidify; this, together with temperature decrease, accounts for Nichols' high values and Minakami's rapid increase away from the crater rim.

TABLE 42.—*Viscosity, temperature, and distance from crater rim for Oo-sima 1951 eruption*

Distance from crater (meters)	Temperature °C	Viscosity (poises)
10	1125	5,600
—	1108	18,000
—	1083	71,000
400	1038	230,000

TABLE 43.—*Viscosity measuremetns on fused basalts*

Olivine basalt (Kani)		Basalt (Volarovich, Tolstoy, and Korčemkin)	
°C	poises	°C	poises
1400	137	1400	353
1350	176	1340	650
1300	296	1320	820
1250	656	1280	1340
1200	3180	1220	2960
1150	37900	1160	6820
		1160	13600 crystallizing

From the above figures the best guess is that basaltic magma with 1–2 per cent of contained volatiles would have a viscosity comparable to dry basalt melts at a temperature 100°C. higher, or approximately 3000 poises. The viscosity for the Stillwater magma at 1100°C.–1125°C. almost certainly must be in the range 300 to 30,000 poises, and the most probable value is considered to be 3000 poises. Settling velocities for plagioclase, pyroxene, and chromite in this viscosity range are given in Table 44. Figure 30 shows settling velocities at 3000 poises for the same minerals. By use of the correction scale, values for other desired viscosities can be read. The table and figure 30 bring out clearly a point which is also obvious from inspection of Stokes' equation—namely, that settling velocities vary as the *square* of the radius but only as the first power of density difference or viscosity. Thus a chromite with a ¼-mm radius settles more slowly than a plagioclase with a 1-mm radius, or a pyroxene with a ½-mm radius settles more slowly than a plagioclase with a 1½-mm radius. If the relative sizes of the plagioclase and pyroxene were suitable one might get rhythmic layering in which plagioclase at the bottom of the layer grades upward into pyroxene. Size sorting is probably more common in layered rocks than has been appreciated inasmuch as it is relatively inconspicuous and difficult to observe in the field (Cornwall, 1951). It was not looked for during the field work on the Stillwater nor was it seen. It has since been noted in some of the sequences of specimens col-

TABLE 44.—*Settling velocities of crystals in meters per year*

Magma density 2.58

	Plagioclase				Pyroxene				Chromite				
	0.1				0.7				1.4				Density crystals less density magma
Viscosity in Poises ↓	0.5	1.0	1.5	2.0	0.1	0.5	1.0	2.0	0.1	0.25	0.5	1.0	Radius in mm
300	57	229	515	916	16	400	1602	6408	32	200	801	3204	
1000	17	69	153	274	4.8	120	481	1922	9.6	60	240	961	
3000	5.7	23	52	92	1.6	40	160	641	3.2	20	80	320	Preferred values
10000	1.7	6.9	15	27	.48	12	48	192	.96	6	24	96	
30000	.57	2.3	5.2	9.2	.16	4	16	64	.32	2	8	32	

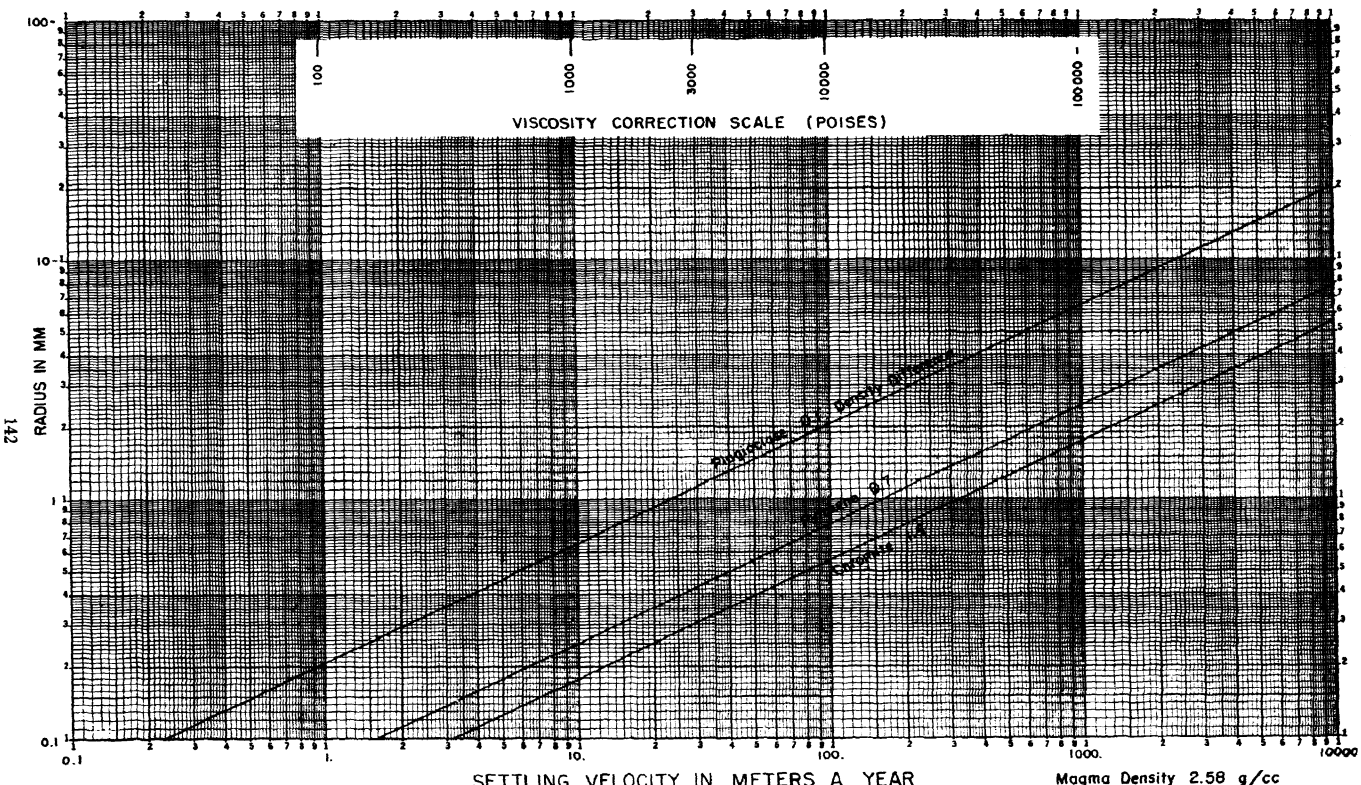


FIGURE 30.—*Settling velocity of spheres in magma with viscosity 3000 poises*

Upper curve for plagioclase, 0.1 density difference; middle curve for pyroxene, 0.7 density difference; lower curve for chromite, 1.4 density difference.

lected for laboratory study. In the Norite Zone where rhythmic layering is most pronounced density and grain size are working to cause maximum differential sorting because the orthopyroxene crystals are invariably larger than the associated plagioclase. For 25 norites from this part of the complex the average grain intercept[11] from traverses of thin sections is 1.14 mm for the orthopyroxene and 0.61 mm for the plagioclase. This would mean a settling velocity for the pyroxene 25 times that of the lighter and smaller plagioclase, if the density difference (crystals–magma) is 7 times as great for pyroxene as for plagioclase. In the upper half of the Stillwater complex rhythmic layering is virtually absent, but the plagioclase crystals are larger here and the pyroxene smaller than in the Norite Zone.

The igneous lamination in the upper portion of the complex is well developed, which is clear evidence that the various phases settled from the magma, but apparently their settling velocities were such that little differential sorting took place. Rhythmic layering in the Bushveld and Skaergaard is also limited to their lower portions, presumably for the same reason. The rather common association of rhythmic layering with norites or troctolites in which either orthopyroxene or olivine is the only ferromagnesian phase that settles suggests the following hypothesis. If only one ferromagnesian phase is separating from the melt it forms relatively large crystals. If, however, two ferromagnesian phases are separating roughly twice as many nucleii of crystals of ferromagnesian phases form, and they compete for the common ions such as Fe^{+2} and Mg^{+2}. For the same content of ferromagnesian minerals this means twice as many crystals each of half the volume of those in the monoferromagnesian type of crystallization. It is rather difficult to test this idea by referring to the literature because commonly the peri-euhedral ferromagnesian phases which have settled from the magma are not distinguished from anhedral ones which have presumably crystallized from interstitial liquid afterward.

Referring again to Stokes' law, the settling velocity varies inversely as the viscosity. Viscosity has no effect on the *relative* rates of settling of different sorts of crystals suspended in a fluid (within the viscosity range here under discussion and for velocities of less than 1 cm per sec.). If the viscosity were too high, the settling velocities might be so low that there would be little settling in the time available for crystallization (of the order of magnitude of 10^5 years). Viscosities greater than 30,000 poises would certainly fall into this category, and it would seem rather difficult to effect settling of plagioclase crystals of 0.6 mm radius at a viscosity as high as 10,000 poises even if convection is assumed to accomplish most of the downward transfer of the crystals.

Field evidence suggests that single crystals may not be the units settling, but that perhaps it is, at least in some cases, glomeroporphyritic aggregates of several similar crystals. Some gabbros have a spotted appearance caused by clusters of several plagioclase crystals. The small chromite crystals, ¼ mm or less, would have such a slow settling rate that it is surprising that they can form a sharp gravity stratified layer with the chromite at the bottom. These chromite crystals are linked together in chains, sometimes branching, by the octahedral chromite crystals touching point

[11] Average grain intercept gives a figure which seems to be close to the mean radius of the crystals, excluding, of course, the poikilitic or irregular interstitial grains.

to point. The separated crystals if placed in a liquid such as glycerine also have a strong tendency to line up in such chains. Perhaps they settled in the magma linked in chains. A detailed investigation of size sorting must precede any specific conclusion as to the possibility of the settling of glomeroporphyritic units.

B. THERMAL RELATIONS

The orthopyroxene defines the temperature range in which the exposed rocks of the Stillwater complex crystallized. The initial crystallization of bronzite took place below the temperature of the clinoenstatite-enstatite inversion. Bowen and Schairer's (1935) data indicate that the inversion temperature for a pyroxene with an Mg:Fe ratio of 85:15 is about 1130°C. The actual temperature of crystallization of the early Stillwater bronzite must have been very close to the inversion curve, or approximately 1125°C. Near the top of the exposed complex the Mg:Fe ratio reaches 70:30, and the orthopyroxene changes from the "Bushveld type" to the "Stillwater type" with coarse augite plates parallel to the {001} plane of a monoclinic pyroxene crystal. The latter type of orthopyroxene is clearly derived from pigeonite by inversion (Hess, 1941; Edwards, 1942; Chapter III of this report). The inversion temperature for this composition is approximately 1100°C.; thus the magma temperature must have crossed the inversion curve at approximately this point in the crystallization of the complex. Judging from the work of Wager and Deer (1939), the remainder probably crystallized between 1100°C. and 955°C.; in this range in the similar Skaergaard intrusion a temperature at 98½ per cent crystallized was obtained from a wollastonite–hedenbergite inversion.

Some very interesting quantitative conclusions could be drawn if the rate of heat loss during crystallization and hence the time required for crystallization could be estimated accurately. Approximations of the rates of settling of crystals have already been made. A thickness of something between 1 and 3 m for the crystal mush on the floor at any given time is probable. If one can estimate how long this mush remained in contact with the overlying magma, the effectiveness of ionic diffusion in modifying the pore-space liquid of the mush can be evaluated. This is an important factor in the differentiation of the complex and in the explanation of many petrographic details.

To solve the problem of the rate of heat loss from the magma and hence the rate of accumulation of crystals on the floor the effects of several factors must be evaluated. These include the thickness of the cover over the complex, convection in the magma, superheat, the partition of heat loss to the roof and floor as the floor moves upward with accumulation of crystals, conductivity of the floor rocks at the high temperatures obtaining, etc. A fairly rigorous treatment of these factors is being made by G. B. Hess and the writer which will be published in the near future. A rough approximation is presented here based on Lovering's (1935) curves.

The thickness of the roof over the Stillwater complex is not known. Roof rocks are exposed in outcrops along the northern side of the East Boulder valley north of the complex. An estimate of at least 1 km thickness seems a reasonable minimum. Computations show that the rate of heat loss with a 100-meter roof will be 10 times

as fast as for a 1-km roof, but increasing thickness above 1 km has a comparatively smaller effect.

Convection will occur in the magma. This can be taken care of by greatly increasing the conductivity of the magma in making the computations. In the estimate made below this was accomplished by putting in a higher value for diffusivity.

Inasmuch as we are interested only in the rate of heat loss during the interval from the start of crystallization to 60 per cent crystallized, certain simplifying assumptions may be made which permit a fairly satisfactory answer directly from curves published by Lovering (1935). The thermal curves which he gives for the lava flow fit the case in hand reasonably well, inasmuch as the cover postulated is only as thick as the sill. After a short interval of adjustment to remove the extreme differences in temperature between the wall rocks and the magma immediately after intrusion, it is assumed that there will follow a long period during which the rate of heat flow from the intrusion into the wall rocks will reach a steady state. After this steady rate of heat flow has been established, the dominant factor will be the ability of the wall rocks to conduct this heat away (their diffusivity). Thus it seems possible to estimate the difference in time between two closely spaced incidents in the cooling history with somewhat greater accuracy than the absolute position of these points in time from the instant of intrusion.

Constants used in computing heat loss

Thermal constants estimated from Lovering (1935), Birch *et al.* (1942), and Daly (1933).

d = 8200 meters ..... Thickness of sheet
h^2 = .008 ..... Diffusivity of magma and wall rocks
t_1 = 1225°C..... Initial temperature of magma
t_w = 25°C ..... Initial temperature of wall rocks
t_2 = 1125°C ..... Temperature at start of crystallization
t_3 = 1100°C ..... Temperature after 60% of complex crystallized
s = 0.3 cal/gm °C. ..... Specific heat of basalt
L = 90 cal/gm..... Latent heat of crystallization of basalt.

1. Heat loss per gram in cooling 1225°C, to 1125°C
$$25 \times 0.3 = 30 \text{ cal}$$
2. Heat loss per gram in cooling 1125°C to 1100°C
$$25 \times 0.3 = 7.5 \text{ cal}$$
3. Latent heat released 1125°C. to 1100°C. (60% of magma crystallized)
$$90 \times .60 = 54 \text{ cal/gm}$$
4. Total heat loss 1225°C. to 75°C.
$$1150 \times 0.3 + 90 = 435 \text{ cal/gm}$$
5. Per cent heat loss 1225°C. to 1125°C.
$$30 = 7\%$$
6. Per cent heat loss 1225°C. to 1100°C.
$$\frac{7.5 + 54 + 30}{435} = 21\%$$

Note: From 1125° to 1100°C. 88% of heat loss is attributable to the latent heat of crystallization.

Using Lovering's curves for the lava flow a number of points (x_1, x_2, x_3 $\cdots$) were chosen at regular intervals from the roof to the floor. For each point the $2h/t$ values

for .93 (7% heat loss) and .79 (21% heat loss) were determined. These figures were then substituted in Lovering's Figure 1 using $d = 8200$ m and various values for the time interval in years were found. The results were then multiplied by 1.5 to correct for the diffusivity assumed. Averaging the resulting values gives 50,000 years for the time required for 60 per cent of the complex to crystallize. Thus the average rate of accumulation of crystal mush on the floor would be of the order of magnitude of 3 years for 1 foot (10 cm/yr).

Some petrologists might object to the assumption of superheat. To the present writer the presence of a moderate amount of superheat seems reasonable. If a basalt magma were generated at a depth of 100 km, or thereabouts, and at the time of generation were in equilibrium with various crystalline phases at that depth, sudden transfer of the liquid to a depth of 10 km or less should result in a considerable amount of superheat upon arrival in the new environment. Melting temperatures increase 3°C. or more per km of depth. Omitting superheat the rate of heat loss after 20 per cent crystallized would be practically the same as given above.

The partition of heat loss to the roof and floor may be approximated by comparing Lovering's curves for a dike (the same as a sill with infinite cover) and a flow. In this case it is found that the flow loses heat twice as fast as the dike. Thus if the loss to the floor is half the loss from a dike to both walls, the loss upward from the flow is 3 times that to the floor. This ratio nicely fits the Stillwater observations. Heat loss from the roof permits crystallization and deposition on the floor. The crystal accumulate is thought to have 25–29 per cent interstitial liquid. Heat loss to the floor would just take care of crystallization of this interstitial liquid. Field evidence indicates that the mush on the floor maintains a fairly constant thickness. This means that the bottom of the mush moves upward at the same rate on the average as does the top of the mush.

The above treatment of the heat-flow problem has many obvious weaknesses, but the more precise calculations now underway are substantially in agreement with the results here presented.

C. EFFECTIVENESS OF DIFFUSION

In the discussion of disturbed layering it was pointed out that the thickness of mush on the floor ranges from about 1 to 3 m. It is here assumed that a slow rate of crystal sedimentation results in a mush thickness of 1 m, an average rate 2 m, and a fast rate 3 m.

The conclusion drawn from estimating the rate of heat loss during crystallization was that 10 cm per year of crystal mush would be deposited on the floor. Taking the average thickness of the mush as 2 m, the bottom of this layer of mush would have had about 20 years in contact with pore-space liquid at depths ranging from zero at the start to 2 m after 20 years. At the slower rate of accumulation 1 m of mush would have been in contact with the liquid for 20 years, and at the fast rate 3 m. This is not strictly true because where the mush was thinner, and hence diffusion more effective, the pore space would be diminished by faster crystallization, and *vice versa* where the mush was thicker. The rate of heat loss from the floor would remain a constant, and just as much crystallization per unit time would take place

in the mush from this cause regardless of thickness. Nevertheless the 1- and 3-m figures and 20-year time interval can be used for an approximation.

The only pertinent experimental work on diffusion in silicate melts is that carried out many years ago by Bowen (1921). These measurements are for the rate of diffu-

TABLE 45.—*Estimated effectiveness of diffusion* (k = .25)

1 Time in years	2 Maximum limit affected by diffusion	3 Strongly affected, 1/2 saturation	4 90 per cent effective
⅔	33 cm	7 cm	1.1 cm
1	41	8.8	1.4
4	82	17.5	2.8
9	123	26.3	4.0
16	165	35	5.5
20	183	39	5.7
64	330	70	11

TABLE 46.—*Estimated effectiveness of diffusion in Stillwater mush*

Time in years	Strongly affected 50 per cent saturation	90 per cent effective
1	1.6 meters	.25 meters
4	3.2	.50
9	4.8	.75
16	6.4	1.00
25	8.0	1.25
64	12.8	2.00

sion of diopside melts into plagioclase melts of various compositions. The figures for diffusion of diopside into labradorite (An_2Ab_1) represent a fair analogy to the case at hand as calcic plagioclase and pyroxene are the main constituents of both the crystal mush and the surrounding liquid. For "average diffusivity" of diopside into labradorite Bowen chose the value k = .25 per cm² per day and with this figure estimated the rates at which certain geological processes might be expected to proceed. The calculations are based on a probability integral derived from Fick's law. The figures in Table 45 were picked from the curves given in Bowen's Figure 7.

Table 45 shows that a meter in 20 years is at least within the range at which diffusion can operate. The effectiveness dies out very rapidly with distance, and to double the effective distance one must square the time. If it were assumed that diffusion were 90 per cent effective for a distance of 1 m in the Stillwater mush, then at 3 m the effectiveness would drop to about 60 per cent, which is approximately what the petrographic evidence indicates. The specific figures taken from Bowen's data are not strictly analogous to the conditions in the Stillwater mush. Slow circulation of the magma could maintain a constant composition above the mush.

Crystallization of pore liquid from the bottom of the mush upward would force some residual liquid upward as the pore space diminished. To produce crystallization of plagioclase, for example, in the pore space it is not necessary to diffuse all the constituents into the space but largely to change the ratio of Na to Ca by about 20 per cent. Finally the top of the mush layer is moving upward so that diffusion is always relatively effective in the first few cm. One would have to integrate the effect of diffusion from the starting condition of zero time and zero depth to a final condition of X time and Y depth. This calculation would be further complicated by the rate of crystallization. If no crystallization occurred there would be no composition gradient, or conversely the greater the rate of crystallization the steeper the gradient and hence the higher the rate of diffusion. Considering all these factors qualitatively an estimate of the effectiveness of diffusion in the Stillwater mush is made in Table 46. The discrepancy between Tables 45 and 46 cannot be reconciled satisfactorily at the present time. Perhaps diffusion rates are faster for the conditions of the Stillwater problem. Perhaps the cover on the Stillwater was much thicker than postulated so that the rate of accumulation of the mush on the floor was slower. At any rate doubling or tripling the thickness of mush on the floor by more rapid accumulation of crystals would have a marked result in the effectiveness of diffusion and could account for the observed fluctuations in the composition curves of the constituent minerals.

C. CONVECTION CURRENTS

Any motion produced in fluids by differences in temperatures, and hence in density, may be called convection. It may be in the form of regular cells or it may be random or irregular in pattern. Convection cells ideally have a horizontal dimension slightly larger than their vertical dimension. Containers of such dimensions favor development of regular cells. The Skaergaard intrusion has a shape well adapted to a regular convection-cell system such as Wager and Deer postulate and demonstrate to have probably existed.

Lopoliths are flat lenses with a horizontal dimension perhaps 20 to 50 times their vertical dimension and are probably slightly convex downward. Conceivably a mass of liquid of this shape could break up into a series of cells side by side horizontally, but it might also behave in a much less regular manner. For example a layer of liquid might develop below the roof which by loss of heat and impregnation by crystals became denser than the underlying liquid. Although mechanically unstable it might develop and remain in this position for a time, just as in winter a layer of colder water sometimes exists metastably for a short time at the surface of the sea. Such a condition would be more readily imaginable in a viscous liquid than in water. Eventually overturn of the unstable mass of liquid will occur. One might expect that some inhomogeneity in the layer would cause it to get started at one point first. A downward bulge would form, the nose of which would accelerate rapidly. The velocity would be enormously greater than the settling of crystals. A single descending column of denser liquid might draw off the whole of the layer from below the roof. It would spread rapidly over the floor and come to rest. Alternatively one might postulate tear-droplike masses starting downward from many points, and at different

times, from the lower surface of the denser liquid. Industrial flotation processes are inverted analogs of such a system. The semiregular spacing of salt domes on the Gulf Coast is another inverted analogy of an unstable less dense layer moving upward.

