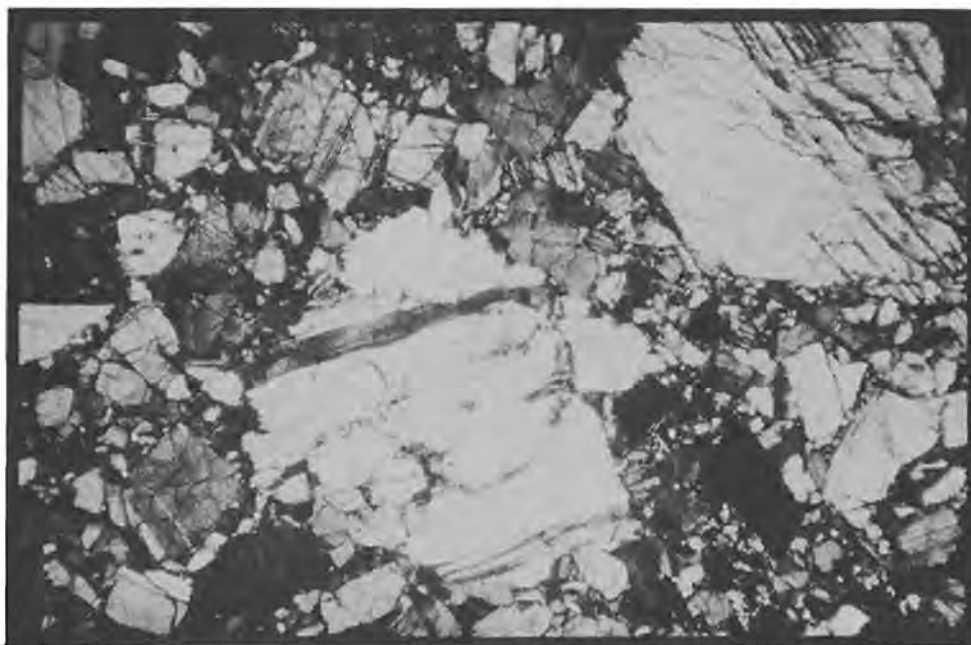


WORKSHOP ON
**MAGMATIC PROCESSES OF EARLY
PLANETARY CRUSTS:**
**MAGMA OCEANS AND STRATIFORM
LAYERED INTRUSIONS**



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WORKSHOP ON
MAGMATIC PROCESSES OF EARLY PLANETARY CRUSTS:
MAGMA OCEANS AND STRATIFORM LAYERED INTRUSIONS

Editors:
D. Walker and I. S. McCallum

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Cover: Cataclastic anorthosite 60025 in crossed polarizers in thin section under a petrographic microscope. Long dimension of photo is 3 mm.

Contents

Introduction	7
Program	3
Summaries of Technical Sessions	7
Session on Terrestrial Layered Intrusions	7
Session on the Petrology, Chemistry, and Chronology of the Lunar Highlands	11
Session on Massif Anorthosites	20
Session on Processes in the Formation of Cumulates	22
Session on Mineralization in the Stillwater Complex	27
Session of General Discussion and Proposed Research Collaboration	31
Abstracts	35
<i>"Unconventional" magmatic processes in layered intrusions—an introduction</i> A. L. Albee	37
<i>Proterozoic anorthosite massifs: a review</i> L. D. Ashwal	40
<i>Investigations of the Stillwater Pt/Pd horizon, Minneapolis adit area: stratigraphic relations, geochemistry and genesis</i> S. J. Barnes	45
<i>The chemistry of pristine lunar samples</i> D. P. Blanchard	49
<i>The size of the Stillwater complex: an estimate from gravity data</i> W. E. Bonini	53
<i>Fine scale rhythmic layering in the Stillwater complex, MT.</i> A. E. Boudreau	56
<i>The main platinum zone Stillwater complex, MT.—evidence for bi-metasomatism and a secondary origin for olivine</i> A. E. Boudreau	59
<i>Layered intrusions: a mini review</i> I. H. Campbell	62
<i>Some isotopic, compositional and thermal aspects of the evolution of magma chambers in the crust</i> D. J. DePaolo	66

<i>Petrological and geochemical characterization of the St. Urbain anorthosite massif, Quebec: summary of initial results</i> L. P. Gromet and R. F. Dymek	72
<i>Some geochemical considerations in constraining anorthosite genesis</i> L. A. Haskin and M. M. Lindstrom	75
<i>Pyroxene-feldspar composition trends in achondrites: parallels to Stillwater and lunar highlands</i> R. H. Hewins	80
<i>Plutonic processes in spreading ocean-ridge magma chambers, illustrated by the Samail ophiolite, Oman</i> C. A. Hopson and J. S. Pallister	83
<i>The efficiency of in situ fractionation: viscosity parameters</i> R. H. Hunter and L. A. Taylor	86
<i>Multicomponent phase diagrams and the phase equilibria of basalts</i> J. Longhi	90
<i>Fe-Ti oxides of the Stillwater complex, Montana: mineralogy, chemistry, and implications</i> W. B. Maze and R. R. Carlson	95
<i>Physical aspects of differentiation</i> A. R. McBirney	99
<i>Magma ocean evolution: raw material and finished products</i> G. A. McKay	101
<i>An example of igneous layering on a meteorite parent body</i> H. Y. McSween, Jr.	106
<i>Cooling history: can we tell a magma ocean from a magma puddle?</i> G. L. Nord, Jr.	109
<i>Radiometric ages and isotopic systematics of pristine plutonic lunar rocks</i> L. E. Nyquist	114
<i>Petrogenesis of Archean anorthosites</i> W. C. Phinney	121
<i>Basic magmatism of the Southern Oklahoma aulacogen</i> B. N. Powell	125
<i>Mineralogy and petrology of layered intrusions: a review</i> L. D. Raedeke	128
<i>Modal and chemical variations in the ultramafic zone of the Stillwater complex</i> L. D. Raedeke and I. S. McCallum	135

<i>Doubly diffusive convection and solute banding in magmas</i> A. Rice	138
<i>Compositional variation within and among pyroxenes and olivines in lunar ferroan anorthosite 60025</i> G. Ryder	141
<i>Limits on the origin of the lunar Mg-suite</i> G. Ryder	144
<i>Pristine lunar highland rocks: hypotheses of origin</i> G. J. Taylor	147
<i>Monomineralicity and oikocrysts: keys to cumulus cooling rates?</i> G. C. Ulmer and D. P. Gold	154
<i>Petrology of pristine nonmare rocks, survivors of the lunar highlands bombardment</i> P. H. Warren	155
<i>Similarities in the regional geology of the midcontinent rift in the Lake Superior region and the Tertiary geology of the North Atlantic</i> P. W. Weiblen	161
<i>Geologic synopsis of the stratiform Fongen-Hyllingen Basic complex, Scandinavian Caledonides, Norway—comparative studies of the lunar magma ocean intrusions</i> J. R. Wilson and K. H. Esbensen	165
Field Guide to the Stillwater Complex L. D. Raedeke and I. S. McCallum	169
Informal Guidebook for the Precambrian Rocks of the Beartooth Mountains, Montana-Wyoming J. Wooden, P. Mueller and D. Bowes	195

Introduction

Basic propositions are periodically reexamined in the course of scientific inquiry. This rarely occurs of its own accord, but rather as the result of progress which calls basic propositions into question.

A basic proposition of igneous petrology has been that large masses of magma cool and grow crystals, and that those crystals and the remaining liquid are distributed principally according to their buoyancy; layered intrusions containing cumulate-textured rocks have been traditionally presumed to have formed in this manner. The cumulate-textured pristine rocks of the early lunar crust sampled in the highlands are presumed to have formed by the same processes as cumulate-textured terrestrial rocks. The hypothesis that the magma from which these pristine samples accumulated formed a primordial, globe-encircling ocean has been a staple of the lunar science literature for more than 10 years. This hypothesis was formulated and extensively embellished in the absence of detailed knowledge of the field relations for these samples and prior to the recognition of many of the complexities that apparently influence the formation of terrestrial cumulates. These complexities have been recognized through detailed petrographic and chemical analysis of samples taken from geologically known terrestrial environments as well as through application of increasingly complex theoretical considerations of heat and mass transport.

The magmatic processes workshop was convened to reevaluate the significance of the lunar highland pristine cumulate samples with the aid of the additional insights provided by the geologically constrained terrestrial investigations of the last decade. This exercise involved a review of the state of knowledge about terrestrial and lunar cumulate rocks as well as an enumeration and reevaluation of the processes hypothesized to have been responsible for their formation, both classically and at present. The workshop was convened in southwestern Montana so that field trips to the Stillwater complex could form an integral part of the workshop. The Stillwater complex is a fairly well exposed, accessible example of a major stratiform layered intrusion. Inasmuch as it is an object of active study by univer-

sity and U.S.G.S. petrologists and geologists, and by mining companies anxious to exploit the significant economic mineralization found there, many knowledgeable parties were able and willing to lead the field excursions.

The extent to which this workshop may have stimulated a reevaluation of the significance of cumulate rocks will not be known for some time, but it was very clear from the discussions that each of the different constituencies represented at the workshop had much to learn from the others. For instance, it was very sobering for lunar chemical stratigraphers (and not a few of their terrestriality-oriented colleagues) to be confronted with virtually the entire range of pristine lunar rock compositions in a single outcrop. Likewise the novelty of a number of processes suggested to be operative in cumulate formation provided food for thought for all. Furthermore it is instructive for all to touch base with the real complexity of outcrop geology. In that the workshop provided these experiences for the participants, it may be judged to have successfully fulfilled its function.

This volume contains the report from this workshop in three major parts. First is a summary of the presentations and deliberations that occurred at the workshop sessions. Second is a collection of abstracts from the invited keynote speakers and additional contributed abstracts. Finally, there is a set of guides to the field trips. This report is submitted for distribution not only to document the activities of this workshop but also to remind the workshop participants and other interested parties of the opportunities for collaborative research in this field. Our colleagues studying terrestrial rocks have undertaken studies of cumulate rocks which deserve as wide a recognition as possible in the lunar sample community. In return, these colleagues may benefit from an appreciation of the level of comprehensiveness and sophistication that has been achieved in the characterization of extraterrestrial materials and that might usefully be employed in terrestrial geology. Clearly it is hoped that this workshop and this report will be a stimulus to the two communities to pool their resources in addressing problems of mutual interest.

Program

The following program of keynote speakers, session chairmen, and summarizers constituted the presentation and discussion portion of the workshop. This portion of the program was held at the Ramada Inn in Billings, Montana and occupied three days. Summaries of and commentaries on these sessions will be found after the program.

The program committee consisted of D. Walker and I. S. McCallum, conveners, and L. Haskin, O. B. James, P. Jones, B. R. Lipin, L. Nyquist, R. J. Phillips, L. Raedeke, J. L. Warner, P. W. Weiblen, and J. Wooden.

Monday, August 3

Introduction to Workshop

***D. Walker and *I. S. McCallum**

Terrestrial Layered Intrusions

Chairman: A. R. McBirney

Summarizer: P. W. Weiblen

Layered intrusions: A mini review

*Campbell, I. H.

Mineralogy and petrology of layered intrusions: A review

*Raedeke, L. D.

Layered rocks in ophiolite complexes

*Hopson, C.

Petrology, Chemistry and Chronology of Lunar Highlands

Chairman: G. L. Nord

Summarizer: G. Ryder

Petrology of pristine non-mare rocks, survivors of the lunar highlands bombardment

*Warren, P. H.

†Pyroxene-feldspar composition trends in achondrites: Parallels to Stillwater and lunar highlands

*Hewins, R. H.

The chemistry of pristine lunar samples

*Blanchard, D. P.

*speaker

†informal contributed presentation

Radiometric ages and isotopic systematics of pristine plutonic rocks

*Nyquist, L. E.

Pristine lunar highland rocks: Hypotheses of origin

*Taylor, G. J.

Discussion: O. B. James, Leader

Areal extent and depth of lunar anorthosites with implications for size of parent magma body

*Longhi, J.

Compositional variation within and among pyroxenes and olivines in lunar ferroan anorthosite 60025

*Ryder, G.

Anorthosites

Chairman: R. H. Hewins

Summarizer: J. L. Warner

Petrogenesis of Archean anorthosites

*Phinney, W. C.

Proterozoic anorthosite massifs: A review

*Ashwal, L. D.

Petrological and geochemical characterization of the St. Urbain anorthosite massif, Quebec

*Gromet, L. P., and Dymek, R. F.

Some geochemical considerations in constraining anorthosite genesis

*Haskin, L. A., and Lindstrom, M. M.

Tuesday, August 4

Processes in the Formation of Cumulates - Part A

Chairman: M. J. Drake

Summarizer: E. Stolper

Multicomponent phase diagrams and the phase equilibrium of basalts

*Longhi, J.

Magma ocean evolution: Raw material and finished products

*McKay, G. A.

Cooling history: Can we tell a magma ocean from a magma puddle? Evidence from microstructures

*Nord, G. L.

Some isotopic, compositional and thermal aspects of the evolution of magma chambers in the crust

*DePaolo, D. J.

Heat and mass transfer

*Walker, D.

Processes in the Formation of Cumulates - Part B

Chairman: A. L. Albee

Summarizer: T. L. Grove

Physical aspects of differentiation

*McBirney, A. R.

Convective fractionation in igneous processes

*Rice, A.

Magmatic infiltration metasomatism: A test of the model

*McCallum, I. S.

†*Fine scale rhythmic layering on the Stillwater complex, Montana*

*Boudreau, A. E.

†*Xenoliths in the Lodgepole intrusion, Montana: Implications for possible northern extension of the Stillwater complex*

*Brozdowski, R.

†*Measurement of f_{O_2} of layered intrusions: Stillwater and Bushveld complexes*

*Ulmer, G. C.

Thursday, August 6

Mineralization in the Stillwater Complex

Chairman: R. Cooper

Summarizer: B. Lipin

Investigations of the Stillwater Pt/Pd horizon: Minneapolis Adit area. Part 1. Stratigraphy, structure and mineralization

Bow, D., Turner, A., *Wolfgram, D., Barnes, S., and Boudreau, A.

Investigations of the Stillwater Pt/Pd horizon: Minneapolis Adit area. Part 2. Geochemistry and genesis

Bow, D., Turner, A., Wolfgram, D., *Barnes, S., and Boudreau, A.

Geology, mineralogy and chemical variations associated with the J-M reef of the Stillwater complex

*Todd, G. S., Mann, E. L., Kuth, D. W., Schissel, D. J., and Irvine, T. N.

**The main platinum zone, Stillwater complex, Montana – Evidence for bi-metasomatism and a secondary origin for olivine*

*Boudreau, A. E.

Summary Session

Chairman: R. Brett

Summarizer: D. Stewart

Poster Presentations

Basic magmatism of the Southern Oklahoma aulacogen

Powell, B. N.

Modal and chemical variations on the ultramafic zone of the Stillwater complex

Raedeke, L. D., and McCallum, I. S.

Geologic map of the Stillwater complex, Montana

Seegerstrom, K., and Carlson, R. R.

Acknowledgments

Logistic and administrative support for this workshop has been provided by P. H. Jones (Project Manager, Lunar and Planetary Institute). This Technical Report has been prepared under the supervision of D. Theiss (Technical Editor, Lunar and Planetary Institute).

Summaries of Technical Sessions

Prepared by: T. L. Grove, O. B. James, B. R. Lipin, I. S. McCallum, G. Ryder, D. B. Stewart, E. M. Stolper, D. Walker, J. L. Warner, and P. W. Weiblen

Session on Terrestrial Layered Intrusions

The opening session of the workshop served as a general introduction to the geology of cumulate-textured terrestrial rocks. Their nomenclature, minerals, textures, ages, compositions, and variability were reviewed. Although this review was principally focused on individual layered intrusions on the continents, such as the Stillwater, Bushveld, Skaergaard, etc., the record of ophiolites and mid-ocean ridges was also considered. The cumulate rocks of Phanerozoic ophiolites and the present oceanic crust comprise the overwhelming volume of cumulate rocks in the Earth's crust today and cannot be ignored in a review of such rocks purporting to have significance for comparative planetology.

Structural setting and geometry of layered intrusions

The structural setting of most layered intrusions is not well-studied, but there is a growing body of evidence that many formed in extensional (rift) environments. This is scarcely surprising in that the prerequisite for a layered intrusion, the delivery of a large volume of magma to the crust, is much more easily realized in a tensional than in a compressional tectonic environment. The Bushveld is apparently an exception to this. Formation of the eastern compartment of the Bushveld can be related to a compressional regime which produced the Transvaal basin by progressive downwarping. Field evidence from ophiolites is beginning to provide insights into the complexity of the dynamic processes of magma chamber formation in what clearly are rifting environments. It appears that periodic replenishment of magma at crustal spreading axes can produce magma chambers with thicknesses of up to 5 km at the spreading axis, thinning out in funnel or half-graben geometry to 10–20 km half-widths. Geophysical and geological data on the Skaergaard, Kiglapait, Muskox, and Duluth complexes also suggest a similar half-graben geometry. A characteristic of both ophiolite sequences and layered intrusions in continental settings is the fact that the margins dip more steeply than the layering. The difference indicates that the rifting in the continental setting is incipient rather than continuing as in the ocean basins. Models of the dynamic steady state of ophiolite magma chamber evolution may shed light on the origin of magma chambers of layered intrusions. The tectonic setting of the magma body(s) parental to the pristine lunar highland samples is entirely conjectural.

Age of layered intrusions

Terrestrial layered intrusions of the continents have an age range from upper Archaean to lower Tertiary. The lower Tertiary age limit is probably a reflection of the time required to expose a magma chamber at the Earth's surface. The evidence from mid-ocean ridges shows the Earth is still able at present to deliver sufficiently large batches of magma to the crust-forming magma chambers and cumulate rock sequences. That layered intrusions older than upper Archaean are unknown may simply be a consequence of an imperfectly preserved record. I. Campbell suggested that early Archaean crust had not stabilized sufficiently to support the loads induced by large volumes of basic liquid. An important point, regardless of the reasons for the age limits, is that the age range of terrestrial layered intrusions covers a very substantial fraction of the Earth's geological record. This contrasts with the limited age range of pristine lunar highland cumulate rocks, which were formed within a few 100 m.y. of the formation of the Moon.

A few secular trends in the characteristics of terrestrial layered intrusions were noted at the workshop. Older intrusions tend to be larger, more reduced, have more economic mineralization, and have orthopy-

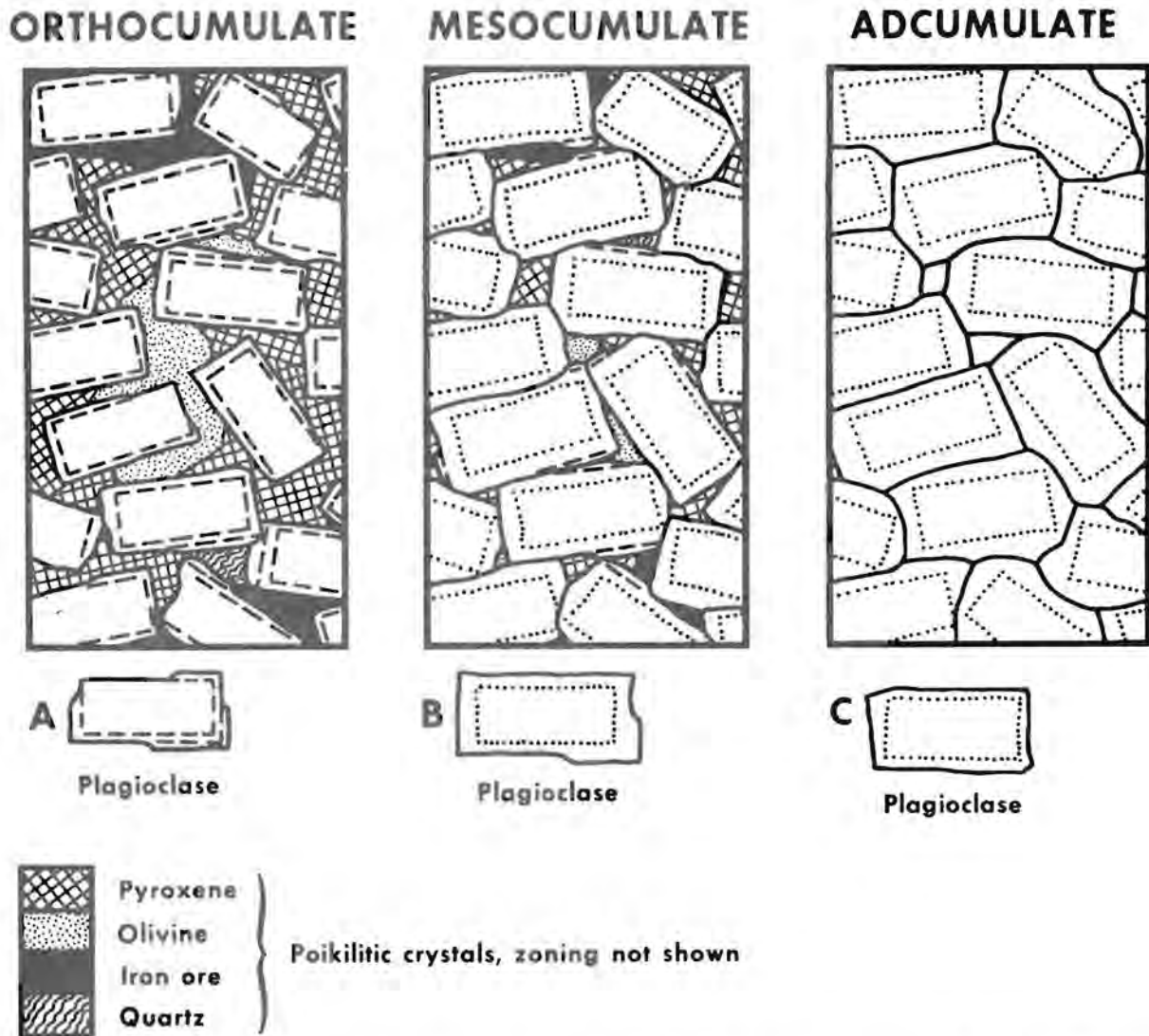


Figure 1 Diagrammatic representation of plagioclase cumulates formed from basaltic magma. Magmatic infiltration metasomatism, double-diffusive fractional crystallization, and adcumulus growth in the muskox intrusion and other layered intrusions, by T. N. Irvine. In *Physics of Magmatic Processes*, edited by R. B. Hargraves, Princeton Univ. Press, 1980. pp. 325-383. With permission of Oxford University Press.

roxene before clinopyroxene in the crystallization sequence. The significance of these trends was not clear to the workshop participants nor was it universally acknowledged that these trends are established. No similar trends in the lunar cumulate samples were identified.

Layering nomenclature

The classic definitions of layering nomenclature appear to be appropriate without significant change in the current textural descriptions and models of crystallization of layered intrusions. In particular the definition of ortho- and adcumulus textures, introduced by Wager and others, was repeated many times

during the workshop and their schematic illustration used to preface all the presentations on crystallization models.

I. Campbell emphasized the significance of the distinction between ortho- and adcumulus textures in light of new studies of the distribution of these textures in the Skaergaard, Kiglapait and Jimberlana intrusions. He has found a close correlation between rhythmic layering and zones of adcumulus textures and uses this correlation as support for an *in situ* mode of crystallization of the layers. It will be of interest to see to what extent future investigations confirm this correlation, and how it is incorporated in models of *in situ* crystallization versus models of gravity segregation.

The textural nomenclature of layered igneous rocks was developed when gravity segregation was considered a major process in the formation of layering. It is reassuring to note that the older textural observations remain valid and that it is the models which are in a continual state of revision. Conversely, the applicability of old terms in the new models emphasizes that correspondence of textures of rocks from different intrusions may not mean that they formed by an identical set of processes.

Magma composition

As in the case of igneous rocks in general, the parent-magma composition of layered intrusions varies with tectonic setting. The intrusions mentioned in the session are all broadly tholeiitic. The well-studied layered intrusions do not commonly have unequivocal chilled margin rocks so there is no way of directly measuring parent-magma compositions. Efforts to determine parent-magma composition from summation and mass balance calculations were mentioned but not discussed in detail.

A persistent, underlying theme discussed in the workshop sessions was the evidence for repeated injections of magma. This dynamic process is central to the interpretations of modal and mineral compositional layering in the models of McBirney and Noyes, Irvine, and Campbell. These models go a long way toward explaining physical and compositional aspects of layering in local sequences of specific intrusions, but they do not explain outstanding stratigraphic problems such as the origin of thick anorthositic units or the reversed sequence in the lower zone of the Stillwater. In view of the fact that in the past magma chambers were considered as more-or-less static crucibles, the direct evidence of the results of growth of magma chambers in ophiolite sequences provides a new perspective for investigating unresolved stratigraphic problems in continental layered intrusions. Hopefully future studies will combine concepts of magma replenishment with those of dynamic growth of magma chambers.

The descriptions of macrorhythmic layers in the Stillwater complex, and ophiolite sequences and observation of such layers on the field trip, prompted questions (discussed further below) concerning the details of processes which produced the successive batches of magma. An unresolved topic was whether the composition remained constant or varied. In addition to the macrorhythmic units, the composition of the magma for the anorthositic units in the Stillwater complex still poses a major problem for study.

Crystallization sequences

L. Raedeke summarized the crystallization sequences for the layered intrusions on which there are sufficient data. In intrusions with exposed ultramafic zones, macrorhythmic units are well-developed near the bottom. In these units, regular repetitions of the crystallization sequences can be attributed to fractional crystallization of consanguineous magmas. The mechanism by which such magmas are replenished to repeat the crystallization remains unexplained. If episodic melting or remelting of a source region is involved, this raises the question of how the rates of magma generation, emplacement, and crystallization are synchronized. This was discussed in conjunction with the formation of ophiolites but no completely satisfactory explanations were offered.

Changes in crystallization sequences have been documented for the Stillwater complex and Muskox intrusion and suggested for other intrusions. Raedeke pointed out that these changes require changes in magma composition or changes in parameters which affect the position of phase boundaries. As new experimental data more closely bracket actual compositions and conditions of crystallization it may be possible to decipher the latter effects.

Reversals in crystallization sequences found in the complex basal subzones of the Stillwater complex and other layered intrusions remain an enigma. By implication, the existence of such crystallization sequence reversals clouds the link between inferring parental liquid composition from textural crystallization sequence. Accordingly the estimates of lunar magma ocean composition made by this method must be regarded with yet a larger measure of uncertainty.

Variations in mineral compositions in layered intrusions

The much-expanded data base on variations in mineral compositions in layered intrusions which has become available in recent years has served more to illuminate the complexity of crystallization processes in layered intrusions than to provide general explanatory models. Raedeke made the important point that observed mineral compositions are the product of a number of processes. Preservation of mineral compositions, in equilibrium with the main magma, established at the time of crystallization can occur only if post-cumulus equilibration with intercumulus melt or subsolidus reequilibration effects are limited. Raedeke and Irvine have demonstrated that modal proportions strongly affect the results of the latter two processes. It is clear that development of criteria for recognizing reequilibration effects are badly needed before valid data on magma compositions can be deduced from cumulates. A further critique of simplified models of equilibria between melts and crystals was provided by Haskin in a review of REE data on anorthosites.

Raedeke pointed out that despite the complexities in variations of mineral compositions alluded to above, a summary of the current data base on variations in plagioclase and mafic minerals for many layered intrusions and lunar samples shows that the complexities can be modelled as exceptions to a general trend related to fractional crystallization. It also appears that despite the complexity of the common occurrence of macrorhythmic units, most layered intrusions show a normal differentiation sequence upwards with late-stage differentiates appearing only in the upper units. This observation indicates that late-stage melts accumulate upward as macrorhythmic units form. Hopson pointed out that in ophiolite sequences late-stage differentiates are also found at the outer and upper regions of the half-graben magma chambers.

Crystallization processes in crystal-melt interfaces

Campbell outlined the current objections to gravity segregation as the major process in formation of cumulates in layered intrusions. He showed photomicrographs of chains of plagioclase and pyroxene crystals in cumulates as direct evidence of self-nucleation in the melt-crystal interface. The texture is analogous to readily observable chains and bundles of crystals in glass in experimental runs, but it is easily overlooked in coarse-grained rocks. Viewed in two dimensions in thin sections it is obscured by intercumulus minerals and adcumulus growth. It is one texture that early workers did not see (except for dramatic examples of harrisitic olivine); consequently, its distribution and extent in even well-studied and well-documented intrusions is not presently known. Because the chain or bundle texture is good evidence of *in situ* crystallization more petrographic data on its occurrence is clearly needed in studies of crystallization in the melt-crystal interface.

There appears to be only rare evidence of Stokes law behavior in cumulates. On the other hand there appears to be abundant evidence to support models of crystallization in the melt-crystal interface as opposed to homogeneous nucleation in the bulk magma and subsequent accumulation. In this regard, a

question was raised concerning the superheat required to inhibit homogeneous nucleation. Discussion of this point indicated the need for caution in ignoring processes of crystal/melt segregation not involving gravity in the formation of cumulates. It was agreed that this neglect has been instinctive for too long.

Session on the Petrology, Chemistry, and Chronology of the Lunar Highlands

This session, chaired by G. L. Nord, was confined almost exclusively to discussions of pristine highland rocks. Polymict breccias and soils were referred to only in passing. The session centered around three invited papers summarizing the petrology, chemistry, and isotopic systematics of pristine rocks; one invited paper summarizing hypotheses for the origin of these rocks; and a discussion devoted to important unsolved problems. The session also included several short informal presentations and was punctuated by comments, questions, and arguments.

Petrology

The first invited presentation, by P. H. Warren on the petrology of pristine rocks, began with a definition of "pristine": produced by endogenous processes. Warren then discussed the first-order problem of recognizing which rocks are pristine. Several criteria can be used, the most common and perhaps definitive of which is a chemical one, that of siderophile-element abundances low enough to indicate a lack of meteoritic contamination. After describing typical pristine rocks, Warren demonstrated that pristine rocks can be divided into two series—the Mg-suite and the ferroan anorthosites—on the basis of mineral compositional variation. From basic geological considerations, he suggested that the Mg-suite rocks are from deeper levels in the crust than the ferroan anorthosites.

R. H. Hewins gave a short presentation on the mineralogy of achondrites and demonstrated that pyroxene-plagioclase variation trends in these rocks are similar to the trends observed in lunar pristine rocks and in rocks from the Stillwater banded zone. With reference to the En versus An diagram on which mineralogic variation trends have been plotted for the lunar and Stillwater samples, diogenites (which mineralogically are orthopyroxenites) have a "normal" sloping trend like the lunar Mg-suite rocks and Stillwater norites and gabbro-norites, and cumulate eucrites (which mineralogically are norites) have a vertical trend like the lunar anorthosites and Stillwater plagioclase-rich rocks. J. Longhi pointed out a critical difference from the Stillwater and lunar trends—the achondrites that show the vertical trend are norites, not plagioclase-rich rocks. L. D. Raedeke briefly explained the model she and I. S. McCallum have proposed for the origin of the sloping and vertical trends and pointed out that the model requires large modal proportions of cumulus plagioclase relative to interstitial liquid in rocks that show the vertical trend. J. L. Warner pointed out that the Stillwater ultramafic rocks show a horizontal variation trend (constant En but variable An) and suggested that the diogenite data more closely correspond to such a trend than to a sloping trend.

Chemistry

The second invited presentation, by D. P. Blanchard on the chemistry of pristine rocks, began with a functional definition of "pristine": rocks that were formed by endogenous processes, are not mixtures, have not been remelted, and predate the episode of intense meteorite bombardment that took place 3.9 b.y. ago. Highly restrictive criteria are commonly used to distinguish pristine rocks. Low siderophile-element content is the single criterion most frequently applied, but A. E. Ringwood and others have argued that endogenous rocks can contain high abundances of siderophiles. Other criteria, such as low incompatible-element abundances, coarse grain size, and homogeneity of mineral compositions, bias the list in favor of accumu-

lates. Thus, the list probably contains only a limited subset of pristine rocks, and models for early lunar history should not be constructed assuming that the list covers the complete range of pristine rocks. Blanchard's presentation was briefly interrupted for a discussion of the pristine-rock criteria and their significance. D. Walker suggested that the criteria might exclude early volcanic rocks. Both Warren and G. Ryder repeated that the list was never intended to be an all-inclusive list of pristine rocks, and in fact, KREEP basalts are included on the list. M. D. Norman emphasized the necessity for using multiple criteria to determine whether or not a rock is produced by endogenous processes by citing the example of impact-melt rocks, which in some cases are not texturally distinguishable from volcanic rocks but are distinguishable by compositional criteria.

The summary was continued by reviewing trace-element distribution coefficients, characteristic REE patterns for important lunar minerals and the effects of fractionation on the REE patterns of liquids. He emphasized an important point with respect to Eu contents of the various pristine rocks; because the distribution coefficient for Eu in lunar plagioclase is close to one and most other minerals contain virtually no Eu, the Eu content of any pristine rock divided by the mole fraction of plagioclase in the rock roughly equals the Eu content of the parent liquid. He then went on to review the compositional characteristics of the pristine rocks, grouped as follows: 1) ferroan anorthosites; 2) troctolites; 3) norites and gabbros; 4) KREEP and quartz monzodiorite; and 5) dunite, spinel troctolite and feldspathic ilmenite. The first four of these groups show systematic, progressive compositional variations. The amount of trapped liquid, total REE content, and Eu content are lowest in the ferroan anorthosites, increase progressively through the troctolites and norites, and are highest in KREEP and quartz monzodiorite. Despite this progression, the various rock groups show differences in slope of their REE patterns that make it difficult to model them as products of a single parent liquid; the differences in Eu contents also suggest different parent liquids. Blanchard suggested that there were several parent magmas, all of which evolved by similar types of fractionation. The rocks of the fifth group all appear to be anomalous and do not fit the trend defined by the other groups; Blanchard suggested that these are nonrepresentative, plagioclase-poor samples from heterogeneous parent rocks.

Blanchard went on to illustrate the Ti/Sm systematics of the pristine rocks. On a plot of Ti/Sm versus $Mg/(Mg + Fe)$, the ferroan anorthosites have Ti/Sm ratios that cluster at values like those in chondrites, whereas the Mg-suite rocks (and KREEP basalts) have much lower Ti/Sm ratios. This compositional dichotomy between the rocks of the two suites further reinforces interpretations that the suites are unrelated. The data have been interpreted as indicating that, prior to crystallization of the Mg-suite rocks, their parent liquids were depleted in Ti relative to Sm by fractionation of ilmenite. Blanchard then illustrated the problems inherent in calculation of the REE patterns of liquids in equilibrium with rock samples. He cited data for norite 78236, where the equilibrium liquid calculated from the compositions of separated minerals is not the same as the equilibrium liquid calculated using whole-rock data in conjunction with mineral modes.

In closing, Blanchard discussed the question of the average composition of the lunar crust. He proposed that the composition of impact-melt rocks produced by basin-forming events is a good approximation of the average lunar crustal section extending to the maximum depth of excavation by the impact. He cited as support, data on terrestrial impact-melt rocks demonstrating that such melts have compositions that are averages of the material excavated. Sample 73215, which has been hypothesized to be Serenitatis fragment-laden impact melt, was cited as a typical example representing the proposed average crustal composition.

Discussion began with G. J. Taylor commenting that Blanchard's proposed average crustal composition is very different from that derived by S. R. Taylor and A. E. Bence, being much richer in incompatible trace elements. Haskin pointed out that data on soil chemistry, both from returned sample studies and orbital geochemical analyses, indicate that no more than 60% of the material in the soils can be ferroan anorthosite

and that a component having a composition like that proposed by Blanchard *must* be important in the upper crust. J. L. Warner suggested that KREEP volcanic rocks extruded relatively late in highland history formed a large volume of the upper crust just prior to the basin-forming impacts, biasing the average composition of impact-excavated material toward that of KREEP.

McCallum, referring back to the problem of calculating equilibrium liquids from REE data for individual rock samples, requested a clarification of the *in situ* crystallization model. He asked if the process of *in situ* crystallization was viewed as completely efficient, with no liquid being trapped. The answer provided by Campbell and McBirney was that during *in situ* crystallization liquid could indeed be trapped, but that the liquid would be a sample of the liquid-crystal boundary layer, not the liquid as a whole, and its composition would be determined by diffusive rather than mechanical processes. McCallum then emphasized that using trace-element modeling to derive the compositions of parent liquids is a very difficult task because of the possibility that trapped liquids are not equivalent to parent liquids; rocks are formed by processes far more complex than those assumed in current modeling techniques. McCallum lamented the tendency to abandon classical ideas of mineral segregation by gravity separation; he cited the great thickness of Stillwater anorthosites—600 m—as evidence that they could not have formed by processes such as *in situ* crystallization and that gravity separation must have been important.

The discussion then turned to the topic of the observed differences in Ti/Sm ratios between the ferroan anorthosites and the Mg-suite rocks. McCallum provided the information that in the Stillwater complex the Ti/Sm ratios in the anorthosites are *lower* than those in the norites, the opposite of the lunar case. Longhi referred back to the discussion on *in situ* crystallization and asked whether such a process could lead to the types of Ti/Sm variations that are observed or if ilmenite fractionation is necessary, as has previously been believed. L. Haskin responded that there are many mechanisms other than ilmenite crystallization that can fractionate Ti from Sm in a melt, but no specific processes were discussed.

Chronology

The third invited paper, by L. E. Nyquist on the chronology and isotopic systematics of pristine rocks, was a peripatetic balance of synthesis and review. Nyquist first summarized current views of the chronology of lunar crustal evolution, assigning fairly large, conservative estimates for the time interval for each episode. Lunar formation took place about 4.55 b.y. ago and was followed by moon-wide differentiation, in the approximate interval 4.4–4.55 b.y. ago. The pristine plutonic lunar rocks that have been studied crystallized between 4.2 and 4.5 b.y. ago. An episode of formation of large meteorite-impact basins took place ~3.9 b.y. ago and may have extended back to ~4.2 b.y. ago. KREEP volcanism occurred in the interval 3.8–3.9 b.y. ago, and mare volcanism 3.1–3.8 b.y. ago. Subsequently, the Moon has been relatively inactive. Apart from the addition of detailed discussions of recent recommendations for revision of decay constants and interlaboratory biases, most of the remainder of Nyquist's presentation followed his abstract closely (this volume). For the Mg-suite rocks, there are abundant K-Ar, Sm-Nd and Rb-Sr data. Interpretation of these data is difficult, however, because different methods have yielded different ages for single rocks. Rb-Sr ages cluster around 4.3 and 4.5 b.y., whereas Sm-Nd ages cluster around 4.2 b.y., with a few near 4.4 b.y. Nyquist discussed at length the question of whether the Mg-suite rocks crystallized ~4.2 or ~4.4 b.y. ago. He cited evidence from Rb-Sr model ages and a plot of initial $^{87}\text{Sr}/^{86}\text{Sr}$ versus initial $^{143}\text{Nd}/^{144}\text{Nd}$ supporting an age of 4.4–4.5 b.y. He suggested a possible alternative, however; that the Mg-suite rocks have a range of ages and some or all of them crystallized from parent magmas formed by remelting of cumulates produced during the primordial lunar differentiation. He closed with a suggestion for research that would help resolve the question of whether or not some of the measured ages have been affected by prolonged subsolidus reequilibration after recrystallization. Needed is detailed knowledge of the P-T conditions under which the rocks formed and

their subsequent P-T histories. This information, in conjunction with knowledge of diffusion parameters of daughter nuclei, should permit an assessment of whether or not the measured age of a given rock is truly its crystallization age.

The discussion began with M. D. Norman citing the 4.3 b.y. Sm-Nd whole-rock isochron obtained by R. W. Carlson and G. W. Lugmair for Mg-suite plutonic rocks and KREEP basalts and asking what mechanisms, other than crystallization from the same magma system, could produce such a linear array. Nyquist did not discuss any specific mechanisms but countered with the example of the Apollo 12 mare basalts, where whole-rock points define a fairly good 5.1 b.y. linear array, as an example of a case where a linear array has no time significance. The discussion then shifted to the isotopic data for ferroan anorthosites. Warren mentioned the coarseness of exsolution lamellae in the inverted pigeonites in the ferroan anorthosites and asked if this characteristic could be used to infer very long periods of subsolidus reequilibration or perhaps to quantify the extent of such reequilibration. McCallum observed that equally coarse exsolution lamellae are present in pyroxenes in the Stillwater rocks, which cooled rapidly compared to the time intervals that could be resolved by the isotopic techniques. D. DePaolo asked how hot the lunar anorthosites were and how deep they formed in the crust. Nyquist did not profess any expertise in this matter but indicated that they would have had to have been at temperatures of 1000–1100°C for significant periods of time if their isotopic systems remained open after they crystallized. O. B. James shifted the topic back to the Mg-suite ages by commenting that the 4.2 b.y. Sm-Nd internal isochron ages for feldspathic ilmenite 67667 and norite 73255,27,45 are best interpreted as crystallization ages, not as the result of partial resetting by subsolidus reequilibration. Textures and mineralogic variations in these two samples indicate much less subsolidus reequilibration than in most Mg-suite plutonic rocks, and both have been interpreted as shallow-level intrusive rocks.

Origins

The fourth invited paper, by G. J. Taylor, summarized the hypotheses for the origins of pristine rocks, concentrating on the magma ocean concept and its historical development (abstract, this volume). The magma ocean concept dates back to the study of Apollo 11 samples, when plagioclase-rich particles in the soils were interpreted as highland samples by J. A. Wood and J. V. Smith and their co-workers. These investigators proposed that the enrichment of the highlands in plagioclase indicated the presence of a large-scale magma body early in lunar history, and that plagioclase floated to the top of this “magma ocean” to form the highlands. Taylor went on to list other evidence for a magma ocean (detailed in his abstract, this volume); he pointed out that this evidence is compelling for early large-scale differentiation but not quite so compelling that this differentiation was by means of a global total-melting event. He outlined most of the magma-ocean hypotheses that have been proposed. Some of these models are qualitative, some describe single important factors affecting crystallization of large magma bodies, and others are quantitative. The latter involve calculations of lunar crystallization using phase equilibria and trace-element partitioning constraints and incorporate physical effects such as the depths of crystallization and magma mixing. Models that produce both the Mg-suite rocks and the ferroan anorthosites from the same magma body are extremely complex.

The question period began with a lengthy discussion between E. Stolper and J. Longhi of Longhi’s modeling of magma-ocean crystallization. Stolper pointed out that a magma ocean of bulk-moon composition would not be plagioclase saturated until a very large fraction of the magma had crystallized, and he found it difficult to envision a process that would permit a large amount of plagioclase to be retained at the top. Longhi suggested that strong radiative cooling in the upper layers would cool the magma rapidly near the lunar surface, causing rapid crystallization and cutting off convection. Stolper pointed out that the upper-

most layers crystallized in this manner should have a bulk composition close to that of the bulk magma. M. J. Drake stated that this rapidly crystallized material formed at the top of the ocean would subsequently sink as blobs into the hotter, still-liquid ocean beneath, and a discussion ensued as to whether plagioclase could be separated from these blobs and concentrated upward as the blobs sank; no consensus was reached. Drake extended the discussion of quenched magma-ocean crust to a possible terrestrial analog, quenched crust on lava lakes. He referred to the visual observations that show quenched crust on lava lakes being rapidly subducted and asked whether this crust is preserved at depth or completely redissolved. Both T. L. Grove and D. B. Stewart responded that the quenched material is indeed preserved as slabs of basaltic hornfels and some of these slabs are overlain by sheets of settled olivine.

D. Shirley then gave an impromptu presentation on a model he is currently developing for the early evolution of the lunar crust. In his model, there was never a moon-wide magma ocean but instead only magma bodies developed at loci of upwelling of mantle-derived partial melts, a situation analogous to that developed at terrestrial mid-ocean ridges. The crystallized tops of these magma bodies are now preserved as the ferroan anorthosites. The Mg-suite rocks crystallized from bodies of mantle-derived melts that separated from the mantle at a slightly later time.

Discussion

After the invited presentation, the session turned to a discussion of major unsolved problems concerning lunar pristine rocks. The discussion centered around a list of topics formulated by the discussion leader, O. B. James (Table 1).

The first set of questions that was addressed concerned the degree of alteration of ferroan anorthosites by metamorphism, shock, and impact-induced granulation. This topic points up one of the major problems in studying lunar rocks—determining their original igneous characteristics is very difficult because the samples are typically very small and not necessarily representative of their parent rocks, there are no data on field relations other than spatial association, and nearly all the samples have been modified to one degree or another by impact processes. James began the discussion by illustrating the textural effects of impact-induced granulation and emphasized that such granulation can produce mixing without necessarily introducing meteoritic contamination. She stated that granulated rocks must be examined very carefully to assess whether or not their bulk compositions and mineralogies are indeed representative of a single parent rock.

The group then discussed the relict textures of the ferroan anorthosites. James illustrated the coarse mosaic texture observed in many anorthosites and offered the opinion that this texture could be formed by igneous or metamorphic processes and is not definitive of origin. A prolonged debate on triple junctions and their significance as petrogenetic indicators then ensued. The presence of triple junctions is commonly used as evidence for metamorphic processes. Nord, however, pointed out that formation of triple junctions is also possible by crystallization of melt and cited the presence of triple junctions in Stillwater adcumulates as an example. McCallum stated that the triple junctions in the Stillwater rocks had previously been considered as subsolidus in origin but agreed with Nord that they might indeed be igneous. Walker added that triple junctions must be metamorphic because once liquid is absent the rock enters the metamorphic realm. G. C. Ulmer cited industrial literature that demonstrates that formation of triple junctions requires the presence of a monomolecular liquid layer along grain boundaries to promote diffusion; the liquid need not be silicate melt, however. Nord disagreed, stating the opinion that a rock can recrystallize wholly in the solid state to produce triple junctions, and that grain boundaries between like phases can straighten and regularize to form lower energy boundaries during subsolidus reequilibration. Drake suggested that the extent of subsolidus modification of textures might be assessed by measurement of interfacial angles around triple junctions;

these angles average 120° but apparently the spread around this value decreases as the extent of subsolidus reequilibration increases.

Ryder pointed out that there are some anorthosites that clearly have been metamorphosed, 15415 for example. James expanded upon this observation by illustrating the texture of 15415, which indicates that the rock was fractured and subsequently recrystallized, but with probably little or no accompanying displacement of mineral grains or mixing. She illustrated the presence of similar recrystallization textures in several Apollo 16 rocks—such as 67075, 62236, and 62237—but these rocks were extensively granulated after they recrystallized and whether the bulk samples are polymict or monomict is difficult to determine.

The debate on triple junctions led into a general discussion of adcumulus character in cumulates. McCallum stated that Stillwater anorthosite layers, which on a 1–10 m scale contain 10–20% postcumulus material, contain hand-specimen-sized domains that appear to be wholly adcumulate. Such gross heterogeneity greatly affects interpretation of the trace-element data. D. Shirley commented that a large proportion of the returned samples of lunar ferroan anorthosites appear to be adcumulates suggesting that they must be adcumulates; if domains rich in trapped liquid existed they would be represented among the returned samples. Norman then asked whether the Raedeke-McCallum hypothesis for formation of the vertical mineralogic variation trend could be reconciled with the adcumulate nature of the lunar anorthosites; the hypothesis involves the presence of small variable amounts of trapped liquid to account for the iron enrichment of the mafic minerals, but the REE data indicate that the rocks contain little or no trapped liquid.

James closed the discussion of ferroan-anorthosite textures by emphasizing that different anorthosites have clearly had different histories and this fact must be taken into account in interpreting the data on isotopic systematics, bulk composition and mineralogic variations. She suggested that detailed studies of rock textures are needed to better understand the histories of these rocks and aid interpretation of the compositional and isotopic data.

James introduced a discussion of the mineralogies of the ferroan anorthosites by illustrating once again the vertical variation trends of En versus An and Fo versus An in these rocks. The gap between the anorthosite and Mg-suite trends has been cited as evidence that the suites are not cogenetic, but both James and G. J. Taylor have found that olivine in sample 64475 has compositions bridging the gap. Taylor and Warren pointed out that 64475 is a dimict breccia—a mixture of anorthosite and VHA-basalt impact-melt rock; although the Mg-rich olivines appear to be within the anorthositic fraction, it has not yet been clearly proven that they were not introduced by the melt.

James asked whether the large variation of En and Fo contents in mafic minerals in single lunar anorthosite samples indicates that the rocks are polymict or if this variation can be explained as a characteristic of the original parent rock. Ryder responded that the variation was primary and illustrated his point with a short presentation of new data for sample 60025 (abstract, this volume). 60025 is a ~1800 g ferroan anorthosite. The part of the sample originally studied was a slab containing only very small amounts of mafic minerals (pyroxenes) whose compositions cluster near the middle of the range shown by the ferroan anorthosite suite as a whole. Ryder's recent inspection of 60025 has indicated that it locally contains more Mg-rich mafic minerals, both pyroxene and olivine, concentrated in clumps 2–3 cm across. His new data show that this single sample contains mafic minerals and plagioclase having ranges of En, Fo and An contents nearly as great as the range shown by the entire ferroan anorthosite suite, and that the most calcic plagioclase is spatially associated with the most magnesian mafic minerals. Weiblen commented that the observed positive correlation of Ca content in plagioclase with Mg content in mafic minerals is very important, as this correlation has not previously been observed in the ferroan anorthosites. McCallum commented that in the Stillwater anorthosites such a large local variation in $Mg/(Mg + Fe)$ of mafic minerals is not observed; the range in En or Fo contents in mafic minerals in a single hand specimen is generally no greater than 1 mole

percent. McCallum suggested that perhaps the lunar and Stillwater anorthosites formed by different processes. James pointed out, however, that the Raedeke-McCallum hypothesis for anorthosite formation could indeed produce a large range of $Mg/(Mg + Fe)$ over very small distances. She suggested that in lunar anorthosites, which would have crystallized under anhydrous conditions, cessation of silicate-melt crystallization might have effectively isolated mafic-rich patches from subsolidus reequilibration with each other, but in the terrestrial anorthosites, the presence of a hydrous phase during subsolidus reequilibration might provide a means for homogenization of compositions over large areas.

James then directed the discussion to consideration of the processes that were involved in genesis of the ferroan anorthosites. She asked whether the rocks were truly cumulates and how such an origin might be proved. Two other processes were mentioned (though not necessarily advocated) as potentially being able to form anorthosites: crystallization of highly feldspathic liquids (Ryder), and metasomatism (C. Hopson). The consensus was that crystal accumulation was far more likely than either of these processes; metasomatism, especially, is unlikely given the anhydrous conditions on the Moon. McCallum solicited any suggestions as to how such thick anorthosite bodies as occur in the Stillwater complex could possibly be made *without* crystal accumulation, but got no immediate offers.

The discussion then moved to a consideration of the questions of the original depths of formation of lunar anorthosites, their thicknesses, and their spatial distribution over the lunar surface. Longhi gave a short presentation in which he contended that anorthosite forms a layer at least 5–6 km thick that immediately underlies much of the lunar surface. He cited orbital geochemical data (primarily Mg/Al ratios) as indicating that the material at the lunar surface ranges from about 60% anorthosite near multi-ring basins on the lunar nearside to about 80% locally on the lunar farside. Laser altimetry indicated 5–8 km of relief in areas where the soils contain 80% anorthosite, indicating that in these areas there is at least 5–6 km of anorthosite. By analogy with the Stillwater complex, 5–6 km of anorthosite would require an original magma body at least 35-km thick. As further evidence supporting his contention that the primordial magma bodies were very thick, Longhi cited the fact that no sample consisting of mafic cumulate rock complementary to the ferroan anorthosite has ever been found. Even the most conservative estimates indicate that the basin-forming impacts excavated materials from depths of 20–40 km in the crust, so the complementary cumulates must lie even deeper.

Haskin gave an impromptu presentation in which he contested many of the points Longhi had made. He disagreed that anorthosite was so abundant at the surface on the lunar farside and estimated that the maximum amount of ferroan anorthosite in the *average* highland soil is only slightly over 50%. The major other soil component, derived from compositional modeling, has noritic bulk composition and has been designated "Highlands Olivine Norite" (HON) by his group. Haskin feels that this norite might be the mafic cumulate complementary to the ferroan anorthosites, or perhaps more likely, equivalent in composition to the parent magma. He suggested that the bulk of the crust has the composition of this norite and anorthosite is greatly subordinate; anorthosite forms such a large volume of the soil only because it is concentrated at the top of the crust.

Drake returned the discussion to a point emphasized by Longhi by asking how deep the Moon did, in fact, melt. He cited as the most compelling evidence for early large-scale lunar melting the observation that primary mare basalt magmas, which are derived from several hundred kilometers deep in the Moon and did not appreciably fractionate during ascent, have negative Eu anomalies. Experimental studies have demonstrated that these basalts cannot have residual plagioclase in their source regions, so plagioclase must have previously been removed from the source regions by some means. The simplest hypothesis for such plagioclase removal is that the mare basalt sources are the complementary mafic cumulates to the ferroan anorthosites. Ryder objected, stating the $Mg/(Mg + Fe)$ inferred for the mare basalt source regions is

0.80–0.85 but for the anorthosites is only 0.70; if the basalt source regions were the complementary mafic cumulates, they should be less magnesian, not much more magnesian, than the anorthosites. He cited evidence from 60025 to the effect that the mafic minerals in the ferroan anorthosites are primary cumulus phases and their $Mg/(Mg + Fe)$ ratios are primary.

When discussion terminated, many points had been aired, and numerous concepts new to a large number of workers had been emphasized and debated. Disagreement was more evident than agreement, except that there appeared to be a consensus that anorthosites are igneous cumulates in their basic characteristics. The discussions provided considerable groundwork for future research on ferroan anorthosites in particular and pristine rocks in general, by delineating some of the questions which must be correctly asked before they can be correctly answered.

Table 1 Discussion topics for lunar session.

I. The ferroan anorthosites—processes involved in their genesis and history

How much have the ferroan anorthosites been altered by processes such as metamorphism, shock, and impact-induced granulation? What characteristics (textural, mineralogical, bulk compositional) were determined by processes that were involved in genesis of the rocks and what characteristics have been altered? (With respect to mineralogy: why do some anorthosites show a large range in mafic mineral compositions? Does this variation indicate that the rocks are polymict or does it reflect the scale of compositional heterogeneity within the parent rocks?) How different were the histories of the different samples? Have any samples escaped alteration?

What processes were involved in the genesis of these rocks? Are they truly cumulates? If so, did they form as orthocumulates, mesocumulates, or adcumulates?

How should we interpret the isotopic data for these rocks? To what extent do the measured ages represent events such as original igneous crystallization, subsequent metamorphic reheating, and impact-induced excavation? What is the significance of the 4.1–4.2 AE ages?

What were the original depths of formation of the rocks? What parameters can we use to help evaluate this question (mineral homogeneity/heterogeneity? grain size? exsolution phenomena?)?

What is the spatial distribution of the ferroan anorthosites over the lunar surface? How similar/dissimilar are ferroan anorthosites at individual landing sites and from different landing sites?

II. The Mg-rich rocks—processes involved in their genesis and history

How much have the rocks of the Mg-rich suite been altered by processes such as metamorphism, shock, and impact-induced granulation? Can we assess the effects of such processes and estimate to what extent the textures, mineral compositions and bulk compositions have been altered? How different were the histories of the different samples? Have any samples escaped alteration?

What processes were involved in the genesis of these rocks? If they are truly cumulates, did they form as orthocumulates, mesocumulates or adcumulates?

What were the postcrystallization thermal histories of these rocks? What were their original depths of formation? (Can we arrive at a relative depth scale? Can we calibrate such a scale?) What parameters can we use to help evaluate these aspects of rock history?

How many compositional subgroups of Mg-suite rocks are there at each of the landing sites we have sampled? How similar/dissimilar are Mg-suite rocks from different landing sites?

How should we interpret the isotopic data for these rocks? To what extent do the data reflect original igneous crystallization, subsequent metamorphic reheating, impact-induced excavation, or later shock events? To what extent does the wide range of ages inferred for these rocks, from 4.08 AE to 4.5 AE, reflect a real range in crystallization age?

Are these rocks related to KREEP? Are they related to highly evolved rocks such as felsites and quartz monzodiorite? If there is a relationship, what is its nature?

III. Relationships between the two suites

Are the rocks of the two suites related? What is the evidence bearing on the presence/absence of a relationship (mineralogic variations, trace-element contents and ratios, Sr isotopic ratios, spatial association)?

What do we know about the composition(s) of the parent magma(s), with respect to major-element, minor-element, and trace-element characteristics? What constraints can truly be placed on the compositions of parent magmas (e.g., rare-earth patterns) using the analyzed compositions of the lunar rocks and their constituent minerals?

What is the significance of the Ti-Sm-Sc variations observed in the pristine rocks? What processes might lead to the observed variations?

Where do samples like the dunite 72415 and the Apollo 16 gabbros and lherzolite fit into the two suites? Mineralogically these are Mg-suite rocks, but either calculated REE patterns of the parent liquids and/or Ti-Sc-Sm systematics suggest a link with the ferroan anorthosite.

What is the vertical and lateral distribution of the ferroan anorthosites and the Mg-suite rocks in the lunar crust?

IV. The lunar magma ocean

Was there a lunar magma ocean? What is the evidence for extensive early differentiation of the Moon? To what depth did this differentiation reach? Is total melting to this depth required, or could partial melting processes account for the data equally well?

What size magma bodies did the ferroan anorthosites crystallize from? Are they relics of a primordial moonwide magma ocean, or did they form in smaller bodies? What are the complementary mafic cumulates like, mineralogically and compositionally? Where might they be in the Moon? Do we have any samples that might represent these cumulates? If not, why not?

What size magma bodies did the Mg-suite rocks crystallize from? Are they relics of a primordial magma ocean, a primordial magma lake or later intrusions into a primordial crust?

What is the composition of the lunar interior? Of the bulk moon?

Session on Massif Anorthosites

Anorthosites are igneous rocks consisting of greater than 90% plagioclase feldspar. There are three general classes of anorthosite bodies known on Earth: (1) anorthosite portions of layered igneous complexes, (2) anorthosite massifs of Proterozoic age; and (3) anorthosite bodies of Archean age. In addition, anorthosites form one of the major types of coarse-grained igneous rocks returned from the Moon.

A major distinction among the terrestrial anorthosite classes is in the bulk composition of the complexes in which they are found. Anorthosite-bearing layered igneous complexes have roughly basaltic bulk compositions. The anorthosites may have been derived by some form of crystal-liquid fractionation. In distinct contrast, Archean and Proterozoic massif anorthosites are found in complexes that are themselves highly feldspathic, raising the possibility that feldspathic parent liquids may have been involved. Either these anorthosites formed from feldspathic magmas of uncertain origin, or they accumulated from cotectic magmas and have subsequently been separated from mafic and/or ultramafic rocks that are needed to balance their highly feldspathic composition.

The session on massif anorthosites was devoted to a review of the characteristics of massif anorthosites to seek genetic links and/or similarities among them that might shed light on the origin and development of their parent silicate melts.

Table 2 contains a set of characteristics of anorthosites to aid in the comparison among Archean, Proterozoic, and lunar examples. Data for this chart have been extracted from the papers in this volume by Phinney and Ashwal for the Archean and Proterozoic anorthosite massifs.

The identification of rocks cogenetic with the massif anorthosites is a central issue in determining the bulk composition of the whole complex. Ashwal pointed out that Proterozoic anorthositic massifs have associated suites of orthopyroxene-bearing granitic rocks that are of about the same age as the anorthosites themselves. These granitic rocks always intrude the anorthosites and may represent independent melts produced as a consequence of the intrusion of hot anorthosite massifs. Proterozoic anorthosite massifs also contain a very minor amount of mafic dikes and cumulates. Although many workers have assumed that the orthopyroxene-bearing granitic rocks are genetically associated with the anorthosites, Ashwal argues that this is incorrect. He feels that the very minor mafic dikes and cumulates are geochemically related to the anorthosites whereas the abundant granitic rocks are not geochemically related.

Phinney and Ashwal pointed out the curious compositional and textural differences between Archean and Proterozoic anorthosites. Plagioclase in Archean anorthosites typically forms equant euhedral crystals of An_{90-90} . In contrast, plagioclase crystals in Proterozoic anorthosites are tabular with lower An contents (typically An_{50}). No hypothesis was offered to explain these differences.

A critical contribution to the session was the paper presented by Gromet (Gromet and Dymek, this volume) on the St. Urbain massif anorthosite of Proterozoic age. Gromet described the St. Urbain complex with its unique occurrence of anorthositic dikes that cut a texturally and chemically distinct anorthosite. The host is a labradorite anorthosite (85–95% plagioclase of An_{58-65}). The dikes range in thickness from 1 cm to over 10 m and consist of andesine anorthosite (over 98% plagioclase of An_{35-45}). The presence of apophyses off

Table 2 Comparisons of characteristics among Archean, Proterozoic, and lunar massif anorthosites.

Characteristic	Archean	Proterozoic	Lunar
Bulk composition	highly feldspathic	highly feldspathic	highly feldspathic
General origin	igneous, mantle derived	igneous, mantle derived	igneous, mantle derived
Plagioclase composition	An _{80 ± 10}	An _{50 ± 10}	An _{96 ± 2}
Plagioclase texture	euhedral (golf balls)	tabular	crushed
Age	3000–2700 m.y.	1700–1200 m.y.	over 4000 m.y.
Area of outcrop	1–1000 km ²	1–50,000 km ²	unknown
Isotopic composition	I _{Sr} = .700 – .701 ε _{Nd} = 0 – +5	I _{Sr} = .703 – .706 ε _{Nd} = 0 – +5	I _{Sr} = .699 – .700 ε _{Nd} = ~0
Geologic occurrence	ensimatic (?)	ensialic	ensimatic (?)
Associated rocks	minor mafic and ultra- mafic cumulates	minor mafic dikes and mafic to ultramafic cumulates; abundant orthopyroxene-bearing granitic rocks	relatively minor mafic cumulates

these dikes, rotated xenoliths in the dikes, and gradations from dikes to anastomosing veinlets clearly point to the emplacement of dikes as anorthositic magma. Similar results were obtained by Wiebe who described anorthositic dikes in the Nain complex. The Gromet, Dymek, and Wiebe data represent evidence for the existence of silicate liquids of anorthositic composition. These data force us to address the problem of the origin and subsequent development of such magmas within the crust of both Earth and Moon.

The primary unresolved problem of massif anorthosites concerns the origin and subsequent development of their parent melts. What is the composition of those magmas? If highly feldspathic compositions are required, how did such magmas originate? How were the magmas emplaced, as crystal mushes or crystal-free liquid? Does the difference in plagioclase morphology and composition between Archean and Proterozoic anorthosites have any bearing on the crystallization history of the parent magmas?

Ashwal and Phinney, in discussing the Archean and Proterozoic anorthosites, suggested similar models of origin. Mafic magma (basalt or komatiite) generated within the mantle rises to the mantle-crust boundary where the magma is ponded due to a density barrier (magma density is less than mantle density so it rises through the mantle, but is greater than the density of the lower crust so its upward migration is stopped). The ponded magma fractionates at the crust-mantle boundary to produce feldspathic magmas that have lower density and may rise into the continental crust. A similar model has been proposed for anorthosites by Emslie. The major problem with this model is that it requires that all anorthosite massifs have left large volumes of mafic to ultramafic cumulate at the mantle-crust boundary.

The contribution by Haskin and Lindstrom addressed the problem of the chemical composition and crystallization history of the parent magma of massif anorthosites. Haskin and Lindstrom evaluated a petrologic model for anorthosites that calls for cumulus plagioclase plus residual magma with the residual magma expelled by a combination of adcumulus growth and filter pressing to yield a rock with as much as 98% plagioclase. They pointed out that with such a process there should be a “base level” trace element concentration, but plots of elemental histograms show that no such base level exists. Also, with such a process the elemental composition should be explained by a linear combination of plagioclase plus mafic minerals, but the rare-earth element pattern for anorthosites commonly approximates the predicted pattern for plagioclase alone, based on trace element distribution coefficients. Yet we know that there are other

components present besides plagioclase. Haskin and Lindstrom argue that the mismatch between predicted and observed element distributions is evidence that the existing petrologic models for anorthosite formation are not adequate to explain the data. More complex models should be generated to explain the data.

Multiple processing of feldspathic magmas may be efficient enough to yield rocks that are virtually pure plagioclase. One example of multiple processing involves a starting system of 90% plagioclase cumulate and 10% mafic component. The system is partially melted to produce 2% melt and half that melt (1% of the system) is extracted. The 2% melt will contain a large fraction of trace elements as dictated by D-values and by extracting half of that melt the remaining system is greatly depleted in trace elements. If the 2% melting with partial melt extraction is repeated about ten times, the remainder is a rock that is virtually pure plagioclase—that is, an anorthosite. For lunar anorthosites, the multiple remelting could be caused by meteorite impacts, but no such simple energy source is available for multiple remelting of terrestrial anorthosites. The concept of multiple processing avoids awkward suggestions such as the requirement for absolute filter pressing to remove all but the final 1% of residual melt. It was generally accepted that we must seek complex processes to explain the quantitative geochemical data.

Haskin's study highlights a feature of many geochemical investigations. The closer the details of a problem are examined, the more often complexities present themselves. Even something as superficially simple geochemically as a pure plagioclase rock does not turn out to be simple on close inspection. The implications of this for studies of lunar rocks, which are characterized by a wealth of detailed data, are disturbing.

Session on Processes in the Formation of Cumulates

The mineralogical and chemical variations upwards through a cumulate pile should reflect the chemical and physical evolution of the magma body in which the cumulates formed. One of the goals of detailed study of a series of cumulates, whether from the Earth or the Moon, is to infer from the observable characteristics of the cumulates the composition and origin of the magma (or magmas) delivered to the magma chamber and the processes responsible for its subsequent evolution. The focus of this session was a consideration of some of the factors affecting the evolution of magma bodies that may influence the observable characteristics of the cumulate pile. These considerations provide a framework within which observations on cumulate sequences can be interpreted.

Parent-magma composition

One of the most readily observable features in a cumulate sequence, and potentially from a single cumulate hand specimen, is the order in which different mineral phases crystallized from the liquid in the magma chamber (i.e., its "crystallization sequence"). The chemical composition of the liquid delivered to the chamber exerts a fundamental influence over the crystallization sequence. The parent liquid composition depends upon the composition of the mantle source region that was melted to form the magma, the depth at which the melting took place, the degree of partial melting in the source region, and the degree to which the liquids were modified during transit from their source regions to the magma chambers. Understanding what the crystallization sequence can tell us about the composition of the parent liquid and what this in turn can tell us about the characteristics of the mantle source regions and their temporal and spatial variations is one of the more important goals of the study of layered rocks from the Earth and the Moon.

In this session, John Longhi reviewed the available data on the phase equilibria of basaltic systems and how they relate to the bulk compositions and crystallization sequences of terrestrial and lunar layered

igneous rocks. The weak link in our understanding of the relevant phase equilibria is clearly the poor state of knowledge of the characteristics of liquids that would be produced by melting of realistic mantle source rocks at the high pressures at which melting is believed to occur in most planetary mantles. Until our understanding of high-pressure phase equilibria improves substantially, it is unlikely that the promise of inferring something about mantle conditions from study of cumulates will be realized.

One important observation made by Longhi is that there appears to be a crude correlation between age, size, and crystallization sequence among terrestrial layered intrusions. Older intrusions (e.g., Stillwater, Bushveld) are larger and crystallized low-calcium pyroxene before high-calcium pyroxene. Younger intrusions (e.g., Skaergaard, Kiglapait) are smaller and crystallized high-calcium pyroxene before low-calcium pyroxene. One intriguing possibility is that this change in characteristics indicates that mantle temperatures were higher in the Archean, but more detailed, high-quality data on the phase equilibria of basaltic systems are needed before this hypothesis can be evaluated.

Intrusion depth

The crystallization sequence of a magma depends not only on the bulk composition of the magma, but also on the pressure at which the magma crystallizes. Various subtleties of the crystallization sequences of layered intrusions may reflect the depth of intrusion. Observed crystallization sequences are typically interpreted in the framework of liquidus phase boundaries for a total pressure of 1 atmosphere, yet the intrusions probably crystallized at pressures of 1–5 kilobars. Indeed, it may ultimately be possible to determine the depth at which an intrusion formed by examining aspects of its crystallization sequence. Unfortunately, as shown by Longhi in his review, there are few available data on phase equilibria in basaltic systems in the 1–5 kilobar pressure range.

Aside from understanding crystallization sequences in layered intrusions, why else might knowledge of the depth of intrusion be important? The thermal and mechanical characteristics of the country rock are strongly dependent on their depth and these characteristics, in turn, exert influence over the amount of contamination of the magma body and the rate at which it crystallizes. The shape of the intrusion strongly affects the fluid-dynamical evolution of the magma chamber, and the shape may be controlled in part by depth. The issue of the depth of intrusion of magma bodies begs a more fundamental question: Why do magma chambers form at all? Do magma chambers form, as suggested at the workshop, simply because the parent magma has ascended to a level where its density is similar to that of the surrounding country rocks and it can rise no further? Or is it related to the lack of availability of suitable pathways by which magma can continue to ascend, reach the surface, and erupt? Without more detailed and systematic information on the depths at which layered intrusions have formed, on the densities of silicate liquids in the crust, and on the properties of the country rocks, these questions cannot be answered.

Thickness of the intrusions

The question of the depth at which an intrusion forms is, of course, ambiguous because most of the intrusions considered in the workshop are thick enough that they evolved over substantial depth intervals. The lunar magma ocean may have been hundreds of kilometers thick, although as pointed out by G. L. Nord in this session, just how thick it may have been is unresolved. The terrestrial layered intrusions such as the Bushveld complex are as thick as 9 km and may be even greater in some cases. Although hardly discussed in this workshop, if there was a lunar magma ocean there must have been a terrestrial magma ocean which might have been hundreds of kilometers thick. These thicknesses translate into pressure differentials ranging from a few kilobars to tens of kilobars between the bottom and the top of the intrusions.

Consideration of this pressure gradient may be important in understanding the evolutions of these magma bodies. Crystallization sequences will differ at the top and bottom of the intrusion. Again, the limited

data available on phase equilibria above a pressure of 1 atmosphere make even qualitative consideration of this effect difficult. Nevertheless, it is clear that different phases may be crystallizing at the top and bottom of a magma chamber at the same time. This will, of course, only be important if there is a substantial thickness of magma at any one time.

Determining the thickness of magma that existed at any particular time is a difficult problem. The model of sulfide formation in the Stillwater complex presented by the Anaconda group in the session on mineralization requires that a zone of magma at least several kilometers thick was present at the time of Pt-group element concentration.

The nature of convection within a magma body (i.e., laminar versus turbulent) may depend upon the vertical dimension of the molten zone as will the very existence of convective instabilities. This, in turn, will strongly influence the way in which the intrusion solidifies (e.g., from the top or the bottom) and stratifies. The amount of assimilation of country rock will also be dependent on the size and shape of the intrusion.

Other means of determining magma body thickness must be developed, not only because of the important petrological implications of this parameter. Nord discussed how textures, diffusion controlled concentration gradients, and exsolution lamellae thicknesses might be used to establish the depth within the Moon at which a highland cumulate may have cooled, thus setting a lower limit on magma ocean thickness. In principle, this approach could also be applied to terrestrial intrusions. The evidence from terrestrial cumulate sequences is that lunar-scale exsolution textures could be developed on rather small magma bodies. Nord emphasized, however, that considerably more data on diffusion coefficients and on reaction kinetics in plutonic rocks will be necessary before this problem can be approached quantitatively.

The effects of assimilation

The final topic that received attention during this session is the extent to which magma composition, and hence crystallization sequences, of layered intrusions are influenced by contamination of the melt by country rock. D. J. DePaolo discussed some of the important factors involved in assimilation processes. He then presented a model of these processes and explored some of the expected consequences of the model. It was emphasized that country rock assimilation takes energy; the energy is made available by the release of latent heat accompanying the crystallization of the melt. An important consequence of this is that the melt produced by an assimilative process will not, in general, simply be a mixture of the initial melt and the contaminant compositions. Some of the important factors controlling the extent of contamination include the size and shape of the intrusion, its depth, and the volatile contents of the melt and the country rocks.

These considerations have several interesting and important implications. Assimilation can produce effects very similar to those of the more simple and widely considered partial melting and fractional crystallization processes. Thus, inferences concerning the petrogenesis of melts in which assimilative processes have not been considered may have to be carefully reexamined. Assimilation may be, in part, accomplished by physical mixing of melted country rock and fractionated parent magma; unfortunately, this process cannot be easily modeled at present because of a lack of knowledge of how melts mix. This same problem was implicit in other sessions in the workshop where difficulties in mixing fractionated and unfractionated melt played an important role in models of magma chamber evolution that are currently under development.

A final consideration concerns the timing of assimilation. By examining chemical variations upwards in a cumulate pile, especially isotopic (e.g., Nd, Sr) variations, it is in principle possible to determine when assimilation actually took place. Interestingly, the two studied cases of assimilation in layered intrusions display very different patterns. The Stillwater magma may be contaminated based on its Nd isotopic characteristics, but if so, it experienced all of its contamination prior to intrusion. On the other hand, the

Kiglapait magma appears to have experienced no contamination until the intrusion was 90% crystallized, at which point it became progressively contaminated by country rock. More work of this sort will help to establish whether any patterns in the mode and style of assimilation of country rock by melts in layered intrusions can be discerned.

Lunar magma ocean cumulates?

G. McKay addressed the question of whether the known pristine, cumulus-textured, lunar highland lithologies could be comagmatic cumulates produced during the solidification of a magma ocean. He reviewed the trace element characteristics of the FAN, Mg-rich, and KREEP suites and the processes commonly thought to operate in models constructed to account for the contrasting characteristics. He concluded that simple models employing conventional processes are inadequate to explain the variations within the framework of a magma ocean model. He noted a number of the more complex combinations of semi-conventional processes that have been proposed to preserve the viability of the magma ocean model. He emphasized that it would be useful to try to understand the trace element distributions between comagmatic cumulates in a geologically-well-characterized, terrestrial layered intrusion before coming to conclusions about the various complex models proposed for the relation between magma ocean cumulates. This remains to be done.

Controversial processes

All of the processes discussed in the earlier part of this session are not particularly controversial, although the relative importance of the specific processes is not generally agreed upon. There is general agreement that these processes can have strong influences on the final crystallization products of large intrusions. In contrast, the latter part of the session was devoted to the discussion of processes which have been suggested only relatively recently as possibly being important in the petrogenesis of cumulate rocks, and have not been universally recognized as operative.

All of the processes, controversial and prosaic, involve heat and mass transfer. D. Walker reviewed the basic mechanisms by which heat and mass are transferred and emphasized that the transfer processes are coupled. That is, mass transport usually is mechanism of heat transport as well, and *vice versa*. This is particularly important in the geologically relevant situation of density differences existing in a gravity field. Buoyancy-driven mass and heat transport (convection) plays a very important role in the classical notions of magma solidification and cumulate formation. Walker pointed out that buoyancy-driven flows could arise through at least three separate effects that produce density variations on magmas. Thermal expansion of an otherwise homogeneous substance produces density differences. A phase change such as partial crystallization of a liquid system may also produce density variations. And chemical variations, for instance those induced in the residual liquid of a crystallization process, also produce density variations. A system with density variations in a gravity field may be quantitatively characterized with respect to whether convection will occur or not. The Rayleigh number is a well-known non-dimensional parameter combination of use in characterizing a system with buoyancy induced by thermal expansion. Walker generalized this formulation for the other sorts of buoyancy. But he pointed out that all these formulations are for substances with Newtonian rheology, a condition applicable to few geological materials. Consequently the characterization of magmatic systems which may be non-Newtonian can only be done approximately at present. This complication is generally ignored in most treatments of magma solidification, the work of McBirney and Noyes, and Hardee being notable exceptions.

The relative importance of temperature and chemistry on magmatic liquid density was examined in some detail by A. R. McBirney who demonstrated that changes in chemistry of liquids produced by crystal

fractionation are much more dramatic than the changes in density produced by the temperature changes driving the crystal fractionation. Consequently, he postulated a mode of magmatic convection driven by the buoyancy of residual liquids produced by crystal fractionation at the walls and roof of a magma chamber. Cold material, paradoxically, can rise because of its chemical buoyancy in calc-alkaline systems. In tholeiitic systems the reverse could be true; residual liquids might sink due to Fe enrichment. Fractionated liquids generated in side boundary layers will tend to pond at the top or bottom of the chamber as they are generated.

McBirney noted an additional complication that may be important. If colder, chemically buoyant material should be generated as a layer overlying warmer, unfractionated, denser material, a special form of convection may develop. The layers would tend to equilibrate by chemical and thermal diffusion. However, these two processes occur at very different rates—chemical diffusion is slower. Consequently the lower layer would heat the base of the cooler, upper, chemically buoyant layer with relatively little chemical exchange occurring at first. Reciprocally the upper layer would cool the top of the lower layer's top. Thus an unstable density gradient with respect to thermal expansion would be generated within each of the two layers adjacent to their mutual boundary until instability occurred and convection began within each of the layers. This mode of convection, on which separate layers connect internally and separately, is one manifestation of a phenomenon known as doubly-diffusive convection. It can occur when two or more of the mechanisms for producing density variations operate together and have markedly different homogenization rates.

McBirney showed some remarkable movies of experiments designed to model the behavior of a solidifying magma chamber, including the effects discussed above. It was clear that buoyant boundary layers and multi-layered, doubly-diffusive convection can produce spectacular results, which are unanticipated in less imaginative treatments of magma solidification and cumulate formation.

A suggestion made by McBirney and amplified by A. Rice was that the convecting layers of contrasting composition encountered in doubly-diffusive convection may be the precursors to the chemical banding seen in cumulate rocks. This is, of course, a radically different view than the classical one that such banding is the result of episodes of gravitational settling and hydraulic sorting of crystals. In this alternate view, the banding is produced by convection in the liquid state and then imperfectly preserved by crystallization *in situ* of inhomogeneous liquid. Many of the failings of the classical model (e.g., lack of hydraulic correspondence of the proposed settled crystals; impossibly repetitious episodes of saturation changes, etc.) are avoided in this alternate model. However, many people noted that it has yet to be demonstrated that the particular modes of doubly-diffusive convection proposed do in fact scale properly to the magmatic environment.

Rice further emphasized the importance of scaling analysis in characterizing the behavior of dynamic fluids. He gave numerical examples of the behavior that might be expected during the transient decay of doubly-diffusive systems. Unfortunately, a number of the parameters necessary to apply these treatments to magmatic systems are poorly known at present. (Consider that the complex density versus differentiation relations discovered in common tholeiitic systems by Sparks *et al.*, and Stolper and Walker are a very recent enlightenment.) Nonetheless the possible range of phenomena is staggering and will undoubtedly occupy igneous petrologists for quite some time. Rice suggested that such geological researchers had much homework to do in the engineering literature where many of the possible phenomena are already partly quantified.

An additional factor that may complicate the coupled nature of heat and mass transport was noted by Walker. The form of the heat and mass diffusion equations is the same, and formally they could be written with cross terms as a comprehensive system. For instance, one can include a term for mass flow induced by the diffusion of heat down a temperature gradient, or *vice versa*. The phenomena to which these terms would correspond are the Soret and the Dufour effects, respectively. Walker showed the results of experimental

work which demonstrated that the Soret effect is intrinsically large in magmatic silicate liquids. For a given temperature difference, the Soret effect can produce a differentiation of silicate liquid comparable to that generated by crystal fractionation. For some elements the Soret effect produces an even larger fractionation. If the Soret effect fails to be important in magmatic systems it is not because it is a small effect, but rather it is because it develops by chemical diffusion which is a slow process. Bowen pointed out that because heat diffuses more rapidly than mass, any temperature contrasts driving Soret diffusion are likely to dissipate themselves before much diffusion can be accomplished. However, Walker noted that in certain doubly-diffusive systems, the process may operate successfully along with crystal fractionation and cited compositional variation in cumulate mid-ocean ridge gabbros as having the footprint of the Soret effect in evidence. McBirney cited the same sort of evidence (too little Fo variation in olivine for the amount of An variation in coexisting plagioclase) from margins of the Skaergaard complex and also attributed this anomaly to the operation of the Soret effect in conjunction with more normal crystal fractionation.

Yet another class of complications in understanding the constitution of cumulate rocks may be encountered in a partially consolidated cumulate section. I. S. McCallum discussed the possible effects of advective and diffusive percolation of intercumulus liquid through crystalline cumulates, in response to compaction of the section. McCallum provided evidence for such a process (magmatic infiltration metasomatism) from the Stillwater and also from the Muskox intrusion. The operation of this process is best observed in cyclic units where mineral precipitation sequences and chemical variations of solid solutions are both preserved. For instance, in cycles beginning with olivine and proceeding to cumulus olivine + pyroxene one expects and finds a general decrease in $Mg\# = 100 \cdot Mg/(Mg + Fe)$ of the cumulus minerals upwards in the cycle. The start of a subsequent cycle is the return to a rock with only cumulus olivine, presumably reflecting a fresh recharge of primitive magma. One might expect that this mineralogical start of the cycle would also correspond to the return of high $Mg\#$ in the cumulus olivine. In a number of cases, however, the reversion to high $Mg\#$ in the cumulus phases occurs above the first reappearance of cumulus olivine and the change in $Mg\#$ is not abrupt but gradational. These effects are thought to be the result of the upward percolation of low $Mg\#$ intercumulus liquid from the upper levels of the lower cycle into the lower levels of the upper cycle during compaction of the cumulus pile.

