WORKSHOP ON THE GROWTH OF CONTINENTAL CRUST

◖ LPI Technical Report Number 88-02

LUNAR AND PLANETARY INSTITUTE   3303 NASA ROAD 1   HOUSTON, TEXAS 77058-4399

WORKSHOP ON THE GROWTH OF CONTINENTAL CRUST

Edited by
Lewis D. Ashwal

Convened by
Lewis D. Ashwal
Stephen Moorbath
Paul N. Taylor

Sponsored by
The Lunar and Planetary Institute

Hosted by
The University of Oxford
Oxford, England
July 13-16, 1987

This report may be cited as:
Ashwal, L. D., ed. (1988) *Workshop on the Growth of Continental Crust.* LPI Tech. Rpt. 88-02. Lunar and Planetary Institute, Houston. 174 pp.

Papers in this report may be cited as:
Author A. A. (1988) Title of paper. In *Workshop on the Growth of Continental Crust* (L. D. Ashwal, ed.), pp. xx–yy. LPI Tech Rpt. 88-02. Lunar and Planetary Institute, Houston.

This report is distributed by:

ORDER DEPARTMENT
Lunar and Planetary Institute
3303 NASA Road 1
Houston, TX 77058-4399

Mail order requestors will be invoiced for the cost of postage and handling.

Contents

Cover

A selection of crustal growth models. Diagram is a modified version of one given in *Reymer and Schubert* (1984).

        Key:

F	=	Fyfe (1978) *Chem. Geol., 23*, 89 –114.
A	=	Armstrong (1981) *Philos. Trans. R. Soc. London, A301*, 443 –472.
B	=	Brown (1979) In *Origin of Granitic Batholiths, Geochemical Evidence* (Atherton, ed.), 106 –115, Shiva.
R & S	=	Reymer and Schubert (1984) *Tectonics, 3*, 63 –77.
D & W	=	Dewey and Windley (1981) *Philos. Trans. R. Soc. London, A301*, 189 –206.
O'N	=	O'Nions et al. (1979) *J. Geophys. Res., 84*, 6091 –6101.
Al	=	Allègre (1982) *Tectonophysics, 81*, 109 –132.
M & T	=	McLennan and Taylor (1982) *J. Geol., 90*, 347 –361.
N & D	=	Nelson and DePaolo (1985) *Geol. Soc. Am. Bull., 96*, 746 –754.
P & A	=	Patchett and Arndt (1986) *Earth Planet. Sci. Lett., 78*, 329 –338.
V & J	=	Veizer and Jansen (1969) *J. Geol., 87*, 341 –370.
H & R	=	Hurley and Rand (1969) *Science, 164*, 1229 –1242.

Preface

On the evening of June 28, 1985, Paul Taylor and I were sipping a glass of wine in the Qôrqut Hotel in West Greenland. Through the dining room window, across a fjord, we could see a sheer face of Qajûta Mountain, composed of 3.0 Ga Qôrqut granite with enclaves and inclusions of 3.8 Ga Amîtsoq gneisses. It was in this setting, among the world's most ancient continental crust, that we discussed the merits of a workshop on the general topic of crustal growth. A year later I went to Oxford and discussed this idea further with Stephen Moorbath and Paul Taylor. The Department of Earth Sciences at the University of Oxford seemed an appropriate location for such a workshop, considering that this is where Stephen Moorbath and colleagues carried out much of the pioneering work on the world's oldest rocks and developed crustal growth models that are still being tested and modified.

Largely through Stephen and Paul's efforts, and with the help of Pam Jones of the LPI, a workshop on the growth of continental crust was organized for July 1987. The objective of the meeting was to consider and discuss constraints and observations on a fundamental unsolved problem of global scale relating to the growth of planetary crusts. We scheduled the workshop to fall between the schedules for other meetings in Britain in order to facilitate participation of a diverse group of scientists. As with previous activities of the Early Crustal Genesis Project, we encouraged cross-fertilization of the planetary and terrestrial science communities by inviting both groups to attend, and this was quite successful. The workshop was attended by 49 scientists from seven countries. For four days we presented and discussed our results and ideas relating to crustal growth on all of the terrestrial planets, focusing on Earth's continental crust. The entire workshop was recorded, and an extensive summary of presentations and discussions prepared from these tapes appears in this volume, in addition to the submitted and invited abstracts. A special issue of *Tectonophysics* to appear in 1988 will contain full papers on the workshop topic by many of the participants. The workshop was highly successful, and I hope this report will serve as a source for those interested in pursuing further research along these lines.

Lewis D. Ashwal
Houston
December 1987

Program

Introduction to the Workshop

Historical Aspects of Modern Studies of Crustal Growth
S. Moorbath (Keynote)

Session I. Extraterrestrial Crustal Growth and Destruction
Chairman: S. R. Taylor

Growth of Planetary Crusts
S. R. Taylor (Keynote)

Growth of the Continental Crust: A Planetary-Mantle Perspective
P. H. Warren

Discussion

Early Intense Cratering: Effects on Growth of Earth's Crust
W. K. Hartmann

The Complexity, Depth, and Rapidity of Processes That Formed the Lunar Crust
G. Ryder and E. Dasch (Keynote)

Chronology of Early Lunar Crust
E. J. Dasch, L. E. Nyquist, and G. Ryder

Discussion

2:00-6:10 p.m.

Session II: Constraints from Observations and Measurements of Terrestrial Rocks
Chairman: K. Burke

A. Geology

An Illustration of the Complexity of Continent Formation
K. Burke (Keynote)

Growth of Early Archaean Crust in the Ancient Gneiss Complex of Swaziland and Adjacent Barberton Greenstone Belt, Southern Africa
A. Kröner, W. Compston, A. Tegtmeyer, C. Milisenda, and T. C. Liew

Discussion

Crustal Growth in the Archean: The Geochemical Evidence
S. R. Taylor

Structure of Southwestern Michipicoten Greenstone Belt, Ontario: Evidence for Archean Accretion?
G. E. McGill and C. H. Shrady

Discussion

Metamorphic Constraints on Mechanisms of Crustal Growth and Reworking
S. L. Harley (Keynote)

The Kerala Khondalite Belt of Southern India: An Ensialic Mobile Belt
T. Chacko, G. R. R. Kumar, J. K. Meen, and J. J. W. Rogers

Discussion

Mid to Late Proterozoic Magmatism Within Northeastern North America and Its Implications for the Growth of the Continental Crust
J. M. McLelland

Metamorphic P-T Paths and Precambrian Crustal Growth in East Antarctica
S. L. Harley

Discussion

Tuesday, July 14, 1987
9:00-11:20 a.m.

Session II (continued): Constraints from Observations and Measurements of Terrestrial Rocks
Chairman: N. T. Arndt

B. Geochemistry

Rate and Mechanism of Continental Growth—Petrological and Geochemical Observations
N. Arndt (Keynote)

Genesis of Archean Granitoids in the Pilbara Block and Implications for Crustal Growth
W. J. Collins

Discussion

Trace Element Differences Between Archaean, Proterozoic, and Phanerozoic Crustal Components—Implications for Crustal Growth Processes
J. Tarney, L. E. A. Wyborn, J. W. Sheraton, and D. Wyborn

Crustal Evolution and the Eclogite to Granulite Phase Transition in Xenoliths from the West African Craton
S. E. Haggerty, D. V. Hills, and P. B. Toft

Discussion

Lithospheric Evolution of the Northern Arabian Shield: Chemical and Isotopic Evidence from Basalts, Xenoliths, and Granites
M. Stein

11:40 a.m.-5:00 p.m.

C. Isotopes
Chairman: W. Compston

The Growth of the Continental Crust: Constraints from Radiogenic Isotope Geochemistry
P. N. Taylor (Keynote)

Role of Zircon in Tracing Crustal Growth and Recycling
W. Compston, I. S. Williams, R. A. Armstrong, J. C. Claouè-Long, P. D. Kinny, J. J. Foster, A. Kröner, R. T. Pidgeon, and J. S. Myers

Discussion

The Earliest Crustal History of Greenland as Recorded in Akilia Sediments
P. D. Kinny, W. Compston, and V. R. McGregor

Late-Archean Crustal Growth in the Lewisian Complex of Northwest Scotland—Diachroneity in Magmatic Accretion and Implications for Models of Crustal Growth
M. J. Whitehouse

Discussion

The Timing and Mechanism of Depletion in Lewisian Granulites
A. S. Cohen, R. K. O'Nions, and M. J. O'Hara

Construction of Mass-Age Curves for the Continental Crust: An Empirical Model and an Example from the Western U.S.
D. J. DePaolo, G. S. Schubert, and A. Linn (Keynote)

Discussion

Changing Styles of Crustal Growth in Southern Africa: Constraints from Geochemical and Sr-Nd Isotope Studies in Archaean to Pan African Terrains
F. McDermott, C. J. Hawkesworth, and N. B. W. Harris

Geochronological and Isotopic Evidence for Rapid Crust Formation in Middle-Proterozoic Time: The Labrador Example
U. Schärer

Discussion

5:10-5:50 p.m.