Inasmuch as no evidence was found for a regular convection-cell system in the Stillwater or Bushveld complexes in contrast to the Skaergaard intrusion, a random plummetting downward of the denser liquid from the vicinity of the roof is postulated. This might well be periodic; a certain degree of instability is required to overcome the cohesiveness and inertia of the denser layer.

After crystallization has started in a large body of magma such as the Stillwater one might expect that for every calorie of heat lost there would be an equivalent amount of crystallization. The layer under the roof would become denser essentially because it contained suspended crystals. Change in temperature and hence change in density of the liquid itself would be exceedingly small during this period. Magma with 1 per cent of suspended pyroxene crystals would be .006 gm/cc denser than magma without crystals. A pyroxene with a mean radius of 1 mm would settle in magma of viscosity 3000 poises, 160 m per year. A spherical mass of magma .006 gm/cc denser than surrounding magma and 1 m in radius would settle at a rate 10,000 times as fast once it reached terminal velocity (5 cm/sec). This is the "two-phase" convection originally suggested by Grout (1918b).

A consideration of the above figures suggests that convection is inevitable in large bodies of basaltic magma.

Appendix II deals with the differentiation of dolerite sills. These differ from lopoliths in absence of layering and igneous lamination. The difference is attributed to a steeper temperature gradient across the contacts of thinner sills such that crystallization takes place directly on the contacts and adhering to them while the liquid circulates by pure convection rather than two-phase convection.

Chapter IX

Complementary Evidence from the Bushveld Complex, Great Dyke, and Skaergaard Intrusion

A. COMPARISONS

1. GENERAL CONSIDERATIONS

Comparatively few large basaltic intrusions show extreme differentiation. The Bushveld complex is the closest analogy of the Stillwater. The probably related Great Dyke is also petrographically similar. The Skaergaard intrusion has some features in common with the Stillwater but differs greatly in shape and considerably in the composition of its original magma. The Duluth gabbro (Grout, 1918a) apparently resembles the Skaergaard in composition (although no analysis of the border facies is available so that the composition of the magma may only be inferred) and the Stillwater in shape. Relatively little is known of the Sierra Leone norite (Dixey, 1922; Junner, 1930) inasmuch as most of it probably lies beneath the sea. In the writer's opinion the Bay of Islands complex in Newfoundland (Cooper, 1936) may be composite, and the gabbroic and peridotitic portions may be separate intrusions. No doubt other such large complexes exist, but the above are the only ones sufficiently well described to be recognized.

2. THE MAGMAS

Border-facies analyses of the Stillwater, Bushveld, and Skaergaard are given in Table 47. None of these analyses describes the magma accurately. The Stillwater analysis has no obvious deficiencies, but more than one analysis would make the evidence much stronger. The Bushveld analysis appears to be about 0.25 low in Na_2O and 0.10 low in K_2O. The normative plagioclase is An_{74}, which is too calcic for an ordinary unmodified basaltic magma; Daly (1928) records the modal plagioclase as An_{65}. In this type of rock the norm and mode should agree very closely. Furthermore the specimen does not represent the original magma, but the residual magma after 10–20 per cent of the intrusion had crystallized. In the discussion of the Bushveld it will be shown that this specimen is from the floor, border facies, at a location where the magma had transgressed outward during crystallization of the lopolith. In spite of these two minor deficiencies the analysis does describe the general character of the magma. The Skaergaard border-facies analysis is a composite of two analyses which differ considerably. The plagioclase content in one is 65.5 per cent and in the other 57.2. While neither analysis looks like an undifferentiated basalt, the composite seems to be fairly satisfactory. Nevertheless the differences between the two analyses throw some degree of doubt on the validity of the composite. The feldspar content is too high, and the MgO:FeO ratio is high for basalt.

The remarkable similarity between the Bushveld and Stillwater magmas is amply demonstrated by the analyses. Both have the same peculiarities as to minor constituents—low K_2O, low Fe_2O_3 and TiO_2.

	Stillwater	*Bushveld*	*Skaergaard*
Fe_2O_3	0.26	0.28	1.18
K_2O	0.24	0.14	0.19
TiO_2	0.45	0.34	1.40

TABLE 47.—*Analyses of Border Facies considered representative of the magmas*

	Stillwater (Peoples, 463G70)	Bushveld (Daly, 1928)	Skaergaard (Wager and Deer, 1939, XIIIa)
SiO_2	50.68	51.45	47.92
Al_2O_3	17.64	18.67	18.87
Fe_2O_3	.26	.28	1.18
FeO	9.88	9.04	8.65
MgO	7.67	6.84	7.82
CaO	10.47	10.95	10.46
Na_2O	1.87	1.58*	2.44
K_2O	.24	.14**	.19
H_2O+	.42	.34	.41
H_2O-	.06	.03	.10
P_2O_5	.09	.09	.07
TiO_2	.45	.34	1.40
Cr_2O_3	.04	—	.00
MnO	.15	.47	.11
CO_2	—	—	.06
S	—	—	.27
ZrO_2	—	—	.00
SrO	—	—	.20
BaO	—	—	.02
CuO	—	—	.007
NiO	.04	—	.00
	99.96	100.22	100.21
S. Gr.	3.03	3.003	2.96
Analyst	R. B. Ellestad	E. G. Radley	W. A. Deer
Q	.00	2.76	.00
Plag. Or	1.39	1.11	1.11
Plag. Ab	15.72	13.62	20.69
Plag. An	39.06	43.09	39.75
Plag.	56.17	57.82	61.55
Di Wo	5.04	—	4.99
Di En	2.60	—	2.90
Di Fs	2.31	—	1.85
Di	9.95	8.47	9.74
Hy En	16.70	—	7.50
Hy Fs	15.12	—	4.42
Hy	31.82	29.52	11.92
Ol Fo			6.37
Ol Fa	—	—	4.59
Ol			10.96
Mt	.46	.46	1.86
Il	.91	.61	2.66

TABLE 47. (*Continued*)

Ap	.20	.34	.20
H_2O	.48	.37	.51
Cal			.10
Py			.51
	99.99	100.35	100.01
	An_{68}	An_{74}†	An_{62}) normative
	En_{59}	En_{56}	En_{68}) molecular ratios

* Estimated to be 0.25 low
** Estimated to be 0.10 low
† Mode An_{65}

The Skaergaard shares only one of these similarities—low K_2O. The Bushveld modal values of An_{65} and En_{56} compare well with the Stillwater values of An_{68} and En_{59} particularly when it is remembered that the Bushveld values in each would be a little higher for the initial undifferentiated magma. The basalt of the Skaergaard is not at all similar to the Bushveld and Stillwater. It is undersaturated and has comparatively high Fe_2O_3 and TiO_2. In spite of these initial differences in the magma there is comparatively little difference in the trend of crystallization or in the resulting rocks except in the greater amount and longer persistence of olivine in the Skaergaard. Initial composition of the basaltic magma does not seem to play a major role in the course of differentiation of such magma by fractional crystallization.

B. THE BUSHVELD COMPLEX

1. INTRODUCTION

Several weeks were spent on the Bushveld complex to collect a suite of typical specimens and to see enough of the general relations to be able to read the literature with the proper understanding. The most important objective was to examine the upper portion of the gabbroic sequence and particularly its upper contact relations. For guidance in the field especially along this upper contact the writer is much indebted to Dr. J. C. Strauss. The writer also wishes to express his appreciation to Dr. L. T. Nel, Director of the Geological Survey of the Union of South Africa, for the many courtesies extended during the course of this work.

The writer's understanding of the Bushveld complex is almost wholly derived from the literature and from conversation with some of the men who wrote that literature. Above all the classic memoir of A. L. Hall (1932) is the foundation of field description upon which all understanding of the geology of the complex is based. Hall expressed his indebtedness to Daly (1928) for his insight into some of the major petrological problems of the complex. Many others contributed to the mapping and reasoning which went into Hall's memoir—E. T. Mellor, G. A. F. Molengraaff, P. A. Wagner, and H. Kynaston, to mention a few.

B. V. Lombaard's 1932 and 1934 papers stimulated the writer to make a similar mineralogical study of a section across the Stillwater complex, and Lombaard is directly or indirectly responsible for most of the petrological literature on the complex since 1934. Among the theses done under his direction are H. J. Nel (1940),

G. S. J. Kuschke (1939), J. C. Boshoff (1942, Unpub. ms.), J. M. van der Westhuizen (1945, Unpub. ms.), J. Berning (1941, Unpub. ms.), C. M. Schwellnus (1939, Unpub. ms.), A. F. Lombaard (1950), C. J. F. van der Walt (1941), R. Liebenberg (1942), E. R. Schmidt (1952), J. J. van der Berg (1946), and F. H. S. Vermaas (1949, Unpub. ms.). An excellent summary of many of these papers is given by Burri and Niggli (1945). Additional papers by Niggli and Lombaard (1933), Strauss (1943), and a lengthy paper by van Biljon (1949), which expresses a different if somewhat unlikely concept of the origin of the complex, may all be noted.

2. OPINIONS

From the general relations in the Bushveld complex as described in the literature, from comparison with very similar relationships exhibited in the Stillwater, and from observation of a few of the critical points in Bushveld in the field, the writer has arrived at the following opinions:

The gabbroic portion of the Bushveld complex represents fundamentally one great intrusion of basaltic magma which has differentiated essentially in place. (This is in disagreement with recent views on the subject most of which present a case for successive intrusions of changing composition together with differentiation *in situ* of each pulse as a secondary modifying effect.)

Two modifying conditions must be appended to the above broad opinion.

(a) The Merensky Reef indicates a quantitatively small addition of *undifferentiated* basaltic magma to the magma chamber after the crystallization of the Critical Zone. Presumably this addition is of essentially the same composition as the original magma.

(b) The magma chamber soon after intrusion was deeper (thicker) and of smaller horizontal dimensions than the complex is today. Either gradually or spasmodically the magma layer extended itself horizontally and diminished in thickness so that the "sea" of magma transgressed over its "shore lines" bringing stratigraphically higher and higher horizons in contact with the floor as it progressed farther from the limits of the original magma "basin".

The gabbroic portion of the complex contains within it all the products of its differentiation. The granitic rocks above the gabbroic facies are not genetically related to the gabbroic rocks but represent a separate intrusion of granitic magma.

3. CONCEPT OF SUCCESSIVE INTRUSIONS VS. DIFFERENTIATION IN SITU

Let us now consider the concept of differentiation of the magma in depth and the emplacement in the lopolith by successive surges. The idea was first advanced by Lombaard (1934)—"The evidence points to the emplacement of the lopolith by intermittent supplies of magma which was continually changing its composition .... Further differentiation took place *in situ* by gravitative accumulation of crystals."

Burri and Niggli (1945, p. 383–384) after reviewing the recent literature on the subject state their concept of the process as follows:

"Was wir sehen ist nicht Differentiation in einem als Ganzes nach der Intrusion einheitlich and völlig flüssig gewesenen Magmaherd. Intrusion Art der Magmenauffüllung und Differentiation stehen unter sich im engsten Zusammenhang; der Hauptteil der Variation von unten nach oben was zuden in der Tiefe Vorgebildet und wurde an gewissermassen sekundärer Stätte zur Intrusionsfolge."

To satisfy the hypothesis of differentiation in depth the magma chamber in depth must be at least several times as large in volume as the gabbroic portion of the Bushveld complex if we consider merely the derivation of these gabbroic rocks. It must not only supply the magma from which the gabbroic rocks of the complex were formed, but also it must use up an even larger proportion of its bulk on crystallization at depth in order to produce the successively more iron-rich and alkali-rich fractions of residual magma. If the Bushveld gabbroic rocks represent a sheet about 7 km thick, the magma chamber below might be 15 or more kilometers thick. Its base must be 30 or more km below the surface. The temperature in this environment is estimated to be 600°C. or higher. Under these conditions the time necessary to conduct the heat away from this deep magma chamber and permit crystallization would be vastly longer than the time necessary to conduct a similar amount of heat away from a lopolith whose upper contact was virtually at the surface of the Earth. Inasmuch as crystallization proceeds at the rate at which the heat of crystallization can be removed from the magma by conduction through its walls, crystallization and hence differentiation in the deeper magma chamber in the hotter environment will proceed at a small fraction of the rate at which it would go at the Earth's surface. If the order of magnitude of time necessary to crystallize the gabbroic portion of the Bushveld complex is 100,000 years, the deeper magma will take several times as long. Thus if successive surges from the deep magma chamber were to be fed into the lopolith, either the surges would have to consist of undifferentiated or virtually undifferentiated basalt, or there would have to be long time intervals between each surge, and the surges would have to be intruded above the products of the previous surge long after it had solidified and frozen to its roof rocks. The first of these alternatives eliminates the hypothesis of successive surges with differentiation in depth. The second alternative should produce features detectable but undetected in the field evidence, and it would require an unacceptable series of coincidences whereby each surge would invariably choose the solid contact between the top of the last surge and its roof as a locus of intrusion. If we assume the further hypothesis that the Bushveld granitic rocks were also a differentiation product of the same deep magma, the quantitative difficulties become even more insuperable. The magma chamber would have to be greatly larger than the figure given above, and the percentage of crystallization in the chamber would have to be higher. It does not seem possible to accommodate so large a chamber within the limited confines of the Earth's crust.

Lombaard, early in his work on the Bushveld, devised three "controls" to study the relations between the various crystalline phases with increase in stratigraphic height in the complex. These are as follows:

The feldspar control $\frac{\text{An}}{\text{An} + \text{Ab}}$

The pyroxene control of "pr" $\frac{\text{orthopyroxene}}{\text{orthopyroxene} + \text{clinopyroxene}}$

The pyroxene composition control $\frac{\text{Mg}}{\text{Mg} + \text{Fe}}$ in the orthopyroxene.

He examined the variations in these controls in sections across the Bushveld gabbroic rocks from the floor upward. One section was made through Brits on the southern margin of the complex, and one along a line westward from Lydenburg on the eastern margin of the complex. These studies involved only a dozen or so specimens on each traverse, and they gave rather erratic variation curves. A great many more sections and more specimens were later studied by Lombaard's students. In general the values of all three controls decline with progress upward in the sections. The downward trends in the controls are not steady but tend to be oscillatory; all three controls move as a rule sympathetically. This caused Lombaard to advance the concept of successive surges of magma from a larger magma chamber below in which differentiation was taking place. Reversals in the general downward trend of the controls or "turning points" are often marked by pyroxenites or pyroxene-rich horizons. Kuschke (1939) gives an excellent summary of these relations and the deductions to be drawn from them.

The writer set out to make the same sort of study in the Stillwater complex. An important item in deducing the causes of the oscillations in the controls is spacing of the specimens at the correct intervals so that there are a number of points on each oscillation and so that the changes of trend can be accurately defined. During the 1935 field season one suite of sample was collected a foot or less apart across 50 feet of a rhythmically layered section. These proved to be too closely spaced and showed virtually no change in the controls through the entire section except for the effect of the interstitial liquid on constituents present in small amount as discussed elsewhere. The collection of specimens across the complex from the floor to the highest exposure was made at about 300-foot intervals (stratigraphically). This spacing is somewhat too wide to get all the detail necessary. After these samples had been studied Peoples and Howland collected an additional suite of specimens spaced at about 30-foot intervals for 1200 feet—the Dead Tree section. It was found that the oscillations were related to the rates of sedimentation of crystals on the floor which in turn, it was postulated, were the result of variations in the vertical velocity of motions in the overlying magma. The relations were found to be very complex with smaller oscillations superimposed on larger. Completion of this portion of the investigation resulted in the great mass of previously unrelated facts—the compositions of the crystalline phases, the ratios of the phases to one another, the grain sizes, and the textures—falling into a single connected and explainable pattern.

Considering the difficulties involved in the postulate of differentiation in the deep magma chamber and the fact that the oscillations in the controls can be explained satisfactorily by an internal process in the lopolith, it seems more reasonable to the writer to conclude that: the changes in composition of the main crystalline phases and the changes of the phases with stratigraphic height were accomplished by fractional crystallization *in situ* of a single intrusion of undifferentiated basaltic magma; the oscillations in the controls may be accounted for by an internal mechanism related to the rate of sedimentation of crystals on the floor.

4. MERENSKY REEF

The Merensky Reef, a layer, is persistent for hundreds of miles along the strike and in this respect differs from a rhythmic layer which ordinarily extends horizon-

tally for only about 1000 times its thickness. The Merensky Reef corresponds closely to such persistent layers as the titaniferous magnetite and chromite horizons and no doubt has a similar origin. Layers of titaniferous magnetite and chromite represent points in the history of the magma when it became saturated with these constituents and precipitated them. If the residual magma were homogeneous at all times then there probably would be only one chromite horizon or one magnetite horizon. In so large a magma chamber some portions of the magma crystallize more rapidly than other portions although these differences are small and are no doubt being continually eliminated by convective circulation. The differences, however, are sufficient so that chromite or magnetite may be precipitated several times forming several individual horizons as each batch of liquid reaches the saturation point in various portions of the overlying magma. The chromite layers are grouped stratigraphically close together, as are the magnetite layers higher in the complex. The total number of magnetite or chromite layers varies in different sections of the complex. Probably no one horizon is persistent for great horizontal distance, but the group of horizons as a whole is. It appears that the Merensky Reef, a relatively thin layer, is more persistent than the individual chromite and magnetite layers. It has the further peculiarity of representing sudden precipitation of not one but of two new phases simultaneously—chromite and sulfides. These facts strongly suggest that it is not a normal precipitate in the course of differentiation of the magma. The facts could be explained nicely if at this point in the history of the complex there were a small addition of undifferentiated basalt. Chromite had already been precipitated from the main residual magma. Presumably, as in the Stillwater, an immiscible sulfide liquid separated soon after emplacement of the main magma. This would be in the bottom of the present saucer and not visible in the outcrops along the rim. If a small amount of undifferentiated basalt were added, chromite and sulfides would immediately be precipitated upon mixing of the new magma with the main residual magma which had already reached its saturation point for these two constituents. Assuming that this has happened and that the new magma had the same composition as the original Bushveld magma, one can estimate the relative volume of the addition roughly as follows:

$$\frac{\text{volume of new magna}}{\text{volume of original magna}} = \frac{\text{volume chromite in Merensky Reef}}{\text{volume chromite in chromite horizons}}$$

This ratio suggests that the addition to the magma represents a volume equal to only a few per cent of the bulk of the original magma.

The other features associated with the Merensky Reef, such as the pyroxenites, are mechanical rather than chemical results of the addition of new magma. The motion imparted to the magma by the physical act of intrusion of more magma, and by the fact that the new magma was hotter and not in thermal equilibrium with its surroundings, would result in an increase of velocity of currents in the magma as a whole. As was pointed out in the discussion of the mechanics of rhythmic layering, this would almost inevitably result in the slowing down of the settling of plagioclase more than it would pyroxene, and a pyroxenite layer would be formed.

Examination of the controls to learn if the crystals above the reef become more

magnesian and more anorthitic leads to no definite conclusion. The amount of magma added is probably too small to produce an effect noticeable above the background of small oscillations related to other causes. In fact, if only a few per cent of new magma were added, it can readily be calculated from the probable Mg:Fe ratio and An:Ab ratio of the new and the residual magma at this stage that the effect on the controls would be too small to detect.

5. PROGRESSIVE TRANSGRESSION OF THE GABBROIC MAGMA

The uppermost gabbroic rocks of the Bushveld are similar chemically and mineralogically wherever they are encountered. This is not true of the floor rocks. Both the lower "chilled" marginal facies and the first settled crystals above the level which has been affected by the "chilling" show marked changes along the rim of the complex.

The lowermost ultramafic rocks—bronzitites and harzburgites—are along the western and northeastern rim, but along the border zone, either on the southern or northern margin, it appears that progressively higher horizons are in contact with the floor. Proceeding for example from Rustenburg (Kuschke, 1939) to Brits (B. V. Lombaard, 1934) to Pretoria (Nel, 1940) the lower ultramafic rocks thin and finally disappear not far beyond Brits. The so-called Critical Zone thins and apparently disappears northwest of Pretoria. North of Pretoria, Main Zone rocks are in contact with the floor. Farther east in Blood River valley the Main Zone also forms the floor (A. F. Lombaard, 1950), but the trend is reversed to the northward along the eastern rim of the complex (B. V. Lombaard, 1934; Schwellnus, 1939, Unpub. ms., Univ. of Pretoria). By plotting on a sketch map of the Bushveld the anorthite content of the most calcic plagioclase near the base of each of the above sections, and contouring the same, the spatial relations can be visualized (Fig. 31). The same may also be accomplished by plotting the Mg content of the most magnesian orthopyroxene at the base of each section (Fig. 32). The number of points on each figure is too small for reliable details of the contouring. It is also quite clear from the available petrographic descriptions that the chilled border facies becomes more ferriferous and more albitic as the overlying settled crystals change in the same direction. The evidence is reasonably clear that the magma spread laterally as differentiation proceeded so that the products of progressively later residual magmas form the floor as one goes outward toward the margin of the original lopolith (Fig. 33).