A latent thread in the session on controversial processes in the formation of cumulate rocks was the extent to which recognition of such processes is restricted to the terrestrial environment. Clearly the geological context is important in aiding recognition of these processes. Yet it is almost self-evident that if such processes do operate, then they must be considered in the petrogenesis of the pristine lunar highland cumulate rocks also. The sorting out of the merits and significance of the various cumulate-forming processes is an enterprise most usefully attempted in the terrestrial examples of cumulate rocks. The messages in lunar cumulate rocks will receive reliable illumination only when these processes are sorted out. However, the level of analytical attention enjoyed by lunar rocks could be directed towards the terrestrial analogs with beneficial results.

Session on Mineralization in the Stillwater Complex

Four talks, three scheduled and one unscheduled, were presented at the Stillwater Mineralization session. All the presentations concerned the platinum-bearing zone of the Stillwater. Among the scheduled talks, two were presented by the geologists of the Anaconda Platinum Project and one was presented by the geologists associated with the Johns-Manville Platinum Project. The unscheduled presentation was given by

A. Boudreau (Anaconda/Univ. of Oregon) concerning some textural features associated with the platinum-group-element (PGE) zone.

The first Anaconda presentation, delivered by D. Wolfgram, was a description of relations within the Minneapolis adit. According to mapping by Anaconda geologists, the PGE zone, called the H-P reef in honor of A. L. Howland and J. W. Peoples, occurs at about 400 m above the top of the ultramafic zone. At depths of 30–35 m below the mineralization, the regular units (composed of successive layers of plagioclase-bronzite cumulates, plagioclase cumulates, and plagioclase-bronzite-augite cumulates) lose their uniform character and begin to exhibit slump structures, rolls, and other features associated with turbulent deposition. The PGE zone overlies these disturbed layers and is generally composed of a plagioclase cumulate overlain by an olivine cumulate. Above the olivine cumulate is a unit identified by Anaconda as “mixed rock,” which is a mixture of plagioclase cumulate, olivine cumulate, and plagioclase-olivine cumulate. This mixed rock is chaotic, and individual units cannot be traced laterally for any distance. The mixed-rock units are commonly coarse grained and characterized by amoeboidal olivine, which in places forms the cores of large orthopyroxene oikocrysts. Most PGE mineralization is in the mixed rock. A plagioclase-olivine cumulate containing fine- to medium-grained olivine lies above the mixed rock and is in gradational contact with the plagioclase cumulate that defines the upper contact of the H-P reef. Along strike in Anaconda’s section of the H-P reef, the PGE mineralization is erratic and may occur in any lithologic unit within the mixed rock. The PGE’s are concentrated mostly in cooperite and braggite-vysotskite.

Wolfgram’s presentation and her answers to the questions afterward served to point out the extreme lateral variability of cumulates. Lateral variation of texture and mineralogy, although usually not as pronounced as in the H-P reef, is common in the Stillwater complex; therefore, one must be careful to avoid placing too much emphasis on interpretation of a narrow vertical section.

The next presentation was by S. J. Barnes (Anaconda/Univ. of Toronto), who summarized the petrology and geochemistry of the H-P reef and offered an hypothesis for its origin. In general, the reef as exposed in Anaconda’s Minneapolis adit, consists of sulfide droplets in an olivine cumulate containing oikocrystic bronzite. The sulfides are mainly pyrrhotite-pyrite-chalcopyrite-pentlandite, a typical magmatic assemblage. The textures of the sulfide blebs are indicative of an immiscible sulfide liquid. The strong positive correlation of PGE concentrations with sulfur is strong circumstantial evidence that the PGE’s are mostly within the sulfides. Analyses of separated sulfide shows the sulfide liquid contained about 6500 ppm Pd and 2000 ppm Pt, over 100 times more than normal magmatic sulfides, whereas Ni and Cu contents are well within normal limits. Thus, the principal extraordinary chemical feature of the sulfides in the H-P reef is their PGE concentrations.

Barnes noted that any hypothesis must take into account: 1) the exceptionally high PGE concentration in the sulfides; 2) the reappearance of olivine and minor chromite; 3) the similarity of stratigraphic position of the H-P reef in the Stillwater complex and the Merensky reef in the Bushveld complex (both are a few hundred meters above the appearance of plagioclase as a cumulate phase).

First, the easiest way of explaining the reappearance of olivine and chromite is by a new pulse of magma entering the Stillwater chamber. Barnes believes that this liquid would be less dense than the fractionated liquid that was crystallizing mainly plagioclase and bronzite. This buoyant liquid would rise above the residual liquid in the chamber. The new liquid would start to cool, and olivine + chromite would precipitate along with the immiscible sulfide liquid. These olivine-chromite-sulfide-liquid “boulders” then would start to settle through the old liquid and would incorporate some suspended plagioclase and scavenge PGE’s from the preexisting, fractionated liquid.

The exceptionally high PGE concentration in the sulfides is thought by Barnes to be the result of a fairly small amount of sulfide liquid from the new pulse equilibrating with and scavenging the PGE’s from a large

volume of residual magma at least 1-km thick and more probably 4-km thick. Thus, the residual liquid would have to be close to saturation with respect to some PGE alloy.

Barnes explained the stratigraphic position of the H-P reef, and the similar position of the Merensky reef, as the result of changing density of a liquid as it goes through fractional crystallization. The density of a silicate liquid decreases as olivine and bronzite are separated, but begins to increase as plagioclase is separated. Therefore, a primitive liquid entering the magma chamber during the time that olivine and bronzite were crystallizing (as in the ultramafic zone) from the liquid in the chamber would be more dense than the residual liquid and would spread out over the bottom. However, after plagioclase began to precipitate, the density would rise and a new magma entering the chamber would be buoyant. Thus, the first magma pulse would have risen rather than sunk in either the Stillwater or Bushveld complexes, if enough plagioclase had crystallized to sufficiently raise the density of the residual liquid.

Discussion of this hypothesis centered on questions of kinetics: could a fast-rising plume of liquid, rising to the "top" in a matter of hours, and then later "boulders", sinking through the residual liquid at equally rapid rates, equilibrate with the residual magma? Others inquired if the reef contains any physical evidence of the "boulders" hitting bottom after falling 1-4 km.

The next talk was presented by S. Todd of the Johns-Manville Corp. Todd and his co-workers presented a very different picture of the H-P reef on Johns-Manville property, which is west of Anaconda's. The reef is in the same stratigraphic position in the complex; however, much greater regularity of layers is evident in the vicinity of the H-P reef on Johns-Manville's property. Todd showed that in the lower part of the Stillwater complex (up to the vicinity of the H-P reef), the crystallization sequence of silicates was olivine-orthopyroxene-plagioclase-augite. In passing upward from the ultramafic zone to the banded zone where plagioclase makes its first appearance as a cumulus phase, the pattern holds up. Bronzite cumulate at the top of the ultramafic unit is overlain by a plagioclase-bronzite cumulate which is, in turn, overlain by a plagioclase-bronzite-augite cumulate (with some minor anomalies that may be due to mechanical sorting). About 360 m above the ultramafic zone, olivine reappears as a cumulus phase. Nine such olivine-bearing layers are present in an 80-90 m section, and the PGE-bearing rocks are in the zone containing the fifth olivine-bearing layer. In the olivine-bearing zones, the order of crystallization was plagioclase-olivine-clinopyroxene-orthopyroxene. Mapping and drill-hole information reveal that this later crystallization sequence is fairly regular around and within the H-P reef. This regularity is in sharp contrast to the highly irregular and laterally discontinuous structures described by the Anaconda group. The sequence of rock types resulting from the new crystallization order is anorthosite-peridotite-troctolite (from bottom to top) and less commonly anorthosite-olivine norite-norite. The PGE-bearing sulfides are most commonly within the troctolite member of these cyclic units.

Chemically, the H-P reef is characterized by a sharp increase in S, Cu, Ni, Te, and, of course, PGE's. Above the reef, sulfur concentrations are higher than below, suggesting that the liquid was at or near sulfide saturation at all stages above the reef.

As the Anaconda group did, Todd and his co-workers proposed that the most reasonable way to account for the reappearance of olivine and a zone of turbulence below the H-P reef is by a new magma pulse. Likewise, the Johns-Manville and Anaconda groups agree that the PGE's are contained in an immiscible sulfide liquid as a result of the new magma pulse. However, agreement between the groups ends there. Todd envisions that the liquid pulse was a very different composition than the magma responsible for the ultramafic zone (see Fig. 2). Furthermore, this new magma batch would have been rich in PGE's and would have had different Ni and Cu concentrations than the residual liquid but would have had Mg/Fe and Ca/Na ratios similar to those of the residual liquid because of the relatively small change of these ratios in ferromagnesian and plagioclase minerals.

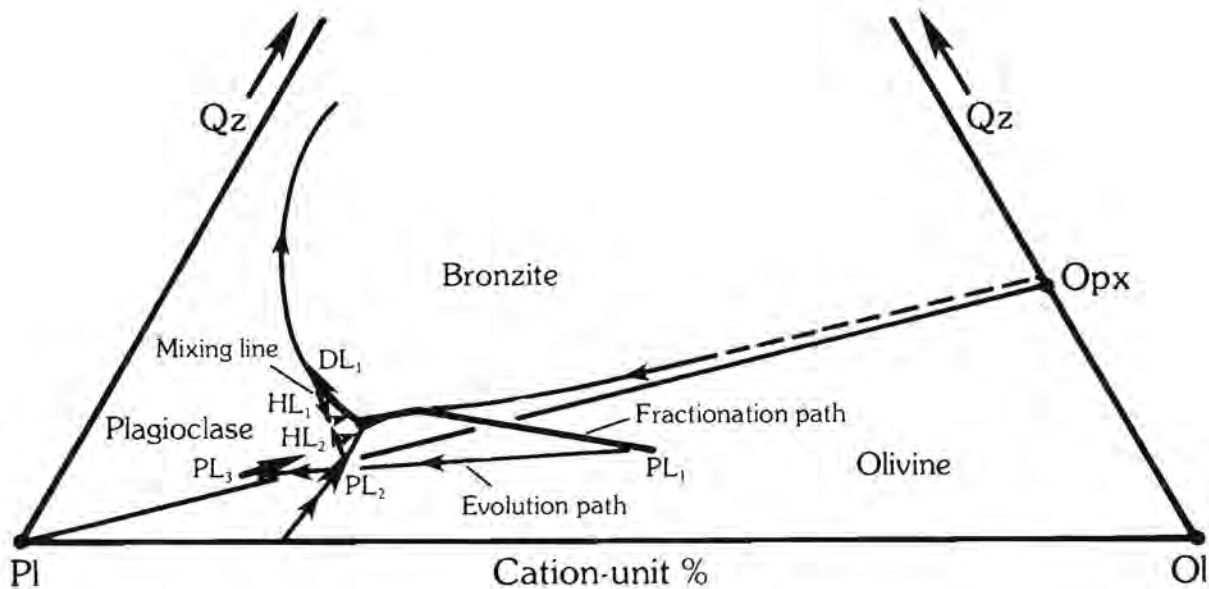


Figure 2 Schematic diagram based on the system $Mg_2SiO_4-CaAl_2Si_2O_8-SiO_2$ and liquidus relations deduced for tholeiitic magmas, illustrating possible effects of mixing a liquid DL_1 differentiated from a primitive liquid PL_1 with another primitive liquid, PL_2 . Hybrid liquids HL_1 and HL_2 are produced. In terms of crystallization order, liquid PL_1 could be parental to the ultramafic zone of the Stillwater complex; PL_2 and the hybrids could be parental to rocks in the vicinity of the H-P reef; and PL_3 could be parental to the thick anorthosite layers that occur stratigraphically higher in the intrusion. From Todd et al. (1979) *Geophysical Laboratory Yearbook (1978-79)*, p. 468.

The new and residual liquids then mixed readily, the new liquid bringing a fresh supply of PGE's to the chamber while the residual liquid, which was close to saturation with respect to sulfides, precipitated the immiscible sulfide liquid.

Some interesting questions are raised by the Johns-Manville hypothesis, such as: 1) Why would liquids evolve from olivine rich to plagioclase rich with time? This trend is on the opposite direction of the change in minimum melt compositions with increasing pressure in the $CaAl_2Si_2O_8-MgSiO_4-SiO_2$ system, which is toward more olivine-rich compositions, and implies progressive melting of an Al_2O_3 -rich mantle source. 2) Did the Bushveld complex also experience an influx of compositionally different liquid at about the same point as the Stillwater during its evolution?

In a fourth, unscheduled talk, A. Boudreau (Anaconda/Univ. of Oregon) pointed out some enigmatic textures in and around the H-P reef, such as large patches of bronzite and olivine surrounded by plagioclase, that Boudreau interprets as the result of recrystallization and remobilization by PGE-bearing volatile components from the ultramafic zone. Boudreau postulated that the volatile components altered the primary silicate mineralogy and chemistry and deposited the PGE-bearing sulfides.

This hypothesis is an interesting and novel approach to the origin of the H-P reef. The hypothesis raises two disturbing questions for igneous petrologists: 1) Can post-crystallization processes fundamentally alter the mineralogical and chemical nature of large areas of layered intrusions? If the answer to question 1) is "yes", then; 2) How much can be deduced confidently about magmatic processes from studying cumulate rocks?

Ideas for research

1) A fundamental difference between the Anaconda and Johns-Manville hypotheses is that one (Anaconda) suggests that the immiscible sulfide melt was provided by the new pulse of magma and this scavenged PGE's dissolved in the residual liquid of the old pulse, whereas the other (Johns-Manville) interpretation is that the new liquid provided the PGE's and the residual liquid provided the sulfur. Thus, it is apparent that we still do not understand adequately the evolution of minor and trace elements in a fractionating liquid. Toward this end, more information is needed on the conditions under which immiscible sulfides melt will separate from a basic liquid and the partition coefficients of the siderophile elements between silicate minerals, silicate liquid, and sulfide liquid.

2) Another question brought out by the two different interpretations of the origin of the H-P reef is, "How readily can two different magmas mix?" Indeed, this was an unanswered question throughout the workshop. Theoretical papers on the subject were cited during the discussions. However, a large body of data (not brought to the attention of those attending the workshop) on magma mixing exists for Hawaiian volcanoes. These data indicate that during periods of active venting, compositionally different magmas mix quite rapidly (weeks), but during inflation (presumably when much less turbulence is involved) compositionally different magmas will sit side by side for years without any substantial chemical exchange. Perhaps it would be very instructive to reexamine magma mixing data on present-day magma chambers.

3) Throughout the workshop, mention was made of the fundamental difference between older (Precambrian) and younger layered intrusions. Older layered intrusions tend to be larger than younger intrusions. Also, older intrusions are orthopyroxene dominant rather than clinopyroxene dominant. In addition, economic concentrations of chromium and PGE's are found in the older intrusions. Possibly, we could learn much about the evolution of the mantle by studying the differences between layered intrusions of various ages.

4) The enrichment process responsible for the PGE mineralization must be complementary to a depletion that must be occurring elsewhere. This depletion process is of considerable interest to those groups studying siderophile element abundances in pristine lunar highland rocks. Much acrimonious controversy has developed about the indigenous level of siderophile element abundances in the Moon. That is, are the low levels observed in some samples the true baseline values uncontaminated by meteoritic debris, or are they artificially depleted by magmatic separation of metal or sulfide phases? It came out in the discussion that the baseline values and the extent of depletion of PGE and analogous siderophile elements was not well characterized for such examples as the Stillwater, where a strong concentration obviously has occurred. We do not know if the bulk of the Stillwater complex, exclusive of the H-P reef, is extraordinarily depleted in PGE and other chalcophile and siderophile elements. Clearly there exists an opportunity for making some basic measurements which could benefit both the lunar and terrestrial community.

Session of General Discussion and Proposed Research Collaboration

R. Brett, as session chairman, remarked that problems as complex as Stillwater petrology naturally lend themselves to the interdisciplinary approach. A thorough understanding of processes operating during crystallization of the Stillwater complex represents good science with wide application to other areas and the Moon, and with application to problems of strategic mineral exploration.

Session summarizer D. B. Stewart suggested that a comprehensive and representative suite of Stillwater samples be carefully curated in a national repository so that investigators could analyze the same

samples using as many techniques as possible. This suite might be a core collected under cooperative government-industry-academic auspices, or might be from a specially prepared outcrop. P. Weiblen suggested that a meter-wide face along the traverse at Mountain View towards the "G" chromitite would be one appropriate locality. Raedeke noted that the average spacing of layers or zones was 10 to 20 meters, and McCallum estimated that to get sufficient details on chemical variations within the mineral phases to resolve whether or not the magma chamber was replenished by new pulses of magmas would require sampling at 2 meter intervals.

Brett stimulated renewed discussion of the processes that could have operated during the crystallization of layered intrusives. The relative contribution of each process to the formation of a given zone, many of the details of each process, and the variation of processes from place to place within the pluton remain enigmatic. There was no consensus as to how units such as the thick anorthosites in the Stillwater complex could have formed. Stewart noted that there had been no discussion at the workshop of the ultimate limits of fractional crystallization in either terrestrial layered intrusions or in the lunar suites. Granitic compositions have been reported to be associated with the Dufek, Bushveld and Duluth complexes.

Much discussion centered around the question of whether layered intrusives result from a single large magma injection, the continued injection over time of the same magma, or the injection of a series of magmas having compositions that varied with time. J. L. Warner argued that evidence from volcanoes is that batches of magma are supplied to the magma chamber over a much longer period of time than is required for any batch of magma to crystallize, and the wide variations in composition over time are so common as to be the rule. P. Weiblen argued that the results of mineralogical and geochemical studies of layered bodies and the mid-ocean ridges indicate that the terrestrial magma chambers enlarge with time, evolving into funnel shaped structures as different pulses of magma are injected. McCallum indicated he would be disturbed by models that *required* that there be a number of magma pulses, and would find it hard to believe that the change in composition of any new magma pulses with time could be so drastic as to completely reverse the order of appearance of liquidus phases, such as causing plagioclase to precede olivine. He asked why is the olivine in Stillwater zone OBZ-I so relatively iron-rich? It obviously did not come from magma of the same composition as the one that produced the ultramafic basal zone. To produce this olivine, new magma of differing composition must have constantly come in, or there must have been instantaneous mixing of new magma of original composition with the residual magma in the chamber, which he did not believe.

In further discussion, I. Campbell suggested that there was a relationship between the number of pulses and the size of the pluton: Skaergaard had 2, Jimbalaya 5, Stillwater 15-20, and Bushveld 30. He regards the iron enrichment to be a secondary process that occurs within the magma chamber and noted that filter-pressing (infiltration metasomatism) can change $Fe/(Fe + Mg)$ by 2%.

G. Ryder noted that the people working with lunar rocks have carefully measured the trace elements in samples and tried to calculate the compositions of their parental magmas. If one did this for samples from near the base of a layered body, from the middle, and from the top, and found consistent results one would know whether there had been any major changes in the magma. If the magma had been replenished or had been mixed with other magma, there could be major changes in the trace elements of the minerals precipitated. L. Haskin reported that these were the very ideas he had had in 1972, when he analyzed a small series of rocks representing a fair span of the Stillwater stratigraphic section supplied by D. Jackson. Unfortunately the results were not interpretable in terms of simple models, and a more thorough study of this sort still needs to be carried out.

Evidence of very voluminous flood basalt flows was cited to argue that single large pulses of magma could be emplaced in short times, but it was not resolved whether crystallization of such a magma body would form a layered intrusion. A. Rice reiterated that pulses are not a prerequisite to get layering, and Brett

noted that every way to account for the observed variation should be sought before invoking new magma pulses. L. Nyquist commented that the fact of a kilometer thickness of orthopyroxenite required that there be a large magma chamber or that it be continually replenished on a rapid time scale; in either case there must have been a large amount of magma present.

The discussion then focused on how much magma might have been present at any one time. Raedeke's data show that the stratigraphic thickness of zones showing the Fe-enrichment trends is 10–20 m, and McCallum using the Rayleigh model calculated that 100–200 m of magma were required to produce these trends. He recognized, however, that only the lower part of the magma chamber may be fractionated at one time. Lipin, arguing that the processes that create ore deposits in layered intrusions are not essentially different from those that create concentrations of other minerals in these bodies, cited the formation of the 'G' chromitite layer as an example. If the 1.2 meters of chromite (~50% Cr₂O₃) crystallized from magma having the estimated Cr₂O₃ content of the bulk of the pluton (0.25%), then ~240 meters of magma must have been present and depleted of Cr₂O₃ to form this one layer.

I. Campbell pointed out that the horizontal extent of the body as well as its thickness were important. A turbulent plume of magma in a 200-meter-thick layer could sustain an 80° spread, but layered bodies were 50 to 100 miles in horizontal extent so that formation from an original 4–6-km-thick layer of melt would not be disproportionate. J. Longhi pointed out that, if the magma body were laterally extensive, they could contain a number of convective cells. What we interpret as multiple intrusives may just be the crystallization front reaching another cell of liquid.

L. Raedeke said she had tried to calculate whether the crust could support the Stillwater magma if it all was present at one time. Although it was difficult to bound such calculations, the result seemed well beyond the ability of the crust to support, so she favored multiple pulses of magma. Earlier in the meeting I. Campbell had ascribed the lack of occurrence of layered igneous complexes prior to 2.75 AE as due to lack of adequate rigidity of the crust in the first 1.75 AE, due to high temperatures, so that it could not support such complexes. Brett speculated that this might be a sampling problem—layered igneous complexes of any age are rare, as are any rocks older than 2.75 AE. Therefore, the probability of finding a very old layered igneous complex is low.

The observation of a constant Fe/(Fe + Mg) trend over great thickness of the Stillwater complex would seem to call for an enormous amount of magma—an amount so large as to be impossible according to McCallum. He stated that, although they studied samples from every available outcrop on several traverses, in fact, the sample spacing in the field is quite wide. The data for small-scale periodic replenishments may be obscured, so that crystallization of an assemblage appears to continue for long periods before significant fractionation becomes apparent. He reported that new data for the orthopyroxenite section does indeed show the presence of thin olivine-bearing layers that suggest magma replenishment. He estimated that samples spaced at about 2 meters would be necessary to give sufficient details on Fe/(Fe + Mg) variation to resolve individual magma pulses.

The concept of a magma ocean on the Moon was subjected to many objections during the workshop, notably from several groups who have proposed that at least one suite of pristine lunar rocks crystallized in intrusions that postdate the primordial lunar crust. P. Warren pointed out that if a lunar magma ocean was present, there must have been a terrestrial magma ocean also, so that this topic was of profound importance to the group. Brett remarked that alternatives to the magma ocean concept had not been as clearly developed and articulated as objections to it.

Phinney and Weiblen reviewed the geologic evidence from the Duluth complex to try to answer the question of an origin from "puddles or an ocean." The geometry of individual intrusions is mostly unknown because of low relief and poor exposures, but individual intrusions cover hundreds of square miles, and the

complex is a combination of layered complexes and multiple intrusions. Overall, the complex appears to fit a dynamic model like that proposed for the mid-ocean ridges, with pulses coming successively to build a large funnel-shaped feature.

Raedeke wondered how the heat is localized to partially melt as much of the mantle as the mass of the Stillwater represents. She disregards the meteorite hypothesis as too unlikely.

Other important questions were put forward, but time constraints precluded their full consideration. A few examples are:

Lipin had pointed out that chromite is sufficiently abundant to form economic deposits only in pre-Jurassic mafic igneous complexes. Brett questioned whether this could be due to a change in the oxidation state of the upper mantle with time. If the older mantle were more reduced, Cr^{2+} would readily enter a partial melt, and as the melt became oxidized in the crust, the Cr^{2+} would oxidize to Cr^{3+} . The lack of high Cr in younger complexes could be ascribed to most Cr being present as Cr^{3+} in the source region. Brett used this speculative example to emphasize the point made by some other speakers that study of layered igneous complexes can teach us much about changes in mantle chemistry with time.

Brett pointed out what seems to be a delicate balance in the formation of sulfide ore deposits in mafic layered igneous complexes. If bulk S is too low, no ore deposit forms. If it is too high, Ni, Co and the platinum elements are diluted by iron in the sulfide phase and a low grade deposit results. He stated that the bulk S content of the Stillwater complex seems to be very low compared to MORB. R. Wendlandt suggested to Brett earlier that this could be because the layered complexes are derived from a deeper source within the mantle than MORB, because Wendlandt has observed a decrease in S solubility in some silicate melts with pressure.

Brett further pointed out that isotope geologists talk of an isotopic disturbance without fully understanding the physical conditions that give rise to the disturbance. Petrologists discuss metamorphic effects without understanding what happens to the isotopes. Understanding is increasing, but research involving an isotope geochemist, an experimental petrologist and an electron microscopist would do much to clarify the interpretation of much of the isotopic data.

Finally, the session chairman wondered aloud if there was any hope of seeing evidence from the Moon for some of the known complex terrestrial processes. He hoped that the interactions of the lunar research community and the terrestrial one could be a long term affair and not just a one week stand.

This session ended abruptly with the unceremonious departure of the workshop participants on the later field trips for Nye, Montana.

ABSTRACTS

"UNCONVENTIONAL" MAGMATIC PROCESSES IN LAYERED INTRUSIONS --
AN INTRODUCTION

Arden L. Albee, Division of Geological & Planetary Sciences
California Institute of Technology, Pasadena, California 91125

The "conventional" magmatic process in layered intrusions forms the title of Chapter IX of Bowen's The Evolution of the Igneous Rocks -- "Rocks whose composition is determined by crystal sorting" -- and he made it clear that this sorting is primarily due to gravity. Bowen also assessed a number of "unconventional" processes and demonstrated to the satisfaction of several generations of petrologists that they were relatively unimportant. The succeeding years brought a succession of more and more detailed studies of layered intrusions (Wager and Deer, 1939; Wager, 1960; Hess, 1960; Jackson, 1961, 1967; Wager and Brown, 1968; Irvine, 1970, 1974; Irvine & Smith, 1967; Loney, et al 1971; McBirney & Noyes, 1979) and led to attempts to better understand the details of the process of crystal sorting. A wide variety of sedimentary structures were recognized and attributed to a number of depositional and erosional processes. However, the repetitive cycles of layers and various anomalies in relative density, mineral composition, and grain size provided difficulties for detailed models. Nevertheless the common explanation for the nature of layered intrusions involved crystal settling and the intermittent formation of density (turbidity) currents spreading out along the floor of the magma chamber (see Irvine, 1980, for a recent review).

In recent years it has become clear that the abundant petrographic, chemical and isotopic data can not be adequately modeled using the major assumption that cumulate crystals settled through the melt to the floor. Some of the "unconventional processes" that have been suggested will be reviewed in this session.

Irvine (1980) has modeled many puzzling details of the Muskox intrusion by an "infiltration metasomatism" model involving an intercumulus melt that is expelled from the accumulating crystal mush by compaction and "diagenesis." The overlying cumulus minerals are changed in composition by reaction with the upward infiltrating intercumulus melt in a process of "infiltration metasomatism" (Korzhiński, 1965). Various dikes and irregular masses in layered rocks may result from such magmatic reaction along relatively restricted channels. A number of papers have argued that the process of "double diffusive convection," known from oceanographic, was an important magmatic process in layered intrusives. (Turner and Gustafson, 1978; McBirney and Noyes, 1979; Chen and Turner, 1980; Irvine, 1980; Huppert and Sparks, 1980; Turner, 1980). This new understanding is based both on detailed computer modeling of the process and on experiments utilizing salt solutions. For example, it has been shown that a high-temperature magma injected into a less-dense, lower temperature magma does not undergo large-scale overturning and mixing, but it spreads out across the floor as an independent layer. With continued cooling from the top, near-horizontal, internally convecting layers will develop. The interfaces between these layers will be maintained because the heat diffusivity is much greater than the chemical diffusivity. Crystallization and the resulting density changes in the residual melt (Stolper and Walker, 1980; Huppert and Sparks, 1980) can produce fluid flows that enhance the vertical density and compositional gradients or can eventually produce large-scale overturns that mix the entire chamber.

There may be other important consequences. The Soret effect was dismissed by Bowen (1928, p.5) as an unimportant magmatic mechanism. However, Walker et al

Albee, A. L.

(1981) have argued that Soret fractionation in a melt may be significant if it operates in conjunction with some form of "double diffusive convection" at a boundary layer. McBirney and Noyes (1979) proposed a mechanism for the formation of fine-scale layering based on a phenomena of chemical oscillation arising because of different rates of thermal and chemical diffusion during crystal growth at a solid boundary.

Similar new ideas may arise from current thermal and fluid dynamical modeling of the whole process of magma genesis and transport and investigations into crystal growth, nucleation and diffusion in silicate melts (see Hargraves 1980, for a review). The "unconventional" processes of magma mixing and silicate liquid immiscibility are being reassessed (Anderson, 1976; Roedder, 1980). The discovery that abundant meteoric water was drawn into layered intrusions by hydrothermal convection (Taylor and Forester, 1973) has not been fully assessed, and there is no clear understanding as to whether such an influx was an important part of the magmatic process.

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Albee, A. L.

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PROTEROZOIC ANORTHOSITE MASSIFS: A REVIEW Lewis D. Ashwal,
Lunar and Planetary Institute, 3303 NASA Road 1, Houston, Texas 77058

Massif-type anorthosites occur as intrusive masses ranging in size from small plutons to huge complexes up to 3×10^4 km² in area. They are generally found within the high grade metamorphic terrains of Proterozoic orogenic belts (1), such as the Grenville Province of North America (2) or the Eastern Ghats of India (3). Many have been metamorphosed, although primary features are preserved to a surprising degree. Notable exceptions are the anorthosites of the Nain Province, Labrador, which are undeformed and unmetamorphosed (4). Shapes of anorthosite massifs range from funnels (5) to lopoliths (6), although many are sheets about 5 km thick with one or more vertical feeder pipes (7). The domical form of many massifs and the existence of large-scale nappe structures involving anorthositic rocks is attributable to post-solidification horizontal compression and diapiric uprise (8).

Plagioclase in anorthositic rocks generally ranges between An₄₀ and An₆₅ mole % (9), although in the Nain complex there is a gaussian distribution of compositions with a peak near An₅₀ and extremes of An₃₄ and An₉₀ (10). Oscillatory, normal and reverse zoning have been reported (11), but compositional homogeneity over wide areas is the rule rather than the exception. In addition to huge volumes of true anorthosite ($\geq 90\%$ plagioclase), the typical massif anorthosite *suite* contains minor facies containing larger proportions of mafic silicates, Fe-Ti oxides and apatite (12). Extreme concentrations of these minerals give rise to minor ultramafic rocks, in addition to the ilmenite-magnetite ore deposits whose association with anorthosite massifs is well known (13,14). Some massifs (e.g., Adirondacks, N.Y.; Morin, Que; Egersund, Norway) contain pyroxenes as the major primary mafic silicate phases, with Fe-rich olivine occurring only as a minor constituent of extremely late-stage differentiates, if at all (12). Other massifs (e.g., Nain; Harp Lake, Labrador) contain relatively Mg-rich olivine as a major constituent of most rock types (4-6,11). Primary compositions of coarsely exsolved augites and inverted pigeonites have been reconstructed using clever microprobe techniques, and suggest minimum crystallization temperatures of 1100°-1200°C (12,15). Pyroxene compositional trends in anorthosite massifs are broadly similar to the iron-enrichment trends of the mafic layered intrusions (12,16). However, the most Mg-rich pyroxenes associated with anorthosite massifs rarely exceed mg^* [molar $Mg/(Mg + Fe)] = 0.70$ (12,16-18), and are thus more Fe-rich than mafic silicates in typical primary mantle-derived melts such as MORB. Giant (up to 50 cm across) pyroxene megacrysts with *exsolved* plagioclase lamellae and mg^* up to 0.81, recently discovered in many anorthosite massifs (19) may represent aluminous pyroxenes originally crystallized at high pressure, and which were carried by ascending magmas to shallower depths where they exsolved feldspar upon cooling. Similar Mg-rich pyroxenes with garnet (H. W. Jaffe, pers. comm.) or spinel (R. F. Emslie, pers. comm.) lamellae have also been found.

Spatially associated with nearly all anorthosite massifs is a suite of orthopyroxene-bearing silicic rocks ranging in composition from granite (charnockite), through monzonite (mangerite) to monzodiorite (jotunite). It has often been suggested that these silicic rocks were comagmatic with the anorthosites, and related by differentiation of a parent magma of roughly granodiorite composition (20). Field evidence (13) and geochemical relations, including REE (21) and isotopic studies (22) are persuasive that the silicic rocks represent independent melts, chemically unrelated to the anorthosites. Instead, it is the minor mafic and Fe-Ti oxide rich facies, and not the silicic rocks which represent late-stage differentiates of the anorthosites (12,23). The silicic suites were probably produced by partial melting of quartzofeldspathic supra-

Ashwal, L. D.

crustal rocks into which the hot anorthosite massifs were intruded (24).

Anorthosites are difficult rocks from which to obtain direct radiometric ages, because they exhibit little variability in Rb/Sr, and rarely contain zircons. As a result, nearly all anorthosite "ages" have come from the spatially associated silicic rocks which in most cases are not comagmatic with the anorthosites. Most published ages of anorthosite massifs range between about 1700 and 900 m.y. (4), although individual massifs as old as 2300 m.y. (25) and as young as 502-540 m.y. (26,27) have been reported. A Sm-Nd whole rock isochron of about 1200 m.y. has been determined for a layered sequence (oxide-rich pyroxenite + leuconorite) from the Marcy massif of the Adirondacks (28). This is 200-300 m.y. older than that given by Rb-Sr isotopic analyses of the same samples, and about 100 m.y. older than zircon ages of nearby charnockites (29). The Sm-Nd whole-rock isotopic technique thus offers great promise for direct determination of anorthosite ages.

The Anorthosite Problem

The problem of the origin of massif-type anorthosites is nearly a century old, and can be elaborated in to the following questions, none of which have satisfactory answers:

What was the composition of the parental magma of the anorthosites? The class of hypotheses which chemically relate the anorthosite and silicic suites, hence implying a roughly granodioritic parental magma, can be eliminated from consideration in most cases. Excluding the silicic rocks, the bulk compositions of the massifs approximate leucogabbro or leuconorite (75-90% normative plagioclase component). Since evidence (direct or indirect) for the existence of large volumes of buried complementary ultramafic cumulates is completely lacking, the possibility of very feldspathic parental melt compositions must be seriously considered. Supporting evidence for the existence of such magmas comes from sills (30) and dikes (31) of leucogabbro or leuconorite, and from chilled margins of the Nain complex with 82-90% plagioclase and 21-25 wt.% Al_2O_3 (32). A satisfactory mechanism to produce such melts, however, has not been demonstrated. An attractive model has been proposed (33) whereby high-Al basalts were produced by high pressure fractionation of olivine and aluminous pyroxene from mantle-derived olivine tholeiite magmas which accumulated near the base of the continental crust. The aluminous melts were then emplaced to higher crustal levels where they precipitated only plagioclase. This model accounts for the relatively low mg* of anorthositic rocks and the occurrence of aluminous pyroxene megacrysts, but is somewhat unsatisfying in its requirement that all anorthosite massifs have left large volumes of mafic cumulates near the Moho. Chilled margins of some massifs (Michikimau, Hettasch, Kiglapait, Harp Lake, Labrador) are high-Al basalts with 18-19 wt.% Al_2O_3 (4,34,46). These complexes show many features intermediate between anorthosite massifs and mafic layered intrusions (35), suggesting the possibility of a continuum of magma compositions from basaltic to anorthositic.

In what manner did these magmas crystallize? With the possible exception of prior high-pressure crystallization of aluminous pyroxene, available textural evidence suggests that plagioclase was the first and only phase to precipitate throughout much of the crystallization history of the massifs. That the true anorthosites are crystal cumulates is a virtual certainty, but the sink/float relationship of crystallizing plagioclase is difficult to evaluate, in large part because field evidence is ambiguous (36). Kinetic considerations such as isothermal supersaturation of plagioclase caused by nucleation of mafic minerals (37,48) have great prospects for solving some of the anorthosite problems, and will likely receive much attention in the years to come. Based on field and textural evidence, and major element chemistries of minerals and whole rocks, it was suggested (38) that the Adirondack and Morin, Que. massifs crystallized

Ashwal, L. D.

only plagioclase until about 65% solidified, when augite and pigeonite precipitated. Fe-Ti oxides (hemo-ilmenite and magnetite) formed at 80% solidified, and at 92%, apatite formed. The very last liquids formed oxide-rich gabbro dikes, from which Fe-rich olivine crystallized. Similar crystallization histories have been suggested for other massifs (39), but their details depend on accurate knowledge of parental magma compositions.

Where and how were these magmas generated? There is little doubt that the magmas which gave rise to the anorthosite massifs, whatever their composition might have been, were ultimately derived from a mantle source. Evidence includes their overall basaltic mineral assemblages, and for the Marcy massif of the Adirondacks, the ϵ_{Nd} value of +5 (28), which implies a light REE depleted source, similar to that of MORB (40). If basaltic melts are the ultimate parents of massif-type anorthosites, it is easier to understand how they may have been generated than how they may have lost the large volumes of mafic cumulates which do not seem to be presently associated with the massifs. More feldspathic primary melts would seem to require atypical mantle source compositions or unusual melting mechanisms. Experimental studies at high pressures have shown dramatic shifts of cotectic compositions toward plagioclase components in both anhydrous (41,46) and water saturated (42) systems. Decompression of such melts, for example by sufficiently rapid ascent would result in plagioclase supersaturated liquids. The absence of primary hydrous phases in nearly all anorthosite massifs, particularly in those rocks thought to have crystallized from late-stage residual liquids is suggestive of anhydrous parent magmas.

When, where and why were these magmas emplaced? Massif-type anorthosites are almost uniquely Proterozoic. To date, no Archean examples have been reported; the Sept Iles, Que, and Wilmington, Delaware complexes appear to be lower Paleozoic (26,27). The idea of a unique, catastrophic anorthosite *event* (43) is unreasonable. A detailed understanding of the apparent temporal restriction of massif anorthosites to the Proterozoic must await further *direct* determinations of anorthosite crystallization ages.

The massifs were emplaced into continental crust. Since the likelihood is high that they were ultimately derived from the upper mantle, their formation represents a major process which contributed to the growth of Proterozoic continents. Levels of anorthosite emplacement varied from lower crustal depths of about 25-30 km for the Adirondack and Morin massifs (44) to shallower depths of about 7-12 km for the Nain complex (45). Differing pressures of crystallization may account for the variable mineralogy among massifs, for example, pyroxene-bearing anorthosites may have crystallized at higher pressures than olivine-bearing ones (46). For deep complexes such as the Marcy massif of the Adirondacks, the metamorphic reconstitution including the production of abundant garnet, may be attributed to slow, isobaric cooling at lower crustal depths, followed by rapid cooling, possibly associated with regional uplift (44,47). Although many massifs are metamorphosed, high-grade metamorphism was not a causative factor in anorthosite genesis (48).

Many authors have speculated on the significance of the apparent spatial restriction of anorthosite massifs to linear belts, and the continuity of these belts on paleogeographic reconstructions of the continents (1,43,39). Proposed tectonic environments for anorthosite genesis range from continental rifts (1,4,45) to convergent plate margins (50), to hot spots (51). An understanding of the origin of Proterozoic anorthosites will provide a link between contemporary and Archean tectonic processes. This should be considered an ultimate goal of anorthosite research.

Ashwal, L. D.

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Ashwal, L. D.

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INVESTIGATIONS OF THE STILLWATER Pt/Pd HORIZON,
MINNEAPOLIS ADIT AREA : STRATIGRAPHIC RELATIONS, GEOCHEMISTRY
AND GENESIS. S.J.Barnes, Dept. of Geology, Univ. of Toronto,
Toronto, Ont. M5S 1A1.

Accumulations of platinum group element (PGE) enriched base metal sulfides occur within a distinctive stratigraphic interval within the Lower Banded Zone of the Stillwater Complex. This interval occurs about 400m above the top of the Ultramafic Zone, is characterized by the reappearance of olivine as a cumulus phase and the first significant appearance of anorthosite, and is defined as olivine-bearing subzone I (OBZ I) by McCallum *et al.* (1980).

Rapid lateral changes in mineralogy and thickness of OBZ I were observed over 700m of underground development. The package pinches and swells, mainly at the expense of hanging wall norites, ranging from 30m to less than 5m in thickness. The thicker portions contain numerous layers and irregular pod-like bodies of coarse grained olivine cumulates rimmed by anorthosite. With decreasing thickness of the OBZ I package, the overall proportion of olivine decreases, and the extent of postcumulus resorption of olivine increases. Irregular pods of very coarse grained gabbro pegmatites are developed within gabbros and norites in the immediate footwall of OBZ I.

Sulfides are irregularly distributed within OBZ I with very little regard for host lithology. Pods of 1 to 3% sulfides occur scattered through the thick, olivine-rich portion of the zone, becoming more regular and stratiform as the zone thins. Texturally, sulfides show clear evidence for an origin as immiscible sulfide liquid droplets, occurring interstitial to cumulus olivine grains. Where extensive postcumulus resorption of olivine has occurred, sulfide blebs are often found enclosed within bronzite oikocrysts. The mineralogy of the blebs is that of a typical magmatic Fe-Ni-Cu sulfide assemblage. PGE are present mainly as the sulfide minerals cooperite and braggite - vysotskite, at the margins of sulfide blebs; a wide variety of PGE tellurides is also found.

Barnes S.J.

A strong correlation is observed between concentrations of sulfur and the PGE, indicating that all the PGE were originally present in the sulfide liquid. This sulfide liquid contained about 6500 ppm Pd and 2000 ppm Pt ; these concentrations are over 100 times greater than the maximum observed in normal magmatic Ni-Cu sulfide deposits. The Ni and Cu contents, however, are typical of sulfide liquids in equilibrium with a normal basaltic magma.

Microprobe investigation of silicate mineral compositions indicates that a trend of increasing Fe/Fe Mg in pyroxenes exists from footwall gabbro-norites through to hanging wall norites, and that no pronounced step back to more Mg-rich compositions is evident within OBZ I. Ni contents of olivines in contact with sulfides are unusually high, and exhibit a strong correlation with Fe content. This is ascribed to equilibrium Fe-Ni exchange between olivine and sulfide liquid, as the olivine became more iron rich during crystallisation of trapped intercumulus liquid.

An hypothesis for the origin of the PGE zone must take into account the following essential features : The exceptionally high PGE content of the sulfides; the reappearance of olivine, sulfide liquid and minor chromite as cumulus phases; the occurrence of pegmatoidal textures; and the stratigraphic position of the zone within the Complex. There is a marked similarity between the stratigraphic level at which PGE mineralization occurs within the Stillwater, and that at which the Merensky Reef occurs in the Bushveld Complex, both being a few hundred metres above the first appearance of cumulus plagioclase.

The major problem is the high PGE content of the sulfides. There is no evidence to suggest that the Stillwater parent magma contained 100 times more Pt and Pd than normal continental tholeiite magmas. It therefore appears that the partition coefficient for Pt and Pd between the Stillwater magma and the sulfide liquid must have been exceptionally high : 10^5 to 10^6 , compared with the values of 1000 - 1500 which appear to apply

Barnes S.J.

in the case of "normal" magmatic sulfides associated with tholeiitic and komatiitic magmas (Naldrett and Duke, 1980). Furthermore, the sulfide liquid must have equilibrated with a large volume of magma , specifically a column of magma at least 1 km thick. Abnormally high partition coefficients might occur if the magma at this level was at or close to saturation with respect to a PGE alloy phase. Efficient scavenging of PGE from a large volume of magma by a sulfide liquid could be the consequence of the injection of a buoyant plume of primitive magma into the chamber.

It has been shown that, during fractional crystallization, the density of the residual magma decreases when olivine is the fractionated phase, and increases when plagioclase becomes a liquidus phase (Sparks et al., 1980). During crystallization of the Ultramafic Zone macrorhythmic cycles, the primitive magma introduced into the chamber would have had a higher density than the differentiated magma in the chamber, and would have formed a layer at the base of the chamber (Huppert and Sparks, 1981). However, at the level of OBZ I, after crystallization of a significant amount of plagioclase, the magma in the chamber would be denser than the primitive magma. A new batch of primitive magma introduced into the chamber at this point would form a bouyant plume, rising through the chamber and spreading out to form a distinct layer at the top (Sparks et al., 1980) or at some intermediate level in the case of a compositionally stratified body of magma . Cooling of this layer of primitive liquid would lead to rapid crystallization of chains of olivine crystals, and to separation of an immiscible sulfide liquid. Composite "boulders" of olivine, sulfide liquid and interstitial primitive magma would settle through the differentiated liquid beneath the primitive liquid layer, accumulating on the floor of the chamber to form olivine cumulates. Plagioclase grains suspended within the differentiated liquid may become entrained in the primitive magma during ascent of the plume, and incorporated into the "boulders". The widespread

Barnes S.J.

development of pegmatoidal textures is attributed to late post-cumulus recrystallization under conditions of elevated volatile contents in the last few percent of the intercumulus liquid.

The volume of primitive magma injected into the chamber must have been small relative to the volume of the chamber, to account for the absence of any major discontinuity in mineral compositions across OBZ I. The high PGE concentrations in the sulfide liquid are due to scavenging of PGE from the differentiated liquid during settling of the sulfide liquid from the primitive layer.

The main attraction of this hypothesis is that it offers an explanation for the stratigraphic position of PGE mineralization in both the Stillwater and Bushveld Complexes, the essential feature being the injection of a primitive magma into a denser differentiate produced by plagioclase fractionation. This hypothesis can also account for the irregular geometry of the olivine cumulates in OBZ I, the extensive resorption of olivine, and the presence of rounded plagioclase inclusions in olivine grains.

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THE CHEMISTRY OF PRISTINE LUNAR SAMPLES

D.P. Blanchard, SN2, Curatorial Branch, JSC-NASA, Houston, Texas.

What is a pristine lunar sample? For purposes of this work a pristine lunar sample is one which:

1. Has not been mixed with meteoritic or other lunar material.
2. Has not been remelted by shock or thermal events since its consolidation.
3. Has retained the products of radioisotope decay intact and in place.
4. Is older than 3.9-4.0 b.y.

Many mare basalts and melt rocks with igneous textures meet the first three criteria. However, if we limit the candidates to those that were formed before the major era of lunar basaltic volcanism (i.e., with ages >3.9-4.0 b.y.), the resulting suite of rocks are those that are commonly called "pristine".

From these old, pristine samples we can get the least complicated look at the materials that were the original crystallization products of the magma ocean. Their ages establish the time line for the early crustal evolution; their phase relationships and compositions put constraints on time-temperature-pressure models. Ultimately we seek the original composition and structure of the moon to relate it to our models for solar system evolution and planetary accretion.

The first difficulty in studying pristine lunar samples is to document their pedigree. As in most exercises of this type there is controversy over the criteria for selection and validation. The following indicators are most often used:

1. Low abundances of siderophilic elements.
2. Evidence for cumulate mineralogy and texture.
3. "Cumulate" chemistry
 - a. Primitive mineral chemistry
 - b. Relatively unevolved trace element distributions.

The most quantitative and least ambiguous criteria for pristinity is the siderophile abundances. Morgan (1972) has estimated indigenous lunar abundances of siderophile elements from clearly pristine rock 15445. Warren and Wasson (1977, 1978) used a somewhat different set of indigenous abundances and with six other discriminants assessed the pristinity of large number of lunar rocks. Delano and Ringwood (1978) argue on analogy to their theories for fractionation of siderophiles in the early, evolving earth that the low levels of siderophiles are far too exclusionary. Those arguments aside, the list of pristine samples compiled by Warren and Wasson (1977, 1978) constitute a nucleus of high certainty, pristine rocks with which we can structure the present discussion.

Norman and Ryder (1979) compiled their own list of pristine rocks which largely overlapped the list of Warren and Wasson and attempted a classification into five groups:

Blanchard, D.P.

1. Ferroan Anorthosites
2. Troctolites and Anorthositic Troctolites
3. Norites and Gabbros
4. Spinel Troctolites, Dunites, and Feldspathic Iherzolite
5. KREEP

By their own admission and as can be seen from ranges of mineral compositions in table 1, the classification is only marginally successful, especially as the rocks become less and less anorthositic.

Rare earth element abundances in these pristine samples show a progression of low abundances ($<1.0x$ chondrites) with sharp positive Eu anomalies (10-20x chondrites) for anorthosites; through somewhat higher abundances (2-8x) and similar Eu abundances (10-20x) for troctolites; through norites with, again, similar Eu abundances (10-20x) but now with sufficient total REE (7-50x) that both positive and negative Eu anomalies are observed; finally to very enriched and fractionated KREEP samples ranging from about 200x (La) up to quartz monzodiorite at about 700x (La) with Eu abundances between 20-40x resulting in sharp negative anomalies. Throughout the progression, the REE patterns become increasingly fractionated (larger La/Lu) with increasing total REE abundances.

Spinel troctolites, dunites, and feldspathic Iherzolites do not fit into this progression. They have widely varying Eu content (1.0x - 10x) and both positive and negative overall slopes. From both mineral chemistry and trace element chemistry this category seems to be a collection of lithologic misfits. Perhaps with the increasing efforts to study smaller rocks and pristine fragments in complex breccias these samples may also be put into a meaningful classification context.

Given the increasing inventory of pristine lunar rocks, what constraints do their compositions offer with respect to the early evolution of the moon and the details of the crystallization of the magma ocean?

A first order observation is that anorthosites and troctolites have large positive Eu anomalies while more evolved lithologies have large negative anomalies. This is consistent with the overall theory of the magma ocean which has Eu-concentrating plagioclase crystallizing early leaving an ever more Eu-depleted liquid from which later lithologies crystallize.

There is evidence from An vs. Mg^* diagrams of plagioclase and coexisting mafic mineral compositions that there are at least two distinct evolutionary trends, the ferroan anorthosites and the magnesian suite, that may not be compatible with a simple, single liquid model. Bulk composition data plotted on variation diagrams designed to clearly show mineral fractionation support this dicotomy. The ratio of Ti/Sm is often used to demonstrate the effects of ilmenite (high D_{Ti}) fractionation. The ferroan anorthosites have nearly chondritic Ti/Sm ratios while the Mg-suite has lower Ti/Sm values. The implication is that the Mg-suite has lost ilmenite from its liquid before it crystallized but that the ferroan anorthosite liquid has seen no ilmenite fractionation.

Another ratio, Sc/Sm, is sensitive to pyroxene fractionation. A plot in Norman and Ryder (1980) suggests that both the Mg-suite and the ferroan anorthosites have lost pyroxene from their parental liquids (Mg-suite somewhat more so than the ferroan anorthosites).

Blanchard, D.P.

Warren et al (1980, 1981) have made an interesting observation regarding the areal distribution and compositions of pristine samples. They argue that otherwise similar groups of rocks show a correlation of Eu/Sm, Sc/Sm, and Ti/Sm with longitude. These rather intriguing correlations suggest that inhomogeneities existed on a major scale in the magma ocean and that simple, single magma evolution models cannot account for these observed differences. This work also reminds us painfully of the inadequacy of the sampling of the lunar surface.

Other modeling difficulties suggest that there is yet much to be uncovered by the careful, insightful investigator. For example, as a general observation, mixing models which derive soil and breccia compositions from endmember lithologies have been largely inadequate. We simply do not know all of the potential end member lithologies nor can we judge which are quantitatively important. In particular we seem to lack detailed information on trace-element rich pristine lithologies (e.g., Wasson et al, 1977).

In summary, there is clear evidence for mineral fractionation in the magma ocean involving plagioclase and pyroxene and, in only some cases, ilmenite. Furthermore, there is evidence that the magma ocean was not homogeneous on a global scale. Finally, there is the real possibility that the processes involved are far more complex than we have modeled and what we now interpret as evidence for separate magmas and large scale inhomogeneities are simply artifacts of our incomplete sample suite and our feeble understanding of early planetary processes. The work on the Stillwater complex by Raedeke and McCallum (1980) seems especially significant in this regard.

However incomplete the sampling and however inadequate our models, the lunar collection continues to be our best tool (perhaps our only tool) to directly study the processes operating in the earliest evolution of planetary bodies.

Table #1: Ranges of Composition of Pristine Rock Groups (after Norman and Ryder, 1979)

	Al ₂ O ₃	Mg*	Fo	Au	Rb
Ferroan Anorthosites	30+			95-98	
Troctolites	20-30	70-80	83-90	95-96	0.2-3.7
Norites and Gabbros	10-23	56-70	65-80	80-95	0.5-13
Spinel Troct. et al	1.2-16	61-85	71-97	89-98	.8
KREEP	13-18	40-80	2-85	50-94	1-40

Blanchard, D.P.

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THE SIZE OF THE STILLWATER COMPLEX: AN ESTIMATE FROM GRAVITY DATA. W. E. Bonini, Department of Geological and Geophysical Sciences, Princeton University, Princeton, N. J. 08544

The Stillwater complex produces a prominent gravity high on, and east of, the outcrop area, where it dips steeply on the north east flank of the Beartooth Range. Results from a field gravity study (1,2) in the Beartooth Range and the Crazy Mt. and Bighorn basins to the north, east and south, reveal anomalously high gravity (Fig. 1). This strongly suggests that eastward of the outcrop area the complex extends in the basement as a relatively flat sheet under sedimentary cover.

The southern limit - a continuation of the steep gradient seen over the complex - extends east-southeast along the Nye-Bowler lineament from outcrop 125 km to Bowler, then changes direction to north-northeast towards Billings for 70 km with a high closure to the west (Fromberg high) and a low closure (Pryor Creek low) to the east. East of the Pryor Creek low is a circular high 15 km in radius (Beauvais Creek high), which is possibly also Stillwater. The steepness of the gradients along the south and east edges suggests that the complex is thick and terminates either tectonically, by a later less-dense intrusion, or is naturally thick at these edges. On the northern edge the complex may decrease in thickness or greatly increase in depth, since the steep gravity gradient is not well-represented.

Gravity Modelling: A gravity model (Fig. 2) was constructed from the outcrop area across the Crazy Mt. Basin, the two highs (Fromberg and Beauvais Creek) and the intervening low (Pryor Creek). The model was made earlier, and new models incorporating more recent ideas are in process. However, this earlier model serves to point out qualitatively, at least, the potential size and volume of the complex. Hess (3) "estimated that another 40 per cent of the complex and its roof are hidden beneath Paleozoic rocks," is iron enriched, and that the average composition of the "hidden zone" is that of a quartz ferronorite. The model assumes densities (gm/cc) calculated from Hess' mineral norm averages for each zone as follows: border zone (3.03); ultramafic zone (2.89); above ultramafic zone (2.87); "hidden zone" (3.03).

A 3rd-degree polynomial surface fitted regional was subtracted from the Bouguer anomaly to produce the residual. The anomaly related to the complex reaches values of +40 mg. East of the Fromberg fault a much greater density (3.15) was needed to produce the steep gradient between the Fromberg high and the Pryor Creek low. A dike-like feature was placed in the model, which would reproduce the magnetic anomaly. It is possible that this feature is a feeder zone. The Beauvais Creek high is modelled as part of the complex but separated from it.

Area and Volume: These results suggest that the Stillwater complex falls in the size range of intermediate mafic complexes: 3000-5000 km² (Table 1), and that less than 5 per cent of its extent is exposed. The volume, assuming a constant thickness over the area (4400 km²) is 22,000 km³, if only the exposed 5 km thickness is used; but if Hess' "hidden zone" is included for a total thickness of 8.3 km, the volume would be 36,500 km³.

W. E. Bonini

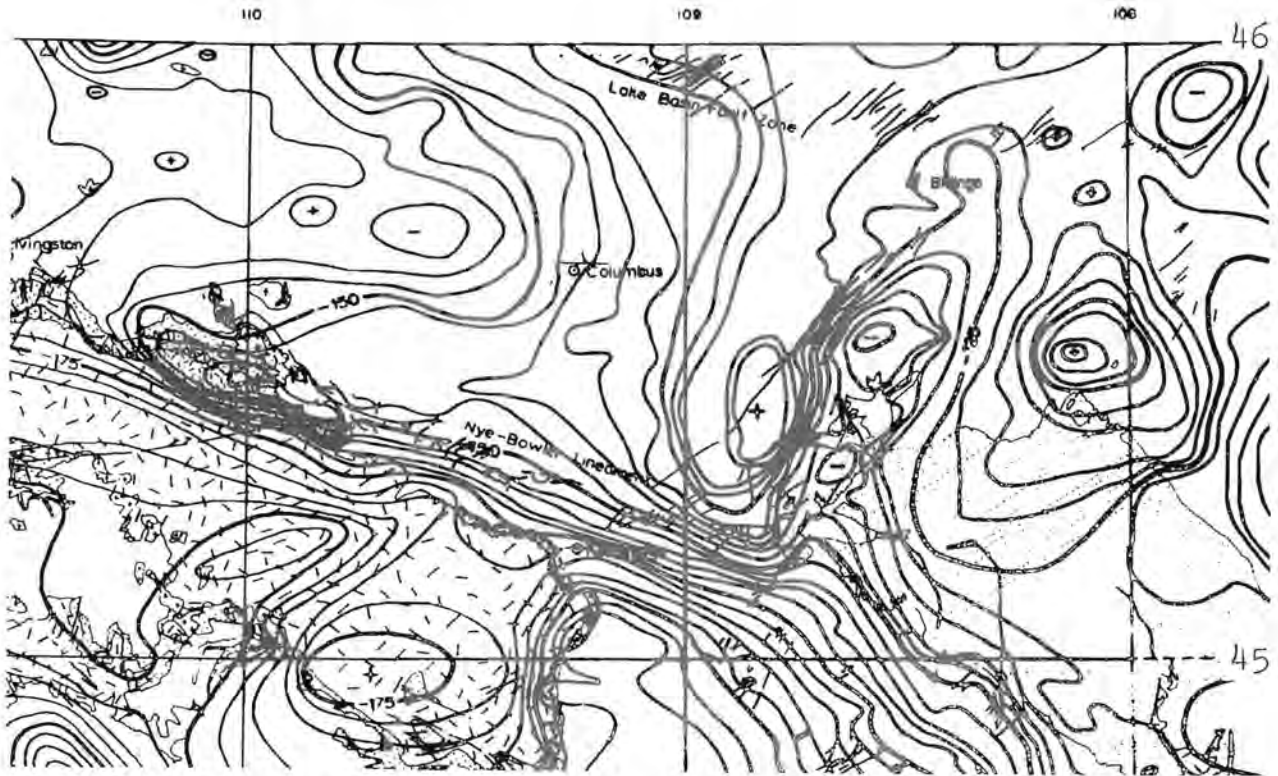


Fig. 1: Bouguer gravity anomaly map, northern Wyoming-southern Montana. Contour interval = 5 mg, distance between Longitude lines = 76 km. Stillwater complex is at Lat. 45.4° and Long. 110.0°

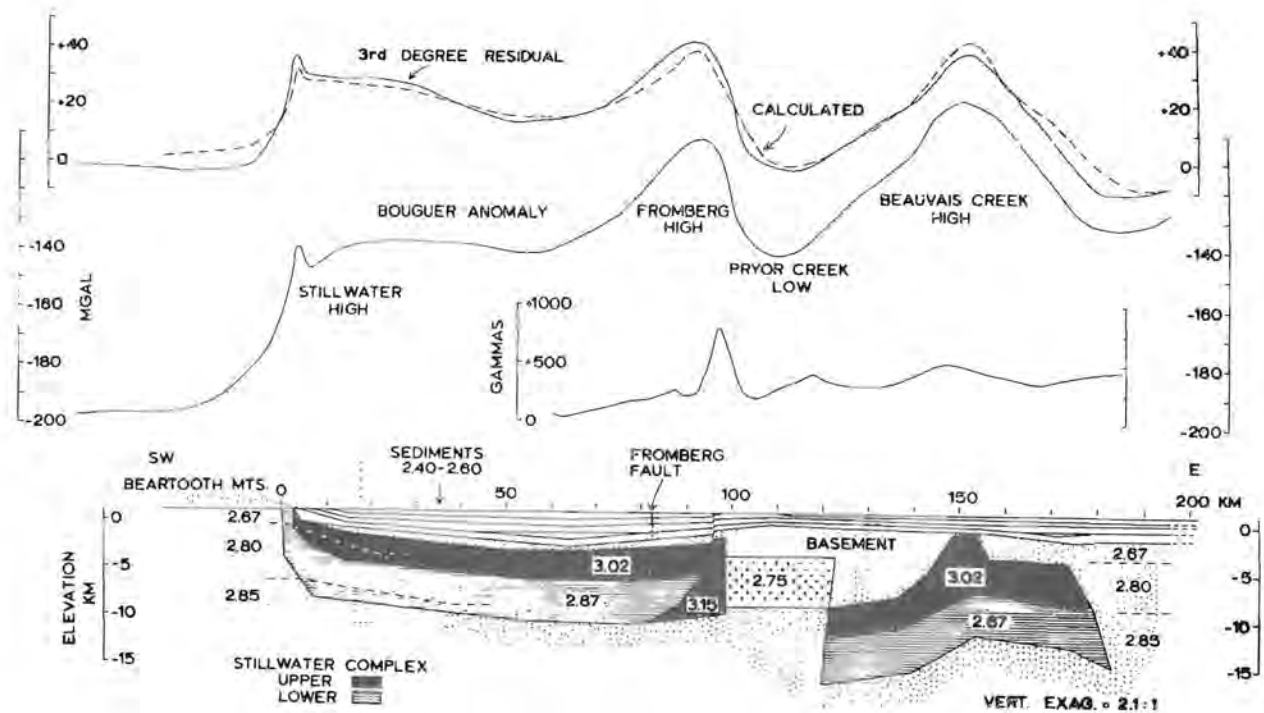


Fig. 2: Gravity cross-section and model. Magnetic anomalies also shown. See Fig. 3 for section line, text for discussion.

W. E. Bonini

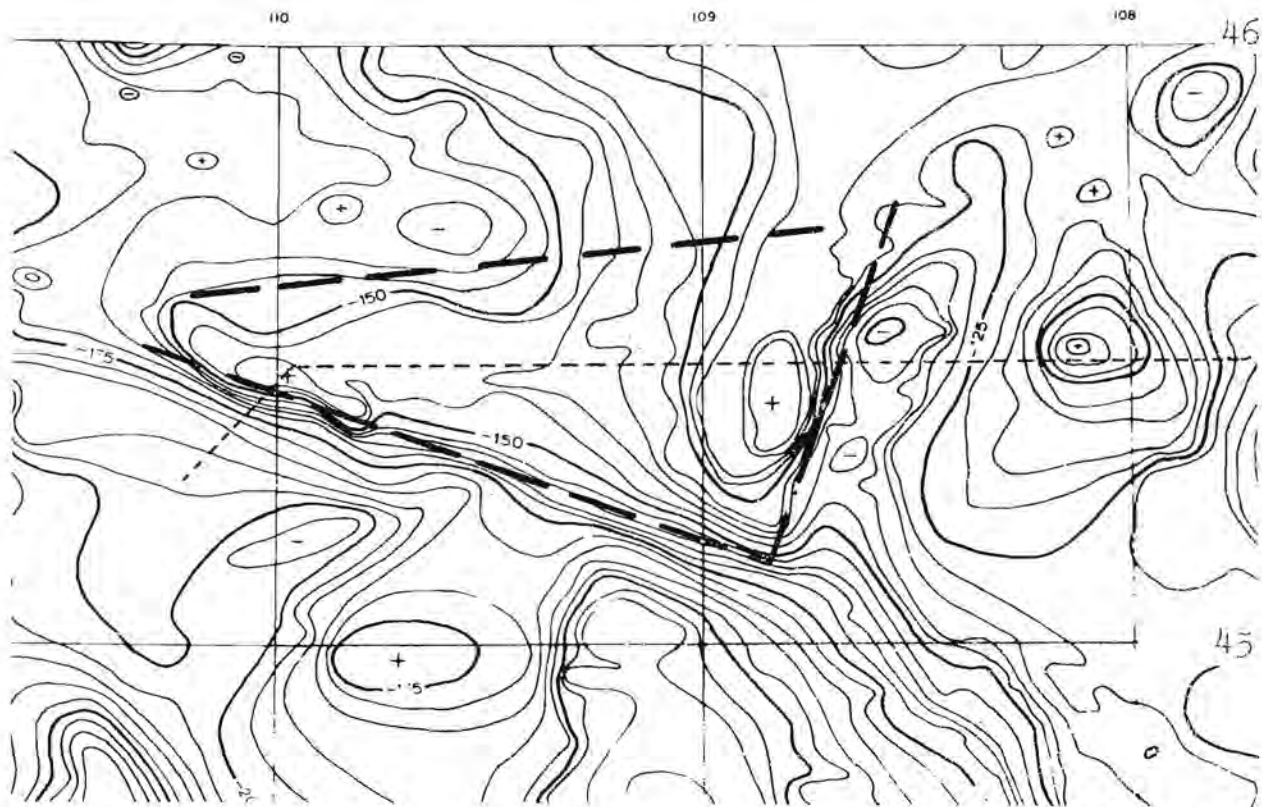


Fig. 3: Bouguer gravity anomaly map, same area and legend as in Fig. 1. Heavy dashed lines enclose area of proposed areal extent of the Stillwater Complex. Cross-section is dotted line.

Table 1: Areal extent of some mafic layered complexes, km² revised after Hunter (4)

Bushveld, South Africa	66,000
*Dufek, Antarctica	50,000 (5)
Duluth, U. S.	4,715
*Stillwater, U. S. (this paper)	4,400+
*Muskox, Canada	3,500
Great Dyke, Rhodesia	3,265
Sudbury, Canada	1,342
Stillwater (exposed only)	194
Skaergaard, Greenland	104

*includes area inferred from geophysical evidence

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FINE SCALE RHYTHMIC LAYERING IN THE STILLWATER COMPLEX, MT; A. E. Boudreau, Dept. of Geology, Univ. of Oregon, Eugene, OR 97403

Fine scale rhythmic layering, or inch scale layering, is characterized by the laminar segregation of mineral phases in igneous rocks on a millimeter to centimeter scale. The extreme regularity characteristic of many examples of fine scale layering has led several workers (1,2,3 and others) to propose a periodic nucleation mechanism for their origin. The theory of periodic nucleation is most fully developed by (3). Because nucleation requires a finite amount of supersaturation in the liquid before it can occur, it is proposed that the nucleation of new crystals must occur at a distance away from previously nucleated crystals.

The theory of periodic precipitation, however, does not explain such phenomenon as the inch scale doublet layering seen in the Banded zone of the Stillwater complex. This unusual layering consists of paired orthopyroxene layers which occur as regularly spaced doublets in an anorthosite host. Another feature not predicted by the theory of periodic precipitation is the observed correlation of grain size with layer spacing, as shown in fig. 1. In this figure, the spacing of layers in rocks of overall norite or gabbro-norite composition from the Banded zone of the Stillwater complex is plotted against the grain size of the orthopyroxene crystals which make up the mafic layers. The graph suggests three correlations: 1) In general, for rocks of a given modal abundance of pyroxene, layers in coarse grained rocks are further apart than are layers in fine grained rocks. 2) In general, for rocks of a given grain size, the spacing of mafic layers decreases with increasing modal abundance of pyroxene. 3) Finally, the overall trend of the graph extrapolates back to the origin. This last observation suggests that there was no pattern (i.e. layering) present at the time the crystals originally underwent nucleation.

In contrast to the classical Liesegang theory of periodic nucleation to explain layer formation are the analysis of rhythmic behavior and pattern formation in biological systems (4), banding in metamorphic rocks (5), and, most relevantly, the analysis of pattern formation in solutions which began as uniform suspensions of homogeneously nucleated crystallites (6,7). The theory presented in all of these papers involves the initiation of minor perturbations (in grain size or chemical concentrations) in an initially uniform system, which then continues to grow and propagate itself in a harmonic manner through the system. In the experiments of (6) and (7), patterns were formed in initially uniform suspensions of PbI_2 crystallites in gel solutions. In the analysis of (7), the authors explained the pattern formation as a consequence of aging: Because bigger crystals require less solution concentrations than do smaller crystals, any minor perturbations in grain size in the initial crystal assemblage will cause some crystals to grow at the expense of surrounding crystals. However, dissolution of the surrounding smaller crystals causes the solution concentrations to increase above the initial ambient value as they attempt to buffer the solution to their ever-diminishing crystal size. These higher solution concentrations lead to diffusion of material both toward the site of the original size perturbation and also in the other direction, to crystals yet further away. The influx of solution components then experienced by crystals at a distance causes them to grow and begin the cycle anew. Thus local maxima in grain size or solution concentrations can cause nearby minima and vice versa. The system is un-

Boudreau, A. E.

stable and any minor fluctuation from the uniform state will not only continue to grow but propagate in a rhythmic manner. For non-uniform systems, as would be expected in a growing front of crystallization in a magma chamber, any minor gradients present in the physical/chemical conditions will initiate and control the symmetry of the developing pattern. In slowly cooled igneous intrusions, the pattern develops as layering perpendicular to the direction of crystallization.

In a much abbreviated analysis, the mathematical evaluation of this process begins with a mass balance equation:

Time rate of change of concentration of a liquid species in a given volume. = Net rate of flux of species into or out of volume. + Net rate of precipitation or dissolution of crystals within the volume element.

or,

$$dC/dt = -\nabla J - Q \quad (1)$$

where

C = liquid concentration, assumed buffered by crystal size.

J = mass flux of component.

Q = net rate of crystallization/dissolution.

The mass flux term, J , is given by Fick's law as $-DdC/dx$ (assuming mass transport in x -direction only). Assuming n spherical grains initially of radius R in a unit volume, the crystallization/dissolution rate, Q , can be expressed as

$$Q = \rho nR^2(dR/dt) \quad (2)$$

where

ρ = mass balance (density) constant.

By the assumption of local equilibrium, C is a function of crystal radius. Furthermore, the crystal radius can be expressed as a minor perturbation about an average grain size:

$$R = R_{avg} + r \quad (3)$$

Using a number of simplifications (i.e. linear dependence of C with crystal radius, mass transport in x -direction only, and R_{avg} constant or changing linearly with time and distance), then eq. (1) reduces to an equation describing the time/space evolution of the grain size perturbation (ignoring constant term):

$$dr/dt = g(d^2r/dx^2) \quad (4)$$

where

$$g = D(dC/dR)/(dC/dR + \rho nR^2)$$

Note that the term g consists of experimentally attainable parameters. Eq. (4) has solutions of the form

$$\begin{aligned} r(t,x) &= \exp(zt + ikx) \\ &= \exp(zt)(\sin kx + \delta) \end{aligned} \quad (5)$$

where z and k are proportional to the variable g :

$$z = -gk^2$$

It is thus apparent that the solution to eq. (4) has a harmonic form. More complete analysis (as done for salt solutions by (7)) show the initial pattern length to be dependent on the initial crystal number density: the initial pattern wavelength is short in systems with an initially high par-

Boudreau, A. E.

title density. Furthermore, because coarser crystals always continue to grow at the expense of less developed secondary maxima, the pattern length will grow larger as the crystals continue to coarsen with time. Both of these theoretical predictions are consistent with the observed relations shown in figure 1. Within any developing layer, crystal growth competition can continue and lead to the formation of 'doublets'. Once non-favored crystals fall below their critical radius they will dissolve completely, resulting in a series of monomineralic layers. Also, crystal competition within a given layer should result in a polygonal pattern developed within the plane of layering. Such polygonal patterns have been observed in three instances in fine scale layering at the Stillwater complex. Finally, it should be noted that layering need not be regular - layers developing independently at several different horizons may interfere to give a more complex spacing of layers.

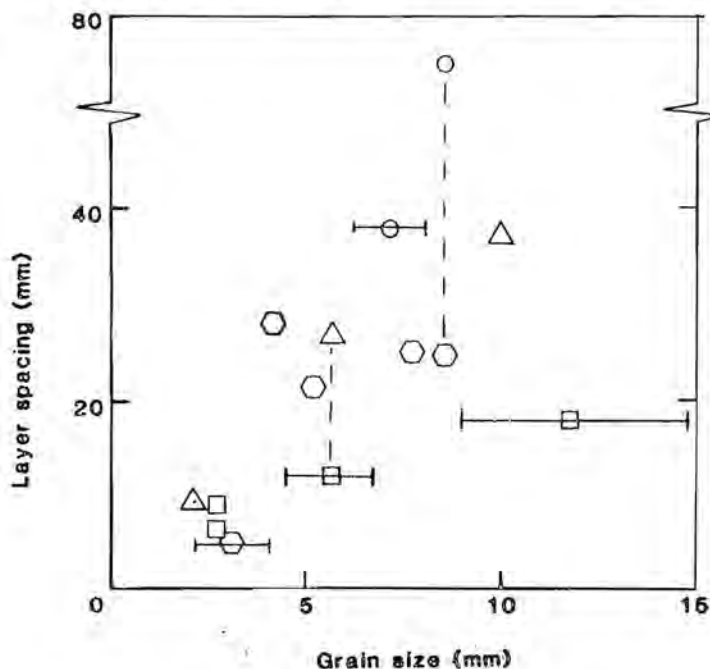


Fig. 1. Layer spacing vs. grain size in layered norites and gabbronorites, Stillwater complex. Overall pyroxene modal abundance is denoted as follows: \circ = 5-15% pyx, \bigcirc = 15-25% pyx, \triangle = 25-35% pyx, \square = >35% pyx. Dashed lines denote layer spacing within and between those layers showing 'doublet' development.

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THE MAIN PLATINUM ZONE, STILLWATER COMPLEX, MT - EVIDENCE FOR BIMETASOMATISM AND A SECONDARY ORIGIN FOR OLIVINE; A. E. Boudreau, Anaconda Copper Co., Nye, MT 59061.

Economic quantities of platinum group elements occur within the first olivine-bearing subzone of the Banded zone of the Stillwater complex (OBZI, stratigraphy after (1)). Most existing theories on the return of olivine in the stratigraphic sequence of the Stillwater complex and the origin of the accompanying platinoid/sulfide mineralization have required the introduction of a new pulse of primitive magma into the partially fractionated magma chamber (2,3, and 4). These interpretations have largely ignored the details of the petrologic associations which strongly point to the role of volatile activity both as a carrier for the mineralization and as a means by which olivine can be formed by the recrystallization of an originally gabbroic or noritic crystal assemblage. Evidence for a secondary origin for some olivine occurrences in the Banded zone has been presented earlier by (5). This paper will review evidence which suggests that the olivine occurrence in OBZI is the result of bimetasomatic reaction caused by localization of a late evolving volatile fluid within the crystal pile.

The stratigraphic sequence in the vicinity of OBZI consists, from bottom to top, of norites and gabbronorites of Gabbronorite subzone I, followed by the olivine-bearing rocks (with variable modal proportions of olivine, pyroxene and plagioclase), anorthosites and norites which make up OBZI, and finally the predominantly noritic rocks of Norite subzone II. The Mg/Mg + Fe ratio (mg) of cumulus pyroxenes throughout the lower Banded zone show a normal fractionation trend which continues without apparent interruption through OBZI (2,6). OBZI is of variable thickness and locally olivine is absent from the stratigraphic sequence. Mapping during underground development at the Anaconda Copper Co. Minneapolis adit shows OBZI to thicken eastward at the expense of hanging wall lithologies. Furthermore, rock types will often change abruptly along strike, with little evidence that the olivine was filling basinal structures in the footwall.

The olivine-bearing rocks are characterized by coarse to pegmatoidal textures and may contain significant biotite. Associated with the more olivine-rich rocks are anorthosites poor or lacking in mafic phases which develop asymmetrically on both sides of the olivine layers. These anorthosites thicken and thin as their associated olivine-rich layer thickens and thins. Olivine may also occur in coarse to pegmatoidal podiform concentrations, about which are often developed anorthositic halos or coronas. These texturally and mineralogically heterogeneous layers are analogous to the pillow troctolites of (7). It is within the olivine-bearing rocks or their associated anorthosites that platinoid/sulfide mineralization is most commonly found.

The olivine/anorthosite association and other anomalous features seen in OBZI are difficult to explain by any reasonable magmatic or crystal settling processes. Instead it is proposed that the mafic-depleted anorthosites which occur about the coarse to pegmatoidal olivine-bearing layers are the result of a bimetasomatic reaction in which olivine is being formed at the expense of pyroxene in the surrounding crystal assemblage. Specifically, it is proposed that a volatile fluid, evolved during the final solidification of the intercumulate liquid, can become confined to a horizon within the

Boudreau, A. E.

crystal pile. (A similar model for the late evolution and migration of a volatile fluid has been proposed by (8) to explain geochemical variation in Icelandic basalt flows.) The experimental work of (9) and (10) show that water and certain other volatile components can increase the olivine phase field relative to orthopyroxene and plagioclase and, further, that silica is soluble in hydrous fluids at magmatic temperatures. It is thus possible for a localized volatile accumulation to initiate a bimetasomatic reaction process - pyroxene, being unstable in the presence of a migrating volatile fluid, would be resorbed along sharp reaction fronts and its mafic components transported down chemical potential gradients to the high volatile regions where olivine is forming. Silica evolved in the destruction of pyroxene would be lost with the eventual escape of the volatile phase.

The process described above is bimetasomatic in the sense that material is being transported in two directions - the volatile component diffuses outward while the mafic components diffuse inward to form the olivine layer. If enough components are mobil, the result of such a bimetasomatic alteration process can be a series of monomineralic layers with sharp contacts. This is often the observed case, the olivine layers terminate sharply against anorthosites which in turn have sharp contacts with norites or other surrounding rocks. Furthermore, in more typical hydrothermal bimetasomatic systems, such as the copper porphyry system at Butte, Mt, the alteration zones associated with two-way mass transport may reach up to 10 m in thickness. This is about the same dimension as the largest anorthosite seen in OBZI.

Several other lines of evidence support a secondary origin for the olivine of OBZI: (a) The data of (2) shows the mg ratio of olivine in the olivine-bearing rocks to be similar to that of the cumulus pyroxenes in the nearby norites and gabbro-norites. This would be expected if olivine was formed at the expense of pyroxene by simple loss of silica. (b) Hydrous phases, especially biotite, are most commonly found associated with the more olivine-rich rock types. Further, the occasional chromite grains found in the olivine-bearing rocks invariably contain spherical inclusions which contain biotite. The presence of hydrous phases suggests that highest volatile activity was in the olivine-bearing rocks and that volatiles can stabilize chromite as well as olivine (opx in surrounding norites contains an average 0.4 wt. % Cr_2O_3). (c) No feeder system of dikes is seen cutting footwall rocks, as might be expected if OBZI were the result of a fresh magma injection. However, within OBZI, olivine-bearing rocks are found as apophyses injected into or crosscutting surrounding rocks. This would be possible if the forming olivine layers were plastic or perhaps even partially molten. (d) The presence of discordant pegmatitic zones which may contain massive sulfide concentrations are evidence that late volatile fluids can transport platinoid-bearing sulfides. (e) Finally, the data of (2) shows little convincing geochemical evidence to support the notion of a new magma pulse.

The most likely source for the fluids required to drive the bimetasomatic alteration process is the Ultramafic zone of the Stillwater complex, as it is anomalously high in platinoids (11). It is envisioned that fluids exsolved during the final solidification of the Ultramafic zone would migrate upwards until it became trapped at a horizon where sufficient intercumulus melt was present to absorb the fluid influx. Further migration would be by diffusive transport until volatile saturation was again reached by continued

Boudreau, A. E.

fluid influx and falling temperatures. Dilution of the fluid phase with intercumulus melt may also be the cause of sulfide precipitation - mixing of the platinoid/sulfide-bearing fluid with magma would cause disruption of the complexing agent responsible for carrying sulfide in the fluid phase. Once the volatiles are confined to a horizon within the crystal pile the alteration process can proceed as previously described. A cartoon of the development of an olivine layer is shown in fig. 1.