Session III: Models of Crustal Growth and Destruction

Thermal Models Pertaining to Continental Growth
  P. Morgan and L. Ashwal (Keynote)

Discussion

Wednesday, July 15, 1987
9:00 a.m.-12:45 p.m.

Session III (continued): Models of Crustal Growth and Destruction
Chairman: G. Schubert

Physical Processes in the Growth of the Continental Crust
  G. Schubert (Keynote)

Constraints on Continental Accretion from Sedimentation
  D. Abbott

Discussion

Volcanic Contribution to Crustal Growth in the Central Andes: A New Estimate and a Discussion of Uncertainties
  C. A. Wood and P. Francis

Discussion

Sm/Nd Constraints on the Growth of Continental Crust
  F. Albarede

Decoupled Rb-Sr and Sm-Nd Isotopic Evolution of the Continental Crust
  S. L. Goldstein

Does Subduction Zone Magmatism Produce Average Continental Crust?
  R. Ellam, C. J. Hawkesworth

Discussion

2:00-5:40 p.m.

Session IV: Processes of Crustal Growth and Destruction
Chairman: J. Veizer

Solid Earth as a Recycling System
  J. Veizer (Keynote)

Pb Isotope Constraints on the Extent of Crustal Recycling into a Steady State Mantle
  S. J. G. Galer, S. L. Goldstein, and R. K. O'Nions

Discussion

Continental Crustal Formation and Recycling: Evidence from Oceanic Basalts
  A. D. Saunders, J. Tarney, and M. J. Norry

Recycling of Lower Continental Crust Through Foundering of Cumulates from Contaminated Mafic Intrusions
  N. T. Arndt and S. L. Goldstein

Discussion

Evidence for Crustal Recycling During the Archean: The Parental Magmas of the Stillwater Complex
  I. S. McCallum

Mineralization Through Geologic Time: Evolution of Continental Crust
  J. Veizer, P. Laznicka, and S. L. Jansen

Discussion

Pluton Emplacement and Magmatic Arc Construction: A Model from the Patagonian Batholith
  R. Bruce, E. Nelson, and S. Weaver

Discussion

Thursday, July 16, 1987
9:00 a.m.-12:40 p.m.

Session IV (continued): Processes of Crustal Growth and Destruction
Chairman: A. Kröner

A Tectonic Approach to the Growth and Evolution of the Continental Crust
  J. Dewey (Keynote)

Ridge-Trench Collision in Archean and Post-Archean Crustal Growth: Evidence from Southern Chile
  E. P. Nelson and R. D. Forsythe

Discussion

Gravity Anomalies, Plate Tectonics, and the Lateral Growth of Precambrian North America
  M. D. Thomas, R. A. F. Grieve, and V. L. Sharpton

Discussion

Continental Growth Through Time by Underplating of Subducted Oceanic Crust: Evidence from Kimberlites in South Africa and the SW Pacific
  L. A. Taylor and C. R. Neal

The Petrogenesis of "Oceanic Kimberlites" and Included Mantle Megacrysts: The Malaitan Alnoite
  C. R. Neal

Discussion

Growth of the Lower Continental Crust
  R. L. Rudnick

Discussion

2:00-5:20 p.m.

Session IV (continued): Processes of Crustal Growth and Destruction
Chairman: J. Tarney

Crustal Underplating in the North American Craton: Evidence from the Petrogenesis of Anorthositic Rocks in the Middle Proterozoic Midcontinent Rift System
  P. W. Weiblen and J. D. Miller, Jr.

The Growth Continents and Some Consequences Since 1.5 Ga
  D. G. Howell

Discussion

Summary of the Workshop

Print Only Abstracts

Diapirs of the Mediterranean Ridge: The Tectonic Regime of an Incipient Accreted Terrane
  Y. Mart

The Tectonic Setting of the Seychelles, Mascarene Amirante Plateaus in the Western Equatorial Indian Ocean
  Y. Mart

The Tectonic Setting of Archean Anorthosites
  D. A. Morrison, W. C. Phinney, and D. E. Maczuga

Archean Crustal Evolution of the Northern North China Craton
  X. Qian, Y. Chen, and J. Liu

Evolution of the Dharwar Craton—A Terrain of Early Archaean Crustal Stability, Long Term Orogenic Cycles, and Large Scale Palaeobiological Activity
  R. Srinivasan and S. M. Naqvi

Summary of Technical Sessions

L. D. Ashwal

This summary of presentations and discussions is based on recordings made at the workshop. Discussion summaries are printed in italics. In most cases I have identified those participants who asked questions or made comments, but this was not always possible. I apologize to anyone I have misidentified, misquoted, or misinterpreted.

INTRODUCTION TO THE WORKSHOP

Stephen Moorbath summarized historical aspects of crustal growth studies. In the decades before plate tectonic theory, it was widely believed that continents grew from small cratons by consolidation of marginal geosynclines. Even as late as the 1950s, continents were considered to be fixed and the ocean basins to be floored by sunken continental crust. Ancient gneiss complexes were thought to be remnants of the Earth's earliest contintental crust, perhaps dating from the beginning of geologic time. The concept of early, complete production of continental crust with subsequent repeated episodes of intracrustal recycling and reworking cannot, however, be supported by modern radiogenic isotope studies. This view is different from modern steady-state models of continental growth (such as that of R. L. Armstrong), which take advantage of plate tectonic theory, and involve recycling of continental material back into the mantle. Detailed mapping in the ancient gneiss complexes has been crucial to our developing understanding of how continents grow. In the early 1950s Sutton and Watson used mafic dikes to separate tectonothermal events in the Lewisian of Scotland. This approach was later used in West Greenland by McGregor to establish the key geological relationships that led to the recognition of the world's oldest rocks.

The idea that ancient continents were thin because of higher heat production during the Archean has largely been negated by recent petrological studies that show that even the oldest supracrustal rocks record evidence for metamorphism at high pressures and temperatures. This gives credence to uniformitarian models of crustal growth, and to the concept that plate tectonics may have been operative very early on planet Earth. Pioneering work in the 1960s by Hurley and colleagues on Sr isotopes led to the idea that there were periodic episodes of relatively short duration (~200 Ma) when major additions to the continental crust took place. The existence of so-called "Crustal Accretion-Differentiation Superevents" (CADS) within continents has been confirmed by later Pb and Nd isotopic work, but the idea that they were synchronous worldwide is clearly too simplistic. It seems doubtful that there were long periods of relative quiescence during which little or no continental material was produced. Moorbath ended by expressing his preference for models involving no significant continental crust prior to 3.8 Ga, followed by slow, but irreversible growth of crust subsequently.

SESSION ONE: EXTRATERRESTRIAL CRUSTAL GROWTH AND DESTRUCTION

This session was opened with a keynote talk by Ross Taylor, who pointed out that the crusts of the 8 planets and ~60 satellites in our solar system all differ in composition from each other and from that of the primordial solar nebula. Taylor divided planetary crusts into three types, based on differences in mode and rate of formation. Primary crusts are those formed during or just after accretion by substantial melting of planetary bodies. Examples include the lunar anorthositic crust (which formed in about 100 Ma by plagioclase flotation from an early magma ocean) and those of some of the heavily cratered icy satellites such as Callisto and Europa (composed of water ice perhaps dating to the time of original accretion). Secondary crusts are those that formed later during partial melting of planetary interiors. Examples include the basaltic crusts of the Earth, Moon, Mars, and possibly Venus, as well as some clearly younger water ice crusts on outer planet satellites such as Ganymede. Tertiary crusts, the only known example of which is the Earth's continental crust, form from more extreme processes of differentiation and may require the presence of liquid water at the surface. Taylor sees no evidence for crustal recycling anywhere in the solar system.

Paul Warren discussed the growth of early crusts from a planetary mantle perspective. He believes that the Earth had a substantial volume of continental crust prior to 3.8 Ga, but this crust did not survive because of vigorous convection and impact degradation. Although the composition of this crust is likely to have been tonalitic rather than anorthositic, these two types of crust are not substantially different when viewed from the perspective of primordial mantle compositions. The Earth, therefore, may well have had an early magma ocean, although the greater pressures at shallower depths in a terrestrial compared to a lunar magma ocean would have resulted in extensive garnet and Al-rich pyroxene crystallization, and this would have prevented the formation of an aluminous crust thicker than 45 km (a conservative estimate). Warren concluded that the relative volume of continental-type crust is more a function of planetary size than time.