The mechanics of intrusion of diabase sills below the floor of the lopolith and the chemical relations of the sills to the Bushveld differentiation sequence can also be accounted for by the reasoning of the last paragraph (Fig. 33). The lowest sill in the figure was intruded at an early stage soon after the initial intrusion of magma into the lopolith and thus is relatively magnesian and anorthitic compared to younger sills. Each higher sill, farther out toward the rim of the lopolith, represents a transgression later in the differentiation of the complex, and each becomes more ferriferous and albitic. Lines of "xenoliths", which are common within the lower portion of the gabbroic rocks of the complex, are the septa of sedimentary rocks shown in the diagram and were originally connected in the present up-dip direction to the wall-rock framework of the complex. The transition from rows of "xenoliths" within the

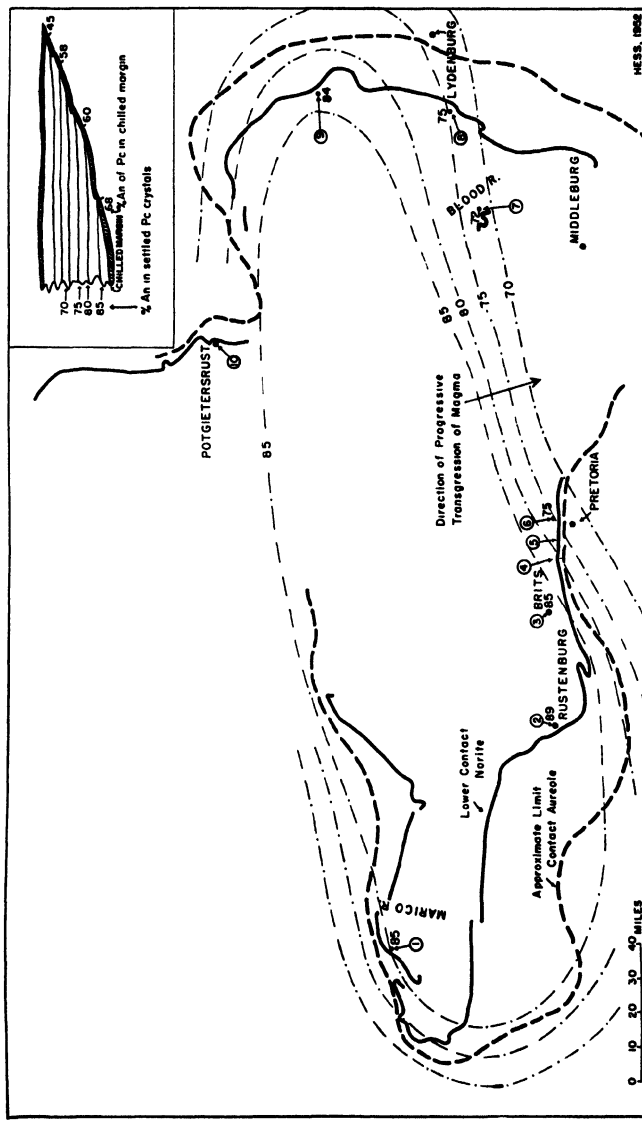


FIGURE 31.—*Composition of earliest accumulated plagioclase for various parts of Bushveld complex*

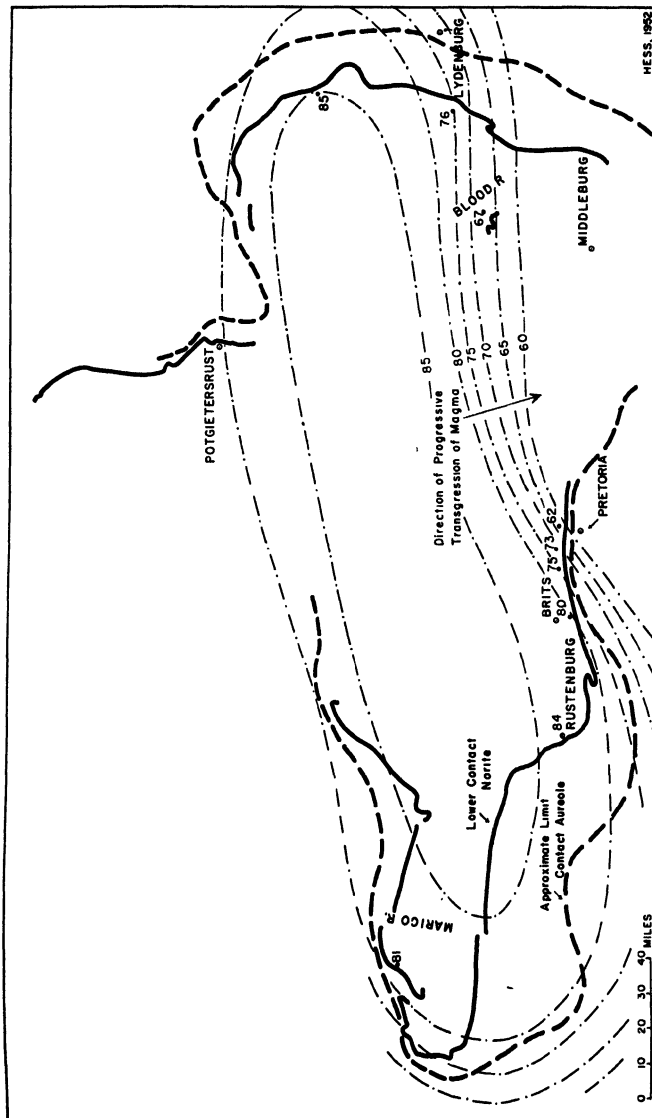


FIGURE 32.—*Composition of earliest accumulated orthopyroxene for various parts of Bushveld complex*

complex to a point at which they become the floor of the complex with a diabase sill below can be seen at a glance in the diagram. The present erosion surface across the rim of the complex can be represented by a line such as A-A in Fig. 34. The Stillwater complex by way of comparison has been steeply upturned and deeply

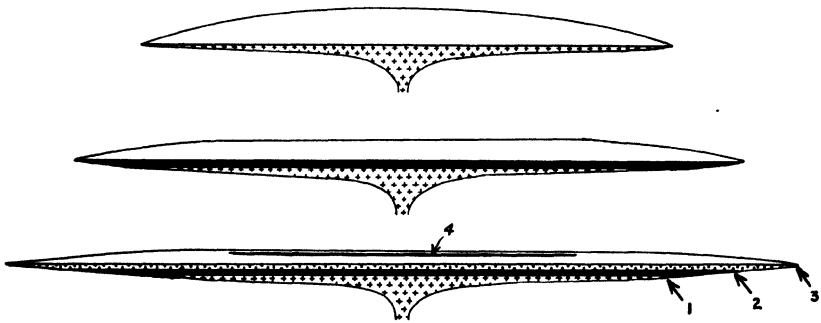


FIGURE 33.—*Diagram illustrating transgression of Bushveld magma during crystallization* 1, 2, 3, and 4 indicate successive magma compositions.

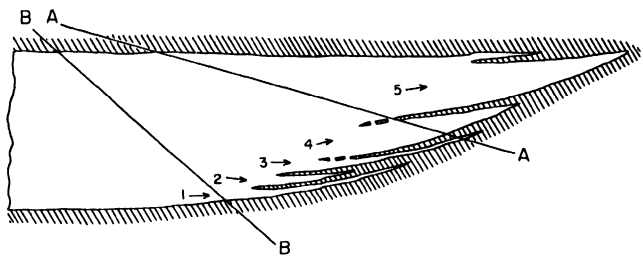


FIGURE 34.—*Cross section illustrating transgression of magma during crystallization and structural relationship of sedimentary septa and sills in floor to main intrusion*

A-A, level of erosion Bushveld complex; B-B, level of erosion Stillwater complex. Arrows point to five successive outward transgressions of magma.

eroded to a section such as represented by B-B; thus the sill and "xenolithic" marginal features are not found.

6. UPPER CONTACT OF THE BUSHVELD GABBROIC FACIES

Probably no feature of the Bushveld complex has been so misinterpreted as the contact of the Upper Zone with its overlying roof rocks. Nevertheless the relations are reasonably clear where exposures permit examination of this contact. Where seen by the writer a fine-grained, "chilled" dark facies of the gabbro was in contact with an overlying highly metamorphosed sedimentary sequence—the Rooiberg series. Angular fragments of the overlying rocks are found in the gabbro. Chemical analyses of the upper border facies are given in Table 48.

The roof rocks consist largely of metamorphosed quartzo-feldspathic sediments. The rather abundant granophyres have been misinterpreted to be igneous rocks and specifically late-stage differentiates of the gabbroic magma. The field relations—

TABLE 48.—*Chemical analyses on liquid line of descent*

	Stillwater 463 G 70	Great Dyke		Bushveld	
		GD 29	GD 31	BV 41	BV 52
SiO_2	50.68	52.55	50.19	47.23	49.86
Al_2O_3	17.64	15.38	14.79	15.71	11.17
Fe_2O_3	.26	.70	1.57	5.05	2.37
FeO	9.88	10.26	12.47	13.36	14.70
MgO	7.67	6.84	6.12	4.11	3.44
CaO	10.47	10.30	10.36	8.03	10.59
Na_2O	1.87	2.70	2.42	2.60	2.46
K_2O	.24	.22	.23	.50	.91
H_2O+	.42	.41	.40	.67	.50
H_2O-	.06	.06	.06	.07	.08
TiO_2	.45	.30	.94	2.17	2.02
P_2O_5	.09	.01	.01	.09	1.20
Cr_2O_3	.04	.01	.01	.02	.00
MnO	.15	.21	.23	.27	.28
NiO	.04	ND	ND	ND	ND
Cl	ND	.01	ND	.01	.16
S	ND	.03	.19	.26	.08
CO_2	ND	ND	ND	ND	.06
BaO	ND	ND	ND	ND	.05
	99.96	99.99	99.99	100.15	99.93
Less O for S and Cl.....		−.01	−.08	−.09	−.07
		99.98	99.91	100.06	99.86
Analyst.	Ellestad	Oslund	Oslund	Oslund	Oslund

Norms

Q		—		—		—		1.38		3.48
Pc	Or	1.39	56.17	2.22	53.78	1.39	50.60	2.78	54.54	5.56	42.92
	Ab	15.72		22.79		20.44		22.01		20.96	
	An	39.06		28.77		28.77		29.75		16.40	
Di	Wo	5.04	9.95	9.34	18.56	9.28	18.56	3.83	7.71	11.83	24.21
	En	2.60		4.20		4.00		1.50		3.40	
	Fs	2.31		5.02		5.28		2.38		8.98	
Hy	En	16.70	31.82	12.80	26.00	11.20	25.86	8.70	23.75	5.10	18.17
	Fs	15.12		13.20		14.65		15.05		13.07	
Mt		.46		.93		2.32		7.42		3.48	
Il		.91		.61		1.82		3.50		3.80	
Ap		.20		—		—		.34		2.85	
Cc		—		—		—		—		.10	
Py		—		—		.36		.48		.15	
H_2O		.48		.47		.46		.74		.58	
		99.99		100.35		99.98		99.86		99.74	

TABLE 48.—*Continued*

	$An_{68.2}$ $En_{59.4}$	An_{62} En_{55}	$An_{55.5}$ En_{50}	$An_{52.5}$ $En_{42.5}$	$An_{52.5}$ En_{34}
	Modes				
	Stillwater 463 G 70	Great Dyke		Bushvveld	
		GD 29	GD 31	BV 41	BV 52
Pc	53.5	45	44	58	46.8
C-Px	7.9	14	21	9	25.6
O-Px	37.3	40	29	18	6.6
Opaque	1.3	1	6	14	8.7
Q				.8	2.6
Bi				.4	—
Hb				—	5.2
Ap				—	4.5
	100.0	100	100	100.2	100.0

463 G 70. Border facies, Stillwater complex (Peoples, 1932, Ph.D. thesis, Princeton Univ.)

GD 29. Roof border facies, Great Dyke, Gwelo Salisbury Road. *See* Figure 1, Hess (1950)

GD 31. Roof border facies, Great Dyke, Gwelo-Salisbury Road. *See* Figure 1, Hess (1950)

BV 41. Roof border facies, Bushveld complex east of Potgietersrust. Collected with Strauss. *See* his Figure 1 (1943). Location small lenticular patch of norite just north of his section A-B a little to right of center

BV 51. Roof border facies, southeastern Bushveld road outcrop on route to Tauteshoogte. Collected with Strauss

intercalation of granophyric rocks with quartzites and hornfelses—suggested a metasedimentary origin to the writer. Stronger evidence is, however, available from recent petrographic studies. Vermaas (1949, M.Sc. Thesis, Univ. Pretoria) analyzed the zircons from Bushveld granites, granophyres, leptites, and quartzites. The last three have practically identical zircon suites with a length to breadth ratio less than 2 to 1; the granites have zircons with ratios between 5 to 1 and 2 to 1.

Above the Rooiberg metasedimentary sequence is the main Bushveld granite, and apophyses of granite intrude both the metasedimentary rocks and upper gabbro. The relations along the contact are well described by Strauss (1943) for an area west of Potgietersrust. The metamorphism by the gabbro here has been sufficiently intense to mobilize some of the quartzose sediments forming rheomorphic breccias in the immediate vicinity of the contact.

Another circumstance that has led to confusion is the fact that rocks on either side of the contact have been called "diorites". Metasedimentary rocks above the contact consisting of microperthite, oligoclase, quartz, and minor ferromagnesian minerals have a composition which may be referred to as "diorite". These rocks may have a granitoid texture. Below the contact the rock that the writer has chosen to call "gabbro" has andesine as its plagioclase and thus might also be called "diorite". Its texture and mineralogy (two pyroxenes, titanomagnetite, and andesine, An_{45}) show clearly that it is a variant of the normal underlying Bushveld gabbro. Below this border facies and above the uppermost magnetite horizon are rocks con-

sisting of andesine and ferrohortonolite, which have also been classed as "diorites". The textures are typical of crystal accumulates having peri-euhedral plagioclase and olivine as main constituents with a minor amount of interstitial pyroxine as a rule. These might be called ferrohortonolite andesinites as the similar rocks in the Skaergaard were called by Wager and Deer.

Still another factor that has led to misinterpretation is the gradation in composition of several of the mineral phases up to, across, and in apparent disregard of, the contact. Thus the An:Ab ratio of the plagioclase, the Mg:Fe ratio of the pyroxene, and Fo:Fa ratio of the olivine may be traced across the contact with fairly continuous downward trends. This is not so strange inasmuch as the physical conditions of temperature and pressure were gradational across the contact at the time of crystallization or recrystallization so that there was a strong tendency for the mineral phases to become adjusted in composition to this gradient. There is, however, a marked change in the proportions of the mineral phases present and in textural relations upon crossing the contact.

The writer has little confidence in the proposition that consanguinity of a rock suite can generally be demonstrated by means of chemical-variation diagrams. A great many competent petrologists do look upon this device with favor, however, so the evidence so derived cannot be disregarded. Niggli and Lombaard (1933) give such a diagram for the entire Bushveld suite which purports to show a continuous sequence through the differentiates of the gabbroic lopolith all the way to the Bushveld granite. The differentiation within the gabbroic rocks is fairly well portrayed, but not nearly so clearly as can be done by using curves of variation of mineral compositions and proportions with stratigraphic height (Lombaard's controls). In the chemical-variation diagram the jump from ferrogabbro far up the scale to the granite is accomplished only by including some analyses of intermediate types found in the roof rocks above the gabbro, which the writer regards as metasedimentary rocks.

7. RELATION TO BUSHVELD GRANITIC ROCKS TO THE BUSHVELD GABBRO LOPOLITH

Many have regarded the Bushveld granite and its related felsites as differentiates wholly or in part of the basaltic magma of the lopolith. The quantity of granite is vastly too great to have been derived by differentiation of the gabbroic rocks of the lopolith. The derivation of the granite and gabbro by differentiation of a huge mass of basaltic magma in depth also seems quantitatively unlikely.

Let us consider the general case of deriving a granitic magma by differentiation of basalt. That granitic material may be such an end product of crystallization of basalt is suggested by the common occurrence of a small amount of interstitial micropegmatite as a last stage in the crystallization of dolerites.

The bulk composition of the lower 60 per cent of the Stillwater has been determined, and the composition of the residual magma at this stage calculated. This calculated residual magma is very close in composition to the exposed portion of the Skaergaard intrusion, and probably it resembles even more closely the upper portion of the Bushveld gabbro (Table 48).

Under the most favorable circumstances one could derive a volume of granite from the Stillwater residual magma equivalent to about 5 per cent of its mass.

This would amount to only 2 per cent of the original amount of basaltic magma intruded in the lopolith. Wager and Deer arrive at a very similar figure in their computation of the amount of granitic material which could be derived from the Skaergaard magma. The amount of granitic residue which might be produced by the Bushveld basaltic magma can be set between the limits of 3 and 0 per cent. Petrographic examination of a large number of Bushveld gabbroic rocks suggests that the lower limit rather than the upper is nearer the correct answer. Probably the small potential residue of granitic material is represented by the small amount of interstitial quartz and alkali feldspar present in the gabbroic rocks. In other words the gabbro lopolith contains within it all the products of its differentiation. In the most extreme cases of fractional crystallization, in which the chances of obtaining a granitic residue would be most favorable, it seems as though little if any granite results. Therefore on general grounds it seems very unlikely that large bodies of granite are ever derived from fractional crystallization of basalt.

The common association of the younger felsic rocks overlying the mafic is far more reasonably accounted for on the basis of Kennedy's and Anderson's (1938) concept of magma generation rather than by the quantitatively awkward means of resorting to magmatic differentiation either in depth or *in situ*. The Kennedy-Anderson process can be further enhanced by assuming partial rather than complete fusion of various crustal layers (Buddington, 1943) in order to generate the magmas. Bowen (1928) obtained his primary basaltic magma by partial fusion of feldspathic peridotite. Granite might similarly be derived by partial fusion of dioritic rock higher in the crust, or diorite might be derived by partial fusion of gabbro. This is further discussed in Appendix 1.

C. THE GREAT DYKE

An excellent general description of the Great Dyke is given by Lightfoot (1940). It is about 325 miles long and averages about 4 miles wide. It is largely harzburgite and bronzitite, but relatively small relics of feldspathic rocks rest upon it at its widest portion. The present writer (1950) noted what he considered to be remnants of an upper chilled margin. The layering dips gently inward near the walls and is nearly horizontal in the middle of the Dyke. No chilled margin against the wall was present at the one exposure of the contact seen by the writer.

The rocks and mineral phases are practically identical to those of the Bushveld complex, so it seems very likely that both are related to the same igneous episode and that both crystallized from the same sort of magma. A slight difference in mineralogy of the pyroxene may be correlated with the extremely slow cooling of the enormous vertical column represented by the Dyke. The cooling was so slow that it apparently allowed Ca ions to migrate to the margins of orthopyroxene and escape. Thus the total CaO content of these orthopyroxene crystals is somewhat lower than in any others of the Bushveld type which have crystallized from a basaltic magma (Table 3, no. 7704). In some cases so much CaO has been lost that lamellae of diopside are absent (although not all the CaO has escaped). Complementary to this loss of CaO from the orthopyroxene, very large poikilitic chrome diopsides have developed in some of the bronzitites.

Vertical variation of the mineral compositions and phases in the Great Dyke

were discussed by the writer (1950), but one modification in these data is necessary. The plagioclase is shown as decreasing in anorthite content to An_{44}; actually the minimum is slightly greater than An_{80}. An analysis of a plagioclase is given in Table 10.

The writer is indebted to F. E. Amm, J. W. Wiles, and J. G. Stagman for background information on the Great Dyke and for their guidance in the field.

Previously the writer suggested that the Great Dyke was the feeder for a much

TABLE 49.—*Stillwater residue after 60 per cent crystallization compared to Bushveld chilled roof facies and Skaergaard second and third liquids*

	Stillwater Residue	BV 41	BV 52	Skaergaard liquids (calculated)	
				2d	3d
SiO_2	52.0	47.7	50.4	46.7	45.7
Al_2O_3	14.8	15.8	11.3	15.3	12.7
Fe_2O_3	5.1	5.1	2.4	2.9	3.6
FeO	12.0	13.5	14.8	12.9	18.2
MgO	3.4	4.1	3.5	5.9	3.5
CaO	8.3	8.1	10.7	9.9	8.3
Na_2O	2.39	2.60	2.50	2.83	3.16
K_2O	.51	.50	.92	.29	.44
TiO_2	.90	2.20	2.04	2.18	2.36
P_2O_5	.19	.09	1.21	.42	.92
MnO	.23	.27	.28	.16	.22
	99.82	99.96	100.95	99.48	99.10
	An_{53} En_{40}	$An_{53.5}$ $En_{43.5}$	An_{37} En_{34}	An_{51} En_{52}	An_{48} En_{30}

larger eruptive mass. If the magma were basaltic, the ultramafic rocks of the Dyke would require a feldspathic complement 10 times as large as the volume of ultramafic rocks. The presence of the upper chilled border indicates a total thickness of feldspathic rocks over the Dyke of only 2000–3000 feet. Apparently the magma spread laterally over a vast area so that the sheet formed was much thinner than the Bushveld lopolith.

This raises the question as to whether the Bushveld and other lopoliths have a "Great Dyke"-like feeder beneath them too. If there were such under the Bushveld one would suppose it to be oriented west-southwest–east-northeast along the longer axis of the present elliptical area of outcrop. The mineralogical data of Figures 31 and 32 would also favor such an orientation. Weiss' (1940) gravimetric traverse across the Great Dyke gave a very sharp positive anomaly of 50 milligals. The feeder for the Bushveld if present might well be found by gravity measurements even though buried 12–15 km below the present surface. This would be a relatively easy project to carry out.

D. STILLWATER-BUSHVELD-GREAT DYKE LIQUID LINE OF DESCENT

As shown in Figures 31 to 34 a series of chilled-border samples would represent successively later residual liquids as fractional crystallization proceeded. Daly's

analysis of the lower border zone itself represents a liquid after about 10 per cent of the complex had crystallized. Fortunately rocks representing essentially the end product—final liquid—are available, and analyses of two from the Bushveld and two from the Great Dyke are given in Table 48. Inasmuch as the Stillwater and Bushveld initial magmas appear to have been identical (Table 47), the above four analyses are compared with the Stillwater lower border zone in Table 48.

It was quite unexpected to find an upper border in both the Bushveld and Great Dyke which represented a final residuum chilled against the roof. One would expect that initially the floor and roof border zones represented the original magma. No doubt such a border zone must have once been present along the roof. In some manner it has been eliminated. Perhaps after the roof rocks became heated it was removed by stoping, or, if enough initial superheat was present, the convecting magma below may have remelted it. From then on magma was in contact with very hot roof rocks and did not chill or become so viscous that a new border zone developed. Only when a small percentage of liquid was left and the temperature of crystallization began to drop rapidly was a fine-grained border zone again developed. It is these rocks which have been sampled and analyzed in Table 48 (GD 29, BV 41, BV 52). Actually the very last liquid was trapped immediately beneath this border zone. A sample of the residuum beneath the roof border zone of the Great Dyke is represented by GD 31 in Table 48. It is not significantly different from GD 29.