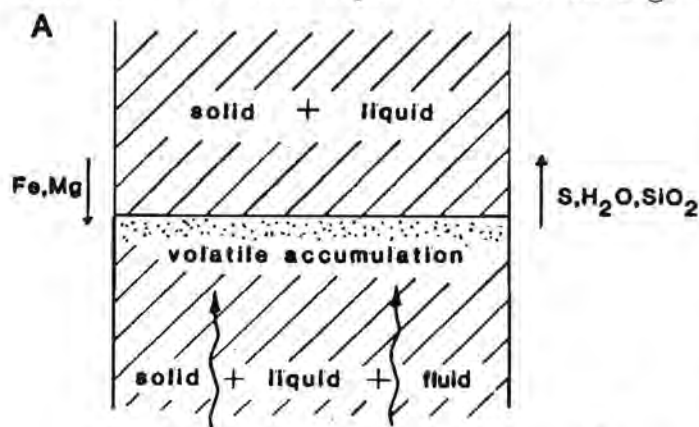
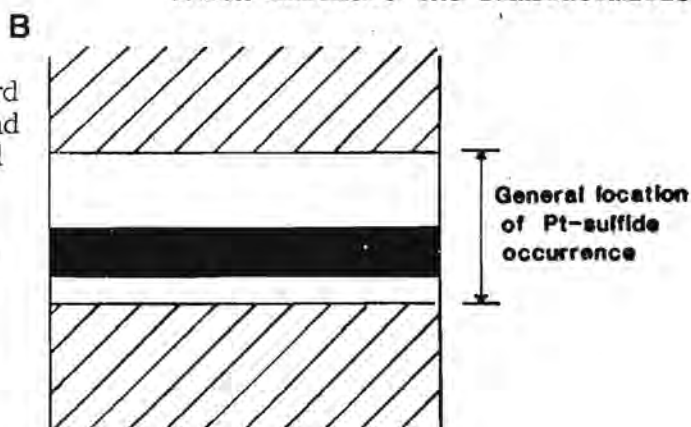


Figure 1. Simplified cartoon showing the formation of olivine/anorthosite layers within an initially gabbroic crystal assemblage. (A) Migrating volatile fluids, exsolved during final solidification of the crystal pile, becomes confined to a horizon within the pile. This leads to locally steep chemical potential gradients which initiate the bimetasomatic

reaction: Fe and Mg diffuse toward the zone of volatile enrichment and water, sulfides and silica evolved by the destruction of pyroxene move outward. (B) The eventual result is an anorthositic selvage developed about the secondary olivine layer. Platinoid/sulfide mineralization is found anywhere within the reaction zone.



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LAYERED INTRUSIONS: A MINI REVIEW; I.H. Campbell, Earth and Plan. Sciences, Room 3032, Erindale College, University of Toronto, Mississauga, Ontario L5L 1C6

Introduction: It is recognised by geochemists that much of the chemical variation seen in volcanic rocks is due to fractional crystallisation. Layered intrusions offer the opportunity to study the mechanism of fractional crystallisation directly; yet, despite their obvious importance, have received comparatively little attention in recent years. The purpose of this review is to summarise some of the main features of layered intrusions, paying special attention to the mechanism of fractional crystallisation.

Structural setting: The structural settings of layered intrusions are, for the most part, poorly defined. Many have formed in zones of crustal extension, perhaps the most favourable settings being failed and successful rift systems. The Duluth Complex and Noril'sk gabbro-dolerites are believed to be examples of magma chambers which have formed in a failed rift environment(1). Many layered intrusions near continental margins, such as Skaergaard and Rhum, probably formed during the development of a successful rift.

Age: Layered intrusions range in age from Tertiary to late Archean. The lower age limit is constrained by the time it takes tectonic and erosional processes to bring a magma chamber to the surface. The oldest intrusions are of special importance because they created the first opportunity for high level fractional crystallisation. These early magma chambers, which include the Bell River Complex in Quebec, the Kam-Kotia Sill in Ontario and the Stillwater Complex, date from about 2.75 AE. Significantly, magmas older than 2.80 AE rarely show Eu anomalies whereas they are common in younger flows. The absence of layered intrusions prior to this time suggests that the crust was not rigid and/or thick enough to support large magma chambers.

Shape: Layered intrusions were originally classified as lopoliths(2), which were defined as large centrally sunken sills with a semi-horizontal roof and with internal layering parallel to the floor. However, detailed observations of the margins of intrusions by Wager and others showed that they dip inwards at a greater angle than the internal layering of the rocks, indicating that the form of the intrusion is funnel-shaped rather than lopolithic. The angles at the bases of layered intrusions are variable and influence the shape of the layering. Wide angle intrusions, such as the Muskox and Stillwater Intrusions, have low angle layering whereas the layering in acute angled intrusions, such as Jimberlana and the Great Dyke is "U" shaped and dips steeply at the margin(3).

Composition of Magmas: The magmas which feed layered intrusions show a wide range in composition which can include komatiites, tholeiites, high alumina basalts, calc-alkaline basalts, alkali basalts and anorthosites, depending on how one chooses to define the term layered intrusion. If we confine our discussion to basic intrusions then most of the important ones (e.g. Bushveld, Stillwater, Kiglapait and Skaergaard) are tholeiitic. They fractionate with a strong Fe enrichment trend which may be accompanied by weak Silica depletion. The MgO content of the parent magmas is generally placed within the range 10-14%, somewhat below the 18% suggested by O'Hara(4) for MORB.

The crystallisation sequence varies with the composition of the melt but a typical sequence is olivine + Cr-spinel followed by bronzite then plagioclase + augite + hypersthene. If the intrusion follows an Fe enrichment - silica depletion trend, orthopyroxene is replaced by Fe-rich olivine during the final stages of fractional crystallisation (e.g. Skaergaard). If the

I.H. Campbell

magma follows a silica enrichment trend olivine does not reappear (e.g. Jimberlana).

Types of layering: Geologists who have not worked on layered intrusions probably have a false impression of the frequency with which rhythmic layering is found in intrusions. Rhythmic layering is actually a minor component in many layered intrusions, though its genetic importance cannot be overemphasised.

There are numerous types of layering but only the three most important will be discussed here.

(i) Phase layering: major stratigraphic layering resulting from the appearance or disappearance of a cumulus phase. Layered intrusions are mapped and subdivided on the basis of this type of layering.

(ii) Cyclic layering: a macrorhythmic unit in which the cumulus mineral sequence can be identified as the fractional crystallisation order of the magma(5). These cyclic units are 30-300 m thick; a typical unit having olivine cumulates at its base which pass abruptly or gradationally into bronzite or plagioclase cumulates at the top. This sequence may be repeated many times, each cycle representing the entry of a new pulse of magma into the chamber (6,7,8,9). If each macro cycle represents a new pulse of magma, the mantle beneath the intrusion must undergo multiple melting events. Each time the mantle melts it will become more refractory making further melting more difficult. A given volume of mantle may melt twice but a third melting event is unlikely. Multiple melting of the mantle is facilitated if previously melted material is removed by mantle convection. This is most likely where layered intrusions are located above a zone of upward mantle convection, lending support to the failed rift-layered intrusion connection.

(iii) Rhythmic layering: fine scale layering is due, in most cases, to variations in the modal proportions of the cumulus minerals. Contacts between layers may be sharp or gradational. Graded layers usually have the most dense material at the bottom and the lightest at the top, but true hydrological sorting, taking both size and density into account, is rare. In fact many graded layers are hydrologically upside-down (10,11,3). The layering at Duke Island (12), which shows excellent size-density sorting, is an important exception.

Cumulate textures: A cumulate rock is made up of two parts, a liquidus or cumulus component and the trapped intercumulus liquid. This is the most important property of cumulates for it is this property which makes fractional crystallisation possible. If the fractionation process is efficient, most of the trapped liquid is expelled from the crystal pile at the liquidus temperature and the rock is said to have an adcumulate texture. If it is inefficient, much of the liquid is trapped between the cumulus grains and an orthocumulate texture results. In adcumulate sequences the Rayleigh equation, $C_L/C_I = F(D'-1)$, can be used to model the effects of fractionation but in orthocumulate sequences this equation should be replaced by $C_L/C_I = F(D'-1)(1-P)$, where P is the fraction of trapped intercumulus liquid.

In recent years a number of authors (13,14,15,16) have questioned the widely held belief that cumulate textures result from crystal settling in the magma chamber. There are numerous objections to the cumulate theory only two of which will be discussed here. The first and most important objection concerns the buoyancy of plagioclase in tholeiitic liquids. Both calculations (17,18) and experiments (19) suggest that plagioclase should float in the Fe-rich liquids of the Skaergaard, Bushveld and Kiglapait intrusions. The second objection concerns the mechanism of nucleation (14). The cumulate theory assumes that crystals nucleate homogeneously and settle as individual grains. The activation energy required for this type of nucleation is high

I.H. Campbell

but it is greatly reduced if crystals nucleate against a pre-existing crystal (heterogeneous nucleation), especially if the crystal is of the same type (self nucleation). In natural processes heterogeneous nucleation is the dominant nucleation mechanism. Thus nucleation theory predicts that new crystals should nucleate in situ against pre-existing crystals at the floor and sides of the intrusion rather than form as individual crystals in the centre of the intrusion which subsequently gravitate to the floor of the chamber.

The distribution of cumulate textures in layered intrusions: Any explanation of cumulates should account for the distribution of adcumulate and orthocumulate textures within layered intrusions. Detailed studies of the distribution of adcumulate and orthocumulate textures have been made for the Skaergaard (20), Kiglapait (21) and Jimberlana (13) intrusions. From these studies the following generalisations can be made:

(i) The textures at the base of the Kiglapait and Skaergaard intrusions are orthocumulate but gradually become more adcumulate as the top of the intrusions are approached. In Jimberlana the reverse is true; the percentage of trapped liquid in cumulates increases systematically from the bottom upwards. The viscosity of the Kiglapait and Skaergaard melts, which fractionate with an Fe enrichment trend, are believed to decrease with increased fractionation (22). The Jimberlana magma, on the other hand, follows a silica enrichment trend and the viscosity of the melt increases with increased fractionation. These observations suggest that adcumulate textures are favoured by melts of low viscosity, whereas high viscosity melts favour the formation of orthocumulate textures.

(ii) As a given layer is traced towards the margin of an intrusion the residual porosity increases and the texture of the rock becomes more orthocumulate (13, 23). In this case a correlation can be made between textural type and the direction of heat loss at the crystal-liquid interface. At the margin of the intrusion heat loss is downwards through the crystal pile, whereas in the centre it is upwards into the overlying melt.

(iii) The appearance of a new cumulus phase results in the bottom of the new layer being more adcumulate than the top of the underlying layer. This is believed to be due to the melt having to become supersaturated with the new phase before it can begin to crystallise (24,21).

(iv) In the Jimberlana intrusion there is a close correlation between rhythmic layering and zones of adcumulate texture. This correlation can be extended in a general way to other layered intrusions. The gabbros of Kiglapait, Skaergaard and the Bushveld, which are predominately adcumulate, are well layered whereas those of Jimberlana and Sudbury, which are predominately orthocumulate, are not.

There are two hypotheses for the formation of adcumulate textures. The first is that the intercumulus liquid is squeezed out by filter pressing (16). The second is that the development of the different textural types is controlled by the balance between nucleation and crystal growth at the crystal-liquid interface (14). The observations cited above support the second hypothesis. Low melt viscosity assists the transport of material to and from the growing crystals, favouring the continued growth of existing crystals and hence the development of adcumulate textures. Heat loss through the crystal pile has the reverse effect. It increases the amount of supercooling at the crystal-liquid interface, resulting in the nucleation of new grains and the formation of orthocumulate textures.

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SOME ISOTOPIC, COMPOSITIONAL AND THERMAL ASPECTS OF THE EVOLUTION OF MAGMA CHAMBERS IN THE CRUST, D.J. DePaolo, Department of Earth and Space Sciences, University of California, Los Angeles, California 90024.

Introduction: Understanding of the evolution of magmas in the crust centers on questions related to the location of the magma source and its properties, the composition--both isotopic and chemical--of the parent magma, the crystallization of the magma (especially the nature and extent of the isolation of crystals as they form), the supply of fresh parental magma during crystallization, and the exchange of heat and mass between the magma and the country rocks. Information on all of these can be obtained through radiogenic isotope studies in conjunction with petrologic and geophysical studies. In order to extract the full amount of petrologic information possible, it is necessary to construct appropriate models which describe material and heat transfer in detail comparable to that obtainable by petrologic studies of mineralogy, texture and mineral composition. It is the purpose of this paper to briefly describe some models that may be applicable and to give some examples of the effects expected. The large layered mafic intrusions of the earth, like the Skaergaard, Stillwater, Kiglapait, Muskox, and Bushveld are excellent testing grounds for such models because they are relatively simple bodies and have been thoroughly studied.

Assimilation-Fractional Crystallization (AFC). The chemical composition of a magma body can obviously be modified by assimilation of country rock, which can effect the differentiation path or at a minimum, modify the proportions of crystallizing minerals. Furthermore, assimilation can modify the initial $^{143}\text{Nd}/^{144}\text{Nd}$, $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$ ratios. This will affect inferences about the nature of the magma source based on these ratios. In general, the assimilation of country rock, which requires heat, will be aided by concurrent crystallization of the magma, which releases latent heat. For the layered intrusions this is a moot point because crystallization has obviously been the most important aspect of the evolution of the magma chamber. In any event, the residual magma will see the affects of both assimilation and crystallization at the same time.

If the mass rate of assimilation \dot{M}_a is a fraction (a) of the rate of crystallization (\dot{M}_c) the concentration of an element in the residual magma (C_m) is given by [1]:

$$C_m/C_m^0 = F^{-z} + \frac{a}{a-1} \frac{C_a}{zC_m^0} (1-F^{-z}) \quad (1)$$

where:

$$z = \frac{a + D - 1}{a - 1}$$

D = bulk solid/liquid distribution coefficient for the crystallizing phases

C_a = concentration of the element in the contaminant

F = mass of magma remaining; as a fraction of the mass of the original magma

For $a = 1$:

DePaolo, D.J.

$$C_m/C_m^0 = \exp(-DM_a/M_m) + \frac{C_a}{DC_m^0} [1 - \exp(-DM_a/M_m)] \quad (2)$$

where $M_a = \int_0^t \dot{M}_a d\tau$.

For an isotopic ratio (ϵ), the shift due to contamination is:

$$f = \frac{\epsilon_m - \epsilon_m^0}{\epsilon_a - \epsilon_m^0} = 1 - (C_m^0/C_m) F^{-Z} \quad (3)$$

where ϵ_a is the isotopic ratio of the contaminant. For $a = 1$:

$$f = \frac{\epsilon_m - \epsilon_m^0}{\epsilon_a - \epsilon_m^0} = 1 - (C_m^0/C_m) \exp(-DM_a/M_m) \quad (4)$$

The value of D depends on the phases crystallizing and their proportions and may vary continuously or discontinuously during the course of crystallization; the parameter "a" may also vary. Note that for any element in the magma, regardless of whether it is radiogenic, f is the fraction of the element that was derived from the assimilated material.

Figure 1 shows the variation of selected trace and minor elements in a basaltic magma undergoing AFC for $a = 1$ (dry, ~ 8 kb) and $a = 0.2$ (water saturated at ~ 5 kb), for the liquidus phase proportions given. The contaminant was assumed to have the trace element composition of average upper crust [2] and the major element composition of andesite. The starting magma was assigned the major and trace element composition of average mid-ocean ridge tholeiitic basalt. The patterns calculated are not sensitive to differences in liquidus phase proportions if the magma remains dry. When hornblende becomes a liquidus phase, however, the patterns are highly sensitive to its proportion and somewhat uncertain due to variability of the distribution coefficients for rare-earths. Figure 2 shows the changes of $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$, with arbitrary values assigned the endmembers. Relative to simple fractional crystallization, the primary difference is that the magma becomes more enriched in those elements that are enriched in the contaminant. An important point, however, is that the patterns are undistinguished. The same patterns could be generated, using a batch partial melting model and appropriate source characteristics, or a partial melting plus fractional crystallization model. Consequently, the fact that trace element patterns can be modelled by these latter processes in no way rules out assimilation.

Another parameter that can be useful in detecting contamination is $\delta^{18}\text{O}$. Assuming $C_m = C_a$ and $D = 1$ for oxygen yields:

$$\delta_m - \delta_m^0 = (\delta_a - \delta_m^0 - \Delta/a)(1 - F^{-Z}) \quad (5)$$

and for $a = 1$:

$$\delta_m - \delta_m^0 = (\delta_a - \delta_m^0 - \Delta)[1 - \exp(-M_a/M_m)], \quad (6)$$

DePaolo, D.J.

where $\Delta = \delta_{\text{crystals}} - \delta_{\text{magma}}$. The shift of $\delta^{18}\text{O}$ is sensitive to the ratio Δ/a . As an example, for layered intrusions "a" may in general be small (0.1 or less; see below). Consequently, if Δ is +0.3, Δ/a is 6 for $a = 0.05$. In this case, even though $\delta_a - \delta_m^0$ may be positive, δ_m will decrease during crystallization-assimilation, i.e. change in the direction away from $\delta^{18}\text{O}$ of the contaminant.

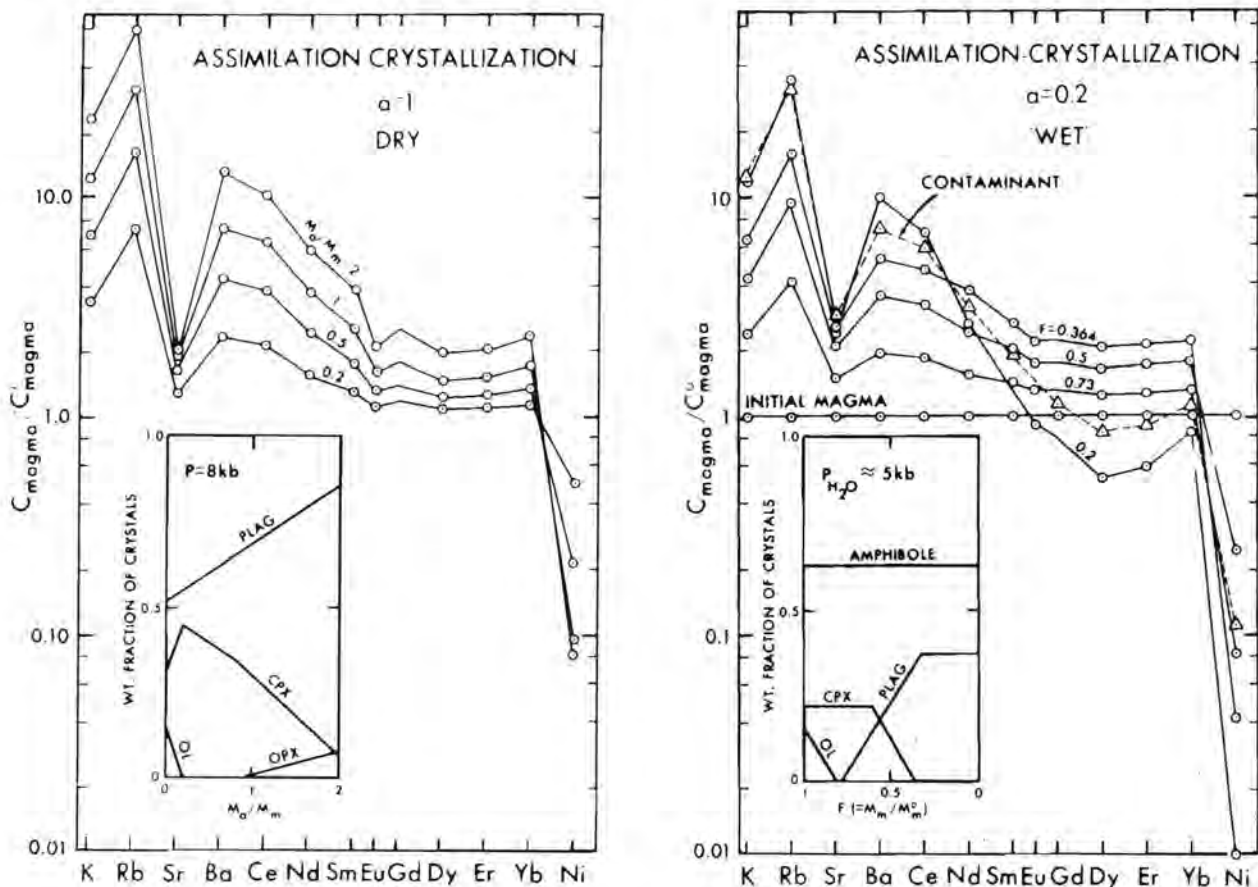


Figure 1

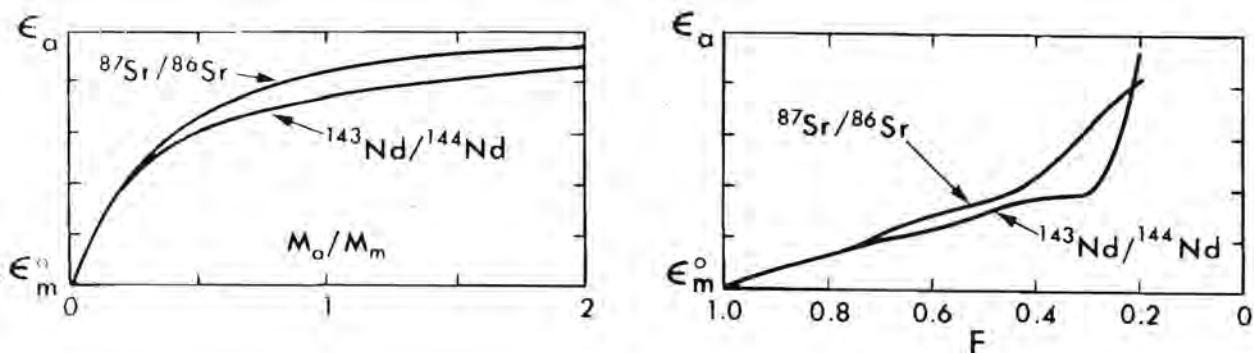


Figure 2

DePaolo, D.J.

Magma recharge, assimilation and crystallization. Many layered intrusions show evidence that the magma chamber was periodically (or continuously) recharged with fresh magma during crystallization. If recharge is episodic, the problem can be treated sufficiently well with the equations given above; each pulse of new magma will simply mix with the evolved magma causing a discontinuous shift in properties. This would produce a distinct type of variation in the initial $^{87}\text{Sr}/^{86}\text{Sr}$, etc. as a function of stratigraphic level. For continuous recharge at a rate equal to a fraction (r) of the crystallization rate, the concentration of an element is given by:

$$C_m/C_m^0 = F^{-z_r} + \frac{a(C_a/C_m^0) + r(C_r/C_m^0)}{z_r(a+r-1)}(1 - F^{-z_r}) \quad (7)$$

where $z_r = \frac{a+r+D-1}{a+r-1}$ and C_r is the concentration in the recharging magma. The differential form for the isotopic ratios is:

$$\frac{d\epsilon_m}{d \ln F} = (a+r-1)^{-1} \left[a \frac{C_a}{C_m} (\epsilon_a - \epsilon_m) + r \frac{C_r}{C_m} (\epsilon_r - \epsilon_m) \right] \quad (8)$$

where ϵ_r is the isotopic ratio of the recharging magma. This process presents the possibility that the isotopic ratio of the magma could be held during crystallization at a value that is constant and displaced from that of the magma source, by balancing the assimilation and recharge rates such that,

$$a/r = - (C_r/C_a) \frac{\epsilon_r - \epsilon_m}{\epsilon_a - \epsilon_m} \quad (9)$$

Limitations on assimilation; compositional effects. A limiting case for which the crystallization rate can be related to assimilation rate is for a magma of eutectic composition and temperature T_m surrounded by perfectly insulating walls. If an incremental mass of solid contaminant (dm_a) at temperature $T_a < T_m$ is added to the magma it will cause crystallization of an incremental mass of crystals (dm_c), with the two related by:

$$\left(\frac{dm_a}{dm_c} \right)^e = a_{\max}^e = \frac{L_m}{c_a(T_m - T_a)} \quad (10)$$

where L_m is the latent heat of crystallization of the magma and c_a is the heat capacity of the solid contaminant. For a eutectic magma losing heat by conduction to cooler wallrocks, the value of "a" must be smaller than a_{\max} and can be expressed in terms of the total heat loss of the magma during crystallization (dq_m) and the heat used for assimilation (dq_a) by:

$$a = a_{\max} (dq_a/dq_m) \quad (11)$$

The parameter "a" can therefore be formally treated as dependent upon a thermodynamic term (a_{\max}), which can be estimated in many circumstances, and a complex rate term which depends upon the detailed physics of the magma-wallrock interface and may be difficult to predict. However, information on this term can be derived if "a" can be estimated from isotopic effects.

For the general case of bulk assimilation by a non-eutectic magma the following relationship is obtained:

DePaolo, D.J.

$$a_{\max} = \left(\frac{dm_a}{dm_c} \right)_{\max} = \frac{L_m + c_m (\vec{X}_c - \vec{X}_m) \cdot \text{grad } T_m(\vec{X}_m)}{L_a + c_a (T_m - T_a) + c_m (\vec{X}_a - \vec{X}_m) \cdot \text{grad } T_m(\vec{X}_m)} \quad (12)$$

where L_a is the latent heat of melting of the assimilated material, c_a is the heat capacity (assumed identical whether solid or liquid) and T_m is the liquidus temperature, a function of magma composition. The vectors \vec{X}_m , \vec{X}_a and \vec{X}_c give the compositions of the magma, assimilated material and liquidus crystals respectively in terms of any convenient units (e.g. oxide weight fractions), and $\text{grad } T_m(\vec{X}_m)$ is the gradient of the liquidus temperature. In general, the dot product in the numerator is positive in value while that in the denominator is negative. Consequently, a_{\max} can be greater than for the eutectic case. Of particular interest is the situation where dry magma assimilates wet country rock. Because dT_m/dX_{H_2O} can be of the order $-10,000^\circ$, a_{\max} may become very large, so even if dq_a/dq_m is small, a may be large.

Example: Kiglapait Layered Intrusion. A possible example of crystallization accompanied by assimilation is provided by the Kiglapait layered intrusion in Labrador [3]. Isotopic studies of Sr show that during the crystallization of 90% of the volume of the intrusion the initial $^{87}\text{Sr}/^{86}\text{Sr}$ is constant at about 0.7040 [4]. During the last 10% of crystallization, initial $^{87}\text{Sr}/^{86}\text{Sr}$ increases monotonically to 0.7066. Neodymium isotopes vary in similar manner from $\epsilon_{\text{Nd}} = -1$ to $\epsilon_{\text{Nd}} = -5$ [5]. Using these data, it can be calculated that the value of the parameter "a" was zero during the first 90% of crystallization, then jumped to about 0.1 during the next 5% of crystallization and dropped off to ~ 0.02 for the final 5%. These values are estimates; they depend on the properties of the contaminant, which are unknown. The cause of the initiation of assimilation at 90% solidified is unknown. One possibility is that the solidified roof (border) zone of the intrusion foundered or cracked at that time allowing roof rocks (probably partly molten) access to the magma chamber. Another possibility is that assimilation did occur at a small rate during the first 90% crystallization, but was overwhelmed by the semi-continuous influx of fresh magma (i.e. the ratio a/r was small). The 90% solidified level could correspond to the cessation of recharge ($r \rightarrow 0$), allowing the assimilation effects to be manifested in the isotopic ratios thereafter. $\delta^{18}\text{O}$ in the Kiglapait appears to decrease slightly during crystallization [6] from +6.4 to +5.9. From eq. (5), however, it can be seen that for the small values of "a" associated with this intrusion, the $\delta^{18}\text{O}$ shifts are dominated by the Δ value for the crystallizing phases.

The constraints on the assimilation rates provided by the isotopic data suggest that the ratio $dq_a/dq_m \approx 0.02$, indicating that only a small part of the heat released during crystallization was used for assimilation. This may be a general feature of magmas emplaced at relatively shallow crustal levels (10 km or less). Other evidence suggests that magmas having a substantial residence time at deep crustal levels assimilate much more material, especially in orogenic zones where the country rocks may be preheated [7].

The Stillwater intrusion exhibits no detectable variation of initial ϵ_{Nd} [8]. Its crystallization-assimilation history may not, however, have been different from that of the Kiglapait, because 1) the top 10% of the Stillwater is not exposed, and 2) the country rocks of the Stillwater are

DePaolo, D.J.

not very old relative to the intrusion; their ϵ_{Nd} values did not differ from that of the intrusion by an amount as large as that for the Kiglapait.

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PETROLOGICAL AND GEOCHEMICAL CHARACTERIZATION OF THE ST. URBAIN ANORTHOSITE MASSIF, QUEBEC: SUMMARY OF INITIAL RESULTS: L. PETER GROMET, Dept. Geol. Sci., Brown Univ., Providence, RI 02912 and ROBERT F. DYMEK, Dept. Geol. Sci., Harvard Univ., Cambridge, MA 02138

INTRODUCTION - The St. Urbain anorthosite is a relatively small massif (Fig. 1) located in the central high-grade granulite terrain of the Grenville Province, 120 km NE of Quebec City (1,2,3). It is comprised almost exclusively of anorthositic rocks, dominantly andesine anorthosite of several textural types, but including some labradorite anorthosite, very minor leuconorite, and a few small Fe-Ti oxide deposits. The massif lacks penetrative deformation, corona textures and other evidence of significant post-magmatic modification. Similarly, the preservation of delicate igneous textures and compositional zoning in various minerals indicates that magmatic processes can be directly studied at St. Urbain. Moreover, this massif displays numerous features and critical relationships which are relevant to the origin and evolution of a classic massif anorthosite, including locally abundant dikes of andesine anorthosite cutting labradorite anorthosite, plagioclase lamellae-bearing orthopyroxene megacrysts in certain lithologies and local compositional layering in anorthosite/leuconorite. Some preliminary results of our integrated field, petrological and geochemical studies of the St. Urbain massif are summarized below.

ANORTHOSITE DIKES - One of the most intriguing aspects of this massif is the juxtaposition, with sharp (3 mm) contacts, of texturally and chemically distinct anorthosite types. Such relationships are best displayed on Mont due Lac des Cygnes (MLC), a locality (see Fig. 1) which we have studied in detail (4,5,6). On MLC, andesine anorthosite dikes (~1 cm to > 10 m wide; ~98% antiperthitic plag; An_{35-45}) crosscut labradorite anorthosite wall rocks (85-95% plag; An_{58-65}). The presence of apophyses off these dikes, rotated xenoliths in the dikes and gradations from dikes to anastomosing veinlets clearly point to the emplacement of dikes as anorthositic magmas, although the original proportions of plagioclase crystals and melt are uncertain. In addition to the differences in plagioclase composition, our studies have documented other contrasts between these two lithologies: Cpx in wall rock (Mg_{77-87}) is strongly zoned in Al, Ti, Na, Fe^{3+} and Mn (decreasing core to rim), but in dike rocks, less abundant Cpx (Mg_{68-87}) has uniformly lower contents of minor elements (Fig. 2); wall rock has an accessory assemblage including hemoilm + mgt + pleonaste + bio + hb, whereas dike rock has traces of hemoilm + bio + myrmekite; wall rock has lower Sr contents and $^{87}Sr/^{86}Sr$, but higher $^{87}Rb/^{86}Sr$ than dike rock (see Figs. 3,4). The isotopic distinction between wall and dike rocks is preserved even for closely sampled (70 cm) pairs. The trend on Figure 3 for the dike rocks (slope > 7000 Ma!) and the scatter among the wall rocks reflect heterogeneities present in these rocks at the time of crystallization.

These data, particularly the isotopic data, appear to preclude a direct comagmatic relationship between the dike and wall rocks, including an origin for the dikes by partial melting of the wall rocks (7). The REE characteristics of the wall and dike rocks also appear distinct (Fig. 5), but analyses of mineral separates from the labradorite anorthosite demonstrate that this difference is simply due to the greater mafic contents (Cpx > Opx) of the wall rock. Although the petrological and Rb-Sr relationships point to different origins for the wall and dike rocks, they appear to have formed from magmas with similar REE characteristics.

Gromet, L. P. and Dymek, R. F.

RECONNAISSANCE OF ST. URBAIN - A broader-based but much less complete study of the massif as a whole confirms that most of it is comprised of andesine anorthosite (An₃₅₋₅₅), and that it has the distinctive Sr concentration, Sr isotopic, and REE characteristics (Fig. 4,5) defined for the dike rocks on MLC. These dike rocks were almost certainly derived from this much larger mass and intruded into pre-existing wall rock of less certain association. In the few leuconorites studied so far, mineral compositions and geochemical features indicate a close relationship to the more leucocratic andesine anorthosites (Figs. 4,6). Again, mineral separates indicate that the REE in the leuconorites differ from those in the andesine anorthosites due to a higher mafic content; REE pattern shapes in leuconorites differ from those in labradorite anorthosite due to Opx>>Cpx in the leuconorites.

IMPLICATIONS FOR CRYSTALLIZATION PROCESSES AND THE NATURE OF PARENT MAGMAS - The andesine anorthosites, the major lithology of the St. Urbain massif, are unusually poor in mafic components (>98% plag). Despite the unequivocal intrusive relationships on MLC, it is difficult to explain the very low mafic content, high Sr contents and very large Eu anomalies without invoking efficient crystal segregation and/or adcumulate growth processes. The pyroxenes in these rocks and in the leuconorites appear to be of intercumulus origin, but available geochemical data suggest that they grew largely by adcumulus/heteradcumulus processes rather than simple orthocumulus crystallization of a trapped melt component (cf. 8,9). In general, the amount of trapped melt in these rocks, including the more mafic leuconorites, appears small.

In addition to the exceptionally plagioclase-rich nature of the andesine anorthosites, their highly fractionated REE patterns and extremely low heavy REE contents distinguish them from other anorthositic rocks of the Grenville and Nain provinces (e.g. 8,9,10; see Nain data, Fig. 6). The extent to which these distinctions are due to differences in the mafic content of the sampled rocks needs to be assessed. Interestingly, leuconorites from St. Urbain have REE patterns which are quite similar to those of the Nain Complex. However, plagioclase from leuconorites have REE patterns similar to those of the andesine anorthosites, as does plagioclase from several Norwegian massif anorthosites (11). We emphasize that the highly fractionated REE patterns of the andesine anorthosites cannot be explained by local subsolidus exchange with mafic minerals because whole rocks were in fact analyzed. Furthermore, zoning in minerals indicates that such exchange is limited to no more than a few mm, and less than 150 μ m in most cases (4,12). We conclude that the andesine anorthosites and leuconorites appear to require a parent magma with highly fractionated REE characteristics. Whether a magma with similar REE characteristics could be parental to other massif anorthosites (cf. 8,9,10) needs to be considered.

The nature of the parent magmas of the St. Urbain massif remains enigmatic, but features such as the coexistence of sodic plag with relatively Mg-rich mafic minerals, the oxidizing conditions indicated by primary hemoilmenite, the highly fractionated REE patterns and the high Sr contents do not correspond to the properties expected of cumulates from mafic melts derived from mantle sources. The possibilities of deriving the parent magmas by partial melting of basaltic and/or crustal terrains and subsequent extensive Fe-Ti oxide fractionation at high P need to be carefully evaluated.

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Gromet, L. P. and Dymek, R. F.



Fig. 1
Generalized geologic map of the St. Urbain area. Extent of the lab. ano. unit may be much less than shown (after 1,2).

Fig. 2
Pyroxene compositions - MLC. Note differences in minor elements in cpx for andesine and labradorite anorthosites.

Fig. 3
Rb-Sr isotopic relationships - MLC. Dashed line connects a dike-wall rock pair separated by 70 cm. 1100 MY reference line indicates isotopic evolution over maximum likely age of the massif.

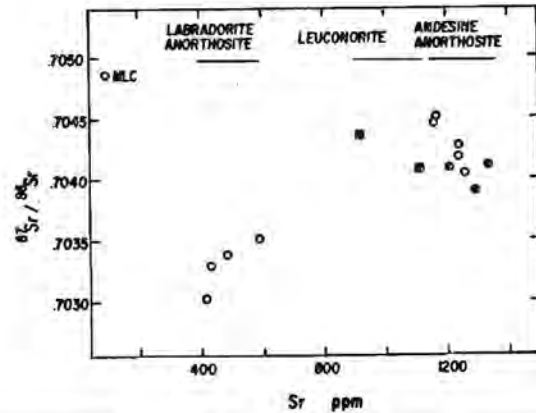
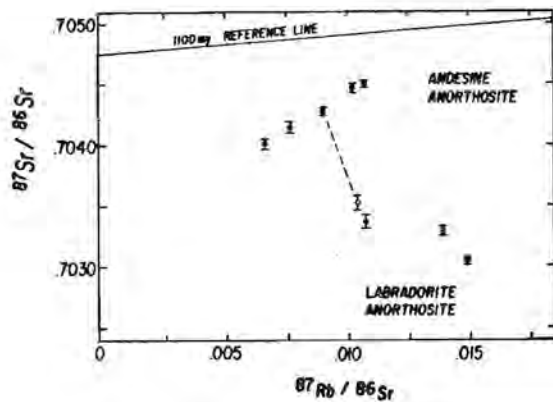
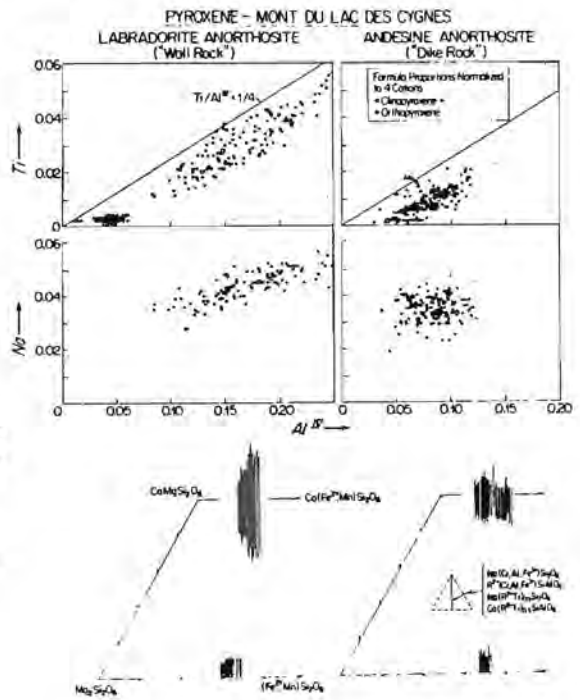
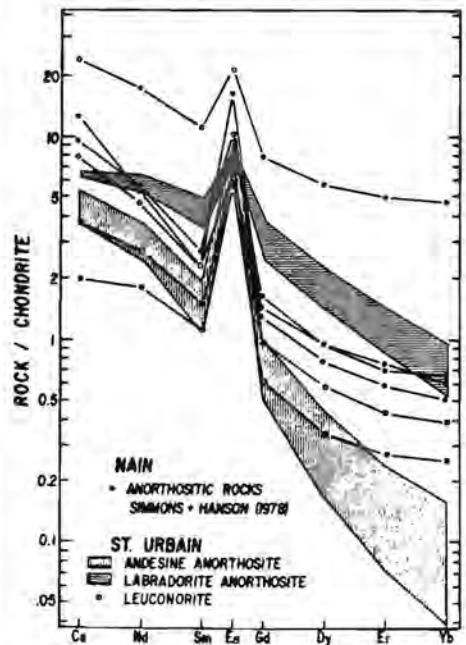
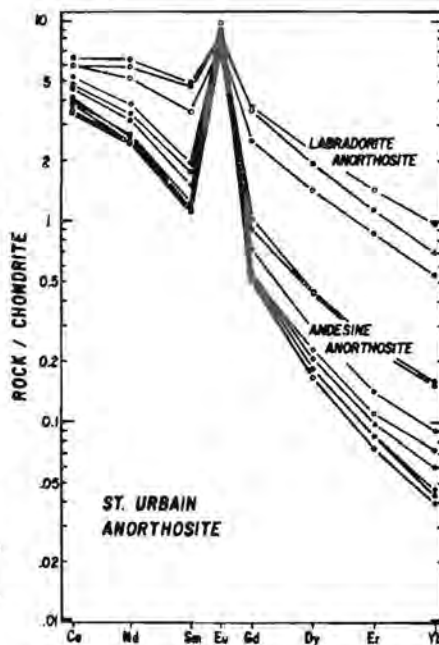


Fig. 4
 $^{87}\text{Sr}/^{86}\text{Sr}$ vs. Sr for St. Urbain lithologies. Open symbols are from MLC. Note similarities between leuconorites and and. ano. (plag from ln. has Sr contents equal to andesine anorthosite).

Fig. 5
REE in lab. ano. (all from MLC) and and. ano. (5 from MLC; 3 not).

Fig. 6
Anorthosites and leuconorites from St. Urbain compared to some anorthositic rocks from Nain. Note similarity between the ln. and the Nain rks.



SOME GEOCHEMICAL CONSIDERATIONS IN CONSTRAINING ANORTHOSITE GENESIS;
 Larry A. Haskin and Marilyn M. Lindstrom, Dept. of Earth and Planetary
 Sciences and McDonnell Center for the Space Sciences, Washington University,
 St. Louis, MO 63130.

Despite considerable field, petrographic, and compositional data to the contrary, the following notions about how anorthosites form seem ingrained. REE data have been used to support these ideas; we believe that this support is poorly founded.

1. Anorthosites begin as massive accumulations of plagioclase in a cooling magma. Evidence exists for magmas of anorthositic composition (e.g., 1) but magmas of gabbroic or intermediate composition are more often considered (e.g., 2, 3, 4). Most such magmas would precipitate plagioclase and one or more mafic minerals simultaneously, requiring a mechanism for separating the plagioclase from the coaccumulating mafic minerals.

2. Substantial quantities of residual liquid from the parent magma would initially fill the interstices among the plagioclase crystals. The bulk of this liquid is lost from the anorthosite by the processes of adcumulus growth and filter pressing. These processes are regarded as adequate to produce large amounts of anorthosite that are as pure as 98% plagioclase (5).

3. The bulk of the mafic constituents of most anorthosites and related materials is solidification products of interstitial residual liquid from parent magma that was not expelled by filter pressing and adcumulus growth. Entrapped coaccumulated mafic minerals have been observed in some massifs (e.g., 6) but are rare.

4. The bulk of the residual liquid from the parent magma of the anorthosites (which would have to be abundant to provide for the presumed adcumulus growth if a mafic magma is assumed) is either represented by the mafic, intermediate, or granitic rocks associated with massif anorthosites, or is missing and presumed lost.

REE abundances have been taken as supporting the above scenario and even constraining it on the basis of the following features:

A. Most anorthosites (but not all, e.g., 7) have large positive Eu anomalies as expected for plagioclase cumulates (from magmas with small or no Eu anomalies).

B. For many anorthosites and related rocks (gabbroic anorthosites, anorthositic gabbros, etc.), REE concentrations (especially those for HREE) increase with increasing modal abundances of non-plagioclase minerals (e.g., 8). This is expected because REE (except Eu) are incompatible with plagioclase so that entrapped residual liquid would contribute disproportionately higher concentrations.

C. The size of the positive Eu anomaly decreases as the trapped liquid component increases, a corollary of B.

D. By dividing the REE concentrations in plagioclase that has been separated from anorthosites by values for plagioclase distribution coefficients (D), REE concentrations of the residual liquid of the parent magma can be calculated. The limiting factor of such calculations is usually taken to be uncertainties in D values, especially for Eu, but slopes of calculated REE patterns are usually regarded as accurate. Steep patterns calculated for residual liquids of terrestrial anorthosites have been interpreted to imply garnet in the magma source region (e.g., 3, 7).

E. "Minor" discrepancies between expectations and observations for REE seem to be dismissed as unimportant effects of poor analysis, poor sampling, lack of adequate D values, REE mobility during alteration, other

Haskin, L.A. and Lindstrom, M.

specific second-order events, or the notion that real rocks are just too complex to describe quantitatively.

It is also assumed that the petrological processes believed to produce anorthosites on Earth are fully adequate to produce them on Moon despite some apparent major differences, which include the following:

i. Lunar ferroan anorthosites (FAN) are much more calcic (An 96-98) than common terrestrial ones (see 9 for summary of FAN characteristics).

ii. To the extent that our sampling is representative, lunar anorthosites with plagioclase contents $> 95\%$ are the rule (usually $\geq 98\%$) and associated igneous rocks with substantially lower plagioclase concentrations have not been recognized as abundant. On Earth, anorthosites with $> 95\%$ plagioclase (e.g., 10, 11, 12, 13) are rare, most tend to be small portions of less pure anorthosite bodies, and most bodies contain or are associated with substantial quantities of related rocks with much higher modal proportions of mafic minerals.

iii. Sr initial ratios in lunar anorthosites are so low that those rocks must have been isolated from significant amounts of Rb since very early in solar-system history ($< 10^8$ years, 19). Terrestrial anorthosites are much younger.

iv. Presumed primary igneous textures have been obliterated from lunar anorthosites, presumably a consequence of episodes of tectonic stress (meteoroid impacts). Many terrestrial anorthosites retain primary textures (e.g., 9, 13).

v. Lunar anorthosites are an areally major constituent of the lunar crust (e.g., 15). Terrestrial anorthosites seem volumetrically minor (e.g., 2).

We suggest that REE data (among other types) do not support the above scenario for those anorthosites studied in greatest detail. Consider the following examples:

I. Marcy anorthosite and related materials (16; see also 3, 8 17). Some 80 samples from a 12,500 km area have been analyzed by INAA and divide into two well separated compositional groups. The low REE group ($0.3 < Sm < 5$, $> 52\%$ plagioclase) has increasing concentrations of MgO, FeO, and incompatible elements, low but steeply increasing concentrations of Cr, Co, and Sc, decreasing Na₂O concentrations, and increasing Sm/Eu, decreasing La/Yb and La/Sm, all with increasing REE concentrations. The high REE group ($8 < Sm < 15$, 42-90% plagioclase) has increasing concentrations of FeO, Co, and Sc, decreasing Na₂O concentrations and increasing Sm/Eu and La/Sm with increasing REE concentrations. It is compositionally more variable than the low REE group and has overlapping MgO, FeO, Ba, Rb, Co, Sc, Th, Ta, and Hf and overlapping or lower concentrations Na₂O, Cr and Sr. Nearly all samples from the interior of the Marcy body as well as many from the margins belong to the low REE group and nearly all samples from the high REE group come from the margins. The high REE group cannot be low REE samples plus a high REE contaminant because of its low Cr concentrations. The two groups converge on variation diagrams to a common region, that represented by the samples with highest modal % plagioclase. If the trends are assumed to result from mixtures of cumulate plagioclase plus trapped liquid, compositions for two liquids can be derived. The REE pattern for neither liquid matches that of the presumed residual liquid calculated from the pattern for pure plagioclase. Nor does any analyzed sample contain a representative portion of either hypothetical liquid.

II. Samples from a single Marcy hand specimen (18). A slab of AA-10 ($\geq 95\%$ plagioclase) some 5 cm x 2.1 cm x 1.3 cm was cut and cleaned.

Haskin, L.A. and Lindstrom, M.

Seven wafers, 0.6 mm thick, ~ 0.5g each, were sliced off at 7.5 mm intervals for INAA. Facing thin sections were prepared for each wafer, separated from the analyzed wafers only by 0.3 mm of kerf. Plagioclase modes ranged from 83.5 to 98.5%, pyroxene from 1.2 to 10.1%, and opaque minerals from 0.3 to 6.4%. No correlation was observed between REE concentrations or patterns and modal composition as interpolated from the facing thin sections. Six of the samples gave a good Cr-Sm correlation but the slope was distinctly different from that for both the high REE and low REE trends found for the Marcy whole rock samples. FeO concentration correlated roughly with pyroxene content; no correlation between FeO and Co or Cr was observed. No simple interpretation in terms of cumulate plagioclase and trapped liquid component or cocumulate minerals could be made.

III. Lunar ferroan anorthosite, FAN. Most samples of FAN contain only sparse pyroxene or pyroxene plus olivine, and only traces of other minerals except plagioclase, which is often > 98% abundant (e.g., 9). REE concentrations vary by more than an order of magnitude, La/Sm more than a factor of 2, and Sm/Yb more than a factor of 5 (e.g., 18). Sm/Eu varies by more than a factor of 5 but, unlike Sm/Yb or La/Sm, the majority of samples have low Sm/Eu (< 0.1), and many approach the ratio of D values (~0.05) determined experimentally for lunar oxygen fugacities (19,20). For a parent liquid with no Eu anomaly or a slightly negative one, 5% of trapped residual liquid would raise Sm/Eu from the value for plagioclase of < 0.05 to a value > 0.10. Thus, the commonly reported > 2% of pyroxene in the least mafic FAN samples (e.g., 9) might not represent the > 5% of trapped liquid that would correspond to that amount. It is not clear that adcumulus growth can be presumed to produce massive amounts of anorthosite as free from mafic minerals as FAN, nor is adcumulus growth implicated in producing the Nain anorthosites (13). In terrestrial layered intrusions, anorthosites have > 10% trapped liquid component (11,12) such that mesocumulates, not adcumulates, are the rule (11).

Such efficient removal of residual liquid from FAN in a single stage by a process such as adcumulus growth would require conditions enabling the process to advance to an extreme extent. A multistage process such as partial remelting of mafic mineral bearing anorthosite and partial loss of successive liquids can remove both mafic mineral components and incompatible elements efficiently without extreme expulsion of any one liquid.

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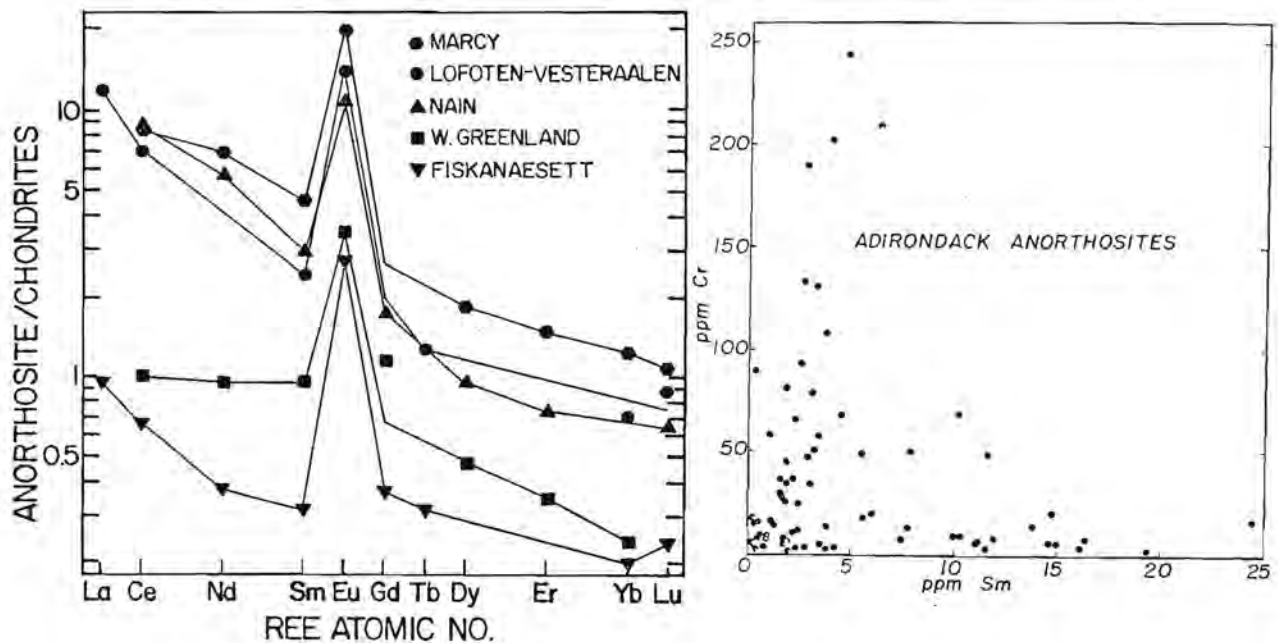


Fig. 1. Typical terrestrial anorthosites; Marcy, ref. 3; Nain, 3; Greenland, 7; Fiskanaesett, 21; Lofoten, 22.

Fig. 2. Cr-Sm variation diagram for Marcy anorthosite and related rocks; figure from ref. 16.

Haskin, L.A. and Lindstrom, M.

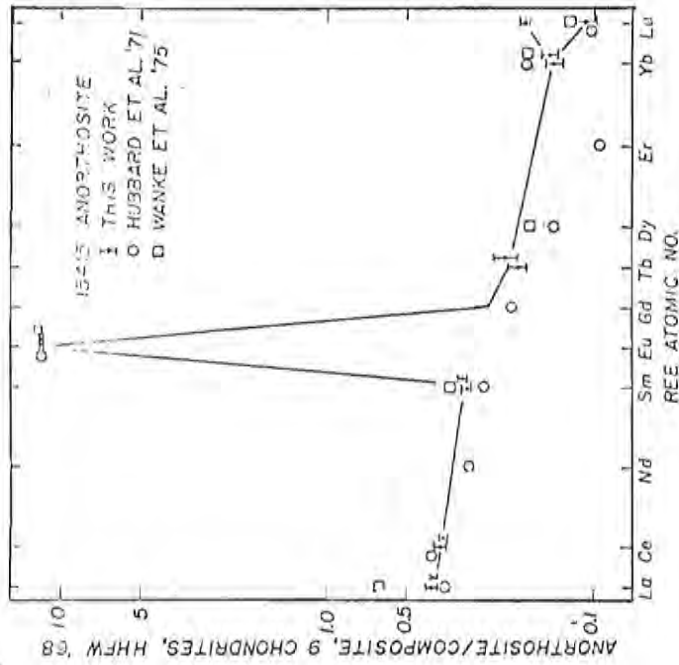


Fig. 4. Typical lunar ferroan anorthosite; figure from ref. 18.

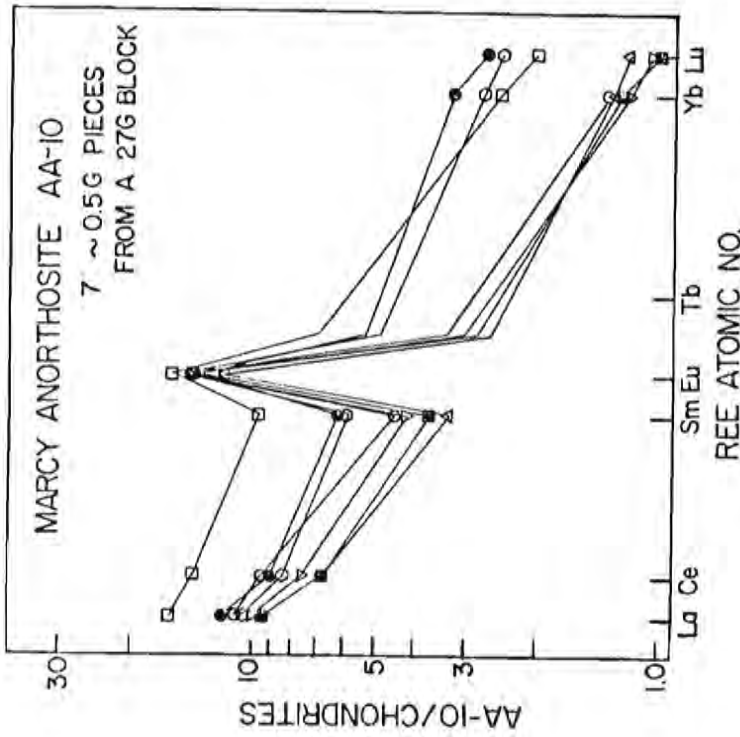


Fig. 3. REE distributions in seven wafers of Marcy anorthosite slab; figure from ref. 18.

PYROXENE-FELDSPAR COMPOSITION TRENDS IN ACHONDRITES: PARALLELS TO STILLWATER AND LUNAR HIGHLANDS. Roger H. Hewins, Geological Sciences, Rutgers University, New Brunswick, N.J. 08903.

Given the complexity of processes operating in large magma bodies, it is very difficult to deduce igneous evolution where field relations are lacking, as in samples collected in a regolith. Meteorites are more difficult to interpret than lunar highland rocks, because first the case for origin on a common parent body must be made. The close relationship between diogenites (orthopyroxenites) and cumulate eucrites (norites) is inferred from the presence of clasts of similar material in polymict breccias (howardites) and a large body of geochemical data is consistent with this generally accepted idea. However, the many attempts to define the precise nature of this relationship have had limited success.

Stolper (1977) showed that cumulate eucrites could not have been produced from a liquid equivalent to the common basaltic eucrites and that a common origin for diogenites and cumulate eucrites by fractional crystallization of a more Mg-rich melt is plausible. However, Mittlefehldt and Drake (1981) showed from modelling of major and trace element fractionation that these rocks could not be comagmatic. Indeed, Ma and Schmitt (1979) using similar models showed that there may be no simple relationship between the known cumulate eucrites.

Phase composition data for diogenites (olivine-chromite-plagioclase orthopyroxenites) are relevant to the question of complex relationships between diogenites and cumulate eucrites. Hewins (1981a) reported subsolidus equilibration in diogenites to temperatures of about 900°C, based on olivine-chromite and orthopyroxene-augite compositions. This is similar to closure temperatures for pyroxene exsolution in cumulate eucrites (Takeda et al., 1981) and is consistent with similar thermal histories for both types of rock.

There is evidence of fractionation within diogenites, consistent with sequential cotectic crystallization. This includes variation in Fe/Mg and Cr/Al in chromite from diogenite to diogenite and also from cumulus to inter-cumulus chromite in one diogenite. The fractionation sequence, from high temperature to low, is Allan Hills A77256, Roda, Garland. For the same samples, compositions of coexisting orthopyroxene and plagioclase have been plotted in Fig. 1. There is a range from En₇₇An₉₂ to En₆₇An₈₅, with the crystallization order of these rocks the same as for the chromite sequence.

The compositions of plagioclase and pigeonite from cumulate eucrites (Duke and Silver, 1967; Lovering, 1975; Garcia and Prinz, 1978; Harlow et al., 1979a) have also been plotted in Fig. 1. The two separate trends for diogenites and cumulate eucrites form a pattern strikingly familiar from work on the lunar highlands pristine rocks and the Stillwater complex. The achondrite data are replotted in Fig. 2 along with lunar and Stillwater data taken from Raedeke and McCallum (1980).

The lack of correlation between the composition of plagioclase and pyroxene in cumulate eucrites is similar to the situation in lunar anorthosites which accumulated by plagioclase flotation. This suggests that plagioclase may have been carried in suspension in the magma for some time before deposition along with cumulus pigeonite as cumulate eucrite.

Hewins, Roger H.

In an attempt to formulate a relationship between diogenites and cumulate eucrites, it would be tempting to assume that the plagioclase in cumulate eucrites crystallized from the same melts as the most Mg-rich pyroxene in diogenites. However, unlike Stillwater, the achondrite and lunar data show gaps suggesting pairs of distinct magmas (or magma groups) which is supported by modelling (Raedeke and McCallum, 1980; Mittlefehldt and Drake, 1981). Following ideas expressed by Stolper (1977) and Raedecke and McCallum (1980), a sequence of basaltic eucrite flows, cumulate eucrite plutons, and diogenite plutons may be proposed, as a result of progressive partial melting in the achondrite parent body. In the light of the great diversity of basalt clasts recently discovered in the polymict achondrites (howardites) by Delaney et al. (1981) and in the few "not-quite-monomict" cumulate eucrites and diogenites (Garcia and Prinz, 1978; Hewins, 1981b), there is no need for these plutons to be restricted and simple. Hence the suggestions of different types of diogenite (Heyse, 1975; Hewins, 1980) and unrelated cumulate eucrites (Ma and Schmitt, 1979) may be realistic.

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Hewins, Roger H.

Fig. 1.

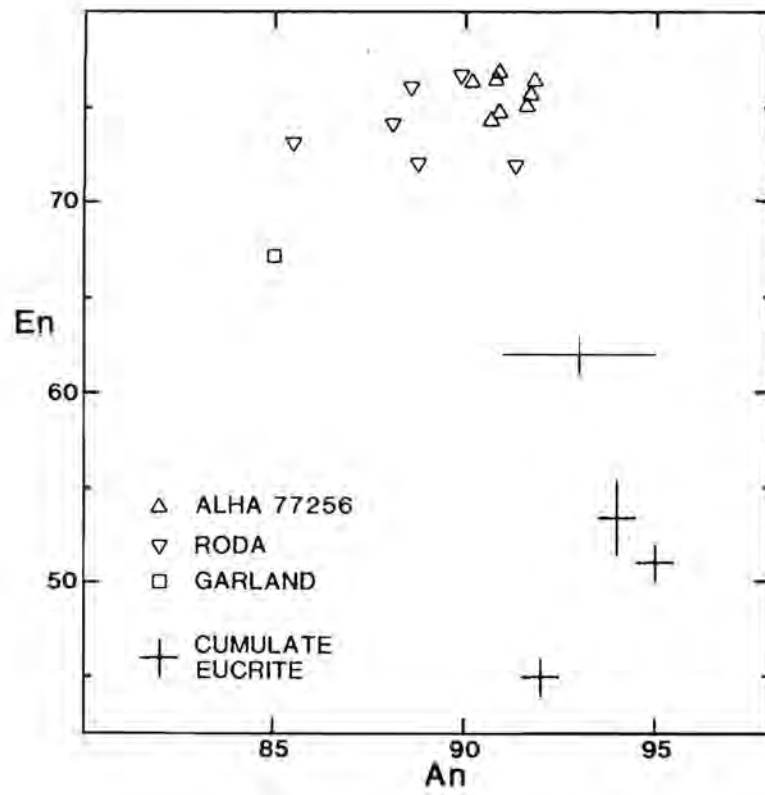
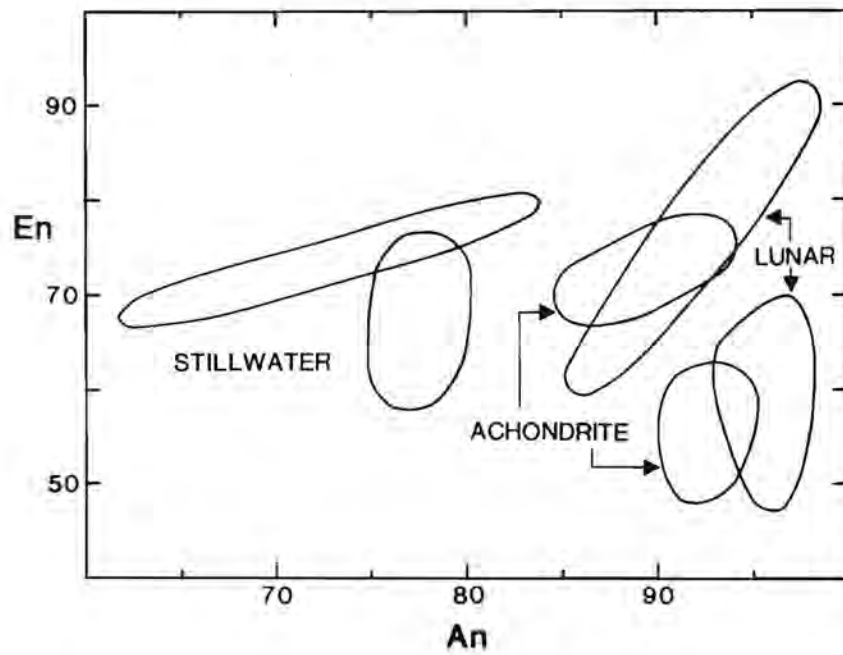


Fig. 2.



PLUTONIC PROCESSES IN SPREADING OCEAN-RIDGE MAGMA CHAMBERS, ILLUSTRATED BY THE SAMAIL OPHIOLITE, OMAN. C. A. Hopson and J. S. Pallister, University of California, Santa Barbara, Calif. 93106, and U. S. Geological Survey, American Embassy, APO New York 09697.

The Samail ophiolite (95 m.y.), exposed for >450 km along the Oman Mountains, reveals a stratigraphically intact oceanic crustal and upper mantle sequence, from peridotite and dunite tectonite through layered and high-level isotropic gabbros, diabase sheeted dike complex, to basaltic submarine lavas overlain by oceanic sediments (1,2). Origin at a spreading ocean ridge is indicated by: 1) the occurrence of sheeted dikes everywhere overlying the plutonic sequence, demonstrating a continuously dilating and accreting roof above the magma chamber, 2) MORB chemistry for the sheeted dikes and lavas (1,3,4), and 3) radiolarian-bearing pelagic and metaliferous hydrothermal sediments (MnFe umbers) overlying and intercalated within the lavas (5), suggesting an open-ocean setting with ridge-crest type hot-spring activity accompanying volcanism. The ophiolite sequence has the thickness and seismic velocity structure of normal oceanic crust and uppermost mantle (6).

The plutonic sequence displays the crystallization and fractionation products of a steady-state magma chamber that existed beneath the ridge crest. This sequence in the Ibra area (southeastern Oman Mountains), determined from four measured stratigraphic sections each spaced ca. 5 km apart normal to the sheeted dike trend (ridge axis direction), show the following upward vertical progression: 1) dunite (chr-ol cumulates) 0-200 m thick, grading up from a transition zone with harzburgite tectonite, 2) interlayered wehrlite-melagabbro-gabbro (cpx-ol and ol-cpx-pl cumulates) 0-100 m thick, 3) layered gabbro (chiefly ol-cpx-pl cumulates but including recurrent intervals of cumulus wehrlite and melagabbro) 2.6-5.5 km thick, 4) planar laminated non-layered gabbro (chiefly ol-cpx-pl cumulates) 100-400 m thick, grading to 5) texturally variable non-cumulus (ol)-hb-cpx gabbro 200-800 m thick, 6) small, discontinuous diorite to plagiogranite bodies near the top of the gabbro (7). Cumulus textures (adcumulus > mesocumulus), planar lamination, and cumulus layering (phase, ratio, and grain-size layers at mm to 10-m scale, commonly graded) within this sequence show that the crystals accumulated from the base of the magma chamber upward to within a few hundred meters of the top; downward solidification from the roof was minor. Cut-and-fill structures and local foreset lamination within the gabbro layers have probably resulted from magmatic current action. Cyclicity within the cumulus sequence is represented by the recurrence of olivine-rich intervals (wehrlite, melagabbro) up to high stratigraphic levels, and by hundreds of phase-graded layers that are olivine-rich at the base. Cryptic variation with very steep zig-zag trends and the limited compositional range of cumulus olivine (FO_{69-90}), plagioclase (AN_{62-95}) and clinopyroxene (EN_{40-54} , FS_{4-16} , WO_{37-49}) through 3-5 km of section require open system crystallization (periodic replenishment). Zig-zag normal and reverse cryptic variation differentiation trends indicate that replenishment took place in pulses, where periods of magma mixing alternated with crystal fractionation. Recurrent olivine-rich intervals commonly coincide with reverse cryptic variation trends. They are thought to result from fresh draughts of primitive magma, which commonly had olivine and chromite as the only liquidus phases prior to mixing with the more fractionated resident magma within the chamber. This main magma body rarely fractionated beyond the ol-cpx-pl cotectic until a very late stage of crystallization, as shown by the predominance of ol-cpx-pl cumulates throughout most of the section, and the absence of cumulus orthopyroxene. The inability of the melt to fractionate beyond the olivine reaction point during the build up of 3-5 km of cumulates is further evidence that open system fractionation was maintained to a very

Hopson, C. A. and Pallister, J. S.

late stage (7). Exceptions are found in the northern Oman Mountains, where cumulus opx appears early in some of the layered gabbro sections (8).

Field evidence, including continuity of layering over large areas, distinctive layer sets that are mapped for several km along strike, and the lack of cross cutting (chamber edge) contacts indicates that crystallization of the Samail ophiolite plutonic sequence took place in a single, large, long-lived magma chamber rather than in small ephemeral chambers (7,8). The chamber shape, controlled by floor growth upward greatly exceeding roof growth downward as the chamber halves diverged from the paleo-spreading axis, must have been funnel-shaped in cross section; i.e., the floor sloped inward to the center from a sandwich horizon that developed just beneath the roof at both sides. This conclusion is based on the occurrence of cumulates up to a very high level in the stratigraphic sections, and on geologic mapping which shows that cumulus layering locally approaches the base of the cumulus section at an angle (7). This vergence angle is $<20^\circ$; moreover, the scarcity of slump structures and layer disruption within the main layered sequence also verifies that the floor sloped rather gently. Thus, the half-width of the chamber must have exceeded its vertical dimension of about 3-5 km. Chamber half-widths of about 11-18 km are calculated from a preferred vergence angle of about 15° (7). The cumulates crystallized within this spreading chamber from a magma body that was repeatedly replenished by primitive magma from a central feeder zone at the spreading axis. High-level gabbro, diorite, and plagiogranite crystallized from stagnant residual magma at the outer edges of the chamber, beyond reach of replenishment from the center (7).

The plutonic zones of other ophiolites that originated at oceanic spreading centers vary considerably in stratigraphic thickness, composition, internal structure, sequence of cumulus phases, and many other details. A common characteristic, however, is that gabbroic and ultramafic cumulates tend to be significantly thicker than the overlying non-cumulus gabbros, diorite, and plagiogranite. This implies funnel-shaped rather than bell-shaped magma chambers, if the chambers were steady state. Also, cyclic units appear to be common where the layered cumulates have been closely studied (8,9,10), and further, the main cumulus phases typically display very steep cryptic variation trends (8,11,12). Thus, open system crystallization appears to be normal in ophiolites. A current point of controversy concerns the origin of the "cumulus" layering itself. Is most of this layering the product of gravitational crystal settling, or has much of it been produced by in-situ crystallization in a static boundary layer along the floor and walls of the chamber, as proposed by McBirney and Noyes (8,13,14)?

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THE EFFICIENCY OF IN SITU FRACTIONATION: VISCOSITY PARAMETERS.

Robert H. Hunter and Lawrence A. Taylor, Department of Geological Sciences,
University of Tennessee, Knoxville, TN 37916

Introduction. Perfect crystal-liquid fractionation (Rayleigh fractionation) requires complete isolation of the growing crystal from its host liquid. If the crystal - liquid density contrast results in a stress great enough to overcome the yield strength of the liquid, the crystal may settle under the influence of gravity, at a rate controlled by Stokes' Law, and become isolated from that liquid. This process may only be effective in liquids of basic composition or when the viscosity of a more evolved liquid is lowered by the presence of certain volatile elements. Rayleigh fractionation may also occur when isolation of the crystals is effected by displacement of the liquid, i.e., filter pressing.

In some cases, the crystals remain at the site of liquid fractionation because of rapid crystallization, melt viscosity, and/or size of magma chamber. This "in situ fractionation" is controlled largely by nucleation, growth, and other kinetic considerations. The crystallization must be sufficiently rapid so that existing crystals do not have adequate time for back-reaction and become rimmed with new compositions. This leads to an effective isolation of the crystal interiors and the fractionating liquid.

The factors which control the nature of the fractionating mechanisms i.e., mechanical versus in situ, are: melt volume, cooling rate, gravity, viscosity (melt composition), and other fluid dynamic parameters. The degree to which fractionation occurs, will depend on the extent to which the crystals and melt are isolated. The two factors which will have the greatest effect on this "efficiency" are the melt viscosity and the cooling rate. The viscosity is primarily a function of the ratio of bridging to non-bridging oxygens and therefore, increases with increasing silica content of the melt. Water acts as a network-breaker, depolymerizing the melt and lowering the viscosity. The cooling rate depends upon the ambient thermal gradient and is a function of the size and depth of the magma chamber.

In considerations of fractionation, it is necessary to define a parameter reflecting the degree of attainment of completeness. Thus, the efficiency is defined as the degree of perfect fractionation that is attained. As a measure of this efficiency, chemical parameters such as ΔX between successive liquids can be used, where X is some differentiation index, eg., SiO₂ or mg.

In situations where the effects of gravity are negligible and heat loss is primarily by conduction (i.e., in small intrusions), in situ crystallization is the dominant process, and fractionation is controlled largely by kinetic processes. The efficiency of fractionation is dependant on the rate of crystal growth and diffusive transport. Given a uniform thermal gradient, the viscosity of the melt becomes the primary influencing factor on the diffusion coefficients and consequently, the efficiency of fractionation.

In situ fractionation. In this paper we will present a case study of in situ fractionation which will allow us to evaluate the efficiency of such crystallizing systems. The case study describes aspects of the Carrock Fell complex of Cumbria Co., Northern England. The detailed field relations and petrogenesis are described elsewhere (1). Three sub-vertical sheets, Units B2, B3, and B4, represent successive batches of fractionating magma from depth. There is a systematic modal, cryptic, and bulk-chemical variation laterally within each sheet that has resulted from in situ fractionation within an asymmetric thermal gradient. The overall scheme of fractionation is from ferrogabbro to ferrogranophyre, but each sheet shows overlap in variation. Unit B5 is a homogeneous granophyre that represents the latest stages of differentiation of the intrusion.

Plagioclase, Ca-rich cpx, and FeTi oxides occur throughout the sequence; apatite, amphibole, and zircon appear in Units B2, B3, and B4 respectively.

Hunter R.H. and Taylor L.A.

Fig. 1.

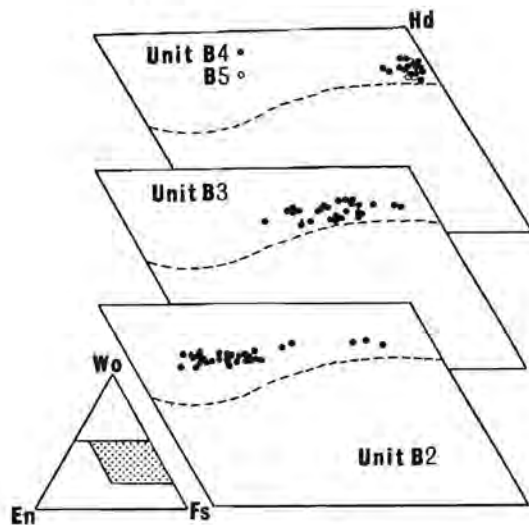
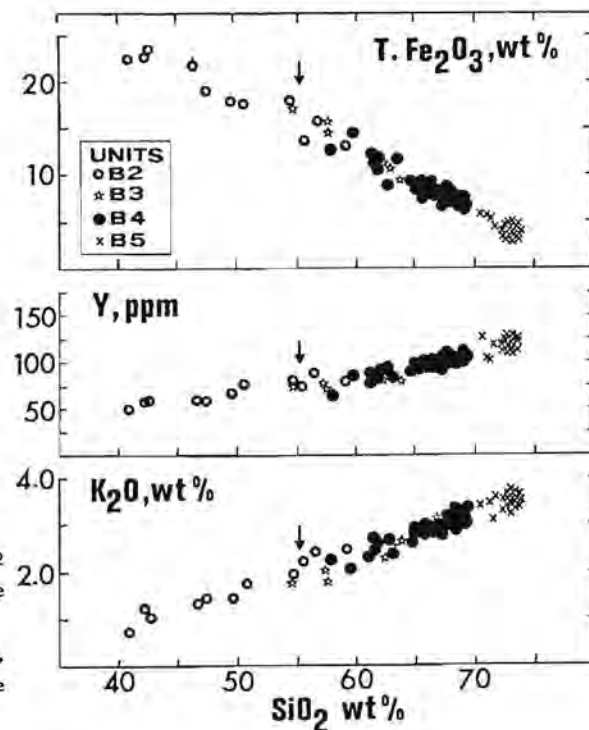


Fig. 1. Pyroxene crystallization trends for Units B2 to B5 of the Carrock Fell intrusion. Dashed line is the Skaergaard trend for comparison.

Fig. 2. Variation in bulk-rock total Iron (as Fe_2O_3), K_2O , and Y vs. SiO_2 for the Carrock Fell intrusion. The arrow indicates the estimated parental composition.

Fig. 2.



Modally, each sheet shows an increase in granophytic intergrowth from one margin to the other with a concomitant decrease in silicates and oxides. There is also a systematic change in crystal morphology with fractionation, leading to the development of pegmatitic facies within the more evolved portions of the sheets.

The variation in pyroxene chemistry is shown in Fig. 1. The overall variation is from $\text{Wo}_{41}\text{En}_{33}$ to $\text{Wo}_{44}\text{En}_2$. Zonation within pyroxenes from individual samples is negligible, but the variation within units is substantial. Plag shows an overall variation from $\text{An}_{56}\text{Or}_{10}$ to $\text{An}_{20}\text{Or}_{10}$. The range of bulk-rock variation in total iron, K_2O , and Y is shown plotted versus SiO_2 (Fig. 2). The liquid composition that has resulted in a particular bulk-rock composition lies on a mixing line between the bulk-composition of the crystallizing phases and the composition of the trapped residual liquid. Because the overall trends for most elements form straight lines between B2 and B5, the successive liquids produced during fractionation are defined by that line. The approximate composition of the parental melt (basaltic andesite, $\approx 55\% \text{SiO}_2$) has been shown in Fig. 2.

The efficiency of the fractionation process in the Carrock Fell intrusion is indicated by the changes in differentiation indices; $\Delta \text{mg} \approx .54$, $\Delta \text{An} \approx 36$, and $\Delta \text{SiO}_2 \approx 20\text{wt}\%$. The overall scheme is similar to the latter stages of fractionation of many, much larger, tholeiitic intrusions. Indeed, the variation at Carrock Fell that has occurred over horizontal distances of 20-50m, is equivalent to that which has occurred over 500-1000m vertically in the Skaergaard intrusion, for instance. The dependence of the efficiency of in situ fractionation, to a large extent, upon melt viscosity has been established above. Therefore, we must conclude that in order to achieve a similar efficiency within such small volumes of magma, in the absence of convective or gravitational processes, the diffusion coefficients must have been increased by lowered viscosity. In this respect, the development of pegmatitic facies within the units is of relevance, indicating that volatile saturation was achieved during the

Hunter R.H. and Taylor L.A.

crystallization of each sheet. It has been postulated that the intrusion crystallized at a total pressure of 1 kb, and the estimated parental composition corresponds to 75% fractionation of what might be considered a parental melt composition of mantle derivation. Fig. 3 shows the degree of crystallization of anhydrous phases required to achieve saturation for a given depth and initial water content. At 1 kb and seventy-five percent crystallization, the melt may have contained as much as 4 wt% H₂O before saturation. Efficiency is high in saturated systems where gas-transport is a contributing factor to diffusion rates. The role of depth on saturation levels and the stability of hydrous phases are limiting factors to the increase in water content in fractionating systems; however, further discussion will be restricted to water undersaturated systems and the effect of water content on the efficiency of fractionation.

Viscosity. To a large extent, diffusion coefficients can be equated with melt-viscosity; thus, an evaluation of the role of water in lowering viscosity is necessary in order to assess the efficiency of *in situ* fractionation. In evolved tholeiitic magmas, the three principal compositional parameters affecting viscosity are FeO, SiO₂, and H₂O. Melts of high iron content can have low viscosity due to their relatively depolymerized nature. During the intermediate stages of fractionation, when Fe increases and Si remains constant or even decreases, we may expect the viscosity to

Fig. 3.

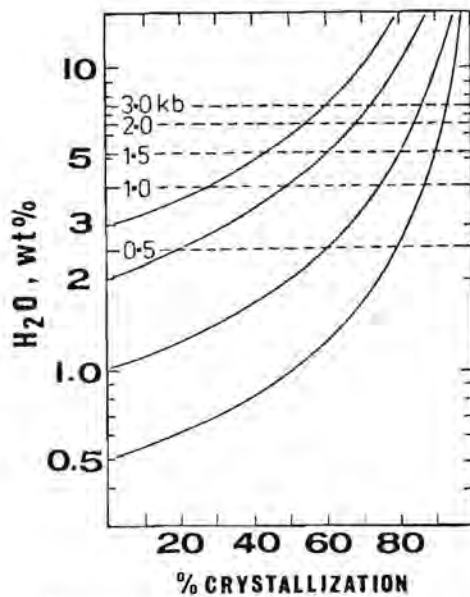


Fig. 3. The calculated change in wt% H₂O (solid lines) with degree of crystallization of anhydrous phases for initial concentrations of 0.5, 1.0, 2.0 and 3.0 wt% H₂O. Dashed lines indicate water saturation levels at various total pressures (kb) in intermediate to granitic compositions (saturation data are from (4) and (5)).

Fig. 4.

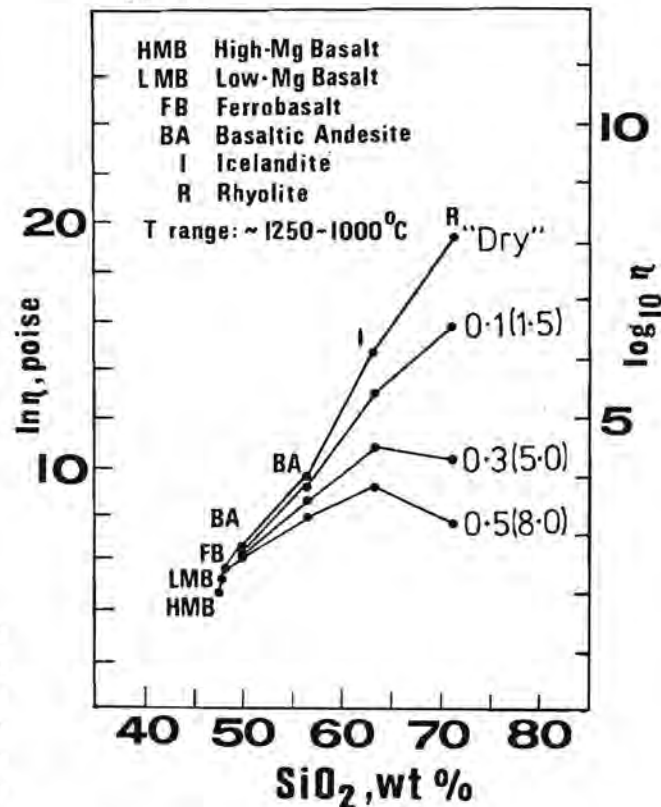


Fig. 4. Change in calculated viscosity (η) with SiO₂ content for a suite of tholeiitic lava compositions from East Iceland (data from (2)). FB corresponds to 65%, BA to 70-75%, I to 87%, and R to 93% fractionation from the HMB parent. The curves correspond to viscosity changes during fractionation with initial H₂O contents of zero, 0.1, 0.3, and 0.5 wt%. The numbers in parentheses indicate water contents (wt%) after 93% fractionation. (Liquidus temperatures over the crystallization interval have been assumed, and the effects of water saturation and hydrous phase formation have been omitted).