In discussion, Warren reiterated the role of garnet fractionation in preventing a thick anorthositic crust from forming from a terrestrial magma ocean. He also suggested that the internal pressure of Mars, being intermediate between Earth and Moon, would inhibit the formation of such a crust there. The possibility was raised that neither Mars nor the eucrite parent body was ever molten. Warren was asked why his model could not allow continental crust to start forming at 3.8 Ga instead of 4.5 Ga, and he responded that it would be difficult to avoid making crust this early. Those who favor growth models starting at 3.8 Ga must specify what was present before this time. This raised the semantic question of what is meant by "continental crust," primordial or otherwise. Some, such as Warren, consider basaltic crust as "continental crust," broadly defined; from the

perspective of the mantle, the difference between "granitic" and "basaltic" crust is small. The issue was raised of possible early fractionation between the Earth's upper and lower mantle, and how this would affect early crust formation. Warren assumed no such mantle fractionation but stated that in either case this would not substantially affect his model. There was some discussion that Warren may have been inappropriately comparing different types of crust as defined by Taylor in the previous talk, although some support was offered for the significance of a negative correlation between planetary radius and fraction of the surface covered by "continental" crust.

The session continued with W. K. Hartmann's talk in which he discussed the disrupting effects of early intense meteorite bombardment on Earth's proto-crustal evolution. He emphasized that we should not consider the Earth's impact history as a discrete phase separate from an early crust-forming event, and also that the end of the impacts was not a singular event that destroyed all previously formed crust.

In the next talk, G. Ryder described the character of the lunar crust in detail, emphasizing its complexity. The lunar highland crust is almost entirely made up of plutonic fragments, and is composed of (1) ferroan anorthosites, (2) an Mg-rich suite of norites, troctolites, and troctolites, and (3) KREEP. Regardless of whether the lunar crust was formed by serial magmatism or from a magma ocean, a lot of complex things happened very rapidly even though the Moon is a rather small body.

E. J. Dasch underscored this point in his talk, which summarized the chronology of lunar rocks. The oldest "pristine" (i.e., lacking meteoritic contamination or admixed components) lunar rock, recently dated with Sm-Nd by Lugmair, is a ferroan anorthosite, with an age of 4.44 + 0.02 Ga. Ages of Mg-suite rocks (4.1–4.5 Ga) have large uncertainties, so that age differences between lunar plutonic rock suites cannot yet be resolved. Most mare basalts crystallized between 3.1 and 3.9 Ga. The vast bulk of the lunar crust, therefore, formed before the oldest preserved terrestrial rocks. If the Moon accreted at 4.56 Ga, then 120 Ma may have elapsed before lunar crust was formed.

In discussion it was pointed out that some lunar Rb-Sr ages are older than Sm-Nd. Dasch suggested that this might be an artifact caused by Rb leaching during sample preparation. Hartmann was asked about the possibility that terrestrial crust extant before 3.8 Ga was recycled back into the mantle. Hartmann suggested that for both Earth and Moon, the probability of impact-induced crustal destruction continuously decreased between 4.6 and ~4.0 Ga. An additional factor that must not be ignored is the vigorous convective recycling on Earth during that period. Discussion then focused on the "subductibility" of various types of terrestrial crust, but no concensus was reached. Dasch was asked by Hartmann if there was any clustering of lunar ages before 4.0 Ga that might indicate isotopic resetting during a major basin-forming event. This brought up the issue of the degree to which rock ages can be reset by impacts. There is some evidence that isotopic clocks are little disturbed by impacts as large as that of the Imbrium

basin (diameter = ~ 1,000 km), although much material is ejected. Hartmann was asked why samples from the lunar mantle were not excavated by giant impacts. This question bears on the timing of the fractionation event(s) that formed the lunar crust. Hartmann responded that the major fractionation event took place before the time when surface rocks had a 50% chance of surviving impact punishment. Ross Taylor commented that a major iron-removing event must take place before a crust began to form from the lunar magma ocean. Ryder asked the general question of why the lunar highlands crust age of 3.9 Ga is so close to the age of the oldest terrestrial rocks, if this coincidence is not a result of cratering effects. No convincing answer was offered.

SESSION TWO: CONSTRAINTS FROM OBSERVATIONS AND MEASUREMENTS OF TERRESTRIAL ROCKS

The first part of this session (Geology) was opened by Kevin Burke, who pointed out that a consensus may be emerging in crustal growth models, considering the clustering of most growth curves (see cover figure) and their uncertainties. Curves most distant from this clustering represent models involving extensive recycling of continental material back into the mantle, but Burke wondered if geochemical signatures for this would be recognizable considering the lack of evidence from seismic tomography for discrete mantle reservoirs, and the likelihood of core-mantle interaction based on recent high pressure experiments. Unreactivated Archean rocks represent only 2% of present continental area, and Burke was uncomfortable about basing inferences on what the early Earth was like on such a small amount of information. Burke feels that the hypothesis of continental assembly that needs testing is that of banging together of island arcs, such as in Indonesia today. As an example of how complex this process can be, Burke described the geology of the Caribbean arc system, which shows evidence for reversals of subduction polarity, numerous collisional events, and substantial strike-slip movements. It seemed unlikely to Burke that Archean examples would have been less complicated, and, therefore, we should expect only to be able to recognize broad general environments such as island arc, ocean floor, Andean, or Tibetan environments.

Alfred Kröner then discussed the relationship between early Archean greenstones and high grade gneisses in the Ancient Gneiss Complex (AGC) of Swaziland and the neighboring Barberton greenstone belt in Southern Africa. New high precision zircon analyses reveal, among other things, a complex history in individual zircons from tonalitic orthogneisses, with ages as old as 3644 + 4 Ma. This suggests the presence of continental crust prior to the formation of the supracrustal rocks of the Barberton greenstone belt, which have been previously considered to be the earliest rocks in the area. Kröner suggested that these data are incompatible with the intraoceanic settings that have been widely accepted for this terrane, and favors either a marginal basin or rift environment. By using the detailed age information obtained from zircons in combination with $^{40}Ar/^{39}Ar$ and paleomagnetic measurements, Kröner and colleagues have estimated that plate velocities for this part of the African craton were about 10–70 mm/yr, during the period 3.4–2.5 Ga. This

is not incompatible with the idea that Archean plate velocities may have been similar to those of today.

In discussion, Kröner was asked to specify the time interval considered in determining the Archean plate velocities, as rate estimates will be inversely proportional to the time resolution. Kröner responded that the time resolution was good, on the order of 50 –100 Ma, especially for the older events. Kröner was then asked why he considered the AGC to represent the basement for the greenstone belts if the two were tectonically juxtaposed. Kröner answered that there is one locality at which the two are in direct, nontectonic juxtaposition, and that there was evidence for detritus in greenstone belt sediments having been derived from the AGC. It was pointed out that this detritus is not as old as some of the rocks in the AGC. Kröner responded that at the very least, the new age data rule out the hypothesis of deriving the 3.55 –3.65 Ga granitoids of the AGC from the ~3.45 Ga greenstone belts. Regarding a point Burke made in his talk, it was pointed out that if, following accretion, water was added to the Earth from elsewhere in the solar system, then the Moon should also be expected to have received some. There was some discussion about whether the Moon would retain water or other volatiles received in this way. No consensus was reached.

The session continued as Ross Taylor discussed his well-known crustal growth model, which is based, among other things, on trace element and isotope systematics of naturally produced widespread crustal samples such as loess and mature sedimentary rocks. Taylor and colleagues still maintain the basic premise of their model, which holds that a major period of crustal growth and intracrustal melting took place in the late Archean, and that there has been little change in upper crustal composition since then. However, even though the Archean upper crust was dominated by a bimodal suite of basaltic and tonalitic-trondjhemitic igneous rocks, new REE data on Archean sedimentary rocks from Mt. Narryer (Australia) and Limpopo (South Africa) show negative Eu anomalies, which were previously thought to be restricted to post-Archean samples. These, Taylor believes, were derived from localized minicratonic areas rich in potassic granites. He pointed out that only high-grade Archean terranes contain samples with typical post-Archean REE patterns. Taylor recommended a minor modification of his crustal growth model, adding small step-like increments to the otherwise smooth Archean growth curve to account for the existence of local K-rich crust as old as 3.3 – 3.4 Ga.

G. E. McGill then discussed the results of his detailed mapping in the well exposed (by a large fume kill) Archean Michipicoten greenstone belt of Ontario. He described numerous structural features, including soft-sediment deformation, thrust faults, isoclinal folds, and sill intrusion, which apparently formed prior to the earliest flattening cleavage. He inferred that all of these events may have taken place before the rocks were completely lithified, and raised the question as to whether early soft-sediment and tectonic deformation were coeval. If so, the overall picture would permit (but not compel) recognition of a convergent

oceanic environment such as an accretionary wedge or forearc basin.