Both Great Dyke samples show less extreme enrichment in iron than do the Bushveld samples. Assuming the original Bushveld and Great Dyke magmas were the same these two pairs give us two points on the liquid line of descent. In Plate 11 the liquid line of descent as represented on the MgO-FeO-alkalies diagram is given for the Bushveld, Skaergaard, and Stillwater. All are very much the same except for the last couple of per cent of liquid in the Skaergaard where great enrichment in alkalies occurs. This does not seem to happen in the Bushveld-Great Dyke and probably did not in the Stillwater. The alkali enrichment may be due to initially higher K_2O in the Skaergaard, or it may be attributed to contamination by incorporation of gneiss in the Skaergaard magma. Wager and Deer postulate that part of the granophyric residuum is the result of contamination.

In Table 49 the calculated residue of the Stillwater after 60 per cent crystallization is compared to the upper border zone analyses from the Bushveld and the Skaergaard calculated second and third liquids. The general similarity of all five analyses may be noted. The Stillwater residue and BV 41 are close except that SiO_2 is lower in BV 41. The Stillwater residue has less than half as much TiO_2 as any of the other four. This may be an error in computation of the Stillwater residue. BV 52 is uncommonly rich in apatite, and P_2O_5 is 1.21 per cent in the analysis. One might suspect that the apatite has been introduced—perhaps carried upward by volatiles from below. The Skaergaard magma started out with lower silica, and the second and third liquids have lower silica than the Stillwater residue or Bushveld final liquid.

Appendixes

Material contained in Appendices 1 and 2 rather than in the main body of this report includes consideration of certain corollary problems on which the main part of the investigation throws some light. Much of the material is in the form of opinions which the writer cannot substantiate to a degree which would warrant inclusion in the main portion of the report. Such opinions may be useful to others in guiding or stimulating further research, and their presentation in this form gives the writer greater freedom in expressing ideas which otherwise he might hesitate to include.

When the Stillwater investigation started there was virtually no reliable data on optical properties of natural calcic plagioclase. Most of the chemical analyses then available were of poor quality or were made on altered or impure material. At the suggestion of James Gilluly, Alexander H. Phillips and Hess began a reinvestigation of this group. Phillips completed many of the analyses before his death, and Hess began to tackle the problem of greater precision of measurement of the optical properties on these and on the pyroxene group.

In the early 1950's Smith took over the problem of working out the methods and making the measurements of the optical properties of plagioclase concentrating on the low-temperature type found in slowly cooled rocks. With skill, ingenuity, and hard work he carried this investigation to a most successful completion. His results are given in Appendix 3.

Appendix 1. Products of Differentiation by Fractional Crystallization Compared to the Calc-Alkaline Magma Series with Comments on Generation of Magmas and a Resulting Composition Gradient in the Crust

1. Conditions favoring extreme fractional crystallization of a basaltic magma by crystal sorting
   (a) Long time interval during which crystallization proceeds
   (b) Low to moderate viscosity
2. Trends of change in a basaltic magma which result from a high degree of fractional crystallization
   (a) Major variables in the Stillwater, Bushveld, and Skaergaard trends
   (b) MgO:FeO: total alkalies (weight per cent) diagram as an indicator of fractional crystallization
   (c) Differentiation of basaltic volcanic sequences, dikes, and thin sheets
3. Discussion of origin of basaltic magma and the calc-alkaline magma series by selective fusion in depth
   (a) Generation of basaltic magma
   (b) Generation of calc-alkaline magma series
4. Generation of magmas and development of a composition gradient in crust

1. CONDITIONS FAVORING EXTREME FRACTIONAL CRYSTALLIZATION OF A BASALTIC MAGMA BY CRYSTAL SORTING

Relatively few examples of extreme fractional crystallization of basalt comparable to the Stillwater complex are known—the Bushveld complex and related Great Dyke, the Skaergaard intrusion, and the Duluth gabbro. Only in the Skaergaard have the late-stage products been thoroughly described, but a certain amount of information is also available on the Bushveld complex.

Let us consider the prerequisites for extreme differentiation such as is found in these intrusions.

(a) LONG TIME INTERVAL DURING WHICH CRYSTALLIZATION PROCEEDS

The magma must cool very slowly to permit crystal sorting. Slow cooling is a function of (i) size of the magma body and to a lesser extent (ii) its shape, (iii) its depth, (iv) the temperature and thermal properties of its wall rocks, and (v) its internal thermal properties.

(b) LOW TO MODERATE VISCOSITY

High viscosity would prevent crystal sorting (considering normal viscosity to be that which is normal for basalt). High viscosity in basalt may result from abnormalities of composition such as above-average Or + Ab content, high SiO_2 or Al_2O_3, or too rapid cooling. Low viscosity (below normal) may be brought about by an increase of dissolved volatiles and by relatively high FeO, MgO, and CaO.

The Sudbury lopolith is large. The mafic portion of this intrusive shows surprisingly little differentiation in spite of its size and hence long time available for differentiation. In the succeeding section of this report it will be shown that the MgO-FeO-total alkalies diagram is an excellent indicator of the degree to which fractional crystallization by crystal sorting has occurred. Figure 35 shows the plotted results for 82 chemically analyzed Sudbury igneous rocks. These fall into two distinct groups—the "norites" and the "granitic" types. The norites show practically no variation along the lines of fractional crystallization as discussed in connection with Plate 11. Apparently no crystal sorting has occurred. An examination of the composition of the border-facies analysis of the Sudbury lopolith shows an abnormally high Or + Ab content, although the MgO:FeO ratio is normal for basalt. Possibly, therefore, the high Or + Ab has increased the viscosity enough to inhibit crystal sorting. (Recently Williams and Thompson have suggested that the Sudbury igneous rocks are not a lopolith.)

The Duluth sills (Schwartz and Sandberg, 1940) show a remarkable degree of differentiation in spite of their comparatively small thickness. They appear to be of normal basaltic composition. In this case the history includes considerable igneous activity preceding their intrusion, and the high degree of differentiation may have been due to relatively high wall-rock temperatures, which permitted slow crystallization of the magmas.

The thin picrite-teschenite sills of Scotland (Flett, 1930; 1932) and the Lugar Sill (Tyrrell, 1951) also show a marked degree of crystal sorting. Seemingly an uncom-

monly high volatile content lowered the viscosity sufficiently to facilitate rapid crystal sorting.

Finally the Bushveld, Stillwater, Skaergaard, and Duluth igneous massifs represent fairly normal basaltic magmas in so far as composition and viscosity are concerned, but size was the determining factor in permitting extreme differentiation.

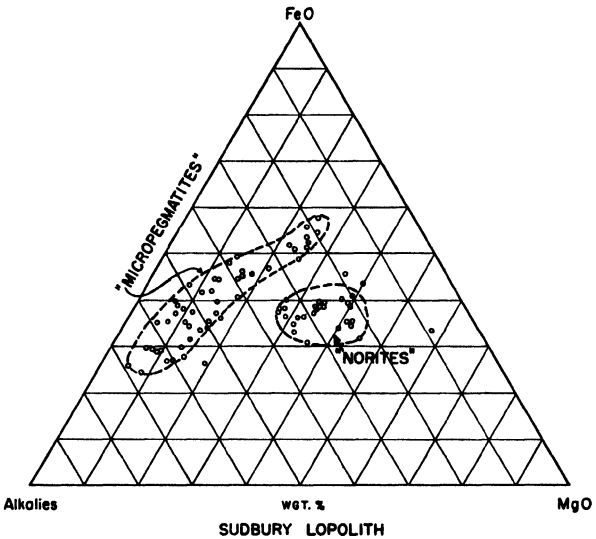


FIGURE 35.—*MgO-FeO-alkalies diagram for Sudbury analyzed rocks*

2. TRENDS OF CHANGE IN A BASALTIC MAGMA WHICH RESULT FROM A HIGH DEGREE OF FRACTIONAL CRYSTALLIZATION

a. MAJOR VARIABLES IN THE STILLWATER, BUSHVELD, AND SKAERGAARD TRENDS

More than 98 per cent of such magmas as the Stillwater and Bushveld consist of potential plagioclase plus pyroxene. Both the norm and mode of the Stillwater border facies contain more than 98 per cent of these constituents, which is an expression of the phases that will develop with no fractionation. With extreme fractionation the situation is very much the same. Plagioclase and pyroxene are the preponderant constituents of the products. Olivine forms in place of pyroxene early in this sequence and also at a very late stage, judging from the Bushveld. The crystallization of olivine will result in the development of some free quartz. Titaniferous magnetite crystallizes in a very late stage of the Bushveld sequence. This indicates oxidation of some of the FeO as differentiation proceeds; the crystallization of such oxides should also free some SiO_2 (which would otherwise have been combined in pyroxene) to form quartz. The initial 0.24 per cent K_2O of the Stillwater magma and the fact that the calcic plagioclase carries approximately 0.6 per cent Or makes it doubtful whether K feldspar will appear as an independent phase; if is should appear, the quantity must be small. The Bushveld suggests that K feldspar did not form, but it

does appear in the last stage of crystallization of the Skaergaard. Contamination by granitic material accounts for a considerable part of, if not all, the K feldspar in the Skaergaard. With a higher initial K_2O content than the Stillwater, Bushveld, or Skaergaard, basaltic magmas undergoing fractional crystallization would produce K feldspar as a late product. The interstitial micropegmatite of the great majority of differentiated dolerites is an indication of this, although even in this case contamination by sediments or granitic material accounts for at least part of the residue.

The two main variables in the Stillwater type of differentiation by extreme fractionation are the trend from calcic to sodic in the plagioclases and the trend from magnesian to ferriferous in the pyroxenes. The third most important variable is the ratio of plagioclase to pyroxene. The evidence available suggests that these two constituents separate from the magma in very nearly the same ratio as they are present in the magma. In the earliest stages olivine and bronzite crystallize so that the feldspar to pyroxene ratio rises in the magma. After the start of crystallization of plagioclase the ratio remains almost constant for much of the course of crystal fractionation. Judging from the uppermost Bushveld rocks the plagioclase ratio drops somewhat in the closing stages. In the Skaergaard according to Wager and Deer's estimates of the successive liquids the plagioclase ratio falls slowly until about 95 per cent of the crystallization has taken place. A fourth possible variable is the change in the ratio of femic CaO to MgO + FeO. This, in other words, is the ratio of clinopyroxene to orthopyroxene in the rocks developed. A glance at the norms of the Skaergaard and Stillwater successive liquids shows comparatively slight variation in this ratio with crystallization after an initial stage before clinopyroxene appears as settled crystals. In this initial stage the magma becomes slightly enriched in clinopyroxene with respect to orthopyroxene.

Thus the course of crystallization with strong fractionation may be summed up as resulting essentially in the plagioclase becoming more albitic and the pyroxene becoming more ferriferous as fractionation proceeds. The ratios of plagioclase to pyroxene and clinopyroxene to orthopyroxene are relatively undisturbed, and the production of new phases is quantitatively exceedingly small.

Although extreme differentiation has taken place the rocks and magmas produced for the first 95 per cent of the differentiation in the Skaergaard and perhaps 100 per cent of the Bushveld and Stillwater may be considered of basaltic or gabbroic composition. There is however relatively great variation in bulk chemical composition within the series.

b. MgO:FeO: TOTAL ALKALIES (WEIGHT PER CENT) DIAGRAM AS AN INDICATOR OF FRACTIONAL CRYSTALLIZATION

The $MgO:FeO:Na_2O + K_2O$ weight per cent diagrams furnish an exceedingly useful simplification of the trend of change in the magma which nevertheless embodies all the more important variations. It was used effectively by Wager and Deer (1939), Walker and Poldervaart (1949), Tilley (1950), Cornwall (1951), and Hotz (1953), and its qualities are described by Poldervaart (1949).

Although the trends of change should be considered from the point of view of changes in the magma with crystallization, sufficient information to calculate such a complete trend is available only for the Skaergaard. For extrusives, rapidly quenched

samples of original magmas may be obtained, but this is rarely the case for intrusives. Fortunately, although liquids cannot be calculated, rock analyses follow so closely the same trend that they may be used. Care must be taken only to choose analyses of rocks with normal feldspar : femic ratios. Those with greatly disturbed ratios such as analyses representative of a portion of a layer in a rhythmically layered rock cannot be used. Furthermore the early differentiates that are more magnesian and more anorthitic than the magma lie on an extension of the liquid trend projected downward in the general direction of the MgO corner. A comparison of the trends obtained from calculated liquids and from the actual rocks of the Skaergaard is given in Plate 11 A. The close parallelism of the trends may be noted as well as the trend of the early formed rocks towards the MgO corner. It will, therefore, be possible to consider the trends based on rock analyses of such igneous bodies as the Bushveld complex, the Duluth gabbro, the Stillwater complex, the Palisades diabase, etc.

Plate 11 B shows the MgO:FeO:total alkalies diagram for the Bushveld complex. This is based on selected rock analyses, but three points on the liquid line from the chilled border facies are also available. Its form is very nearly the same as for the Skaergaard. The Duluth gabbro trend is shown in Plate 11 D based upon rock analyses only. Again the trend is similar. The Stillwater trend is based on the initial magma as indicated by the analysis of the floor border facies and two calculated liquids computed by subtracting the composite analyses from the border-facies analysis. The trend obtained is identical with that found for the Bushveld. In the Skaergaard the abrupt change in trend toward the alkali corner takes place when only a few per cent of liquid remains. It seems somewhat doubtful whether the Stillwater or Bushveld trends will go much beyond the peak in iron enrichment. The entire intrusion may completely crystallize without the development of a granophyric residue because of the initial low K_2O content. The Duluth gabbro on the other hand shows a more abundant development of the granophyric facies than the Skaergaard, and the available chemical analyses suggest a higher initial K_2O content.

From E to H of Plate 11 are shown the MgO:FeO:total alkalies trends for thick dolerite sheets. The form of the curves is similar to those for the lopoliths, but the peak in iron enrichment is lower and comes slightly earlier in the differentiation. The trend from the initial magma to the peak is straight but less steep. The difference between the dolerite sheets and lopoliths is interpreted as a difference in the degree of fractional crystallization—there is less in the dolerites. In the sequence from E to H, the Tasmanian dolerites, Dillsberg sill, Palisades sill, and Nipissing sills also exhibit a decreasing degree of fractionation from left to right, and this is also a sequence of decreasing thickness. In the lopoliths fractionation was largely accomplished by crystal settling and in the dolerites probably by inward crystallization from the floor and roof plus a diffusion process discussed in Appendix 2.

In Plate 11, I and J represent volcanic suites in which fractional crystallization has been either slight or negligible. The beginning of the trend for the Hawaiian suite shows an initial slight rise which might be attributed to fractional crystallization. Points on the right side of the diagram below the curve probably represent magmas enriched in early formed olivine and/or augite crystals. It seems very unlikely, however, that the main trend which is directed toward the alkali corner can

be attributed to fractional crystallization. The Crater Lake suite trend shows no evidence of the effects of fractional crystallization. The points show considerable scatter and could perhaps be better represented by a band rather than the line given on Plate 11 J.

In Plate 11, L presents the MgO:FeO:total alkalies diagram for a typical calc-alkaline plutonic suite. This is the Garabal Hill-Glen Fyne sequence from Nockolds (1941). It very closely resembles the Crater Lake trend (Williams, 1942) and the calc-alkaline trend plotted from Daly's averages in K. Thus one should hesitate to call upon fractional crystallization to account for the sequence represented by calc-alkaline magmas.

c. DIFFERENTIATION OF BASALTIC VOLCANIC SEQUENCES, DIKES, AND THIN SHEETS

In oceanic-island volcanic suites, basaltic dikes, and in thin sheets the differentiation generally trends directly toward the alkalies corner. The question is: What process would produce such a trend? In the Hawaiian suite there certainly was some early crystal sorting whereby olivine and augite tended to be enriched in some fractions of the magma and extracted from others. This seems, however, to be a subordinate factor in the differentiation of the Hawaiian lavas. The clue to the main process operating is given in various papers by Walker (1930; 1935; 1952) and a recent paper by Vincent (1950). These investigations give the bulk analysis of a rock and the residual glassy fraction. The glass is always displaced from the rock in the direction of the alkali corner of the diagram (Plate 11 K, trends A and B). This means that, if a magma is allowed to crystallize for a period of time with the crystals remaining suspended in, and in equilibrium with, the liquid and if by some process the liquid and crystals are separated entirely or in part, then the two products lie along a line trending from the rock composition to the alkali corner. Some may regard this as fractional crystallization, but there has been only one fractionation, and for the most part the crystals have remained in the magma and in equilibrium with it until the separation into two portions occurred. This process seems characteristic of basaltic suites of oceanic islands. As Edwards (1942) points out, the differentiation products may be a function of the shape of the magma chamber. Under the oceanic islands one might postulate vertical magma columns rather thoroughly stirred by convection during the first half of crystallization so that the crystals remain in contact with the liquid for a long time. Filter pressing or quiescence to allow the crystals to settle might then produce a residual magma displaced in composition toward the alkali corner. Continental basalts tend to form sheets in which flow is restricted after emplacement, and fractional crystallization has a better chance to operate. In thin sheets or dikes where fractional crystallization does not have much opportunity to operate, any process which can draw off the residual liquid after partial crystallization will produce a residuum on the trend toward the alkali corner.

3. DISCUSSION OF ORIGIN OF BASALTIC MAGMA AND THE CALC-ALKALINE MAGMA SERIES BY SELECTIVE FUSION IN DEPTH

Hybridization may also account for some calc-alkaline igneous rocks. A minute amount of them may be the final product of extreme fractional crystallization. None of these processes seems satisfactory to account for the many large bodies of uniform

composition of calc-alkaline intrusives. Inasmuch as generation of basaltic magma represents a special case it will be discussed first, and members of the calc-alkaline series other than basalt later.

a. GENERATION OF BASALTIC MAGMA

The world-wide uniformity of primary basaltic magma has been somewhat exaggerated because the modal and normative minerals are always essentially pyroxene plus plagioclase. Actually considerable variation in Fe:Mg ratio may be noted. Many of the basalts of the Columbia Plateau are as high in total Fe and as low in MgO as the Stillwater residual magma after half of the complex had crystallized. Alumina may range from less than 13 to 19 per cent. The TiO_2 content may be from less than 1 to more than 3 per cent. High TiO_2 takes FeO from the magma to form ilmenite rather than having the FeO combine with SiO_2 to form pyroxene. This results in the appearance of free SiO_2 in the norm when a glance at total SiO_2 in the analysis might suggest an undersaturated rock.

On the other hand differences between the great flood basalts of continents and the primitive basalt of oceanic islands have been emphasized until recently probably because of the differences between the end products of differentiation of the two groups. Tilley (1950) compares the Koolau series of the Hawaiian islands to the New Jersey Triassic basalts and concludes that "the tholeiite series represents the primary magma group of the ocean basins".

The minimum depth of generation of basaltic magma may be estimated from its temperature on arrival at the surface. Present uncertainties as to the geothermal gradient in the Earth rule out setting a specific figure on the depth. Generally basaltic magma reaches the surface or near surface at a temperature slightly above 1100°C. Some measurements of lavas as high as 1200°C. have been reported. From the temperature it is at least clear that generation of basaltic magma takes place well below the M discontinuity. Some of the lower estimates of the geothermal gradient would place it deep within the mantle, perhaps hundreds of kilometers below the surface. Pressure raises the melting points of the silicates constituting basalt. This effect might be estimated to be about 3°C. per kilometer depth. Thus if basalts were generated at 100 km depth a temperature near 1400°C. would be required.

Bowen (1928) derives primary basaltic magma by *partial fusion* of feldspathic peridotite at depth. Ross, Foster, and Myers (1954) have recently studied peridotite nodules brought to the surface in basaltic volcanoes. These all consist of four phases: olivine (predominant), enstatite, chrome diopside, and chromite. The compositions of the four phases are practically identical whether from oceanic or continental volcanoes. The composition of the host volcanic rocks varies over a considerable range although the rocks are generally olivine or melilite basaltic types. Most of the xenoliths show fabrics that suggest strong deformation and solid flow (Ernst, 1936; Turner, 1942). The writer agrees with Ross, Foster, and Myers that the nodules are fragments of the material from below the M discontinuity and are therefore probably representative of some part of the Earth's mantle. It is interesting to note that the material has the same composition under continents and oceans. An analysis of a typical nodule is given in Table 50.

Whether this rock type is the material from which basaltic magma is derived by partial fusion or whether it may be the solid residue from partial fusion could be debated.

The great difference in composition between basalt and the xenoliths indicates that the percentage of partial fusion to give basalt liquid must have been very small.

TABLE 50.—*Peridotite xenolith from basalt, Ludlow, San Bernardino County, California*
Analyst, R. B. Ellestad

		Norm		
SiO_4	44.35	Ab 1.57	8.80 Pc	
Al_2O_3	2.97	An 7.23		
Fe_2O_3	.67			
		Wo 2.20	4.13 Di	
FeO	7.59	En 1.80		
MgO	40.80	Fs .13		
CaO	2.55			
		En 16.30	18.54 Hy	
Na_2O	.20	Fs 2.24		
K_2O	.01			
H_2O+	.06	Fo 58.85	67.11 Ol	
H_2O-	.03	Fa 8.26		
CO_2	.00	Mt .93		
TiO_2	.14	Il .30	1.83	
P_2O_5	.02	Cr .60		
Cr_2O_3	.41			
MnO	.13		100.41	
NiO	.31			
	100.24			

The percentage of alkalies in the liquid for example is more than ten times that of the xenolith. Thus if the xenoliths were the residue, addition of a few per cent of basalt to get the original bulk composition would change their composition very slightly.