Hunter R.H. and Taylor L.A.

remain constant or to decrease slightly. At some stage during fractionation, dependant upon the oxygen fugacity, the FeTi oxide stability field is intersected. Beyond this point the structure of the melt changes, becoming more polymerized with increasing Si and decreasing Fe content. The rate of Si increase will be a function of the plagioclase to pyroxene ratio in the crystallizing fraction, but we may expect SiO₂ to increase linearly. The increasing Si content, coupled with the effects of decreasing temperature, should lead to an increase in viscosity and therein a decrease in the efficiency during continued fractionation. However, this may not be the case in hydrous systems; an increase in water, because of its network-modifying characteristics, acts to decrease melt-viscosity. Fig. 3 shows the water content to increase exponentially with crystallization of anhydrous phases. At early to intermediate stages of fractionation, the change in concentration, and hence effect of water, will be negligible. However, as crystallization proceeds, the role of water becomes more pronounced, and its effects on viscosity are substantial.

Wood (2) has presented bulk-rock and mineral analyses of a suite of tholeiitic lavas, ranging in composition from high-Mg basalt to rhyolite, from E. Iceland. He interpreted the variation in the context of varying degrees of fractionation, using least squares modelling techniques. The water contents at varying degrees of fractionation may be estimated from Fig. 3, assuming an initial content. Using these various data, and the method of Shaw (3), we have calculated the changing viscosity with fractionation at different water contents. This permits the evaluation of the combined effects of FeO, SiO₂, and H₂O on the efficiency of fractionation processes. The results are shown in Fig. 4. For the purpose of calculation, we have estimated the liquidus temperatures for the various compositions. During the first 65% of fractionation, there is relatively little change in the viscosity or silica content. The subsequent increase in both these parameters over the interval 65-75% fractionation results from the crystallization of FeTi oxides. In the cases with higher water contents, there is an inflection in the viscosity curves at approximately 60-65% SiO₂ (icelandite). This corresponds to the point at which the contribution of SiO₂ to increasing viscosity is balanced by the counteracting effect of H₂O. With further crystallization, viscosity decreases as the role of H₂O becomes dominant.

For efficient in situ fractionation to occur, effective diffusion rates must be maintained with progressive crystallization. This places limits on the increase in viscosity during fractionation. If the initial water content is greater than 0.3 wt%, the melt viscosity does not increase by more than two to three orders of magnitude, and indeed, may decrease during the latter stages of fractionation of a tholeiitic melt. Thus, efficient fractionation is not limited to large, slowly cooled magma volumes at depth. We conclude that the effect of water on the viscosity of a melt may be of sufficient magnitude, for small volumes of magma to achieve efficient in situ differentiation, comparable in degree to large layered plutons at depth. The Carrock Fell intrusion provides an illustrative case in point.

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MULTICOMPONENT PHASE DIAGRAMS AND THE PHASE EQUILIBRIA OF BASALTS.

John Longhi, Dept. of Geology and Geophysics, Yale Univ., P.O. Box 6666, New Haven, CT 06511.

Although more than 10 components may be necessary to describe the chemical variation of basalts completely, most of the chemical variation of basalts and their phase equilibria can be depicted by relatively simple phase diagrams consisting of four mineral components: olivine ($(\text{Mg,Fe})_2\text{SiO}_4$), wollastonite (CaSiO_3), plagioclase ($(\text{Na}_x, \text{Ca}_{1-x})\text{Al}_{2-x}\text{Si}_{2+x}\text{O}_8$) and silica. The topological basis for constructing the phase boundaries of natural liquids lies within the simple system forsterite (Mg_2SiO_4) - wollastonite - anorthite ($\text{CaAl}_2\text{Si}_2\text{O}_8$) - silica (Fig. 1). Most of the liquidus boundaries at low pressure have been accurately located by microprobe analysis of multiply-saturated quenched liquids (eg. 1). Most of the liquidus boundaries at high pressures up to 20 kb, relevant to basalt genesis, can be drawn as a result of the work of 2 and 3. Comparison of experimental results from work on natural systems (4,5,6) shows that the positions and general topologies of liquidus boundaries of lunar and terrestrial basalts are similar to those in the simple four component system at both low and high pressure when the analyses of the multi-component experimental liquids are recast into the four major mineral components. In other words the addition of components such as FeO, alkalis, O_2 , TiO_2 and Cr_2O_3 in the amounts found in the more primitive natural basalts has relatively little effect on the topologies of the liquidus curves. Addition of these components does, of course, cause the curves to shift and lowers temperature. Therefore, once the positions of the liquidus boundaries are known in the simple system, positions of natural liquidus boundaries can be determined experimentally and expressed in an analogous manner. This principle has been employed for some time by petrologists (eg. 7 and 8) and recent experimental data has led merely to refinements of these earlier efforts.

Even four components, however, present a graphical problem, as suggested by Figure 1. Therefore, projections of 3-dimensional features onto 2-dimensional planes are required. Figures 2, 3 and 4 illustrate the three projections within the model four component system needed to represent the liquidus equilibria of basalt. Light solid lines are from the simple 4 component system; heavy lines are drawn from experimental data on terrestrial tholeiites at low pressure (5). Liquidus boundaries obtained from lunar compositions give similar results (4) but are not shown for the sake of clarity. Also for the sake of clarity the low-Ca pyroxene phases--protoenstatite, orthoenstatite and pigeonite--are lumped together in the natural case. Experimental work (4) has shown that the pigeonite field expands and the protoenstatite and orthoenstatite shrink with decreasing $\text{MgO}/(\text{MgO}+\text{FeO})$ ($\equiv \text{Mg}$), such that protoenstatite disappears at $\text{Mg} < 0.75$, and that pigeonite coexists with olivine, anorthite and liquid at $\text{Mg} \leq 0.75$ in lunar liquids and at $\text{Mg} \leq 0.5$ in tholeiitic liquids. Figures 3 and 4 are projections of a 3-dimensional saturation surface from a saturating phase with constant composition (at least in terms of the projections) and behave much the same as ternary liquidus diagrams. Projecting from CaSiO_3 rather than $\text{CaMgSi}_2\text{O}_6$ is preferred in Figure 2 because high-Ca pyroxene does not remain constant in composition along the diopside saturation surface and projecting from $\text{CaMgSi}_2\text{O}_6$ causes a greater dispersion than does projecting from CaSiO_3 .

Perhaps the feature most important to basalt genesis on these diagrams is the thermal divide (TD-B) along the olivine + anorthite + high-Ca pyroxene cotectic (2). Liquids which reach the cotectic at the high silica side of the

Longhi, J.

divide produce a tholeiitic crystallization sequence: early olivine, plagioclase and high-Ca pyroxene with low-Ca pyroxene crystallizing later. Liquids which reach the cotectic at the low-silica side of the thermal divide fractionate towards lower SiO_2 and higher CaO contents and never crystallize low-Ca pyroxene--these are the alkali-olivine basalts. Figures 2-4 show the migration of the 5 phase assemblage olivine - 2 pyroxenes - an aluminous phase (plagioclase, spinel or garnet) - liquid as a function of pressure in both the simple (3) and natural (6) systems. Given current prejudices about the composition of the earth's upper mantle (point "M"), these projection points approximately fix the composition of primary magma during the first 15-25% partial melting under anhydrous conditions. More advanced partial melting, either equilibrium or fractional, at pressures ≤ 25 kb will lead to low-Al, high-Ca liquids saturated with olivine and two pyroxenes--these liquids resemble komatiites (point "K" in Figures 2 and 3).

One of the simplest and most important conclusions to be drawn from this rough analysis is that high-Ca pyroxene ought to precede low-Ca pyroxene (if low-Ca pyroxene crystallizes at all) during crystallization at pressures ≤ 5 kb of any primitive magma produced by low to intermediate degrees of partial melting. Plotted in Figures 2-4 are chilled margin or estimated bulk compositions from several terrestrial layered intrusions (8,9,10,11). These compositions correspond to primitive to somewhat evolved tholeiites. Despite a general similarity of bulk compositions, there are significant differences in observed crystallization sequences between the various intrusions that would not be expected from examination of the phase diagrams. There is a crude correlation between crystallization sequence, size of the intrusion and roof relationships which may be the key to understanding how these magma bodies crystallize. In two of the smaller bodies (Skaegaard, Kiglapait) high-Ca pyroxene precedes low-Ca pyroxene and there is a marginal upper border group. In the larger bodies (Bushveld, Great Dyke, Stillwater) low-Ca pyroxene precedes high-Ca pyroxene and where the roof is exposed (Bushveld) there is no upper border group, but rather ferrosyenites and melanogranophyres, some of which may represent fused country rocks (9). In the Muskox intrusion (8), which is of intermediate size, high-Ca pyroxene crystallizes before low-Ca pyroxene initially, but the order reverses approximately 1600 m above the base; the roof of the Muskox contains a granophyre, but no upper border group.

Irvine (8) proposed that contamination of the Muskox parent magma with fused country rock was responsible for the change-over in crystallization sequence. The ability to fuse and assimilate country rock ought to increase with the size of the magma body, and in this respect the crystallization sequences of the Bushveld, Great Dyke and Stillwater are consistent with the assimilation hypothesis. Sm-Nd measurements on the Stillwater rocks are also consistent with crustal assimilation (12). Field relations, however, require that the assimilation took place before appreciable crystallization took place. The path of assimilation may be approximated by drawing a line between point C" (here assumed to be the composition of partially fused crust as well as the terminus of fractional crystallization) and some primitive magma composition. Assimilating crustal material into the chilled margin compositions (Muskox, Bushveld, Stillwater) will produce a composition which projects from olivine into the low-Ca pyroxene field. It should also be noted that assimilating C" into a high pressure minimum melting composition might also produce the chilled margin compositions themselves. From the point of view of phase equilibria the exact nature of the assimilation process--mechanical or diffusional (13)--cannot be determined.

Longhi, J.

An alternative process to assimilation that is also consistent with the correlation of size and crystallization sequence in layered intrusions is pressure increase due to down-warping of an overloaded crust. In Figures 2-4 it can be seen that the low-Ca pyroxene field expands with pressure. If the magma composition were already evolved beyond the chilled margin compositions (Bu, St, Mk) towards silica, then downwarping the crust several kilometers could produce a change in the crystallization sequence and cause low-Ca pyroxene to crystallize before high-Ca pyroxene. In the case of the Bushveld, Stillwater and Great Dyke it would be necessary for the main feeder dike to have fed the chambers from the side and nearer to the top than to the present bottom of the intrusions in order to have this mechanism work.

There are some similarities between proposed intrusions in the ancient lunar crust (14) and the larger terrestrial intrusions that beg comparison. Assimilative processes have been invoked for the Mg-rich series of lunar dunite troctolites, norites and gabbro to explain fractionation of Sm/Ti, Sm/Sc and the REE (15,16) and low-Ca pyroxene appears to have crystallized well before high-Ca pyroxene. First it should be noted that lunar intrusions (if that is what they were) took place in an anorthositic rather than sialic crust: assimilation of plagioclase would have had relatively little effect upon the crystallization sequence of pyroxenes. Second, in the larger terrestrial intrusions gabbro is a major rock type in the plagioclase-cotectic series despite the early crystallization of the low-Ca pyroxene, whereas lunar gabbros are sparse and do not appear to be a significant portion of soils or breccias either. These differences lead to the conclusion that the parental magmas of the lunar Mg-rich series had significantly lower wollastonite contents than their terrestrial analogs. The lunar parental magmas may have been produced by large degrees of partial melting (more so than komatiites), which produced the low wollastonite contents, followed by assimilation of anorthosite and a KREEP-like component upon intrusion into the crust (1).

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Longhi, J.

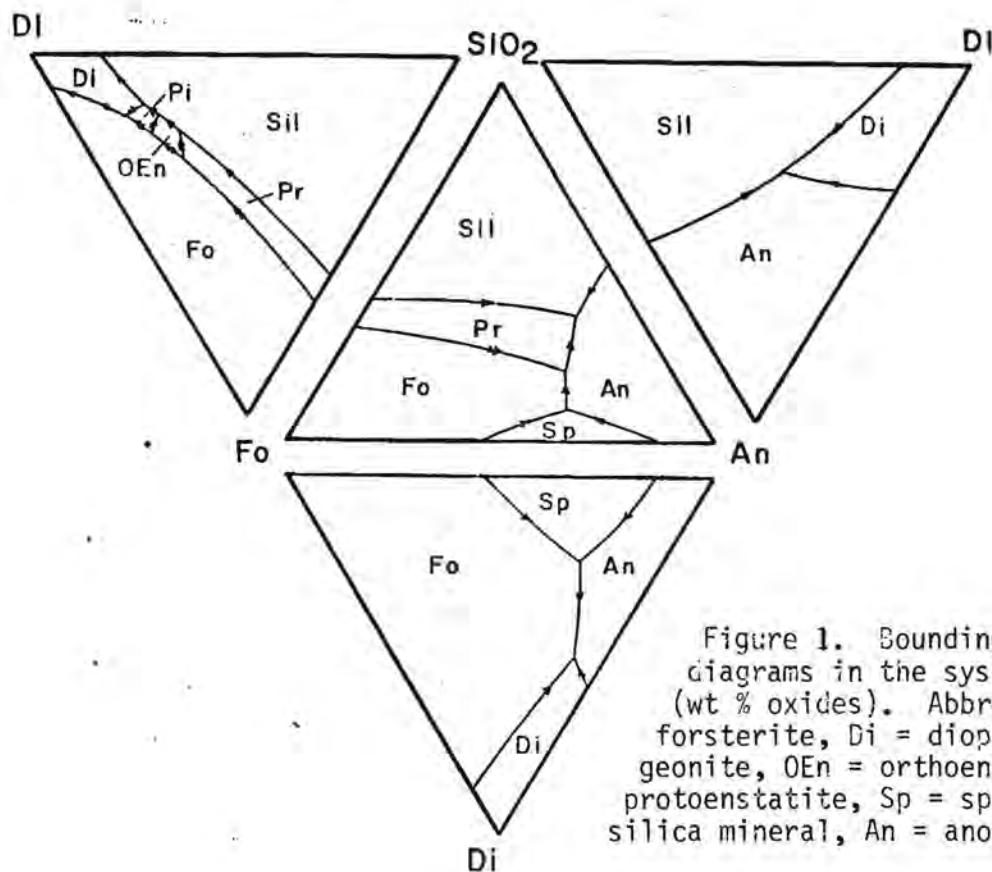


Figure 1. Bounding ternary phase diagrams in the system Fo-Di-An-SiO₂ (wt % oxides). Abbreviations: Fo = forsterite, Di = diopside, Pi = pigeonite, OEn = orthoenstatite, Pr = protoenstatite, Sp = spinel, Sil = silica mineral, An = anorthite.

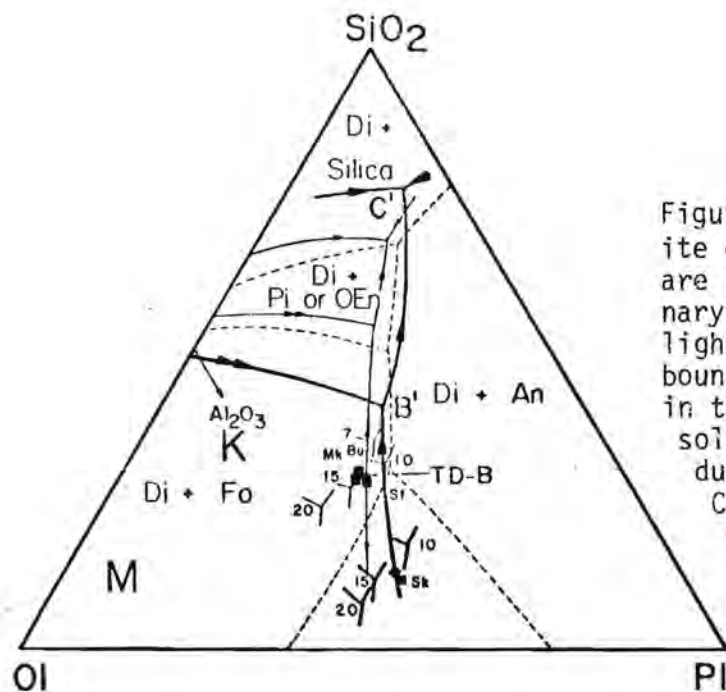


Figure 2. Projection from wollastonite component. Light dashed curves are liquidus boundaries in the ternary system Fo-An-SiO₂ (mole %); light solid curves are liquidus boundaries saturated with diopside in the Fo-An-Wo-SiO₂ system; heavy solid curves are tholeiitic liquidus boundaries saturated with high-Ca pyroxene from (5). Numbers refer to pressure in kilobars of minimum melting composition for model mantle of composition "M". K is composition of typical komatiite. Solid squares are chilled margin compositions of some layered intrusions: Sk = Skaergaard; Bu = Bushveld; St = Stillwater; Mk = Muskox. Solid diamonds are thermal divides on liquidus boundaries.

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Longhi, J.

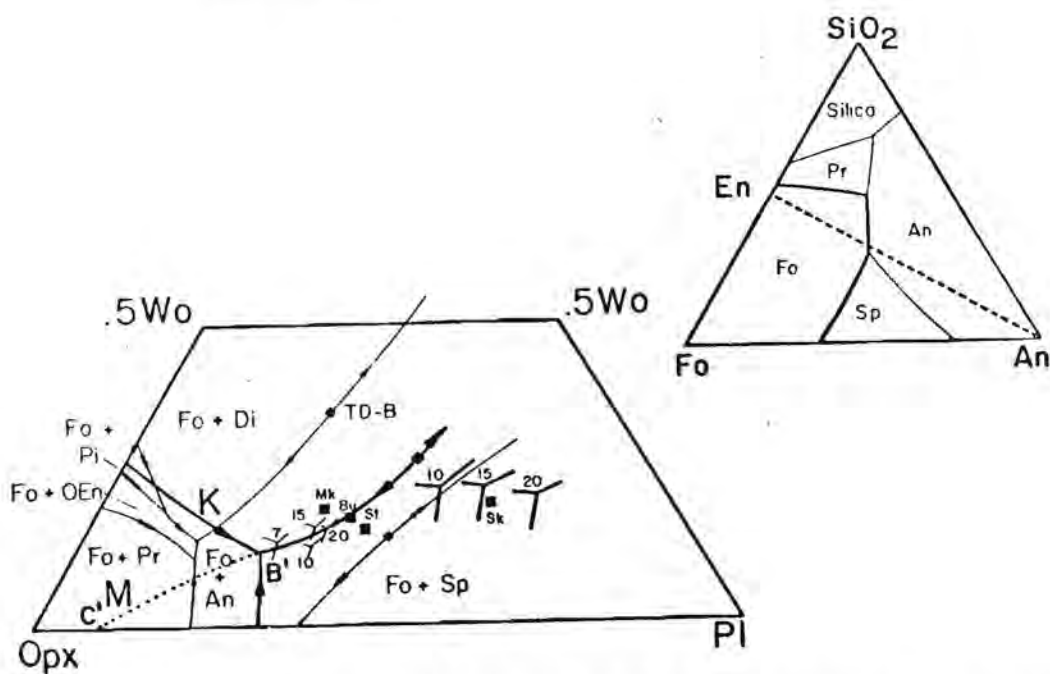
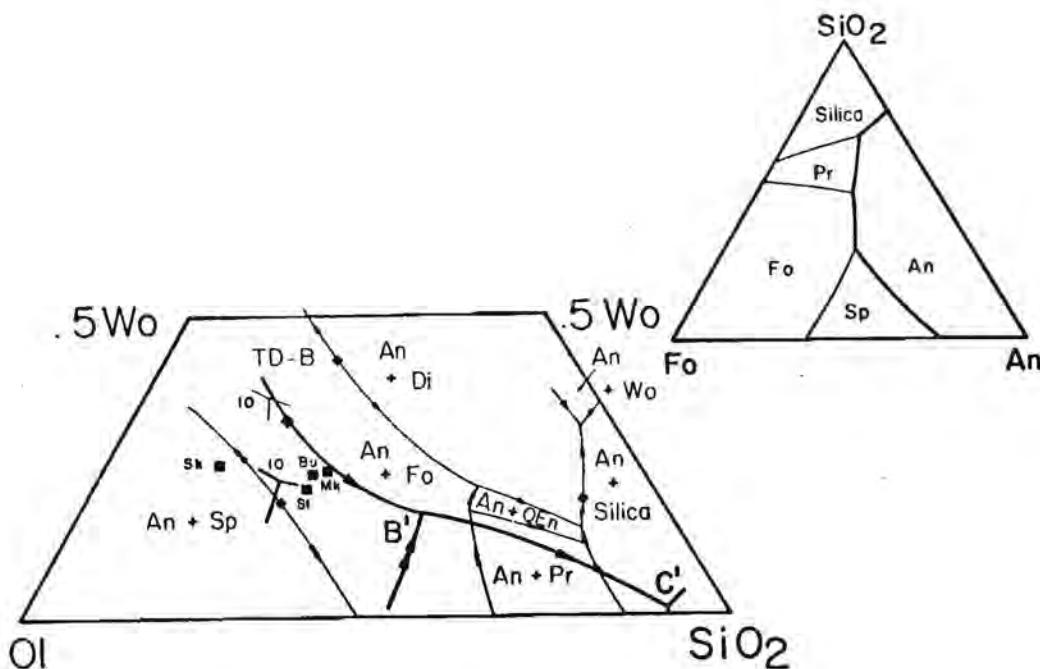


Figure 3. Projection from forsterite (simple system) or olivine (natural system) onto the orthopyroxene (enstatite - plagioclase (anorthite) - wollastonite plane of the olivine saturation surface. Symbols as in Figure 2. Dotted curve is the projection of the two pyroxene + plagioclase cotectic (no olivine). Inset shows the trace of the forsterite saturation surface (heavy curve) in the Fo-An-SiO₂ system.

Figure 4. Projection from anorthite (simple system) or plagioclase (natural system) onto the olivine - wollastonite - SiO₂ plane. Symbols same as in Figures 2 and 3. Inset shows trace of the anorthite saturation surface (heavy curve).



FE-TI OXIDES OF THE STILLWATER COMPLEX, MONTANA: MINERALOGY, CHEMISTRY, AND IMPLICATIONS. W.B. Maze, Princeton University, Dept. Geological and Geophysical Sciences, Princeton, NJ 08544. and R.R. Carlson, USGS, Golden, CO 80401

Fe-Ti oxides in the Stillwater Complex are not well known. Prior to the present study (1), the only reported occurrence was the brief mention by Hess (2).* Thus this study has concentrated on defining the distribution and chemistry of Fe-Ti oxides and associated silicates in the Stillwater Complex in the hope of defining parameters that controlled crystallization in the upper zones.

The Fe-Ti oxides examined for this study came from two widely separated horizons: 1) the top 500m of the exposed Upper Banded Zone (UBZ) between Picket Pin Mtn. and the west side of the Boulder River (9 miles to the west) and 2) at least one thin horizon approximately 350m above the base of the lowest Norite unit. These horizons appear to be laterally continuous on a gross scale, although there are many breaks in the fine-scale layering and mineralogy. Within these horizons there are many layers with Fe-Ti oxides, but they apparently never exceed modal concentrations of several percent.

Within the two horizons there are four distinct modes of oxide occurrence: 1) 0.5-1mm interstitial ilmenite grains within the lowest Norite unit, 2) approximately 1cm magnetite-ilmenite grains in 1-3 cm thick pegmatoid layers of the UBZ with pyroxene, plagioclase (~An 60), and minor quartz and amphibole, 3) discrete interstitial grains (<5mm) and oikocrysts (~2cm dia.) within coarse-grained anorthosite (An 55-66) of the UBZ that contains 5-10% interstitial clinopyroxene and orthopyroxene as well as minor amounts of myrmekite, quartz, sulfides, and apatite (A cumulus phase that has not been previously reported.), and 4) 1-10 μ m oxide grains within "glass" inclusions in the plagioclase of the second and third types above. Field and petrographic observations lead us to believe that the oxide horizons are primary features. However, the compositions of the oxides suggest that subsolidus reequilibration has continued to very low temperatures. There is usually <1wt% Ti in the magnetite and the hematite component of the ilmenite is less than 5% (excluding fine hematite lamellae). The oxide inclusions in the plagioclase apparently retain some of their primary composition. The oxygen fugacity and temperature calculated from these compositions range from log f_{O_2} = -13 and 800°C to log f_{O_2} = -11 and 970°C. These oxides are within 10-60 μ m inclusions in the plagioclase that look like glass, yet have a composition almost identical to that of the orthopyroxenes in the same rocks. Because of their location and isolation these oxides were unable to chemically exchange with nearby oxides (cf. Duchesne (3)). The other three types of occurrence are interstitial to the silicates and thus may have been more susceptible to fluid migration. Similar variations in oxide compositions have been reported from the Bjerkrem-Sogndal Massif of Southwestern Norway and have been attributed to deuteric alteration (3).

* During the field trips run for the Workshop on Magmatic Processes of Early Planetary Crusts it was pointed out by L.D. Raedeke and I.S. McCallum that similar oxides also occur in both the first and second anorthosite units. In the field they appear similar to the oxides at the top of the UBZ. More discussion of the oxides in the lower anorthosites will be found in the PhD dissertation "Petrogenesis of the Stillwater Complex" by L.D. Raedeke of University of Washington, Seattle.

Maze, W.B. and Carlson, R.R.

Evidence for a fluid phase comes from both the presence of amphibole with and without pyroxene cores in the pegmatoid layers, and from fluid inclusions in the interstitial quartz from the coarse-grained anorthosite and the pegmatoid layers. Preliminary microthermometric measurements of the aqueous brine inclusions in this quartz indicate maximum inclusion formation temperatures of approximately 250°C (pressure corrected for 1kb) and approximately 20 wt% NaCl equivalent (although densities vary considerably). This temperature is consistent with the occurrence of chlorite, and perhaps with the serpentinization seen in the ultramafic Stillwater rocks.

The Fe-Ti oxides in the coarse-grained anorthosites (plagioclase grains 1-1.5cm long) are the most common. They can constitute 1-2% of the anorthosite in the uppermost horizons north of Picket Pin Mtn. 5-10% of the rock is pyroxene (both cpx and inverted pigeonite) with minor amounts of apatite, Cu-Fe, Fe-Ni, Zn-(Fe?), and Fe sulfides, and interstitial quartz. The concentration of sulfides may be higher in the pegmatoid layers than in the coarse-grained anorthosite. Complex sulfide-sulfide relationships exist. In general, as the modal content of oxide in the rock increases the modal pyroxene decreases. Both pyroxene and oxide appear interstitial in the field, but petrographic examination shows that some pyroxenes are euhedral to subhedral and may have been cumulus phases.

With few exceptions the magnetite and ilmenite in the anorthosites and pegmatoids of the UBZ occur as intergrowths rather than separate grains. However discrete ilmenite grains appear in the lowest Norite unit. A typical oxide grain from the UBZ contains a granule of ilmenite in, or on the edge of, a magnetite grain. The magnetite is cut by several coarse ilmenite lamellae and the ilmenite has μm -width hematite lamellae. The ilmenite lamellae often have more Mn and Ti than the granules.

Zinc spinels ($\sim 5-10\mu\text{m}$) are found in the magnetites, and rod-like zircons up to approximately $50\mu\text{m}$ in length occur in the ilmenite grains. Whole rock Pd values from both the pegmatoid (30ppb) and the coarse anorthosite (7ppb) are above average Stillwater background. Whether this is related to the oxides or sulfides is unknown. Platinum group minerals associated with oxides in the Dufek Intrusion (4) are an indication that there may be more platinum upsection in the Stillwater Complex.

Although cumulus apatite may have formed earlier in the Stillwater, a comparison of the silicate mineralogy here with that of the Bushveld, Skaergaard (5), and Dufek Intrusions (6) shows that all four crystallized Fe-Ti oxides under similar conditions. In the oxide horizons of the UBZ, the Stillwater plagioclase is $\sim \text{An } 60$ and the pyroxene compositions are WO:EN:FS= 35-46:31-40:14-27 for cpx and 2-10:52-64:33-42 for opx (including inv. pigeonite)(see figs.1 and 2). In the other 3 intrusions, the lowest oxide horizons have plagioclase compositions of approximately An 60 and pyroxene compositions that range from WO:EN:FS= 3-9:56-69:27-37 for orthopyroxenes (including inv. pigeonite) to WO:EN:FS=36-40:41-48:15-21 for the clinopyroxenes. If the Stillwater's initial bulk composition was similar to that of the other 3, perhaps no more than 50% of the original Stillwater is exposed.

Chemical trends in the silicates and oxides at the top of the exposed UBZ are presented in figures 1 and 2. Plagioclase compositions remain relatively constant, while pyroxene compositions increase in Mg/Mg+Fe (atomic) with increasing stratigraphic height. This is a similar situation to that described by Raedeke and McCallum (7) for the Middle Banded Zone (MBZ) in that plagioclase compositions are constant. However, in contrast to what they observed, the pyroxene Mg# increases upsection in the UBZ oxide

Maze, W.B. and Carlson, R.R.

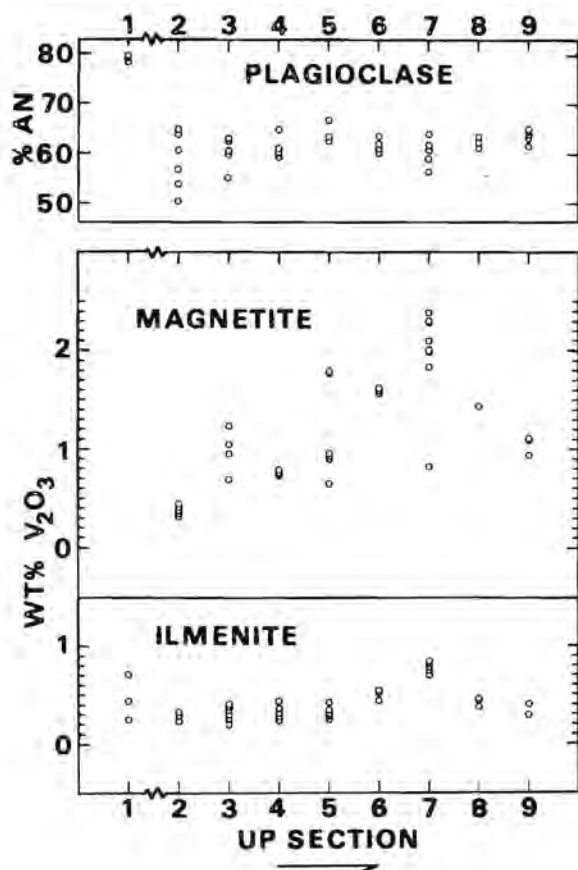


Figure 1.

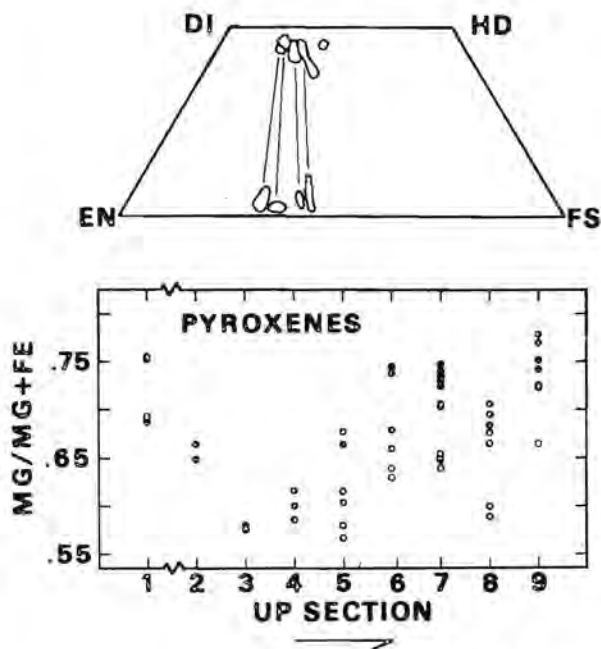


Figure 2.

Figure 1. An composition of plagioclase and V₂O₃ in ilmenite and magnetite vs. stratigraphic height (plag by EDA and V by WDA using a Ti correction). Number 1 is from the lowest Norite unit and the other 8 are from the top of the UBZ nearly 4500m upsection from number 1.

Figure 2. Pyroxene quadrilateral showing UBZ and Norite zone pyroxene compositions and Mg/Mg+Fe vs. stratigraphic height (all analyses by EDA). The open circles represent orthopyroxenes and the filled circles represent clinopyroxenes. Host phases and lamellae were probed separately. The variation in Mg/Mg+Fe (Mg#) in the clinopyroxenes at each stratigraphic height is probably caused by the inclusion of lamellae in the analyses. This is verified by a higher Ca content in the clinopyroxenes that have a higher Mg#. Thus the overall increase in Mg# with height is best defined by the clinopyroxene (and orthopyroxene) analysis with the highest Mg# at each horizon.

Maze, W.B. and Carlson, R.R.

horizons. They proposed that plagioclase compositions were buffered by equilibrium crystallization of "trapped" intercumulus liquid in a plagioclase-rich crystalline mush. However, in the oxide horizons pyroxene compositions may have been controlled by the presence of Fe-Ti oxides. The oxides crystallized when the liquid had more Fe than could partition into the pyroxenes. We suggest that the continued growth of the oxides depleted Fe in the liquid, causing a reverse pyroxene-Mg# trend. Rough mass balance calculations indicate that a small amount of magnetite could have a large effect on the pyroxene compositions. Similarly, fig.1 shows that the V content of the oxides reaches a maximum in sample 7, and decreases in 8 and 9. This suggests that all the V is removed by a small amount of oxide. If this is so, then some process other than mass diffusion must have been responsible for the V trend. Similar trends for Mn, Al, and Si in the oxides indicate that the V trend is the result of primary crystallization and not later mobilization. This suggests to us that the interstitial phases in the oxide horizons were not isolated, but were able to communicate with the main reservoir as they crystallized. There is apparently no modal effect on the compositions as has been described by Raedeke and McCallum (7,8).

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Any magma that crystallizes in a gravitational field will tend to become compositionally zoned while still in the liquid state. This is a result of the density changes that stem from cooling and from the compositional evolution of the liquid.

Because the interior temperature of a crystallizing liquid tends to be buffered uniformly whenever the liquid is in equilibrium with similar crystal assemblages at its margins, the temperature difference between the roof and floor may provide only a very weak driving force for normal thermal convection. Cooling under such conditions may lead to Rahm-Wahlin segregation in which the cooled liquid at the margins descends and remains ponded on the bottom. Because it does not gain enough heat from the floor to rise again, it continues to accumulate and crystallizes essentially in place.

The compositional changes that accompany crystallization have a far greater effect on densities than normal thermal contraction and may either reinforce or oppose thermally driven convection. Depending on whether iron and other heavy components are preferentially fractionated into the crystals or are enriched in the remaining liquid, the boundary layer adjacent to a front of crystallization may either increase or decrease in density. In the case of calc-alkaline magmas, it becomes strongly buoyant and may rise and accumulate under the roof. Tholeiitic and alkaline magmas, however, may become progressively lighter during the early stages of differentiation before plagioclase crystallizes and then become denser during the main stage of strong iron enrichment. In their latest stages, when the rate of enrichment of silica and alkalis exceeds that of iron, the density passes through a maximum, and subsequent liquids become lighter again. Depending on whether or not it has passed over the "density hump", the evolving liquid may flow either to the top or base of a chamber. Many associations of volcanic and intrusive rocks can be explained in terms of these relations.

The process by which a fractionated boundary layer flows along its wall as it rises or sinks may provide a very efficient mechanism of liquid fractionation. Relatively large amounts of differentiated liquid can accumulate in a short time and with only a limited amount of crystallization. The flow regime of the boundary layer will have a strong effect on the rate of accumulation and patterns of differentiation. Turbulent flow results in more rapid rates of accumulation and linear covariance of fractionated components due to back-mixing of the boundary layer with adjacent liquid of the interior. In laminar flow, however, the rate of accumulation is probably slower, and diffusion plays a more important role. The patterns of compositional variations, especially of trace elements, will differ from those predicted for Rayleigh-type fractionation models that ignore diffusion.

Thus, four different types of liquid fractionation are possible, depending on whether the boundary layer sinks or rises and whether it is turbulent or laminar. Most large gabbroic bodies are probably of the sinking - turbulent type; whereas most granitic plutons are of the rising - laminar type. Andesitic compositions seem to be rising - turbulent.

When the gradient of temperature has a configuration that would produce a vertical density profile opposed to that caused by compositional zoning, a distinctive type of convection may develop. This phenomenon, known as double-diffusive convection, leads to a step-like thermal and compositional stratification similar to that of the oceans. Such a mechanism may be responsible for certain types of layering. Differing rates of nucleation, diffusion, and growth of the various mineral phases must also be important and may account for some of the wide variety of types of layering

McBirney A. R.

commonly found in large gabbroic intrusions.

Owing to the different patterns of heat transfer under the various possible convective regimes, rates of solidification may differ widely from one part of an intrusion to another. Depending on its size, composition, and rate of cooling, each body has a certain equilibrium shape, and the liquid fraction will tend to attain this form by crystallizing or melting at different rates on the roof, walls, or floor. This may account for the tendency for most tholeiitic magmas to have a low aspect ratio of height to width, while calc-alkaline bodies are taller and more dome-like.

MAGMA OCEAN EVOLUTION: RAW MATERIAL AND FINISHED PRODUCTS.

G. A. McKay, Code SN7, NASA Johnson Space Center, Houston, Texas 77058.

Introduction. The most widely accepted hypothesis for early lunar history is that the moon underwent global differentiation through crystallization of a magma ocean several hundred km deep (1,2). Primary lunar rock types are commonly interpreted in terms of this hypothesis. Thus, according to the hypothesis, the lunar crust consists largely of feldspathic material concentrated at the magma ocean's surface by floatation, while the lunar mantle consists largely of sunken mafic cumulates. Mare basalts are commonly thought to represent liquids formed by partial remelting of these cumulates. The highly enriched incompatible elements in rocks called KREEP are believed to have been derived from late-stage magma ocean residual liquid.

One of the major remaining tasks in lunar science is gaining a better understanding of the processes involved in the lunar global differentiation. Trace elements can provide important constraints on the nature of these processes. Towards this end, a number of workers have modeled evolution of trace element abundances during crystallization of the magma ocean and remelting of cumulates to form more basalts. These models will not be reviewed in detail, and the reader is referred to Longhi (3) as a comprehensive example. Two of the major issues involved in such modeling are initial composition of the magma ocean, which is the starting point for most models, and the identification of lunar samples whose compositions reflect the trace element evolution of the magma ocean and are therefore capable of constraining the models. This abstract will address some of the implications of these issues for trace element evolution models and constraints which several recent models place on the nature of early lunar global differentiation processes.

Magma Ocean Composition. The major element composition of the magma ocean (generally assumed \approx bulk moon) is usually taken to be that of a feldspar-bearing peridotite. Major questions involve its content of involatile trace elements, including U and REE, and of involatile major elements, including Ca, Ti, and Al. At issue are the overall abundance levels of these elements (relative to chondrites) and their relative abundance patterns (chondritic or otherwise). The most widely accepted estimates for the bulk composition of the moon (4,5,6) derive U abundances of 2-5X chondritic from heat flow data and assume the same chondritic relative abundances for other involatile elements.

The issue of chondritic versus enriched magma ocean involatile trace element abundances is important for trace element evolution models because an ocean with chondritic abundances must undergo significantly more crystallization than a 2-5X enriched ocean to produce cumulates or residual liquids having the same abundance levels. Crystallization products of the chondritic ocean will therefore have had greater opportunity to develop interelement fractionations.

Petrologic arguments (e.g., 7,8) support enrichment in involatile Ca and Al (and hence the involatile trace elements) relative to more volatile Si, Fe, and Mg. According to these arguments, such enrichment is required for formation of troctolites and spinel troctolites observed in the lunar crust. However, several workers (e.g., 9, 10,11) have questioned the involatile element enrichment of the magma ocean. For example, Warren and Wasson (11) argue that (i) uncertainties in interpretation of heat flow data permit chondritic U abundances, (ii) the moon's oxygen isotope composition is close to that of ordinary chondrites, and unlike that of involatile element-rich meteoritic material (CAI's), and (iii) high pressure phase equilibria permit formation of troctolites and spinel troctolites from a magma ocean of chondritic

McKay, G. A.

Ca/Si. The issue of chondritic versus enriched magma ocean abundances is currently unresolved.

The question of chondritic versus fractionated relative abundance patterns is obviously important for trace element evolution models in that it determines how much interelement fractionation must be produced during magma ocean crystallization. Some earlier petrologic arguments (e.g., 7) cast doubt on the assumption of chondritic relative abundances for involatile trace elements in that they required subchondritic Ca/Al in the magma ocean in order to avoid abundant high-Ca pyroxene in the lunar crust, contrary to observation. However, recent experimental determination of pyroxene phase equilibria (12) invalidated the earlier arguments, eliminating this requirement. On the other hand, most models for mare basalt formation (e.g., 13-21) are incapable of generating the observed HREE depletion of mare basalts from a magma ocean having chondritic relative REE without invoking garnet fractionation. Such fractionation would be petrologically implausible in the currently popular shallow magma ocean (e.g., 22) but could be accommodated in a deep magma ocean (e.g., 16). In contrast to most workers, Nyquist et al. (17-21) assume non-chondritic relative REE abundances in the magma ocean. Highland rocks also hint at non-chondritic magma ocean abundance patterns. Inferred parent liquids (see below) of dunite 72417 and several anorthosites (see below) have HREE-depleted abundance patterns which have not been successfully modeled starting with chondritic relative magma ocean abundances (3). Moreover, the inferred dunite parent liquid of McKay et al. (23) has subchondritic Ca/Al which is apparently inconsistent with chondritic magma ocean Ca/Al. It is not yet clear whether the inability of existing trace element evolution models to reconcile initially chondritic relative abundances in the magma ocean with compositions of mare basalts and highland rocks results from inadequacy of current models, early petrogenetic processes not yet accounted for, or an initially non-chondritic moon.

Calculated Versus Observed Abundances. For the highland rocks, a major question is which, if any, samples reflect trace element abundances at some stage during magma ocean crystallization and can therefore be used to constrain evolution models. It is generally believed that the most likely candidates are the pristine anorthosites, but some workers have also attempted to accommodate the Mg-rich pristine plutonic rocks in their models (e.g., 3,8,12). If samples from either or both of these suites crystallized from the magma ocean, and if trace element abundances in the parent liquids of these samples can be correctly inferred, then a successful magma ocean evolution model must produce residual liquids matching these inferred abundances.

Two methods are commonly used to infer trace element abundances in the parent liquids of cumulate rocks. In one method, parent liquid abundances are obtained simply by dividing abundances measured for mineral separates by appropriate crystal/liquid distribution coefficients. This method is valid only if (i) the sampled volume of rock remained closed to trace element exchange after the trapped intercumulus liquid ceased communicating with the main body of liquid, (ii) no intergranular redistribution of trace elements occurred after the volume became closed, and (iii) the mineral separates consist entirely of cumulus crystals and contain no orthocumulus material which crystallized from trapped interstitial liquid. Of these conditions, (ii) is unlikely to be satisfied and (iii) is extremely unlikely to be satisfied for all samples other than very pure adcumulates, because it requires that the mineral separation procedure reject all crystal rims. Thus, parent liquid compositions inferred by this method are likely to be in error.

The second method was developed by Paster et al. (24). In this method a bulk rock sample is considered to have initially consisted of cumulus crystals

McKay, G. A.

of one or more minerals and trapped intercumulus liquid which subsequently crystallized. If the proportions of each cumulus mineral and trapped liquid are known, parent liquid abundances are readily obtained from observed sample abundances using appropriate crystal/liquid distribution coefficients and mass balance considerations. This method is valid only if (i) condition (i) above holds, and (ii) the trapped liquid is the same composition as the liquid from which the cumulus crystals grew. This method has the advantage that it is completely insensitive to redistribution of trace elements provided that scale is small relative to the sample size. In practice, it is often difficult to estimate the proportions of various cumulus phases and trapped liquid. These can often be constrained by major element mass balance, and most workers report a range of parent liquid abundances corresponding to the uncertainties in phase proportions.

It is not clear how often conditions (i) and (ii) for the second method are violated. Haskin et al. (25) have questioned whether condition (i) is satisfied for lunar anorthosites. A series of adjacent slabs from a sample of Marcy Anorthosite showed extreme trace element variations which defied interpretation. Neither could analyses of lunar anorthosites be interpreted in terms of a mixture of cumulus plagioclase and a single mafic component. Furthermore, the parent liquids calculated by the above methods would have fractionated trace element patterns, including positive Eu anomalies, which these authors considered improbable. Hence, to explain these features, as well as the absence of significant volumes of trapped intercumulus liquid which the authors would expect in lunar anorthosites, they proposed that these rocks had suffered partial melting and melt extraction. Of course, such a process would violate condition (i) above, and hence, invalidate parent liquid calculations. However, the arguments presented by Haskin et al. (25) that parent liquid calculations for lunar anorthosites are not meaningful are highly model dependent. Until the method is tested with more terrestrial cumulate rocks, the question of the validity of inferred lunar parent liquid compositions remains unsettled.

Implications for Magma Ocean Evolution. Parent liquid compositions have been inferred using the second method above for a number of pristine lunar cumulates (23,26-31; see Longhi (3) for detailed summary). Several general features are important for magma ocean trace element evolution: (i) All parent liquid patterns show HREE depletion. This feature was apparently established at an early stage, because it is present in liquids having the lowest overall enrichments ($Ce \sim 6X$ chondrite, anorthosite and dunite parents). (ii) Parent liquids for dunites and anorthosites have positive Sr and Eu anomalies. (iii) Parent liquids for troctolite and norite are quite enriched in incompatibles ($Ce = 30-150X$ chond.), despite low Fe/Mg. (iv) Parent liquids for troctolite and norite are depleted in Sc and Ti relative to chondrites, anorthosites, and dunite.

None of these features are consistent with simple models of magma ocean evolution involving only crystallization and accumulation (e.g., 3). In particular, there is a recurring juxtaposition of features indicating evolved trace element abundances and primitive major element compositions. Sc and Ti depletions in norites and troctolites were interpreted by Norman and Ryder (32) as indicating clinopyroxene and ilmenite fractionation. Based on experimental phase equilibria, Longhi (3,12) concluded that ilmenite would begin to precipitate only after the magma ocean was 95% crystallized, at which time the major element composition of the residual liquid would be incapable of producing norites and troctolites. Some workers (e.g., 11,32) have inferred from the above that these samples crystallized from a separate magma not directly

McKay, G. A.

related to the magma ocean. Longhi and coworkers (3,12) have instead invoked mixing of primitive and highly evolved magmas to produce the norite and troctolite parent liquids. The only models addressing the positive Eu and Sr anomalies and HREE depletion in anorthosite and dunite parent liquids are those of Longhi and coworkers (3,12). They appeal to assimilation of plagioclase by hot primitive magma to explain the former. They attempt to model the latter by mixing of primitive and evolved magmas, but these attempts are not completely successful.

Compositional features of KREEP are also inconsistent with simple magma ocean crystallization. Ti depletions, together with high degrees of ilmenite undersaturation, were interpreted by Hess et al. (33) and McKay et al. (26) as indicating that KREEP petrogenesis involved assimilation of large amounts of crustal material. Hence, the petrogenesis of virtually all pristine lunar crustal rocks requires complex processes.

Methods used for modeling trace element evolution in the lunar magma ocean could profitably be tested on a large terrestrial ultramafic intrusion such as the Stillwater. Geological and stratigraphic control, known relationships between samples, and a single well-constrained parent magma would provide an excellent opportunity for testing and refining these methods.

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McKay, G. A.

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AN EXAMPLE OF IGNEOUS LAYERING ON A METEORITE PARENT BODY.

Harry Y. McSween, Jr., Department of Geological Sciences, University of Tennessee, Knoxville, TN 37916.

Igneous layered complexes probably are not restricted to the earth and moon, as these rocks may represent common crust-forming materials on many planetary bodies. Layering phenomena have occurred on at least one meteorite parent body, that of the shergottites (1-3). Reconstructing coeval sequences of magmatic cumulates from meteorites may be even more difficult than from returned lunar samples; however, a newly discovered Antarctic shergottite, Elephant Moraine A79001, will be of considerable benefit in this task. This large (7.94 kg) achondrite contains the only known extraterrestrial example of an igneous (non-brecciated) contact between two lithologies. A79001 has been shock-metamorphosed, resulting in impact-induced phase transformations and the production of veins and pockets of melt glass; however, this report will focus on the igneous layering features.

Petrographic variations in relation to stratigraphy of A79001 are summarized in Figure 1. Lithology A consists of zoned phenocrysts and glomerocrysts of olivine, orthopyroxene, and chromite in a fine-grained, equigranular groundmass of randomly oriented, zoned crystals of pigeonite, augite, maskelynite, and minor ilmenite, chromite, sulfide, whitlockite, and mesostasis. Orthopyroxenes are commonly rimmed by pigeonite of the same composition as in the groundmass. Lithology B is coarser-grained than the groundmass of A and contains no phenocrysts (although the cores of some clinopyroxenes may be primocrysts). It consists of pigeonite, augite, and maskelynite, though in different modal proportions from A, and accessory ilmenite, magnetite?, sulfide, mesostasis, and a silica polymorph. Clinopyroxene and maskelynite in B have core compositions that are similar to those in the groundmass of A, but rims are more strongly zoned to Fe-rich and Na-rich compositions respectively.

The planar contact between lithologies A and B appears sharp in hand specimen, but in thin section, textural changes and modal proportions of pigeonite, augite, and maskelynite are gradational between A and B over a distance of about one cm. The transitional contact argues against either lithology being a xenolith within the other or an intrusive relationship between the two lithologies.

Bulk chemical analyses (4) of lithologies A and B (stars in Figure 1) are shown in Table 1. Lithology B is more highly differentiated than A, as evidenced by its higher Na_2O , K_2O , TiO_2 , P_2O_5 , and lower Cr_2O_3 values relative to A. Petrologic mixing calculations indicate that composition B can be reproduced from A by removal of approximately 28 wt. % orthopyroxene, 16% olivine, and 0.6% chromite, with low residuals (Table 1). The mineral compositions used in this calculation are the same as the average compositions of phenocrysts in A, although the olivine/orthopyroxene ratio in the natural phenocryst assemblage is > 1 . It thus appears possible that lithology B could be formed from a magma of composition A by fractional crystallization of the observed phenocryst assemblage of A. This model is probably too simplistic, however, as phenocrysts of this composition could not have crystallized from A in these proportions. Orthopyroxene phenocrysts are not in equilibrium with the groundmass, as evidenced by the coronas of pigeonite. Also, olivine and orthopyroxene crystals are strongly zoned, and compositions as magnesian as those required in the mixing calculations would be difficult to form during continued crystallization of A. In any case, the lithologies are probably related by a more complex fractionation scheme than suggested by the calculations above.

McSween, H. Y., Jr.

The role of gravitational segregation of crystals in the petrogenesis of this meteorite is difficult to evaluate. Because of the disequilibrium between phenocrysts and groundmass in A, it is possible that some of these large crystals are primocrysts formed from a different liquid and subsequently settled into A or transported in by convection. The mean grain sizes of silicate phenocrysts and glomerocrysts decrease away from the A-B contact, as do the sizes of chromite phenocrysts (Figure 1). This observation suggests that these crystals settled together, but preliminary calculations indicate that these two types of phenocrysts are not hydraulically equivalent. Furthermore, if B is stratigraphically above A, as would be presumed from the fact that B is more highly differentiated, then the larger phenocrysts should have sunk deeper during a given time interval and their sizes should increase downward, the opposite of the observed trend. By analogy with other shergottites (1), the cores of some smaller groundmass clinopyroxenes in both A and B may also be primocrysts. Phenocrysts (except those that settled as glomerocrysts) do not have grain-to-grain contacts. The fine-grained groundmass consisting of zoned crystals and minor glassy mesostasis suggests relatively rapid crystallization, which might preclude efficient crystal settling.

Both lithologies appear to have crystallized primarily from phenocryst-bearing melts rather than formed strictly by accumulation. A possible alternative model may therefore be that A and B crystallized from gravitationally stratified liquids, as proposed by (6). The continuing consortium effort in studying this meteorite will place more constraints on petrogenesis of its constituent layers.

Table 1. Chemical analyses of lithologies A and B by (4). Residuals are the mismatch between B and the calculated composition of B resulting from fractionation of olivine, orthopyroxene, and chromite of the compositions and proportions specified below from lithology A. Calculation was made using the method of (5).

<u>Oxides</u>	<u>Lithology A</u>	<u>Lithology B</u>	<u>Fractionating Phases</u>	<u>Residuals</u>
SiO ₂	48.52	49.03	olivine - 16%	0.01
TiO ₂	0.70	1.23	(Fo ₆₈)	-0.01
Al ₂ O ₃	5.68	9.93		-0.03
Cr ₂ O ₃	0.58	0.14	orthopyroxene - 28%	0.02
FeO*	18.98	17.56	(Wo ₅ En ₆₈ Fs ₂₇)	0.02
MnO	0.52	0.47		-0.02
MgO	16.59	7.32		0.02
CaO	7.10	11.00	chromite - 0.6%	0.11
Na ₂ O	0.84	1.68	(Mg ₂₂ Fe ₈₄ Cr _{1.6} Al _{.29})	-0.09
K ₂ O	0.05	0.09		0.01
S	0.18	0.22		0.06
H ₂ O(+)	0.1	0.1		-
H ₂ O(-)	0.09	0.09		-

*Total Fe as FeO

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LAYERING IN METEORITES

McSween, H. Y., Jr.

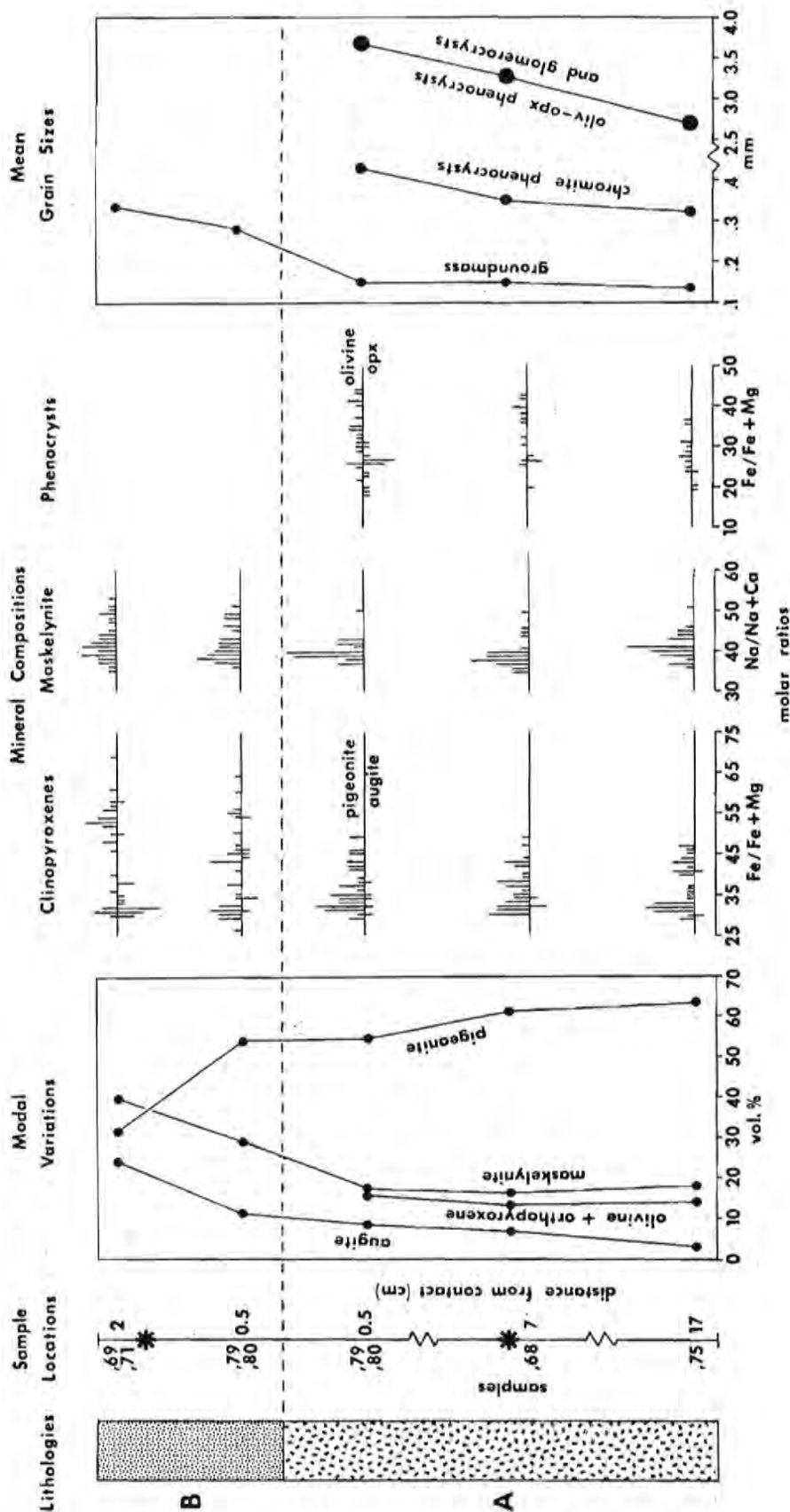


Figure 1. Petrographic variations in different portions of Elephant Moraine A79001 meteorite. Sample locations are approximate distances from the A-B contact; stars represent locations of bulk chemical analyses (Table I). Modal variations were determined from counts of > 1200 points; optically indistinguishable clinopyroxenes were counted together and augite/pigeonite ratios were calculated from microprobe surveys. The smallest vertical bars in mineral composition histograms represent one microprobe analysis. Mean grain sizes for groundmass were calculated from the number of grain contacts encountered during traverses of fixed distance and different orientations; phenocryst sizes are averages of measured diameters in two perpendicular directions.

COOLING HISTORY: CAN WE TELL A MAGMA OCEAN FROM A MAGMA PUDDLE?
 EVIDENCE FROM MICROSTRUCTURES Gordon L. Nord Jr. 959 National Center,
 U.S. Geological Survey, Reston, VA, 22091

In order to adequately define the cooling history of an igneous body one must determine the position of the starting point and a number of subsequent points along its cooling and decompression path in terms of temperature, pressure and fugacities and activities of components. In addition, one must determine the time it takes to advance from one point on the cooling path to another. Some points on the path followed by slowly cooled bodies may be defined by equilibrium crystallization experiments. In more rapidly cooled bodies however increasing the rate of undercooling increases the deviation from equilibrium (1). Each phase and even separate parts of the crystals of each phase (i.e. core, rim) may have different cooling histories.

Points along the cooling path are classically determined by comparing the chemistry of phase assemblages in the rock to equilibrium phase assemblages defined experimentally in the laboratory. When a match is found the rock is said to have equilibrated at such and such a pressure, temperature, f_j and a_j . In reality, such a comparison only measures the temperature, pressure, f_j and a_j where the transport of materials from phase to phase and crystal to crystal is no longer able to keep up with the cooling rate. This most commonly happens at the loss of the melt, vapor or aqueous grain boundary phase through which diffusion is most rapid.

Within the last fifteen years and especially with the impact of the Apollo Program, a great deal of research has concentrated on use of microstructures in minerals as indicators of geological history especially in the absence of any intercrystalline exchange or interaction. What happens in single crystals may in some instances be all that happens during cooling of very dry rocks. In order to gain access to the information locked up in mineral microstructures we must carefully characterize microstructures, structurally and chemically, determine their mechanisms of formation, fields of stability or metastability and most important, their kinetics of formation. The pyroxene and plagioclase systems are particularly appropriate for slowly cooled plutonic rocks.

REACTION RATES AND DIFFUSION

The rate of any thermally activated reaction is described by the Arrhenius equation

$$k = A e^{\frac{-E_a}{RT}}$$

where A is a preexponential term that depends on the reaction mechanism and E_a is the activation energy necessary to overcome the barrier between the initial and final states of the reaction (i.e. breaking bonds). The activation energy for the coarsening reaction of clinopyroxene lamellae in clinopyroxene host is approximately 99 ± 2 Kcal/mole (2,3). The driving force for the coarsening reaction is the reduction of the surface free-energy, accomplished by increasing the size of some lamellae at the expense of others, thereby decreasing the total free-energy of the system. Coarsening is mainly dependent on diffusion and McCallister (2) has suggested that the 99 Kcal/mole activation energy is a measure of the energy barrier associated with diffusion, in this case that of calcium, the largest cation.

Nord, Jr., G.L.

Huebner and Nord (4) have used the 99 Kcal/mole activation energy to show the dependence of diffusion coefficients in pyroxene on temperature (Fig 1) in an attempt to present measured diffusion data on pyroxenes. The dependence of the diffusion coefficient D on temperature is given by the Arrhenius equation

$$D = D_0 e^{\frac{-Q_a}{RT}}$$

where D is the diffusion coefficient, D_0 is a preexponential and Q_a is the activation energy. The equation of the lines in Figure 1 is

$$\text{Log } D = \frac{-Q_a}{2.303RT} (1/T) + \text{Log } D_0$$

The position of the $\text{Fe}/(\text{Fe} + \text{Mg}) = 0$ line is constrained by the self-diffusion of ^{45}Ca (5) and ^{85}Sr (6). A second line $\text{Fe}/(\text{Fe} + \text{Mg}) = \sim 0.3$ intersects a data point determined for Ca-(Mg,Fe) interdiffusion rates in more Fe-rich pyroxene that was incorporated in shock-produced lunar melt rock (7). This fragment-laden pigeonite basalt, 77115, contained augite xenocrysts that reacted with melt to form replacement rims of pigeonite. A comparison of the amount of matter transported to form rims on olivine in the same sample led to the conclusion that the Ca-(Mg,Fe) diffusion rate in these pyroxenes was similar to that of Mg-Fe in the olivine. The values for olivine (8) are also plotted on Fig. 1. The increase in diffusion rate with increase in $\text{Fe}/(\text{Fe} + \text{Mg})$ is also consistent with the order of magnitude increase in coarsening rates between exsolution lamellae in Fe-free clinopyroxene ($\text{Wo}_{25}\text{En}_{75}\text{Fs}_0$) and Fe-rich clinopyroxene ($\text{Wo}_{25}\text{En}_{31}\text{Fs}_{44}$) (9).

A word of caution is necessary at this point. The increase in diffusion coefficient with Fe content is due to an increase in D_0 . The preexponential, D_0 , is essentially a measure of the number of paths and defect sites available for diffusion to proceed (10). Many impurities may act as potential generators of defect sites (i.e. vacancies), particularly multivalent cations such as Fe, Cr, and Mn. Therefore D_0 is also a function of oxygen fugacity and probably a number of other unknowns (11). The $\text{Fe}/(\text{Fe} + \text{Mg}) = \sim 0.3$ line may in fact indicate a maximum value for D_0 obtainable for Ca-(Mg,Fe) diffusion in pyroxene.

The critical temperature for exsolution reactions in pyroxenes decreases as the pyroxene solvus crosses the quadrilateral from the Fe-free join to the Mg-free join. This temperature drop has the opposite effect on reaction rates when compared with that of the increase in the diffusion coefficient with Fe content. We can assess the net effect by plotting the critical temperature, T_c , for the pigeonite eutectoidal reaction (PER), $\text{Pig} + \text{Opx} + \text{Aug}$ on both $\text{Fe}/(\text{Fe} + \text{Mg}) = 0$ and 0.3 lines. D increases more than an order of magnitude between 1300°C (Fe-free eutectoid) and 1150°C ($\text{Fe}/(\text{Fe} + \text{Mg}) = 0.3$ eutectoid) (12). This change illustrates the point of this discussion, that diffusion and therefore the growth of exsolution microstructures increases with the addition of Fe as well as with the addition of impurities. Furthermore, the largest pyroxene exsolution microstructures known to this author are ~ 0.1 mm in width in Moore County pigeonites ($\text{Fe}/(\text{Fe} + \text{Mg}) \approx 0.5$), suggesting that the drop in reaction temperature is more than offset by the increase in diffusion rates.

ESTIMATES OF COOLING RATES FROM EXSOLUTION LAMELLAE WIDTHS

Miyamoto and Takeda (13) have attempted to estimate absolute cooling rates of eucrites from the width of exsolution lamellae of augite in pigeonite. They assumed that growth rate depends only on the calcium diffusion rate, took into

Nord, Jr, G.L.

account the temperature dependence of diffusion and used a lamellae growth coefficient which varies with the changes in lamellae-host compositions during cooling along the solvus limbs. The observed widths of the lamellae could be matched to depths of burial during cooling of an imagined single achondrite body through the temperature range 1200°C - 850°C. Augite lamellae widths of 40-55 μm in Moama and Moore County pigeonites gave cooling rates of 1°C/10⁴ years for depths of 12 km and 9 km respectively. Miyamoto and Takeda used an activation energy for diffusion of 28 Kcal/mole and a D of $4.5 \times 10^{-19} \text{cm}^2/\text{sec}$. This data is plotted in Figure 1 and is several orders of magnitude smaller than the diffusion rates suggested by the synthesis in Figure 1 (4). Other attempts to estimate cooling rates from lamellae widths (14,15) also point out the need for better control over the variables, especially diffusion data and the exact configuration of the solvus.

TYPES OF EXSOLUTION REACTIONS IN PYROXENES

The main types of exsolution reactions in pyroxenes are shown in Fig. 2. Transmission electron microscopy studies have shown that even optically homogeneous pyroxenes usually contain exsolution microstructures and detailed reviews of exsolution mechanisms and textures from both the light optical and electron optical viewpoint can be found in (16). In terms of cooling history, the temperature of the formation of exsolution lamellae in pyroxenes is still not known accurately because of the poor constraints on the positions of the subsolidus pyroxene phase boundaries in P-T-X space. In general TEM studies have uncovered microstructures that provide evidence for particular exsolution and growth mechanisms as well as the presence of metastable phases which form more easily than the equilibrium phases. These metastable phases have been identified in orthopyroxenes and include a transition phase (probably with the structure and composition of pigeonite) and Guinier-Preston (G.P.) zones. G.P. zones are small disks up to several orthopyroxene unit cells in width with a calcium-rich composition but retain the structure of orthopyroxene. Little is known about the kinetics of formation of these metastable phases except that they are only found in metamorphic and plutonic orthopyroxenes.

The inversion of primary pigeonite to orthopyroxene is a particularly complex reaction and can go by three possible mechanisms only one of which is capable of retaining a two dimensional relationship between parent pigeonite and product orthopyroxene. These reactions are discussed in Buseck et al. (17) but no kinetic data is available. Kinetic data would be particularly useful since the inversion of pigeonite to orthopyroxene is widespread, occurring even in the exsolved pigeonite lamellae in augites.

ESTIMATES OF NUCLEATION TEMPERATURES IN CLINOPYROXENE

Recently a very useful geothermometer has been developed for determining the nucleation temperature of clinopyroxene lamellae in a clinopyroxene host (18,19,20). The thermometer is based on minimizing the strain between the host and the nucleus by exactly fitting the two structures, which have slightly different lattice parameters due to the difference in composition, along a plane of best dimensional fit. Because the lattice parameters of pigeonite change more rapidly with falling temperature than those of augite the plane of best fit also changes with falling temperature.

Robinson et al. (20) described five different sets of lamellae in a Bushveld augite; opx on (100), pigeonite on "100" and three sets of pigeonites

Nord, Jr, G.L.

on "001" (the symbol "hkl" denotes, approximately parallel to hkl). The nucleation temperatures for the lamellae were suggested to be about 1000°C, 850°C, 800°C and 600°C. Using the transmission electron microscope, Champness and Lorimer (21) found two more sets of pigeonite with suggested nucleation temperatures of 700°C and 400°C. In one sample therefore, a fairly complete record of nucleation temperatures is retained. If the nucleation data can be combined with growth models similar to that proposed by Miyamoto and Takeda (13) a complete cooling history could be obtained from a single clinopyroxene crystal.

Champness and Lorimer (21) expanded the Robinson model by observing that "001" (or "100") nucleation is favored over "100" (or "001") at a particular temperature. They have shown this by calculating the strain that builds up in the (010) plane of the pigeonite lamellae as the temperature decreased. Growth will be favored on orientations with the least strain build-up. Their results indicate that growth on "001" is favored between 1050°C - 950°C, on "100" at about 900°C, on "001" between 800°C - 750°C and equally on "001" and "100" below 700°C. At temperatures below 700°C, however, they suggest that "100" is favored because the $P2_1/c$ pigeonite structure is stable and would fit more easily in $C2/c$ augite with the lamellar interface parallel to the Si-O chains.

EXSOLUTION RATES IN CALCIC PLAGIOCLASE

The exsolution mechanism in slowly cooled plutonic pyroxenes is that of nucleation and growth whereas the exsolution mechanism in plagioclases from similar environments is that of spinodal decomposition and coarsening. Nucleation processes in plagioclase appear to be kinetically very difficult probably due to the very low diffusion rates of the couple $NaSi \leftrightarrow CaAl$. Grove (22) studied labradorite - bytownite plagioclases from volcanic, plutonic and metamorphic environments and compared the observed exsolution microstructures with calculated cooling rates for the samples. The cooling rates ranged from $5 \times 10^{-3} \text{ } ^\circ\text{C/yr.}$ to $5 \times 10^{-6} \text{ } ^\circ\text{C/yr.}$ and the microstructures ranged from poorly developed 5-10 nm wide lamellae for the rapidly cooled rocks to 200-500 nm wide lamellae for deep crustal plutonics and metamorphic rocks. If coarsening rates could be determined experimentally, exsolution microstructures in bytownites would be a useful cooling indicator.

CONCLUSION

Can we tell a magma ocean from a magma puddle from cooling history? Compare the cooling history of a magma puddle, 20 km deep, constrained to crystallize near the planetary surface, and that of a magma ocean 200 km deep. One should be able to place a representative sample suite from the ocean or puddle into a sequence of decreasing cooling rate and thus increasing depth, utilizing the many geothermometers available including mineral microstructures. Unfortunately, we can not yet place absolute cooling rates on the members of the series and consequently calculate a depth of formation from heat flow models. This leaves us with the capability to estimate relative differences in size (depth) of magma bodies if the differences are large, but only limited capability to characterize the bodies as oceans or puddles. The key to this distinction is the quantification of the kinetics of reaction and diffusion rates involved the mineral systems found in plutonic rocks.

Nord, Jr, G.L.

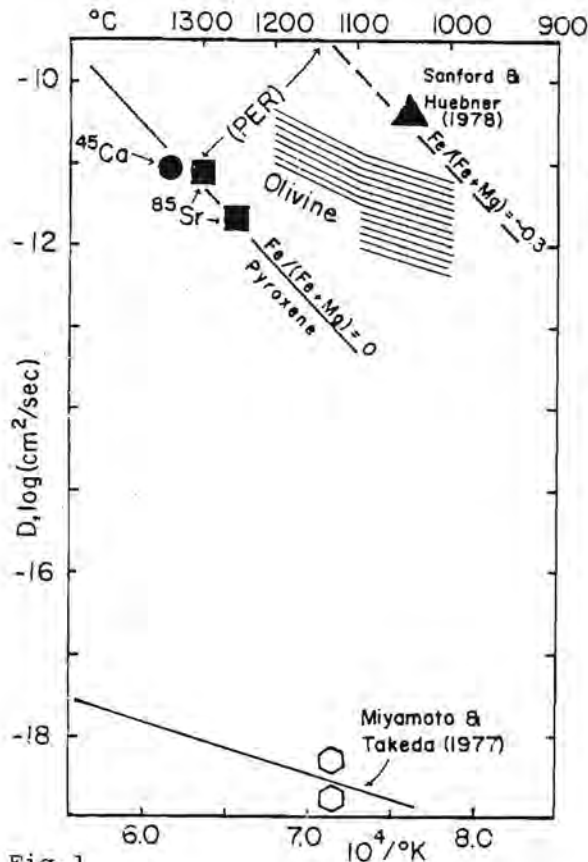


Fig.1

(Figure 1 after Fig.6 of Huebner and Nord, 1981)

FIGURE 2 SUBSOLIDUS REACTIONS IN PYROXENES

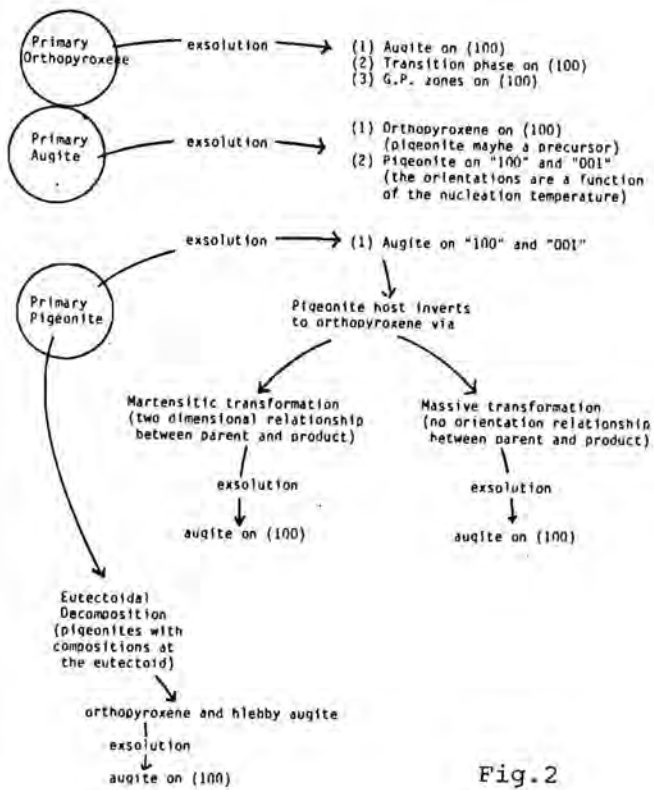


Fig.2

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RADIOMETRIC AGES AND ISOTOPIC SYSTEMATICS OF PRISTINE PLUTONIC LUNAR ROCKS. L. E. Nyquist, NASA Johnson Space Center, Houston, TX 77058

The primary issues to be addressed via radiometric ages and isotopic data for pristine plutonic lunar rocks (PPLR's) include: (i) Do some PPLR's date from the earliest stages of the global lunar differentiation? (ii) If so, can the time interval required for global differentiation be determined? (iii) Can the oldest rocks give unique information about the processes of global differentiation? (iv) Do some PPLR's date from later, presumably local, magmatic processes? (v) Do such "young" rocks yield unique information about magmatic processes occurring in the lunar crust during the first ~ 500 m.y?

Resolution of these primary issues is hindered by a number of secondary factors including: (i) The violent meteorite bombardment history of the Moon has affected the composition and ages of most crustal rocks. (ii) Meteorite bombardment may have affected ages of compositionally pristine rocks so that secondary events may be mistakenly identified as crystallization ages. (iii) Some or all radiometric systems may have remained open for significant periods following crystallization due to subsolidus isotopic equilibration occurring at considerable depth in an initially hot lunar crust. (iv) The required isotopic measurements are analytically difficult.

Radiometric ages are summarized in Tables 1-4. When multiple values are available usually only one has been chosen. Readers are referred to extensive summaries of early work in references (1-5) and to the abstracts and proceedings of the Lunar and Planetary Conferences for additional data. The ages in Tables 1-4 are based on decay constants recommended by Steiger and Jäger (6).

Fig. 1 is a histogram of ages >3.9 AE for PPLR's as determined by the Ar-Ar, Rb-Sr, and Sm-Nd methods. The PPLR's constitute a drastically reduced sample of the total population of returned highland samples (4,5,7,8). When multiple "ages" are exhibited in an Ar-Ar age spectrum, the "high temperature" age has been selected as most representative of the crystallization age. The Ar-Ar data clearly show the ~ 3.9 AE age peak which is attributed to intense bombardment of the lunar surface accompanied by formation of several lunar basins at about this time. There is only a hint of the ~ 4.2 AE age peak which is considerably more prominent among non-pristine rocks. Four samples show ages of 4.3-4.4AE. Assuming that these ages are not artifacts, some anorthosites (A) and norites (N) are indicated to have formed very early in lunar history. The ~ 4.2 AE ages are for two troctolites (T). It is unclear from the Ar-Ar data whether the ~ 4.2 AE ages are to be interpreted as crystallization ages or as somehow related to the lunar bombardment history.

The ~ 3.9 AE age peak does not appear in the Rb-Sr and Sm-Nd data. The two anorthosites with ~ 3.9 AE Ar-Ar ages have no Rb-Sr or Sm-Nd isochron ages. However, norite samples 15455 and 72255 with Ar-Ar ages ~ 3.9 AE have older Rb-Sr isochron ages. Rb/Sr isotopic systems for all PPLR's show some disturbances. Compston et al. (9) suggest that the crystallization age of 72255 could be older than the nominal value obtained from the mineral isochron. The Sr isotopic data for 15455 (10) shows comparatively minor deviations outside of analytical uncertainty from the mineral isochron. This characteristic is also true of the Sr-isotopic data of troctolite 76535 (11) and dunite 72417 (12). The Rb-Sr data for norites 78236 (13) and 77215 (14) show more scatter. The ages reported here rely on data selection criteria for which the reader is referred to the original work.

Apparently undisturbed Sm/Nd mineral isochrons have been reported for 67667 (15), 73255 (16) and 77215 (14). Isochrons for which one or more analyses depart from the best fit line by more than analytical uncertainty have been reported for troctolite 76535 (11), norite 78236 (13,16) and norite 15455 (10). Carlson and Lugmair (16) preferred an age of ~ 4.34 AE for 78236 including all

Nyquist, L. E.

Sm/Nd data; Nyquist et al. (13) argue for an age of ~ 4.43 AE based on concordant Rb-Sr and Sm-Nd ages obtained for one pyroxene analysis and the Ar-Ar age spectrum. The ~ 4.43 AE age is preferred here and is shown in solid outline in Fig. 1, whereas the ~ 4.34 AE is shown in dotted outline. The Sm/Nd age reported in Table 3 and plotted in Fig. 1 for 15455 is based on pyroxene and whole rock analyses only (13).

The interpretation of even Sm/Nd ages of PPLR's is ambiguous. Recently Carlson and Lugmair (16) emphasized the view that the 4.23 ± 0.05 AE obtained for a norite clast from 73255 represented a crystallization age and implied lunar crustal formation over an ~ 350 m.y. time period. This interpretation can be questioned, particularly in light of several ~ 4.2 AE ages obtained by the Ar-Ar method for non-pristine clasts from the same impact breccia (17). Carlson et al. (15), however, also consider the possibility that the ~ 4.2 AE age of 67667 could be due to sustained subsolidus equilibration of isotopic systems deep in the lunar crust. A similar scenario was previously put forward by several authors (11,18,19) to explain the isotopic data on troctolite 76535. Most recently Caffee et al. (20) have reported Pu-Xe ages of 4.26AE for plagioclase and 4.49AE for olivine from 76535 lending further support to the concept that some isotopic and mineral systems in this rock remained open for a significant time after its crystallization. Nyquist et al. (13) have suggested that some of the disturbances observed in all isotopic systems of norite 78236 is due to formation and slow cooling deep in the lunar crust. They have attempted to quantify this suggestion by considering the lunar crustal thermal regime and trace element diffusion rates.

There is a general paucity of PPLR ages determined by other methods of radiometric dating (Table 4). However, two observations from U-Pb studies are of particular note. The first is that analyses of lunar crustal rocks (not necessarily pristine) are aligned on a "cataclysm isochron" on the U-Pb concordia diagram (21). The lower intersection of this isochron with the concordia curve occurs at ~ 3.9 AE and has been interpreted as reflecting a major episode of meteoritic bombardment of the lunar surface at that time. The upper intersection with the concordia curve occurs at ~ 4.42 - 4.47 AE and has been interpreted as reflecting global differentiation of the Moon at that time (21,22). Another noteworthy observation is that among the pristine rocks the U-Pb data of 77215 lie within error of the concordia curve at ~ 4.4 AE and the $^{207}\text{Pb}/^{206}\text{Pb}$ age of this rock is 4.42AE (23) concordant with Rb-Sr and Sm-Nd ages (14).