Most of the ensuing discussion focused on the shapes of crustal growth curves. S. L. Goldstein expressed his dismay about some of the previous speakers' assumptions that most of the Earth's crust was made during the Archean. Some of the early studies leading to this conclusion were carried out before the acceptance of continental drift, and therefore should be considered with caution. Goldstein also objected to Burke's implication that the clustering of growth curves represented a consensus of opinion, and suggested that Veizer and Jansen's curve could indeed be correct. He then pointed out that S. R. Taylor's conclusion that 70% of the crust was extant by the end of the Archean is based on locally derived sediments of greenstone belts, which may not be representative of Archean crust in general. Neodymium model ages are not consistent with this conclusion. Taylor responded that he feels the Archean sedimentary record, especially that in the terranes of low metamorphic grade, samples a sufficiently wide provenance to justify his conclusions. Chairman Burke moved the discussion to other topics, and recognized S. E. Haggerty, who commented that there is substantial evidence from diamond inclusions for the growth of depleted lithosphere down to 200 km depth possibly by as early as 3.5 Ga. This would seem to imply that at least 80% of the cratonic nucleii were in place by about 3.3 Ga. Burke commented that our inferences about the Archean may be skewed toward the unrepresentative if Paul Morgan's inference about selective preservation of low heat production Archean crust is correct. J. Veizer emphasized that the Archean record is not unrepresentative because Archean cratons are inherently stable and will tend to be preserved. Veizer also commented that his growth curve, which had been discussed extensively, actually represents one extreme in a family of curves.

Following the afternoon tea break, Simon Harley summarized the metamorphic constraints on crustal thicknesses in Archean and post-Archean terranes, and possible implications for tectonic processes. It is important to recognize that P-T estimates represent *perturbed* conditions and should not be used to estimate steady-state geothermal gradients or crustal thicknesses. He cited the example of the Dora Maira complex in the French Alps, where crustal rocks record conditions of 35 kbar and 800°C, implying their subduction to depths of 100 km or more, followed by subsequent uplift to the surface. Therefore such P-T estimates tell us more about processes than crustal thicknesses. Of more significance, according to Harley, are determinations of P-T *paths*, particularly coupled with age measurements, because these may provide constraints on how and when perturbed conditions relax back to steady-state conditions. Harley then illustrated P-T paths that should be expected from specific tectonic processes, including Tibetan-style collision, with and without subsequent extension, rifting of thin or thickened crust, and magmatic accretion. Growth of new crust, associated with magmatic accretion, for example, could possibly be monitored with these P-T paths.

Tom Chacko then described the Proterozoic (?) Kerala Khondalite belt of the Southern Indian Shield, a belt dominated

by granulite grade (750°C, 5–6 kbar) supracrustal rocks whose protoliths included arkoses and shales with cratonic provenances. REE and other geochemical signatures suggest a "granitic" source for these metasediments, possibly the spatially associated charnockite massifs. The presence of intercalated mafic gneisses, interpreted as basalts, implies a cratonic rift basin rather than a foreland basin setting. Chacko argued that the Kerala, as well as other early Proterozoic mobile belts formed during abortive continental rifting without major additions of new crust.

In the following discussion, the derivation and usage of so-called steady-state reference geotherms in thermal modeling were questioned. Harley reiterated the uncertainties associated with geotherm determination and emphasized that, in most cases, they are used for reference only. Burke reminded everyone of the difference between conductive and convective geotherms. Schubert asked about the role of fluid transport in P-T paths. Harley responded that fluids could potentially change the distribution of temperatures in metamorphic terranes and the rates at which heat is dissipated by convective heat transfer. Regarding the suggested collaboration of isotope geochemists and metamorphic petrologists on deducing P-T trajectories of rock units, Harley was asked about the relative closure temperatures of mineral such as garnet for Fe-Mg exchange versus isotopic systems such as Rb-Sr or Sm-Nd. He responded that there is a great need for more determinations of closure temperatures of minerals for a variety of isotopic systems. Paul Morgan asked Harley to justify the 20 Ma isobaric heating period that is commonly used in compressional thermal models. Harley responded that there was some evidence from Oxburgh's work in the Alps for a 20 Ma lag time between the estimated onset of collision and the first indications of uplift in the sedimentary record, although this has recently come into question. Harley admitted that evidence for this 20 Ma initial period was tenuous at best.

Jim McLelland then discussed late Proterozoic magmatic rocks from the eastern Canadian Shield, including anorthosite, charnockite, mangerite, and alaskite. He believes these were emplaced under anorogenic or mildly extensional conditions, and drew an analogy among these and the Younger Granites of Niger-Nigeria, and the White Mountain magma series of New Hampshire. Using areal estimates for these new additions to the crust, McLelland computes growth rates of 0.025% to 0.04% per Ma, assuming an emplacement interval of 300–500 Ma. These rates are essentially equivalent to "middle of the road" crustal growth curves.

Simon Harley ended the first day sessions with a discussion of Precambrian metamorphism in East Antarctica. Granulites of the Archean Napier Complex show near-isobaric cooling, and together with age data, suggest a model involving extension following collisional thickening. Within the Napier Complex, Proterozoic shear zones associated with granulites of the Rayner Complex show a near-isothermal decompression history, presumably associated with a collisional event in which the Archean was underplated by new Proterozoic material.

In discussion, S. Klemperer pointed out that the anorthosites and related rocks that McLelland discussed are thought to be only 3–4 km thick, and therefore his growth rates may have been greatly overestimated. McLelland responded that he assumed a similar ratio of new magmatic material to basement throughout the crust. Klemperer reemphasized that some estimates of volumes must be made, and consideration of just areal distributions of rock types could be misleading in determining crustal growth rates. Some questions were then raised about Harley's implication that Proterozoic shear zones were indicative of uplift of the Archean Napier Complex by underplating. The session was quickly adjourned by Chairman Burke, who encouraged participants to carry out further discussions over wine in Professor Dewey's office.

On the second day, the next part of Session Two (Geochemistry) was opened with a talk by Nick Arndt on isotope and trace element geochemistry of Precambrian mantle derived rocks and implications for the formation of the continental crust. Epsilon Nd values of Archean komatiites are variable, but range up to at least +5, suggesting that the Archean mantle was heterogeneous and, in part, very depleted as far back as 3.4–3.5 Ga. This may be taken as evidence for separation of continental crust very early in Earth history. If these komatiite sources were allowed to evolve in a closed system, they would produce modern-day reservoirs with much higher epsilon Nd values than is observed. This implies recycling of some sort of enriched material, perhaps subducted sediments, although other possibilities exist. Archean volcanics show lower Nb/Th than modern volcanics, suggesting a more primitive mantle source than that observed nowadays. However, Cretaceous komatiites from Gorgona island have similar Nb/Th to Archean volcanics, indicating either (1) the Archean mantle source was indeed more primitive, or (2) Archean magmas were derived from a deep ocean-island source like that proposed for Gorgona. If the latter is the case, then melts derived from N-MORB type sources are not represented in Archean terranes, and the Archean mantle may have been even more depleted than current estimates, requiring earlier or faster continental growth than current estimates.

In discussion, John Tarney asked if some process other than separation of continental crust could account for the positive epsilon Nd values of Archean volcanic rocks. Arndt responded that all the data require is separation of some sort of enriched material, although he strongly believes this to be continental crust in light of the discovery of 4.2 Ga zircons. Tarney then asked about the possibility of contaminating the N-MORB source with ocean-island basalts. Arndt admitted that this is a possibility that must be tested by quantitative modeling.

Bill Collins then discussed the origin of the Mount Edgar Batholith of the Pilbara Block in Western Australia, and other Archean granitoids that show relatively primitive isotopic characteristics but more evolved bulk geochemistry. He presented geochemical evidence in support of a multistage process of crustal ripening, involving partial melting of a thick basaltic crust to produce the tonalitic/dacitic sources of the more

evolved granitoids. These events must produce stable continental crust in less than 200 Ma and are different geologically and geochemically from those which produced Phanerozoic granitoids in Andean margin settings.