While the above discussion gives the framework of the problem, there certainly are too many missing quantities to pursue it further at the present time.

b. GENERATION OF THE CALC-ALKALINE MAGMA SERIES

The generation of primary basaltic magma is a rather special case inasmuch as it is the only magma generated below the M discontinuity (unless certain peridotites may also be intruded as liquids). All other magmas of the calc-alkaline series are either generated above the M or are considered to be differentiation products or hybridization products of basaltic magmas. Where the M discontinuity is at shallow depth below the oceans (roughly 5 km below the sea floor), basalt is the only primary magma coming to the surface, and the other calc-alkaline rocks are absent. In the

writer's opinion most calc-alkaline rocks are derived from partial fusion of the continental crust, and so they should not be expected in oceanic areas (except for rare occurrences of minute amounts which can be attributed to differentiation of basalt). Kennedy and Anderson (1938) in a most important paper discuss the generation of magma in the Earth's crust. Their general idea is that downbending of the normally

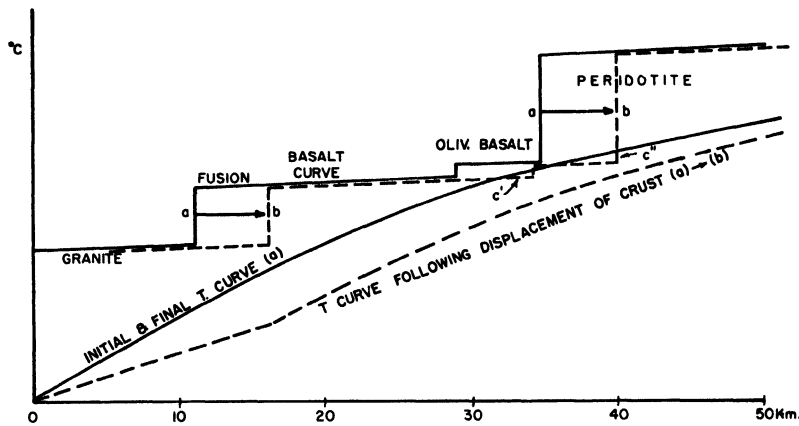


FIGURE 36.—*Effect on isotherms of displacement of crust from a to b and later rise of isotherms to cause partial melting at c'' and c'.*

horizontally layered crust results in bringing rocks from higher to lower levels. With bending of the crust the isothermal lines are also bent downward but after a considerable time lag return to an equilibrium position parallel to the Earth's surface. As this happens, certain layers of the crust will be brought above temperatures at which fusion begins. This concept is diagrammatically shown in Figure 36 in which displacement of the crustal layers downward from position (a) to position (b) is indicated by the solid and dashed lines respectively. At the same time the temperature curve is displaced from the initial solid line to the dashed line. As temperature equilibrium is re-established the temperatures again approach the solid curve and when this happens points *c'* and *c''* in the crust are above their fusion temperatures.

Partial fusion rather than complete fusion must be the rule in the crust, and this concept of a melting interval is incorporated in Buddington's (1943) consideration of the same problem.

Much the same result would be produced by any process that would tend to raise the relative positions of the isothermal planes in the crust. For example a heat source from below would have the same general effect as depressing crustal layers.

The continental crust is probably much more heterogeneous than the simple layered concept of structure by Kennedy and Anderson or Buddington. Nevertheless with increasing depth there must be in general a sequence from lower-melting to higher-melting rocks, from more acid to more basic, and from more felsic to more mafic. This does not change the situation qualitatively but would increase the num-

ber and decrease the height of the breaks in the fusion curves of Figure 36 except that the break at the M discontinuity would still be great.

Inasmuch as all the rocks of the crust must have a considerable melting interval, let us consider what would happen if the temperature were raised to a point at which melting began at some horizon. Even under the most drastic conditions the iso-

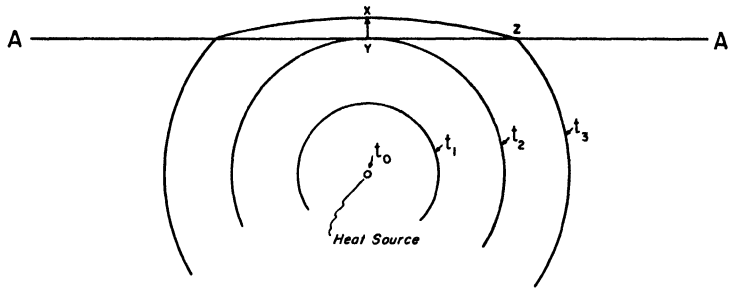


FIGURE 37.—*Diagram to illustrate slow rise of isotherms from Y to X vertically and rapid spread horizontally from Y to Z with partial melting beginning at level A-A*

thermal planes cannot depart very far from horizontality. In other words, the thermal gradients horizontally at any level must be exceedingly small (aside from vicinity of contacts of intrusions). Thus if partial fusion starts at one point there will be a strong tendency for the zone of partial fusion to expand laterally rather than for a limited volume to melt completely. The latent heat of fusion will be extracted at the point at which melting starts, thus retarding further rise of temperature, but to either side of that point the temperature will continue to rise until melting begins (Fig. 37). Assuming a deep heat source so that the isothermal lines rise in a regular manner as smooth curves, the retarding of their rise compared to the lateral spread of melting can be approximately calculated for such a case. If the specific heat (Cp.) of the rocks is 0.3 cal/gm and the latent heat of fusion of a layer at A-A is 90 cal/gm and we assume further that 2 per cent melts per degree C. rise in temperature, this would have a net effect similar to a specific heat above A-A equal to 1.8 cal/gm plus 0.3 cal/gm or 2.1 cal/gm. The temperature rise above A-A while melting proceeded would then be one-seventh as fast as below. The lateral migration of the beginning of melting along A-A depends largely on the shape or curvature of the isothermal lines as they approach A-A, so no specific figure can be given. Velocities 25 to 50 times the velocity of rise above A-A were obtained for several reasonable constructions.

Melting will start with the lowest-melting constituents of the rock, and as the temperature rises progressively more refractory materials will melt. However, the melting interval of most rocks is such that it seems unlikely that any particular zone can reach a stage of complete fusion. In other words, it seems unlikely that the temperatures are ever high enough to melt the last, perhaps 25 per cent, of the most refractory constituents of any given rock at any given level. The exception to this might be granitic rock at a rather moderate depth in the crust particularly in the presence of water.

Thus we might expect, as Bowen did, that basalt forms from partial fusion of feldspathic peridotite and that partial fusion of gabbroic rocks would produce diorite magma or, of diorite, granite magma.

In orogenic belts the time sequence of intrusion of calc-alkaline magmas is com-

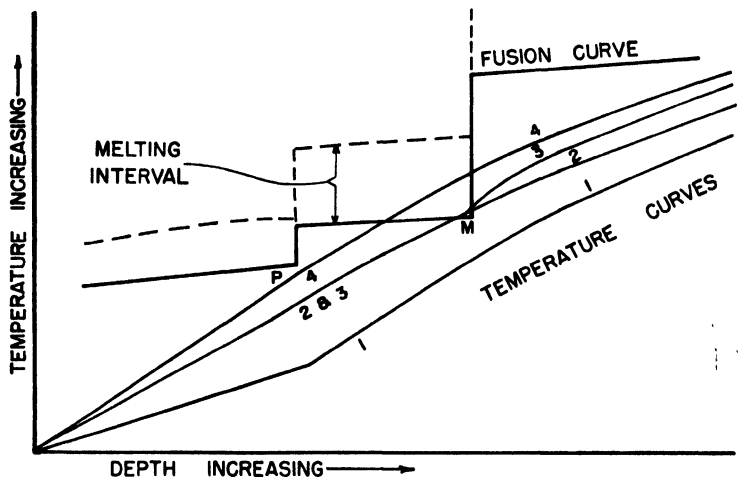


FIGURE 38.—*Effect of latent heat in retarding rise of isotherms at M*

monly from mafic to felsic. Commonly also in nonorogenic areas intrusions of basaltic magma are followed and overlain by granitic magma, as for example the Bushveld, Sudbury, and probably many others where the relationship has been ascribed to magmatic differentiation. The above associations are nicely accounted for by the process of generation of successive magmas by partial fusion. If a given area is disturbed by being bent downward as in orogenic belts, then a situation as illustrated in Figure 38 is developed. Temperature curve 1 indicates the relations immediately after the crust was deformed; it is taken from the similar curve in Kennedy and Anderson (1938). When the curve rises to position 2 partial melting starts at *M*. If the material at *M* melted at a fixed temperature, the temperature curve would be held at this point, while the heat of fusion was supplied to melt all the material at *M*. Curve 3 from *M* to the surface would not change during this interval. Substituting a rock with a melting interval at *M* would have the effect of not quite stopping the temperature rise but of retarding it seriously at this point. This would result in generation of magma at point *M* as a rule before any melting could take place at horizons above *M*. After the lag in rise of the curve produced by extraction of latent heat at *M*, the curve may rise above *M* and develop another site of partial fusion higher in the crust at *P*. The rise of the curve above *P* will be retarded as before, and so on. This will produce magmas at successively higher levels in the crust and might be expected to result in a magma sequence ranging from mafic to felsic. The process would operate in a nonorogenic area in much the same manner if a heat

source were available such as might be supplied by basaltic magma rising from the mantle or convective overturn in the mantle.

Two processes may act as brakes on the generation of separable magmas in the crust by partial fusion. In the first place partial fusion probably must go to at least one-third of the volume before a liquid can be separated from the crystal mush.[12] This also limits the compositions possible for that liquid. An upper limit on partial fusion, the refractory brake, might be placed in the neighborhood of 75 per cent liquid because temperatures will rarely reach a high enough value at any given level to melt the most refractory of the remaining crystals. Complete melting at one level would mean raising the temperature at that level by at least 250° or 300°C., and this seems excessive. Taking the above two limits into consideration an intrudable magma formed by partial fusion of gabbro cannot depart very far from quartz diorite in composition. The possibilities for variation in the composition of magmas produced by partial fusion lie in the variation in the depth at which melting takes place, and consequent variation in the rock type undergoing partial fusion, and in the variation in percentage melted between the upper and lower limits mentioned above.

The process of partial fusion permits of much less variability in the resulting liquids than does crystallization differentiation. There can be no differential sorting of crystals for example. The liquids at the time of removal from the crystal mush would have been in equilibrium with the crystals. In crystallization differentiation the crystals can be continuously or repeatedly removed from contact with the liquid (or armoured to achieve the same result) so that they may not react with the liquid. In this manner the liquid can be greatly changed in composition. During the process of partial fusion, liquids cannot progress along such lines. The products therefore should be, in general, less varied.

The compositions of the products of partial fusion will be exactly analogous to the cases of residual glasses in dolerites discussed by Walker (1930; 1935), Walker, Vincent, and Mitchell (1952), and Vincent (1950). Crystallization of 60 per cent of a basaltic magma without fractionation, for example, leaves 40 per cent of residue which would have the same composition as could have been obtained by partially fusing a solid dolerite of the same composition up to the point where 40 per cent was liquid.

To compare fractional crystallization and partial fusion let us suppose that a rock similar in composition to the Border Zone of the Stillwater complex—a dolerite—were to undergo partial fusion at depth. The MgO:FeO ratio of the rock is close to 60:40. The first pyroxene crystals to melt (referring to data by Bowen and Schairer for synthetic pyroxenes) would have an MgO:FeO ratio of 35:65, but this would increase to 45:65 by the time enough liquid was formed to be removable from the crystal mush. The final ratio for complete fusion would be the same as the original rock, but the process of partial fusion might well be expected to be stopped by the refractory brake when a ratio of 55:45 was reached. Thus the total range of liquids

[12] This may not be true for generation of basalt at relatively great depth. The liquid probably would not be removable so long as the crystals form a rigid framework. At high temperatures, high confining pressures, and high stress when crystals themselves would be subject to solid flow a small percentage of liquid might be extracted. If basalt represented only a minute fraction of partial melting of a peridotite such as given in Table 50, the chemical relations would be more easily reconciled.

which might be intruded into higher horizons would lie between MgO:FeO ratios of 45:55 to 55:45 or only a total variation of about 10 per cent. Referring to the Stillwater-Skaergaard trends, a magma of the same composition as the rock fused in the above discussion could produce crystallization differentiates ranging in MgO:FeO ratio from 90:10 to almost 0:100, or a total range of almost 90 per cent. An entirely analogous situation exists as regards the An:Ab ratio with partial fusion as compared to fractional crystallization. Thus it is evident that the liquids formed by a process of partial fusion will vary much less than those produced by fractional crystallization.

Larsen and Switzer (1939) describe a granodiorite xenolith which has undergone partial fusion. The analyses of the whole xenolith and the glass fraction are given. The trend on the MgO:FeO:total alkalies diagram from the rock to the glass is shown on Plate 11 K. It lies almost exactly on Daly's calc-alkaline trend given in the same diagram. This then is a specific example of the results of partial fusion.

A series of liquids formed by partial fusion at varying depths within the crust, and hence varying compositions of the rocks fused, could, it appears, produce the calc-alkaline magma series. On the MgO:FeO:total alkalies diagrams the calc-alkaline series represents a trend suggesting little or no crystal fractionation, and it has been shown that liquids residual from crystallization without fractionation of basalt exhibit a trend comparable to that of the calc-alkaline series. The partial fusion of granodiorite (Larsen and Switzer) produced a more acid magma farther along on the calc-alkaline trend. Here then is a process which is relatively simple in operation and capable of producing rocks of the compositions required. There can be little doubt that it is the explanation for some calc-alkaline rocks, and the writer is inclined to believe that it is the major means by which such rocks are produced. If this is true, then most large bodies of magma which are intruded into high levels of the Earth's crust represent *primary* magmas in the sense used by Bowen.

4. GENERATION OF MAGMAS AND DEVELOPMENT OF A COMPOSITION GRADIENT IN THE CRUST

In the past seismologists have favored a two- or three-layered concept of the continental crust above the M discontinuity. Today most seismologists would retain the Conrad discontinuity separating the crust into an upper "granitic" layer and a lower "basaltic" layer. The validity of this simple arrangement has been questioned by Tatel, Tuve, and Hart (1954). Whether the somewhat higher wave velocities toward the base of the crust gradually decrease upward or are separated from a lower-velocity layer by a sharp discontinuity is very difficult to establish from seismic data.

It seems likely that after deformation, such as in mountain belts, the whole crust will be strongly disturbed and kneaded together as a heterogeneous mixture much the same as it appears at the surface in the Archean terranes of Precambrian shields. This mixture will gradually become adjusted to the normal thermal gradient in the crust. Lower-melting constituents will move upward and thereby a type of equilibrium will be established such that all the rocks will be at depths above that temperature at which their melting begins. Some may, of course, be far above their melting

temperature if magmas that were formed near the bottom of the crust move to a level at or near the Earth's surface. This would be true of all rocks in the upper 15 km or more of the crust where the thermal gradient steepens rapidly as the surface is approached. The result will be a crust progressively impoverished downward in low-melting constituents, but it will still be heterogeneous. It can hardly be a simple layered crust such as postulated in preceding pages for the sake of clarity in discussing partial melting, or a crust of two or more distinct layers as mentioned above. The equilibrium can be disturbed by rise of temperature as a result of influx of heat from below so that the temperature for a given level exceeds the temperature of the thermal gradient with which the crust was originally in equilibrium. Magmas would again be generated. Strong deformation may fold rocks downward to depths greater than those at which the equilibrium was established, and once more magmas can be generated.

How far this process is operative is a function of the thermal gradient in the continental crust. If the gradient is close to present estimates, a good qualitative picture has been presented. If the gradient were much less steep then the process would have operated to a much lesser extent.

Appendix 2. Differentiation of Dolerite Sills

The rocks formed in common dolerite sills are chemically and mineralogically similar to such intrusions as the Stillwater and Bushveld. The differentiation also follows a similar trend in all but the thinnest sills, although with less extreme fractional crystallization. In general the explanations given for the differentiation are based to a large extent on the assumption that crystal sorting has taken place by settling or rising of crystals in the magma. In the cases of unusually thick sills such as that at Insizwa (Scholtz, 1936, about 3000 feet thick) good evidence of settling of olivine can be seen. In sills less than 1000 feet in thickness indications of crystal settling or rising are very rare. The Palisades sill (Walker, 1940), which is just less than 1000 feet thick, shows evidence of crystal settling clearly only at the thin olivine-rich layer just above the base and here only of the larger olivine crystals. Igneous lamination and rhythmic layering which are characteristic of the Stillwater complex and bodies of lopolithic dimensions are not observed in ordinary dolerite sills, nor do their rocks have the textures typical of crystal accumulates. The absence of igneous lamination seems to be particularly significant for such dimensional orientation of the crystals would seem to be inevitable if they were products of crystal sedimentation on the floor of a magma chamber. The writer has, therefore, been led to look for some other process by which fractional crystallizaton could be accomplished without resorting to settling or rising of crystals in the dolerite sills. An analogy between the crystallization of the dolerites and the crystallization of the pore-space liquid in the mush on the floor of the Stillwater immediately comes to mind. The interstitial liquid in its crystallization follows very closely the same trend as do the settled crystals. This is probably accomplished by diffusion of ions between the pore-space liquid and the near-by overlying main body of magma. The question arises: Could not this process be operative by itself without the necessity of a surrounding mush of settled crystals?

To follow this line of reasoning let us consider what might happen at the junction of the magma and the floor during the solidification of a dolerite sill. A thin layer of chilled dolerite will as a rule form along the floor contact; it will correspond in bulk composition to the original magma less perhaps some of its dissolved volatiles. At the top of this layer it might reasonably be postulated that crystals will begin to grow attached to the floor surface. These will be richer in the refractory constituents than the magma—plagioclase richer in An and pyroxene richer in Mg. At the same time a film of magma around them will be relatively impoverished in the same constituents. Diffusion may then begin to operate in much the same manner as in the case of the pore-space liquid of the Stillwater crystal mush. The net effect of diffusion will be a tendency to minimize the differences in composition between the film impoverished in refractory constituents and the overlying main mass of magma. For this process to make much progress in the way of fractional crystallization, it is necessary that the overlying magma be circulating and that the products of diffusion be thoroughly mixed with the main mass of magma. The circulation itself may be able to sweep away the liquid impoverished in refractories although it seems likely

that diffusion must play some part in moving the materials at least a short distance from the crystals because high viscosity might be expected immediately around the crystals as a result both of proximity to the cooling surface and relative enrichment in alkalies. The next layer of crystals to form on top of the older ones would effectively remove the latter from reaction with the magma so that fractional crystallization can be accomplished in the same manner that it was in the case of settled crystals. The bulk composition of the residual magma would be continually changing and would become enriched relatively in Ab and Fe as well as in those constituents which did not enter the earlier formed crystals. It seems perfectly possible and reasonable to suppose that a magma could crystallize inward from its walls and give a final product when solidified that would qualitatively resemble the products of differentiation by crystal settling.

The process might also be expected to operate in exactly the same manner from the roof downward. The writer has found this to be the case in several dolerite sills that he has examined, although crystallization from the roof downward is considerably slower than from the floor upward. There are two reasons for this: (1) The increase in pressure downward slightly favors crystallization at the floor, and (2) convective circulation of the magma tends to keep the hottest portion directly below the roof so that crystallization along the roof contact may be slowed even though more heat is as a rule lost from the upper than from the lower contact. Most descriptions of dolerite sills show that the final residuum of the magma crystallizes about three quarters of the way up between the floor and roof which is an indication of the rate of crystallization inward from both contacts. The compositions of the crystals and sequence of phases separating from the roof downward duplicate in condensed form the sequence from the floor upward. This is well shown in a study of the Lambertville, New Jersey, dolerite sill by Jacobeen[14] (Fig. 39). The sill is 1800 feet thick and at this size some effects of crystal settling might be expected to appear.

In very thin sills or in the border zones of thicker sills rapid heat loss and consequent undercooling result in great increase in viscosity thereby preventing magma circulation. In the initial stages of cooling of a thick sill, heat loss from the floor and roof are the same. It is only after the heat from the roof begins to reach the surface of the Earth that the thermal gradient above the roof becomes appreciably steeper than the gradient below the floor. In the case of a thin sill at moderate depth, the heat conducted away through the roof may not begin to reach the surface until the sill has completely solidified. The energy for convective circulation is mainly derived from the disequilibrium resulting from greater heat loss through the roof than the floor. Thus in thin sills or in the initial stages of cooling of thick sills conditions favorable to convection are lacking. The thickness of the sill, its depth below the surface, and the thermal properties of the wall rocks are, of course, all factors to be considered in this problem. Undercooling and the lack of magma circulation in the circumstances cited above result in the production of a fine-grained, uniform dolerite mass having essentially the same composition as the original magma. Fractional

[14] The Lambertville sill was studied by F. Jacobeen along the east bank of the Delaware River. This sill is probably a westward extension of the Palisades sill. (Bachelors thesis, Princeton Univ., 1949).

crystallization is inhibited. At any given stage in the crystallization the crystals suspended in the magma or growing on the walls differ in composition from the enclosing liquid. But, so long as the crystals remain immersed in the magma from which they formed, they will continually change in bulk composition as crystallization proceeds and when completely crystallized they will have a bulk composition

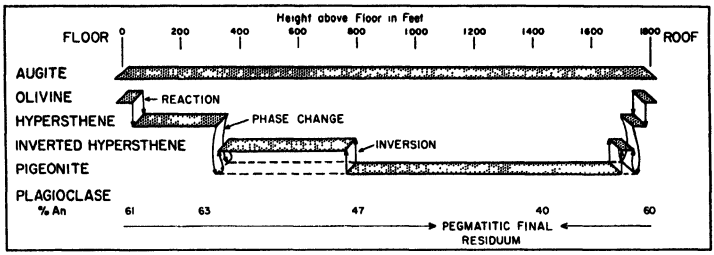


FIGURE 39.—*Crystallization of the Lambertville sill*
After F. H. Jacobeen, Jr., Bachelors thesis, Princeton University, 1949

the same as that of the original magma. In a small sill, if heat loss and consequent crystallization is too rapid, diffusion may be too slow to be effective. Thus this factor as well as lack of convective circulation may play a role in preventing differentiation.

In certain cases the cooling of a dolerite may result in solidification as a mass of crystals with a mesostatis of glass. This glass may be separated and analyzed as has been done by Walker (1930; 1935), Vincent (1950), and Walker, Vincent, and Mitchell (1952) as discussed in Appendix 1. It gives the composition of the residual liquid after partial crystallization without (as a rule) fractionation. The glasses from such rocks indicate a very different sort of change in the successive residual liquids from the changes which result from fractional crystallization of the same type of magma.

Returning once more to the process of fractional crystallization suggested above for dolerite sills, the writer emphasizes that diffusion of ions by itself unaccompanied by circulation of the magma would be quite ineffective in producing fractional crystallization. Diffusion alone would be the old Soret principle which Bowen (1921) has shown would be much too slow to produce the desired results. Diffusion for very short distances, of the order of magnitude of a centimeter or less, plus sweeping away of the diffused ions by circulation of the magma, a modified Soret principle, might work.