Sr and Nd isotopic evolution in PPLR's is illustrated in Figs. 2 and 3 using data from Tables 2 and 3. The measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of anorthosite 60025 is $\leq \text{BABI} = 0.69898$ (27). Because of its very low Rb/Sr ratio, $^{87}\text{Sr}/^{86}\text{Sr}$ in this rock would have changed by only 2 parts in 70,000 over 4.5AE. The very low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of lunar anorthosites has led to their widespread acceptance as early lunar differentiates in spite of "young" Ar-Ar ages of 3.6-4.2AE. Sr-evolution lines for other cumulate PPLR's intersect those of the anorthosites at ~ 4.3 - 4.6 AE. Mineral isochron ages, $T(\text{Sr})$, and initial $^{87}\text{Sr}/^{86}\text{Sr}$, $I(\text{Sr})$, for 72417, 76535, and 15455 cluster within analytical uncertainty at $T(\text{Sr}) = 4.45$ - 4.51 AE, $I(\text{Sr}) = 0.69895$ - 0.69900 . The evolution lines for the cumulate PPLR's (all PPLR's except 72255 and 77215) cluster most tightly at ~ 4.45 AE. The slopes of the evolution lines are proportional to the Rb/Sr ratios of the bulk rocks, which unlike the mineral isochron ages, should be insensitive to subsolidus isotopic equilibration. There is thus a strong suggestion that the true crystallization ages of the cumulate PPLR's is ~ 4.45 AE in spite of some younger mineral isochron ages.

Two hypothetical growth curves are shown in Figure 1. One is for a bulk Moon $^{87}\text{Rb}/^{86}\text{Sr} = 0.05$ (24,28,29). From the intersection of this growth curve with that of individual samples one can calculate model ages $T(\text{Sr}, \text{UR})$ where

Nyquist, L. E.

UR = "uniform reservoir" (30). These model ages are analogous to Sm/Nd ICE ages (25) and are well defined for plagioclase-rich cumulates of low Rb/Sr. They are dependent on the assumed lunar age. $T(\text{Sr}, \text{UR})$ of the cumulate PPLR's except 72417 are ~ 4.45 – 4.55 AE for a lunar age = 4.55AE.

The second hypothetical growth curve in Figure 2 is for $^{87}\text{Rb}/^{86}\text{Sr} = 0.4$ as calculated for the parental magma of cumulate norite 78236 (13). $^{87}\text{Sr}/^{86}\text{Sr}$ evolves very rapidly for this growth curve. There is no evidence for unusually evolved $^{87}\text{Sr}/^{86}\text{Sr}$ in the isotopic data of 78236 suggesting that the parental magma phase was a transient phenomenon in the petrogenesis of this rock. This conclusion seems inconsistent with identification of the parental magma with a late stage in evolution of the lunar magma ocean.

Nd isotopic evolution is shown in ϵ -representation in Figure 3. There is a pronounced tendency to positive ϵ -values. The 77215 datum is so aberrant as to suggest an uncompensated interlaboratory bias. Unpublished JSC data for most clasts of norite and spinel troctolite from breccia 15445 tend to have $\epsilon < 0$ for $T(\text{Nd}) < 4.4$ AE. These latter data are preliminary pending verification that the neutron fluence correction to $^{147}\text{Sm}/^{144}\text{Nd}$ values is the same as for 15455.

Positive ϵ -values of the required magnitude could be produced in the available time in source regions with Sm/Nd ratios similar to those in mare basalt source regions. This observation suggests a model in which the parental magmas of the high Mg suite of PPLR's are formed by melting of early formed mafic cumulates. These magmas could have been intruded into the crust, perhaps assimilated crustal material, and differentiated to form layered plutons. This scenario is similar to that suggested by James (31), differing primarily in that a previously differentiated rather than primitive mantle is the melt source. The Sr isotopic constraints could be satisfied by such a model since mare basalt sources appear to have had $^{87}\text{Rb}/^{86}\text{Sr} \sim 0.01$, similar to that of the cumulate PPLR's.

Figure 4 shows combined Sr and Nd isotopic evolution. The slope and length of a vector in this diagram is proportional to $(\text{Rb}/\text{Sr})/(\text{Sm}/\text{Nd})$ and to time, respectively. Points at 4.2 and 4.4AE ago are calculated from bulk rock isotopic measurements. This calculation assumes all the rocks are at least 4.2AE old. The objective of the exercise is to see whether the Sr and Nd isotopic systematics are consistent with even older ages. For the cumulate PPLR's (all excluding 77215) the points at 4.2 and 4.4AE define vectors which if extended further back in time (not shown) pass nearly within analytical uncertainty of the lunar initial Sr and Nd isotopic composition suggestive of ages ≥ 4.4 AE. However, the time required to reach the bulk Moon values from the rock data at 4.4AE is greater than 0.2AE if the rock growth rate of $^{143}\text{Nd}/^{144}\text{Nd}$ is assumed. A 4.6AE lunar age is assumed primarily for clarity of illustration. A similar argument holds if the bulk Moon initial isotopic values are assumed at 4.55AE (dotted line). This observation suggests that precursors existed for the cumulate norites in which the growth rate of $^{87}\text{Sr}/^{86}\text{Sr}$ was similar to that in the rocks themselves, but the growth rate of $^{143}\text{Nd}/^{144}\text{Nd}$ was slightly greater than the rate in the rocks. The hypothetical bulk Moon growth rates which Nyquist et al. (28) have assumed for their models of mare basalt petrogenesis are shown for comparison to the PPLR data. The hypothetical bulk Moon values at 4.4AE ago differ from those of the PPLR's. This difference can be easily accommodated if the parental magmas of the high-Mg PPLR's were derived by melting cumulate sources of lower Rb/Sr and higher Sm/Nd than the bulk Moon.

In summary, two alternate scenarios for the petrogenesis of the cumulate high-Mg PPLR's are permitted by the isotopic data: (i) The rocks formed very early in lunar history, possibly as crystal cumulates from the magma ocean. Some "young" ages are due to sub-solidus isotopic equilibration and apparently

Nyquist, L. E.

positive $\epsilon(\text{Nd})$ values are due to analytical uncertainties. (ii) At least some of the rocks formed as crystal cumulates from parental magmas formed by remelting the earliest formed mantle cumulates. Positive $\epsilon(\text{Nd})$ values are real and some "young" ($\sim 4.2\text{AE}$) ages are real crystallization ages. A third possibility, including young ages and primitive sources such as suggested by James et al. (31) can apparently only be accommodated by the isotopic data if the PPLR parental magmas assimilated much anorthositic crustal material of low $^{87}\text{Sr}/^{86}\text{Sr}$. The calculated trace element abundance pattern of, for example, the 78236 parental magma (32,33) speaks strongly against this possibility. The author favors (ii) above because the high calculated Rb/Sr ratio of the 78236 parental magma is more consistent with a transient melting episode than with prolonged differentiation in the magma ocean.

TABLE 1. Ar-Ar AGES OF PRISTINE PLUTONIC LUNAR ROCKS

SAMPLE	GROUP	TYPE	AGE(AE)	REF.
15362	FA	Anor	>3.92	(34)
15415	FA	Anor	3.89± .07	(35)
15455c	FA	An Nor	>3.76	(34)
			3.82-4.15	(13)
60015	FA	Anor	3.45± .05	(36)
60025	FA	Anor	4.13± .06	(36)
67075	FA?	Nor An?	3.98± .05	(37)
67435Ac	FA?	Anor?	4.36± .05	(38)
67435Dc	FA?	Anor?	4.34± .05	(38)
72255c	HM	Nor	3.93± .03	(39)
76535	HM	Troct.	4.19± .02	(40)
77215c	HM?	Nor. Br.	3.98± .04	(41)
77115c	HM?	Troct.	4.17± .03	(42)
78236	HM	Anor Nor	~ 4.4	(13)

Abbreviations: C = clast, FA = ferroan anorthosite, HM = high-Mg suite, Anor = anorthosite, An Nor = anorthositic norite, Nor An = noritic anorthosite, Nor = norite, Troct = troctolite, Nor Br = noritic breccia, AE = 10^9 yrs.

TABLE 2. Rb-Sr AGES AND Sr-ISOTOPIC DATA FOR PRISTINE PLUTONIC LUNAR ROCKS

SAMPLE	$\frac{^{87}\text{Rb}}{^{86}\text{Sr}}$	$\frac{^{87}\text{Sr}}{^{86}\text{Sr}}$	T(Sr) (AE)	I(Sr)	REF.
15415	0.0024	0.69914±5	(3.89)	(0.69900)	(43)
15455c	0.0239	0.70054±4	4.48±.12	0.69894±6	(13)
60025	0.0003	0.69896±3	(4.13)	(0.69894)	(44)
61016c	0.0006	0.69800±3			(45)
67667	0.0112	0.69979±5	(4.18)	(0.69905)	(46)
72255c	0.0797	0.70389	4.08±.05	0.69913±7	(9)
72417	0.05	0.70226	4.45±.10	0.69900±7	(12)
76535	0.0058	0.69837±5	4.51±.07	0.69900±7	(11)
77215	0.0777	0.70397±3	4.33±.04	0.69901±7	(14)
78236	0.0240	0.70051±4	4.29±.02	0.69901±2	(13)

Notes: I(Sr) = initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio. Values of T(Sr) and I(Sr) in () are based on Ar-Ar or Sm-Nd ages. $\lambda(^{87}\text{Rb}) = 0.0142 \text{AE}^{-1}$. Where possible, data from a single laboratory have been selected to maintain a maximum degree of self-consistency. Post Rb-Sr data is from the CIT laboratory; the data of other workers have been bias-corrected when necessary.

TABLE 3. Sm-Nd AGES AND Nd-ISOTOPIC DATA FOR PRISTINE PLUTONIC LUNAR ROCKS

SAMPLE	$\frac{^{147}\text{Sm}}{^{144}\text{Nd}}$	$\frac{^{143}\text{Nd}}{^{144}\text{Nd}}$	T(Nd) (SE)	I(Nd)	E(Nd)	REF.
15455c	0.1673	0.51101±2	(4.48)	(0.50604)	(1.0±0.8)	(10, 46)
67667	0.1870	0.51163	4.18±.07	0.50656±8	0.8±0.3	(15)
73255c	0.2380	0.51303±4	4.23±.05	0.50637±7	0.5±0.4	(16)
77215	0.1780	0.51123±7	4.37±.07	0.50604±11	-1.9±0.5	(14)
78236	0.1724	0.51119±3	4.43±.05	0.50609±7	0.8±0.4	(13)
			4.34±.04	0.50626±6	1.1±0.7	(16)

Notes: I(Nd) = initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio, $\epsilon(\text{Nd})$ = deviation from chondritic evolution in parts in 10^4 (25). Age in () is Rb-Sr age. $\lambda(^{147}\text{Sm}) = 0.00654 \text{AE}^{-1}$. Sm/Nd data are normalized to $^{148}\text{Nd}/^{144}\text{Nd} = 0.24308$. UCSD data have been further corrected for the instrumental bias reported by Lugmair and Carlson (50). Thus, all UCSD literature data are multiplied by 0.99831 and all USGS data by 0.99850. Values of $\epsilon(\text{Nd})$ are usually quoted as reported in the original work to avoid affects due to interlaboratory bias. However, data for 77215 and the Juvinas reference (14) do not appear self consistent so in this case $\epsilon(\text{Nd})$ has been calculated relative to present-day $^{147}\text{Sm}/^{144}\text{Nd} = 0.1953$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.511798$ as is the JSC convention (24). Two other sets of self-consistent parameters in use are $^{147}\text{Sm}/^{144}\text{Nd} = 0.1936$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.511769$ ((25) renormalized) and $^{147}\text{Sm}/^{144}\text{Nd} = 0.1967$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.511836$ (26). These values are currently used at the La Jolla and CIT laboratories respectively. For T = 4.3AE, the JSC and CIT choices yielded ϵ values differing by only 0.03 whereas the UCSD parameters yield values lower by 0.4.

TABLE 4
U-Pb, Pb-Pb, and Pu-Xe AGES FOR
PRISTINE PLUTONIC LUNAR ROCKS

SAMPLE	U-Pb (AE)	Pb-Pb (AE)	Pu-Xe (AE)	REF.
15415	~ 4.0	-	~ 2.9	(47)
15455c	-	-	-	(20)
60015	-	~ 3.7	-	(48)
76535	~ 4.0	-	4.26-4.49	(21, 20)
77215	~ 4.55	4.27	-	(21, 49)
	-	4.42	-	(23)

Nyquist, L. E.

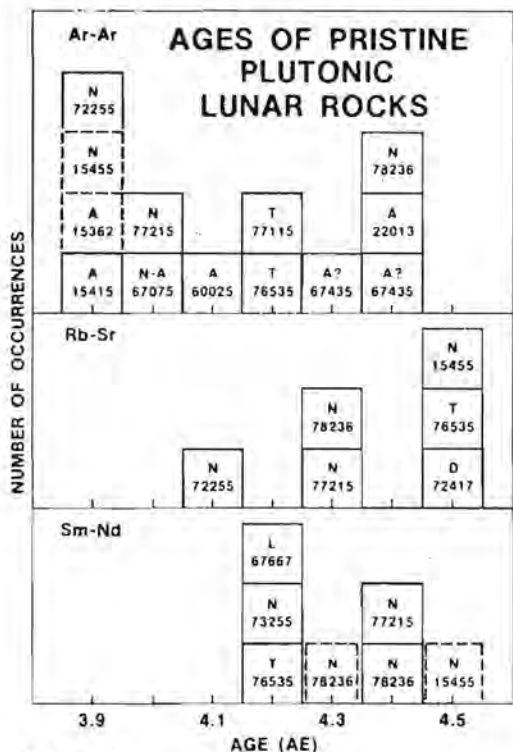


Fig. 1

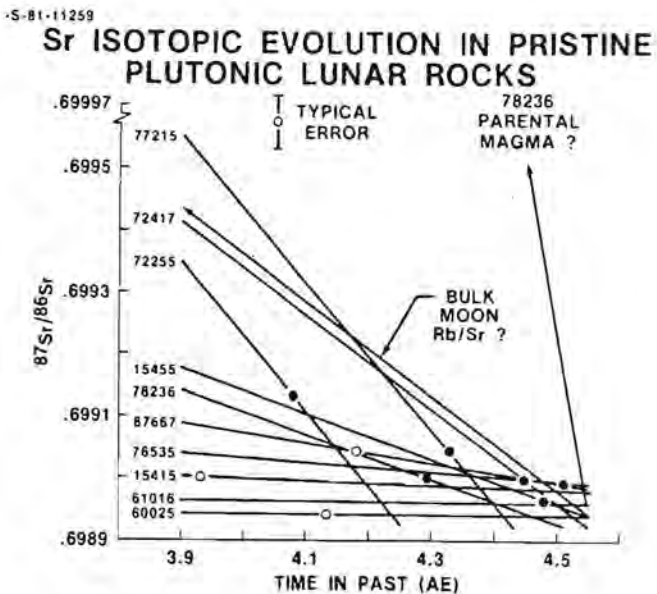


Fig. 2

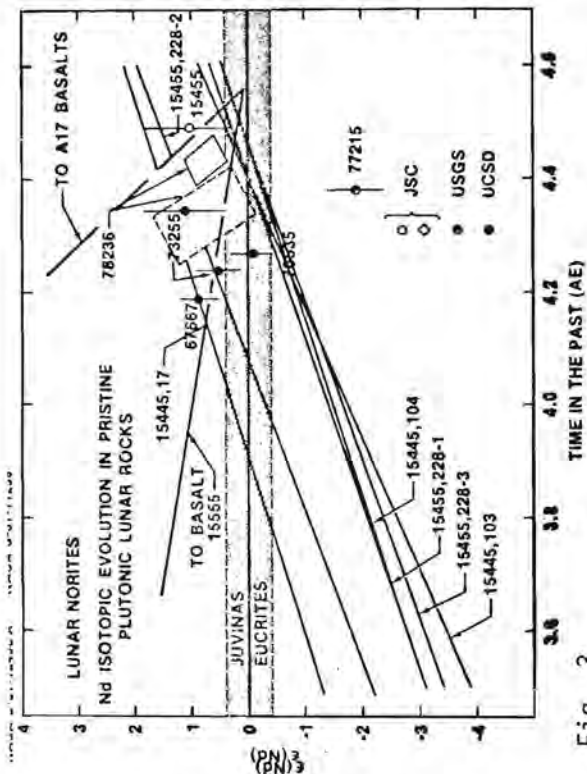


Fig. 3

Sr and Nd ISOTOPIC EVOLUTION IN PRISTINE PLUTONIC LUNAR ROCKS

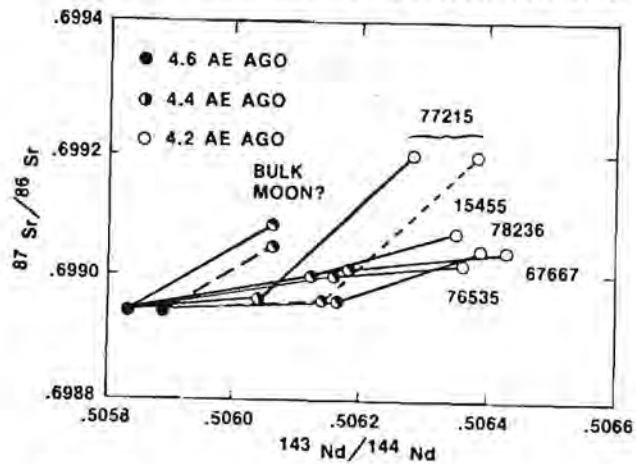


Fig. 4

Nyquist L. E.

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Nyquist L. E.

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PETROGENESIS OF ARCHEAN ANORTHOSITES. W. C. Phinney, SN6/NASA Johnson Space Center, Houston, TX 77058

Archean anorthosite complexes are widespread. They occur in North America, South America, Greenland, Southern Africa, India, Russia and Europe. They display various degrees of metamorphism and states of deformation. All are associated with volcanic units some of which are greenstones while others are higher grade equivalents of volcanic belts which in some cases can be traced into lower grade greenstone belts. There are two striking features of Archean anorthosites: Their texture and their composition. The texture consists of large (up to 20 or 30 cm), equidimensional, euhedral to subhedral plagioclase crystals rather than the lathy to platy plagioclases of most mafic rocks. The composition consists of highly calcic plagioclase usually of An_{80} to An_{90} (rarely up to An_{95}) in marked contrast to the younger Proterozoic complexes of generally An_{40-60} . The euhedral textures and unusually high modal plagioclase content of most complexes indicate that high-Ca plagioclase was the only or principal liquidus phase during most of the crystallization history. The high An contents and the high plagioclase modes lead to very high CaO and Al_2O_3 contents and very low REE abundances with positive Eu anomalies for the bulk compositions of the complexes. The major elements do not concur with any known volcanic compositions and the REE are most compatible with komatiitic melts which should not produce a liquidus phase of plagioclase.

An important clue about the petrogenesis of Archean anorthosites is found in the widespread occurrence in Archean flows, sills and dikes of plagioclase megacrysts of the same texture and composition as in the complexes. Green (1) noted over 30 occurrences of these megacrysts in Canadian greenstone belts. The megacrysts may occur in various proportions ranging from a few crystals through clusters making up 40 to 50% of a unit, to over 80%, as in the case of the football anorthosite sill at Pipestone Lake, Manitoba. Green showed that the megacrysts occur in the upper parts of tholeiitic volcanic sequences and calculated the proportions of mafic phases that should precipitate if one assumes that the sequence of flows represents various stages of fractionation from his most primitive early tholeiites. He found that between 2 and 3 times as much mafic material should be produced as plagioclase. Thus the widespread occurrence of the plagioclase megacrysts suggests not only that they were separated from mafic minerals prior to the emplacement of the flows and sills but also that substantial volumes of complementary mafic separates are not found in association with the volcanics. Another problem is whether plagioclase of up to An_{90} , or even higher, could exist in equilibrium with the flows and sills as a liquidus phase. Comparison of plagioclase compositions and their presumed equilibrium melts indicate that the megacrysts are of higher An content than plagioclases that are in equilibrium with the flows and sills. A similar disequilibrium exists with high Ca megacrysts in recent oceanic tholeiites.

An_{80} to An_{90} is known to be the composition of plagioclase that forms on the cotectic after olivine and pyroxene have separated from mafic or komatiitic melts (Table 1). Such plagioclase is present above the ultramafic zones of the Stillwater, the Bushveld, the Muskox and sills in the Barberton Mountains (2). These occurrences of highly calcic plagioclase combined with the locations of megacrysts in flows suggest a fractionation sequence in which plagioclase separates at some intermediate stage of differentiation. REE patterns for komatiitic and tholeiitic Archean volcanics provide additional evidence for substantial plagioclase segregation. Similar patterns exist in all areas of Archean volcanism; namely the more fractionated melts develop a negative Eu anomaly (Fig. 1). If the komatiitic to Mg-rich melts with no Eu anomalies are assumed as the parent for a fractionating melt then plagioclase separation

W. C. Phinney

could well be the cause of the Eu anomaly. From analyses of a series of flows near the Bad Vermilion Lake anorthosite complex we find very significant negative Eu anomalies in the more fractionated rocks (Fig. 2), again in agreement with substantial plagioclase separation. The data strongly support fractionation of melts in which plagioclase crystallized as well as earlier and simultaneous complementary mafic minerals. The plagioclase occurs commonly as anorthosite but sufficient volumes of complementary ultramafic units are not found in the exposed rocks or subsurface geophysical data. Thus the required volumes of complementary ultramafic rocks must be very deep.

If we compare the densities of melts rising from the mantle, granitic crustal materials, and plagioclase we find that the rising melts in the mantle reach a crust of similar or even lower density, thereby slowing and ponding the melts at the base of the crust (Table 2, Fig. 3A and B). Higher thermal gradients in the Archean would enhance this effect by producing even hotter crusts with lower densities. This assumes that a granitic to tonalitic crust existed before volcanism commenced as appears to be the case in Rhodesia (3). Similar ponding of melts at the crust-mantle boundary has been proposed by others (4).

As the ponded melt cooled it crystallized mafic minerals which formed ultramafic units at the upper mantle boundary (Fig. 3C). If the komatiitic or ultramafic melt, which the Viljoens (5) suggested was widespread under the Archean crust of Southern Africa, was the parent then substantial volumes of ultramafic cumulates could have been added to the upper mantle. Throughout this process, heating of the lower crust should have occurred accompanied by doming of the crust, introduction of volatiles, and metamorphism including partial melting. As fractionation of the melt proceeded it concentrated Al and Ca until high An plagioclase formed as a cotectic phase. Because the plagioclase is similar in density to the melts it may have floated or been kept in suspension (Fig. 3D). Because of the long cooling times for a melt at this depth, substantial volumes of large plagioclase crystals with nearly constant composition could have grown. These crystals would have been carried upwards in any rising melts to produce: flows with megacrysts, sills and dikes with various ratios of megacrysts to melt, and complexes of anorthosite formed by crystals that remained in chambers while much of the melt continued to the surface.

At various stages of fractionation, melts may be tapped by fractures formed in the doming or deforming crust and become sills, dikes, or flows. In the early stages of such fracturing and migration of melt into the crust much heat would be transferred from melt to crust along the rock-melt boundaries thus hindering the passage of early melts to the surface as described in detail by Marsh (6). The fracture systems could have been associated with rift zones formed above upwelling mantle convection cells. As more melts migrated to the surface or into the crust the density of the crust increased thereby allowing melt to rise more readily through the less stable density configuration. Thus the earliest melts would have been the most likely to be ponded at the crust-mantle boundary. In addition the solidification of basaltic rocks with densities of 2.8 to 2.95 above lower density granitic materials would cause the unstable condition necessary for formation of the diapiric gneiss domes that are common in these terrains.

The model suggests anorthosite as a significant unit of the lower crust. This is compatible with known low heat production, seismic velocity data (7), the change in REE patterns of sediments with time (8), and the change in $^{87}\text{Sr}/^{86}\text{Sr}$ in sea water with time (9). One important result of the formation of anorthosite as a major lower crustal unit is the development of a refractory underplating of parts of the crust. This could help initiate long term

W. C. Phinney

stabilization of the continental crusts.

The model outlined above does not appear to explain the origin of calc-alkaline volcanics that may occur late in some volcanic cycles. Their REE patterns are very steep with strong depletion of heavy REE and seem to require garnet fractionation somewhere in their history. The Andesitic volcanics can be added to the above model if they rise through the mantle from a different source or if they are superimposed tectonically from Island Arc environments during closing of an ocean in a Wilson cycle (10).

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TABLE 1

EARLY PLAGIOCLASE CUMULATE COMPOSITIONS

(tabular crystals)

Bushveld Complex	An ₈₀₋₈₅	First cumulate plagioclase after ultramafic zone (Fo ₉₉ , En ₈₉)
Stillwater Complex	An ₈₆	" " " " " " (Fo ₉₀ , En ₈₇)
Muskox Intrusion	An ₈₅	" " " " " "
Gorgona Komatiite	An ₈₃	Fine-grained plagioclase in volcanic flow (Fo ₈₈₋₉₁)
Ultramafic sills of Barberton Mtn. Land	An ₉₀₋₉₅	First cumulate plagioclase after ultramafic zone (Fo ₉₀)

TABLE 2

UNCOMPRESSED DENSITIES

	Crustal Room Temp	Rocks 500°C	Melts	Plagioclase
Granite	2.67	~2.59	basalts at 1250°C 2.64 - 2.75	An ₈₀ at 1200°C 2.67
Tonalite	2.76	~2.68	Komatiite at 1600°C 2.73 - 2.84	
Quartz Diorite	2.81	~2.73		
Gabbro	2.98			

W. C. Phinney

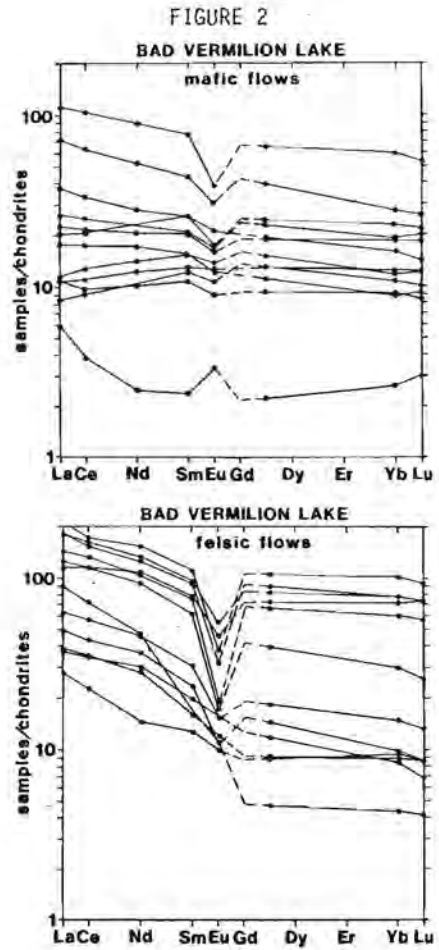
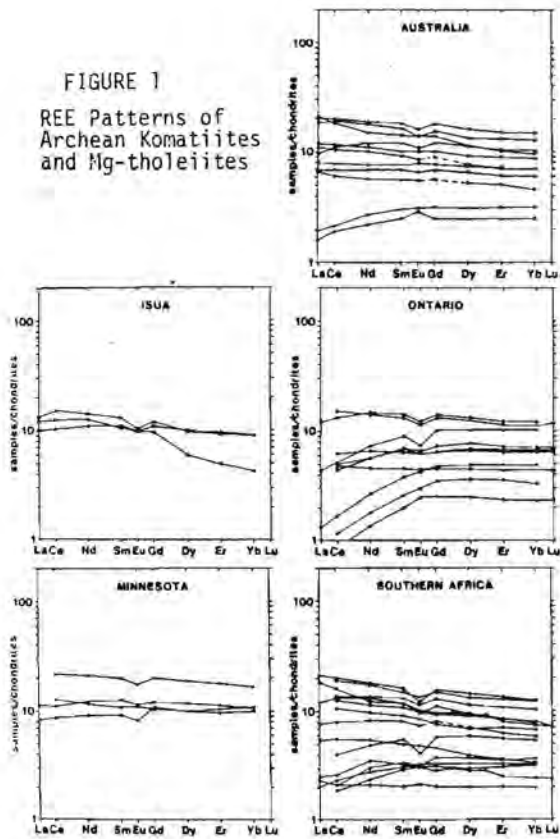
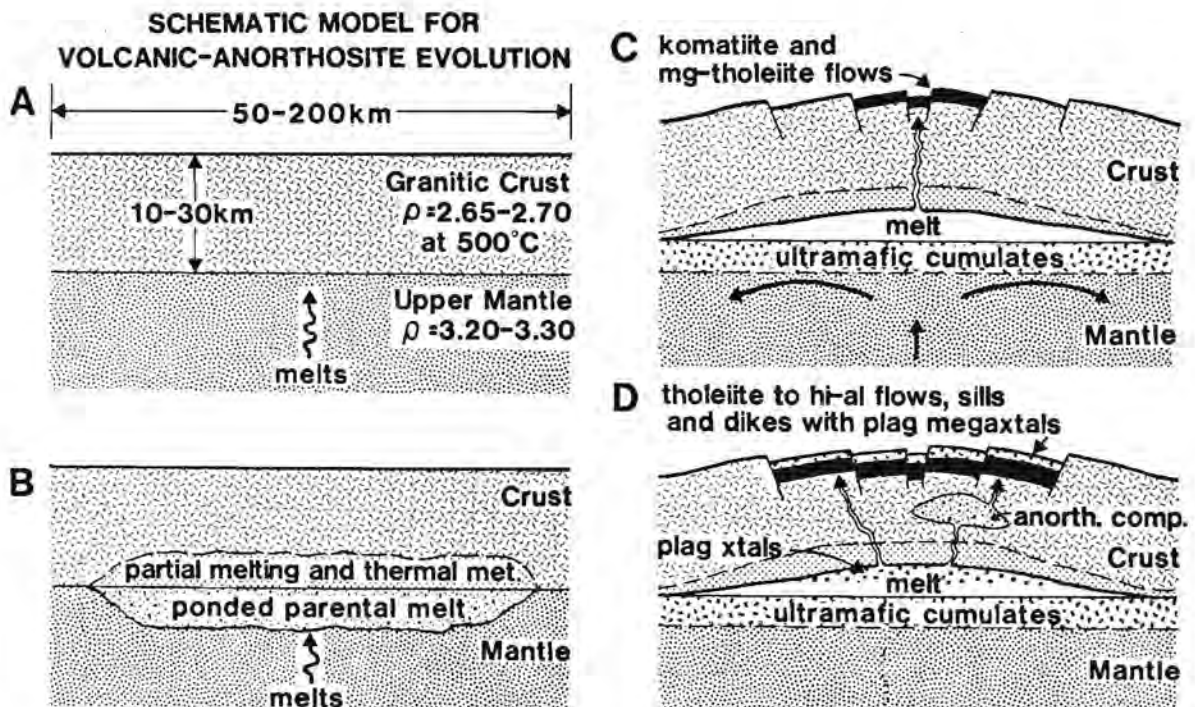


FIGURE 3



BASIC MAGMATISM OF THE SOUTHERN OKLAHOMA AULACOGEN; B. N. Powell, Research and Development, Phillips Petroleum Co., Bartlesville, Okla., 74004

The Glen Mountains Layered Complex (GMLC)[1] of the Wichita Province is the oldest body of a bimodal group of basic plutonic igneous rocks associated with the Southern Oklahoma aulacogen. Intrusive into the GMLC is a suite of amphibole-phlogopite-bearing gabbros collectively known as the Roosevelt Gabbros (RG)[1]. Contrasted compositions imply different parental magmas for the GMLC and RG, which may be unrelated genetically. The magmatic history of the province is further complicated by voluminous younger Cambrian granite and rhyolite which transect and unconformably overlie the older basic rocks and by relatively minor intermediate rocks of uncertain origin.

The GMLC comprises a sequence of gently dipping (5-20°) layered feldspathic cumulates, including anorthosites, troctolites and gabbroic anorthosites. Throughout the limited exposed vertical section (<200m) plagioclase (An78-60) is the dominant cumulus phase, representing 60-95% of the mode[1]. Olivine (Fo75-66) is also cumulus and varies from 0-25 vol.% (Fig. 1). In the upper third or so of the exposed section augite (cpx) (0-40 vol.%) is cumulus, being intercumulus below. Orthopyroxene (opx) (0-10 vol.%) is strictly intercumulus as are accessory ilmenite, magnetite and very rare apatite. Rocks of the GMLC are typically extreme adcumulates and heteradcumulates, with only occasional mesocumulates. Primary magmatic hydrous phases are strictly absent, even in the interstices of mesocumulates. Olivine commonly exhibits postcumulus peritectic rims of hypersthene (Fig. 1) and is locally variably replaced by late-stage symplectic bronzite-magnetite intergrowths, the latter probably reflecting locally increased fO₂ during final crystallization.

The lowermost exposed member of the GMLC, designated the K Zone[1], consists of alternating bands of anorthosite (plagioclase adcumulates) and troctolite (plagioclase-olivine adcumulates), often exhibiting rhythmic layering and igneous lamination. Typically the olivine poikilitically encloses plagioclase crystals that are small and randomly oriented relative to those outside the olivine, a texture seen also in the Rhum intrusion[2]. Although plagioclase clearly nucleated prior to olivine, both are cumulus, the poikilitic olivines having settled as bimineralic aggregates[1,2]. Very widely scattered in the K Zone are ophitic augite crystals, interpreted as strictly intercumulus from the size and orientations of included plagioclase relative to that external to the augite. Minor intercumulus opx, ilmenite and magnetite are also present.

Stratigraphically above lies a more broadly gabbroic zone (L Zone), comprised largely of anorthositic gabbro with more abundant intercumulus ophitic augite (plagioclase and plagioclase-olivine heteradcumulates). Troctolites, with poikilitic cumulus olivine, are less abundant than in the K Zone. Intercumulus opx and magnetite (poikilitic) are widespread. Igneous lamination (sometimes extreme) is common, but rhythmic layering is rare as modal variations are more gradational than abrupt.

The uppermost exposed member (M Zone) is still more mafic than the L Zone although still feldspathic. The principal change in cumulus relations is the appearance of cpx as a cumulus phase. Augite exhibits fine-ophitic textures analogous to those of olivine lower in the section and which are likewise interpreted as settled bimineralic clots. Olivine is more abundant than in lower zones and occurs as discreet (rather than poikilitic) cumulus grains interspersed among cumulus plagioclase crystals. Intercumulus opx, ilmenite and magnetite (the latter poikilitic) continue to be present. Igneous lamination (Fig. 1) and rhythmic isomodal layering are common in parts of the zone.

The exposed section of the GMLC exhibits "normal" cryptic variation in the form of increasing Fe/(Fe+Mg) ratios of olivine and pyroxenes with increa-

Powell, B. N.

sing stratigraphic height. Some indication of subtle Na enrichment in plagioclase may be present, but the interpretation of compositional data is complicated by complex crystallization histories of many grains manifest in complex zoning patterns. Interestingly, plagioclase compositions determined on drill cuttings from 2400m below the exposed level reveal similar An values to those from outcrop samples.

The apparent crystallization sequence in the exposed GMLC, derived from rock textures, cumulus/intercumulus relationships and cumulate sequences, is: plagioclase-olivine-cpx-opx. This sequence is theoretically permitted by crystal fractionation according to liquidus relations in the "system" Ol-Cpx-Plag-SiO₂[3]. The highly feldspathic character of the complex coupled with the early appearance of plagioclase in the crystallization sequence suggest a high-Al₂O₃ parent magma. A tholeiitic affinity is also implied by the presence of opx and cpx and by the peritectic reaction of olivine.

Petrologic modeling of the complex is severely constrained by its fault-bounded character (no chilled margins) and by the limited vertical exposure. Phase compositions suggest the approximate midsection of the complex is exposed. Drilling indicates possibly as much as 2400m of similar rocks below the exposed section. Gravity data and geophysical modeling strongly suggest the presence of a few km of mafic cumulates at depth[4], which implies olivine and/or pyroxene as early liquidus phases (before plagioclase). This appears inconsistent with observed textures in the exposed rocks, for which a number of ad hoc explanations suggest themselves. For example, changes in crystallization sequence could result from magma mixing as proposed for the Muskox intrusion [3,5]. Alternatively, the exposed rocks could represent a single macrorhythmic unit or part of one. Given the limited exposure this must remain conjecture.

The bottom accumulation of plagioclase - strongly suggested by field evidence and textural features - could be explained by a combination of cooperative sinking with mafics and oscillatory nucleation, as proposed by Morse[6] for Kiglapait troctolites. Intermittent density currents as well as bottom crystallization[7] may also have played variably important roles. Complex zoning in plagioclase may have resulted from convective circulation in the magma chamber prior to final accumulation.

By contrast to the GMLC, the Roosevelt Gabbros (RG) are characterized by the presence of primary magmatic hydrous phases. Typically both amphibole and tri-octahedral mica are present in amounts ranging from ca. 1% to 26%. Although late in the crystallization sequence relative to olivine, plagioclase, cpx and opx, the hydrous phases have every appearance of being primary magmatic and not deuteric or hydrothermal. Where relatively abundant they form large poikilitic crystals enveloping unaltered plagioclase, olivine and pyroxene (Fig. 2). In extreme cases pyroxene may be lacking in olivine-rich gabbro, its textural position occupied instead by amphibole and mica. In a single sample olivine may be rimmed by opx, amphibole and/or mica, illustrating variable behavior of the residual liquid in respect to the olivine peritectic reaction. Amphibole compositions range from titanian magnesian hastingsite to kaersutite (TiO₂ > 4.5%). Mica ranges from phlogopite (4-6.6% TiO₂) in olivine-bearing rocks to titaniferous biotite (TiO₂ > 5.5%) in olivine-free rocks. Curious local ultramafic segregations of olivine, magnetite and ilmenite with minor interstitial pyroxene, amphibole, phlogopite, plagioclase and pleonaste spinel show systematically more magnesian silicate compositions compared to enveloping gabbroic materials (Fig. 3). The high Mg values are considered to be essentially liquidus compositions and are attributed to locally elevated fO₂. By contrast, the coexisting ilmenite and magnetite have equilibrated to subsolidus temperatures (700-800°C) and fO₂ values near the FMQ curve, eliminating any fingerprint of initial heterogeneities in magmatic fO₂ values.

Powell, B. N.

Bulk compositions of the RG resemble the high TiO_2 -high P_2O_5 -low Al_2O_3 magmas recognized in the Keweenaw of Upper Michigan[8]. The hydrous condition of the parental magma(s) for the Roosevelt Gabbros could have been derived from melting of hydrous upper mantle source rock or could be the result of crustal contamination. The principal interest lies in their crystallization behavior as hydrous olivine tholeiites and in their chemical contrast to the probable parental magma for the GMLC, thus suggesting a bimodality of basic magmatism that is becoming more widely recognized in other provinces.



Fig. 1. Cumulus olivine and plagioclase (laminated) in the GMLC. Olivine is partly rimmed by postcumulus peritectic hypersthene.



Fig. 2. Primary magmatic kaersutite and phlogopite (dark, interstitial) in olivine gabbro (Roosevelt Gabbro) with associated ultramafic segregations (not pictured).

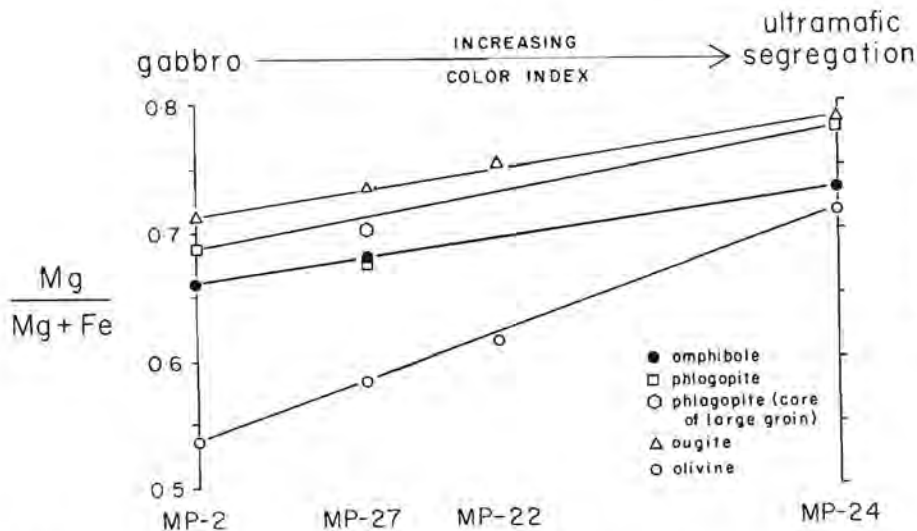


Fig. 3. Variations in mineral compositions in Roosevelt Gabbro with ultramafic segregations. (See text.)

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MINERALOGY AND PETROLOGY OF LAYERED INTRUSIONS: A REVIEW.

L.D. Raedeke, Dept. of Geological Science, University of Washington, Seattle, Washington 98195

Layered basic and ultrabasic intrusions range in age from Archean to Tertiary, in stratigraphic thickness from a sill of a few hundred meters to the Bushveld Intrusion of nearly 9000 meters, and in areal extent from a few km² to the Dufek Intrusion of Antarctica which may be greater than 50,000 km². The most extensively studied layered intrusions include the Bushveld of South Africa (55,6,4,5,29,47,48), the Skaergaard of East Greenland (49,31,32,33,34,40,41), the Stillwater of southwestern Montana (16,28,42,36,43,44,45), the Duluth Complex of northeastern Minnesota (53,54), the Kiglapait of Labrador (37,38,39,20), the Great Dyke of Zimbabwe (15,57,58,29,19), the Muskox in Canadian Northwest Territories (26,27,23,25), the Dufek of western Antarctica (12,13,17,18), the Jimberlana of Western Australia (9,7,8), Rhum of northwest Scotland (3,14,11), and Duke Island of southeastern Alaska (24). For more general review of layered intrusions, see (49,59,56,46). In this paper I cannot attempt to develop all possible or proposed explanations for the observed data on the mineralogy and petrology of these intrusions. I only attempt to report highlights of the observations, to point out some of the complications in interpretation of these data, and where appropriate, to put forward some selected hypotheses on petrogenesis. My intent is to stimulate further discussion of these petrologically and economically important igneous bodies.

As background, it is useful to clarify the terminology of igneous cumulates used in this paper. These rocks consist chiefly of *cumulus* crystals, that is crystals which nucleate and grow predominantly in equilibrium with the main magma body. Cumulus crystals may form *in situ* or they may be transported through the magma by settling, flotation or currents before arriving at their site of accumulation, where they will continue to crystallize. During and/or after accumulation of cumulus crystals, *postcumulus* crystals will grow, typically filling interstices between cumulus grains. Although these definitions are genetic, the distinction between cumulus and postcumulus minerals is made using textural criteria (e.g. 36). For this reason, the use of the terms has recently come under serious question (8, and A.R. McBirney, pers. comm., 1980).

Crystallization sequence: Because these intrusions are layered, it is appropriate to consider their stratigraphy, or more specifically the inferred crystallization sequence of minerals. Generally these intrusions consist of a lower ultramafic zone of olivine + pyroxene and an upper zone of plagioclase + olivine + pyroxene. The generalized crystallization sequences of eight intrusions are shown in figure 1. From the base upward in the Bushveld, Hartley and Stillwater Complexes, a crystallization sequence of [ol, opx + ol, opx, opx + pl, opx + pl + cpx] is well-defined. This crystallization sequence and the modes of coexisting minerals are consistent with those predicted by phase equilibria for a basaltic system undergoing fractional crystallization where all crystallizing phases accumulate together from the floor upward. Similarly, although crystallization sequences vary, all the other intrusions plotted on figure 1 show lithologies representing progressively more differentiated magmas upward in the section. Therefore, although there is on-going debate as to the actual site of crystallization and/or the mechanism of transport and accumulation of the crystals, it seems clear that an important mechanism in the formation of these intrusions is indeed fractional crystallization with accumulation (or crystallization) from the base upward as Bowen (2) suggested. There are however, many complications to this simple picture:

Raedeke, L.D.

1) *Macro-rhythmic units.* Brackets marking sections in figure 1 indicate that the minerals shown actually occur in a series of repeating lithologies, called *macro-rhythmic* or *cyclic* units. In the ultramafic zones of the Bushveld, Stillwater and Hartley Complexes, each cyclic unit is composed of basal dunite, overlain by harzburgite, then by orthopyroxenite with the crystallization sequence [ol, ol+opx, opx]. There are 10 to 20 whole or partial cyclic units superimposed in the peridotite subzones of these intrusions. In the Muskox, the cycle dunite, olivine clinopyroxenite, olivine gabbro repeats in the lower portion with the crystallization sequence [ol, ol+cpx, ol+cpx+p]. More complex cycles occur in the upper Muskox and middle portions of the Stillwater. As first suggested by Cooper (10), the macro-rhythmic peridotite units may result from periodic influxes of undifferentiated (or less differentiated) magma. This can occur in an open system with associated volcanic tapping as suggested by Brown (3), or in a closed system as suggested by Jackson (28). In the latter case, it is postulated that a zone of crystallization at the floor of a magma chamber is effectively isolated from the main magma as a stagnant layer. Within this layer crystallization proceeds to form the dunite to orthopyroxenite sequence. The cycle is restarted or truncated when a major convective overturn within the chamber displaces the stagnant layer and replaces it with the more primitive magma from the main body.

2) *Change in crystallization sequence.* Simple fractional crystallization will yield a crystallization sequence which is predictable by phase equilibria. Mixing of more primitive, consanguineous magma with the fractionating magma will commonly restart or retard the crystallization sequence, but will rarely change it. Nevertheless, a marked change in crystallization sequence is observed in the Stillwater from [opx, opx+p], [opx+p+cpx] in the Lower and Upper zones, to [ol+p], [ol+p+cpx], [ol+p+cpx+opx] in the Middle zone. Also, in the Muskox the sequence changes from [ol, ol+cpx, ol+cpx+p] in the lower zones to [ol, opx, opx+cpx, opx+cpx+p] in the upper portions. Irvine (23) explains this change by a model involving progressive contamination of the Muskox magma with salic material melted from the intrusions's roof rocks. No explanation has yet been proposed for the complex Stillwater case. It can be said, however, that such a marked change in crystallization sequence requires either a change in magma composition or a change in the position of phase boundaries by a perturbation in crystallization conditions.

3) *Reversals of crystallization sequence.* The basal subzones of the Stillwater, Bushveld and Muskox are marked by a reversal in crystallization sequence such that rocks grade from ultrabasic to basic downward in the section. For example, the lowest rock type in the Basal subzone of the Stillwater Complex is typically an opx+p+cpx diabase. This is overlain by a heterogeneous norite, orthopyroxenite and the layered peridotites. The crystallization sequence upward is [opx+p+cpx, opx+p], [opx, ol], which is the reverse of that in the main layered series. There is also an irregular but persistent trend of Mg-enrichment upward in orthopyroxene composition of the Basal subzone (42). All of these data are consistent with a model of contamination of the magma in the contact zone by a partial melt of the country rock as was proposed by Raedeke and McCallum (43). Nevertheless, as Page (42) points out, because of the ubiquitous irregularities and complexities of the Basal subzone, it is unlikely that either mechanical or chemical equilibrium were reached over any appreciable volume.

Mineral compositions: As preface to a discussion of mineral compositions, it is necessary to consider the actual meaning of recorded mineral composition of cumulate rocks. It has been shown that ultimate compositions of coexisting

Raedeke, L.D.

minerals are at mutual equilibrium (35), but these final compositions are the product of a series of processes including: 1) crystallization of cumulus minerals in equilibrium with the main magma body, 2) postcumulus crystallization of the intercumulus liquid, and 3) subsolidus re-equilibration. The ultimate mineral compositions will adequately reflect the original cumulus mineral composition in equilibrium with the main magma body only if processes 2 and 3 do not effectively dominate. Process 2 will become important if the mode of a mineral is small relative to the amount of intercumulus liquid at the time the liquid is effectively trapped (44). Process 3 also becomes important as a function of mode. For example, as Irvine (21,22) demonstrated, final olivine and spinel compositions after complete equilibration will be a function of initial modal proportions, i.e., Fo content of olivine will increase as mode of coexisting chromite increases. Process 3 will become progressively more important as K_d for the parameter of interest (e.g. FeO/MgO) deviates from unity. Therefore, since values of $K_d(\text{FeO/MgO})$ between coexisting opx, cpx and ol are near unity, the major effects on composition of mafic silicates are 1) presence of abundant oxides or sulfides (which will cause Mg-enrichment of silicates since $K_d(\text{oxide/silicate}) > 1$), or 2) very low mode of the mafic silicate (which will cause Fe-enrichment of the silicate through equilibration to the intercumulus liquid, since $K_d(\text{mafic crystal/liquid}) < 1$).

The change of composition of cumulus minerals with stratigraphic height is termed *cryptic variation* or *cryptic layering* by Wager and Deer (50). Systematic cryptic variation over a restricted thickness is seen in the ultramafic portions of the Muskox and Stillwater Intrusions. Detailed study in the Muskox shows Fe-enrichment through macro-rhythmic units with a break to more magnesian mafic minerals between units. This break is translated upward beyond the lithologic break by what is postulated to be infiltration metasomatism (25). The features of one macro-rhythmic unit in the Stillwater Complex are: 1) constant Mg# through the basal dunite, 2) smooth Fe-enrichment from the base of the harzburgite up into the lower orthopyroxenite, and 3) smooth Mg-enrichment followed by constant Mg# upward through the remainder of the unit. There is evidence for a sharp break at or near the lithologic boundaries between units (45). The change from Fe-rich to more magnesian mafic minerals in both the Muskox and Stillwater macro-rhythmic units, is interpreted as being the result of injection of more primitive magma into the chamber followed by mixing (25,45). On the larger scale of the entire ultramafic zone of the intrusions in figure 1, Mg# is seen in most cases to vary only 2-3 mol % En from a mean value. This also may be due in part to periodic influxes of more magnesian magma, however, the lack of systematic Fe-enrichment on the broad scale cannot be explained by this process alone.

In the lower peridotite subzones of the Hartley and Stillwater Intrusions, an exceptional trend of Mg-enrichment of about 10 mol % En is seen. In the Stillwater, the lowest Mg# of opx in the peridotite cumulates is En 76 and is found at the very base of the section. The maximum value of En 86 is found ~600 m higher up. This Mg-enrichment cannot be explained by increasing degrees of partial melting of a mantle source because mantle opx is certainly more magnesian than En 76, nor can it be explained by fractional crystallization of an Fe-rich mineral, because none appear in sufficient abundance to account for the magnitude of the change. The trend could be explained by diminishing olivine loss during continuous magma supply as the rock surrounding the conduit of the feeder system heats up. Alternatively, the trend could be the result of contamination of the Stillwater Complex magma by a partial melt of the country rock coupled with partial equilibration of the magma to

Raedeke, L.D.

the abundant orthopyroxene in the hornfels (En 45) (1). Either of the latter alternatives would have diminishing effect as the magma chamber became established.

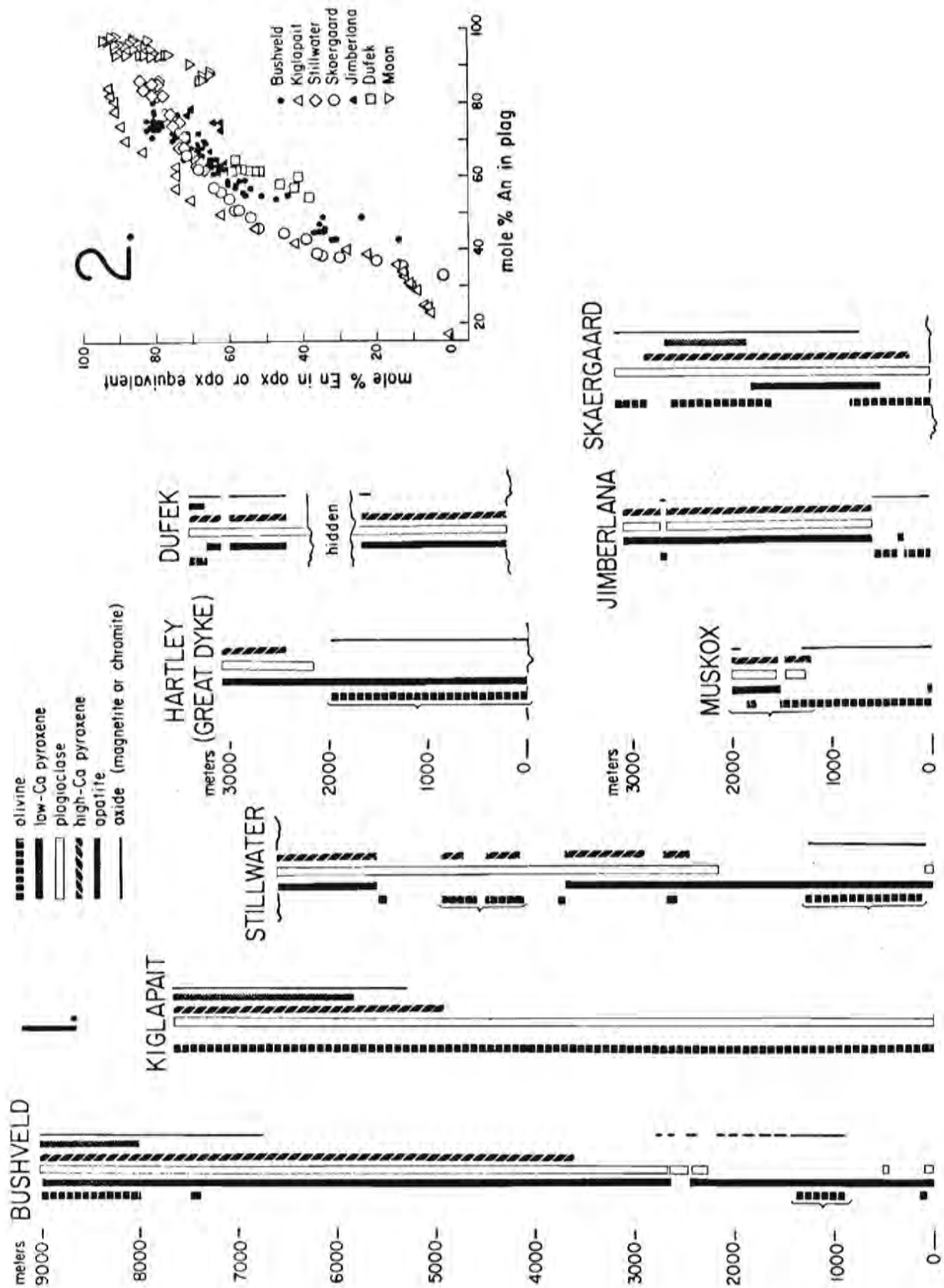
Cryptic variation seen in plagioclase+opx/cpx/ol cumulates of six intrusions is shown on figure 2 (see figure caption for details of the plot). Generally, stratigraphic "up" is from the upper right to the lower left of the figure. All six complexes show a "normal" oblique trend predicted for fractional crystallization with mineral segregation from the base upwards. There are subtle differences between trends which are most likely due to the difference of crystallization sequence from one intrusion to another. The Mg-rich suite of rocks from the lunar crust are also plotted. The oblique lunar trend can be modeled by perfect fractional crystallization (30), but the trend of the model is too steep for crystallization trends defined by any of the other intrusions, with the possible exception of the Dufek. Therefore, although fractional crystallization with mineral segregation from the base upwards is undoubtedly an important mechanism, other processes such as periodic mixing of more primitive magma, or ineffective crystal segregation must also have been operative. These processes and their predicted effects must be modeled for each individual intrusion.

In summary, mineralogical and petrological data indicate that several processes are important in the formation of large layered intrusions. In the ultramafic portions of these intrusions, periodic injection of magma and associated mixing dominate, with continuous fractional crystallization occurring between mixing events. The injected magmas are probably consanguineous but may be of varied compositions, modified by olivine loss during ascent, or by assimilation of the country rock. Although the effect of assimilation of country rock is most easily noted in the basal portions of many intrusions, contamination may also cause a marked change of crystallization sequence elsewhere in the intrusion. Fractional crystallization begins to dominate only after plagioclase begins to crystallize, but ideal fractional crystallization seldom or never occurs. Fractional crystallization may be modified by many mechanisms including imperfect mineral segregation, periodic injection of magma, or complex convection in the chamber. After accumulation or *in situ* formation of cumulus grains, crystallization of the intercumulus liquid and subsolidus re-equilibration continue to alter mineral compositions. Also, although not specifically treated in this discussion, many processes must be considered in the development of reasonable petrogenetic models including: the fluid dynamics of a continuously growing and changing magma chamber, regional tectonics related to chamber growth, the role of volatiles, rate and mechanism of heat loss, among others.

Figure 1. Simplified plot of cumulus minerals vs. stratigraphic height. Broken line at base or top of the section indicates that some portion is missing. Brackets indicate minerals present occur as cyclic lithologic units. Data sources — Bushveld: Basal and Lower zones (5), Critical zone (4,6), Main and Upper zones (47); Kiglapait: (38); Stillwater: Ultramafic zone (45), Banded zone (36); Hartley: (29); Muskox: (23); Dufek: (17); Jimberlana: (7); Skaergaard: (49), Upper Border Group (41).

Figure 2. Mol % An in plagioclase vs. mol % En in coexisting opx. If only cpx or ol data were available, the opx predicted to be in equilibrium with cpx or ol, or "equivalent opx" was calculated using values of $K_d(\text{FeO/MgO})_{\text{ol/opx}} = 1.23$, and $\text{cpx/opx} = 0.819$. K_d values were determined on Stillwater rocks (35, and Raedeke, unpublished). Only rocks with both cumulus plagioclase and cumulus mafic mineral are plotted. Skaergaard data were based on optical determinations, Bushveld data were based on a variety of methods, all other data were obtained by electron microprobe. Data sources — Bushveld (5,47), Kiglapait (38), Stillwater (44), Skaergaard (49), Jimberlana (7), Dufek (17), Moon (51,52).

Raedeke, L. D.



Raedeke, L.D.

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MODAL AND CHEMICAL VARIATIONS IN THE ULTRAMAFIC ZONE OF THE STILLWATER COMPLEX. L.D. Raedeke and I.S. McCallum, Dept. of Geological Sciences, University of Washington, Seattle, WA 98195

The Ultramafic zone of the Stillwater Complex can be subdivided into a Basal subzone of diabase, heterogeneous norite and orthopyroxenite, a Peridotite subzone (PSZ) of olivine (ol) + orthopyroxene (opx) cumulates, and an Orthopyroxenite subzone (OSZ). Two stratigraphic sections through the Ultramafic zone have been completed (figure 1), one at Mountain View in the eastern portion of the Complex, and one at Chrome Mountain ~40 km along strike to the northwest of Mountain View.

Modal layering is well-developed in the Peridotite subzone, but sedimentary structures such as cross-bedding, cut-and-fill, and graded layering are relatively rare. Modal layering occurs on a large scale in macro-rhythmic units of dunite-harzburgite-orthopyroxenite in both sections, and in harzburgites at Chrome Mountain as cm-scale alternating olivine-rich and pyroxene-rich bands. Chromite occurs as disseminated grains in most ol-bearing units, and as chromite seams near the base of many dunites. At Chrome Mountain, fine-grained discordant dunite masses crop out over large areas and appear to replace the cumulates. Dunite is particularly abundant in areas of extensive faulting. Cross-cutting bodies of pegmatitic opx are closely associated with this secondary dunite. Pegmatitic opx is also associated with chromitites at both Chrome Mountain and Mountain View.

Mineral compositions obtained by electron microprobe are superimposed on the standard stratigraphic section in figures 1 and 2. First, note that opx is consistently more magnesian than coexisting ol, which is simply a consequence of Fe/Mg partitioning between ol and opx. As determined from these data, $Kd(FeO/MgO:ol/opx)=1.23$ ($r=0.97$). The cryptic variation through the Ultramafic zone does not show systematic Fe-enrichment as predicted if fractional crystallization with segregation of crystals at the bottom were the dominant mechanism operative. Rather, within the lower PSZ there is systematic Mg-enrichment upward which shows no apparent correlation with lithology. The most Fe-rich compositions found in the Ultramafic zone are actually at the base of the section. In both stratigraphic sections, the Mg-enrichment occurs as three separate chemical cycles with a sharp break to more Fe-rich compositions between cycles. These breaks occur at ~200m and 370m at Mountain View, and 280m and 400m at Chrome Mountain. The Mg-enrichment cycles broadly correlate in thickness, degree of enrichment, and stratigraphic position between Chrome Mountain and Mountain View, but lithologies do not. Some hypotheses to explain this Mg-enrichment are given by Raedeke (1).

Above the Mg-enrichment cycles at Mountain View, there is a zone of relatively wide variation of $Mg/(Mg+Fe)$ (Mg#) which gradually narrows upward to a constant value (En 85) that is maintained throughout the remainder of the PSZ and the entire OSZ. A single macro-rhythmic unit was sampled in detail within this zone (figure 2). The middle harzburgite of the sampled unit is unusually narrow, but otherwise the unit is typical. In figure 2, the following points concerning cryptic variation are noteworthy: 1) With the exception of the olivine point at 780m (which is apparently not in equilibrium with coexisting opx), olivine composition remains constant through the dunite. 2) Smooth Fe-enrichment occurs through the harzburgite into the lower orthopyroxenite. 3) There is a reversal to Mg-enrichment in the lower orthopyroxenite then constant Mg# is maintained upward through the remainder of the unit. 4) Considering only opx compositions or opx compositions predicted by Kd to be in

Raedeke, L.D. and McCallum, I.S.

equilibrium with measured olivine compositions, there is a sharp break of ~ 1 mol % En at the lithologic break into the next overlying cyclic unit. It seems reasonable to extrapolate the cryptic variation seen in figure 2 to other macro-rhythmic units in the upper PSZ at Mountain View (figure 1). The sparse data for these other units seem to support this extrapolation. For example, note that in the dunites at 700m and 1000m, olivine composition remains approximately constant. Also note that the most Fe-rich opx compositions occur within the lower orthopyroxenites, such as at 520m, 625m and 745m.

The lithologic sequence and cryptic variation of the macro-rhythmic units in the PSZ can be explained by a model of periodic influxes of more primitive magma with little or no initial mixing as proposed by Huppert and Sparks (2), with subsequent modification of mineral compositions by infiltration metasomatism (3). In the Ultramafic zone as a whole, the Fe-enrichment predicted if fractional crystallization with bottom accumulation dominates is never seen on a large scale. It is, however, seen on the smaller scale within macro-rhythmic units. Therefore, in general, it appears that periodic injection of magma and associated mixing dominate in the PSZ with continuous fractional crystallization occurring between mixing events. The constant Mg# throughout the OSZ may also result in part from magma mixing, but the lack of systematic Fe-enrichment on the large scale cannot be explained by this process alone.

The most dramatic difference between the two sections is that at Chrome Mountain both the PSZ and OSZ are much thinner than at Mountain View. By contrast, the thickness of the three separate cycles of Mg-enrichment in the lower PSZ are approximately the same at the two locations. If multiple injections of magma were occurring during formation of the Ultramafic zone, the thicker section at Mountain View may have formed in a subsiding basin within the magma chamber. Chrome Mountain, by analogy, must have been a relatively stable "topographic" high, with a lateral transgressive contact between the two locations. If the Mg-enrichment cycles at the base of the PSZ are a primary cumulus feature, then, since these do correlate along strike, the initiation of major subsidence must not have begun until >500 m of peridotites had accumulated.

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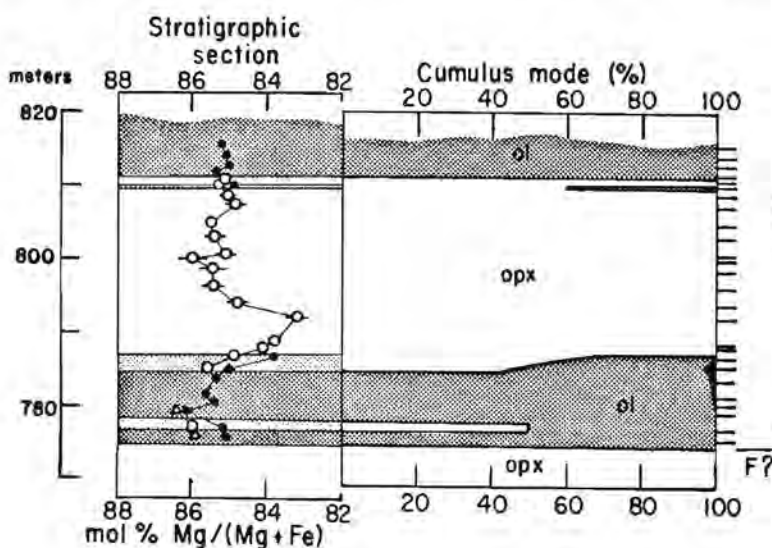
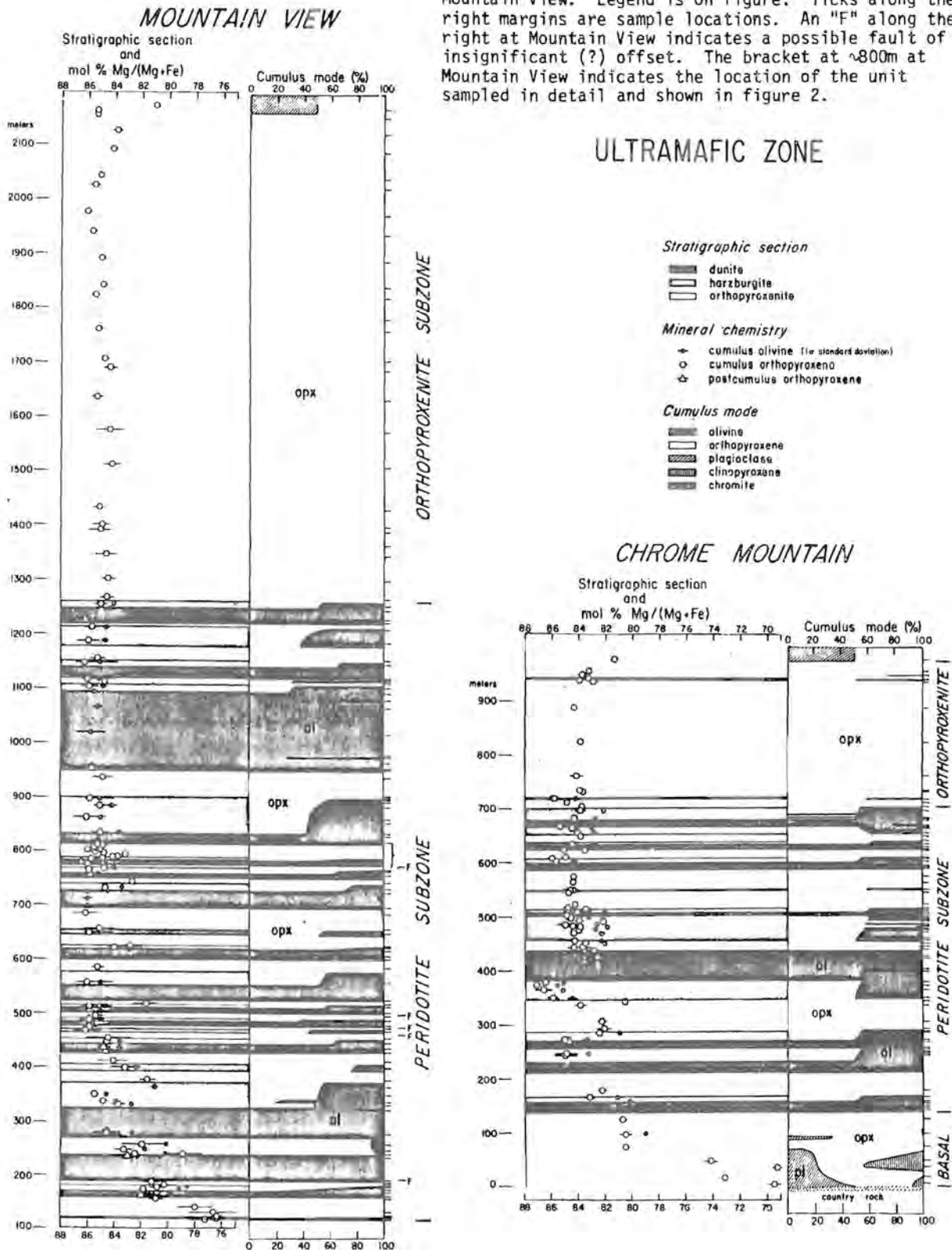


Figure 2. Macro-rhythmic unit #15 at Mountain View. Legend is the same as for figure 1.

Raedeke, L.D. and McCallum, I.S.

Figure 1. Mode of cumulus minerals vs. stratigraphic height (right-hand columns) and standard stratigraphic section with mineral compositions superimposed (left-hand columns). The Basal subzone is not shown for Mountain View. Legend is on figure. Ticks along the right margins are sample locations. An "F" along the right at Mountain View indicates a possible fault of insignificant (?) offset. The bracket at ~800m at Mountain View indicates the location of the unit sampled in detail and shown in figure 2.



DOUBLY DIFFUSIVE CONVECTION AND SOLUTE BANDING IN MAGMAS, A. Rice,
 Dept. of Oceanography, Florida Inst. Technology, Melbourne, FL 32901

Industrial experience indicates two simultaneous processes take place in solidifying melts: 1) solute banding and 2) doubly diffusive or Soret convection. The first process, solute banding, can strongly fractionate a melt that was initially homogeneous throughout. Solute banding occurs as a result of solidification of a vigorously convecting melt. There is strong theoretical evidence that magma chambers can experience vigorous convection. Solute banding manifests itself in two fashions, both of which may occur simultaneously: a) microsegregation and b) macrosegregation. The former is a manifestation of temperature fluctuations inherent in vigorous convection. These fluctuations serve to quench and remelt, quench and remelt in a cyclic fashion crystals to the solidification front which forms the walls and roof of the chamber. If a two component (A and B) fluid is involved and the fluid is on the "A" side of the systems cotectic, the temperature downswing convecting past the solidification front would quench or "plaster" a layer of composition "A" to the ceiling or walls. Partial remelting of this new layer would occur on the temperature upswing with "B" enriched material from resorption re-entering the melt in the solutal boundary layer just beneath the solid-liquid interface. Cotectic shifting accompanies such nonequilibrium processes and the melt immediately beneath the solidification front would have its composition shifted further to the "B" side of the cotectic.

On the next temperature downswing, a solid of composition "B" would be quenched or "plastered" to the roof and walls of the chamber. This process repeats itself cyclicly to form striations in the solid alternating in composition "A" and "B". The obvious application of this process to the geological context is allowing "A" to be plagioclase and "B" to be olivine. Drawing on experimental work in high Prandtl (Pr) number fluids (e.g., Pr = 8000), that is, highly viscous fluids, one may infer periodicities in temperature fluctuations and then from solidification rates obtain theoretical estimates of the band width of the layering. For an intrusion the size of the Skaergaard, theoretical bandwidths run centimeters to meters. Microsegregation banding is accompanied by sharp boundaries, crystal structure oriented with the flow and rezoning in the crystals. The rezoning is indicative of temperature fluctuations during growth. Solute banding has the distinct advantage of providing "sedimentary" features along walls without having to appeal to sedimentation! If solute banding had been operant during the solidification of the Skaergaards, there are reasons for confining it to the border groups.

Macrosegregation represents variations in composition on much larger scales. In the context of the Skaergaards, these would be variations over kilometers rather than meters. While phases "A" and "B" may be cyclicly quenched out onto the solidification front, the first components of "A" and "B" to be precipitated out will be the high temperature components. This will leave residual melt in the upper reaches of the chamber depleted in the high temperature components which have gone into the overlying solid. The initially homogeneous melt becomes stratified with melt closer to the original composition underlying the depleted melt at the top of the chamber. On complete solidification, one would find the initially homogeneous distribution in the vertical shifted such that high temperature components were at the top, low temperature components further down and composition grading over to that of the original melt as one continues down through the chamber. Applying the process to our chosen geological context, one would expect to find the top of the intrusive enriched in An, Fo and enriched in Ab, Fa somewhat further down. Continuing down through the unit, the composition should grade back over to An, Fo. Other complexities can

RICE, A.

diversify this picture.

Once the concentration gradients have set in through macrosegregation, Soret convection can come into play. Concentration gradients as well as temperature gradients must attend Soret processes (hence the appellation "doubly diffusive"). There are two possible situations with two different types of Soret convection attending: I) heavier, hotter fluid underlying lighter, cooler fluid and II) cooler, lighter fluid underlying warmer heavier fluid. The latter case is obviously metastable. However, the former case is not at all necessarily stable for all time. Case I is referred to as simply "diffusive" convection and case II is referred to as "finger" convection. In both cases, the form of convection is distinctly different than the more familiar Benard convection wherein a single convecting system occupies the fluid from top to bottom (only one cell in the vertical). In doubly diffusive convection, layers of convection cells may occupy the fluid from bottom to top. If there is a discontinuous compositional change such as one distinctly different fluid underlying another, i.e., a sharp stratification, these layers of convection cells develop at the interface of the stratification and slowly spread vertically through the fluid. The initial sharp discontinuity in composition through the vertical is replaced with a "staircase" distribution. Each layer of convection cells is of uniform temperature and composition with sharp changes in temperature and composition in the boundary layers separating each layer of convection cells. In the "diffusive" case, temperature and density increase through each band as one proceeds downward. If there is a heat flux through the system, then mixing at interfaces and heating from below and/or cooling at the top will eventually lead to density equilibration between layers or bands before there is compositional or temperature equilibration. When this occurs between two contiguous layers, their mutual boundary fades and they mix together, telescoping one into another. This telescoping spreads in an exponential fashion until the entire system is homogenized. This mixing event is termed rollover in the engineering literature. Systems have been observed to stratify due to convective fractionation, then roll over (homogenize), re-stratify, roll over again, continuing on repetitiously. These processes may be interrupted by freezing, volatile release etc. In seeking to find manifestations of these mechanisms in magma chambers, an assesment of rates and accelerations of solidification must be made. The ability of magmas to convect vigorously until quite slushy (approx. 55% crystallization) and then experience a snap through increase in viscosity by many many orders of magnitude allows the possibility that the structure of the on-going convection may literally be captured in stone. For sufficiently rapid cooling, bands of diffusive convection may be frozen in one after the other from the top down. Roll over or homogenization may be moving upward to this "solidification" front, whereupon solute banding would re-stratify the remaining fluid in the chamber (which would be towards the bottom). These competing mechanisms yield the possibility of cyclic variation in the macrosegregation, on the order of kilometers, with compositional shifts in each cycle. We associate macrosegregation with cryptic variation and there are units (e.g., Bushveld) which show cryptic variations reversing themselves, then returning to general trends with depth. Because of the diffuse nature of the "zebra" banding (which is confined to the main units) of the Skaergaard as opposed to the sharp banding of the Border groups, we attribute the Border groups to solute banding, and the main unit to layers of doubly diffusive convection that were frozen in place from the top down, the doubly diffusive convection brought into play by the macrosegregation associated with the solute banding. Theoretical estimations of doubly diffusive band widths are within those observed geologically. It should be noted that turbulent convection can cause solute banding to pinch

A. Rice

out. The physical implausibilities associated with crystal settling (e.g., convective velocities orders of magnitude greater than settling velocities) do not attend convective fractionation. The vigor of convection likely to attend a mafic magma chamber dictates the chamber to be initially homogeneous. Soret convection can then only come about if solute banding has been a precursor. Attempts to model these processes in the lab with precipitation of crystals by changing equilibrium direction through cooling has no physical relationship to a solidifying melt and is meaningless let alone mismatch in scaling (e.g., density contrasts between crystals and melt and convection velocities) and let alone the roll of glass as a glue for crystals.

Convective fractionation need not be limited to mafic melts. Mafic banding in the outer fringes of granitic batholiths has been reported. Although scaling from the glass industry indicates convective velocities of a cm/sec for silicic melts, the massive viscosity change beyond 55% crystallization may impede further development of Soret effects in purely silicic environments. Soret effects very likely attend zoned chambers, i.e., silicic magma overlying mafic. Computations indicate a chill wave propagating downward into the less viscous mafic portion with attendant development of banded convection at the silicic-mafic interface. Since basalt is close to its liquidus in nature, the chill wave serves to quench, hence cause massive exsolution of volatiles which accelerates the rollover process by frothing up the underlying basalt. The result is almost instantaneous, massive explosive rollover, a problem not uncommon and of profound hazard in the engineering environment. Plinian eruptions generated by massive rollover might show silicic ash first, followed by banded ash, and then finally mafic ash.

The system of equations describing Soret processes in magma chambers are also the same equations describing fractionation columns. Each layer of doubly diffusive convection acts as a stage in a cascade fractionation system. When the banding develops, there is reason to suspect that chemical fractionation (e.g., Shaw) and/or isotope fractionation (e.g., Rice) may occur. Chemical fractionation through Soret processes has obvious application to units such as the Merensky Reef (chromite, platinum banding). Soret fractionation has been used for years to separate isotopes in industry with separation factors greater than 10^5 . Soret fractionation of isotopes is nonlinear and cannot be corrected for by normal mass difference procedures. For liquids, the degree of both chemical and isotope fractionation depends on the degree of polymerization, concentration, etc and all these things evolve with the convecting system. The Soret coefficient which governs these processes is a function of the log of the ratio of temperature extremes and is not a constant throughout the evolution of the convecting system. Trends may suddenly reverse themselves or take on extremely bizarre aspects. Soret transport by diffusion alone is negligible but with convection in attendance, it can be everything as it can be commercially. A theoretically feasible isotope evolution would be a long history of rhyolitic extrusives with isotope ratios suggesting origins in country rock followed on by explosive volcanism yielding silicic ash wherein the precursors to the explosion would be the layers of Soret convection in the magma chamber acting as fractionation stages to deplete the upper silicic portions of the chamber in those isotopes associated with country rock. The final ash would appear more primitive than it really was. Baring what actually happened, such an isotope history seems to attend Long Valley.

"Finger" convection consists of alternating layers of common convection separated by layers of "fingers" which look strikingly similar to crescumulates. Finger convection is always attended by an inversion such as seems to be seen in the Rhum intrusives, e.g., An decreasing as one goes downward.

COMPOSITIONAL VARIATION WITHIN AND AMONG PYROXENES AND OLIVINES IN LUNAR FERROAN ANORTHOSITE 60025. Graham Ryder, Lunar Curatorial Laboratory, Northrop Services, Inc., P.O. Box 34416, Houston, TX 77034.

60025 is a coarse-grained pristine ferroan anorthosite containing green clots or stringers of mafic minerals, a few centimeters long and several millimeters wide. Most analytical work, including microprobe studies (1-4) has been done on pieces taken from a slab cut through 60025 which did not include the mafic clots. Thin sections from this slab (~99% plagioclase) contain iron-rich pyroxenes and lack olivine. A single mafic mineral rich "undocumented" chip, which was loose in the same sample container as 60025, was made into thin sections. Apparently due to a lack of conviction that this chip was a part of 60025, no data were reported for this sample until Warren and Wasson (5) reported mineral analyses. These authors did not allude to the possibility that this sample might not be from 60025, nor to the clear difference in their mineral chemistry (olivines and pyroxenes, more Mg-rich) from the data reported in (1-4).

In conjunction with an allocation to K. Marti for radiogenic isotope studies, grains of the mafic clots were taken for optical and microprobe studies. These grains are olivines and low-Ca pyroxenes, individually larger than 2 mm. Olivine is apparently slightly more abundant. The pyroxenes contain abundant exsolution lamellae. I have made microprobe analyses of these olivines and pyroxenes, and of pyroxenes and olivines in a second thin section of the mafic rich chip analyzed by (5). The object was to ascertain the compositions of the mafic clots, the total variation of mafic mineral compositions in 60025, and the homogeneity of individual grains. The pyroxene analyses were mainly for 9 elements with instrumental, background, and Bence-Albee matrix corrections. Some were 3 element (Ca, Mg, Fe) with a similar correction procedure, and these agreed with the former for the same grains to about half mol per cent En. The olivine data were all 3 element (Ca, Mg, Fe) with similar correction procedures; a few were reanalyzed for 9 elements and agreed to within 1 mol% Fo.

The major element data are shown in Figures 1-3. The separated mineral grains are olivines $Fo_{66.5-67.0}$ and low Ca pyroxenes En_{68-72} . The pyroxenes are individually homogeneous (except for the exsolution lamellae) but show significant differences among grains (Fig. 1). The mafic mineral chip of (5) is dominated by olivines and pyroxenes with the same compositions as the picked mafic grains, hence is undoubtedly a part of the rock (a systematic difference of several mol% in both En and Fo between the analyses reported here and those of (5) is apparent). In addition the chip contains olivines and pyroxenes which are more Fe-rich. The pyroxenes include both high-Ca and low-Ca varieties, and the compositions extend to those reported for the 60025 slab (Fig. 2)(1-4). Olivine has a more restricted total range. Individual grains of both olivine (Fig. 3) and pyroxenes are quite homogeneous.