Roberta Rudnick started the discussion by asking Collins about the fate of mafic residual material produced by extracting granitoids from a mafic protocrust. Collins explained that in his model, partial melting of a 60 km thick basaltic protocrust produces 30% tonalitic melt and an eclogitic residue that founders and sinks back into the mantle, which induces crustal recycling and further partial melting. Dallas Abbott asked why the tonalites could not be derived simply by small degrees of partial melting of basalt. Collins referred to experimental studies that indicate that such melts would be expected to have 60 –63 wt.% SiO_2, far lower than the 70 wt.% observed in the Pilbara. Stephen Moorbath asked Arndt how the epsilon Nd values he discussed were determined and about the possibility of open system behavior. Arndt reiterated that the epsilon values were calculated for individual samples, assuming an age obtained from U/Pb zircon analyses. He admitted that false values may be a problem for terranes affected by metasomatic/metamorphic events, particularly if they took place much later than primary crystallization. We were assured, however, that Arndt's data are from areas unaffected by such events. John Dewey asked Collins to elaborate on the geologic and tectonic environment for the model he proposed. Collins favors rifting of and production of tonalitic melts from a thick basaltic protocrust in response to underplating. This is then followed by foundering of dense eclogitic material, which causes additional partial melting of the tonalitic material to produce the more evolved granitoids. He does not believe that this is a global phenomenon, and sees no evidence for Andean-style settings in the Pilbara. Dewey was unconvinced that there were adequate heat sources for such a scenario.

The session continued with John Tarney's talk about secular differences in trace element geochemistry of granitoids. Archean tonalites and trondhjemites are characteristically Sr-undepleted and Y-depleted, suggesting an origin by hydrous partial melting of basalt. These are similar, except for subtle differences, to those in Phanerozoic Andean margin settings. In contrast, Proterozic and early Paleozoic types are largely Sr-depleted and Y-undepleted, and Tarney infers these to have formed by intracrustal melting, but of subcratonic lithosphere or mafic lower crust. Tarney suggested that tectonic processes giving rise to granitoids have changed in response to Earth's changing thermal evolution.

Steve Haggerty then described a suite of interesting eclogite and granulite facies xenoliths from kimberlite pipes in the Archean Man Shield of West Africa. The xenoliths include lithologies ranging in composition from komatiite to anorthosite and appear to be geochemically, petrologically, and geophysically related. The suite may represent fractionation of felsic material separated from ancient mantle and added to early Archean crust. The samples can be used to define a xenolith geotherm, which may represent an ancient episode of high heat flow. The samples also imply that the crust-mantle boundary

is a gradational and possibly interlayered geochemical, mineralogical, and seismic transition. Haggerty speculated that the depleted subcontinental mantle required by diamond-bearing xenoliths of ultramafic affinity may have developed by coalescence of smaller depletion cells formed by extraction of ancient crustal components. These depleted zones are surrounded by fertile asthenospheric mantle, which may have given rise to later flood basalts such as in the Karroo and Parana Provinces.

In discussion, Roberta Rudnick asked Haggerty if any mantle xenoliths are present in the suites he described. He replied that there were not, and this seems typical of eclogite-dominated inclusion suites. He speculated that the peridotitic component may have been disrupted by a closely associated carbonatitic event, which may also account for the the alteration of olivine phenocrysts in the host kimberlite. Clive Neal asked if Haggerty supports an origin for the eclogite xenoliths by subduction of oceanic crust. Haggerty certainly allows this as a possibility, and in fact there is some support for seawater interaction from oxygen isotopes, but other mechanisms also exist, such as underplating. Paul Weiblen asked about the areal extent of the xenolith populations, and Haggerty responded that since two sites 100 km apart have similar xenolith suites, and both show similarities to other xenoliths elsewhere in Africa, they must be tapping a reasonably widespread and homogeneous source. Dallas Abbott asked about the size of the depletion cells Haggerty referred to. Haggerty responded that DeBeers asked the same question, and that the size of the cells must have been on the order of 100 –200 km, although in his model, they grow with time. Haggerty was asked if any isotope work had been done on the diamonds from these localities, and he responded that this and other isotope work was in progress. Urs Schärer asked about the ages of the diamonds and their inclusions. Haggerty informed us that the ages range between 2 and 4 Ga by a variety of isotopic methods, but the most reliable ages are Sm-Nd ages of 3.2 –3.3 Ga for garnet inclusions in the diamonds. Schärer wondered if these were model ages, and Haggerty assured us that Steve Richardson's work establishes that these are absolute ages and not model ages. Kevin Burke asked if all diamonds are of that age, and Haggerty spoke of eclogitic diamonds from the Premier Pipe with ages of 2.8 Ga. There is some hint that eclogitic diamonds are somewhat younger than ultramafic ones. Bill Compston commented that there are some unpublished ion probe ages of about 1.8 Ga for zircon inclusions in diamonds from the Sloan Pipe of North America. Haggerty commented that this was very important, and that there is some evidence for different ages of inclusions in the centers versus edges of diamonds. Bill McDonough asked if the densities of some of the garnet-bearing eclogite xenoliths might be at odds with Tom Jordan's tectosphere model. Haggerty answered that the density constraints at least allow the eclogites to sink, but it is not clear whether they are capable of entering the asthenosphere. He also speculated that some of the ecolgite xenoliths may have been derived from the asthenosphere.

After a coffee break, M. Stein discussed the evolution of continental lithosphere as revealed by xenoliths from Mesozoic and Cenozoic basalts from the Northern Arabian Shield. Isotopic and other data allow the peridotite xenoliths to be divided into depleted and enriched types. Sm-Nd model ages of the depleted peridotites are 2.2 Ga, and enriched ones are 1.0 –0.8 Ga, but internal isotopic systems show evidence of later heating events, presumably associated with transport of the xenoliths to the surface, among other things. Lower crustal xenoliths have compositions from gabbro to anorthosite, and give Sm-Nd ages of 0.62 –0.70 Ga, interpreted as the time of crystallization from mafic melts. Stein suggested a model in which depleted mantle evolves to enriched subcontinental lithosphere by extensive production of basaltic melts at 1.0 –0.6 Ga. These melts were underplated and crystallized at lower crustal depths, giving rise to rocks ranging from pyroxenite to anorthosite. Later metasomatic events produced enriched peridotites and source regions for the basalts that host the xenoliths.

The first talk in the third part (Isotopes) of Session Two was given by Paul Taylor, who summarized the general constraints that isotopes place on crustal growth models. Constraints on the start-time for crustal accretion are given by the the oldest preserved rock units (3.8 Ga Isua metavolcanics, West Greenland), and even older mineral grains (4.1 –4.2 Ga detrital zircons, Mt. Narryer paragneisses, Western Australia), although positive epsilon Nd values in the oldest mantle-derived rocks imply the existence of earlier continental crust, assuming a chondritic Earth. A uniform rate of crustal growth should result in a present average crustal age of about 1.9 Ga, assuming no recycling and a 3.8 Ga start-time for crustal growth. This agrees well with Goldstein et al.'s Sm-Nd model age of 1.70 ± 0.35 Ga of modern river sediments, which presumably sample broad areas of Earth's presently exposed continental crust, and implies that a maximum of only 38% of the Earth's continental crust had formed by 2.5 Ga. Support for the assumption of no recycling comes from the good correlations between epsilon Nd and epsilon Hf on the one hand, and Pb/Ce versus Ce on the other, among oceanic basalts. Such correlations would be degraded if continental-derived sedimentary material was effectively recycled back into the mantle.

Bill Compston then discussed single crystal ion probe ages of zircons, which allow much better time resolution compared to other geochronological methods, although the technique is not without problems. He described rocks from two localities (Barberton, South Africa and Jack Hills, Western Australia) that contain composite zircon populations, including true magmatic zircons as well as a variety of xenocrystic types. It is often difficult to distinguish these; xenocrystic zircons, for example, cannot always be identified on the basis of morphology alone. Clearly, additional evidence is needed before making age interpretations. Compston also presented evidence of zircon growth long after the original time of crystallization, in some cases apparently at temperatures <300°C. He then gave an account of the spectacular discovery of 4.1 –4.2 Ga detrital zircons in metaquartzites from the Mount Narryer area of Western Australia. Similar zircons with ages as old as 4276 Ma have been found in the nearby Jack Hills area. The source areas or parent lithologies of these zircons have not yet been

determined, but Compston expects that they may be unrecognized or buried antecedents of the K-rich Narryer gneisses. Uranium or thorium concentrations of zircons evidently cannot be used to discrminate between felsic and mafic source rocks.