Differentiation by inward growth from the walls with diffusion of ions and circulation of the magma would differ in its results from differentiation predominantly by crystal settling in the following respects: (1) The dimensional orientation of crystals characteristic of sedimentation of crystals on a floor would be lacking. (2) The characteristic textures of crystal accumulates—peri-euhedral crystals surrounded by anhedral grains from the crystallization of the pore-space liquid—would be absent. (3) Size or density sorting would not be found. (4) The changes in crystalline phases

and compositions of the phases from the floor upward to a point from two-thirds to three-quarters of the way to the roof would be duplicated in sequence from the roof downward. The section from the roof downward would be much less thick than from the floor upward and might be modified somewhat by upward streaming of late-stage volatile constituents.

Appendix 3. Optical Properties of Low-Temperature Plagioclase

By J. R. Smith

A. Introduction
B. Acknowledgments
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   2. Implications and limitations of the data
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A. INTRODUCTION

During his detailed mineralogical study of the Stillwater igneous complex, Hess found that existing determinative curves for plagioclases failed to predict the compositions of chemically analyzed calcic plagioclases with the accuracy necessary for his purposes. He therefore made careful measurements of the optical properties of several samples of calcic plagioclases, with the object of defining more accurate determinative curves.

Since the time Hess began this work, it has been discovered that plagioclases belong to at least two different groups: a high-temperature group, which includes plagioclases synthesized in the dry way and some natural plagioclases of volcanic and hypabyssal origins, and a low-temperature group, which includes plagioclases from some plutonic igneous rocks, pegmatites, and metamorphic rocks. The distinction between the two groups has been made on the basis of optical and structural differences, many of the details of which are not yet known (Tuttle and Bowen, 1950). The discovery of these differences has made it important that the variations of physical parameters with composition in plagioclases of diverse origins be closely studied, partly to elucidate the nature and occurrence of the different modifications, and partly to establish determinative curves for them. The most widely used existing determinative curves, those based on optical properties, must be looked upon as representing averages of the properties of plagioclases of many different origins. Furthermore, the scatter of the points through which the curves have been drawn is in some cases greater than the differences between the low- and high-temperature modifications. Part of this scatter probably represents real differences between the samples used, but an indeterminate part of it is due to errors in the optical determinations and chemical analyses.

With these considerations in mind, the writer undertook to work with Hess in extending his optical studies to more sodic plagioclases, using samples chosen on the basis of similarity of thermal history, insofar as this was possible. Samples covering the entire composition range have now been studied; these include albite and oligoclase from pegmatites and granites; andesine, labradorite, and bytownite from the Stillwater igneous complex and other thick stratiform mafic intrusions; and anorthite from metamorphic rocks. All the samples have been chemically analyzed, most of them by analysts who are well known for their skill. The optical measurements were made by carefully tested techniques, and the properties of numerous grains from each analyzed sample were averaged to take into account variations of composition within the samples.

The results of the study indicate that the samples used do not all belong to one truly isomorphous series, which is in accord with the results of X-ray studies by others. Some ambiguity remains in the definition of "low-temperature" plagioclases and, accordingly, in the curves of optical properties for such plagioclases. The optical data presented here should afford a basis for comparison with plagioclases of other origins when similarly precise data become available for them, and by such comparisons the ambiguities may be resolved.

B. ACKNOWLEDGMENTS

The achievements of this study depend largely on a diligent application of carefully tested techniques of optical measurements. These techniques were developed by Hess, and in part modified by the writer. Hess instructed and supervised the writer in the laboratory work and painstakingly studied the data in order to criticize the conclusions drawn. Responsibility for the general interpretation of the data lies with the writer.

All the costs of the study, including the writer's livelihood during its course, were met by grants from the A. H. Phillips and the Higgins Trust funds of the Department of Geology of Princeton University.

Many of the plagioclase samples were taken from specimens in the Princeton Mineralogical Collection. Most of these specimens were collected by Hess and coworkers from the Stillwater igneous complex and by Hess from the Bushveld igneous complex and from the Great Dyke. Four specimens, two of them analyzed, of plagioclase collected from the Skaergaard intrusion by Professor L. R. Wager of Oxford University were kindly sent by him to us for study. Dr. C. S. Ross of the United States Geological Survey and Professor R. C. Emmons of Wisconsin University generously contributed excellent samples with superior analyses. Other contributors are listed in the text.

The success of studies of this nature depends as much on the accuracy of the chemical analyses as on the accuracy of the optical measurements. In the present study, one interesting fact is that, in a consistent interpretation of the data, the error that need be attributed to the chemical analyses is very small. The skill and diligence of the analysts at the Rock Analysis Laboratory of the University of Minnesota, who are responsible for most of the previously unpublished analyses presented here, is thus gratefully acknowledged; the same credit is due the other analysts involved.

C. DESCRIPTION AND COMPOSITION OF SAMPLES

1. PURITY OF THE ANALYZED SAMPLES

Ten of the 12 samples in the composition range An_{44} to An_{100}[15] were purified for analysis at Princeton. The separations were made first by numerous runs through a Frantz magnetic separator, which removed all the ferromagnesian minerals, as verified by counts of 1000 grains mounted in oils. If apatite was found in the counts, it was removed by sinking it in tetrabromoethane. Most of the plagioclase samples contain a small amount of included material, mostly alteration products, which was reduced to a minimum by repeated runs through the Frantz separator at low tilts and maximum field current. Samples 354, 80165, and 97490 were purified elsewhere; the fractions of the analyzed samples which were studied optically are almost free of impurities and alteration products.

Two of the Skaergaard andesine samples, numbers 4145 ($An_{36.5}$) and 2580 ($An_{40.2}$), were purified and analyzed by Wager and Deer (1939). The optical studies were done on plagioclase separated from chips of the rocks from which the analyzed samples were taken. These two plagioclase samples contain abundant alteration products, but the rarity of occurrence of andesine in stratiform mafic intrusions made it imperative to include them in the study.

[15] All compositions are stated in mol per cent anorthite.

The purity of eight of the samples in the range An_0 to An_{80} is discussed by Emmons *et al.* (1953, p. 17); the samples meet very high standards except for appreciable amounts of quartz in samples 73, 156, and 94; quartz impurities do not affect the

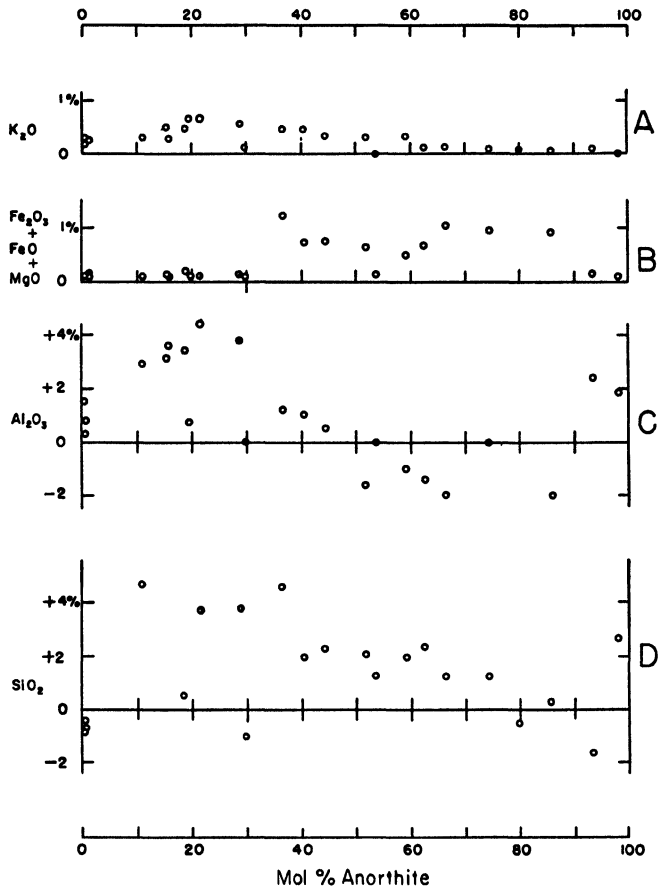

FIGURE 40.—*Deviations from the ideal formula in amounts of certain oxides in the plagioclase samples*

A. Per cent K_2O present in samples versus mol per cent An.

B. Per cent (Fe_2O_3 + FeO + MgO) present in samples versus mol per cent An.

C. Excess (+) or deficiency (−) of Al_2O_3, calculated as per cent of amount of Al_2O_3, versus mol per cent An.

D. Excess (+) or deficiency (−) of SiO_2 calculated as per cent of amount of SiO_2, versus mol per cent An.

results, because the anorthite content of the samples was calculated from the lime and alkali determinations only.

A partial measure of the maximum amount of ferromagnesian impurities which might be present in the samples is given by the amount of Fe_2O_3 + FeO + MgO

in the analyses of the samples; the sum of the percentages of these oxides present in the samples is plotted against mol per cent An in Figure 40 B. They are present in very minute amounts in the samples from granites, pegmatites, and metamorphic

TABLE 51.—*Source of plagioclase samples*

Collector's number	Mol % An	Collector (Contributor)	Locality	Rock type
D-761-1	0.2	C. S. Ross	Ramona, Calif.	Pegmatite
D-761-2	0.4	C. S. Ross	Pala, Calif.	Pegmatite
D-761-3	0.4	J. J. Glass	Amelia, Va.	Pegmatite
10	11.0	Wards (R. C. Emmons)	Peekskill, N. Y.	Two-mica granite
73	15.4	Wards (R. C. Emmons)	Parishville, N. Y.	Biotite granite
156	15.9	S. E. Clabaugh (R. C. Emmons)	Llano Co., Texas	Granite
94	18.7	R. C. Emmons	Tigerton, Wis.	Granite
80165	21.6	F. C. Kracek	Bakersville, N. C.	Pegmatite
97490	28.7	F. C. Kracek	Mitchel Co., N. C.	Pegmatite
D-638	29.7	C. S. Ross	Macon Co., N. C.	Pegmatitic vein in dunite
4145	36.5	L. R. Wager	Kangerdlugssuak, Greenland	Ferrogabbro, Skaergaard intrusion
2580	40.2			
1963	44.2			
C-41	50	A. P. Subramaniam	Salem district, India	Metagabbro
GD 29	51.8	H. H. Hess	S. Rhodesia	Dolerite, Great Dyke
GD 31	55.5	H. H. Hess	S. Rhodesia	Dolerite, Great Dyke
BV 63	59.2	H. H. Hess	Transvaal, South Africa	Gabbro, Bushveld igneous complex
EB 41	62.5	H. H. Hess	Montana	Gabbro, Stillwater igneous complex
EB 42	66.2	H. H. Hess	Montana	
EB 43	74.3	H. H. Hess	Montana	
EB 38	80.0	H. H. Hess	Montana	Norite, Stillwater igneous complex
EB 18	85.8	H. H. Hess	Montana	Norite, Stillwater igneous complex
354	93.3	F. S. Miller (E. S. Larson)	Pala Mountain, Calif.	Hornblende-olivine norite
C 50	98.0	A. P. Subramaniam	Salem district, India	Meta-anorthosite

rocks, whereas the amounts are significantly higher in the samples from the stratiform mafic intrusions. This may mean that iron and/or magnesium entered the plagioclase structure at the higher crystallization temperatures of the mafic intrusions, and that the iron oxides and magnesia do not represent impurities in the ordinary sense of the term.

2. K_2O CONTENT OF SAMPLES

The percentage of K_2O in the samples is plotted against mol per cent anorthite in Figure 40A. A slight increase in the amount is noticeable in succeedingly more sodic samples from the stratiform mafic intrusions (An_{86} to An_{36}). Three of the 10 sam-

ples from granites and pegmatites (An_0 to An_{30}) contain slightly more K_2O than the most sodic samples from mafic intrusions, but the others contain as little or less. K_2O is absent from samples C41 (An_{50}) and C_{50} (An_{98}), which are from metamorphic rocks. K_2O is assumed to be present as K^+ in the plagioclase structure; the total amount present and the relative changes are probably too slight to affect the optical properties to a detectable extent.

3. SOURCE OF THE SAMPLES

Table 51 is a summary of the sources of the samples. For convenient reference, the compositions of the samples, which are discussed in the next section, are also listed.

4. CHEMICAL ANALYSES

Most of the chemical analyses were made by analysts who are well known for their skill. As may be seen in Table 52, the totals representing the percentage of the sample weight accounted for are in all cases very close to 100 per cent.

Table 53 lists the molecular proportions of the major oxides in each sample. These were calculated from the weight percentages of the oxides as reported by the analysts, using the tables of Burri and Niggli (1945). Table 53 also shows the excess or deficiency of the molecular proportions of SiO_2 and Al_2O_3 as compared to the proportions required by the alkalies and lime in an ideal AnAbOr formula. The excess or deficiency of SiO_2 and Al_2O_3 is plotted as percentage of the amount present versus mol per cent anorthite in D and C of Figure 40. The trends may be significant, but the significance is not obvious in view of the fact that the amounts of SiO_2 and Al_2O_3 required by the Fe_2O_3, FeO, and MgO (Fig. 40B) are not known; also, small errors in the alkali determinations would result in six times as much excess or deficiency in the molecular proportions of SiO_2.

The An contents of the samples listed in the last column of Table 53 were calculated from the expression

$$An = \frac{CaO}{CaO + 2(Na_2O + K_2O)} \times 100.$$

where CaO, Na_2O, and K_2O are the molecular proportions of the oxides listed in Table 53. The product of this expression is the same as that of

$$\frac{Ca^{++}}{Ca^{++} + Na^+ + K^+} \times 100.$$

An is therefore the molecular per cent of anorthite in the ideal feldspar molecule, and the compositions of the samples are so quoted throughout this report. The compositions as quoted are subject only to the errors in the lime and alkali determinations; if the sum of the absolute values of the errors in the lime and alkali determinations were $|2\%|$ of the amount present, the error in the quoted composition would be ± 0.5 mol per cent An at An_{50} and less near either end of the series. If the errors in the lime and alkali determinations were both positive or both negative they would partially cancel each other. The fact that very little error need be attributed to any

TABLE 52.—*Chemical analyses of plagioclase*

	D-761-1[a]	D-761-2[a]	D-761-3[a]	10[b]	73[b]	156[b]	94[d]	80165[e]	97490[e]	D-638[e]	4145[f]	2580[f]
SiO_2	68.05	68.40	68.17	66.30	65.94	70.79	75.75	62.54	60.85	60.38	59.03	58.03
Al_2O_3	19.73	19.58	19.62	21.22	21.44	18.40	15.20	23.31	24.44	24.88	25.02	25.81
Fe_2O_3	0.10	0.07	0.08	0.04	0.08	0.02	0.10	0.03	0.06	0.07	0.75	0.38
FeO	—	—	—	0.03	0.04	0.05	0.10	0.04	0.04	—	0.45	0.30
MgO	Tr	Tr	Tr	0.02	0.02	0.02	0.01	0.03	0.03	Tr	0.02	0.04
CaO	0.05	0.08	0.08	2.21	3.03	2.66	2.53	4.38	5.81	6.31	7.14	8.01
Na_2O	11.69	11.58	11.59	9.85	8.99	7.70	5.80	8.33	7.59	8.12	6.73	6.47
K_2O	0.18	0.30	0.28	0.27	0.46	0.25	0.46	0.66	0.59	0.12	0.46	0.46
H_2O+	0.10	0.16	0.11	0.03	0.00	0.02	0.10	0.40	0.43	—	—	—
H_2O-	0.03	0.03	0.01	0.01	0.05	0.03	0.00	0.20	0.25	0.09	—	—
TiO_2	0.01	0.01	0.01	0.01	.03	0.01	0.03	0.00	0.01	0.01	—	—
MnO	—	—	—	0.00	Tr	0.00	0.01	0.01	—	0.03	—	—
SrO	—	—	—	0.06	0.06	0.05	0.06	—	—	—	0.28	0.41
Totals....	99.94	100.21	99.95	99.82	100.15	100.01	100.19	99.93	100.10	100.08	99.88	99.91
Analyst	R. E. Stevens			Lee Peck				E. Chadbourn		R. E. Stevens	W. A. Deer	

	1963	C 41	GD 29	BV 63	EB 41	EB 42	EB 43	EB 38	EB 18	354[g]	C 50[h]
SiO_2	56.91	54.90	55.24	53.58	52.95	51.34	49.19	47.67	46.34	43.77	43.88
Al_2O_3	26.60	28.72	27.63	29.07	29.48	29.98	31.82	33.46	33.36	36.11	36.18
Fe_2O_3	0.65	0.08	0.09	0.08	0.24	0.55	0.55	—	0.54	0.09	0.08
FeO			0.35	0.28	0.32					—	—
MgO	0.10	0.08	0.18	0.12	0.10	0.48	0.41	0.35	0.38	0.70	0.00
CaO	8.94	11.07	10.55	12.04	12.70	13.39	14.95	16.23	17.31	18.73	19.37
Na_2O	6.02	5.24	5.22	4.33	4.12	3.70	2.78	2.19	1.55	0.67	0.22
K_2O	0.36	0.02	0.32	0.36	0.13	0.15	0.09	0.07	0.05	0.11	0.00
H_2O+	0.11	0.04	0.21	0.24	0.10	0.35	0.28	—	0.23	0.40	0.28
H_2O-	0.01	0.00	0.04	0.04	0.00	0.02	—	—	—		0.08
TiO_2	0.11	0.01	0.06	0.05	0.05	—	—	—	—	—	0.00
MnO	(P_2O_5) 0.10	—	—	—	—	—	—	—	Trace	—	—
SrO	0.05	—	—	—	—	—	—	—	—	—	0.01
Totals....	99.96	100.16	99.89	100.19	100.19	99.96	100.07	99.97	99.76	99.95	100.10
Analyst.	E. H. Oslund					A. H. Phillips Alkalies checked by R. B. Ellestad				F. A. Gonyer	Oslund

[a] Analyses provided by C. S. Ross to Emmons *et al.* (1953, p. 19)

[b] Analyses as quoted in Emmons *et al.* (1953, p. 18). Spectrographic determinations not listed above but included in the totals: Sample 10—Ba 0.007; Li 0.0003; Rb 0.0003. Sample 73—Ba 0.007; Li 0.0004; Rb 0.0008. Sample 156—Ba 0.011; Li 0.0007; Rb ND

[c] Analyses provided by C. S. Ross to Emmons *et al.* (1953, p. 19). Determinations not listed above but included in total of Sample D-638: Li_2O 0.02; ZrO_2 0.02; BaO 0.01; SrO .02

[d] Analyses as quoted in Emmons *et al.* (1953, p. 18). Spectrographic determinations not listed above but included in the totals: Sample 94—Ba 0.038; Li 0.002; Rb 0.0005

[e] Analyses as quoted in Kracek and Neuvonen. (1952, Table 3, columns 10 and 11)

[f] Analyses as quoted in Wager and Mitchell (1951, Table B following p. 208). Spectrographic determinations of trace elements are not listed above nor included in the totals

[g] Analysis as quoted by Miller (1935, p. 143)

[h] Analysis as quoted by Subramaniam (1956, Table 8, column 8)

of the compositions calculated from the chemical analyses is more easily understood in the light of these considerations.

D. DETERMINATIVE METHODS AND THEIR ACCURACY

1. GENERAL STATEMENT

The methods of determining refractive index and birefringence used in this study were arrived at only by time-consuming experimentation, much of which was done

TABLE 53.—*Recalculation of chemical analyses of plagioclase*

Note: Columns 1 to 5 are molecular proportions of oxides. Columns 6 and 7 are proportions of SiO_2 and Al_2O_3 over (+) or under (−) the proportions required by the feldspar molecule.

	1 SiO_2	2 Al_2O_3	3 CaO	4 Na_2O	5 K_2O	6 $\pm SiO_2$	7 $\pm Al_2O_3$	8 mol % An
D-761-1	1133	194	0.9	188	2	−9	+3	0.2
D-761-2	1139	192	1.4	187	3	−4	+0.6	0.4
D-761-3	1135	193	1.4	187	3	−8	+1.6	0.4
10	1104	208	40*	159	3	+52	+6	11.0
73	1098	211	54.5*	145	5	+89**	+6.5	15.4
156	1179	181	48*	124	2.5	+324**	+6.5	15.9
94	1261	149	45.5*	93.5	5	+578**	+5	18.7
80165	1040	229	78	134	7	+38	+10	21.6
97490	1011	240	103	122	6	+37	+9	28.7
D-638	1006	244	112	131	1	−10	0	29.8
4145	983	246	130*	108	5	+45	+3	36.5
2580	966	253	146.5*	104	5	+19	+2.5	40.2
1963	947	261	159.5*	97	4	+22	+0.5	44.2
C 41	914	282	197	84.5	0.2	+12	0	53.7
GD 29	920	271	188	84	3.5	+19	−4.5	51.8
BV 63	892	286	215	70	4	+18	−3	59.2
EB 41	881	290	226	66.5	1.5	+21	−4	62.5
EB 42	855	294	239	59.5	1.5	+11	−6	66.2
EB 43	819	312	266	45	1	+11	0	74.3
FB 38	793		290	35.5	0.7	−4		80.0
EB 18	771	327	308	25	0.5	+2	−6.5	85.8
354	728	354	334	11	1	−12	+8	93.3
C 50	731	355	345	3.5	0	+20	+6.5	98.0

* Includes SrO calculated as CaO

** Free quartz in sample

by Hess before the writer undertook to continue the optical work. For the sake of consistency in whatever personal bias has entered the results, all of Hess' original determinations were checked by the writer. During the 3-year course of the study, Hess frequently reread measurements made by the writer. No evidence of personal bias was found in the reruns. Any small instrumental bias in the results is also as nearly as possible consistent throughout the study, since the same instruments and standards were used throughout. Comparisons of measurements made by other laboratories on some of the samples used show acceptable consistency.

In each of the determinative methods used, special care was taken to minimize

nonrandom errors which would bias the means. Error of this type is believed to have been practically eliminated, except from the standards used in the refractive-index determinations, in which it is probably very small.

2. REFRACTIVE-INDEX DETERMINATION

Refractive indices were measured on grains from the crushed samples used for the chemical analyses, by a controlled-temperature universal-stage technique. The essence of the method is this: A grain, illuminated in sodium light, is rotated from a principal orientation toward another principal orientation until the refractive index of the grain exactly matches that of the oil in which it is immersed, at which point it disappears if it is colorless and transparent. Knowing the index of the oil, the angle of rotation from the principal orientation, and the birefringence of the section of the mineral perpendicular to the axis of rotation, two principal refractive indices of the grain can be calculated. If $2V$ is also known, the third principal index of refraction can be calculated. The detailed procedure necessary to make the method practicable and precise is described in the following paragraphs.