The results demonstrate that 60025 contains clots of Mg-rich pyroxene and olivine, and small grains of more Fe-rich pyroxenes. The total range is close to that reported for the ferroan anorthosite group as a whole. While each grain is homogeneous, the differences among grains show that metamorphic equilibration has not taken place over a distance of more than a millimeter. Because of their grain size, the Mg-rich mafics at least must be cumulate in origin, not exsolution or crystallized trapped liquid. The liquid from which these Mg-rich olivines and pyroxenes crystallized had an $Mg' < 45$, so even the more Fe-rich mafic grains are not simply trapped liquid. The rock is probably not polymict, as might be the case for the only other anorthosite with a considerable compositional range, 67075 (6), because the individual mafic clots contain adjoining minerals with different compositions.

Ryder, G.

Any model for the petrogenesis of lunar anorthosites must take into account that at least this one sample is unmetamorphosed, monomict, and contains a wide range of mafic mineral composition including at least some cumulate grains.

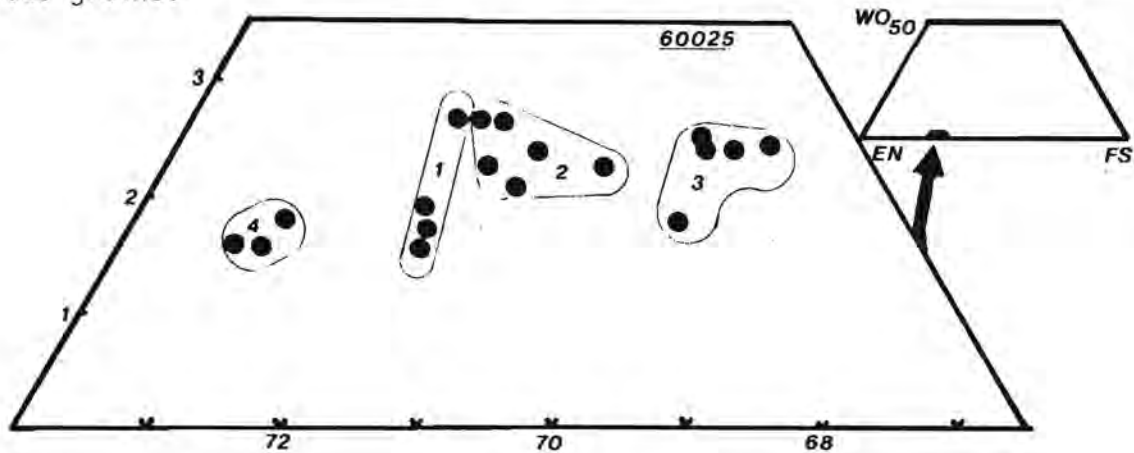


Figure 1. Compositions of low-Ca pyroxene grains picked from mafic clumps in ferroan anorthosite 60025. All contain exsolution lamellae of high-Ca pyroxene. All analyzed in same run, in order enumerated. Each grain is fairly homogeneous and distinct from the others.

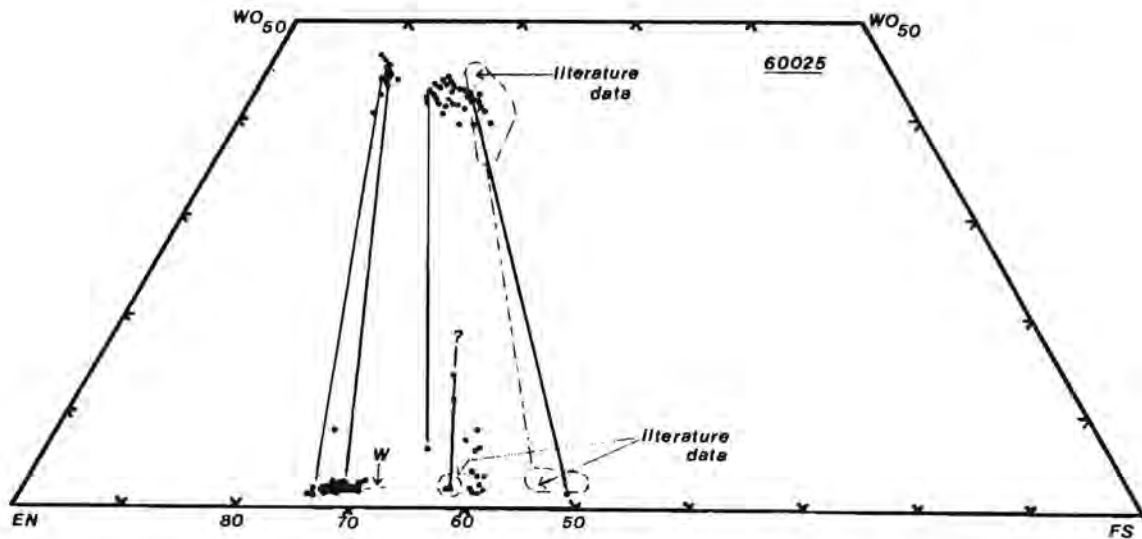


Figure 2. Total range of compositions of pyroxenes in ferroan anorthosite 60025, as analyzed in this study (dots), and literature data (Refs. 1-4). Line marked W is range given by Warren and Wasson (1978)(5) for low Ca pyroxenes in a thin section of the mafic mineral rich undocumented chip from 60025.

Ryder, G.

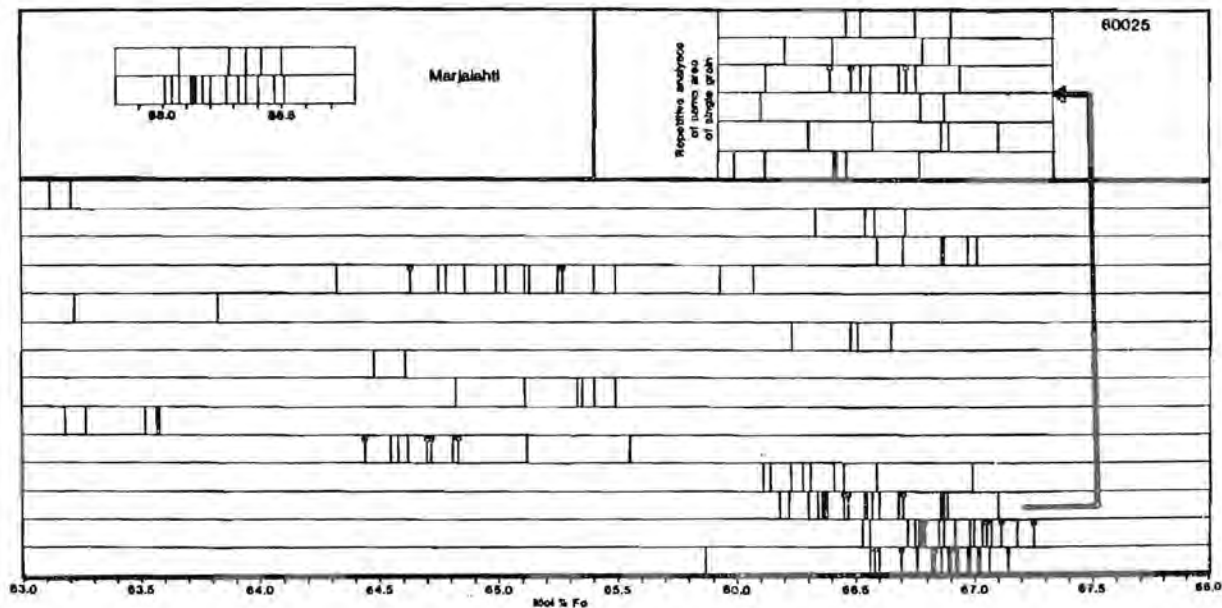


Figure 3. Compositions of olivine grains in ferroan anorthosite 60025, analyzed in this study. Bottom two lines are analyses of grains picked from the mafic clumps; all others are grains in a thin section of the mafic mineral rich undocumented chip. Warren and Wasson (1978)(5) report olivines $Fe_{63.6-64.8}$ with one small zone $Fe_{57.0-64.7}$, for this same chip. The top right inset shows repeated analyses, at separated periods of the analytical run, of the same area of a single grain. The top left inset shows two sets of analyses of Marjalahti (the standard) analyzed as an unknown. Points with the same symbol in a given line are consecutive analyses of a single point. The data demonstrate that individual grains are homogeneous, but that there are significant differences between grains. The most forsteritic grains are also the largest grains.

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LIMITS ON THE ORIGIN OF THE LUNAR MG-SUITE. Graham Ryder, Lunar Curatorial Lab., Northrop Services, Inc., P.O. Box 34416, Houston, TX 77034.

There is little doubt that a lunar Mg-rich plutonic suite, mainly comprising norites and troctolites, is distinct from the ferroan anorthosite suite (1,2,3). However, the relationships among members of the Mg-suite and the characteristics of their parental liquids still require clarification. The ultimate role of the Mg-suite in the early crustal evolution is conjectural: of primary concern is whether the Mg-suite samples had a single, Moon-wide origin, or whether they comprise the relics of more than one pluton. This abstract outlines some factors which limit concepts of the petrogenesis of the Mg-suite, but which have previously been inadequately considered.

Abundance: Pristine igneous highlands samples (2) consist of ferroan anorthosites, Mg-suite members, a few KREEP samples, and a small number of samples which cannot be assigned to any particular group. The Mg-suite forms almost half of this collection (2,4) and is dominated by several norites, most of which are from meter-sized boulders and clasts at the Apollo 17 site. Furthermore, the chemical compositions of highlands breccias are consistent with, and suggestive of, norite at least being both ubiquitous and common as a component of the highlands (5).

In contrast, the total volume of Mg-suite material may not be a very large proportion of the total crust. The calculated parent liquids for the norites (6,7) contain more than 100x chondritic abundances of rare earth elements (REEs). If the bulk Moon contains twice chondritic REEs (a reasonable estimate) then these liquids were less than 2% of the Moon (given that all lunar REEs were not in these liquids then the amount is actually even less). This is equivalent to a Moon-wide shell with a thickness of <12 km measured from the surface. The amount of norite would then be some factor less than this. So far the fate of the residual, high REE liquid has not been identified. The parent liquids of a dunite (questionably Mg-suite) and some troctolites, with less than 20x chondritic REEs (8,9) could be equivalent to the entire lunar crust, but these rocks are much less abundant than norites among the samples. In fact the assignment of much of the crustal budget of REEs to KREEP requires the actual abundances of these samples to be quite small, unless some relationship between KREEP and Mg-suite parental liquids can be established.

Location: Most of the pristine Mg-suite samples are from the Apollo 17 site, and appear to have been a common clastic constituent of the Serenitatis melt sheet. Some are from the Apollo 15 site (possibly from the Imbrium melt sheet), and some rare pieces, mainly spinel troctolites, are from the Apollo 16 site. In contrast, the Serenitatis melt sheet is devoid of ferroan anorthosite, and in fact no ferroan anorthosite *sensu stricto* has been identified in published studies of the Apollo 17 samples, nor is anorthosite required in mixing models of the orbital data from around that site (10). Ferroan anorthosites are common at the Apollo 16 site, and rare at the Apollo 15 site. This implies that ferroan anorthosites and the Mg-suite are not generally intermixed in the lunar crust on a scale of kilometers. Ryder and Wood (11) took this as evidence that the lunar crust had anorthosites on top and Mg-suite rocks below, but in fact this cannot be confirmed. Lateral variations are possible, with the real probability of a lack of anorthosite in the crust around the Apollo 17 site, either above, below, or with the Mg-suite. Further sampling of the highlands, or at least a considerable improvement in orbital geochemical data, is probably required to resolve the regional, lateral variations.

Ti/Sm: The low Ti/Sm (<1000) of the Mg-suite members contrasts sharply with chondritic, and lunar mare and anorthositic ratios (~4000) (3,12,13).

Ryder, G.

This characteristic appears to have no correlary among other solar system samples other than lunar KREEP which is probably in some way related to the Mg-suite. The low Ti/Sm requires substantial differentiation involving the separation of a Ti-rich phase, presumably ilmenite. This is difficult to reconcile directly with the high Mg' (100x molar Mg/(Mg+Fe) of 70-90 of the Mg-suite; hence mixing of a small proportion of a highly evolved liquid with a large proportion of a primitive liquid has been suggested (14, and others). Even if this is the case, the derivation of the low Ti/Sm evolved liquid is unclear. The Kiglapait intrusion did not evolve to such low Ti/Sm liquids, only to ~ 1500 , at Mg' <20 (S.A. Morse, pers. comm.); however, the main decline in Ti/Sm is caused by oxide phases and tends credence to the concept that ilmenite separation in the lunar case might produce adequately low Ti/Sm ratios in an evolved liquid.

Plagioclase enrichment: Norites generally have cotectic proportions of mafic minerals and plagioclase, but some are plagioclase-enriched. Nearly all troctolites and spinel troctolites have plagioclase in excess of cotectic proportions. On average then the Mg-suite is plagioclase-enriched. This suggests that there are complementary mafic mineral (particularly olivine) enriched plutonic cumulates which have not been sampled. The mafic mineral enriched cumulates, dunite 72415-7 and feldspathic lherzolite 67667, do not appear to be a part of the main Mg-suite according to trace element considerations (e.g. 3).

Regularity of sequence: The sequence from spinel troctolite, through troctolite, to norite, is marked by a reasonable regularity of decreasing Mg' and increasing incompatible element abundances in the samples and calculated parent liquids. The trend extrapolates to pristine KREEP materials, most of which, however, appear to have crystallized at 4.0 b.y. rather than the 4.2-4.6 b.y. which encompasses the Mg-suite ages. Among norites themselves, those with higher REEs have lower Mg'. This certainly suggests some kind of genetic relationship, but it cannot be shown that the relationship is very direct. For instance, calculated parent liquids for troctolite 76535 have REEs less than 20x chondritic (9), while REEs in the calculated parent liquid for norite 15455 are greater than 100x chondrites (6); yet the corresponding Mg' variation is only 88 to 82, inconsistent with 80% crystal fractionation.

Conclusions: Several pieces of data are inconsistent with the proposals (14,15) that the Mg-suite rocks are from a set of discrete plutons which intruded a ferroan anorthosite crust, viz. 1) ferroan anorthosites are not found in association with Mg-suite rocks, such as one might expect from impact excavation if they were proximal. The data are perhaps compatible with underplating or overplating of anorthosite with Mg-suite rocks. 2) The presence of plagioclase-enriched Mg-suite rocks and a general lack of complementary mafic mineral enriched cumulates suggests that the lower parts of the igneous body or bodies which make up the Mg-suite have not been reached by impacts. For a series of nested plutons it is unlikely that the lower parts of some were not reached. 3) The regularity of chemistry and rock type are inconsistent with independent liquids, each required to be contaminated or mixed with evolved liquids to the same fortuitous extent. If independent plutons are invoked, still a single liquid must be the source for all these.

The regularities of mineralogy and chemistry of Mg-suite samples which suggest their genetic relationship cannot be explained by simple fractional crystallization of a single starting liquid. Thus we appear to be forced to postulate a complex event which was common for all Mg-suite samples, implying a large scale event. The event requires progressive mixing with, or assimilation of, evolved liquids and a differentiating primitive liquid, or the magma may even have been zoned with respect to trace elements. This

Ryder, G.

differentiating liquid was not tapped at successive stages to produce small plutons, for if so there would probably not be the general characteristic of plagioclase-enrichment. The Mg-suite-producing event might have been Moon-wide, or it could have been restricted to the fairly small portion of the Moon from which the samples were taken. This discussion is not, however, an attempt to define the origin of the Mg-suite but to show some of the constraints which must be, and have not yet been, met by hypotheses.

Finally, the great abundance of Mg-suite samples among the pristine rocks and in lunar breccias, in contrast with the total volume which they compose assuming a chondritic or near-chondritic Moon, suggests that the depth of impact excavation on the Moon has not been great--certainly the Moon might only rarely have had impacts excavating material from deeper than 15 kilometers. If so, then reconstructions of the lunar crust using chemical and petrological considerations are based only on the upper 25% or so of it. What is the material which has not been sampled?

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PRISTINE LUNAR HIGHLAND ROCKS: HYPOTHESES OF ORIGIN, G. Jeffrey Taylor, Inst. of Meteoritics and Dept. of Geology, Univ. of New Mexico, Albuquerque, NM 87131.

Understanding the petrogenesis of the lunar highlands is inherently more difficult than understanding the crystallization of a terrestrial layered intrusive. Not only must one consider the myriad of igneous processes that operated, many of which are poorly understood, but one must also grapple with the following problems:

1) A nearly total lack of field control--No lunar samples were collected in place. Instead, they were picked up from the chaotic regolith that covers the moon's surface or chipped from large boulders that sit in the regolith.

2) Poor sampling--It is not clear that we have sampled all the prominent types of rocks that make up the lunar highland crust. Even if we have representatives of all major lithologies, we do not know if we have them in any semblance of their true relative abundance. This situation is due in part to the limited number, duration and spacial extent of expeditions to the moon and in part to a period of intense meteorite bombardment that transformed much of the accessible lunar crust into a series of complex breccias and impact-melt rocks.

3) The scale of the processes--Although many models for the origin of the moon's crust have been proposed, all consider processes operating on a global scale.

4) "Planetary" processes--Some processes are unique to planetary formation. The most important of these may be continued meteorite impact during the waning stages of accretion. Loss of volatile elements and separation of metallic or sulfide phases may also be important.

These complexities have not stopped lunar scientists from courageously trying to determine how the lunar highlands formed. Attention has been focussed on four main groups of rocks: ferroan anorthosites, the Mg-suite (mostly norites and troctolites), KREEP, and (to a lesser extent) the source regions of mare basalts. See reviews in this volume by P. Warren and by D. Blanchard for details of the petrology and chemistry of the first three types. The origins of mare basalts is still debated, especially with regard to the extent to which their primary magmas were affected by assimilation, but there is general agreement that they were derived by partial melting at depths of ~ 100 to ~ 300 km(1), possibly as deep as 500 km(2), of ultramafic rocks consisting of olivine and orthopyroxene with variable amounts of clinopyroxene and/or ilmenite and/or plagioclase. All four groups of rocks formed early in lunar history (see below). The question is, how are they related to one another, if at all, and how did they form? Attempts to answer this question led to two main categories of hypotheses on highland rock petrogenesis: those evoking the formation of a globe-encircling magma ocean (or total melting of the moon) from which some or all of the highland crustal rocks crystallized and those that do not evoke a magma ocean.

MAGMA OCEANOGRAPHY

The concept that the outer portions of the moon were molten very early in its history was originally developed by Wood et al (3) and Smith et al (4) to explain the apparent overabundance of plagioclase in the lunar highlands. These authors postulated that as the extensive magma system crystallized, olivine and pyroxene sank and plagioclase floated to form the anorthositic crust. The idea is a testament to scientific imagination because it was sparked by the presence of a small percentage of millimeter-sized anorthositic

G. Jeffrey Taylor

rock fragments in the soil returned by Apollo 11 from a mare region of the moon. As detailed below, the concept has really taken hold in lunar science and has begun to be applied to other planets, including the earth (e.g., 5).

Evidence for a magma ocean

Several lines of evidence indicate that the moon experienced a major differentiation event within ~ 200 Myr of its formation. The simplest way to explain these data is to postulate a global magma ocean. The most compelling arguments are the following:

1) Plagioclase-rich highlands crust—The average crust in the highlands contains ~ 25 wt% Al_2O_3 , or $\sim 70\%$ normative plagioclase and is ~ 60 km thick (1,6). It has a fractionated (light-enriched) REE pattern with a positive Eu anomaly. These data indicate that plagioclase has been concentrated into the crust. In fact, if the Mg-suite actually postdates the formation of ferroan anorthosites (7-11), then the latter represents the original lunar crust, which, therefore, must have been even richer in plagioclase than the present highlands crust. As suggested on the basis of the Apollo 11 anorthosites (3,4), such an enrichment was probably caused by a large-scale differentiation of the moon.

2) Complementary Eu anomalies—Most mare basalts have prominent negative Eu anomalies, a property they inherited from their source regions (12,13). Because the source regions have negative Eu anomalies, plagioclase may have been subtracted from them. A straight-forward way of doing that is to form the mare basalt source regions as sunken mafic cumulates from a magma (ocean) that had begun to precipitate plagioclase. This simultaneously explains the positive Eu anomaly in the crust and the negative anomaly in the mare basalts.

3) Early fractionations—Rb-Sr data indicate that the mare basalt source regions experienced a fractionation event ~ 4.5 Gyr ago (14). Anorthosites also seem to record this event (14,15). Sm-Nd data on mare basalts confirmed the Rb-Sr data (16,17), pointing to a very early fractionation event. Sm-Nd work on KREEP (18) indicates that it, too, formed early in the moon's history, but perhaps as recently as ~ 4.4 Gyr. All these data are consistent with the idea that a significant magmatic event took place during the first ~ 200 Myr of lunar history. (Individual ages on pristine Mg-suite rocks, however, are a bit troublesome (see 19), ranging in age from 4.2 to 4.55 Gyr.)

4) Uniformity of KREEP—The isotopic systematics (18) and REE patterns (20) of KREEP are strikingly uniform from site to site on the moon. This suggests that KREEP formed by a global process.

Energy sources for magma ocean

There is no shortage of potential heat sources capable of causing widespread melting on the moon and the terrestrial planets. These have been summarized by Hostetler and Drake (21) and include enhanced solar luminosity, electromagnetic induction, shortlived radioactivities, adiabatic compression of nebular gases, tidal heating, and accretion. To this list one may add the energy provided if the moon formed by fission from the earth. All these mechanisms provide heat in the requisite short time scale, $< \sim 10^8$ yr, but it is not clear which one dominated or if all of them operating together caused wholesale melting of the outer several hundred kilometers of the moon. As discussed below, recent doubt about the efficacy of accretional heating has led to suggestions that the magma ocean never existed.

Magma ocean composition

There is no direct way of determining the composition of the magma ocean. The chill zone, if there was one, no longer exists intact (or at least we do not recognize it). We do not have a complete section through the former ocean, so we cannot compute the composition. Consequently, indirect methods must be

G. Jeffrey Taylor

used. The most commonly accepted assumption is that the ocean had the composition of the bulk moon. Whole-moon compositions have been estimated (guessed?) by numerous investigators (6,7, 22-28; ref. 7 contains a summary), generally using element ratios and cosmochemical principles. The compositions range widely; Al_2O_3 , for example, ranges from 2.7 wt% to 27wt%. However, most recent estimates are ultramafic with Al_2O_3 in the range 3 to 8 wt% and $Mg/(Mg+Fe)$ of 0.8 to 0.9. REE and other refractory trace elements are unfractured relative to chondrites, but somewhat enriched in total concentration.

Size of magma ocean

The depth to which total melting occurred when the moon formed is poorly constrained. Estimates range from ~ 100 km to the entire moon. Geochemical and geophysical constraints (summarized in ref. 29) give a reasonable estimate of ~ 500 km ($2/3$ the volume of the moon) for the depth of the magma ocean. This value has been used in most recent models of magma ocean evolution.

Oceanography

Early models (3,4) of the magma ocean were relatively simple. When samples were returned from the highlands, however, the models had to become more complex. S. R. Taylor and co-workers developed some of the most elaborate, general models of magma ocean evolution (1,6,26,29). These models envisage crystallization of olivine followed by olivine and orthopyroxene. This leads to decreasing Ni and Mg/Fe in the cumulate pile at the base of the magma ocean. Continued crystallization leads to saturation of the magma with plagioclase and clinopyroxene. Plagioclase floats to the top to form the anorthositic crust. Throughout crystallization, the concentrations of REE and other incompatible elements increase in the residual magma, eventually forming a zone beneath the floated crust. The model accounts qualitatively for the source regions of mare basalts, the anorthositic crust, and KREEP (the Mg-suite is not explicitly considered) and provides a very useful framework in which to understand the evolution of the moon. A. B. Binder (30-34) also developed complex models for magma evolution, but considered total melting of a moon of pyrolite composition derived by fission from the earth.

A new era in magma oceanography began when Drake (35) modeled quantitatively the evolution of both major and trace elements during the crystallization of a magma ocean of whole-moon composition. Since then, detailed quantitative models have been developed by Longhi (28,36-38) and by Warren and Wasson (7), with contributions to various aspects of the subject by many others. Longhi and Warren have taken quite different approaches to the problem. Warren emphasizes the importance of pressure in a large magma body and his model oceans crystallize from the bottom up. Longhi, in contrast, emphasizes processes that might have operated at the surface of the magma system.

Several authors have drawn attention to the importance of the large pressure gradient in a large magma body (7,32,39-41). The basic idea is that a large, convecting body establishes an adiabatic temperature gradient. Because adiabats are less sensitive to pressure than silicate liquids (Fig. 1) crystallization takes place first at the base of the system (except for a rapidly-cooled zone at the surface). Because the pressure is greater at the base of the magma (24 kb in a 500-km deep magma ocean), the order of crystallization can change. Specifically, the olivine-pyroxene peritectic curve becomes cotectic and moves towards the olivine-plagioclase base in the olivine-plag-silica system (Fig. 2). Consequently, depending on the depth of the ocean (i.e., on the P) olivine is soon followed by co-crystallization of pyroxene. As crystallization proceeds, the pressure decreases and the liquid follows the path shown in Fig. 2. This is the crystallization behavior modeled by Warren and Wasson (7). Their calculations stop at the point of plagioclase

G. Jeffrey Taylor

saturation at which point the $Mg/(Mg+Fe)$ of olivines in equilibrium with the magma is appropriate for that found in ferroan anorthosites (depending on the initial composition of the magma ocean). Warren and Wasson specifically state that Mg-suite rocks cannot form from the magma ocean.

Longhi has tacitly acknowledged that crystallization could take place at the base of the ocean (37,38), but argues that crystallization of olivine and orthopyroxene will not produce the fractionated REE patterns that the parent magmas of the Mg-suite (which Longhi views as magma ocean products) and the ferroan anorthosites seem to have had (see review by Blanchard in this volume). If the parent magmas of pristine rocks had fractionated REE patterns, more complex processes are required. Consequently, Longhi has focussed attention on processes that may have operated at the surface of a moon-wide magma system. Three main categories of surface effects have been considered: formation of a quenched crust, formation of rockbergs, and the effects of continued accretionary impacts.

1) Quenched crust--There is no doubt that a conducting layer would have formed on top of the magma ocean. Without such a layer, the entire magma would crystallize in ~ 100 yrs (42). Although some thermal models incorporate a very large (~ 10 km) chilled margin (43), most models require that the quenched zone be small, tenths of a kilometer at most (44). The thickness is limited by the simple fact that the quenched layer is denser than the underlying magma and, therefore, will tend to founder and sink (42,44,45). Breaking and foundering would be aided by meteorite impacts and by vigorous convection in the magma. The sunken material would begin to remelt as it descended, thus serving as an effective means of heat transport in the magma system (45). Chunks of frozen crust that sank all the way to the base of the magma could be trapped in the cumulate pile, making these basal layers less barren of radioactive, heat-producing elements (and other incompatible elements as well), a necessary requirement to form mare basalts 0.6 to 1.5 Gyr later (43).

2) Rockbergs--Chronological data suggest it took about 200 Myr for the magma ocean to solidify (see above). Continued foundering of a thin quenched crust leads to solidification times at least an order of magnitude shorter than this (42,44). Consequently, at some stage during the crystallization of the magma ocean, a thicker insulating crust is needed. This crust could form when plagioclase begins to precipitate after it has saturated in the magma ocean (44). Herbert et al (42) provided another mechanism to produce this crust early in the evolution of the magma ocean: the formation of rockbergs by the precipitation and floatation of plagioclase near the surface long before plagioclase is saturated in the main body of magma. The idea is that the foundering and remelting of the crust weakens convection currents at shallow depths and thereby produces a steep (nonadiabatic) temperature gradient near the surface. This leads to cooling of surface layers and early crystallization of plagioclase. The plagioclase floats (while denser mafics sink) and concentrates over the cool areas above descending convection currents. The ultimate result is rockbergs of anorthosite.

3) Bombardment--Hartmann (46) has drawn attention to the possibly profound effects that meteorite impact could have had on the evolving magma ocean and highlands crust. Although the role of impact has been recognized in many models, its petrologic effects have not been investigated in detail. Models usually call upon impacts to smash up the quenched crust, but impacts may also destroy significant amounts of the early anorthositic crust, forming breccias composed of anorthosites and quenched magma-ocean liquid.

Longhi (36-38) has incorporated these surface processes into progressively more complicated models. The models emphasize the need to

G. Jeffrey Taylor

produce liquids that crystallize plagioclase before pyroxenes (necessary for forming troctolites) and that have fractionated REE patterns. Note that Longhi explicitly assumes that the Mg-suite formed in the magma ocean. The essential ingredients of Longhi's models are the formation of anorthosite rockbergs and areas near the surface choked with plagioclase. The plagioclase-rich suspensions undergo extensive fractional crystallization (Longhi has modeled it up to 97% crystallization), then mix with relatively fresh liquid from the main body of the magma ocean. This causes remelting of the plagioclase. The new mixed liquid then fractionates extensively and mixes again (Fig. 3); this process continues for several cycles. Longhi envisages similar processes happening beneath the growing feldspathic crust to give rise to most Mg-suite rocks. His model, in short, involves fractional crystallization, assimilation, and magma mixing, all taking place at the surface of an evolving magma ocean and growing crust.

The quantitative models developed by Longhi and by Warren and Wasson differ in their emphasis, the former presuming that most of the action takes place at the surface of the magma ocean, the latter suggesting that the real action is at the base of the system. The two approaches, however, are not incompatible: A magma ocean could crystallize from the bottom up while forming rockbergs and undergoing complex fractionation and mixing processes at the surface. It seems clear that crystallization would take place at the base of a magma ocean. The fundamental question is, how efficient were the surface processes?

MODELS NOT INVOLVING A MAGMA OCEAN

Many authors have subscribed to the view that the ferroan anorthosites and the Mg-suite are unrelated. Most allow the anorthosites to be products of the magma ocean, but suggest that the Mg-suite derived from later intrusions into the anorthositic crust (7-11,34). These magmas are proposed to have derived from partial melting of either basal cumulates laid down in the magma ocean or of underlying primitive lunar material. If derived in this way the parent magmas to the norites and troctolites may have experienced varying degrees of assimilation of residual magma (rich in incompatible elements) from the magma ocean. These models have not yet been examined quantitatively.

A new school of thought has cropped up relatively recently. It holds that there was no magma ocean. This dreadful development stems from Wetherill's arguments (47-50) that the moon grew by accretion of large (~ 100 km), relatively slow (6-10 km/sec) planetesimals during a period of ~ 100 Myr. The petrologic consequences of this have not been investigated quantitatively. Wetherill (47) suggested that the result would be a series of layered plutons throughout the outer regions of the moon. Simonds (51,52) examined the consequences of slow accretion of large bodies from the standpoint of the effects of impact processes and concluded that the moon would differentiate largely by the formation and subsequent fractionation of partial, rather than total, melts. J. V. Smith (53,54), while not disarding the magma ocean entirely, has recently added another twist. He suggests that the ferroan anorthosites formed from a magma ocean surrounding a proto-moon (about half its present volume); the Mg-suite and mare basalt source regions formed by partial melting and assimilation during subsequent accretion of the other half of the moon's volume. It is not clear that any of these complicated stories can be modeled quantitatively.

G. Jeffrey Taylor

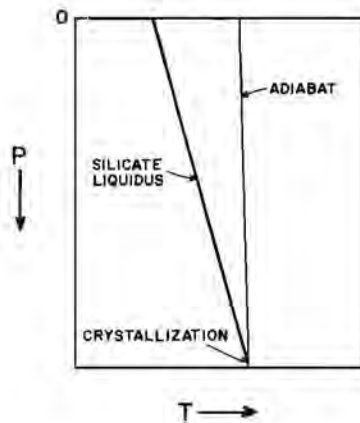


Fig. 1

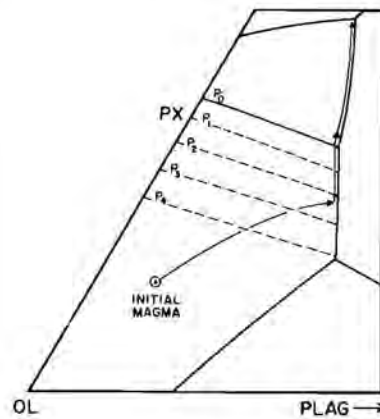


Fig. 2

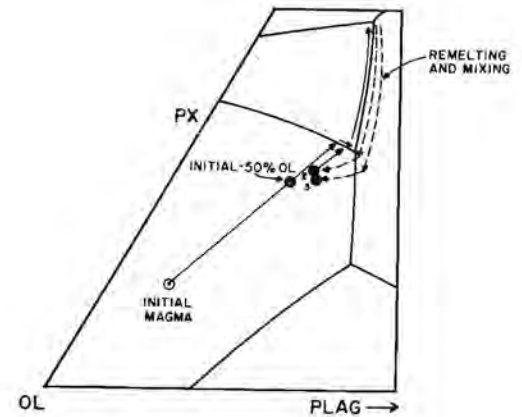


Fig. 3

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G. Jeffrey Taylor

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MONOMINERALICITY AND OIKOCRYSTS: KEYS TO CUMULUS COOLING RATES?

Dr. G.C. Ulmer, Geol. Dept., Temple Univ., Philadelphia, Pa. 19122

Dr. D.P. Gold, Dept. Geosciences, Penn State Univ., University Park, Pa. 16802

Initial accumulation of typical-sized igneous crystals (approx. 0.1 mm to 10 mm) produces layers with substantial interstitial magma (approx. 35% to 20%, respectively). However, monomineralic cumulate rocks with as little as 4% gangue mineral are known. The original magma trapped in the interstices must have been reduced in volume by two main processes:

- (1) more of the accumulating monomineralic phase may have crystallized epitaxially from the trapped magma (adcumulus)
- (2) some of the interstitial magma must have been physically excluded from the forming layer.

Even if mechanism (1) accounts for half the volume of the originally trapped interstitial magma, the rest of the volume of that magma must be physically removed from the forming layer. Overburden pressures, sink/float density mechanisms, and diffusion controlled sintering must all play a role in excluding a volume of liquid large enough to produce 96% monomineralic rocks. By these gradualistic processes the interstitial magma (Figure 1) diminishes to intersertal pools (Figure 2) which in turn become vanishingly small as the dihedral angles of contact between any three adjacent cumulate grains approaches the 120° limit (Figure 3). If solidification occurs too quickly, monomineralicity will be only poorly developed. Even in the 'infamous' inch scale layering of the Stillwater, the monomineralicity is not as well developed as in similar horizons of the Bushveld. This may suggest that the Stillwater was more shallowly emplaced and cooled too quickly for well developed monomineralicity.

Oikocrysts of optically continuous opx upto tens of centimeters in size are common in the Stillwater, but not in the Bushveld. As seen in Stillwater outcrops, it is often clear that the rocks surrounding big opx oiks are devoid of interstitial opx for several cubic meters, as though all the opx accumulated into just a few point locations. We, therefore wonder if these oikocrysts may be large pockets of "gangue" composition that coalesced from the small and numerous intersertal pools in a still mushy layer as the layer was attempting to become more monomineralic. If total solidification of the layer occurs before the spatial purge of this "gangue" material is finished, then in this model the "gangue" material crystallizes to oikocrysts.

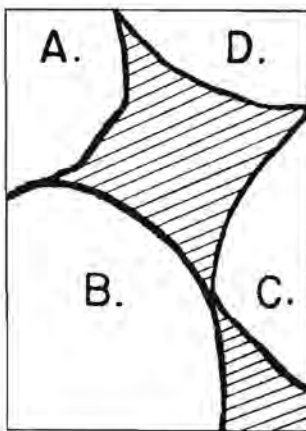


FIGURE 1

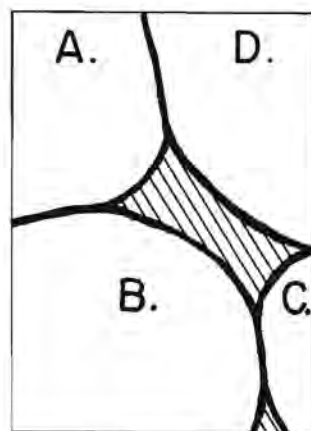


FIGURE 2

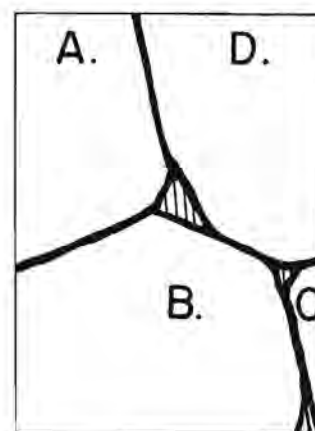


FIGURE 3

PETROLOGY OF PRISTINE NONMARE ROCKS, SURVIVORS OF THE LUNAR HIGHLANDS BOMBARDMENT. Paul H. Warren, Dept. of Geology and Institute of Meteoritics, University of New Mexico, Albuquerque, NM 87131.

Introduction: What they are, how they are identified

"Pristine" in the context of lunar petrology means: produced by endogenous (internal) processes. The term is generally used in a purely compositional sense. On the Moon's highlands, pristine material is rare. Due to its relatively small size, the Moon essentially ceased to produce new rocks, except for volumetrically unimportant basalts, even before it had quite completed accretion, the process by which planets grew during the first ~600 Ma of the solar system, by absorbing smaller bodies. Accretion involved tremendously violent collisions, and the Moon's surface suffered chaotic faulting, to the point of pulverization in places, and heating, to the point of remelting in places. Most of the upper crust was broken up and heated (not necessarily all by one impact), producing new rocks called impact breccias. Impact breccias tend to contain materials of very diverse provenance, particularly if material was subjected to multiple impacts, as was often the case on the highlands. Most of the impact breccias that cover the highlands are polymict (many-component) mixtures of materials that originally formed by separate endogenous processes, at separate places and separate times. It would be difficult to reconstruct what those endogenous processes were, if polymict breccias were our only sources for information.

Most pristine rocks are monomict (one-component-mixture) breccias, not even close to being texturally pristine. Only about ten nonmare samples are known that feature textures (as well as compositions) that might be sanctioned with the "pristine" encomium. A few of these are shown in Fig. 1. In most cases where nonmare rocks show vestiges of pristine textures, the texture is plutonic (i.e., coarse grained, either igneous or granulitic; other signs might be symplectites, coarse exsolution lamellae, high mineral-chemistry geobarometer readings, etc.). This may be because originally volcanic rocks were obviously in unfavorable (shallow) positions to survive the late accretionary bombardment, but it is also probable that the crust is simply richer in originally plutonic material. This is in a sense fortunate, because polymict breccias tend to be fine-grained, and if they appear igneous at all, they tend to appear volcanic. Pristine volcanic rocks are extremely difficult (some might say impossible) to distinguish from clast-free impact melt breccias on a textural basis. However, there is general agreement (except for [1]) that plutonic textures are unlikely to have resulted from any non-endogenous process.

Distinguishing highly cataclastic monomict breccias from polymict breccias can be a tricky business. Usually, polymict breccias contain sharply-bounded "clasts" or "lithic fragments" of material clearly unrelated to the rest of the rock. Occasionally such clasts are pristine, but unless every clast is compositionally alike, and the same as the groundmass, the rock in toto is probably nonpristine. Another method for distinguishing between monomict and polymict breccias is to check for telltale mineralogy. For instance, a monomict rock containing Fo₉₀ olivine will generally not also contain quartz, or even olivine that is much more Fe-rich than Fo₈₀; a rock with plag ranging only from An₉₃ to An₉₄ is generally more likely monomict than one with plag ranging from An₈₀ to An₉₈; and so on. Native Fe-metal is extremely useful as a pristinity indicator. Lunar polymict breccias are products of impacts of meteoritic bodies with the Moon. Most meteorites are rich in Fe-metal that contains 6-7% Ni and 0.3-0.4% Co. Polymict breccias are generally much richer in metal than monomict ones, and the metals in polymict breccias are usually close in composition to meteoritic metal. Among pristine rocks, where metal is present at all it tends to feature much higher Co/Ni, or similar Co/Ni but much lower or higher Ni, compared to meteoritic metal.

Paul Warren

The petrographic approach to distinguishing between monomict (pristine) breccias and polymict ones usually "works" inasmuch as most polymict breccias can be spotted very easily, and most monomict breccias can be distinguished with reasonable confidence. Ambiguities are rare, and they can generally be resolved if a bulk analysis of the material can be performed for siderophile trace elements. Most meteorites are greatly enriched in siderophile elements compared to almost all pristine lunar crustal rocks. Hence, an excellent way of distinguishing pristine material is to check bulk siderophile concentrations. Trace elements are beyond the scope of this review; see [2] for details. Suffice it to say that a low content of siderophiles is usually the best evidence that a sample is pristine.

Cumulates

The popular nomenclature of mafic rocks that are "results of gravity acting on a crystallizing magma," stems mainly from Wager and Brown [3], who labeled such rocks cumulates, a term that obviously connotes formation via crystal settling. Actually, a great deal of the rhythmic layering that was once attributed to differential crystal setting is probably caused by oscillatory nucleation and diffusion-controlled crystallization [e.g., 4, 5, 6]. Another problem with the nomenclature of [3] is that it enhances the common misconception that most cumulates are formed with ~35 % interstitial "trapped liquid." Actually, as [3] themselves emphasized (p. 554), most cumulates are poor in trapped liquid; they contain far more material of cumulus composition than the ~65% expected according to the once orthodox concept of settled crystals. Having given these caveats, I will use the nomenclature of [3].

Although most of the tiny Apollo-11 soil fragments described as cumulates by [7] in 1970 are probably not even pristine, there was justification for [7] to conclude that most nonmare materials were originally cumulates. The average nonmare rock has ~75% plag. Bowen [8] pointed out that we rarely find anorthosites that represent quenched liquids, and moreover that exceptionally high temperatures have to be reached to produce such a liquid. This reasoning led Bowen to invoke crystal "accumulation," and it still applies, even on the Moon. Since the time of [7] a number of pristine nonmare rocks with clear vestiges of igneous textures have been described, and most of them, particularly the volumetrically predominant anorthosites, norites and troctolites, show relict cumulate textures. (From here on I will be citing data and descriptions for specific, numbered pristine samples. Unless specified, these are from the compilations of Ryder and Norman [9, 10], or at least from references mentioned prominently therein.)

Terrestrial petrologists are accustomed to recognizing cumulates mainly by their rhythmic and/or cryptic layering. Unfortunately, even rhythmic layers are generally far too thick to be seen within individual lunar samples (typically 1 to 10 cm), and none of the nonmare samples was taken from an outcrop such that its "cryptic" position in its intrusion is known. We must resort to other methods to recognize lunar cumulates. Adcumulates (those with very little trapped liquid) are often recognizable by their simple mineralogy, e.g., almost pure plag, or pure plag plus one mafic silicate. In contrast, complete in situ crystallization of silicate melts, including trapped liquids, generally produces a broad spectrum of minor and accessory phases, not to mention zoning of the earliest crystals (which might be erased by subsequent annealing). Simple mineralogy might also result from anatexis, but in such a case the texture would be distinctly granulitic. Coarse granularity suggests a cumulate origin. Norite 78235 has plag primocrysts as large as 10x7 mm. Dunite cumulate 72415 has olivines up to 10 mm across. Anorthositic norite cumulate 15455c has crystals up to 5 mm across. Most lunar cumulates are not quite so coarse-grained (e.g., gabbro 61224,6; anorthosites 62236, 62237 and 67915c [11]; troctolite 67435c [12]), but then neither are most terrestrial cumulates

[13]. Not all of the coarsest pristine rocks feature cumulate textures: some (e.g., anorthosite 15415; troctolite 76535) are coarse granulites. It is very commonly assumed that such samples were originally (before thermal metamorphism) cumulates, but this is quite an extrapolation for a polymineralic rock such as 76535. Cumulates should generally have grains at least as coarse as 1 mm [13], but coarse granularity alone is hardly unambiguous proof of cumulate origin.

A more generally applicable method for recognizing lunar cumulates is to look for a pattern of euhedral, usually equant, primocrysts set amidst anhedral "poikilitic" material [3]. If the rock is an adcumulate, the poikilitic material will be identical in composition to the primocrysts, unless the rock is a heteradcumulate, in which case the poikilitic material will be several different phases, but also uniform in composition. If the rock is an orthocumulate (rich in trapped liquid), the poikilitic material will be many phases of variable composition. Sometimes (Fig. 1) this method works well. However, judging from terrestrial cumulate textures [3], it is generally not so easy to tell "primocrysts" from "poikilitic" material.

A few pristine nonmare rocks are definitely not cumulates. The rare KREEPy ones are all volcanic, except for one (15405c) that is hypabyssal. However, for virtually all of the other pristine nonmare rocks the evidence is at least wholly consistent with ultimate genesis as cumulates. The volcanic ones have been reviewed by [14] and [15].

Mineralogy

Pristine nonmare rocks are composed of the same restricted set of anhydrous, ferric-iron-free minerals as nonpristine nonmare rocks. The best evidence regarding the composition of the nonmare crust is obtained from the quintessentially nonpristine highlands soils. Calcic plagioclase is clearly the most abundant mineral (~75% at the Apollo-16 and Luna-20 sites; even higher in the farside crust, based on remote sensing data). Besides plagioclase, all but one or two percent of most nonmare rocks is orthopyroxene and/or olivine (at Apollo 16 orthopyroxene is preponderant; however at Luna 20 olivine is at least equally abundant). High-Ca pyroxene is only occasionally a major constituent (e.g., 61224,6; 67667; 67915c; 76255c), but most anorthosites have small amounts of inverted pigeonite, now orthopyroxene rich with high-Ca pyroxene blebs and lamellae. Ilmenite is ubiquitous in trace-to-accessory amounts. Native Fe-metal is almost ubiquitous, but never in more than trace amounts, among pristine rocks. Many of the noritic and KREEPy rocks contain trace-to-minor amounts of K-feldspar, silica polymorphs, zircon, apatite and whitlockite. Chromite is a very common trace constituent of anorthosites, and Mg-rich "pink" spinel is a common trace-to-minor phase in troctolites.

A milestone in lunar petrology was the discovery by [16] (following a suggestion by [17]) of a dichotomy among pristine nonmare rocks based on silicate compositions (Fig. 2). For terrestrial cumulates, a plot analogous to Fig. 2 generally shows a good correlation between mg ($\text{Mg}/(\text{Mg}+\text{Fe})$) and An, along lines with slope ~ 2.0 . Treated as a single group, pristine nonmare rocks show almost no correlation, and the slope is negative, if anything. This puzzled petrologists for several years, and some imaginative models were developed to try to explain it. The real problem was that not enough pristine rocks were known yet. When [16] found two pristine clasts with comparatively Na-rich plagioclase, the dichotomy finally became manifest: Roughly half of the pristine rocks belong to a class dubbed the "Mg-rich" group by [16], that shows a "normal" correlation on Fig. 2 with slope ~ 2.0 . The other half belong to a separate, apparently unrelated group characterized by uniformly calcic plagioclase but relatively low-mg mafics, called the "ferroan" group after [18], who first emphasized that they are "distinctive." Incidentally, [19] found that the Stillwater intrusion cumulates distribute into two trends very similar to the

ones in Fig. 2. There is no gap separating the Stillwater groups, however, and [19] reject the hypothesis that the Stillwater data imply that the two lunar groups are actually from a single magma.

The mineral-chemistry dichotomy is matched by a dichotomy of modal plagioclase content. The annotations in Fig. 2 are largely self-explanatory: All ferroan samples large enough to give meaningful modes contain at least 83% plagioclase, all "large" Mg-rich samples contain no more than 77% plagioclase. Not shown in Fig. 2 are 8 exceptionally Na-rich samples. Five of these are "alkali-rich" anorthositic rocks [20, 21], which have Fe-rich ($\text{mg} = 41-70$) mafics, but An_{79-88} plagioclase. Alkali-rich anorthositic rocks have been found only in Apollo-12 and -14 samples. They probably represent either an entirely separate group, or else a type of Mg-rich anorthositic rock; they are certainly not closely related to the ferroan anorthositic rocks [21].

Only pristine rocks are shown in Fig. 2. Nonpristine rocks tend to plot in the gap between the trends [22], just as one would expect, assuming they are mixtures of pristine rocks. This assumption is generally accepted today, but the thoroughness and pervasiveness of the mixing was never apparent until ~1977, when enough pristine rocks had been discovered to manifest their great diversity in comparison to nonpristine rocks. Until then, petrologists could only assume that nonpristine rocks had at least usually "inherited" compositions directly from "precursor" pristine rocks. Interpreting nonpristine samples in this way led to erroneous notions that soil-like compositions, namely anorthositic norite and "low-K Fram Mauro," constitute most of the pristine crust. Actually, these compositions are both very rare among genuine pristine rocks [23].

Crystallization processes

In part because the abundance of anorthositic norite was greatly overestimated, many of the early models for producing the nonmare crust (and a few recent ones, too) assume most pristine rocks to be orthocumulates—50-70% cumulus plagioclase plus 30-50% trapped liquid—that floated over a magma ocean. It now seems clear, based on the evidence of Fig. 2, that none of the Mg-rich pristine rocks could ever have floated over any reasonable-density magma [24]. This leaves the ferroan anorthositic rocks as the only widespread group feldspathic enough to be floated ocean cumulates. There is no strong evidence that the anorthositic rocks did not form this way; in fact, they are precisely the sort of rocks we expect to form the "upper border group" of the ocean, based on a fairly simple theoretical model plus the reasonable assumption that the silicate part of the Moon is roughly equal in composition to the silicate part of nearby Earth [25].

None of the pristine rocks appear to contain orthocumulate amounts of trapped liquid, least of all the anorthositic rocks. Fig. 3 shows MgO vs. Sm for ferroan samples, and a mixing line for pure plagioclase plus trapped liquid. The composition assumed for trapped liquid (5% MgO, 0.8 $\mu\text{g/g}$ Sm) has conservatively low Sm (or else high MgO), so some samples, particularly 15465c, plot below the line. The four most MgO-rich anorthositic rocks all plot far above the line; i.e., they all contain too little Sm to be consistent with a trapped liquid origin for more than small fractions of their mafics. Except for the three in the lower right corner (66035c2, 67455,32 and 15465c), all these samples are probably adcumulates.

Mg-rich pristine rocks generally contain only two or three important minerals; they too are probably mostly adcumulates. Assuming that the trend in Fig. 2 was produced by fractional crystallization of their parental magma(s), the magma(s) initially precipitated plagioclase + olivine \pm Mg-spinel \pm opx. Then came a stage of plagioclase + opx, followed by plagioclase + opx + high-Ca pyroxene. Further evolution was probably complex, and not uniform everywhere in the crust. No corresponding sequence can be inferred for the ferroan anorthositic parent

magma(s), but the most mafic-rich ones (15437, 62236, 62237, and especially 60025) indicate that plagioclase, olivine and pigeonite were all cumulus phases, simultaneously.

"Field" geology; inferred stratigraphy

Field geology is almost nonexistent for pristine nonmare rocks, but not quite. Remote sensing data show that incompatible elements in the crust are intensely concentrated in the longitude range 0-50°W. Most of the landing sites are east of this zone, but the Apollo-12 and -14 sites are close to its center. According to several recent models [24, 26, 27], the Mg-rich rocks formed in layered intrusions emplaced in the crust shortly after the magma ocean finished solidifying. It was noted by [24] that much "urKREEP" residuum from the ocean would probably have been assimilated by the parental Mg-rich magmas, in the vicinity of the crust-mantle boundary. The incompatible element enrichments in the 0-50°W region are probably due almost entirely to enrichments in KREEP, but recently [21] showed that even Mg-rich (plutonic) pristine rocks from Apollo 12 and 14 feature huge incompatible enrichments, compared to their eastern counterparts. This observation, which was implicitly predicted by [24], militates against any model that assumes Mg-rich rocks are all products of one magma, even if it was an "ocean".

Another important way that the landing sites differ is in proximity to impact basins. For our purposes, we need only divide the Apollo sites into three groups: the Apollo-11, -15 and -17 sites are all within 1.25 crater radii of the centers of basins; the Apollo-12 and -14 sites are both ~2.5 crater radii away; and the "true" highlands Apollo-16 site is 3.4 crater radii away [28]. In general, samples collected close to basins will more frequently be from deep in the crust than samples collected far from all basins. Bearing this in mind, an inventory of the distribution of the various pristine lithologies among the sites can provide a general picture of the crust's stratigraphy. For this purpose, larger samples are obviously somewhat more significant than smaller ones, and modes of samples smaller than ~3 g are bound to suffer from sampling errors; so they should almost be discounted. Also, samples that are not at least ~90%-probably pristine will not even be considered.

From the intermediate sites, the only pristine samples are all <3 g, and there is the complication that peculiarities of the group may be due more to the western longitude of the sites involved than to depth effects. For the record, the pristine samples from the intermediate group include: 2 or 3 alkali-rich anorthosites, one ferroan anorthosite, ~7 troctolites, one gabbro, one syenite, and one granite. From Apollo 16, all of the ~25 pristine, >3 g samples are ferroan anorthosites, except for 67667, which is a 7.9-g feldspathic ilmenite. The <3 g pristine samples include: numerous ferroan anorthosites, 3 or 4 troctolites, one clast that is technically a norite, but probably is actually an unrepresentative piece of a ferroan anorthosite, and one gabbro. From the sites right next to basins, the pristine, >3 g samples include: only two ferroan anorthosites, 4 or 5 troctolites, 7 or 8 norites, 3 KREEP basalts, and one dunite. The <3 g samples include: 4 ferroan anorthosites, 4 or 5 troctolites, 2 or 3 norites, numerous KREEP basalts (all closely related to two of the >3 g KREEP basalts), one gabbro, and one quartz monzodiorite.

This distribution indicates that ferroan anorthosite is the dominant pristine rock type in the upper crust, but lower down (it is difficult to say how far down; perhaps 20 km--depending on how deep the basins excavated, which is very poorly known) the Mg-rich types, especially norite and troctolite, begin to predominate. This stratigraphy (first noted by [23]) is more consistent with models that have the ferroan anorthosites but not the Mg-rich rocks float over a magma ocean, than with models that attempt to account for both groups via an ocean.

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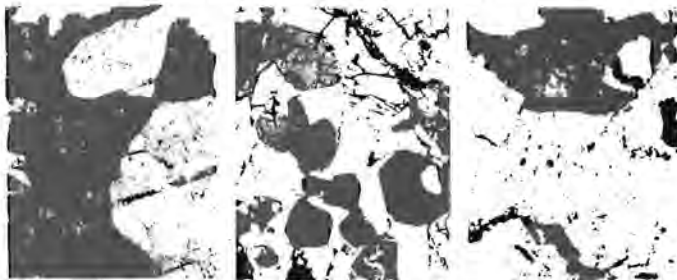
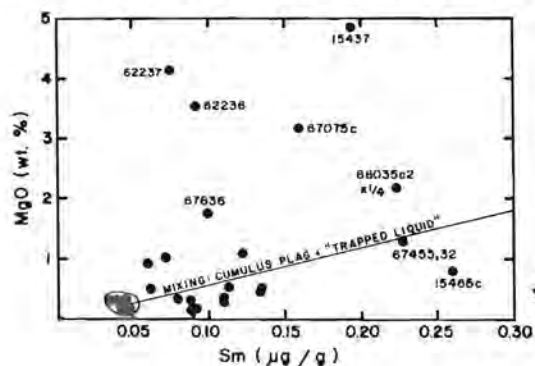
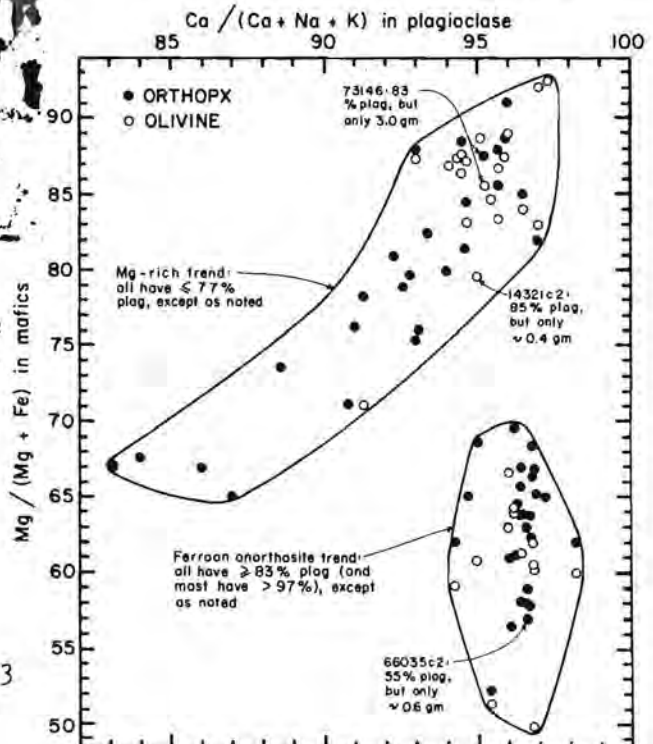


Fig. 1. Cumulates, 2.1x1.6 mm views, 1 to r: plagioclase-pigeonite mesocumulate 66035c2; olivine-spinel heteradcumulate 67435c; plagioclase-olivine adcumulate 67915c.

Fig. 2



← Fig. 3

SIMILARITIES IN THE REGIONAL GEOLOGY OF THE MIDCONTINENT RIFT IN THE LAKE SUPERIOR REGION AND THE TERTIARY GEOLOGY OF THE NORTH ATLANTIC. Paul W. Weiblen, Dept. of Geology and Geophysics, University of Minnesota, Minneapolis, Minnesota 55455.

A compilation of the regional geology of the Lower and Middle Keweenawan (~ 1.3-0.9 b.y.) intrusive igneous rocks of the Lake Superior region (1) reveals a remarkable similarity in the spatial distribution and succession of igneous intrusions and range of magma compositions with the Tertiary geology of East Greenland, Iceland, and the Faeroe Islands. Green (2) has previously noted the similarities between the extrusive rocks of the Lake Superior region and Iceland.

The Keweenawan intrusive rocks in the Midcontinent Rift can be divided into three groups: i) alkaline plutonic rocks; ii) layered gabbroic and associated minor granitic differentiates and; iii) diabasic basaltic and felsic rocks.

- (i) The alkaline rocks are represented by a series of small gabbroic-syenitic-granitic intrusives which trend in a northerly direction at a high angle to the axis of Lake Superior (Fig. 1). The best studied of these is the Coldwell Complex which consists of a series of eccentric, elliptical, funnel-shaped intrusions (3 and 4) typical of those found in rift environments (5 and references therein). A series of nephelinitic-carbonatitic complexes in the southern part of the Kapuskasing Structural Zone have K-Ar ages of 1 b.y. (3) and may also be a part of early alkaline magmatism in the Midcontinent Rift (1 and 5).
- (ii) The diabasic dikes and sills occur as distinct swarms flanking Lake Superior (Fig. 1). At least three crosscutting intrusive relationships have been recognized (6, 7), but there appears to be two major magma compositions: an early high-Ti ferrobasalt and a later olivine tholeiite (Table 1). These two compositions are similar in major elements (Table 1) to mantle plume basalt (8) and MORB (9). Wilband and Wasuwanich (10) have recently found the two compositions in the Baraga dike swarm and associated flows on the south shore of Lake Superior (Fig. 1, Table 1).
- (iii) The plutonic gabbroic and associated differentiates are represented by the Crystal Lake Gabbro (6) in Ontario, the Duluth Complex in Minnesota (11), and the Mellen Complex (12-13) in Wisconsin (Fig. 1). Weiblen and Morey have argued (11) that there is evidence of two periods of magmatism in the Duluth Complex. The high-Ti ferrobasalt composition is considered to be a differentiate of the parent magma of the first intrusive event and olivine tholeiite is presumed to be the parent magma for the younger gabbroic-troctolitic intrusions.

The spatial distribution of the intrusive units referred to above are consistent with a triple junction (5) tectonic setting in the Lake Superior region as shown schematically in (Fig. 1). Brooks (14) has suggested an analogous tectonic setting for the Tertiary rifting in the

Weiblen, P.W.

North Atlantic (Fig. 2). The similarities between the geology of the intrusive rocks in the Lake Superior region and the North Atlantic may be summarized as follows:

- a) The alkaline Kangerdlugssuag intrusion in East Greenland (15) and the Coldwell Complex (4) occupy similar positions relative to the rifting directions in each case and both may be construed to lie within a failed arm of a triple junction (Figs. 1 and 2).
- b) The extensive dike swarm in the Kangerdlugssuag area of East Greenland (16) shows six crosscutting intrusive relationships. However, the major sequences are an early high-Ti ferrobasalt followed by olivine tholeiite of MORB affinity. The high-Ti ferrobasalt swarm has been identified as feeders to plateau basalts of East Greenland and the composition is similar to the Faeroe Island basalts and others in the Wyville-Thomson Ridge (17). These compositions and the areal distribution are similar to those of the early high-Ti ferrobasalts in the Lake Superior region (Figs. 1 and 2). The second generation MORB dikes are similar in composition to the second generation olivine diabases of the Lake Superior region.
- c) Rare earth data (10) on the Keweenaw Baraga dike swarm and associated flows in Michigan show enrichment patterns of the high-Ti ferrobasalt over the olivine tholeiite similar to those found for the Faeroe Islands basalts (18, 19) and MORB (9). The Keweenaw olivine tholeiites are higher in the LREE's and K than MORB (Table 1).

These similarities suggest that comparable tectonic processes and grossly similar magma sources were involved in the formation of the Midcontinent Rift and the initial break up of the North Atlantic. However, the differences in the REE data (Table 1) indicate that the Keweenaw magma sources were enriched relative to the Tertiary sources in light REE's and K or the magma generation, near surface differentiation, or crustal contamination processes differed. The details of these three processes remain to be worked out for both the Midcontinent Rift and the North Atlantic (19-23). The gross similarities outlined here suggest that continued comparative studies of the overall regional geology of the Lake Superior region and the North Atlantic should be fruitful. The fortuitous exposures in each region may be complimentary and provide a more complete record than heretofore recognized of the process of initiation of rifting and the complexities of magma variations through time.

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Table 1. Average Analyses of High-Ti Ferrobasalt and Olivine Tholeiite from the Midcontinent Rift and the North Atlantic.

	1	2	3	4
SiO ₂	48.47	45.42	49.5	48.56
TiO ₂	2.48	1.54	2.9	3.35
Al ₂ O ₃	13.77	16.43	13.5	16.19
FeO	14.21	11.34	14.2	9.30
MnO	0.22	0.17	0.2	0.17
HgO	5.67	7.86	5.7	6.49
CaO	8.76	10.08	10.4	12.66
Mg ₂ O	2.76	2.50	2.5	2.36
K ₂ O	1.02	0.56	0.5	0.09
F ₂ O ₅	0.29	0.10	-	0.15
TOTAL	97.65	96.0	99.4	97.32
La	14.60	12.60	11.1	2.01
Ce	40.86	26.27	-	7.35
Sm	7.02	4.22	5.5	2.97
Eu	2.13	1.17	1.9	1.14
Yb	3.60	2.51	-	3.10
Lu	0.64	0.64	0.33	0.49

Notes:

FeO is total Fe and FeO. REE in ppm.

1. Average of 19 Keeweenaw flows and dikes from the south shore of Lake Superior (10).

2. Average of 31 Keeweenaw olivine tholeiite flows from the south shore of Lake Superior (10).

3. Major for elements are an average of 10 aphyric High-Ti, ferrobasalts from the Færøe Islands (18). REE from (19), Table 1 #11-1).

4. Average of 12 MORB basalts (9, Table 1).

Weiblen, P.W.

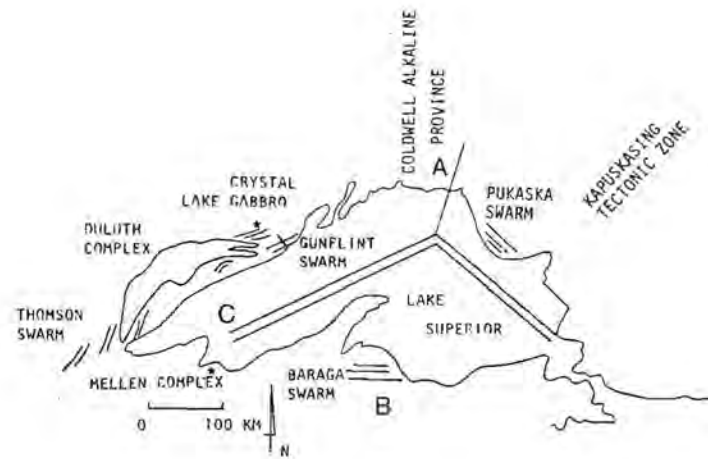


Fig. 1. Schematic map of the distribution of Keweenawan intrusive rocks in the Lake Superior region. (See references in 1 - 7 for detailed geology)

A - Trend of the Coldwell Alkaline Province (3 and 4).

B - Baraga dike swarm, a well-exposed example of high-Ti ferrobasalt (10).

C - Trend of olivine tholeiite dikes and sills and flows of the North Shore Volcanic Group (2, 6, and 7) in the Midcontinent Rift.

Short, solid lines indicate general trends of dike and sill swarms.

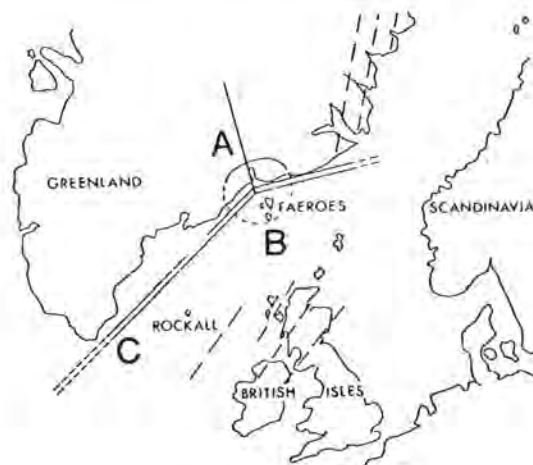


Fig. 2. Schematic map of the breakup of the North Atlantic. From Brooks (14), based on a pre-rift reconstruction (references in 14 and 23).

A - Location of the Kangerdlugssuag alkaline intrusion (15).

B - Position of the Faeroe Island high-Ti ferrobasalt eruption during early stages of the breakup of the North Atlantic.

C - Trend of the Mid-Atlantic Ridge and later MORB volcanism.

GEOLOGIC SYNOPSIS OF THE STRATIFORM FONGEN-HYLLINGEN BASIC COMPLEX, SCANDINAVIAN CALEDONIDES, NORWAY - COMPARATIVE STUDIES OF THE LUNAR MAGMA OCEAN/INTRUSIONS. J. Richard Wilson and Kim H. Esbensen, University of Aarhus, Dept. of Geology, DK-8000 Århus, Denmark.

The 160 km² central Norwegian Fongen-Hyllingen (F-H) complex is an extremely differentiated layered basic intrusion: almost 10 km of dominantly rhythmically layered rocks, commencing with olivine-picotite cumulates and ending with quartz-bearing ferrosyenites. The complex was emplaced, during the Caledonian orogeny, at 5-6 kb, indicating a depth of 15-20 km for the magma chamber.

Mineral compositions cover extreme ranges: olivine Fo₈₆₋₀; plagioclase An_{00-1.5}; Ca-rich pyroxene Wo₄₅En₄₄Fs₁₁ - Wo₄₇En₀Fs₅₃; Ca-poor pyroxene Wo₂En₆₇Fs₃₁ - Wo₂En₁₇Fs₈₁. The pyroxenes define a new type of trend for a layered basic intrusion [1]. The quartz-bearing nature of the final differentiates, the coexistence of two pyroxenes over a very wide crystallization interval, an olivine "gap" in the interval Fo₇₁₋₆₁, and the geochemistry of non-cumulate homogeneous olivine gabbro-norites in a presumed feeder, indicate a tholeiitic parent.

Reversals in phase and cryptic layering with stratigraphic height indicate at least seven major episodes of magma replenishment, and removal of compositional overlaps reduces the present cumulate stratigraphic thickness by over 75%. Reversals are to irregular compositions and successive magma batches are believed to have mixed with residual, differentiated magma from the preceding episode. The magma chamber was closed with respect to the exit of material [2].

Magnetite and ilmenite join the cumulus assemblage at about Fo₅₅, ilmenite persisting into the final differentiates. Successive liquidus phases in the late differentiates are: apatite, biotite, zircon, quartz, K-feldspar (coexisting with albite), and allanite. A brown intercumulus titanian calcic amphibole is present through most of the layered series, becoming a liquidus phase in the late differentiates and reaching ferroedenite in the quartz-bearing ferrosyenites. The sequence in which minerals enter and leave the cumulate assemblage, together with the coexisting mineral compositions in the different influxes, show only slight variations, indicating that the successive magma batches were essentially of uniform composition.

The rhythmically layered, cumulate series is intraformationally interspersed with non-layered, medium- to very fine-grained, homogeneous granular rocks displaying grading relationships to the surrounding layered facies. These rocks fit in the general cumulus stratigraphy; the pertinent liquidus phases (cumulus phases in the enclosing cumulates) make up the main part of these "average rocks" together with a minor amount of brown amphibole. Their textures show no evidence of accumulation; relatively rapid crystallization from the contemporary magma is inferred. The major element chemistry of these "average rocks", together with average cumulates reveal a pattern consistent with tholeiitic differentiation with moderate iron enrichment in the early to middle stages [3]. The final differentiates have K₂O/Na₂O ~ 1. Trace element patterns are analogous: the large cations (K, Rb, Cs and Ba) are all markedly enriched in the final derivatives as are the large, multivalent cations (Nb, Hf, Zr, U and Th). Th/U increases with differentiation as does Nb/Ta; Zr/Hf, however, shows no significant trend. The REE likewise show a prominent enrichment trend relative to the homogeneous feeder rocks which are the best available representation of the new initial magma. Initial Ce/Yb

Wilson, J.R. and Esbensen, K.H.

is approx 1-2 while the extreme differentiates (before the entry of allanite) have $Ce/Yb = 8.8$.

The F-H layered intrusion thus forms an essentially complete expose of the products of fractional crystallisation of basaltic magma under moderately elevated pressure. A recent petrological description of the complex [2] provides a geological basis on which to carry out comparisons with the pristine lunar rocks believed to have formed in a magma ocean (and/or smaller layered intrusions) [4].

Possibilities for analogous studies with the lunar evolution include:

- 1) The early lunar Mg-rich rocks have been interpreted as not belonging to the ocean suite [4] but rather to represent layered noritic to troctolitic plutons. Using their chemical systematics as well as our current petrological techniques it will doubtless prove advantageous to further study the earliest bottom cumulates of the F-H complex which are cogenetic with the rest of the overlying layered series.
- 2) The final derivatives of the F-H complex (including some residual filter-pressed liquids occurring as injections in the country rocks) which represent the most extreme magmatic differentiates are genetically similar to the lunar KREEP, especially "urKREEP", hypothesized by [5] to have been the residuum of the magma ocean. While urKREEP probably was usually modified by later events, the F-H material is essentially in a pristine condition. The two associations exhibit similar geochemical characteristics; e.g., typical KREEP $Ce/Yb = 7.5$ [5].
- 3) There is a possibility of formulating a stochastic cratering model "sampling" the F-H complex. The resulting "dilution pattern" can be compared with the lunar sampling of diverse shallow as well as deep crustal components.

Some of the remaining tasks that can be carried out as a joint venture between lunar and terrestrial scientists are:

- 1) study of F-H mineral chemical trends involving microprobe work and INAA of separated phases from selected samples representing the entire magmatic evolution of the F-H complex.
- 2) INAA and XRF analyses of representative cumulates for comparison with lunar rock.

Similarities between the F-H complex and the moon have been emphasized above, but there is a vital contrast in that crystallisation of the F-H complex took place under moderate and increasing $P-H_2O$ [1, 2].

Further study of this accessible and well-exposed intrusion will improve our understanding of processes responsible for the formation of terrestrial layered complexes. Comprehension of significant similarities/contrasts with the moon will enable our picture of the genesis of the lunar analogues to be refined.

Acknowledgement: We thank Paul Warren, who helped arrange this abstract into final form (and who will be collaborating with us in future F-H studies).

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Wilson, J.R. and Esbensen, K.H.

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Fig. 1

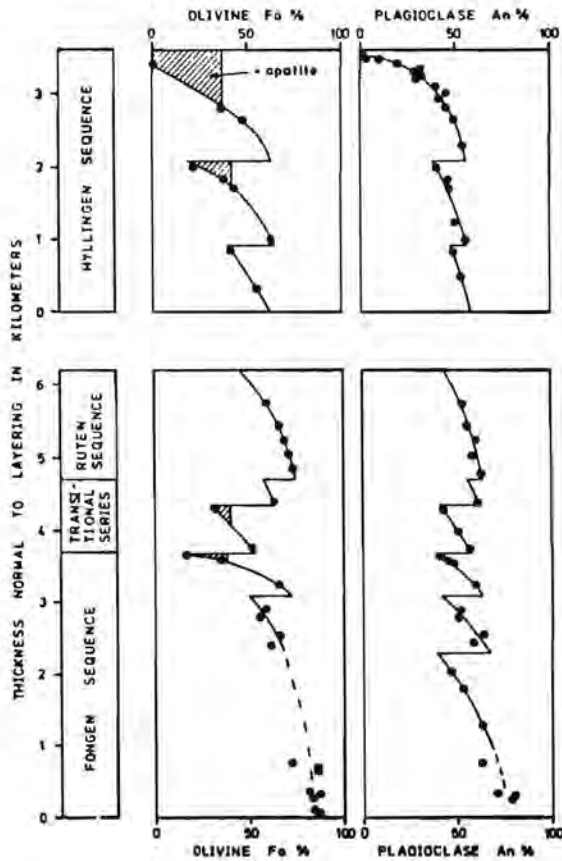


Fig. 2

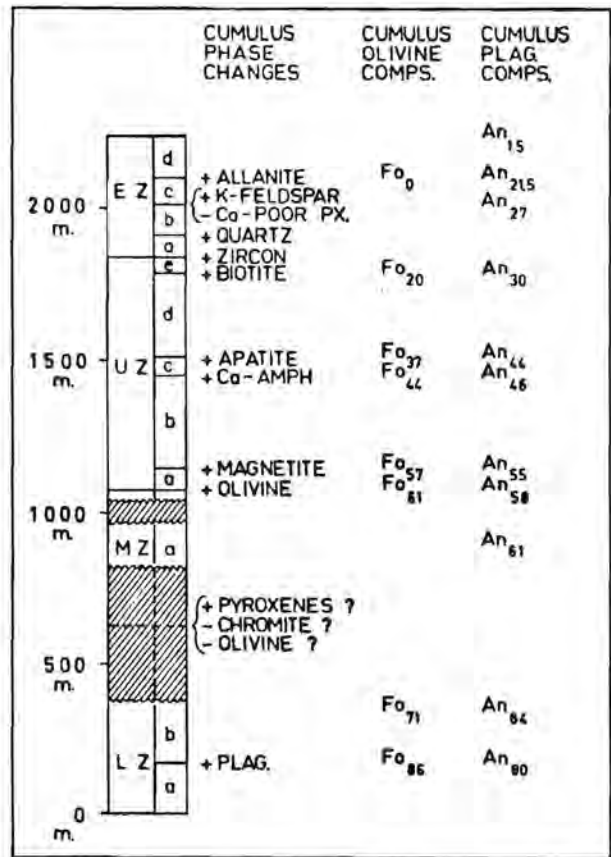


Fig. 1. Stratigraphic variation in plagioclase and olivine composition. Sampling interval approx. 200 m. Abrupt reversals in mineral composition trends here are interpreted as seven magma replenishments (schematic solid trace). An olivine "gap" (Fo_{71-61}) is apparent as is the entry of the cumulus phase apatite. The entry of apatite at virtually identical stages (measured by the olivine composition) indicates successive replenishments of nearly identical chemical composition. Intermediate part of the Fongen sequence is hidden, see Fig. 2.

Fig. 2. "Condensed" cumulus stratigraphy of the Fongen-Hyllingen complex, after removal of the reversals due to magma replenishments. Abbreviations: LZ, MZ, UZ, EZ denote Lower, Middle, Upper and Extreme Zone respectively. Entry of cumulus phases (+); exit of cumulus phases (-), except "+MAGNETITE" denotes both magnetite and ilmenite, the latter persisting throughout EZd and magnetite s.s. disappearing at the base of UZe. Hidden part of LZ and MZ crosshatched.

Field Guide to the Stillwater Complex

Workshop of magmatic processes in early planetary crusts

FIELD GUIDE TO THE STILLWATER COMPLEX

prepared jointly by:

L.D. Raedeke and I.S. McCallum, University of Washington, Seattle
The Anaconda Copper Company — Stillwater Project
Stillwater PGM Resources

INTRODUCTION

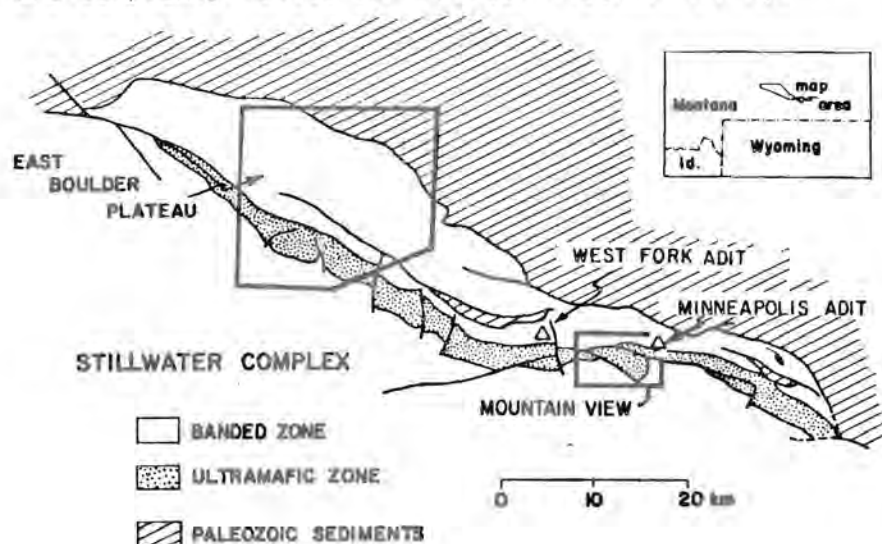
The Stillwater Complex has continued to attract the attention of geologists both because it is an easily accessible, well exposed example of a large layered intrusion, and because it is a significant repository of magmatic ore deposits. The large reserves of chromite (10), the Cu/Ni sulfides in the basal portions of the intrusion (1,16), and the recent discovery of a laterally extensive platinum group element (PGE) mineralized zone (3) combine to make the Complex a major source of mineral wealth.

The age of the Complex as determined by a Sm-Nd mineral isochron is 2701 ± 8 my (4). The general geology of the intrusion and surrounding rocks has been described in numerous publications, most notably those of Jones *et al.* (13), Hess (5), Page (15), Segerstrom and Carlson (22), and Carlson and Segerstrom (2). Petrological and geochemical studies include Hess (5), Jackson (8,9,11,12), Page (16), McCallum *et al.* (14), and Raedeke and McCallum (21). In addition, geologists of the Anaconda Copper Company, AMAX and Stillwater PGM Resources have intensively studied areas of economic potential within the Complex.

The Stillwater Complex is located in southwestern Montana (figure 1). It trends NW-SE, roughly parallel with the northern front of the Beartooth Mountains. As presently exposed the layering most commonly dips 60° to the north, but ranges to being overturned to the south. The lower contact of the Complex is intrusive along most of the exposed strike length, while the upper portion of the Complex is unconformably overlain by Paleozoic sediments. The footwall rocks of the intrusion are complexly folded and faulted Precambrian age

Figure 1.

Location map
simplified from
Jones *et al.* (13).



Raedeke, L.D.

sediments which were metamorphosed during the intrusion of the Complex to a cordierite-orthopyroxene hornfels. Locally, a quartz monzonite (2.7 Gy), which postdates the Complex, was intruded along the contact between the Stillwater Complex and the underlying hornfels. The hornfels contains sedimentary sulfide which predates the Complex, but was significantly affected by basic intrusions which precede the intrusion of the Stillwater Complex.

The Complex can be subdivided into a narrow Basal zone composed dominantly of heterogeneous norite and orthopyroxenite, an Ultramafic zone of olivine ± orthopyroxene rocks, and a Banded zone of predominantly plagioclase-rich rocks. The boundary between the Ultramafic and Banded zones is defined as that horizon where plagioclase first appears as a cumulus phase.