In the following discussion, Simon Harley asked Compston whether the young zircon growth he described could be related to greenschist facies metamorphism or possibly fluid interaction. Compston replied that this result was totally unexpected, and further work will be needed to resolve the problem, although there appears to be no evidence for a superposed greenschist facies event in the area. Francis Albarède asked if there was any additional geochemical evidence on these young zircons, such as Th/U or REE. Compston replied that the new zircons were not geochemically distinct from the older population, but he warned against using ratios such as Th/U as discriminators. In some cases, however, large differences in Th/U exist between zircon cores and metamorphic overgrowths. Gerry Schubert asked Paul Taylor about the possibility that modern-day sediments represent a sampling of young material, and therefore the 1.7 Ga average crustal age is really a minimum age. Taylor responded that this age respresents an upper limit, and that the constraint imposed by this value could be relaxed if, for example, the crust was age-stratified. The age of rocks in the lower crust is difficult to determine, but could be attacked by examining the isotopic signatures of volcanics that have assimilated lower crust by passing through it. Steve Goldstein offered support for Taylor's interpretation by pointing out that in his study, he sampled wind blown dust in addition to river sediments. He found no significant differences between these types of samples, and concluded, therefore, that river sediments were not biased toward a younger age, as would be the case if they preferentially sampled mountain belts. Francis Albarède commented that we really are not able to determine at present whether "deeper" represents "older." His work on Hercynian (France) basement xenoliths and granites derived from the lower crust show Nd residence ages that are not significantly older than surficial sediments. Taylor mentioned a study in Scotland, where volcanics intruded through Lewisian basement (2.7 Ga) have brought up material that is actually younger than the surface rocks. Paul Taylor was then asked whether crustal recycling could take place by delamination rather than sediment subduction. He responded that the Pb-Ce and Nd-Hf data would allow this as a possibility only if average total crust (i.e., unfractionated into upper and lower crustal components) was somehow recycled. Otherwise, the Pb/Ce and Nd/Hf correlations among oceanic volcanic rocks would be degraded.

After the lunch break, the session continued with Peter Kinny's talk about the isotope systematics of some of the oldest samples on Earth from both Greenland and Australia. Kinny has confirmed the antiquity of the 4.1 –4.2 Ga zircons from Western Australia; the model Lu-Hf age of these zircons, as measured with the ANU ion probe is 4.14 ± 0.24 Ga, although the oldest preserved rock units there are anorthosites with a Lu-Hf model age of about 3.73 Ga. He then reported U-Pb ion probe ages of detrital zircons ranging between 2.87 Ga and 3.89 Ga from an Akilia-association quartzite (West Greenland), whose age of

deposition is probably around 3.8 Ga. He argued that the younger ages in this range are discordant because of late Pb-loss, probably associated with a high grade metamorphic event at about 3.6 Ga. He argued that the earliest crust in West Greenland and elsewhere (Labrador, Antarctica) is about 3.9 Ga, but in some places, such as Western Australia, crustal evolution took place much earlier, perhaps starting as far back as 4.3 Ga. This would account for the presence in that terrane of abundant K-rich granitoid, the paucity of tonalitic and trondhjemitic materials, and the existence of Eu anomalies in early Archean sediments.

In discussion of this paper, Stephen Moorbath expressed concern that Kinny's implication of Amîtsoq gneiss as old as 3.9 Ga contradicts field evidence that indicates it to be younger than Isua and Akilia supracrustals, and hence it must be <3.8 Ga. Paul Taylor commented that wherever original (nontectonic) contacts between the two are observed, the relationship is that Amîtsoq gneiss always intrudes Isua (or Akilia) supracrustals. Chairman Compston requested that this discussion be deferred until a later time.

Martin Whitehouse then discussed the Lewisian Complex (Northwest Scotland) in terms of whether the late Archean crust there was produced continuously from 2.9 Ga to 2.65 Ga, culminating in a high grade metamorphic event, or whether events in the area were short-lived and episodic. Sm-Nd model ages of felsic gneisses from the area show a progressive increase from about 2.7 Ga for amphibolite grade gneisses in the northern part of the area to over 3.0 Ga for the Scourie granulites to the south. Whitehouse preferred to interpret this apparent age stratification as representing lateral diachronism rather than magmatic overaccretion, but it is not possible to distinguish whether the dominant tectonic mechanism was movement of mafic protocrust over a mantle plume or marginal accretion of island arcs.

In discussion, John Tarney asked Whitehouse if a similar age progression might result if the samples were taken from a single granulite terrane. Whitehouse responded that there is no evidence that the amphibolite facies gneisses had ever suffered a granulite grade event. Much of the following discussion related to details of samples analysed and methods of plotting isochron diagrams. It was suggested that perhaps detailed field work might be able to distinguish between some of the tectonic environments Whitehouse offered for the Lewisian terrane. He responded that this is difficult due to the high grade metamorphism suffered by the terrane. Kevin Burke suggested that the relatively small areal extent of the Lewisian exposures rendered the possibility of a firm interpretation of the tectonic enrironment rather low. In response to a question, Burke pointed out that he did not build the planet, but was merely trying to describe it! Simon Klemperer commented that the age-depth relationship observed in the Lewisian, which is based on surface exposures, may be misleading when viewed on a larger scale; reflection seismic work shows that lower crust looks very different from upper crust in this area. Lower crustal structures can evidently be traced far to the south across the Caledonian and may indeed be much younger than Archean.

The session continued with A. S. Cohen's talk on LIL depletion in Lewisian granulites. Severe depletions in U, Th, and other LIL have been well documented in Lewisian mafic and felsic gneisses, but new Pb isotopic analyses show little or no depletion in lithlogies with high solidus temperatures, such as peridotite. This suggests that LIL transport in this terrane took place by removal of partial melts rather than by pervasive flooding with externally derived CO_2. The Pb and Nd isotopic data gathered on these rocks show that the depletion and granulite metamorphism are distinct events about 250 Ma apart. Both fluid inclusions and cation exchange geothermometers date from the later metamorphic event and therefore have little bearing on the depletion event, suggesting a note of caution for interpretations of other granulite terranes.

Gerry Schubert then discussed his work with colleagues DePaolo and Linn on the methodology of determining crustal mass-age curves. Problems in doing this include determination of model ages and accounting for mixing of materials of different ages during crust-forming processes. These difficulties can be overcome with some reasonable assumptions and estimates of rock volumes based on areal proportions. Schubert and colleagues used this technique to construct a reasonably well constrained mass-age curve for the southwestern United States based on isotopic measurements on over 100 samples. The results imply that the crust in this area grew episodically at 2.8, 1.8, and 0.1 Ga. They estimate that it would take on the order of $10^4 - 10^5$ individual Sm-Nd isotopic measurements to carry out a similar exercise for continental crust worldwide.

In discussion, Steve Goldstein asked Schubert whether these results from the western United States would necessitate a modification of the Reymer-Schubert growth curve. Schubert responded that the area considered here is not large enough to represent more than a small bump on that curve. Estimates of the crustal growth rate for this area are much higher (about 400 km³ / Ma / km of arc length) than those for modern-day arcs, even considering the uncertainties in some of the assumptions, implying another process of crustal growth besides collision of island arcs. Kevin Burke pointed out that island arcs represent additions to preexisting crust, but Schubert commented that his growth rate considers only a 200 Ma interval around 1.8 Ga. Burke did not seem convinced that growth rate determinations were meaningful, inasmuch as material is added to continents in short, episodic spurts. Paul Weiblen commented that underplating is another process that needs to be considered. John Tarney asked about the growth rate in the Andean arc, where contamination effects may be expected to be much smaller. Schubert estimates a value between 20 and 40 km³ / Ma / km of arc length, which is much less than that for the western U.S. at 1.8 Ga ago. Francis Albarède asked whether we could distinguish a major addition of crust at 1.8 -2.0 Ga from a model involving effective crustal mixing but constant growth rate through geologic time. The latter case would result in a mean crustal age of 1.8 -2.0 Ga, as discussed previously. Schubert assured us that there is strong evidence for a crust-producing

event in the southwestern U.S. at that time, and others were quick to point out events of similar age in Australia and Scandinavia. Burke asked Moorbath if he felt there was a worldwide crust-producing event at about 2.0 Ga. Moorbath responded affirmatively, but pointed out that there were many other such events over the range of geologic time. Bill Collins commented that the 1.7–1.9 Ga event was a major continent-wide event in Australia, during which a large volume of crust was added. Compston expressed some doubt that all rocks of that age represent juvenile crustal additions. Nick Arndt voiced strong support for distinct crust-forming events at about 2.7 and 2.0 Ga in North America, Scandinavia, Australia, and southern Africa. Al Kröner asked Schubert if he thought these episodes of high crustal growth rate represent periods of faster sea-floor spreading, on the order of 10–20 times faster than present-day rates. Schubert said that there are some who believe this, but he prefers to interpret these periods in terms of different processes, major episodes of hot-spot additions. Tarney commented that the different trace element signatures of granitoids favored separate tectonic mechanisms of crustal growth. Schubert commented that the high crustal growth rate in the early Proterozoic could not be explained simply in terms of the Earth's higher heat production then.