In mounting the grains, the difficulty of preventing the grains from rolling when the universal stage is tilted was overcome in the following manner: A thin film of glycol phthalate was applied to an ordinary glass slide by heating a fragment of glycol phthalate gently on the slide over a bunsen burner, and spreading it evenly with a stirring rod after it melted. After allowing the slide to cool until the glycol phthalate could just be dented by a knife point, the grains of the sample (100- to 200-mesh size) were sprinkled liberally over the surface. If the grains were applied before the glycol phthalate had cooled sufficiently, they sank in it; if applied after complete hardening, they were not held by it, and rolled during orientation. Ideally, the lower surfaces of the grains should adhere to the glycol phthalate without projecting into it.

Immersion oil of index intermediate between N_x and N_z of the sample (previously roughly determined or estimated) was applied to the layer of grains, which was then covered with a cover glass. The same oil was used in mounting the slide between the hemispheres of the universal stage, so that no contamination of the immersion oil could result. The possibility of contamination of the immersion oil by reaction with the glycol phthalate was carefully checked by measuring the index of several oils before and after they had been in contact with glycol phthalate and exposed to light for 1–2 hours. No measurable change of index was found. The oils so tested are Shillaber's Certified Index of Refraction Liquids, in the refractive-index range 1.530 to 1.580. It should be noted that oils of different compositions, especially high-index types, probably react with glycol phthalate.

In order to make the method precise, it was necessary to have rigid temperature control during the determinations. This was accomplished by using a Bausch and Lomb universal stage fitted with a lower hemisphere mount through which water was circulated. The water temperature as measured by a thermometer in the circuit was kept equal to the room temperature by an adjustable heating–cooling unit in the circulator, to insure against loss or gain of heat in the lines between the thermometer and the hemisphere, and so that there would be no unknown temperature gradient

between the water in the lower hemisphere and the grain mount. Unless the water is at room temperature, there must necessarily be some unknown temperature difference between the thermometer, the water in the lower hemisphere, and the grain mount. If the water were 10°C. or more different from room temperature, an error of 2°C. in the assumed temperature of the grain mount might easily result and would cause an error of .0008 in the measured refractive index. The present method has the distinct advantage of eliminating temperature gradients by the simple expedient of keeping the temperature of the whole setup very close to room temperature.

To make the actual measurement, a suitable grain was oriented in white light, from which the heat wave lengths were removed by an infra-red filter. The optic plane was made vertical, with either X or Z horizontal and north-south (parallel to the vibration direction of the lower nicol). With a change to sodium light, the grain was rotated on the control axis until its index matched that of the oil. This position is best seen if a medium- or low-power objective lens is used, and if the grain is slightly out of focus. The position was determined and recorded three times on one side of the zero position of the control axis, and then, if possible, on the other side, in order to cancel out the error due to misorientation. The angles of tilt were corrected for difference in index between oil and hemispheres, taking into account the added or subtracted north or south tilt of the inner stage. The average of the six corrected readings was taken as the true reading. This value, the known index of the oil, and the average total birefringence of the sample (previously determined) was used to calculate N_x and N_z of the grain from the convenient graphic solution of ellipses presented by Emmons (1948, Pl. 10).[16]

Knowing $2V$ for the plagioclase of the sample, N_y was calculated from Wright's approximate formula

$$(N_y - N_x) = (N_z - N_x) \sin^2 V_z$$

the approximation being well within the limits of precision of the measurement. The indices of 7 to 15 grains were determined in each sample, the number of grains depending on the range of index in the sample. It should be pointed out that none of the three index values is independent. That is, the determined values of N_x and N_z depend partly on the measured total birefringence of the sample, and the determined value of N_y depends partly on the measured birefringence and $2V$. However, the effect on the index values of errors in the measured values of birefringence and $2V$ is very slight. For instance, an error of .0005 in birefringence results in a maximum error of .0003 in N_x and N_z; an error of 2° in $2V$ results in a maximum error of .0002 in N_y. Inasmuch as the probable errors in the mean values of total birefringence and $2V$ are estimated to be $\pm$.0001 and $\pm$.3°, respectively, their effect on the mean values of the principal indices can be considered to be negligible.

[16] Application of the average total birefringence of the sample, rather than the unknown total birefringence of the particular grain, results in a maximum error of $\pm$.0002 in the measured index of the grain if the range of total birefringence in the sample is .001 and if the angle of rotation from the principal orientation to the position at which the index of the grain matches that of the oil is 50°. In most samples, the range of total birefringence is less than .001, and the angle of rotation was less than 50°, so that the resultant error in an individual determination is almost negligible. It is, of course, averaged out of the mean of measurements on several grains.

The standards used in the index determinations were carefully checked. The oils were calibrated on an Abbe refractometer through which water at room temperature was circulated, and the water temperature was measured with the same thermometer as that used in the plagioclase index determinations. The thermometer was calibrated against a thermometer calibrated by the United States Bureau of Standards. The Abbe refractometer was first checked with a block of optical glass of known index. As there was some doubt as to the accuracy of reported refractive indices of optical glasses, the index of an oil was measured by substituting quartz for plagioclase in the method of refractive-index measurement described above, and running the determination under exactly the same conditions. The refractive indices and birefringence of quartz are known to the fifth decimal place, and therefore the only remaining unknown is the refractive index of the oil, which can be computed. Four readings of the index of the oil had a range of .0001; the average was taken as correct, and the Abbe refractometer was calibrated against the oil. The calibration by this method differed from the calibration by the optical glass block by $-.0003$. There is no way of knowing which is more nearly correct, because the calibration of the refractometer is itself subject to an error of $\pm.0002$. The calibration against quartz was arbitrarily chosen as correct. If the standards, thus checked, are considered to be correct, the estimated error in an individual index determination is $\pm.0004$. This is the sum of the known reproducibility of a measurement on slightly altered and twinned plagioclase ($\pm.0001$), the estimated error due to error in the temperature measurement ($\pm.0001$), and the error in the measurement of the index of the oil ($\pm.0002$). The error in the oil index and possibly the error in the temperature measurement remain in the mean and account for part of the scatter of the points on the index curves in Plate 12.

The method of refractive-index determination described above is precise and rapid, and it is worthy of widespread application. It equals the precision claimed for the double-variation method developed by Emmons (1948) and avoids the errors due to unknown temperature gradients and to the inconvenient variable wave-length light source. It provides a rapid means of measuring in one or two mounts as many grains as are needed; the indices of 10 grains can be measured in 1 hour, which is almost as rapid as the immersion methods used to determine maximum or minimum plagioclase indices to $\pm.001$. If the birefringence of the section of the mineral perpendicular to the axis of rotation is not known, the measurements must be made in two oils of different indices, which doubles the time required for the determination, but which on the other hand gives two principal indices of the sample, independent of other measurements. The indices of plagioclase of unknown composition from intermediate to mafic plutonic igneous rocks can be calculated from the measurements made in one oil by making a first approximation of the index, using an assumed birefringence, then using the birefringence associated with the first approximation of the index in Plate 12 in the second approximation, and so on. A third approximation would generally suffice. This procedure should be used with discretion until the relations of optical properties to composition in other types of plagioclase are known.

3. BIREFRINGENCE DETERMINATION

Most of the determinations of the total birefringence of plagioclase hitherto reported in the literature were made by taking the difference between the measured

values of N_x and N_z. Inasmuch as the best determinations of N_x and N_z by immersion methods are subject to errors of $\pm.0004$, the maximum error in the difference between them is $\pm.0008$, which is about ± 10 per cent of the total birefringence of plagioclase. The precision of the method used in the present study is about ± 2 per cent of the total birefringence, which is comparable to the precision of the refractive index and $2V$ measurements. In the samples studied, birefringence is therefore an independently measured diagnostic property and assumes this status for the first time in any extensive optical study of plagioclase.

The determinative method used has been briefly described by Hess (1949, p. 628–629). In effect, it compared the birefringence of plagioclase with that of quartz, which was taken to be .00918 for white light (Winchell, 1951, p. 247). Briefly, it consists in mounting a mixture of quartz grains and plagioclase grains in hard bakelite discs, from which thin sections, ground flat and semipolished on each side, are cut (Fig. 31). The thickness of any part of the slide can then be determined accurately on the universal stage by measuring the retardation in oriented quartz grains with a Berek compensator, and substituting this value in the relation

$$\text{inclined thickness} = \frac{\text{retardation}}{\text{birefringence } (= .00918).} \qquad (1)$$

The true thickness is equal to the inclined thickness multiplied by the cosine of the corrected angle of tile of the inner stage. The birefringence of a plagioclase grain is given by the relation

$$\text{birefringence} = \frac{\text{retardation} \times \text{cosine of the corrected angle of tilt}}{\text{true thickness.}} \qquad (2)$$

The retardation is measured in the X–Z plane, and the true thickness is interpolated from the true thickness of surrounding quartz grains. The method has the following advantages:

1) Hundreds of grains are available in one mount, among which many are close to the desired orientation.

2) Quartz and plagioclase are intimately mixed, so that several independent determinations of the thickness can be made in the immediate vicinity of one plagioclase grain.

3) Bakelite is far superior to softer mounting media because the mineral grains, very strongly bound by the bakelite, do not pluck out during grinding; on flat steel laps, the bakelite is not preferentially ground and does not flow plastically, so that a flat surface is attainable. If softer mounting media are used, preferential grinding and/or plastic flow of the medium result in irregularities in thickness even between grains side by side in the mount, and the irregularities are of such a magnitude as to destroy the precision of the method.

4) Part of the sample used for the chemical analysis can be mounted, thereby eliminating the possibility that the grains studied optically have a bulk composition different from the chemical analysis.

5) The slides are useful for the measurement of $2V$ of grains from analyzed samples.

The bakelite discs containing the grains were made on a mounting press of the type

ordinarily used for mounting polished sections of ore minerals, according to the following procedure: A mixture of one part plagioclase grains, one part quartz grains, and one part screened bakelite powder was spread evenly over the surface of the lower inner cylinder of the mold. Best results were obtained when the quartz, plagioclase, and bakelite were of approximately equal grain size, between 100 and 200 mesh, and when the layer was made as nearly as possible one grain thick on the surface of the cylinder. The sleeve of the mold was then fitted over the lower cylinder, and care was taken not to disturb the layer of grains. After the bakelite powder was poured into the sleeve over the sample (sufficient to make a disc ¼ to ½ inch thick), and the upper inner cylinder was inserted, the mold was placed carefully on the press; pressure was maintained at 1000 psi until the temperature had risen to 85°C.; it was then raised to 3500 psi. When the temperature reached 140°C., the mold was removed from the press, and the bakelite disc from the mold. At this stage, the sample was a thin layer of grains at or nearly at one surface of the disc. This surface was then ground on a flat steel lap, using American Optical No. 303½ abrasive, until the layer of grains had been well intersected by the surface. The surface was then semipolished on another steel lap, using American Optical No. 305 abrasive.[17] Next, an ordinary thin section mounted on a glass slide was cut from the side of the disc containing the grains. The cut surface of the thin section was first ground until the desired thickness was reached (as estimated by the quartz interference colors), and then semipolished as before. The slide was then ready for use.

All measurements of the birefringence of the samples so mounted were made on the universal stage in white light by the following procedure: A grain of plagioclase was chosen with the optic plane not more than 40° from the horizontal and closely surrounded by three or more quartz grains with their c axes less than 40° from horizontal. A sketch of the part of the field containing the grains was made to any convenient scale, using a grid ocular and grid paper, and each grain was identified by a number (Fig. 41). To determine the thickness of each quartz grain, the c axis was oriented horizontal and exactly in the N. 45° E. position, in which position the angle of compensation was measured with a Berek compensator, reading the angle two or three times on either side of the zero position of the compensator; the inclined thickness corresponding to the observed compensation angle was then read directly from a curve constructed from the compensator tables and equation (1); the inclined thickness was multiplied by the cosine of the corrected angle of tilt to give the true thickness. The true thickness of each quartz grain was recorded on the sketch of the field, and the true thickness of the plagioclase grain was interpolated. The retardation in the optic plane of the plagioclase grain was similarly determined, in this case orienting the optic plane horizontal and Z exactly in the N. 45° E. position. Finally, the true thickness of the plagioclase grain, the retardation, and the cosine of the corrected angle of tilt were substituted in equation (2) to give the total birefringence. Five to 15 independent determinations of birefringence were made on each sample, the number depending on the variation found. The method requires considerable time: with experience, a mount can be made in half an hour (assuming that the thin section can be

[17] Grinding can be done equally well on flat plate glass.

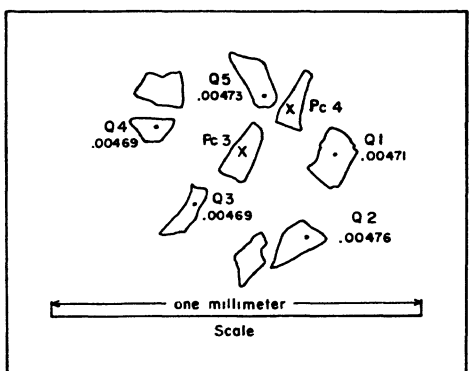

FIGURE 41.

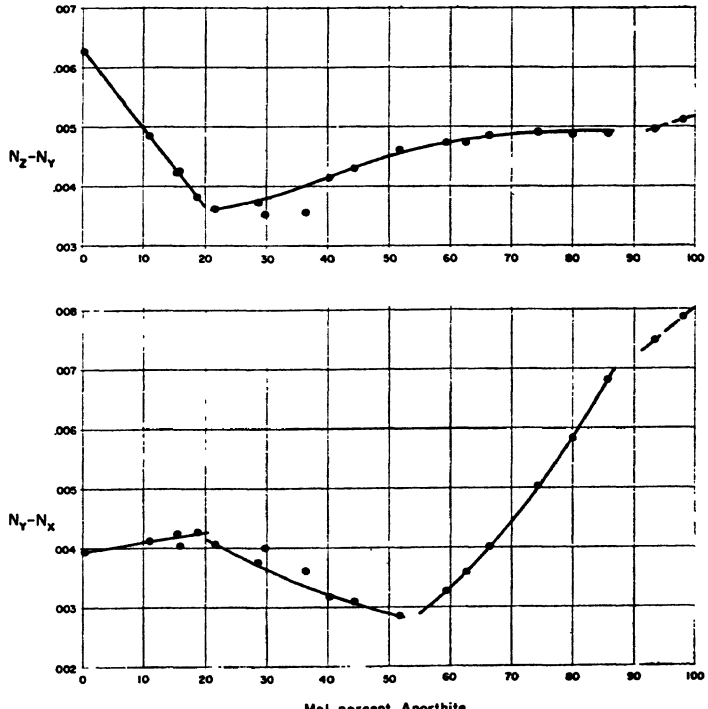

FIGURE 42.

FIGURE 41.—*Illustration of birefringence measurement*

Q = quartz with the thickness in mm as measured; Pc = plagioclase on which birefringence was determined.

FIGURE 42.—*Variations in partial birefringence with composition*

cut immediately), and the birefringence of one plagioclase grain can be measured and calculated in three-quarters of an hour.

Error from several sources enters the determination. First, the sides of the thin section are not perfectly flat planes. Undulations in thickness over distances greater than 1 mm, if they exist, do not affect the result, because the quartz grains of which the thickness is measured are always less than 1 mm from the plagioclase grain. Smaller-scale irregularities not taken into account by straight-line interpolation between the points of known thickness are minimized by the fine grinding; their effect is partly averaged out by measuring the thickness of three or more quartz grains, and it is estimated to cause less than ± 1 per cent error in birefringence, except possibly in one or two erratic cases among scores of determinations. The thickness of any one quartz grain can be reproduced to $\pm .0005$ mm, representing an error of ± 1.4 per cent in birefringence (in a slide .035 mm thick). The thickness of a plagioclase grain, determined from three or more quartz grains, is thus subject to an error of less than ± 2 per cent from these two sources. Misorientation of the plagioclase grain introduces a small error which is always negative in total birefringence measurements. However, the error is much less than the sensitivity of the measurement, as it is unmeasurable if the optic normal is rotated as much as 10° away from the microscope axis. The same Berek compensator, Leitz No. 903, was used for all the determinations. The effect of the unknown error in the compensator is probably negligible, because quartz and plagioclase have closely similar total birefringence, so that the error in the measured plagioclase retardation is cancelled by the error in the measured quartz retardation. This would not be true if the birefringence to be determined were greatly different from the birefringence of quartz, in which case it would be necessary to test the error in the compensator. The mean of two or three readings of the compensation angle can always be reproduced to the limit of sensitivity of the scale reading if a high-power, long-focus objective lens[18] is used in making the measurement, and the compensation band is brought to the same point within the grain for each reading. The sensitivity of the scale reading is .1°, so that the maximum error in the compensation angle is $\pm .05°$, representing an error of ± 1 per cent in birefringence in a slide .03 mm thick or an error of $\pm .6$ per cent in a slide .06 mm thick. For this reason, the slide should be made as thick as the grain size of the sample will allow, which was found to be about .045 mm if the grain size of the sample is 150 mesh. The reproducibility of the measurement of the birefringence of one plagioclase grain is at worst $\pm .0002$, which is somewhat less than the sum of the maximum estimated errors listed above. The mean values are probably correct to $\pm .0001$.

4. OPTIC AXIAL-ANGLE DETERMINATION

Measurements of $2V$ were made mostly on grains from the analyzed samples in the bakelite slides prepared for the birefringence determinations; these grains have a strong preference to lie on their cleavages, and in some composition ranges this leaves very few grains with the acute bisectrix close to vertical. In such cases, as many grains as possible were measured in the bakelite slides, and others were measured in thin sections cut from the specimens from which the plagioclase samples were taken.

[18] A Leitz UM-4 objective was used.

In making the measurements, the standard orthoscopic procedure was followed. The optic plane was oriented vertical, and the extinction position of each optic axis was observed three to five times and averaged, first in one 45-degree position of the microscope stage and then in the other; the average of the two corrected values of $2V$ thus obtained was accepted for the grain. In no case was V measured and doubled to obtain $2V$, a procedure which introduces as much as 2° error, no matter how carefully the grain is oriented. Six to 20 measurements of $2V$ were made on different grains in each sample, depending on the variation found. The reproducibility of the measurement of $2V$ on one grain was found to be 0.5° or better. The range of values from grain to grain in one sample varies considerably from sample to sample. In some samples, the range of $2V$ is that to be expected from a consideration of the range of refractive index for the same samples; in other words, the range of $2V$ in these samples can probably be accounted for by the range of composition within the samples. In other samples, the range of $2V$ is too great to be accounted for in this way; this is especially true of the samples from granites in the composition range An_{10} to An_{20}, and of the samples from the Skaergaard intrusion in the composition range An_{35} to An_{45}. This "excessive scatter" of $2V$ in these samples is thought to be real and to result from varying states of inversion between high- and low-temperature modifications. (*See further* under Interpretation of the Data.)

E. OPTICAL DATA

1. GENERAL STATEMENT

Table 54 lists the results of the individual optical measurements and is more or less self-explanatory. The first column gives the collector's number of the sample, composition in mol per cent anorthite, and the source of the sample. The second column lists the individual determinations of N_x and gives their arithmetic mean and range of variation for each sample. The third column gives the mean values of N_y and N_z for each sample; the means involve the same range as the N_x values. The fourth column gives the individual measurements, arithmetic mean, and range of the total birefringence for each sample. The fifth column gives the mean value of $N_y - N_x$ calculated from the mean total birefringence and mean $2V$ of the sample. The sixth column gives the individual measurements, arithmetic mean, and range of the $2V_z$ values for each sample; in many cases, $2V_x$ was actually measured, but to avoid confusion all the $2V$ measurements are reported as $2V_z$.

2. REJECTION OF DATA

Several chemically analyzed samples were rejected because the range of variation of the optical properties in the samples was so great that an undue expenditure of time would have been required to establish accurate mean values. About 10 individual optical measurements were discarded from the hundreds in Table 54 because their obvious aberrancy suggested erratic error due to unknown causes; if they had been included, the range of variation in the samples involved would have been greater, but in no case would the mean values have been changed beyond their estimated limits of accuracy.