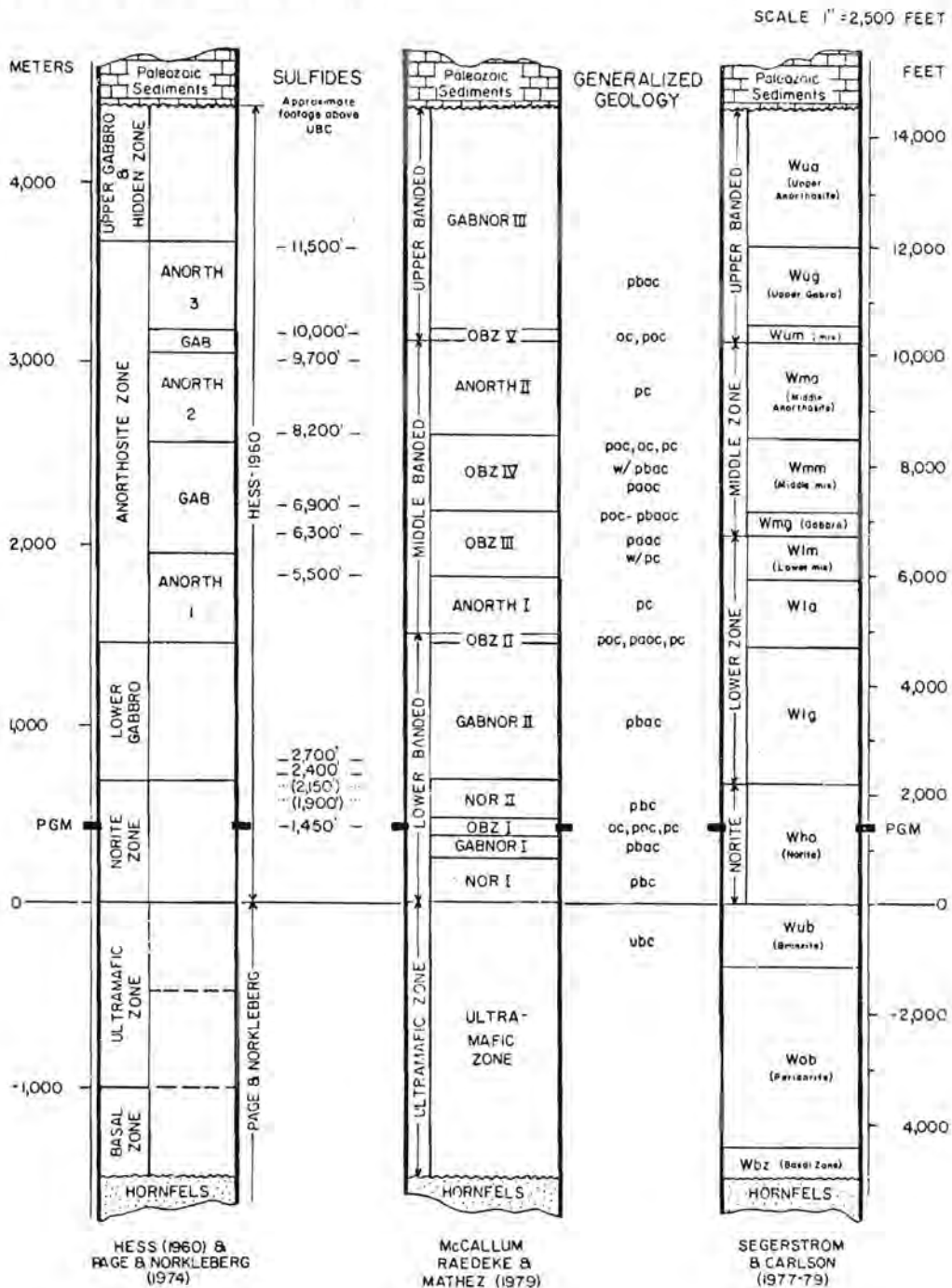
Basal zone: The Basal zone is 18 to 400 meters thick, averaging between 60 and 240 meters. (16). Broadly it is composed of a basal heterogeneous norite, a mixed cumulate member of plagioclase, olivine, olivine-orthopyroxene, and orthopyroxene-rich layers with limited lateral continuity, and a uniform orthopyroxenite. Ni/Cu sulfides are distributed throughout the Basal zone, but, in general, decrease in abundance from the bottom to the top.

Ultramafic zone: This zone can be subdivided into a lower Peridotite subzone or member of olivine ± orthopyroxene cumulates, and an upper Orthopyroxenite subzone. The rocks of the Peridotite subzone occur as cyclic units (macro-rhythmic units) of dunite-harzburgite-orthopyroxenite. Thirteen of the cyclic units contain chromite layers, which are designated A through I from the base upwards, but only the G and H have been mined. The chromitite layers have been traced along the entire 42 km strike length of the Complex.

Banded zone: The Banded zone includes a thick sequence (>4500 meters) of plagioclase-rich rocks. The zone can be further subdivided into a Lower Banded zone (LBZ) of norites and gabbro-norites, a Middle Banded zone (MBZ) of anorthosites, troctolites and olivine gabbros, and an Upper Banded zone (UBZ) of gabbro-norite. The layers of economic interest are three plagioclase-olivine layers which contain Pt/Pd mineralization. The layer with highest Pt/Pd grades is stratigraphically lowest (LBZ). It is being explored by Anaconda and by Stillwater PGM Resources. The middle plagioclase-olivine layer of interest is in the MBZ, and the uppermost, called by Anaconda the Picket Pin/Contact Mountain Horizon, is in the lowest UBZ.

Raedeke, L.D.

Figure 2. Generalized stratigraphic section through the Complex comparing nomenclature and divisions used by several workers. References are Hess (5), Page and Nokleberg (17), McCallum *et al.* (14) and Segerstrom and Carlson (22). Note that the Ultramafic zone has been further subdivided by Raedeke and McCallum (fold-out figure, this report). Abbreviations used: *ubc* = upper orthopyroxene (bronzite) cumulate, *bc* = orthopyroxene cumulate, *pc* = plagioclase cumulate, *oc* = olivine cumulate, *poc* = plagioclase-olivine cumulate, *pbac* = plagioclase-orthopyroxene cumulate, *pbac* = plagioclase-orthopyroxene-augite cumulate, *paoc* = plagioclase-augite-olivine cumulate. Figure compiled and prepared by Stillwater PGM Resources.



Raedeke, L.D.

AUGUST 5: MOUNTAIN VIEWStop 1: Peridotite subzone

L.D. Raedeke and I.S. McCallum

Stratigraphic sections through the Peridotite subzone at Mountain View and at Chrome Mountain on the East Boulder Plateau are provided in the envelope at the back of this volume. Sample locations for the Mountain View section and the field trip traverse are shown on the simplified map of figure 3. We will traverse the macro-rhythmic units from ~700 m to 1000 m in the stratigraphic section.

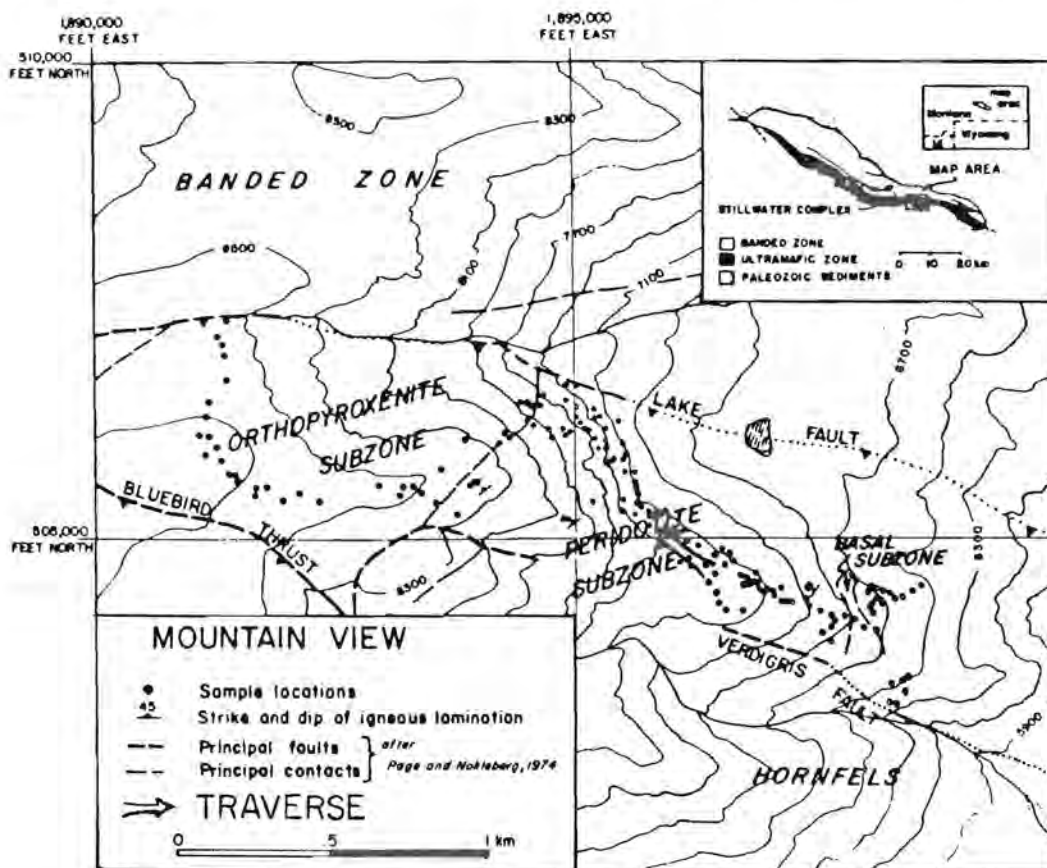


Figure 3. Simplified map of Mountain View area.

Modal layering is well-developed in the Peridotite subzone, occurring on a large scale in macro-rhythmic units of dunite-harzburgite-orthopyroxenite. One such cyclic unit at Mountain View has been sampled in detail and is shown in figure 4. In the macro-rhythmic units, the lower contacts of the dunites are invariably sharp and planar. Postcumulus orthopyroxene occupies >10 volume percent in most dunites, so the majority of the olivine-bearing rocks in the Ultramafic zone must be termed harzburgitic. In the dunites, the orthopyroxene occurs as oikocrysts with rounded and/or embayed olivine inclusions, clearly indicating formation of the pyroxene from olivine by reaction with the interstitial liquid. Modes of the low-Ca pyroxene generally increase upward, and, although it has not been systematically verified, field observations suggest that oikocrysts of orthopyroxene decrease in size and increase in number in the same direction. In many dunites, olivine has grown through

Raedeke, L.D.

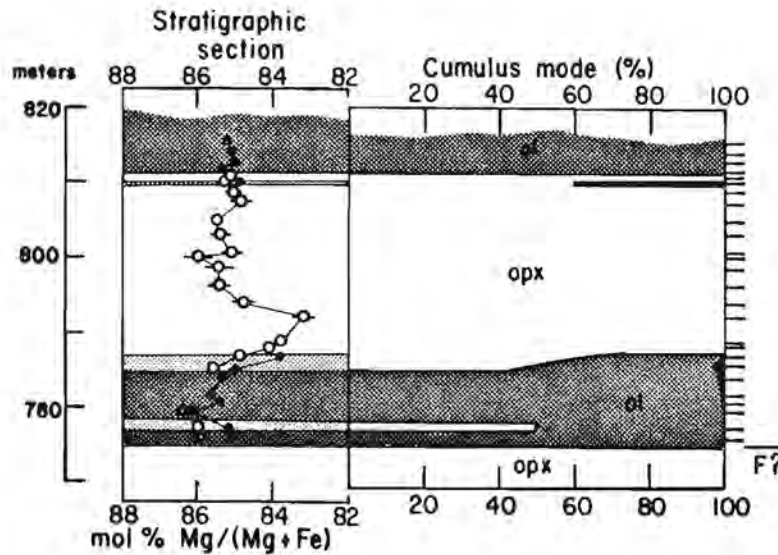


Figure 4. Detailed stratigraphic section through the macro-rhythmic unit at 775 m at Mountain View. Ticks on right are sample locations. "F" indicates probable fault. Closed circles are cumulus olivine compositions. Open circles are cumulus opx compositions, open triangles are postcumulus opx compositions.

secondary enlargement to form a granular texture with well developed triple junctions. It is not uncommon to find domains of secondarily enlarged olivine in the same thin section with domains of olivines which have been resorbed and rimmed by orthopyroxene. Postcumulus plagioclase, occurring usually as interstitial fillings, is present in small quantities (2-10 volume percent) in virtually all the dunites examined from Mountain View. Postcumulus augite, if present, is a minor constituent. Biotite is a minor but common postcumulus phase in dunites. It usually occurs as small (<1 mm) interstitial flakes. Rarely, it forms small oikocrysts. Chromite is commonly disseminated in the olivine-bearing rocks and/or may form seams typically near the base of the dunites.

The upper contact between harzburgitic dunite and harzburgite, the second unit of the cyclic sequence, is marked by an abrupt increase in modal orthopyroxene. Nevertheless, the textural change across this contact is not always sharp. In many cases, the habit of orthopyroxene changes gradually over a stratigraphic thickness of a few meters from poikilitic to granular. Olivine abundances are varied in the harzburgites, ranging from 5 to 78 percent olivine. With few exceptions, however, the olivine/orthopyroxene ratio decreases up section. Although the upper contact with orthopyroxenite is most commonly sharp and marked by the complete disappearance of olivine, this contact may also be gradational over several meters. Textures are complex and varied in the harzburgites. In some rocks, olivine and orthopyroxene are in apparent equilibrium, i.e., both minerals show secondary overgrowths to form a simple granular texture. In other rocks, olivine may be resorbed and/or rimmed by an orthopyroxene overgrowth in an apparent reaction relationship.

The uppermost unit in the cyclic sequence is an orthopyroxenite with postcumulus plagioclase and augite. Orthopyroxene usually shows secondary overgrowths and commonly forms a monomineralic granular texture. Grain size

Raedeke, L.D.

variation with stratigraphic height is common in otherwise uniform orthopyroxenites. Also, sharp breaks between competent and friable orthopyroxenite occur along conformable contacts. Augite occurs as reaction rims or oikocrysts enclosing partially resorbed orthopyroxene crystals. Both plagioclase and augite increase in abundance upward in the orthopyroxenite.

All three units within the cyclic sequence (dunite-harzburgite-orthopyroxenite) are not always present. The dunite is most commonly the missing unit, but orthopyroxenites are also sometimes absent. Harzburgites are never absent in the stratigraphic section described here, but Jackson (8,12) reports rare cases where the middle harzburgite unit is missing from the cycle. The initiation of each cycle can be defined as that horizon where olivine sharply increases (i.e., lithology changes either from an orthopyroxenite of the preceding cycle to a dunite or harzburgite of the next cycle, or from a harzburgite of the preceding cycle to a dunite of the subsequent cycle). Thus defined, there are 19 major and 2 minor macro-rhythmic units shown at Mountain View, and at Chrome Mountain there are 11 major and 9 minor units shown. Although the number of cyclic units is similar for Chrome Mountain and Mountain View, it should not be assumed that they can be correlated. Indeed, detailed mapping (18) shows that the lithologic units thicken, thin, or pinch out along strike. See Raedeke and McCallum (this volume) for further detail on the Ultramafic zone.

Raedeke, L.D.

Stops 2-4

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Stop 2: Peridotite subzone

Inspect mine dump for typical lithologies of the Peridotite member of the Ultramafic zone. The Mountain View mine is in the Mouat block which has been rotated approximately 90° to the regional NW-SE strike of the Stillwater Complex. The Mouat Block is bounded on the north by the Lake Fault and the south by the Bluebird Thrust.

Excellent examples of all parts of a cyclic unit can be found in the dump area and several excellent exposures of olivine cumulate with bronzite oikocrysts can be observed approximately 50 meters back down the road.

The cycles in the Ultramafic zone have been interpreted as being the result of convective overturn of a single magma (8). However, by analogy with the Muskox Intrusion (6) each cycle could represent a fresh influx of magma along the bottom of the intrusion displacing either all or part of the residual magma that produced the previous cycle.

Stop 3: Ultramafic zone - Banded zone contact

This contact is marked by the first appearance of plagioclase as a cumulus phase above the Basal zone. The horizon has been traced along the entire strike length of the Complex. A possible inclusion of the Ultramafic zone in the Banded zone can be seen a short distance above the contact in the outcrop.

Stop 4:

A troctolite which belongs to the Middle Banded zone is exposed in this roadcut. Called the second troctolite or poor man unit by ANCC, this unit may be correlative with one of the OBZ subzones of the stratigraphy of McCallum *et al.* (14). Layering is nearly vertical in this exposure; tops of units face north at this location.

Three distinct lithologies are present:

1. gabbronorite (PBAC)
2. troctolite (POC)
3. anorthosite (PcoA)

The gabbronorite occurs as the footwall unit to the troctolite. It consists of green clinopyroxene, brown orthopyroxene, and white plagioclase. All three minerals appear to be cumulus phases. A lamination defined by the plagioclase crystals is well developed.

The troctolite consists of cumulus olivine and plagioclase. The unit appears to be mineral graded. The lower part of the layer is olivine-rich whereas the upper part of the layer is plagioclase-rich. The layer is slightly sheared. All olivine is altered to serpentine and magnetite, although locally, the olivine is altered to fibrous amphiboles.

The hanging wall to the troctolite is an anorthosite. It consists of cumulus plagioclase and minor interstitial pyroxene. The pyroxene forms coarse-grained oikocrysts. All contacts between units are sharp crystal contacts. They are exceptionally well exposed here. A sulfide horizon occurs in the anorthosite well above the troctolite. It consists of two parallel thin layers (approximately 1 cm thick).

Raedeke, L.D.

Stop 5: West Fork Adit

Stillwater PGM Resources

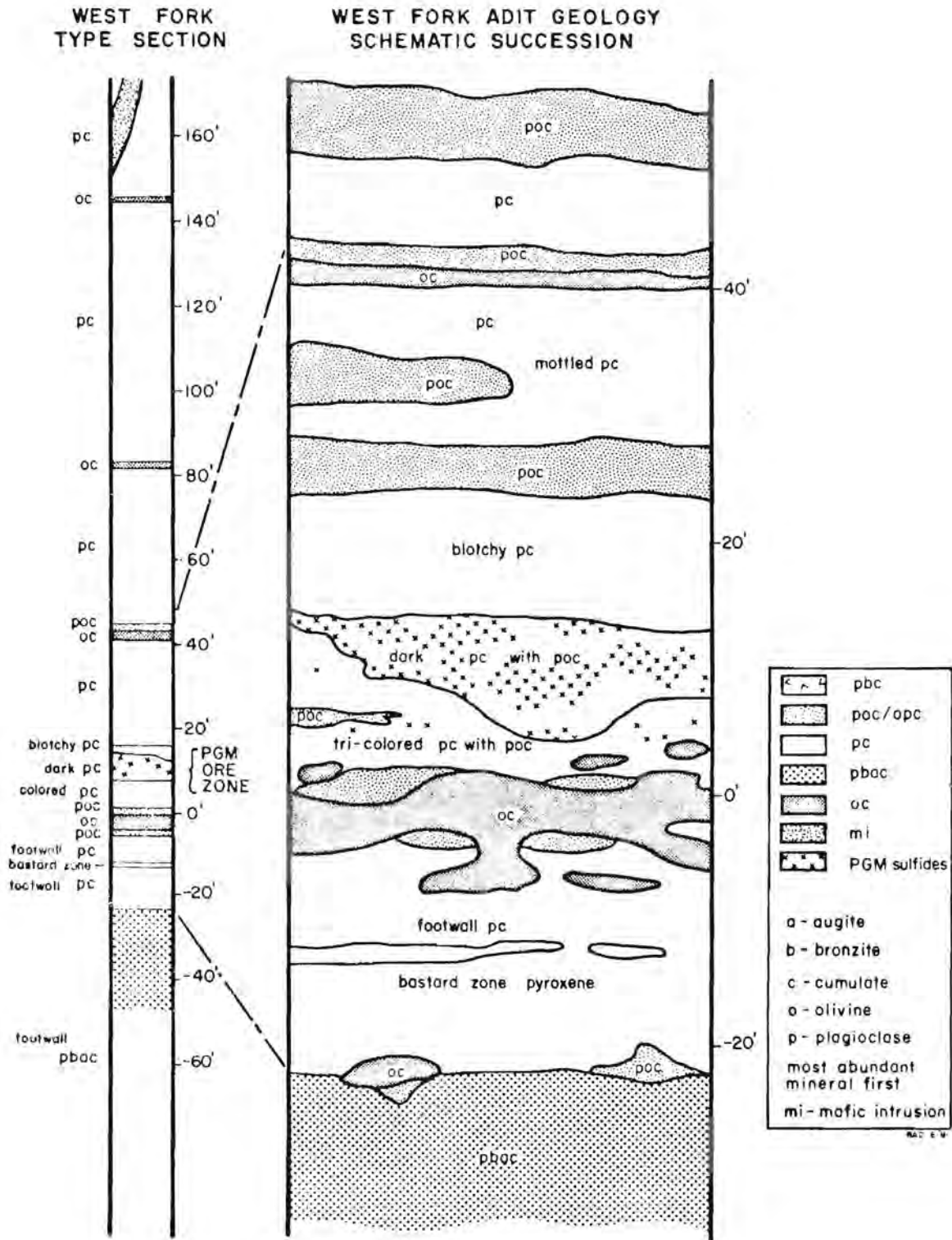
Generalized notes on the stratigraphic succession of the PGM mineralized zone (LBZ) referring to figure 5.

pbc	+240 m	Plagioclase (55%) opx (40%) with some cpx.
pc	+25 m	Plagioclase cumulate with layers of postcumulus opx + cpx oikocrysts or layers of poc or even oc in places (possibly Unit 4).
poc oc (poc)	Unit 3 { .5-1 m	Olivine (70%) cumulate with postcumulus plagioclase. Grades upwards (and, in some places downwards) into plagioclase (70%) olivine (20%) cumulate. Slight geochemical PGM signature. Magnetic.
pc		11-12 m Plagioclase (80-90%) with some indication of pc with oikocrysts or poc.
poc oc (poc)	Unit 2 { .5-1 m	Olivine (70%) cumulate with postcumulus plagioclase. Grades upwards (and, in places downwards) into plagioclase (70%) olivine cumulate. Slight geochemical PGM signature. Magnetic.
pc		1.5-4.5 m Plagioclase cumulate. "Mottled" texture caused by 10-15 cm pyroxene oikocrysts. Also contains some poc layers.
poc	1.5-2.5 m	Plagioclase (75%) olivine (5-15%) cumulate with olivine as widely separated "blebs" and typically altered to serpentine and magnetite with magnetic signature.
pc	1.5-3 m	Plagioclase (70-80%) cumulate. "Blotchy" texture with conspicuous 15-25 cm, superimposed opx (10-15%) and cpx (10-15%) oikocrysts to form dark grey "blotches" against white plagioclase background. Lower contact against dark pc commonly "spotted" with scattered small (5 mm) clots of postcumulus pyroxene. Blotchy texture indistinguishable in outcrop.
pc dark	Zone of Interest (ZOI) { 1-2.5 m	Plagioclase cumulate (gray-green) with up to 50% post-cumulus brown opx and green cpx oikocrysts and up to 5% olivine. Finely disseminated sulfides (+1-2%) which are about 1/3 each of chalcopyrite, pentlandite and pyrrhotite, with associated PGM's especially in upper portion.
pc "tri-colored"		1-2.5 m Overlying patches (5-15 cm) of plagioclase cumulate, green cpx and subordinate brown postcumulus opx. Medium-grained and commonly containing wispy stringers of poc with occasional disseminated chromite or mica. Alteration of olivine generally is to non-magnetic tremolite and talc rather than usual serpentine-magnetite variety. Tri-colored pc contains erratic medium and coarser-grained segregations of sulfides with associated PGM's.

Raedeke, L.D.

poc	Unit 1	0-.6 m	Plagioclase (50-70%) olivine (20-50%) cumulate. Olivine + 5 mm commonly coalescing to form "Vermiform" or "ameboidal" shapes with typical reaction rim textures. Generally lacks PGM values but can contain significant coarse-grained sulfide with PGM's in some areas.
oc		1-1.5 m	Olivine (70%) cumulate with 20% large (5-10 cm) (occasionally 20 cm) pyroxene oikocrysts (with included "sieve" structure olivine and 10-20% postcumulus plagioclase). Typically dense, black, magnetic coarse-grained pegmatoid rock with greenish to purplish tint. Olivine 2-15 mm with good reaction rims generally altering to serpentine and magnetite but in places displaying tremolite and talc alteration. Cumulus chromite (.5-3 mm) and books of reddish-brown divalent chrome mica. Sporadic segregation of interstitial sulfides with PGM's in places.
pc "footwall"		2-2.5 m	"Footwall" plagioclase (80-90%) cumulate. Light gray to white with gray serpentine-filled fractures.
aoc "Bastard"		.15-.3 m	Augite (80-90%) olivine (0-5%) cumulate with 10-20% postcumulus plagioclase. "Bastard zone". Rare occurrences of PGM sulfides. Dip-slip faulting common along plane of Bastard zone.
pc		2.5-3 m	Plagioclase cumulate. Light gray to white with <5% postcumulus pyroxene.
pbac		90 m	Plagioclase (55-60%) bronzite (20-25%) augite (15-20%) cumulate (gabbronorite) with alternating layers of pc. Opx crystals are stubby and brown and augite crystals are smaller, elongated and green. The upper 15 cm is generally coarser-grained.
pbc		240-275 m	Plagioclase (60%) bronzite (35%) cumulate with small amounts of green postcumulus augite oikocrysts. Generally massive norite becoming more banded towards top of sequence.

Figure 5. Stratigraphic section, West Fork adit.



Raedeke, L.D.

Stops 6-7:

Anaconda Copper Company — Stillwater Project

Stop 6: Inch scale layering

At this stop there are several features which have always been difficult to explain by crystal settling. The regular pattern to the repetitive layers of pyroxene and plagioclase suggests a leisegang mechanism for their formation, whereby the regular spacing of layers is the result of periodic nucleation in a temperature/compositional gradient. The "doublet" nature of the orthopyroxene layers is most readily explained by competition between growing crystal grains in the developing layers.

Other features to note are the modal increase of orthopyroxene as the inch scale layering is followed up section, the typical coarse-texture of the inch scale layers as compared with the medium-grained textures typical of the gabbros and norites above and below the inch scale section, and the presence of a crude, hexagonal pattern in the orthopyroxene layers seen on joint surfaces which are parallel to bedding.

Stop 7:

A section through the Pt/Pd zone from footwall gabbros up into hanging wall norites is visible in a road cut on the road leading to the Mouat block Ni/Cu showings.

The base of the exposed section lies within well-layered plagioclase-bronzite-augite cumulates (PBAC) and PBC's with numerous sharply defined bands of pure anorthosite a few centimeters in width. The base of the olivine bearing subzone (OBZ I) is marked by the first appearance of plagioclase cumulate with pyroxene oikocrysts (PcoB/A), containing sporadic development of patches of pegmatoidal pyroxenes with olivine cores. This rock type is designated mixed rock due to its extreme inhomogeneity. Intermixed plagioclase cumulates, plagioclase-olivine cumulates and mixed rock occupy about 10 meters of section above the last PBAC. Characteristic mixed rock textures showing the irregular shapes of the pegmatoidal olivine-bearing bodies may be seen in a number of large fallen blocks just below the edge of the road.

The Pt/Pd mineralization occurs within the succeeding unit; a 3 meter thick layer of olivine-rich "ameboidal troctolite". This rock type consists of large, irregular and resorbed crystals of plagioclase, with interstitial olivine, numerous very large oikocrysts of bronzite and augite, and minor cumulus chromite and interstitial phlogopite. Samples from this outcrop assay up to 4 oz/ton (140 ppm) Pt+Pd. The platinum group elements were originally present within interstitial sulfide blebs (pyrrhotite-pentlandite-chalcopyrite-PGE minerals). Unfortunately the sulfides in this outcrop have been completely oxidized to magnetite, and none are visible.

This unit is succeeded by about 5 meters of section of plagioclase-olivine cumulate, succeeded at a sheared contact by a layer of PcoB/A. The pyroxene oikocrysts stand out as prominent blotches on the weathered surface, and the rock shows a remarkable similarity with the "mottled anorthosite" found at the top of the Merensky Reef cyclic unit in the Bushveld Complex. The exposed section finishes up within uniform plagioclase-bronzite cumulates, at the base of the Norite II subunit. As the Pt/Pd bearing zone is followed up the hill above the road (i.e. towards the west), the whole package thins down markedly, and the overall proportion of olivine appears to decrease. An outcrop of anorthosite with 2-5% PGE-rich sulfides occurs about 150 meters above the road.

Raedeke, L.D.

AUGUST 7: CONTACT MOUNTAIN

A stratigraphic section through the Banded zone at Contact Mountain is provided as a fold-out figure in the envelope at the back of this volume. Sample locations for the Contact Mountain section are shown on figure 6. In addition, a simplified stratigraphic section with mineral compositions is shown in figure 7. Our traverse will follow the main line of section as seen on figure 6 except near the base. The traverse through Norite I, Gabbronorite I and OBZ I at the base will be jointly directed by Raedeke and McCallum and geologists of Stillwater PGM Resources. The stratigraphic sections prepared by the latter are shown in figures 8 and 9. The section prepared by Raedeke and McCallum through Norite I and Gabbronorite I which will be traversed is shown in figure 10, and through OBZ I in figure 11.

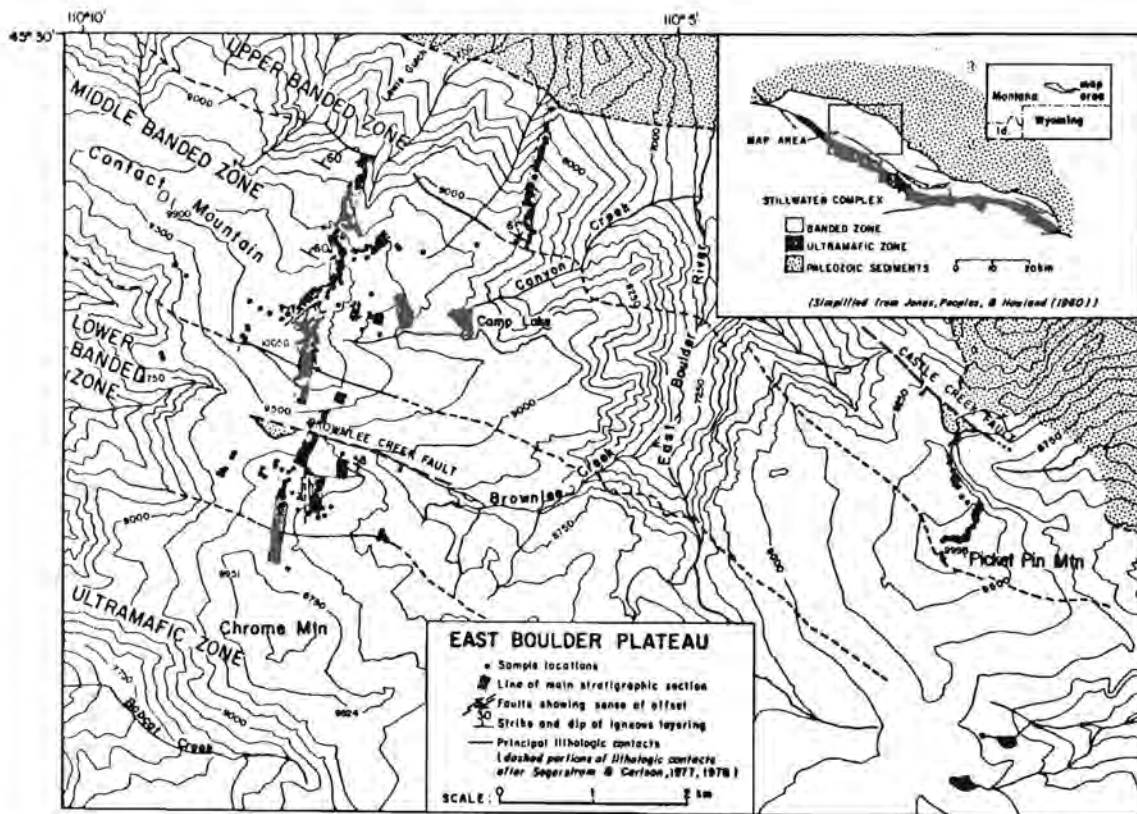


Figure 6. Location map for Contact Mountain and Picket Pin Mountain traverses. After McCallum *et al.* (14).

Raedeke, L.D.

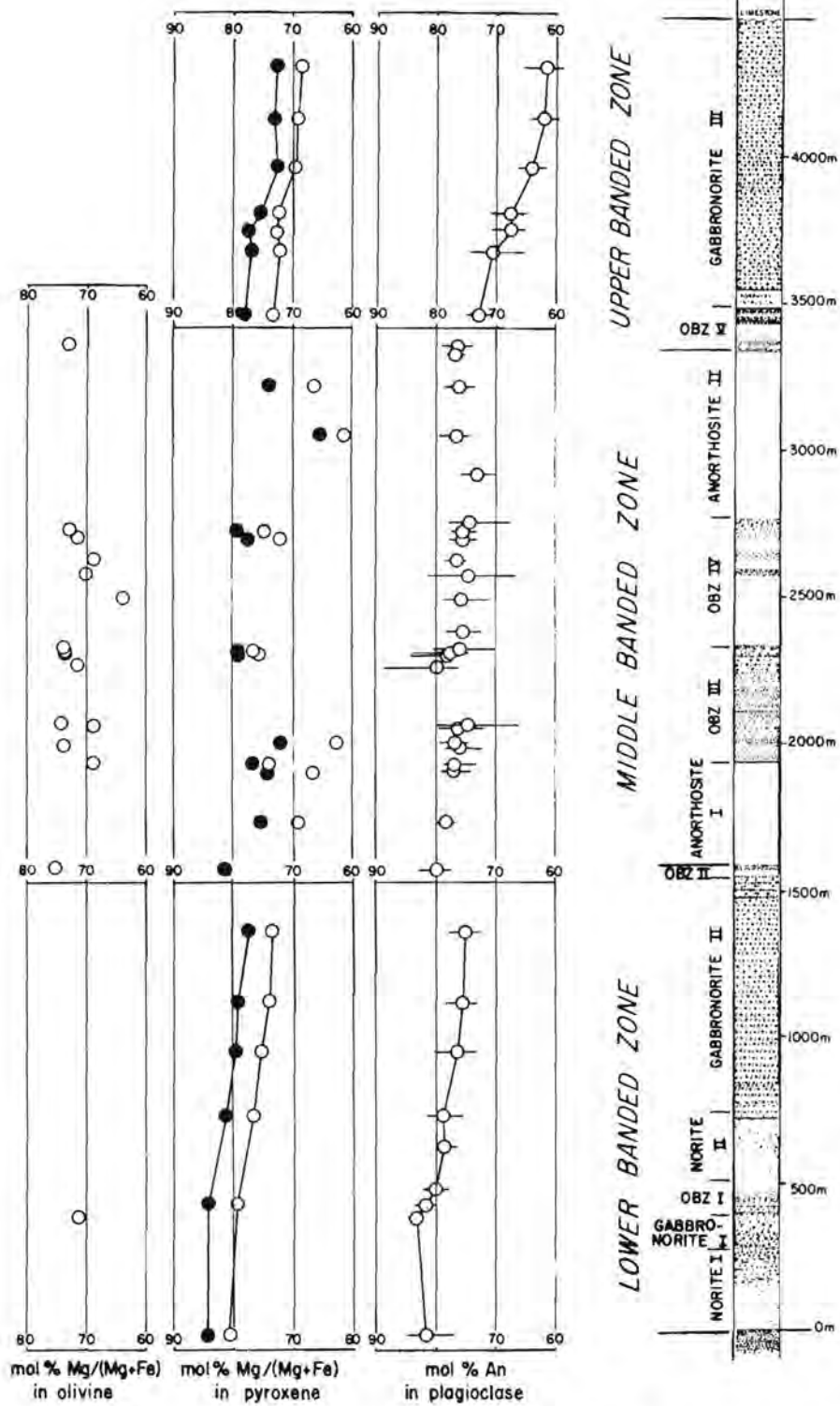


Figure 7. Simplified stratigraphic section through the Banded zone at Contact Mountain, after Raedeke and McCallum (19). Bars on plagioclase composition indicate range of composition within a single thin section. Open circles of pyroxene are low-Ca pyroxenes, filled circles are augites. See text for subzone descriptions.

Lower and Middle Banded zones

L.D. Raedeke and I.S. McCallum

Lower Banded zone (0-1590 meters)

The lower contact of this zone is placed at the horizon marking the first appearance of cumulus plagioclase. This phase contact is sharp and easily recognizable from the abrupt increase in plagioclase mode from 5 percent in the orthopyroxenite to >50 percent in the overlying norite. The upper contact is placed at the base of the first thick anorthosite unit. These boundaries are well defined and can be traced across the entire intrusion. The dominant lithologies in this zone are norite and gabbronorite with minor anorthosite, troctolite and gabbro members.

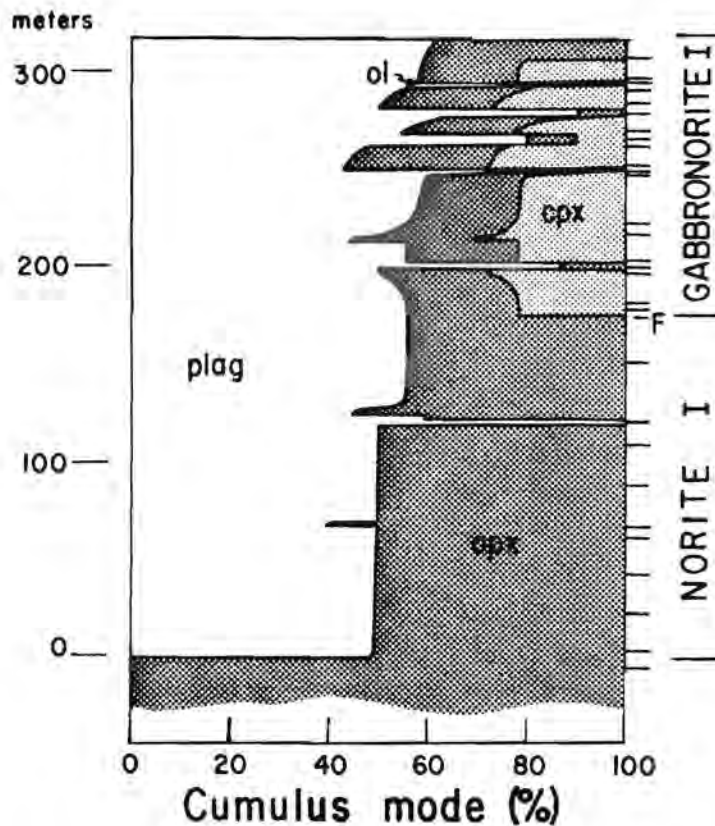
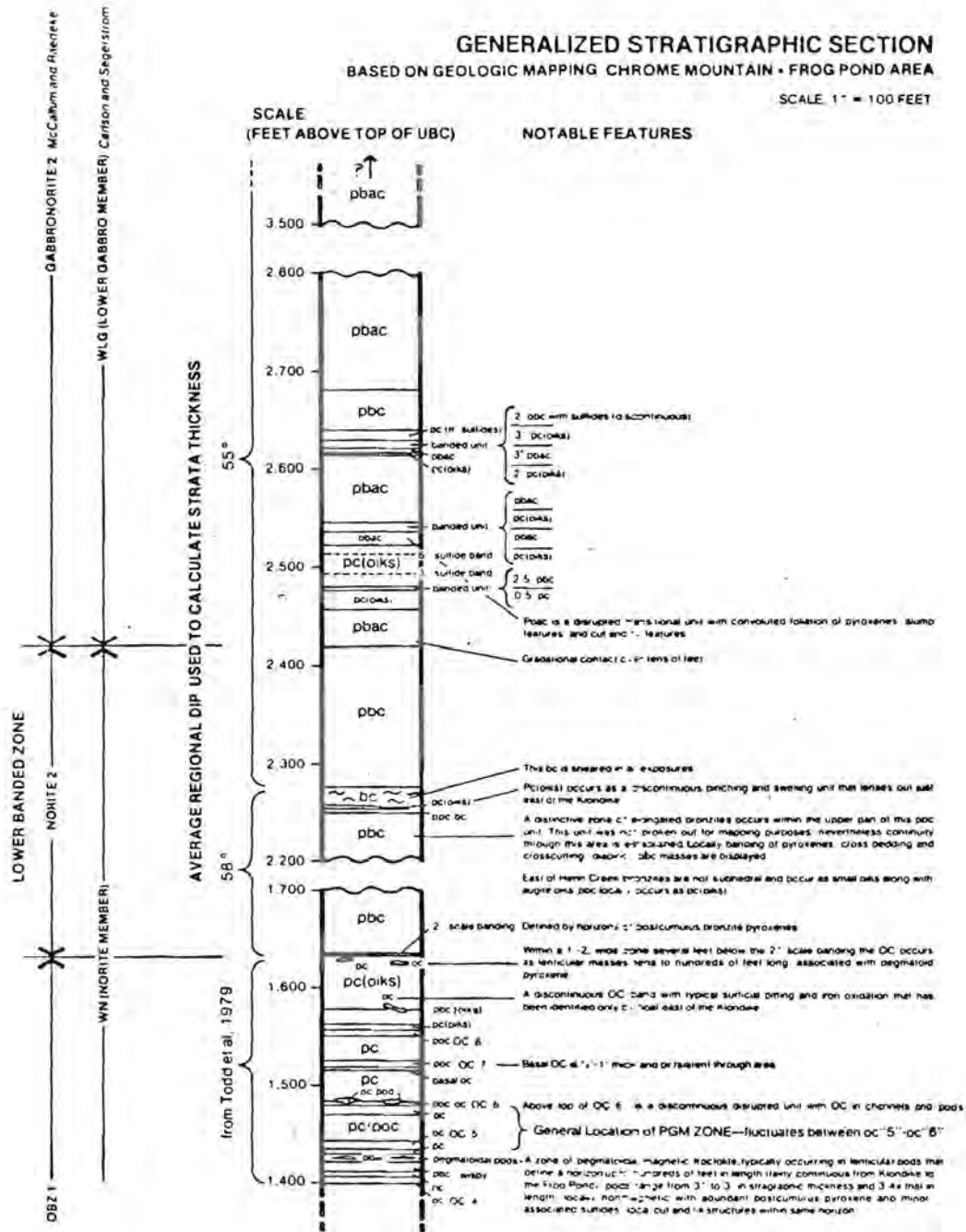


Figure 10. Variation of cumulus mode vs. stratigraphic height through Norite I and Gabbronorite I along field trip traverse.

Norite I (0-270 meters). Orthopyroxene and plagioclase are in approximate cotectic proportions in the lower uniform part of the norite, but plagioclase increases in abundance in the upper 100 meters and layering becomes more pronounced (figure 10). Layering is defined by alternating pyroxene-rich and plagioclase-rich units from a few centimeters to more than two meters thick. The narrower layers commonly bifurcate, are laterally discontinuous and may show right-side-up modally graded layering, scour-and-fill and slump structures. The magma was doubly saturated during formation of the norite and both cumulus phases appear to have accumulated on the floor of the magma chamber. Much of the layering can be ascribed to current action, but

Figure 9. Generalized stratigraphic section through upper OBZ I, Norite II, and lower Gabbronorite II. Prepared by Stillwater PGM Resources.



Raedeke, L.D.

those sharply bounded layers of anorthosite (maximum 3 meters thick), having no complementary mafic layers cannot be readily explained by this mechanism.

Gabbronorite I (270-400 meters). The lower contact of this unit is placed at the first appearance of cumulus augite, i.e., the magma was triply saturated. However, the proportions of orthopyroxene, augite and plagioclase are highly varied and generally non-cotectic principally because of an increase in plagioclase. In the upper part of this subzone there is a complex, laterally extensive unit characterized by highly disturbed layering and irregular intermixing of norite, gabbronorite, anorthosite and coarse-grained pyroxenite. This zone appears to be the result of strong currents and/or slumping at the floor of the magma chamber. Ellipsoidal inclusions of orthopyroxenite, texturally identical to that in the Ultramafic zone, are abundant in the uppermost 30 meters of this unit.

Texturally, the rocks in this subzone are somewhat unusual. Grain boundaries between pyroxenes, particularly clinopyroxenes, are commonly complexly interfingered, indicating extensive secondary overgrowths. Orthopyroxene crystals enclose rounded 0.5-1.0 mm clinopyroxene grains and small plagioclase laths. Similar textures are seen in Gabbronorite III and in olivine gabbros of Olivine-bearing subzone III.

Olivine-bearing subzone I (OBZ I) (400-504 meters). The basal contact of this complex subzone is marked by the reappearance of cumulus olivine. The contact is well defined but highly irregular. Pods of pegmatitic pyroxenite (up to 2 meters across) are sporadically developed along the contact and in the underlying gabbronorite. While this zone can be traced for at least 20 kilometers along strike, individual members are laterally variable, both in thickness and mode. Modal proportions are highly variable and generally non-cotectic. Two stratigraphic sections have been completed through OBZ I (figure 11). Section 11a is the same as that shown in the fold-out figure,

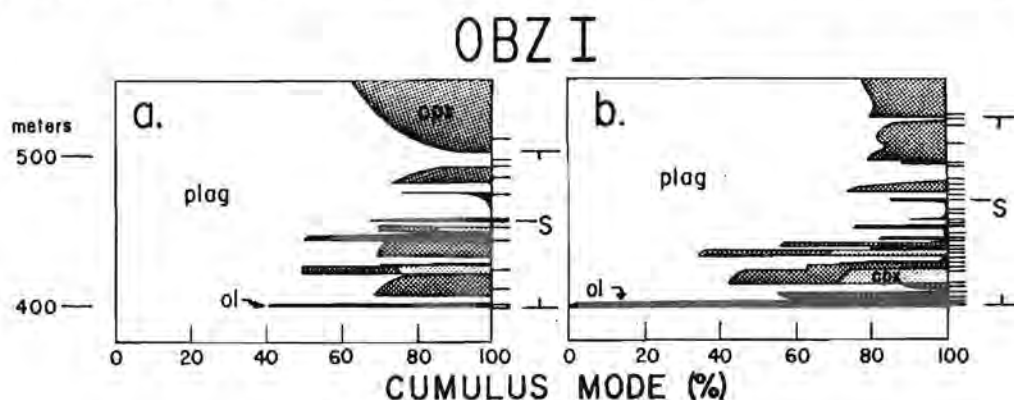


Figure 11. Cumulus mode vs. stratigraphic height for OBZ I. Brackets on right indicate limits of the subzone. See text for discussion.

and section 11b is ~180 meters to the east of 11a. Although the outcrop for 11b was substantially better than for 11a, some real differences exist. In section 11a, the sequence, troctolite-anorthosite-norite-gabbronorite is irregularly repeated. The basal troctolite is locally very olivine-rich while the fourth troctolite contains abundant sulfides. The noritic members in the lower part of this subzone are characterized by a distinctive wispy banding; higher in the subzone, layering within the norites becomes more regular. In

Raedeke, L.D.

section 11b the basal unit is a dunite grading upward to a troctolite. The apparent lithologic repetition of section 11a is not seen in 11b, and troctolites and anorthosites are more abundant in the latter. In both sections, pyroxene-rich rocks at the base give way to more anorthositic rocks near the top, clinopyroxene disappears as a cumulus phase upward in the section, and the PGE-bearing sulfides occur ~60-70 meters above the reappearance of cumulus olivine.

The crystallization sequence through Norite I and Gabbronorite I is that predicted for the basaltic system in which fractional crystallization with bottom accumulation is the dominant mechanism operative. However, in OBZ I the reappearance of olivine (+ sulfides), the unconformable nature of the basal contact together with the occurrence of pegmatitic pyroxenites and orthopyroxenite xenoliths indicate a major perturbation in the conditions of crystallization. The reappearance of cumulus olivine, in particular, is consistent with a hypothesis of one or more injections of olivine-saturated magma followed by a prolonged period of mixing before the magma returned to a relatively uniform composition represented by the overlying norite subzone.

Norite II (504-745 meters). The lower boundary of this subzone is placed at the contact between anorthosite and an anorthositic norite which shows well developed inch-scale layering. The layering is best developed in those norites with a high plagioclase/orthopyroxene ratio. As modal proportions approach cotectic values, the norite becomes correspondingly more uniform. Near the upper part of the norite the habit of the cumulus orthopyroxene changes from subrounded to highly elongate (10-15 mm in length). Grains of orthopyroxene oriented with their long axes in the plane of layering impart a pronounced planar lamination. A distinctive, laterally extensive, mafic layer (2 meters thick) and a complementary anorthosite layer occur ~20 meters below the top of this subzone.

Gabbronorite II (745-1538 meters). The base of this unit is marked by the reappearance of cumulus augite. The lower 100 meters are characterized by alternating layers of anorthosite and gabbronorite. Mineral proportions in the gabbronorite are varied. Anorthosite layers range in thickness from ~10 cm to 15 meters; two are sulfide bearing. The lower sulfide unit is a conformable, laterally continuous layer about 10 cm thick which in outcrop is sharply defined by a characteristic rusty stain.

In the central part of this subzone, mineral proportions in the gabbro-norite are near-cotectic and planar lamination is well developed. The upper 50 meters are composed of five well developed but laterally discontinuous cyclic units showing modally graded layering defined by an upward increase in plagioclase/pyroxene ratio. These cycles may be the result of localized, periodic, density currents interspersed with periods of relatively quiescent crystallization and differential settling.

Olivine-bearing subzone II (1538-1590 meters). The upper 8 meters of this subzone is a remarkable association of gabbro, troctolite and gabbroic pegmatite. The contact between gabbro and overlying troctolite is sinuous and discordant in contrast to the planar and concordant contact between the troctolite and overlying anorthosite. Irregular patches of gabbro are enclosed within troctolite and vice versa. Associated pegmatites contain pyroxene and plagioclase megacrysts up to 25 cm in diameter. Olivine in the troctolite occurs as large (>10 mm) "ameboidal" grains with inclusions of plagioclase

Raedeke, L.D.

set in a matrix of relatively small (1-2 mm) plagioclase crystals. The troctolite does not have the appearance of a typical cumulate,

Middle Banded zone (1590-3338 meters)

The stratigraphic boundaries of this zone are arbitrarily placed at the base of the first thick anorthosite (Anorthosite I), and the top of the second thick anorthosite. Throughout this zone, anorthosites and olivine-bearing rocks are dominant.

Anorthosite I (1590-1939 meters). This subzone is composed entirely of a uniform plagioclase cumulate with postcumulus augite and postcumulus inverted pigeonite. The average grain size of the plagioclase is coarser and more uniform than in the two- and three-phase cumulates above and below. Disseminated sulfides occur in the upper 8 meters of this unit and reach a maximum concentration in the upper 2 meters. This sulfide unit has been traced along strike for more than a kilometer to the east and the west of the main line of section.

Olivine-bearing subzone III (1939-2339 meters). This 400 meters of section is very complex. The predominant rock types are troctolite, anorthositic troctolite, anorthositic gabbro, olivine gabbro, and olivine gabbro-norite. The various members are complexly interlayered, often on a centimeter scale. The lower boundary of this subzone is placed at the base of a banded troctolite unit in which layering is defined by subparallel stringers of olivine grains set in a matrix of cumulus plagioclase. Olivine is present as a cumulus mineral through 80 percent of the subzone, but, except for eight narrow troctolite layers in which it comprises 40 percent, olivine abundances are low, making up between 1 and 10 modal percent of the cumulus mineral assemblage. The major units can be traced laterally but there are substantial lateral variations in the thickness, modes and textures of the thinner members.

While systematic stratigraphic variations are not immediately obvious, two distinct cyclic sequences that are repeated several times in the lower 275 meters of this subzone can be identified. Idealized examples are illustrated in figure 12. Incomplete cycles are common, i.e. all rock units of the idealized cyclic sequences are not always present; however, the order in which they appear does not change. The units overlying the second and third troctolites, respectively, represent good examples of the two cyclic sequences. The base of each cycle is commonly marked by a sharp but irregular contact.

Olivine gabbros and gabbro-norites are generally isomodal with a well defined lamination. A distinctive feature of these rocks, particularly those found high in the section, is the occurrence of (~2.5 X 1.0 X 1.0 cm) polycrystalline aggregates of plagioclase, the long axes of which are aligned parallel to the lamination. By contrast, the anorthositic gabbros commonly show an erratic layering defined by alternating mafic and felsic layers. Between 1990 and 2020 meters there is a laterally extensive zone of disturbed layering. This zone also contains irregular masses of anorthosite and appears to be the result of slumping of a partly consolidated crystal mush. Associated with this zone, but not necessarily genetically related, are troctolite lenses and pipes (up to 2 meters across) which crosscut or are enclosed within olivine gabbro.

Raedeke, L.D.

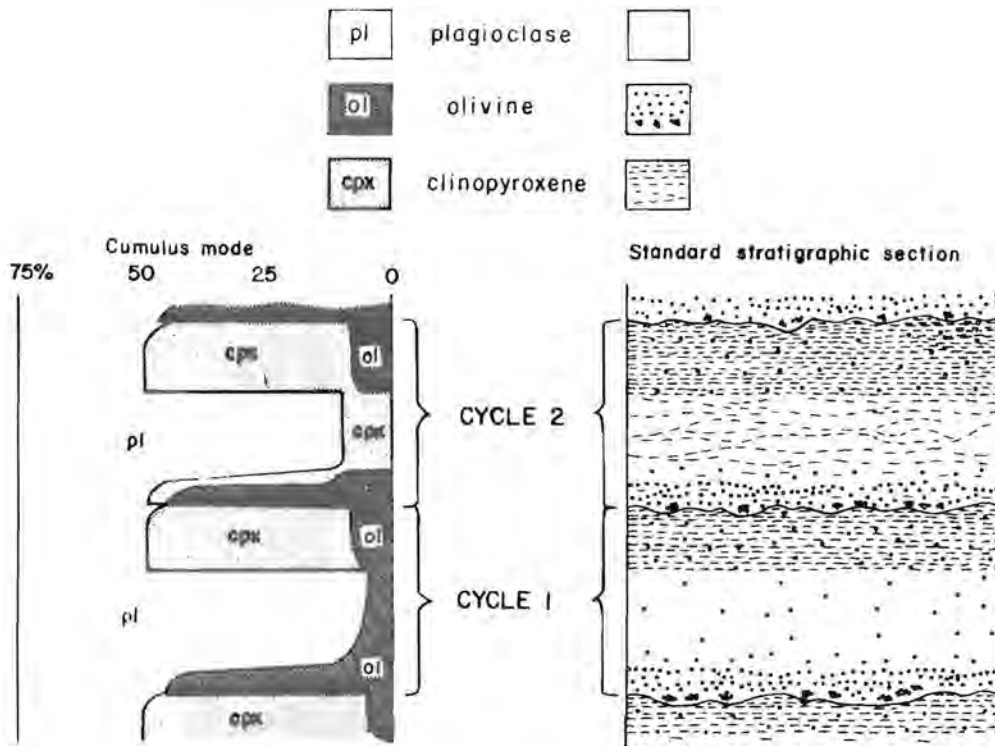


Figure 12. Idealized cycles in OBZ III and OBZ IV.

The upper 125 meters of this subzone are characterized by a four-phase cumulate. In some samples, small embayed olivines are present in the cores of orthopyroxenes, clearly indicating a reaction relationship. In other samples, however, coexisting orthopyroxene and olivine do not show the reaction relationship and both appear to be cumulus minerals. However, the distinction between cumulus and postcumulus orthopyroxene (including that formed by reaction from olivine) is not always obvious and reflects, in part, our limited understanding of the formation of "cumulate" rocks.

Olivine-bearing subzone IV (2339-2768 meters). The lower boundary of this subzone is placed at the base of a well banded troctolite and the upper boundary at the base of the second thick anorthosite. The sequence, troctolite-anorthosite-anorthositic troctolite-olivine gabbro (+ gabbronorite) is repeated three times. The gabbronorite unit at the top of the third cycle is variable in thickness and not always present. The uppermost troctolite may represent the basal member of an incomplete fourth cycle.

In the anorthositic troctolites of the first cycle, orthopyroxene and inverted pigeonite coexist as postcumulus phases with the former occurring exclusively as reaction rims around olivine while the latter form large oikocrysts.

The contact between troctolite and underlying olivine gabbro ranges from gradational to sharp, sinuous and discordant, whereas the upper contact between troctolites and overlying anorthosites is invariably sharp, planar and concordant. The uppermost troctolite is structureless, highly discordant

Raedeke, L.D.

contains ameboidal olivine aggregates, is associated with pyroxene and plagioclase megacrysts and is virtually identical to the troctolite at the base of the first thick anorthosite described previously (OBZ II).

Anorthosite II (2768-3338 meters). This uniform unit is the thickest anorthosite in the Banded zone. Postcumulus augite and inverted pigeonite make up 10 to 12 percent of the rocks and disseminated sulfides are concentrated in two narrow layers near the base and top of this unit. The average grain size of plagioclase is approximately twice that of plagioclase in two- and three-phase cumulates.

Raedeke, L.D.

AUGUST 8: PICKET PIN MOUNTAINUpper Banded zone

L.D. Raedeke and I.S. McCallum

Our traverse will begin in Olivine-bearing subzone IV of the Middle Banded zone and continue through the Upper Banded zone (figure 6). The Upper Banded zone is arbitrarily divided into two subzones, a lower olivine-bearing subzone (OBZ V) and an upper subzone of gabbronorite.

Olivine-bearing subzone V. The basal member of this subzone is a well-banded troctolite containing varied amounts of plagioclase, normally in excess of cotectic proportions. Modally graded layering, cross-bedding and cut-and-fill structures are locally present indicating strong current action during the deposition of this troctolite. Generally, the lithologic sequence upward is TAN/ANG/ANG/ANG where T = troctolite, A = anorthosite, N = norite, and G = gabbronorite. Irregular layering characterized by alternating mafic and felsic layers, is well developed in norites and gabbronorites. As noted earlier, layering is most conspicuous in those members containing "excess" plagioclase. In the lowest gabbronorite there is a conspicuous unit showing complex layering with anorthosite lenses and 2-5 cm "snowball" aggregates of pyroxene. About 20 meters above this complex unit, quartz veins cross-cut the gabbronorite. Sulfides are found in the uppermost anorthosite of the subzone.

Gabbronorite III. The lower contact of this subzone is marked by a sharp increase in pyroxene mode at a conformable contact between the upper anorthositic gabbronorite of OBZ V and Gabbronorite III. Throughout this uniform subzone, plagioclase, augite and low-Ca pyroxene occur in approximately cotectic proportions, and the mineral compositions vary in a regular and predictable manner. While planar lamination is almost always present, there is no preferred mineral orientation within the plane of layering. The most interesting aspect of this subzone is the change in morphology and composition of the low-Ca pyroxenes. Below the mafic dike*, orthopyroxene is clearly a cumulus mineral. Approaching the dike, orthopyroxene occurs as poikilitic crystals (up to 15 cm in diameter) containing abundant "inclusions" of small rounded augites. In the upper part of the section, orthopyroxene occurs as poikilitic crystals containing numerous sets of oriented "001" augite exsolution lamellae. These sets of lamellae outline domains corresponding to original cumulus pigeonite crystals. Similar features have been described by von Gruenewaldt (23) in pigeonites from the Bushveld Complex.

In the transition zone from cumulus orthopyroxene to cumulus pigeonite, poikilitic orthopyroxene may be mistaken for a postcumulus mineral, which is not the case. The apparent textural ambiguities among pyroxenes can be satisfactorily explained by consideration of the subsolidus pigeonite breakdown (7). It is well known that pigeonite replaces orthopyroxene as the liquidus phase as $Fe^{2+}/Fe^{2+} + Mg$ (Fe#) increases in the melt (cf. Hess (5)). In addition, the subsolidus field of pigeonite + augite extends over a progressively wider temperature range with increasing Fe# content in the system

*This mafic dike was intruded after the crystallization of the Upper Banded zone and has had no effect on the pyroxene compositions or habit discussed here. It is used in this discussion simply as a marker bed.

Raedeke, L.D.

((7), figure 1). On cooling, pigeonite exsolves augite (aug) along the solvus indicated in figure 13 until temperature T_a is reached, where it decomposes to opx + aug. Just at the value of Fe# where pigeonite becomes stable as the liquidus phase, the temperature range over which the subsolidus aug-pig field extends is at its most narrow. Exsolution of pigeonite therefore either does not occur or it occurs over a very narrow temperature range before decomposition to opx and blebby augite. As Fe# increases and the field enlarges, the

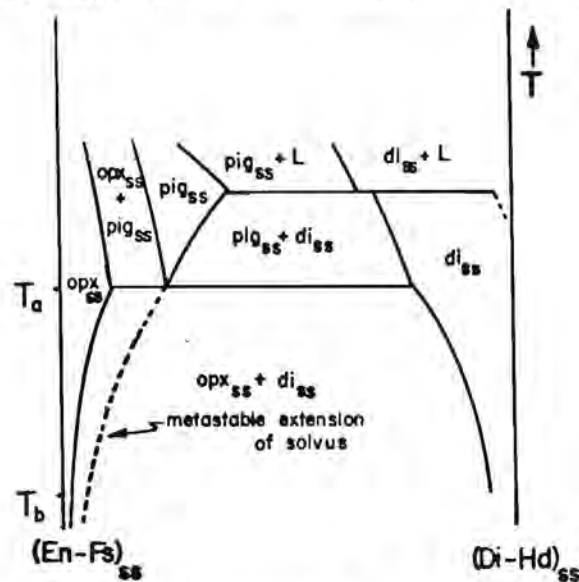


Figure 13. Schematic representation of pyroxene subsolidus relationships. Simplified from Ishii and Takeda (7), figure 4.

the temperature range (ΔT) over which exsolution takes place also increases. As ΔT increases, augite exsolves as (001) lamellae over a correspondingly larger interval. As these lamellae become a significant part of the crystal, they may tend to inhibit inversion from the monoclinic pigeonite structure, thereby allowing exsolution to continue along the metastable extension of the solvus curve, culminating eventually in decomposition to opx + aug at some lower temperature (e.g. T_b). It seems likely that the sequence of textures seen in Gabbronorite III is a result of postcumulus recrystallization accompanying the exsolution and inversion reactions. By this interpretation, the section just below the mafic dike marks the cross-over into the stability field of pigeonite. The zone above where poikilitic orthopyroxene coexists with cumulus augite, represents conditions where ΔT was small and cumulus pigeonite decomposed to opx + aug. Finally, in the upper section, ΔT was sufficiently large that well formed (001) exsolution lamellae could develop. Insofar as there is no significant change in mineral proportions associated with these textural changes, it is concluded that low-Ca pyroxene formed as a cumulus mineral throughout Gabbronorite III, the poikilitic texture being the result of postcumulus recrystallization accompanying the inversion reaction. The change from cumulus orthopyroxene to cumulus pigeonite is complicated by a reversal about midway through the gabbronorite. The change occurs at a thin but complex zone in which the sequence, fine-grained gabbronorite - anorthositic gabbronorite pegmatite - mafic gabbronorite pegmatite, is developed. Above this zone, the transitional change to pigeonite resumes. The upper gabbronorite contains numerous veins and dikes of pegmatitic hornblende-plagioclase-quartz up to 50 cm in width. These rocks may have formed during a late magmatic stage.

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**An Informal Guidebook for the Precambrian Rocks of the
Beartooth Mountains, Montana-Wyoming**

prepared jointly by

**J. Wooden
Lockheed-NASA
Johnson Space Center
Houston, Texas**

**P. Mueller
Dept. of Geology
University of Florida
Gainesville, Florida**

**D. Bowes
Geology Department
University of Glasgow
Scotland**

PREFACE

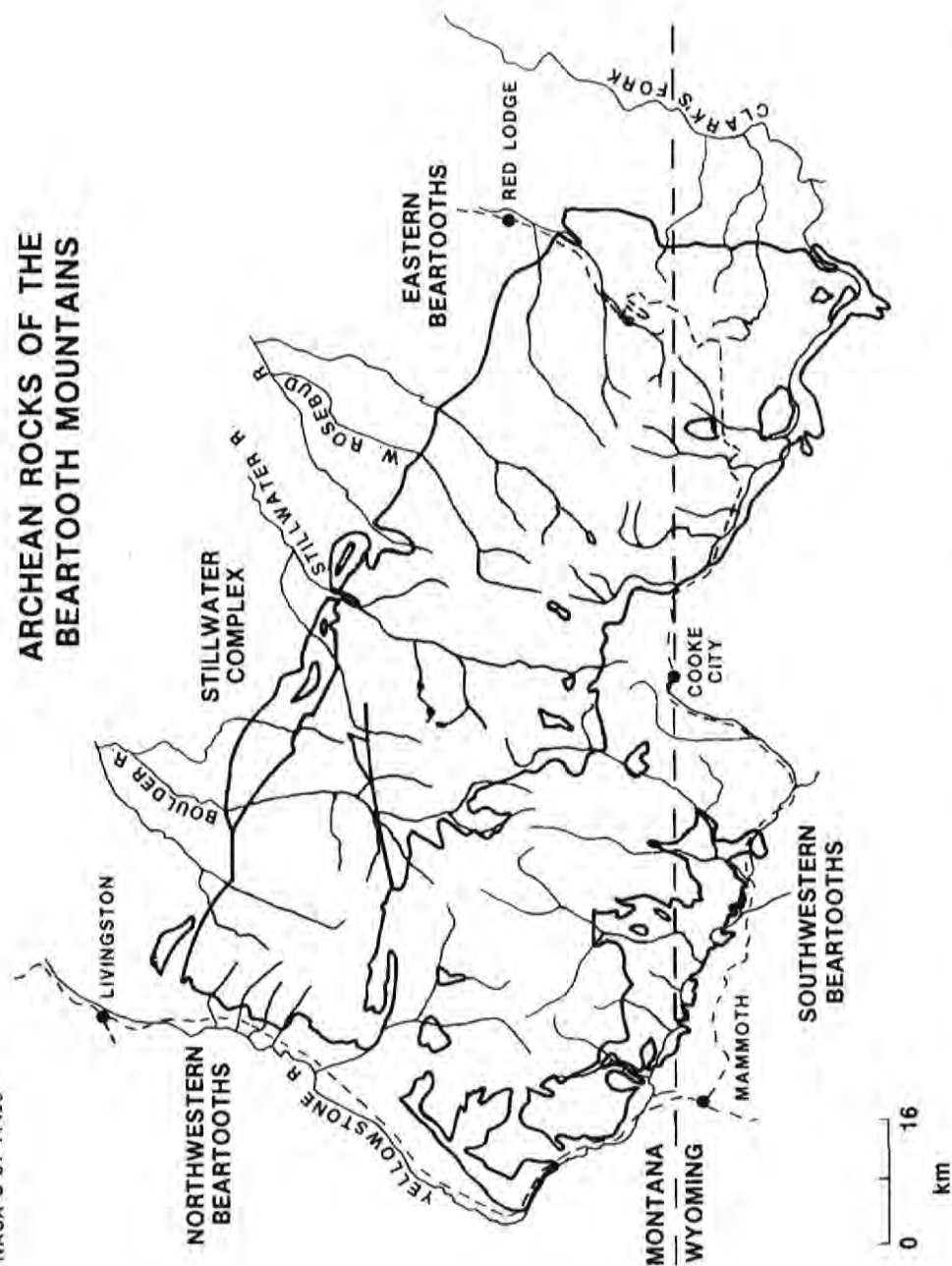
This guidebook for the Precambrian rocks of the Beartooth Mountains consists of figures and tables that summarize the authors' present knowledge of this topic. No formal text was prepared because we felt the information we wished to convey would be assimilated more easily and quickly in a graphical format. The Precambrian geologic history of the Beartooth Mountains is still being actively investigated, and the present summary will certainly be changed as new information is gained. This summary necessarily is biased toward our own data and a geochemical perspective. Our work has not been supported by the necessary detailed field-work from which all projects such as this greatly benefit; unfortunately, field-oriented projects are not favored by graduate students or funding agencies at this time. Other presently active workers in the Beartooths on whose knowledge we have drawn are C. J. Casella, Dept. of Geology, Northern Illinois University, who has conducted a field-oriented and petrologic study of the southwestern Beartooths, and David Mogk, Dept. of Geology, University of Washington, Seattle, who is conducting a field-petrologic-geochemical study of the northwestern Beartooths for his Ph.D. We have also benefited from the earlier work initiated by A. Poldervaart and carried out by R. D. Bentley, J. R. Butler, C. J. Casella, F. D. Eckelmann, R. L. Harris, Jr., L. H. Larsen, M. Prinz, L. Rowan, W. R. Skinner, and E. W. Spencer.

Our research has been supported by NSF, NASA, and the NERC of the U. K.



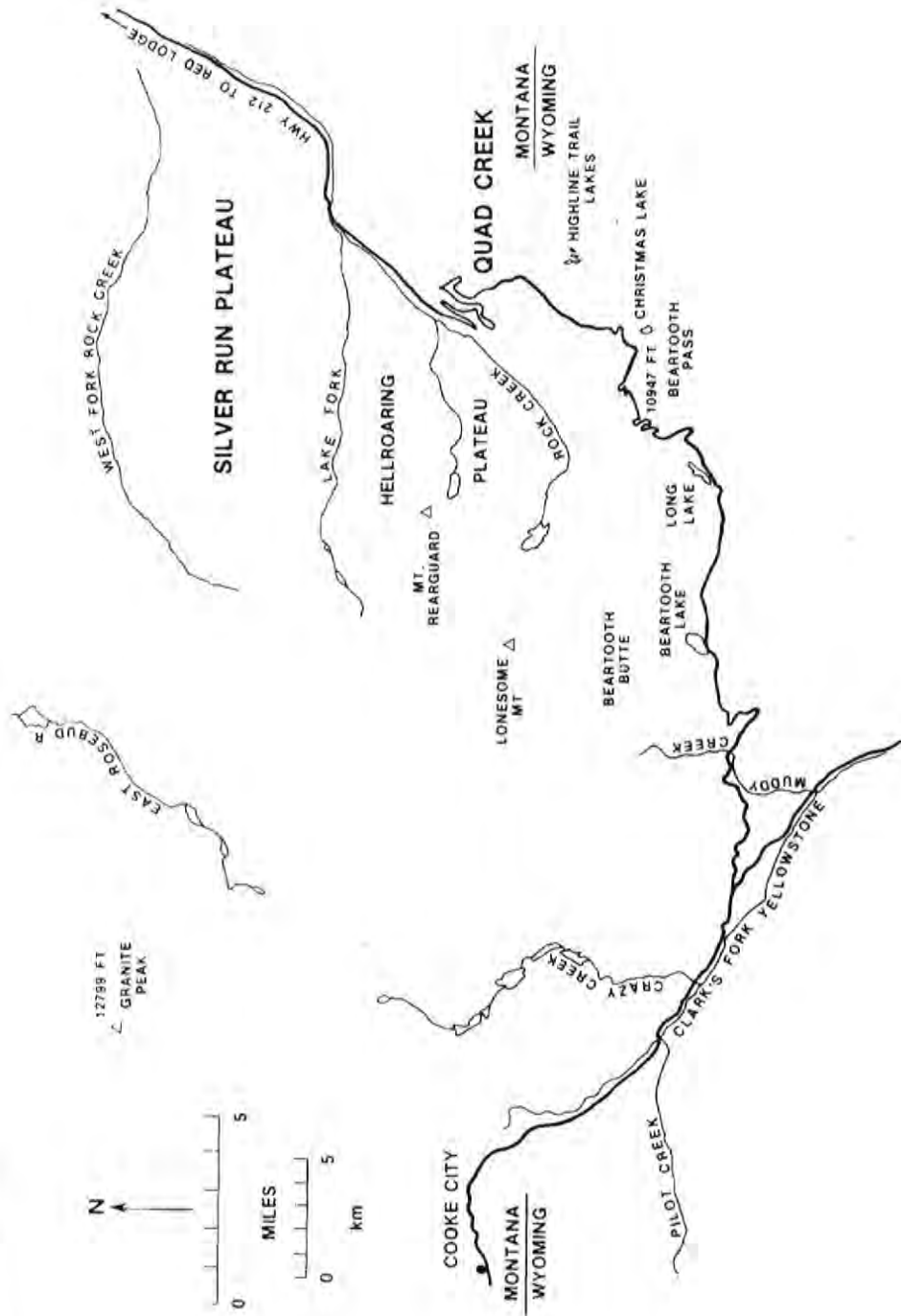
Generalized geologic map showing areas of Precambrian rocks in the northern Rocky Mountain region. The Beartooth Mountains belong to what is commonly termed the Wyoming Archean province.

ARCHEAN ROCKS OF THE BEARTOOTH MOUNTAINS



The Archean rocks of the Beartooth Mountains are divided into four geographic areas just for ease of discussion. The areas are: eastern, Stillwater Complex area, northwestern, and southwestern. The southwestern and northwestern areas are distinctive from the other two areas, and each other, because of the large amounts of metasedimentary rocks found in each area. The eastern area, which is dominated by meta-igneous rocks, is in large part gradational into the Stillwater area. The metasedimentary rocks intruded by the Stillwater Complex may be gradational into similar rocks in the northwestern area.

NASA 1-67-11499



This generalized map focuses on the area in the eastern Beartooths where the Archean field trip will be conducted. Stops will be made along Hwy 212 at rocks exposures at (1) upper and lower Quad Creek along the major, eastern set of switchbacks; (2) the eastern end of Long Lake; (3) the western end of Beartooth Lake; and (4) the large switchback between Beartooth Lake and Muddy Creek. If the weather is clear, additional stops will be made to view the Alpine scenery of the area.

TENTATIVE ARCHEAN-PROTEROZOIC HISTORY OF THE SOUTHEASTERN BEARTOOTH MTNS.

3.3-3.4 AE Supracrustal sequence dominated equally by mafic and low-K silicic rocks with associated sedimentary rocks including graywackes, shales, quartzites, ironstones.

? High T, Moderate P Metamorphism.

2.9-3.1 AE Additional supracrustal rocks deposited. Sequence dominated by andesitic rocks.

? Moderate T Metamorphism.

2.8 AE Large volumes of intrusive granitic rocks composing several magma series invaded older metamorphosed supracrustal rocks. Low-Moderate T Metamorphism.

2.5-2.7 AE Mafic dikes.

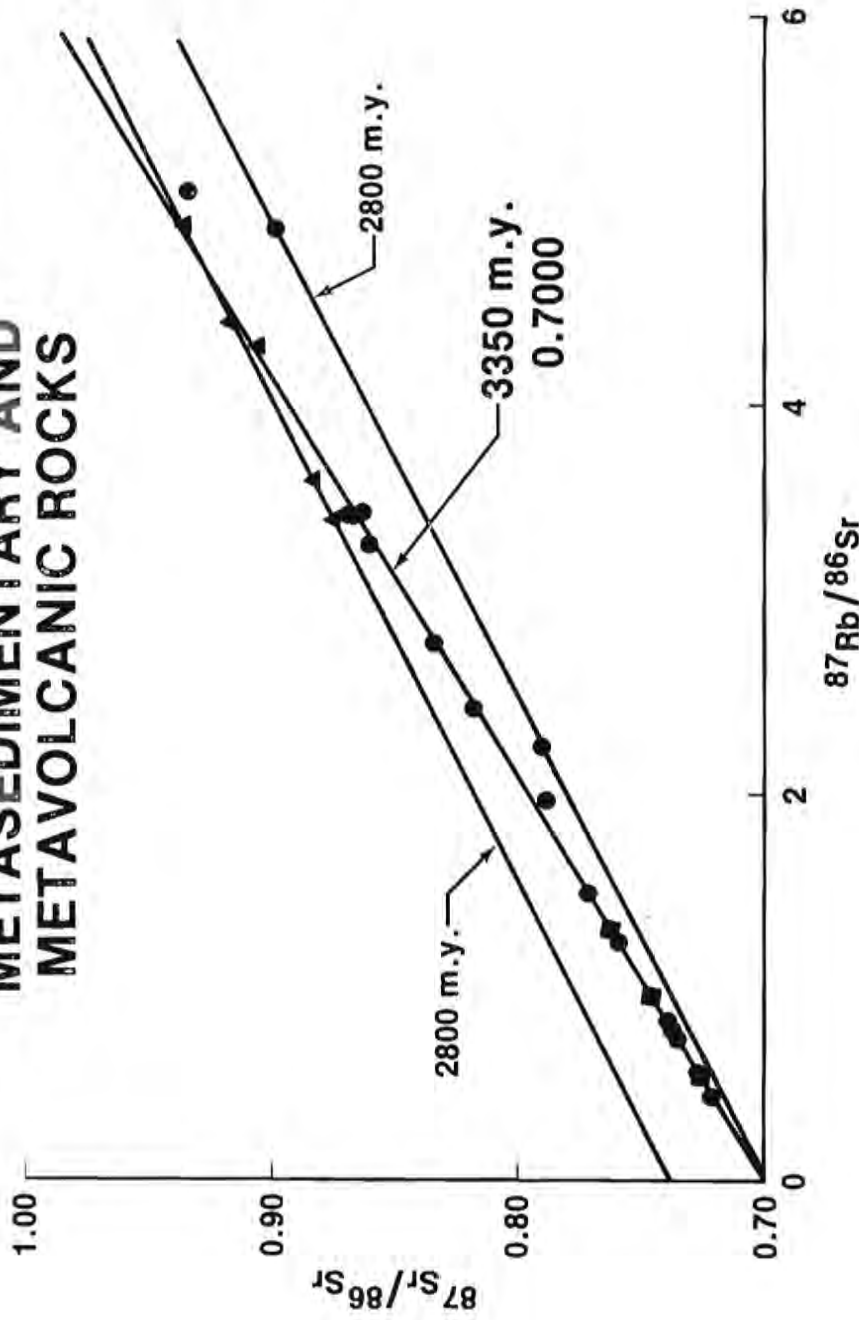
2.0-2.2 AE Mafic dikes.