After a tea break, F. McDermott presented Nd isotopic data for southern Africa in support of episodic crustal growth. Over 50% of the continental crust there had formed before 2.5 Ga, and less than 10% was produced after about 1.0 Ga. The data imply a mean crustal age of about 2.4 Ga for southern Africa, and a higher rate of crustal growth than that derived from Australian shale data, particularly during the Proterozoic. Isotopic data from Damara metasediments imply that there is no need to invoke decoupling of the Rb-Sr and Sm-Nd systems in the continental crust.

Urs Schärer then presented isotopic data in favor of a sharp increase in crustal growth during the mid-Proterozoic in Labrador. U-Pb ages of zircon, sphene and monazite show that, with few exceptions, large volumes of mafic to intermediate plutonic rocks were produced between 1660 Ma and 1630 Ma. The juvenile character of these rocks is supported by initial Pb isotopic compositions in feldspar separates, which show mantle-like signatures. Isotopic systematics in these rocks and minerals have survived variable effects of intense Grenvillian metamorphism at 980–1030 Ma.

In discussion, Lew Ashwal suggested that Schärer's conclusions could be extended to the entire Grenville Province, in which there is no evidence for crust older than about 1.6–1.7 Ga (except for thin zones reworked Archean Superior Province rocks near the Grenville Front). Bill Collins was asked whether the mantle-like initial Pb values of K-feldspar in these granitoids implies rapid reprocessing of mantle-derived rocks. Schärer reiterated his conclusion that the granitoids represented juvenile crust. Ashwal added that the nearby anorthosite massifs of the same age represent voluminous mantle-derived materials. Al Kröner asked McDermott about his preferred tectonic model for the Damaran Belt of Southern Africa. McDermott apparently favors intracratonic rifting followed by a continental

collision. Chairman Compston then requested additional discussion on a contentious point raised by Peter Kinney in his talk. Kinney's work on the Kapisigdlit quartzite (West Greenland), which is presumably an Akilia-association inclusion in Amîtsoq gneiss, implies an older Amîtsoq age (about 3820 Ma) than has previously been measured. Kinney suggested therefore, that this quartzite, and possibly other supracrustal enclaves may predate the Isua sequence. Akilia supracrustals themselves have thus far not proven amenable to age determination. Stephen Moorbath argued that 15 years of field and laboratory work has shown incontravertibly that Amîtsoq gneisses postdate Isua supracrustals. He suggested that Kinney must be extremely careful to understand the field relations of his samples and the textural context of their zircons before making age interpretations. Kinney explained why the old zircons in his sample of Amîtsoq gneiss were interpreted as an igneous population, rather than an inherited or xenocrystic one. Moorbath suggested that Kinney and colleagues return to this locality and attempt to obtain materials to provide an age of the Akilia supracrustals, but Kinney pointed out that igneous lithologies were very difficult to identify among the Akilia suite.

SESSION THREE: MODELS OF CRUSTAL GROWTH AND DESTRUCTION

This session opened with the final talk of the second day by Paul Morgan, who discussed thermal models pertaining to continental growth. Since the Earth has been cooling since its formation, a reasonable expectation would be that tectonic processes have been slowing down. Any nonsmooth crustal growth curve implies that uniformitarianism, in the strictest sense, does not apply to the Earth. There are alternatives to plate tectonic heat loss, however, such as hot spots; these may have dominated during early Earth history. A fundamental outstanding problem is the fate of Earth's primordial crust. If it was more felsic than basalt, impact or tectonic processes would be incapable of destroying it completely. Regarding the comparatively LIL-poor composition of surviving Archean crust, Morgan suggested that lithospheric strength may have been much lower during the Archean, in response to a hotter thermal regime, and that crust high in LIL elements may have been selectively destroyed by reworking. Selective preservation of crust may have decreased later in Earth history as mantle strength became a significant factor in the cooler thermal regime. In order to account for the apparent dichotomy between a hotter Archean thermal regime and the evidence for thick lithosphere required by diamonds of Archean age, Morgan offered a model in which increased heat production in Archean mantle lithosphere resulted in a lithospheric geotherm asymptotic to that of underlying asthenosphere. In this way, heat input into the base of the ancient lithosphere would be minimized, and relatively thick lithosphere could coexist with a hotter Archean asthenosphere.

In discussion, Dallas Abbott suggested that the LIL-poor Archean crust could be accounted for by shallower subduction then, because subducting oceanic crust would melt at a shallower

depth. Morgan agreed that the hotter Archean thermal regime would favor shallower subduction on average, but he pointed out that the angles of modern-day subduction zones are highly variable, and they are likely to have been variable during the Archean. The mean angle of subduction should gradually increase in response to the cooling Earth, and the onset of K-rich granitoid production should be gradual, rather than abrupt, as it appears to be at about 2.7 Ga. Kevin Burke clarified his position that uniformitarianism merely implies that the laws of physics have not changed over geologic time. Morgan and others in the audience felt that uniformitarianism implied that processes as well as their rates had not changed with time. Roberta Rudnick asked if Morgan's model of selective reworking implied an Archean mantle heat flux twice that of the present-day, and if so, how this could be reconciled with the existence of Archean diamonds. Morgan replied that the higher heat flux he discussed was for tectonically active areas and reiterated his thermal model, which attempts to account for the diamonds. Gerry Schubert expressed doubt about the thermal stability of Morgan's model. Morgan replied that if the thermal conditions at the base of the lithosphere were the same as that in underlying convecting lithosphere, no transfer of energy would take place, and the situation would be stable. Schubert did not seem convinced, and Morgan agreed to discuss the matter at a later time.

Paul Warren then described his model of komatiite generation at the base of a relatively thick Archean lithosphere, taking into account the high Archean heat production. Warren feels that the Archean mantle may have been effectively dehydrated in the aftermath of the impact phase of Earth history. This would have raised mantle melting temperatures and produced thicker lithosphere. Paul Toft suggested that Warren's model did not take into account constraints from phase equilibria, which suggest, for example, that CO_2 played an important role in komatiite generation. Warren defended his model by arguing that CO_2 and water have opposite effects on melting temperatures. Nick Arndt asked Warren if the 80 Ma Gorgona komatiites implied that a pocket of dry mantle has survived since the impact stage of Earth history. Warren indicated that the lower MgO content of those samples implies a lower melting temperature than Archean komatiites. Arndt replied that a good case could be made at Gorgona for melts with 22% MgO, implying eruption temperatures of about 1550° C. Warren argued that on a statistical basis, formation of higher temperature melts would be produced more frequently if the mantle was on average dryer during the Archean. Warren admitted, however, that his model would be hard to test. Gerry Schubert commented that Earth degassing models would imply a wetter Archean mantle, and Kevin Burke interjected that this view is not as widely held now as several decades ago. Schubert wondered how the mantle becomes regassed, and Burke suggested subduction as the major process.

Gerry Schubert continued Session Three with the first talk the following day on physical processes of crustal growth. Mechanisms of crustal addition include volcanism and plutonism at plate margins (island arcs), or within plate interiors (hot spots). Mechanisms of crustal subtraction include sediment subduction,

and possibly lithospheric foundering (delamination) and subduction erosion. Island arc additions have dominated during the Mesozoic-Cenozoic (estimated at about 1.1 km^3/yr), but hot spot additions have not been insignificant (about 0.5 km^3/yr). Schubert described three-dimensional convection calculations that show that plumes (hot spots) are a necessary consequence of plate tectonics, because thermal boundary layers are inherently unstable. The relative importance of hot spot versus plate margin magmatism may have been different in the past, but it may be misleading to assume that one mechanism was operating to the exclusion of the other at any period of Earth history. Isotopically determined crustal growth rates from several Precambrian terranes are about an order of magnitude larger than that estimated for modern-day island arcs (20 –40 km^3/Ma/km), suggesting faster Precambrian plate margin and/or hot spot tectonics. If continental freeboard has been constant throughout geological time, for which there is some evidence, then the crust has grown only by about 25% since the end of the Archean. Approximately 6% of potential continental material exists in oceanic plateaus such as the Falklands and Ontong-Java plateaus.

Dallas Abbott then discussed heat loss in the ancient Earth assuming that classical sea-floor spreading was the only mechanism. This may be expressed either as faster spreading or longer total ridge length. These have important implications as to the size and number of cratonic plates in the distant past, the degree to which they are flooded, the kinds of sediments and volcanics that would be expected, and the amount of recycling of continental material taking place. The higher proportion of marine sedimentary rocks and oceanic volcanics in the Archean, and the relative paucity of evaporites and continental volcanics may in part be due to smaller cratonic blocks.