TABLE 54.—*Lists of optical data for plagioclase*

Sample	N_x	N_y / N_z Means	$N_z - N_x$	$N_y - N_x$	$2V_z$
D-761-1	1.5285		.01003		76.4°
An 0.2	1.5285	N_y	.01008		76.4
In pegmatite, Ramona,	1.5286	1.5326	.01018		77.0
Calif.	1.5286		.01022		77.2
	1.5286	N_z	.01023		77.3
	1.5287	1.5388	.01026		77.7
	1.5290		.01033		77.8
	———		.01037		78.0
	Mean		Mean		Mean
	1.5286		.01021	Calculated	77.2°
				.00394	
	Range		Range		Range
	.0005		.00034		1.6°
D-761-2	1.5281				
An 0.4	1.5284				
In pegmatite, Pala, Calif.	1.5286				
	1.5286				
	1.5286				
	1.5287				
	1.5287				
	———				
	Mean				
	1.5285				
	Range				
	.0006				
D-761-3	1.5286				
An 0.4	1.5286				
In pegmatite, Amelia, Va.	1.5286				
	1.5287				
	1.5288				
	1.5289				
	1.5294				
	———				
	Mean				
	1.5288				
	Range				
	.0008				
10	1.5339		.00884		82.3°
An 11.0	1.5341		.00886		82.7
In granite, Peekskill, N. Y.	1.5343	N_y	.00891		84.6
	1.5344	1.5386	.00896		85.1
	1.5344		.00899		85.3

TABLE 54.—*Continued*

Sample	N_x	N_y N_z Means	$N_z - N_x$	$N_y - N_x$	$2V_z$
	1.5344	N_z	.00901		85.6
	1.5345	1.5435	.00918		87.8
	1.5348		———		88.1
	1.5350				———
	1.5351				
	———				
	Mean		Mean		Mean
	1.5345		.00896	Calculated	85.2°
				.00411	
	Range		Range		Range
	.0012		.00034		5.8°
73	1.5364		.00833		88.6°
An 15.4	1.5365		.00843		89.3
In granite, Parishville,	1.5368	N_y	.00843		89.7
N. Y.	1.5369	1.5412	.00851		90.0
	1.5370		.00851		90.8
	1.5370	N_z	.00851		91.4
	1.5370	1.5454	———		———
	1.5370				
	1.5373				
	1.5375				
	1.5375				
	———				
	Mean		Mean		Mean
	1.5370		.00845	Calculated	90.0°
				.00423	
	Range		Range		Range
	.0011		.00018		2.8°
156	1.5361		.00824		86.3°
An 15.9	1.5363	N_y	.00829		86.4
In granite, Llano Co., Texas	1.5364	1.5407	.00831		86.6
	1.5366		.00832		87.8
	1.5368	N_z	———		88.0
	1.5368	1.5450			88.1
	1.5369				88.1
	1.5370				88.6
	1.5377				89.0
	———				89.0
					89.3
					89.3
					89.6
					90.8
					———
	Mean		Mean		Mean
	1.5367		.00829	Calculated	88.4°
				.00403	
	Range		Range		Range
	.0016		.00008		4.5°

TABLE 54.—*Continued*

Sample	N_z	$\frac{N_y}{N_z}$ Means	$N_z - N_x$	$N_y - N_x$	$2V_z$
94	1.5381		.00794		90.3°
An 18.7	1.5384		.00797		91.5
In granite, Tigerton, Wisconsin	1.5385	N_y	.00807		92.0
	1.5386	1.5429	.00808		92.7
	1.5387		.00816		92.8
	1.5387	N_z	.00824		93.3
	1.5387	1.5467	———		93.5
	1.5391				93.6
	———				94.8
					95.6
					95.6
					———
	Mean		Mean		Mean
	1.5386		.00808	Calculated .00427	93.2°
	Range		Range		Range
	.0010		.00030		5.3°
80165	1.5388		.00763		92.8°
An 21.6	1.5388	N_y	.00764		93.0
In pegmatite from Hawk Mica Mine, Bakersville, N. C.	1.5390	1.5431	.00764		93.4
	1.5391		.00768		93.6
	1.5392	N_z	.00782		93.9
	1.5392	1.5467	———		94.2
	———				———
	Mean		Mean		Mean
	1.5390		.00768	Calculated .00407	93.5°
	Range		Range		Range
	.0004		.00019		1.4°
97490	1.5424		.00734		89.6°
An 28.7	1.5424	N_y	.00741		89.8
In pegmatite, Little Rock Creek, Mitchel Co., N. C.	1.5425	1.5463	.00745		90.2
	1.5425		.00750		90.5
	1.5426	N_z	.00750		90.7
	1.5426	1.5500	.00753		90.9
	———		.00758		———
			———		
	Mean		Mean		Mean
	1.5425		.00747	Calculated .00375	90.3°
	Range		Range		Range
	.0002		.00024		1.3°
D-638	1.5433		.00733		91.4°
An 29.8	1.5433		.00735		91.9

TABLE 54.—*Continued*

Sample	N_z	N_y / N_z Means	$N_z - N_x$	$N_y - N_x$	$2V_z$
In pegmatite from Macon Co. N. C.	1.5433	N_y	.00739		92.0
	1.5435	1.5479	.00752		94.0
	1.5437		.00756		94.4
	1.5439	N_z	.00777		94.5
	1.5442	1.5514	.00778		94.6
	1.5444		———		94.7
	1.5444				94.7
	1.5450				94.8
	———				95.1
	Mean		Mean		Mean
	1.5439		.00753	Calculated .00400	93.6°
	Range		Range		Range
	.0017		.00045		3.7°
4145	1.5473		.00681		89.0°
An 36.5	1.5475	N_y	.00698		89.4
In gabbro from Skaergaard intrusion, Greenland	1.5476	1.5512	.00705		89.4
	1.5476		.00710		89.6
	1.5476	N_z	.00718		89.7
	1.5476	1.5548	.00729		90.0
	1.5481		.00742		90.6
	———		.00757		90.7
			———		91.7
					93.7
	Mean		Mean		Mean
	1.5476		.00717	Calculated .00361	90.4°
	Range		Range		Range
	.0008		.00076		4.7°
2580	1.5493		.00715		79.7°
An 40.2	1.5500		.00716		80.0
In gabbro from Skaergaard intrusion, Greenland	1.5501	N_y	.00728		80.8
	1.5501	1.5537	.00737		81.4
	1.5501		.00738		82.3
	1.5502	N_z	.00747		82.9
	1.5504	1.5578	.00750		83.0
	1.5508		———		83.8
	1.5513				84.1
	1.5514				86.8
	1.5516				———
	———				
	Mean		Mean		Mean
	1.5505		.00733	Calculated .00319	82.5°
	Range		Range		Range
	.0023		.00035		7.1°

TABLE 54.—*Continued*

Sample	N_x	N_y N_z Means	$N_z - N_x$	$N_y - N_x$	$2V_z$
1963	1.5506		.00721		78.1°
An 44.2	1.5506		.00722		79.4
In gabbro from Skaergaard	1.5506	N_y	.00736		79.4
intrusion, Greenland	1.5506	1.5547	.00739		79.5
	1.5517		.00750		79.5
	1.5519	N_z	.00754		79.7
	1.5520	1.5590	.00757		79.8
	1.5520		———		79.9
	1.5520				80.2
	1.5521				80.2
	1.5525				82.9
	1.5525				83.6
	———				86.1
					———
	Mean		Mean		Mean
	1.5516		.00740	Calculated	80.6°
				.00310	
	Range		Range		Range
	.0019		.00036		8.0°
C-41			.00733		
An 50 ± 2*			.00736		
In metagabbro, Sittam-			.00739		
pundi Complex, India			.00745		
			———		
			Mean		
			.00738		
			Range		
			.00012		
GD29	1.5540		.00738		75.0°
An 51.8	1.5541		.00740		75.0
In dolerite, Great Dyke, S.	1.5542	N_y	.00748		75.7
Rhodesia	1.5545	1.5575	.00751		75.7
	1.5548		.00752		76.0
	1.5548	N_z	———		76.0
	1.5549	1.5621			76.3
	1.5550				76.5
	1.5558				76.6
	———				76.6
					77.0
					77.0
					77.7
					78.1
					78.2
					———
	Mean		Mean		Mean
	1.5547		.00746	Calculated	76.4°
				.00285	
	Range		Range		Range
	.0018		.00014		3.2°

TABLE 54.—*Continued*

Sample	N_x	N_y / N_z Means	$N_z - N_x$	$N_y - N_x$	$2V_z$
GD31	1.5550		.00755		77.3°
Unanalyzed; normative	1.5551				77.9
composition from rock	1.5552				78.0
analysis	1.5553				78.0
An 55.5	1.5553				78.5
In dolerite, Great Dyke,	1.5554				78.6
S. Rhodesia	———				79.1
					81.6
					———
	Mean				Mean
	1.5552				78.6°
	Range				Range
	.0004				4.3°
BV63	1.5579		.00776		78.8°
An 59.2	1.5580	N_y	.00794		79.0
In gabbro, Bushveld	1.5581	1.5615	.00796		79.0
igneous complex, S.	1.5581		.00813		79.3
Africa	1.5581	N_z	.00814		79.4
	1.5583	1.5662	.00814		81.6
	1.5586		———		———
	———				
	Mean		Mean		Mean
	1.5582		.00801	Calculated	79.5°
				.00328	
	Range		Range		Range
	.0007		.00038		2.8°
EB41	1.5589		.00791		78.5°†
An 62.5	1.5598		.00812		79.0
In gabbro, Stillwater	1.5599	N_y	.00821		79.7†
igneous complex,	1.5600	1.5640	.00823		80.6
Montana	1.5603		.00823		81.5†
	1.5604	N_z	.00825		81.8†
	1.5605	1.5687	.00826		82.0
	1.5607		.00830		82.2
	1.5607		.00831		82.2
	1.5615		.00832		82.2†
	1.5615		.00837		82.6
	———		.00850		82.6
			.00852		82.6†
			.00853		82.7
			.00871		82.8†
			———		82.9
					83.1
					83.3
					83.4
					84.5
					84.5
					———
	Mean		Mean		Mean
	1.5604		.00832	Calculated	82.1°
				.00359	
	Range		Range		Range
	.0026		.00080		6.0°

TABLE 54.—*Continued*

Sample	N_z	N_y / N_z Means	$N_z - N_x$	$N_y - N_x$	$2V_z$
EB42	1.5606		.00852		82.4°
An 66.2	1.5610		.00854		82.8
In gabbro, Stillwater	1.5610	N_y	.00855		83.0
igneous complex,	1.5611	1.5656	.00857		83.2
Montana	1.5613		.00875		83.3
	1.5613	N_z	.00880		83.4
	1.5614	1.5705	.00883		84.2
	1.5616		.00887		84.4
	1.5622		.00888		84.4
	1.5630		.00888		84.5
	1.5633		.00888		85.5
	———		.00907		86.4
			.00922		86.8
			.00923		89.1
			.00943		——
			———		
	Mean		Mean		Mean
	1.5616		.00887	Calculated	84.5°
				.00401	
	Range		Range		Range
	.0027		.00091		6.7°
EB43	1.5641		.00925		88.9°
An 74.3	1.5643		.00952		89.3†
In gabbro, Stillwater	1.5644	N_y	.00964		89.4†
igneous complex,	1.5645	1.5701	.00972		89.6†
Montana	1.5645		.00977		89.6†
	1.5648	N_z	.00981		89.8
	1.5649	1.5750	.00987		89.8
	1.5650		.00989		90.2
	1.5651		.00989		90.5†
	1.5659		.01012		93.0†
	1.5663		.01043		93.7
	1.5668		.01047		93.7
	———		.01068		——
			———		
	Mean		Mean		Mean
	1.5650		.00993	Calculated	90.6°
				.00502	
	Range		Range		Range
	.0027		.00143		4.8°
EB38	1.5664†		.01046		93.7°
An 80.0	1.5668		.01058		93.7
In gabbro, Stillwater	1.5668†		.01061		94.2†
igneous complex,	1.5669†	N_y	.01065		94.4
Montana	1.5669	1.5729	.01070		94.5
	1.5669		.01076		95.1†
	1.5670†	N_z	.01081		96.0
	1.5670†	1.5778	.01088		96.1†
	1.5671†		.01090		96.2†
	1.5671		———		96.6
	1.5671				——
	1.5672†				

TABLE 54.—*Continued*

Sample	N_x	N_y N_z Means	$N_z - N_x$	$N_y - N_x$	$2V_z$
	1.5673†				
	1.5673†				
	1.5673				
	1.5679				
	Mean		Mean		Mean
	1.5671		.01071	Calculated	95.0°
				.00583	
	Range		Range		Range
	.0015		.00044		2.9°
EB18	1.5693		.01128†		97.4°
An 85.8	1.5693		.01139		97.9
In gabbro, Stillwater	1.5694		.01146†		98.5
igneous complex,	1.5695		.01146		99.2†
Montana	1.5695	N_y	.01153†		99.3†
	1.5696	1.5765	.01161		99.8†
	1.5696		.01180		100.2
	1.5696	N_z	.01181		100.3†
	1.5696	1.5814	.01182†		100.8
	1.5696		.01184†		101.0
	1.5700		.01187†		———
	1.5700		.01193		
	1.5702		.01197†		
	1.5705		.01208		
	Mean		Mean		Mean
	1.5697		.01170	Calculated	99.4°
				.00681	
	Range		Range		Range
	.0012		.00080		3.6°
354	1.5713		.01156		100.0°
An 93.3	1.5716		.01213		101.2
In norite, Pala Mountain,	1.5719		.01252		101.4
Calif.	1.5719	N_y	.01317		101.9
	1.5721	1.5798	———		102.9
	1.5721				103.1
	1.5722	N_z			———
	1.5722	1.5848			
	1.5724				
	1.5724				
	1.5725				
	1.5733				
	1.5740				
	Mean		Mean		Mean
	1.5723		.01245	Calculated	101.6°
				.00748	
	Range		Range		Range
	.0027		.00161		3.1°

TABLE 54.—*Concluded*

Sample	N_x	$\frac{N_y}{N_z}$ Means	$N_z - N_x$	$N_y - N_x$	$2V_z$
C50	1.5736		.01270†		101.3°
An 98.0	1.5738		.01283		101.5
In metaanorthosite,	1.5746	N_y	.01290		101.9
Sittampundi complex,	1.5746	1.5825	.01292		102.6
India	1.5747		.01303†		102.8
	1.5747	N_z	.01309†		102.9
	1.5747	1.5876	.01346†		102.9
	1.5748		———		———
	1.5750				
	1.5753				
	———				
	Mean		Mean		Mean
	1.5746		.01299	Calculated	102.3°
				.00788	
	Range		Range		Range
	.0017		.00076		1.6°

* Bulk composition of sample is $An_{83.7}$ by chemical analysis. By modal analysis, sample consists of 12% An_{80} (estimated by refractive index) and 88% An_{80} (by difference). Birefringence was measured on An_{80} grains.

† Measured by H. H. Hess

3. COMPARISON OF MEAN OPTICAL PROPERTIES WITH MEASUREMENTS MADE IN OTHER LABORATORIES

The refractive indices, and in some cases the $2V$ of some of the samples studied, have been measured in other laboratories. A comparison of these measurements is given in Table 55, with the aim of assessing interlaboratory bias. The discrepancies between the refractive-index measurements made at Princeton by the writer and those made at Wisconsin by C. D. Reynolds (samples D-761-1, 2, 3, 4 and D-638) are seriously large. Reynolds measured the refractive indices of only two grains out of each sample, but this does not account for the discrepancies, because the lowest N_x values measured by the writer are higher by .0018 to .0027 than the highest measured by Reynolds in four samples. The writer's measurements of N_x in samples D-761-3 (Amelia albite) and D-761-4 (Bakersville oligoclase) agree well with the values from the literature, most of which were made by the prism method (for reference to reports of the measurements, *see* Chayes, 1952). It would therefore appear that the burden of proof of accuracy rests with Reynolds. The N_x values of samples 10, 73, 156, and 94 measured by the writer agree fairly well with those measured by Crump and Ketner at Wisconsin; the discrepancies here may well be due to the fact that Crump and Ketner measured the indices of only three or four grains from each sample (Emmons *et al.*, 1953, p. 24), because the range of values found by the writer includes Crump and Ketner's values in most cases. The discrepancies between the $2V$ values compared in Table 55 are greater than would be expected from the estimated error in the mean values of $2V$ in the present study, which is $\pm .3°$ for most of the samples. This is probably partly a result of the fact that in some cases Crump and Ketner measured V

and doubled it to obtain $2V$ (Emmons *et al.*, p. 30), which they themselves admit introduces considerable error (p. 30, 32); their measurements of $2V$ were apparently made on cleavage fragments in oil, which, as previously discussed, introduces error unless (1) the grains are perfectly flat plates or (2) the index of the oil is equal to N_y

TABLE 55.—*Comparison of measurements on plagioclase at several laboratories*

Sample	N_x			$2V_z$	
	Smith	Wisconsin*	Mean of "old" measurements	Smith	Wisconsin*
D-761-1	1.5286	1.5258			
D-761-2	1.5285	1.5261			
D-761-3	1.5288	1.5263	1.5291**		
10	1.5345	1.5346		85.2°	81°
73	1.5370	1.5374		90.0°	87°
156	1.5367	1.5373		88.4°	89°
94	1.5386	1.5376		93.2°	94.5°
80165	1.5390	1.5370	1.5387**		
D-638	1.5439	1.5421			

* Mean values of measurements quoted by Emmons *et al.* (1953)

** In each case, mean of five determinations selected by Chayes (1952) from the literature

of the grain. There is therefore little basis for comparing the accuracies of the $2V$ measurements.

F. INTERPRETATION OF THE DATA

1. FITTING THE CURVES

The mean values of the optical properties listed in Table 54 are plotted against composition in Plate 12. The curves were drawn primarily on the basis of the best fit by visual inspection, but partly on the basis of less direct considerations which require some discussion. First, the mathematical relationships between refractive indices, total birefringence, and $2V$ impose mutual requirements on the shapes of the curves in Plate 12. The relationship between $2V$ and refractive indices is more easily visualized in terms of the partial birefringences, which have been calculated from the total birefringence and $2V$ for each sample, and plotted against composition in Figure 42. The relative slopes and the changes in slope of the refractive-index curves in Plate 7 are closely consistent with the requirements imposed on them by the variations of the partial birefringences and the total birefringence. For instance, where the total birefringence is nearly constant (An_{20} to about An_{55}), the curves for N_x and N_z are parallel; where the partial birefringence ($N_z - N_y$) is nearly constant (An_{25} to An_{86}), the curve for N_z is parallel to the curve for N_y; and so on. The breaks in slope for all the curves at An_{20}, An_{52}, and *ca.* An_{90} are also mutually consistent. A slight inconsistency lies in the fact that, where one or more of the birefringence curves is not a straight line, at least one of the refractive-index curves cannot be a straight line, whereas all the refractive-index curves are drawn as straight lines over the whole composition range. However, the departure from linearity re-

quired by the curvatures in the birefringence curves is so slight that, even if it were assigned wholly to one refractive-index curve, that curve would depart less than .0002 refractive-index units from a straight line; if the departure is shared among the three refractive-index curves, it is too slight to plot.

The curves in Plate 7 are therefore consistent to a high degree with the optical data and the chemical compositions (with the exception of the two obviously anomalous $2V$ values, which are discussed in the next section). This does not mean that the curves represent the only possible relations between optical properties and chemical composition for the samples studied, nor that the details of the relations may not be changed when further precise data become available for other plagioclase samples.

2. IMPLICATIONS AND LIMITATIONS OF THE DATA

The results of recent X-ray studies of plagioclase have demonstrated differences in the crystal structures of plagioclase of different compositions from rocks which either crystallized at relatively low temperatures or cooled very slowly from higher temperatures. Such plagioclase is called "low-temperature" plagioclase. Briefly, the differences which have been found in the structures of low-temperature plagioclase are as follows: The unit cell of low-temperature pure albite is face-centered on (001), and has a c dimension of about 7 Å (Cole, Sörum, and Taylor, 1951). Low-temperature albite and oligoclase in the composition range An_6 to An_{17} consist of two submicroscopically discrete phases (Laves, 1954). Low-temperature plagioclase in the composition range $An_{20\pm}$ to $An_{70\pm}$ has a superlattice structure (Gay, 1956). The unit cell of low-temperature bytownite is body-centered, with a c dimension of about 14 Å (Gay, 1953). The unit cell of low-temperature anorthite is primitive, with a c dimension of about 14 Å (Gay, 1953). Inasmuch as optical properties are functions of crystal structure as well as of composition, the discontinuities in the optical-property curves may reflect discontinuous changes in the crystal structure. For example, the discontinuities in the optical-property curves at An_{20} may reflect a change from a "peristeritic" two-phase intergrowth to the superlattice structure of the intermediate plagioclases; this change has been placed near An_{20} by Gay and Smith (1955). However, the four samples of the present study in the composition range An_{10} to An_{20} have been studied by Laves (1954); of the four, the three most sodic were found by him to be two-phase intergrowths, but the most calcic sample, number 94 ($An_{18.7}$), was found to be homogeneous, so that this explanation of the optical discontinuity is apparently not valid. Similar speculations might be applied to the optical discontinuities near An_{52} and An_{90}, but, in the present state of knowledge, no rigorous explanations of the discontinuities can be made. When the variations in crystal structure and optical properties of plagioclases of different origins are more fully understood, the correlations may be more obvious.

The anomalous $2V$ values for samples D 638 ($An_{29.8}$) and 4145 ($An_{36.5}$) may be explained by assuming that these two plagioclases are transitional between high- and low-temperature modifications. According to Tuttle and Bowen (1950), $2V_z$ of high-temperature sodic plagioclases is greater than that of the low-temperature forms, but the difference becomes slight near An_{35} (Bowen and Tuttle, 1950, Fig. 4) and does not explain the magnitude of the departure of sample 4145 from the low-temperature $2V$

curve of the present study. However, many of Tuttle and Bowen's points representing natural plagioclases lie below their tentative low-temperature $2V$ curve, and it may be assumed that the difference between the two modifications at An_{35} is greater than they thought it to be. Further evidence of the nature of sample 4145 was given the writer by J. V. Smith of Cambridge University in a private communication. He found that the subsidiary reflections characteristic of the superlattice structure of low-temperature intermediate plagioclases were absent in single-crystal X-ray photographs of sample 4145, which is strong evidence that the sample is not a low-temperature modification. The fact that the refractive index and total birefringence values for the two samples in question are not anomalous is explained by the very slight differences in these properties for high- and low-temperature modifications in this composition range (unpublished data of the writer). The occurrence of samples D 638 and 4145, the former from a pegmatitic vein in dunite and the latter from a position in the Skaergaard intrusion which is more than 500 m below the roof and 3 km from the contact (Wager and Deer, 1939, Pl. 25), illustrates the fact that the thermal modification of a plagioclase cannot always be predicted from its geological setting. From this it follows that none of the samples of the present study necessarily represents the "maximum" low-temperature state, and that the optical-property curves for a hypothetical maximum low-temperature series of plagioclase might differ in some measure from the curves of the present study.

G. CONCLUSIONS

The structural complexities of plagioclase, which are not yet fully understood, are reflected in the relations of optical properties to composition. At least some of the complexities can be detected by optical and chemical studies if:

(1) The optical measurements and chemical determinations are sufficiently precise.

(2) The widely known fact that the plagioclase of most samples varies in composition is taken into account, and the optical properties of numerous grains from one analyzed sample are averaged to obtain accurate values for plagioclase of the bulk composition of the sample.

(3) The optical properties of extensive groups of samples of comparable thermal histories are considered separately. Averaging the optical properties of plagioclase of different thermal histories will at least in some cases average out the significant details of the relations between optical properties and composition.

The relations between optical properties and composition in the group of samples of the present study are discontinuous. This is interpreted as meaning that the plagioclase represented by the samples does not all belong to one ideal isomorphous series. It may represent a single unique low-temperature series, or parts of more than one low-temperature series. The solution postulated for the relations is in part subject to change when similarly accurate and extensive data become available for plagioclase of different origins and thermal histories.

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