1.33 AE Alkali-olivine dikes.

0.74 AE Quartz tholeiite dikes.

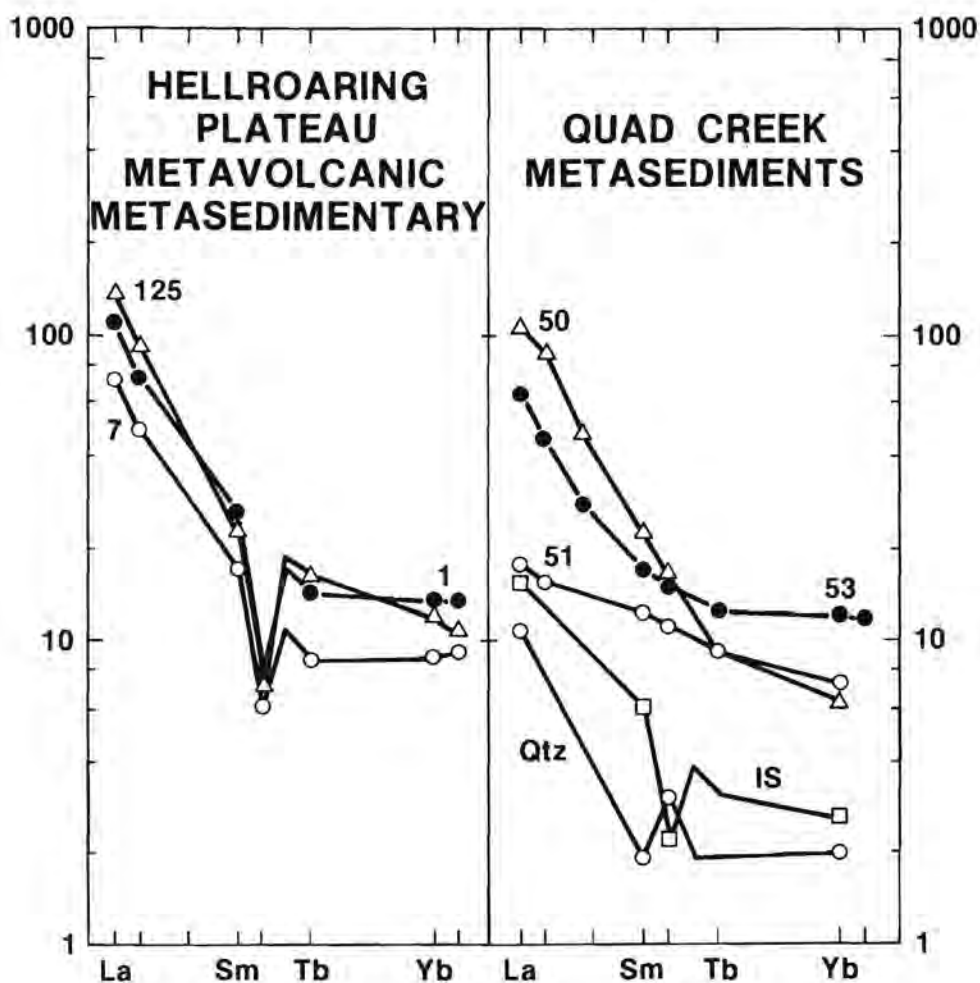
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QUAD CREEK AND HELLROARING PLATEAU METASEDIMENTARY AND METAVOLCANIC ROCKS



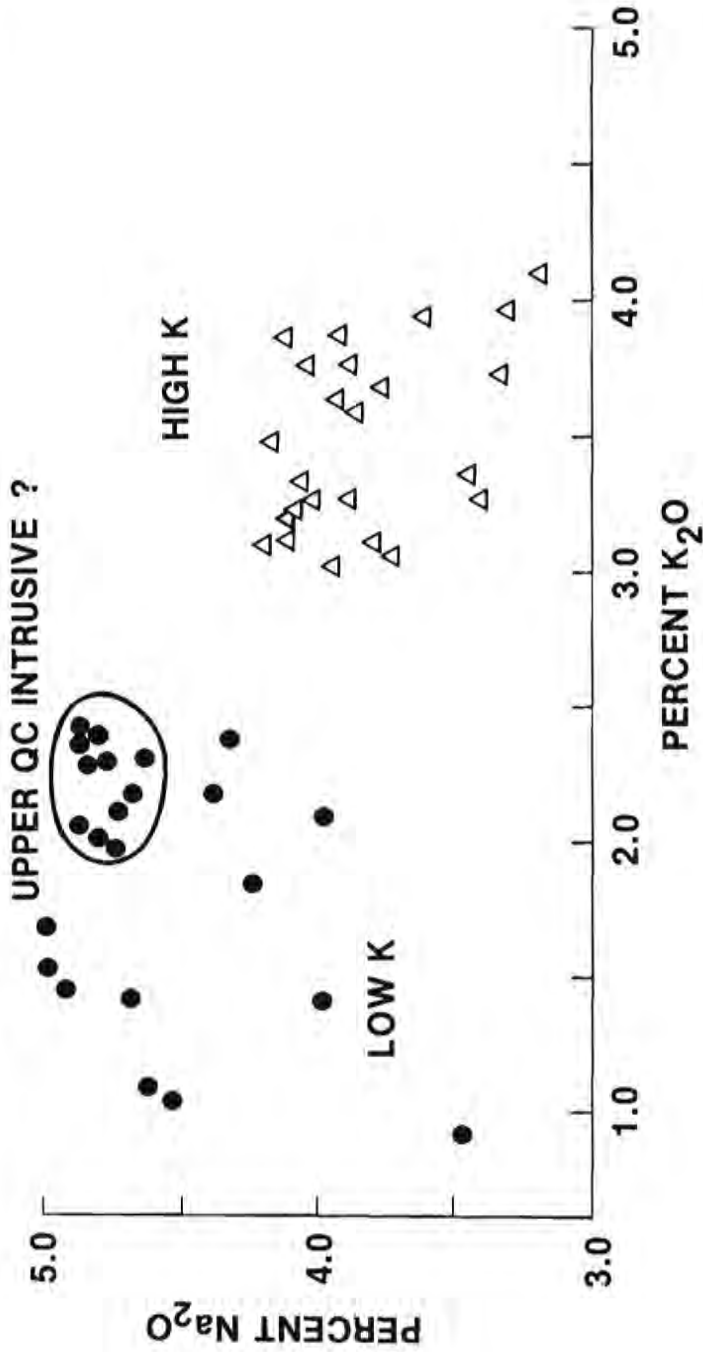
Rb-Sr systematics of the oldest identified Beartooth rocks. The approximate 3.35AE age of these may represent the time of their origin and/or the time of a high grade metamorphic event. The scatter of some of the data away from the 3.35AE line is interpreted as the result of the 2.8AE Beartooth intrusive-metamorphic event. Samples from a single outcrop on Hellroaring Plateau (triangles) appear to have been rehomogenized at about 2.8AE and show an elevated initial Sr ratio.

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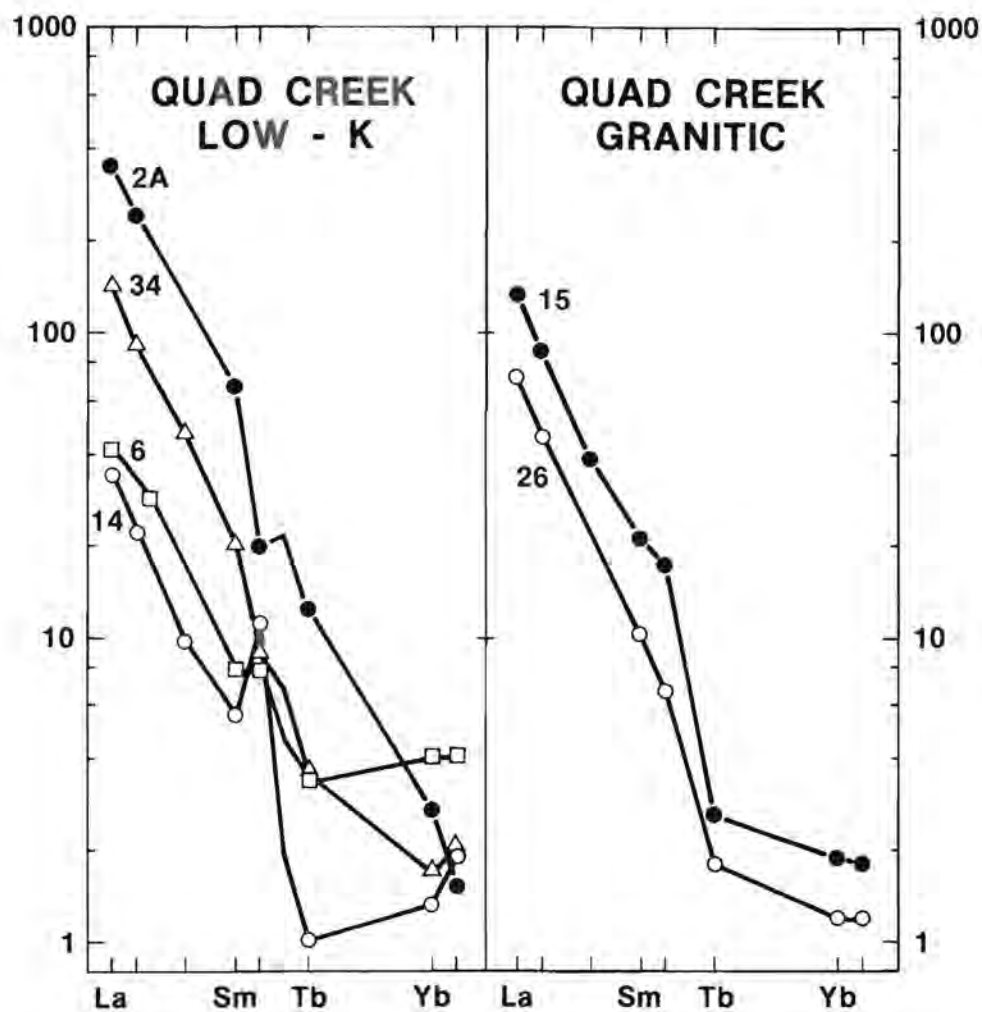
Chondrite normalized REE patterns for some of the oldest (~ 3.3 AE) identified rocks in the Beartooths. For the Hellroaring Plateau rocks nos. 1 and 7 are rhyodacitic and no. 125 graywacke in composition-mineralogy. Compositional data is not available for the Quad Creek rocks but their mineralogies suggest that nos. 50 and 53 are graywackes, 51 a basalt, IS an ironstone, and Qtz a quartzite.

QUAD CREEK QUARTZFELDSPATHIC ROCKS



Na₂O vs K₂O for Quad Creek silicic rocks. Two general groups - low K and high K - are recognizable. Many of the samples in each group have compositions that would be classified as igneous. The low K samples with the lower Na contents, however, are unusual and may be volcanoclastic - sedimentary in origin. Samples labeled as upper QC intrusive represent a homogeneous geochemical group with Rb-Sr systematics that consistently suggest a 2.8AE age. The field characteristics and appearance of some of the high K rocks requires that K metasomatism be considered as a factor in their origin.

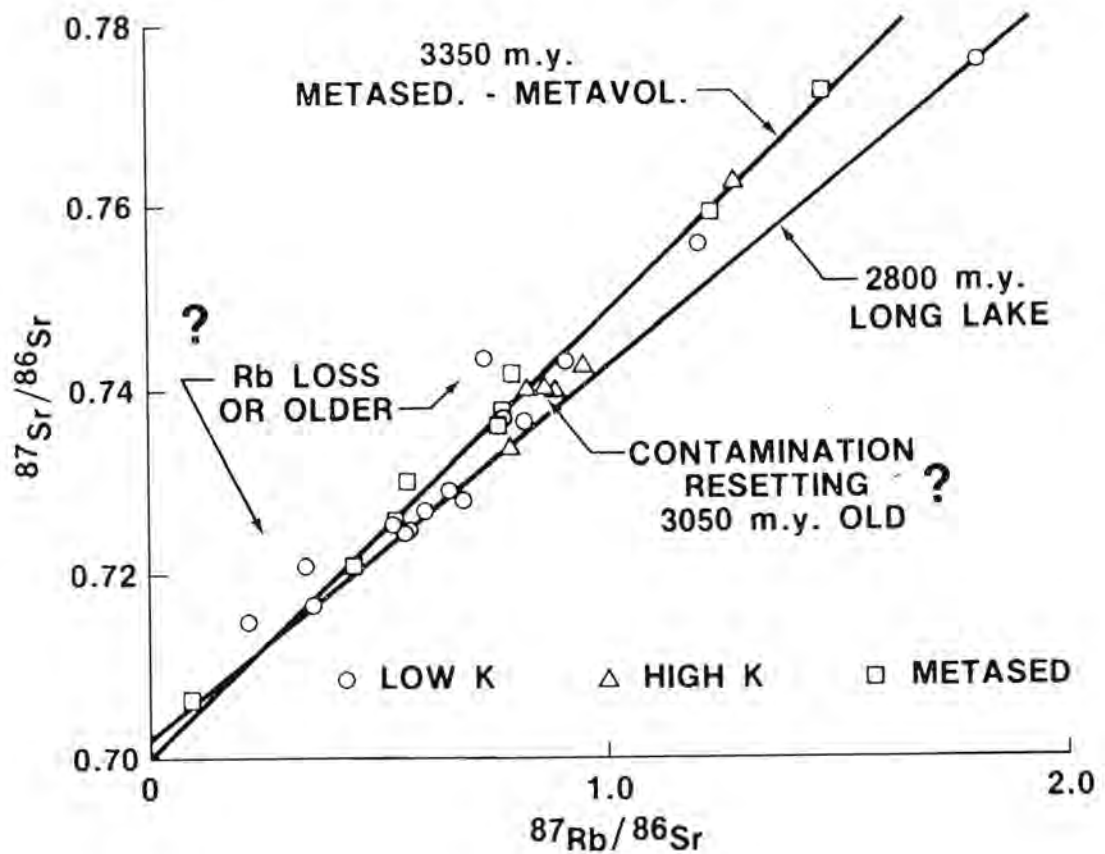
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Chondrite normalized REE patterns for the Quad Creek silicic rocks. The patterns and overall geochemistry of the granitic rocks resemble those of the Long Lake granite. The origin of the low-K rocks is problematical and ranges from igneous, volcanoclastic, sedimentary.

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QUAD CREEK AREA



Rb-Sr isotopic data for rocks from the Quad Creek area. The scatter of the data is not clearly interpretable. One possible interpretation is that a group of metavolcanic-metasedimentary rocks with a minimum age of 3.35AE were intruded at 2.8AE. The 2.8 intrusive rocks were variably contaminated by the older rocks. Some movement of Rb may have occurred as a result of the post 2.8AE tectonism of these rocks.

QUAD CREEK AND HELLROARING PLATEAU SILICIC ROCKS

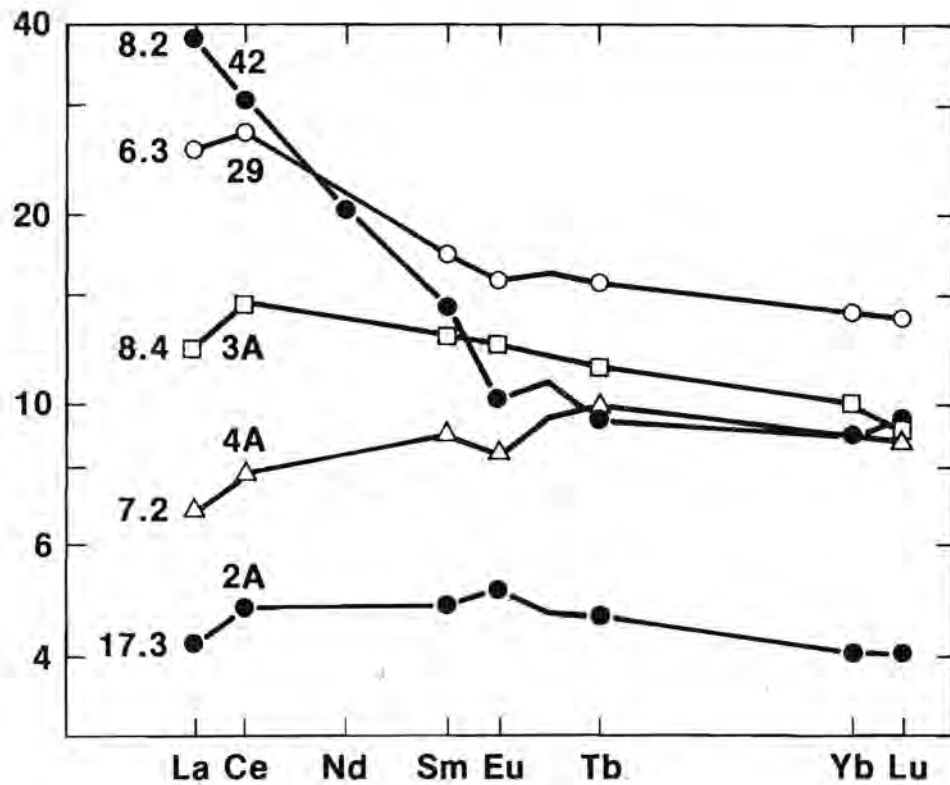
	QUAD CREEK LOW-K			QUAD CREEK HIGH-K			HELLROARING METAVOLCANIC				
	QC ZA	QC 13	QC 14	QC 19	QC 34	QC 15	QC 20	QC 26	QC 30	HP 1	HP 9
SiO ₂	69.8	66.0	76.0	70.7	72.7	73.0	73.8	74.8	74.7	74.9	69.8
TiO ₂	0.10	0.41	0.12	0.32	0.22	0.12	0.11	0.14	0.14	0.12	0.44
Al ₂ O ₃	17.4	16.3	12.7	15.4	15.0	14.0	13.7	13.7	13.5	13.5	15.0
Fe ₂ O ₃	0.38	1.02	0.43	0.72	0.72	0.28	0.21	0.37	0.44	0.69	2.04
FeO	0.66	2.29	0.69	1.48	1.10	0.74	0.70	0.70	0.61	0.50	2.20
MnO	0.02	0.06	0.02	0.02	0.02	0.02	0.02	0.01	0.01	0.03	0.08
MgO	0.63	1.48	0.63	0.80	0.66	0.62	0.50	0.43	0.41	0.40	1.26
CaO	4.70	4.06	3.64	2.93	2.29	2.46	2.53	1.67	1.95	1.55	3.57
Na ₂ O	4.82	4.76	3.54	4.88	4.64	3.30	3.36	4.02	4.40	3.74	3.88
K ₂ O	1.01	1.47	0.82	1.73	2.51	4.00	3.50	4.18	3.20	4.12	1.55
P ₂ O ₅	0.06	0.15	0.02	0.04	0.05	0.03	0.04	0.04	0.03	0.06	0.17
H ₂ O+	0.49	0.91	0.39	0.49	0.80	0.43	0.67	0.37	0.72	0.48	0.81
CO ₂	0.18	0.32	0.09	0.21	0.17	0.05	0.27	0.21	0.26	0.66	0.03
Ba	230	226	214	888	470	2250	1370	1180	1140	340	300
Rb	31	78	32	55	64	103	81	86	74	121	127
Sr	380	381	263	274	262	340	279	219	239	106	188
Zr	51	177	132	218	137	44	59	47	63	74	173

AMPHIBOLITES

	QUAD CREEK		HELLROARING PLATEAU			
	QC 3	QC 42	HA 2A	HA 3D	HA 2C	HA 3A
SiO ₂	49.6	52.8	47.8	49.7	56.3	49.0
TiO ₂	0.73	0.58	0.39	0.42	1.01	0.95
Al ₂ O ₃	13.3	14.7	6.84	9.59	9.72	13.3
Fe ₂ O ₃	2.91	1.99	4.48	2.89	4.05	3.10
FeO	8.97	7.35	6.90	8.21	7.32	8.85
MnO	0.17	0.16	0.19	0.21	0.22	0.17
MgO	9.31	8.22	17.3	13.1	10.4	8.44
CaO	10.2	10.2	11.8	10.9	5.24	10.5
Na ₂ O	2.03	2.24	0.89	1.52	1.60	2.19
K ₂ O	0.27	0.36	0.17	0.64	0.45	0.89
P ₂ O ₅	0.05	0.08	0.04	0.03	0.36	0.08
H ₂ O ⁺	1.56	0.57	2.19	2.09	2.26	1.87
CO ₂	0.09	0.10	0.11	0.10	0.10	0.10
Cr	254	251	1820	1260	600	284
Ni	48	100	450	380	320	70
Sr	88	121	15	79	150	161
Zr	52	79	35	44	233	65

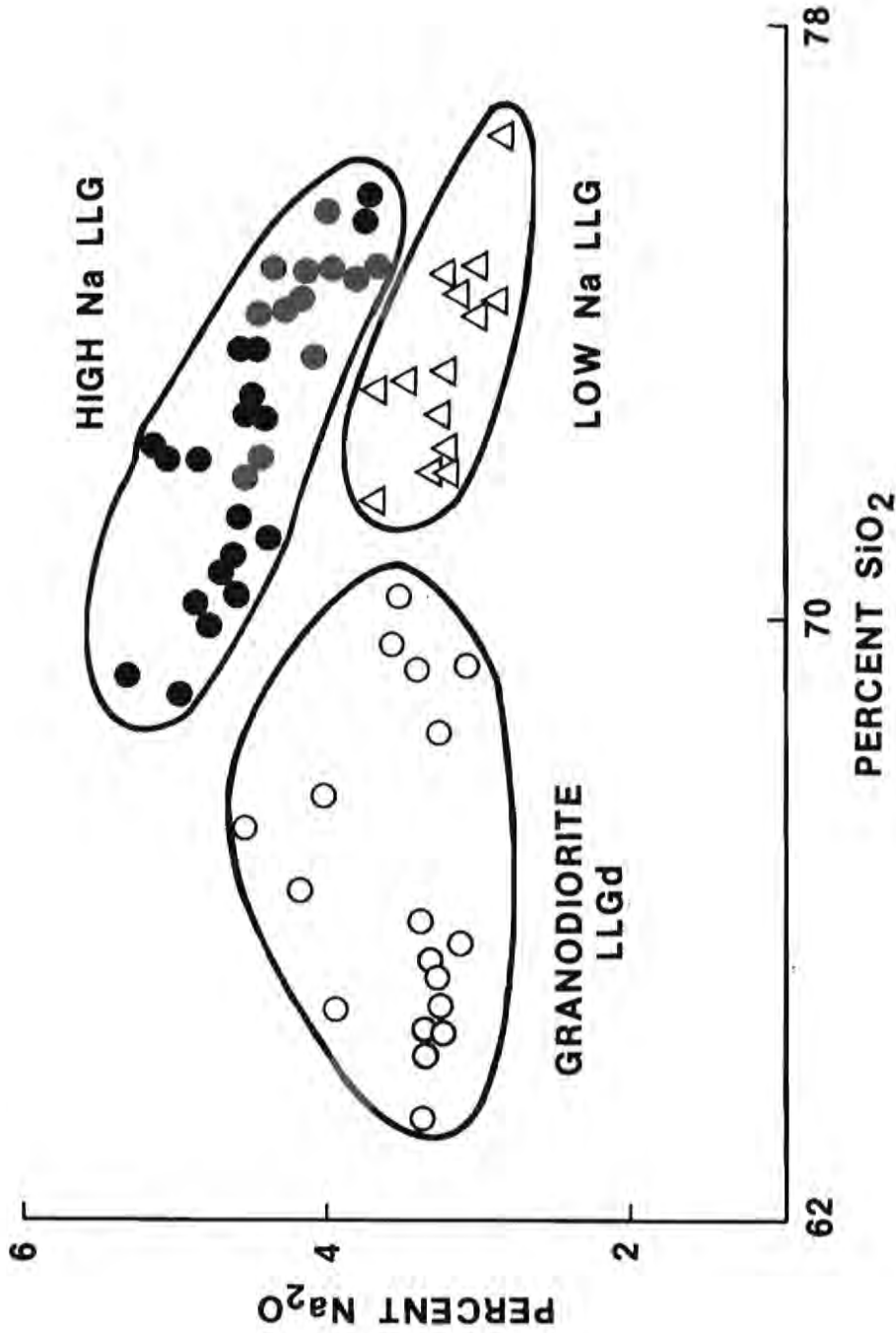
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BEARTOOTH BASALTIC AMPHIBOLITES



Chondrite normalized REE patterns for basaltic amphibolites from Hell-roaring Plateau (2A, 3A, 4A), Quad Creek (42), and Long Lake (29). Other nos. beside patterns refer to their MgO content.

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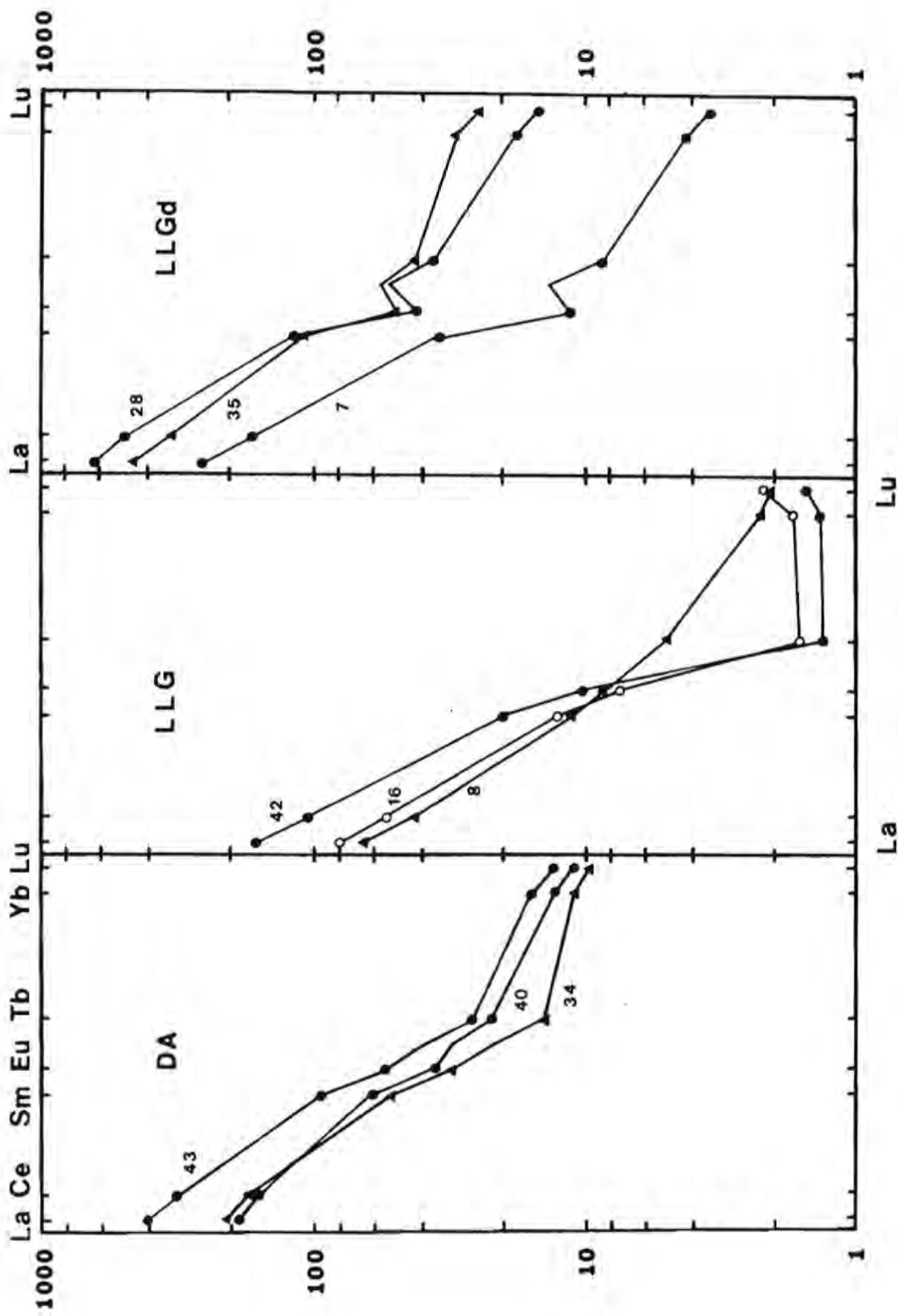
SiO₂ vs Na₂O for magma series in the Long Lake group. Three magma series have been identified among the later intrusive rocks of the interior of the eastern Beartooths. The two Long Lake granite (LLG) groups are 2.8AE in age. The Long Lake granodiorite (LLGd) is found as inclusions in the LLG rocks, but doesn't appear to be older than 3.0AE. Both the high and low Na LLG vary from tonalite to granite in modal mineralogy and geochemistry but maintain a very high SiO₂ content. This feature is a characteristic of other Archean rocks, but is unusual for modern intrusive rocks.

LONG LAKE ROCK SERIES

	HIGH Na GRANITE		LOW Na GRANITE		GRANODIORITE	
	BTR 46	BTR 18	BTR 11	WRS 71-41	WRS 70-31	BTR 35
SiO ₂	71.9	74.7	72.0	74.2	64.8	69.4
TiO ₂	0.25	0.19	0.27	0.21	0.93	0.80
Al ₂ O ₃	13.8	13.7	14.5	14.1	16.1	13.6
Fe ₂ O ₃	1.05	0.67	0.90	0.67	2.47	2.67
FeO	1.01	0.90	1.06	0.76	2.63	2.00
MnO	0.01	0.02	0.02	0.02	0.09	0.06
MgO	0.68	0.56	0.79	0.58	1.53	1.07
CaO	2.42	1.78	2.79	1.25	3.33	3.06
Na ₂ O	4.57	4.16	3.35	2.97	3.30	3.44
K ₂ O	2.60	3.50	4.22	4.91	3.66	3.34
P ₂ O ₅	0.06	0.05	0.11	0.07	0.30	0.31
H ₂ O ⁺	0.52	0.59	0.53	0.56	0.43	0.93
CO ₂	0.21	0.07	0.04	0.06	0.00	0.26
Ba	1190	1150	2170	1930	2620	2460
Rb	72	85	118	145	195	101
Sr	315	227	313	330	510	458
Zr	130	101	149	140	405	301

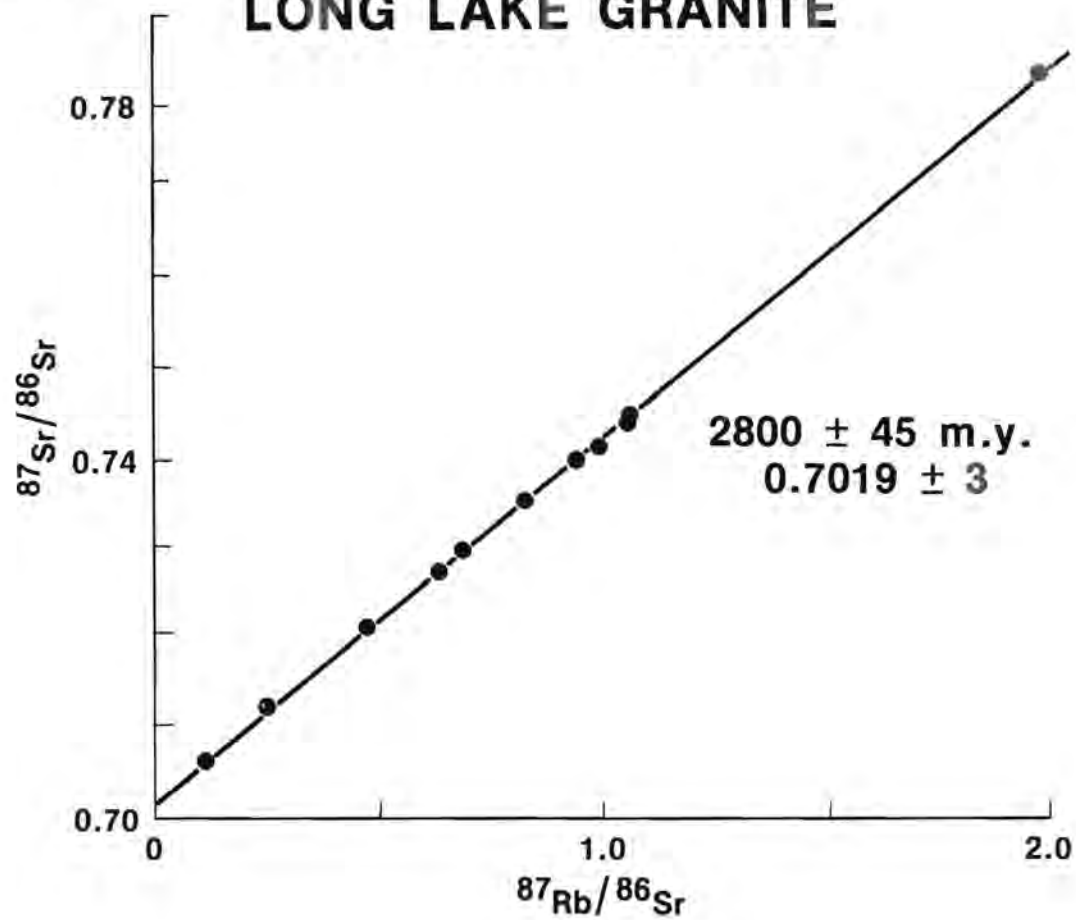
AMPHIBOLITES INCLUDED IN LONG LAKE GRANITIC ROCKS

	DIORITIC			BASALTIC		
	BTR 26	BTR 34	BTR 40	BTR 30	BTR 31	BTR 60
SiO ₂	55.1	56.1	56.3	50.3	51.4	48.3
TiO ₂	0.73	0.84	1.24	1.07	0.70	0.48
Al ₂ O ₃	16.2	15.1	14.5	14.0	8.41	8.24
Fe ₂ O ₃	3.18	3.78	4.96	4.88	4.63	3.53
FeO	4.75	5.22	5.04	8.43	8.98	6.74
MnO	0.11	0.16	0.16	0.23	0.29	0.17
MgO	4.58	4.85	3.34	6.87	10.9	21.5
CaO	7.78	6.77	7.95	9.81	8.71	6.30
Na ₂ O	3.25	3.46	3.13	2.61	0.96	1.48
K ₂ O	1.93	1.55	2.10	0.72	2.39	0.42
P ₂ O ₅	0.29	0.39	0.55	0.11	0.15	0.15
H ₂ O ⁺	1.46	0.88	1.32	1.37	1.51	2.06
CO ₂	0.43	0.10	0.06	0.01	0.02	0.64
Cr	84	140	69	255	834	2020
Ni	47	40	23	79	358	690
Sr	800	440	540	98	78	-
Zr	104	156	190	86	59	54
						107
						81
						97
						75

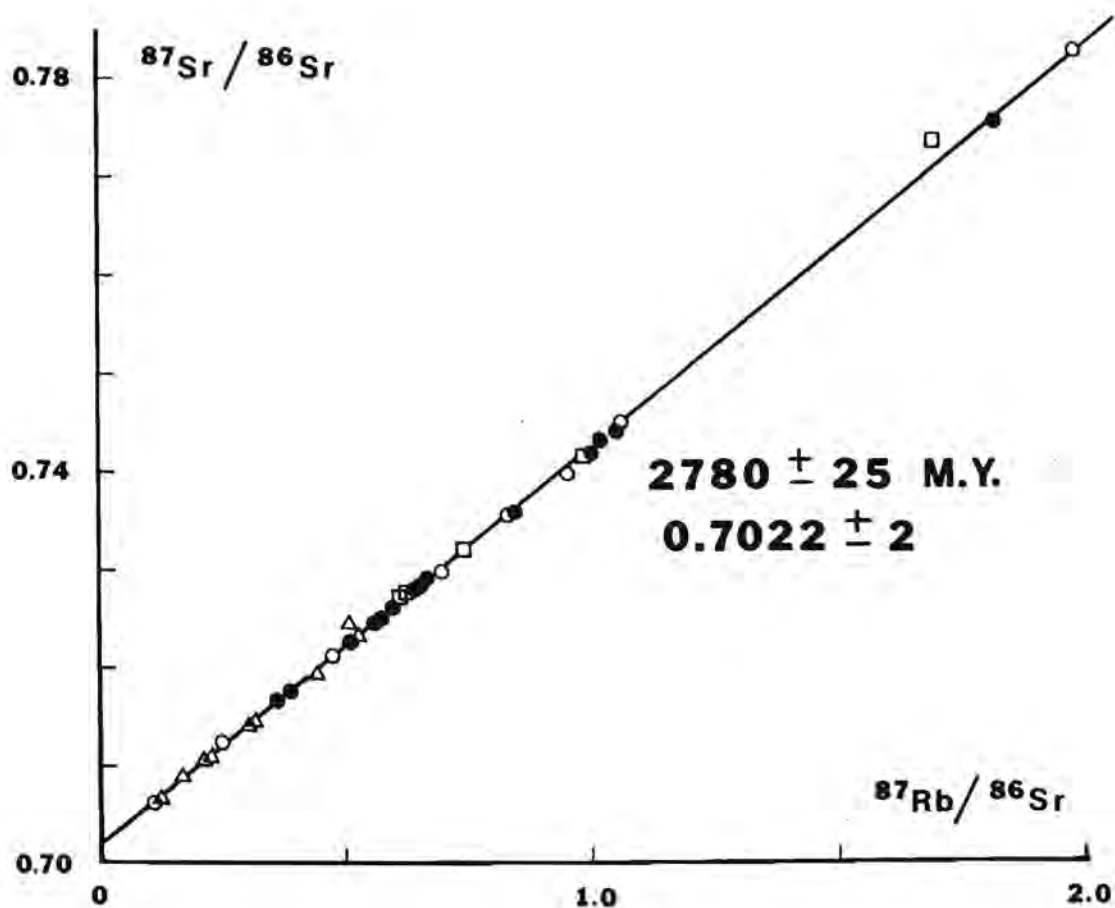


Chondrite normalized REE patterns for samples of dioritic amphibolite (DA), Long Lake granite (LLG), and Long Lake granodiorite (LLGd). These three rock types which dominate the eastern Beartooth area are easily distinguished by their trace element contents. Both the DA and LLGd samples must have been derived from sources enriched in light REE and feldspar has played an important role in the origin of the LLGd samples. LLG samples 16 and 42 are high Na types while sample 8 is a low Na type.

NASA-S-81-11503

LONG LAKE GRANITE

Rb-Sr isochron for rocks from the Long Lake granite suite. The age is interpreted as the time of intrusion. The initial ratio may suggest that crustal recycling has played a role in the origin of these rocks.



Composite Rb-Sr isochron for rocks from the eastern and southwestern Beartooths indicating the widespread nature of the 2.8AE event. Samples may lie on this isochron either because they were intruded at 2.8AE or because they were reset by the thermal pulse accompanying this event. Rb-Sr dating just beginning in the Stillwater and northwestern Beartooth areas indicate that this 2.8AE event was important in those areas also.

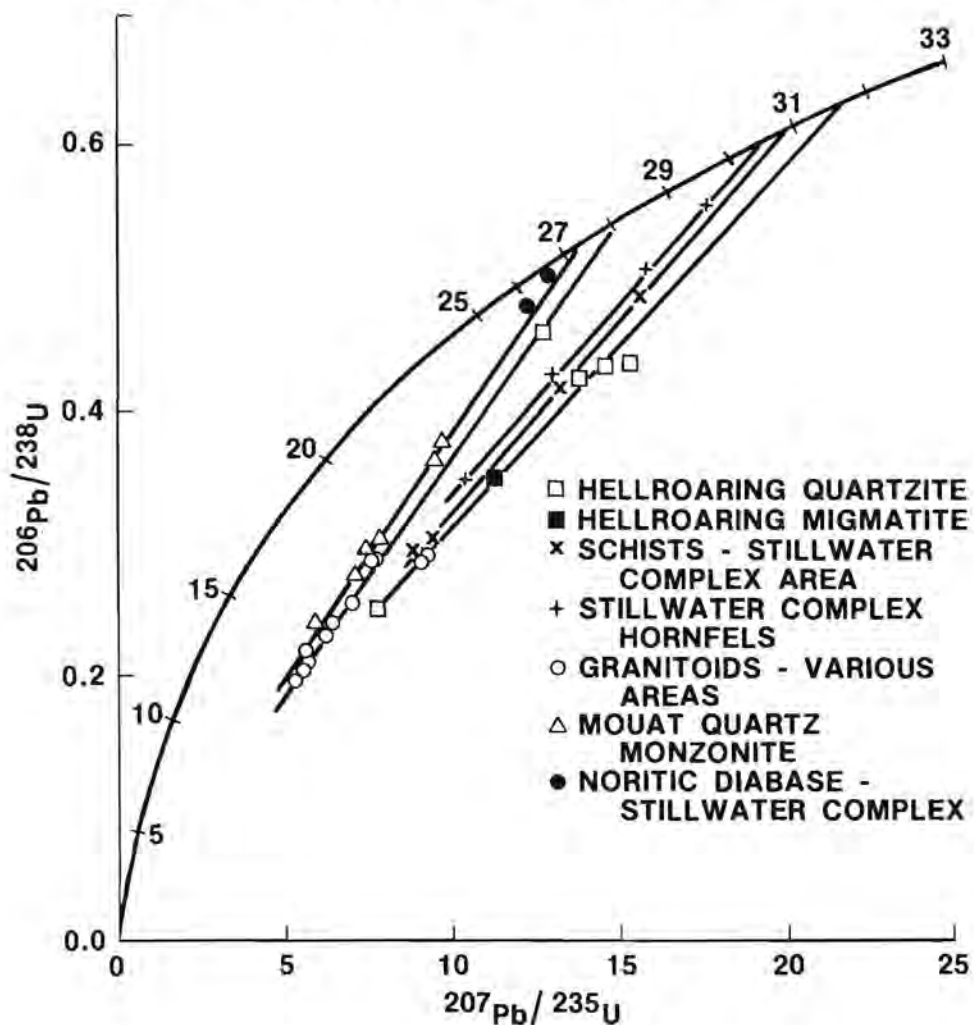
<u>SAMPLE</u>	<u>ROCK TYPE</u>	$^{147}\text{Sm}/^{144}\text{Nd}$	MODEL AGE, 10^9 YR	
			<u>CHONDRITIC</u>	<u>DEPLETED MANTLE</u>
HP-1	Hellroaring Plateau Metavolcanic	0.1228	3.33	3.45
DSC-110	Stillwater Hornfels	0.1207	3.13	3.27
BTR-79	Southwestern BT Metasediment	0.1244	2.92	3.10
QC-2A	Quad Creek Low-K Quartzofeldspathic	0.0965	2.96	3.08
QC-15	Quad Creek High-K Quartzofeldspathic	0.0946	2.95	3.07
BTR-34	Southeastern BT dioritic amphibolite	0.1028	2.91	3.06
DSC-131	Stillwater River Valley granitic	0.1131	2.86	3.01
BTR-16	Long Lake "granite"	0.0958	2.87	3.00
BTR-35	Long Lake "granodiorite"	0.1040	2.84	2.98
QC-34	Quad Creek Low-K granitic	0.0908	2.77	2.91

Notes: DSC-110 and 131 from DePaolo and Wasserburg (1979). All other data from JSC.

Chondritic model ages based on JSC eucrite data of $^{147}\text{Sm}/^{144}\text{Nd} = 0.1953$ and present day $^{143}\text{Nd}/^{144}\text{Nd} = 0.51180$. Depleted mantle ages after DePaolo (1981).

NASA-S-81-11500

BEARTOOTH ZIRCON DATA



The zircon data for this diagram come from a variety of sources including Nunes and Tilton (1971), Catanzaro and Kulp (1964), Mueller et al (in press), and Montgomery (in press). Geologic events interpreted from this diagram are: (1) intrusion of Stillwater Complex at 2.7AE; (2) intrusion of Mouat quartz monzonite at 2.72AE; (3) intrusion, deformation, metamorphism of granitic rocks at 2.8AE; (4) a complex set of events at 3.05-3.20AE that may represent a combination of events such as emplacement of igneous rocks, development of a sedimentary sequence, deformation and metamorphism.

SUMMARY OF CHRONOLOGIC DATA FOR THE STILLWATER COMPLEX AREA

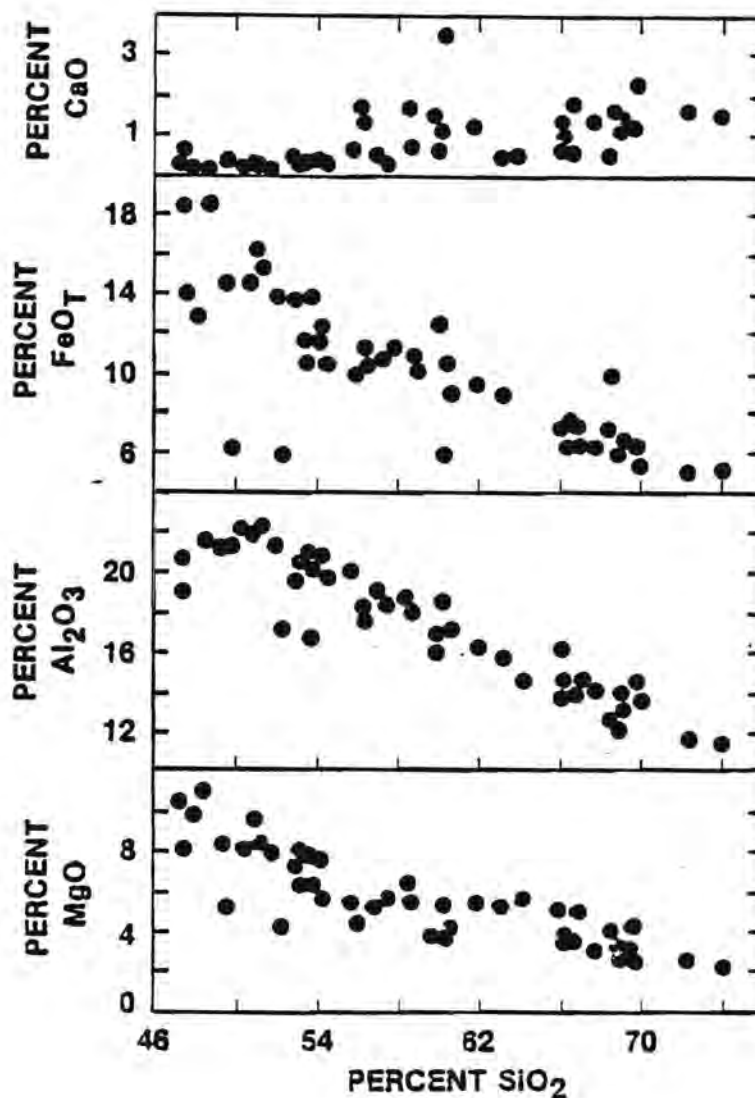
<u>ROCK TYPE</u>	<u>AGE, 10⁶ YR</u>	<u>TECHNIQUE</u>	<u>REFERENCE</u>
Mafic Dikes	1330	Rb-Sr WR	Mueller and Baadsgaard (1975)
	2100-2200		Wooden (1975)
	2500-2700		
Stillwater Complex Igneous Rocks	2701±8	Sm-Nd Mineral	DePaolo and Wasserburg (1979)
	2653±72	Sm-Nd WR	
Stillwater Complex Noritic Diabase	2700	Pb-Pb zircon	Nunes and Tilton (1971)
Stillwater Complex Hornfels	2692±45	Rb-Sr WR	Mueller and Wooden (1976)
Mouat Quartz Monzonite	2730	U-Pb concordia intercept	Nunes and Tilton (1971)
Granitic Gneiss	2805	Pb-Pb zircon	Nunes and Tilton (1971)
Granitic Gneiss	2860	Sm-Nd model age (chond.)	DePaolo and Wasserburg (1979)
	3010	Sm-Nd model age (depleted)	
Stillwater Complex Hornfels	3060	U-Pb concordia intercept	Nunes and Tilton (1971)
Schist	3090	U-Pb concordia intercept	Nunes and Tilton (1971)
Stillwater Complex Hornfels	3130	Sm-Nd model age (chond.)	DePaolo and Wasserburg (1979)
	3270	Sm-Nd model age (depleted)	

ARCHEAN ROCKS OF THE STILLWATER COMPLEX AREA

	FOLIATED GRANITIC ROCKS			INTRUSIVE QUARTZ MONZONITE				
	B-70B 7	XN 19	XW 35A	B-92B 7	51NB 69	59BE 69	57MV 69	57BE 69
SiO ₂	74.4	74.1	74.2	71.6	69.5	70.9	73.1	74.0
TiO ₂	0.21	0.15	0.08	0.27	0.36	0.29	0.13	0.17
Al ₂ O ₃	13.9	14.0	14.7	14.6	15.4	14.9	14.5	13.6
Fe ₂ O ₃	0.24	0.45	0.45	0.74	1.00	1.3	0.64	0.91
FeO	2.27	1.04	1.07	2.19	1.7	1.5	0.96	0.72
MnO	0.03	0.03	0.02	0.05	0.04	0.04	0.04	0.0
MgO	1.00	0.44	0.35	0.85	1.0	0.93	0.41	0.54
CaO	2.17	1.68	1.78	2.64	2.3	2.1	1.5	1.7
Na ₂ O	4.33	4.93	5.52	4.85	3.6	3.6	4.0	3.7
K ₂ O	0.96	2.28	2.00	1.42	3.6	4.2	3.4	4.1
P ₂ O ₅	0.04	0.04	0.05	0.05	0.10	0.09	0.03	0.08
H ₂ O+	0.66	0.72	0.10	0.55	1.2	0.0	0.46	0.28
CO ₂	0.13	0.21	0.13	0.20	0.05	0.05	0.05	0.05
Ba	170	560	440	640	2000	1500	1500	1000
Rb	117	77	147	152	-	-	-	-
Sr	285	310	375	350	300	300	200	150
Zr	100	120	115	335	150	70	70	100

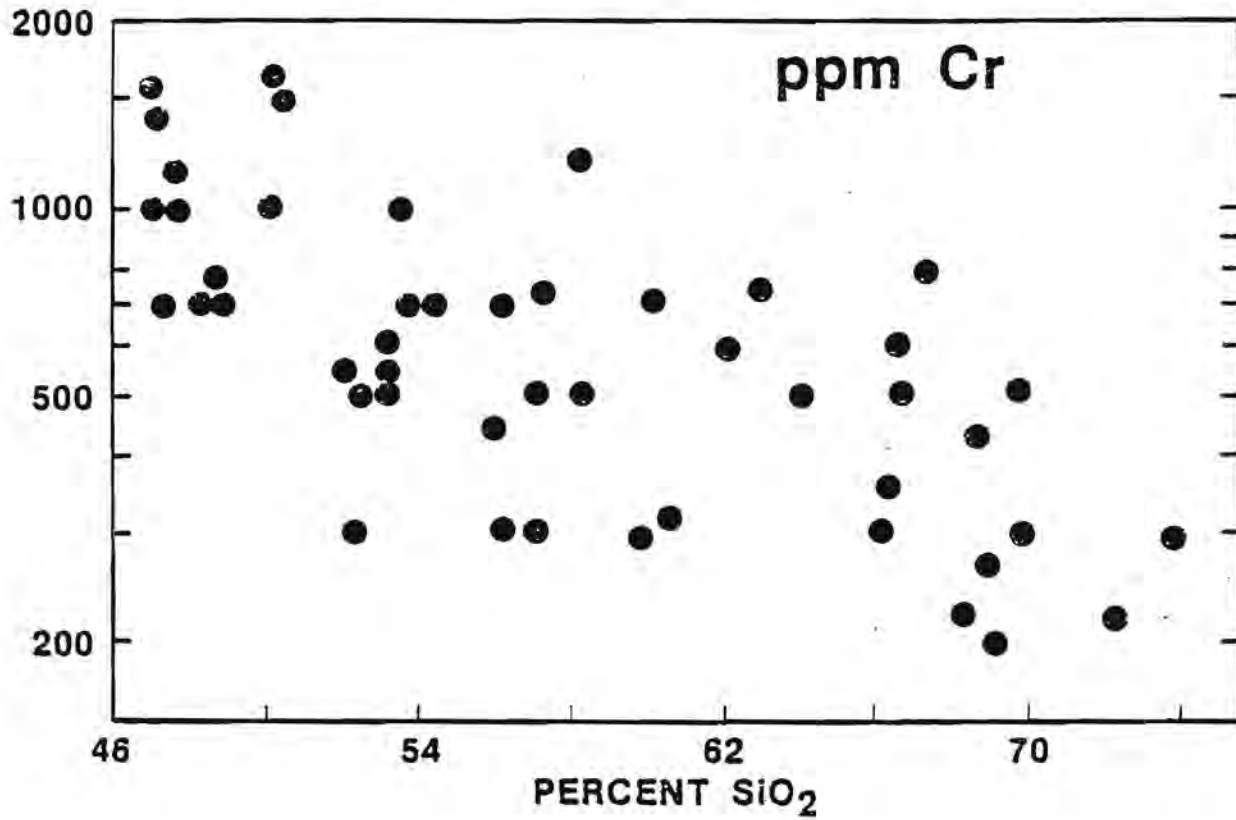
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METASEDIMENTARY ROCKS SOUTH OF STILLWATER COMPLEX



Geochemical data for the hornfels and schists intruded by the Stillwater complex and found as inclusions in the intrusive igneous rocks south of the Stillwater complex. Data compiled from several sources including Beltrame (in press), Page (1977), and unpublished analyses of Bowes. A combination of mafic and quartz-rich sources which were altered, weathered, and mixed during deposition is required for this series of rocks.

NASA-S-81-11494



Cr vs SiO₂, for the schists and hornfels from the Stillwater complex area. These rocks are distinguished by their high Cr and Ni contents which require a mafic component to play a prominent role in the formation of even the most SiO₂ rich rocks.

Mafic Dikes - Beartooth Mountains

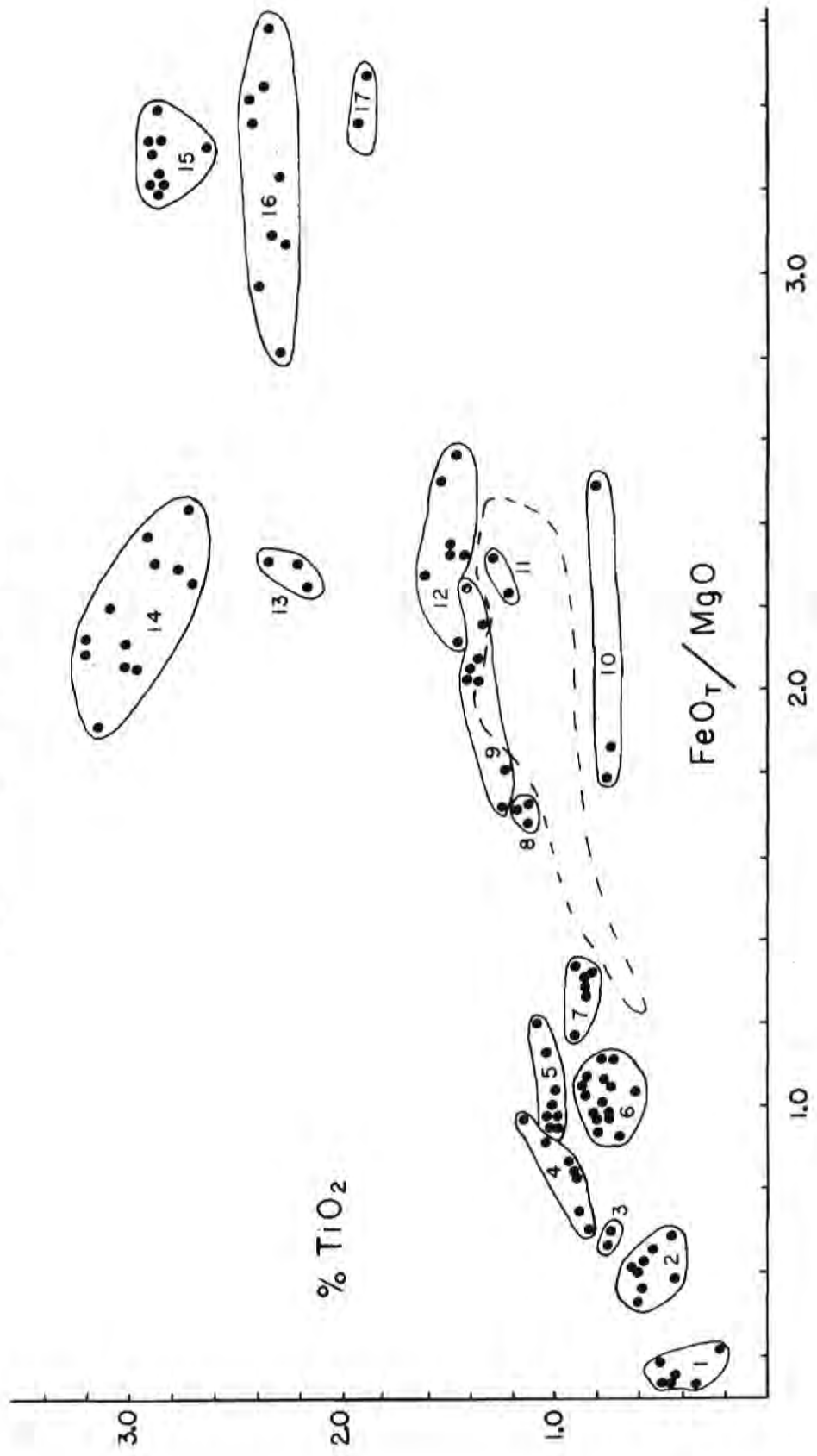
wt. %	1	2	3	4	5	6	7	8	9
SiO ₂	42.8	49.1	51.2	48.3	49.6	50.1	48.8	48.9	52.0
TiO ₂	0.45	0.62	0.75	0.97	1.01	0.81	0.86	1.13	1.35
Al ₂ O ₃	4.86	9.14	7.70	11.8	12.3	11.8	14.6	15.2	12.2
FeO _T	10.2	9.90	13.9	12.2	12.2	10.3	10.2	12.7	12.8
MnO	0.16	0.19	0.23	0.18	0.18	0.19	0.18	0.21	0.20
MgO	29.1	16.9	20.3	14.3	11.9	10.5	7.82	7.45	6.46
CaO	4.38	8.04	4.70	8.73	9.22	10.2	12.2	11.4	9.18
Na ₂ O	0.22	1.54	1.20	1.72	2.67	1.61	1.89	2.47	2.95
K ₂ O	0.31	0.56	0.42	0.85	0.56	0.77	0.39	0.31	0.84

ppm	1	2	3	4	5	6	7	8	9
Ba	83	191	145	215	326	117	104	97	173
Rb	12	20	19	37	16	50	15	13	42
Sr	50	124	186	122	293	121	127	110	165
Ni	1930	411	360	458	462	176	105	152	35
V	113	166	135	256	240	247	254	273	324
Zr	40	76	42	81	100	70	74	86	129

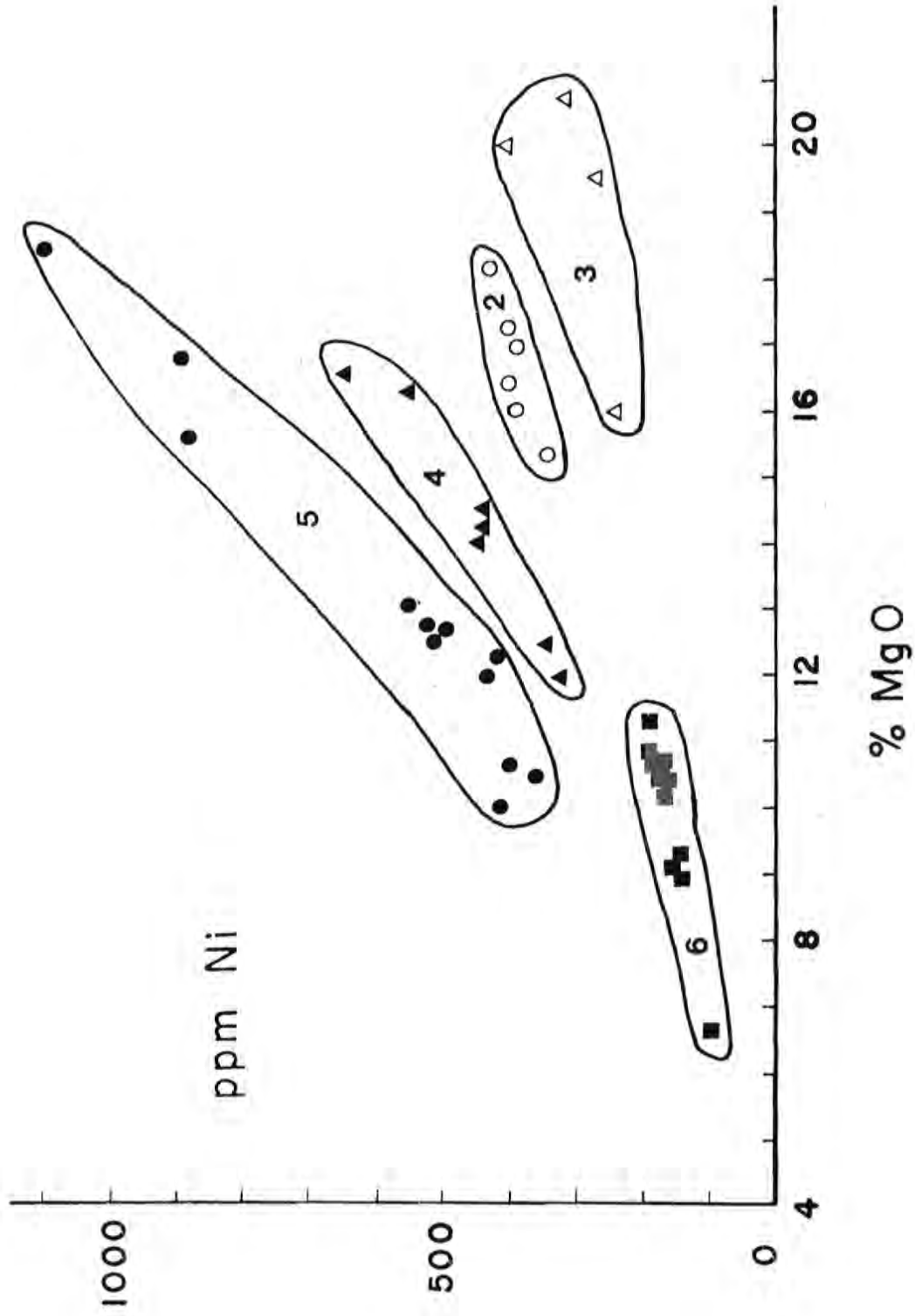
Mafic Dikes - Beartooth Mountains

wt. %	10	11	12	13	14	15	16	17
SiO ₂	56.2	51.3	50.5	49.4	45.2	52.2	51.6	52.5
TiO ₂	0.77	1.25	1.48	2.22	2.96	2.84	2.35	1.89
Al ₂ O ₃	14.6	14.9	13.5	12.1	13.5	12.5	12.6	12.9
FeO _T	9.58	12.2	12.7	15.5	14.4	15.0	14.5	14.9
MnO	0.16	0.19	0.21	0.21	0.21	0.22	0.21	0.19
MgO	4.75	5.40	5.43	6.83	6.70	4.62	4.49	4.38
CaO	8.32	9.79	9.01	7.82	8.90	8.47	8.26	7.66
Na ₂ O	2.37	2.63	2.72	3.77	2.40	2.69	2.61	4.50
K ₂ O	1.79	1.07	0.96	1.21	1.41	1.04	1.25	1.21
ppm								
Ba	395	173	174	310	542	217	375	358
Rb	65	35	36	52	56	41	47	45
Sr	165	137	150	167	309	178	153	409
Ni	59	68	29	41	63	40	47	77
V	183	257	357	389	334	567	387	267
Zr	124	128	136	203	211	188	175	145

BEARTOOTH DIKES



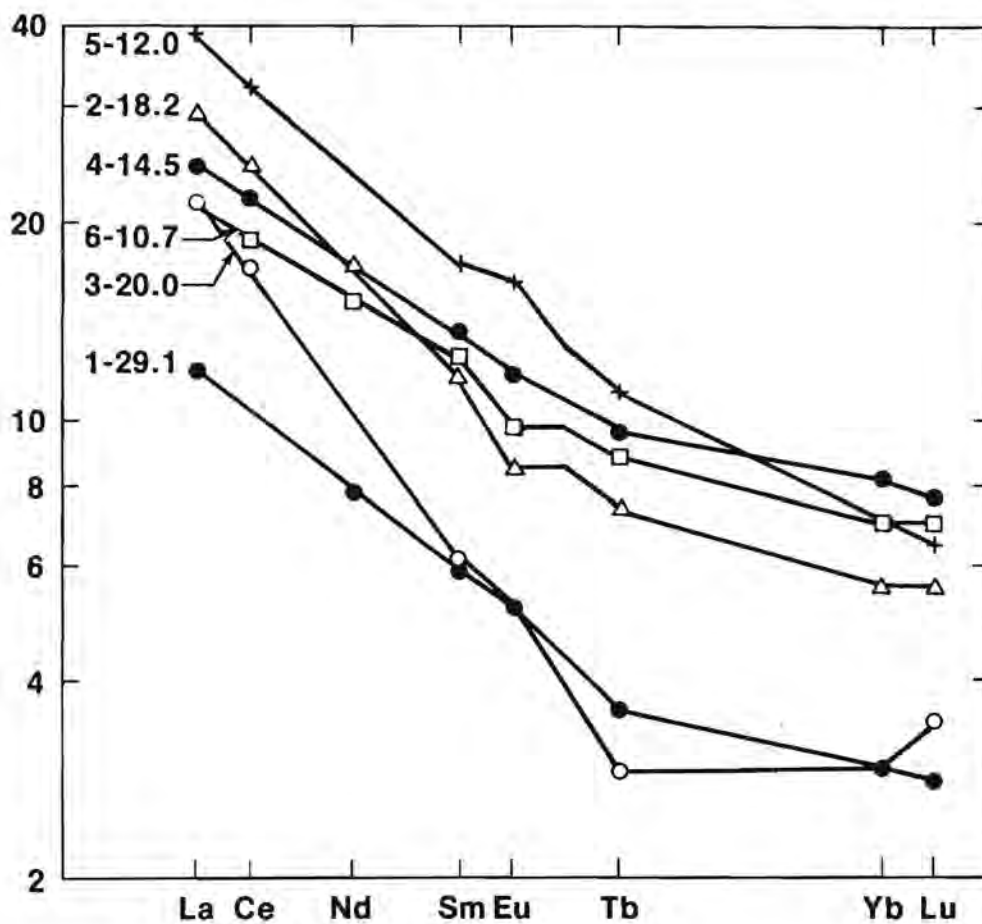
TiO₂ vs the ratio FeO_T/MgO for the mafic dikes of the Beartooth Mtns. At least 17 dike groups are recognized and range in age from 0.74 to 2.8AE. Numbers are keyed to the table of average group compositions. The dashed line encloses the field of a group of post 2.8AE dikes that were metamorphosed to moderate amphibolite grade.



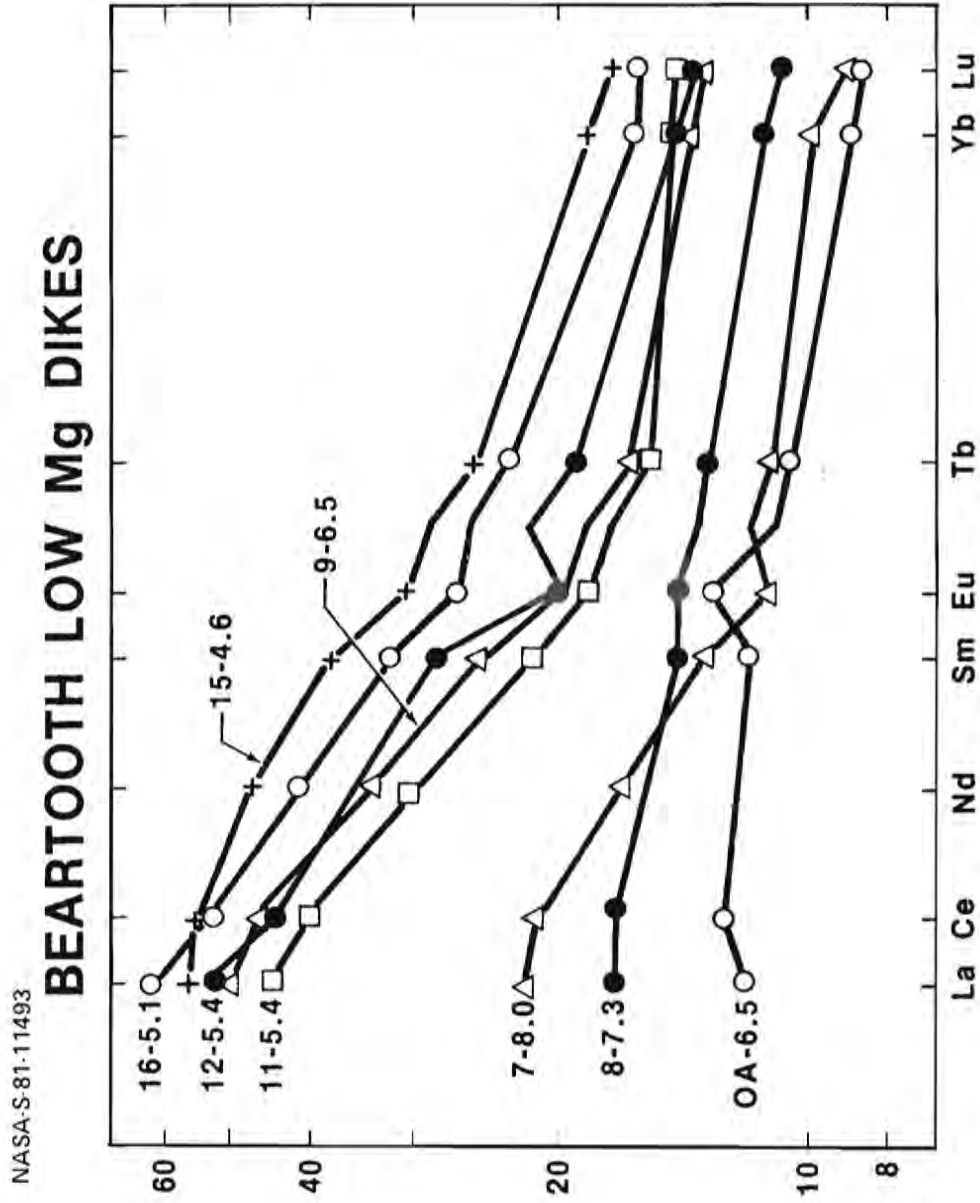
Ni vs MgO for the high Mg dikes of the Beartooth Mountains. Group numbers are keyed to the dike average composition table. The fact that these dikes are dominated by pyroxene and not olivine holds the Ni contents of some of these groups down. The data indicate that no simple fractionation relationship exists among these dikes.

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BEARTOOTH HIGH Mg DIKES



Chondrite normalized REE patterns for Beartooth dikes with MgO contents above 10%. Numbers given beside each pattern are keyed to the table containing the average major element analyses of the Beartooth dike groups. 5-12.0 means analysis no. 5 and this particular sample has 12.0% MgO. These high MgO dikes are noritic with only minor amounts of olivine.



Chondrite normalized REE patterns for the low Mg dikes of the Beartooths. Patterns are labeled as discussed for previous diagram. These samples can be generally classified as Fe-rich tholeiites.

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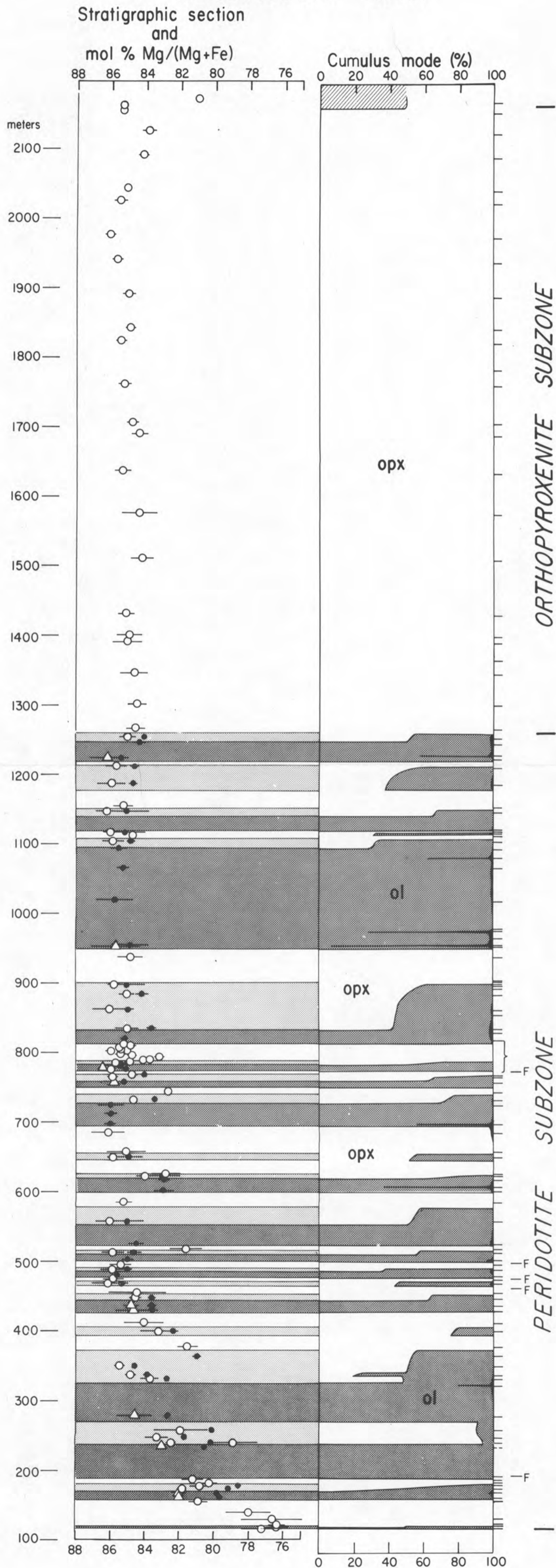
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- Montgomery, C. W. "Preliminary zircon U-Pb dating of biotite granodiorite from the South Snowy Block, Beartooth Mountains."
- Simmons, E. C. and Lambert, D. D. "Magma evolution in the Stillwater Complex, Montana: A preliminary evaluation using REE data for whole-rocks and cumulate feldspars."
- Timm, R. W. "Mineralogy and petrology of some metasedimentary xenoliths in the granitic gneisses of the Broadwater River area, Beartooth Mountains, Montana."
- Casella, C. J., Levay, J., Eble, E., Hirst, B., Huffman, K., Lahti, V., and Metzger, R. "Precambrian geology of the southwestern Beartooth Mountains, Yellowstone National Park, Montana and Wyoming."
- Mueller, P. A., Wooden, J. L., Odom, A. L., and Bowes D. R. "Geochemistry of the Archean rocks of the Quad Creek and Hellroaring Plateau areas of the eastern Beartooth Mountains."
- Wooden, J. L., Mueller, P. A., Hunt, D. K., and Bowes, D. R. "Geochemistry and Rb-Sr geochronology of Archean rocks from the interior of the southeastern Beartooth Mountains, Montana and Wyoming."
- Beltrame, R. J. "Systematic variations in hornfels at the base of the Stillwater Complex, Montana."
- Warner, J. L., Lee-Berman, R., and Simonds, C. H. "Field and petrologic relations of some Archean rocks near Long Lake, eastern Beartooth Mountains, Montana and Wyoming."

MOUNTAIN VIEW



ULTRAMAFIC ZONE

(Raedeke and McCallum, 1981)

Stratigraphic section

- dunite
- ▨ harzburgite
- orthopyroxenite

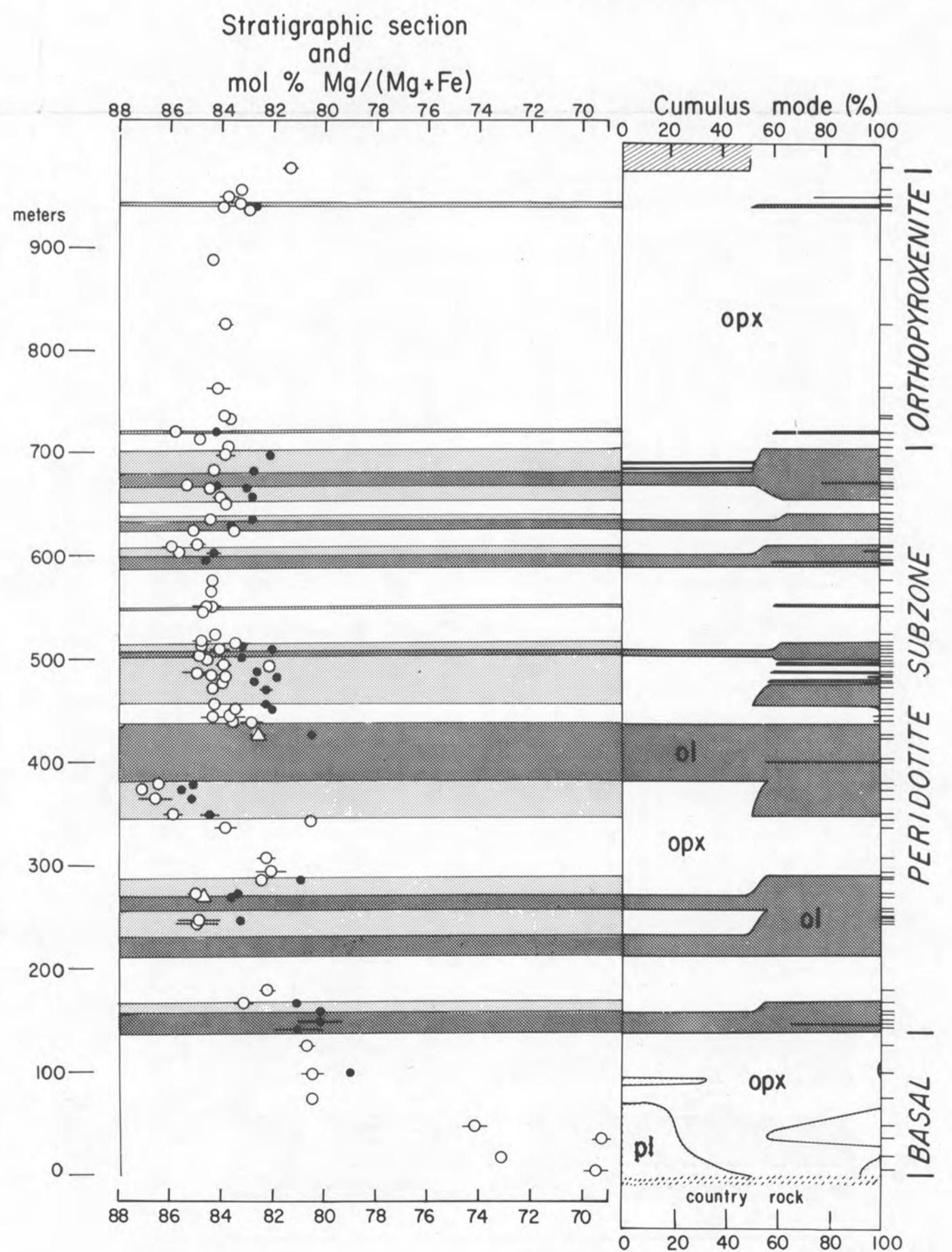
Mineral chemistry

- cumulus olivine (1σ standard deviation)
- cumulus orthopyroxene
- △ postcumulus orthopyroxene

Cumulus mode

- olivine
- orthopyroxene
- ▨ plagioclase
- ▤ clinopyroxene
- chromite

CHROME MOUNTAIN

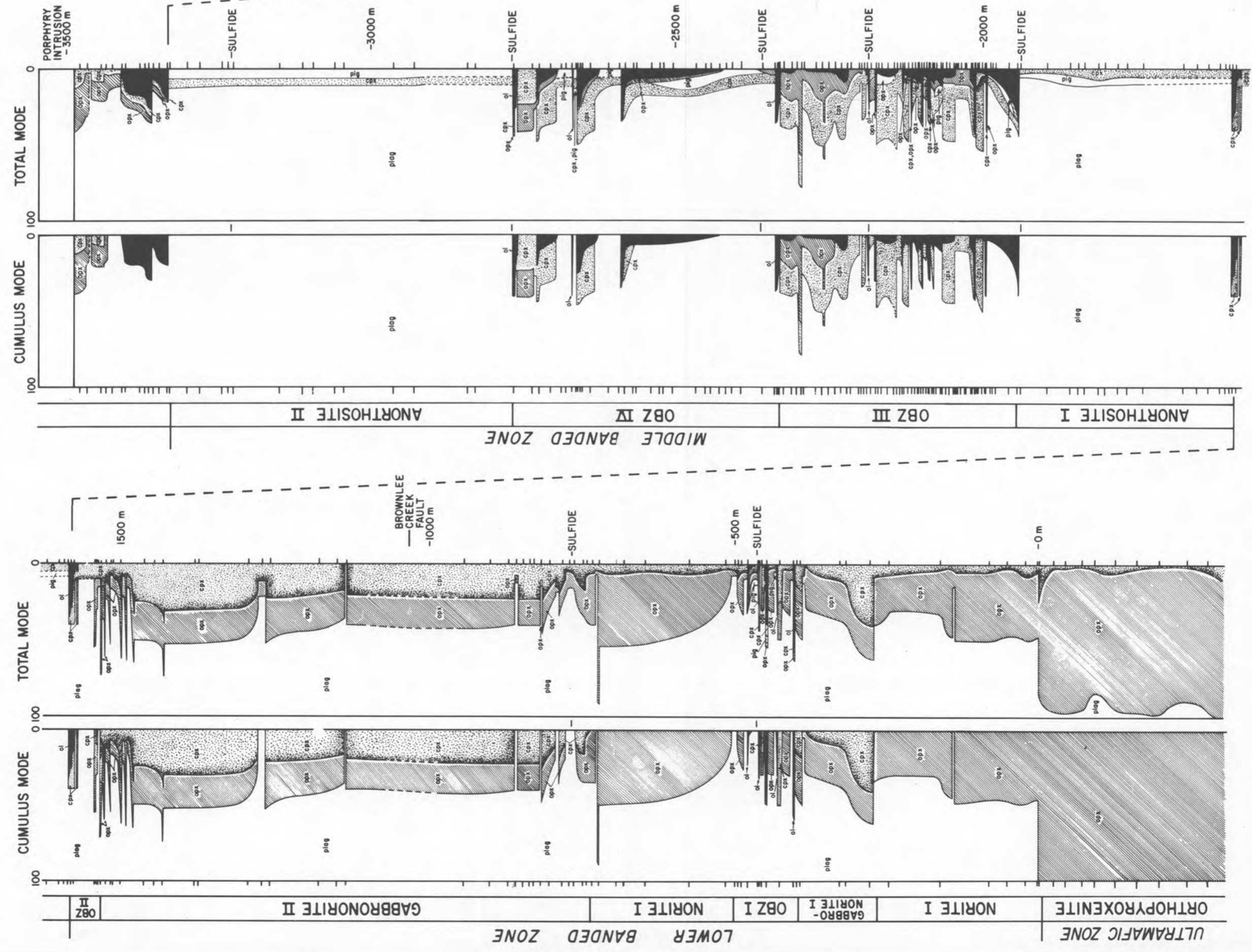


(McCallum, Raedeke and Markez, 1980)
BANDED ZONE, STILLWATER COMPLEX
IGNEOUS STRATIGRAPHY

-  PLAGIOCLASE
-  ORTHOPYROXENE
-  CLINOPYROXENE
-  PIGEONITE (INVERTED)
-  OLIVINE

Sample locations shown by "ticks" on outer margins

CONTACT MOUNTAIN SECTION



PICKET PIN SECTION