In discussion, Bill Collins asked Schubert if he could place any time constraints on the process of lithospheric delamination and subsequent plume generation. Schubert estimated a period of about 30 –50 Ma between delamination and onset of plume activity. Subsequently plumes rise rapidly through the mantle from a deep boundary layer. The delamination process itself would require thickened lithosphere to be placed over a region of low viscosity, and plume activity may actually initiate this. Stephen Moorbath asked Schubert how his freeboard model costrained how much higher continents would stand above mean sea level at 2.5 Ga. Schubert said that continents would stand about 0.5 km higher for models requiring >50% continental growth since 2.5 Ga, and still higher for models with greater growth. Moorbath stated that Archean continents may have been thicker than present-day ones, perhaps about 60 –80 km thick. Several participants pointed out that constraints on Archean crustal thickness based on P-T determinations of granulites may be misleading because these record orogenic events rather than steady-state conditions. Paul Morgan pointed out that Archean crust constitutes only a tiny fraction (1.7%) of present continental area, and this may be unrepresentative. One participant commented that Schubert's plume model was inconsistent with our understanding of modern hot spots, which have great stability with respect to each other and to the Earth's

spin axis. Schubert disagreed and said that the instabilities of the thermal boundary layer that are responsible for plumes will last on the order of 200 Ma. This is equivalent to how far back in time we can trace modern plumes. Regarding Schubert's delamination model, Al Kröner asked how lower crust could sink through relatively rigid mantle lithosphere. Schubert replied that this could be accomplished, in principle, if lithosphere were delaminated first. Bill McDonough asked if Schubert had tried his model with a density contrast between lithospheric and asthenospheric mantles. Schubert agreed that his model would not work if the mantle lithosphere was buoyant, for example, because of being depleted in fertile basaltic components. McDonough then asked about the fate of buoyant depleted mantle, and Schubert replied that it might be circulated through the mantle, depending on how large a region could accumulate. Such material could even be subducted if it did not maintain its integrety on a large scale, or it could be removed by horizontal convection. Dallas Abbott asked why the Canadian Shield does not contain abundant kimberlite pipes if hot spots were a major mechanism of heat loss then. Schubert could not offer an answer.

Paul Weiblen then changed the topic of discussion, and showed a few slides relating to the geology of the Superior Province, in which there is evidence for a three-stage process of basalt generation, subsequent burial and melting to produce tonalitic material, and further burial and melting to produce granitoid rocks. He wondered what the burial mechanisms were, whether or not the Superior Province was pieced together by island arc collision, and if it represents a terrane of extensive recycling. Bill Collins commented that the Pilbara block seems not to represent accreted terranes, but rather to have formed relatively rapidly, perhaps in response to hot spot activity. He admitted that evidence is equivocal as to whether greenstone belts represent arc or hot spot tectonic settings, and that this must be determined before any theorizing will be meaningful. The discussion then shifted to Schubert's statement about the equivalence of heat flux between Io and the Earth. It was pointed out that either Io is much younger than the Earth, or it has a much different way of generating heat. Schubert replied that Io was not younger than Earth, and it indeed generates heat differently—by dissipation of tidal energy caused by its proximity to Jupiter. Steve Goldstein asked Schubert if in his convective model he could account for the geochemical constraint about the existence somewhere in the Earth of an undifferentiated mantle reservoir. Several participants questioned this as a constraint, and Goldstein pointed to the rare gas data that some believe require a primitive mantle source. Schubert expressed his confidence that his model (and others) could be made to fit the data. Francis Albarède asked Schubert how layered convection would affect the time constants of his model. He responded that layered convection would provide additional boundary layers that could act as sources of plumes, and depending on the density contrasts between the layers, the plumes may or may not be able to penetrate through any given interface. Schubert commented that he did not believe geochemistry demands the lower mantle to be a primitive reservoir.

The session then continued with Chuck Wood's talk on estimating the volcanic contribution to crustal growth in the central Andes. Previous estimates of the rate of magmatic additions to the crust in $1° -3°$ areas, based on heights of Andean volcanoes and their ages, range between $1.6 -4.2$ km³/Ma/km of arc length. Based on a census of 1077 Andean volcanoes mapped between 14° and 28°S using Landsat and Space Shuttle photographs, Wood and colleagues have made new growth rate estimates. The growth rate based on visible volcanic deposits (cones + ignimbrite sheets) is about 1 km³/Ma/km of arc, but Wood argued that this is an underestimate because it does not take into account material stranded in magma chambers or dispersed great distances by explosive eruptions. Wood estimated that these so-called "invisible" contributions represent a growth rate of over 16 km³/Ma/km of arc, most of which is from distantly deposited material. About 5 km³/Ma/km of arc of these positive additions to the crust represents reworked crustal material, and must be subtracted, giving a total growth rate of $11 -12$ km³/Ma/km of arc. This value is $3 -7$ times larger than previous estimates.

In discussion, one questioner took issue with Wood's assumption that andesites were 100% mantle derived, and further stated that Wood did not take into account buried mafic cumulates complementary to the andesites. Wood agreed. Al Kröner stated that ignimbrites represent a much smaller fraction of intraoceanic arcs, and that Reymer and Schubert's growth rate values may be appropriate for a global crustal growth rate. Wood replied that ignimbrites were abundant in oceanic arcs. Both agreed that ignimbrites represent <40% of such arcs. Abbott commented that in the Cascades the rate of magmatic additions was roughly proportional to convergence rate and asked if Wood had taken the latter into account in his calculations. Wood replied that he just considered what was produced over the last 20 Ma, regardless of convergence rate. Prior to 20 Ma ago, there was a period of 10 -15 Ma during which there was no volcanism in this part of the Andes, and if this period is included, Wood's growth rate estimate would decrease by a factor of two. Bill Hartmann pointed out that the length of the Andean arc may have been different in the past, and he asked how Wood determined the mean arc length for the period of time he considered. Wood responded that he used the 37,000 km value of Reymer and Schubert, and that his calculations are based on the assumption that the tectonic and magmatic processes operating in the Andes have not substantially changed during the last 20 Ma. Stephen Moorbath pointed out that there are differences in crustal thickness, styles of volcanism, and amount of crustal contribution to the magmas in different parts of the Andes. It is dangerous, he argued, to make general conclusions about the entire Andes from the 14° -28°S region Wood considered. Wood agreed, but justified using the area he studied on the basis of good ground control and availability of high-quality spacecraft imagery. Randy Forsythe pointed out that Wood's estimates may be too high because they include the sub-Andean crustal keel, which may be tectonic in origin. Wood commented that recent studies show the keel to represent a smaller volume than previous estimates. Schubert asked if anyone still subscribed to the idea that the abnormally thick Andean crust was due to tectonic compression. Forsythe pointed out that calculations by the Cornell group indicate that the

entire keel could be produced tectonically in the late Miocene, with a minimal component from magmatic underplating. Moorbath commented that thick Precambrian crust is present below the central volcanic zone. Schubert said that the high calculated crustal production rates would be substantially lowered if the keel were not considered. Jan Veizer commented to the agreement of Chuck Wood that there were very many uncertainties in these sorts of calculations. Moorbath pointed out to everyone's amusement that if the uncertainties were statistically distributed, the answers would be right. Bill Collins asked Wood how much of the Andean granitoid batholith was assumed to be mantle-derived. Whereas a decent case can be made from isotopic measurements that Archean and Proterozoic granitoids represent juvenile crust, it is not clear how much recycled Precambrian crust is present in younger granitoid batholiths. Wood felt that the Andean batholiths contain a substantial fraction of recycled crust and suggested that the only new material may be the basic rocks we now see at the surface. Wood said that in any case he did not consider this problem in his calculations. Dallas Abbott voiced her view that there were likely to have been substantial time variations in the rate of crustal addition by island arc magmatism, and we should consider factors such as hotter mantle and different plate geometries in considering these rates back as far as the Archean. Eric Nelson suggested further that in the past island arc magmas may have been produced by direct slab melting rather than by melting of the hydrated mantle wedge. Bill Collins wondered about the paucity of evidence for the products of island arc environments in Precambrian terranes. Some feel that these are represented by Archean greenstone belts, but Collins stated that such evidence is lacking in Proterozoic terranes. Paul Weiblen suggested that more work is needed in Precambrian shield areas to unravel the isoclinally compressed volcanic terranes commonly present there before we can reliably interpret their tectonic settings. Collins suggested that it is the geochemists rather than the structural geologists who are the main protagonists of island arc settings for Precambrian rock units. Many participants disagreed with this, and Chairman Schubert wisely adjourned the session for a coffee break.

Francis Albarède then discussed Sm/Nd isotopic constraints on crustal growth. In order to constrain Sm/Nd fractionation between continental crust and depleted mantle, Albarède and colleagues compiled an extensive data base of isotopic measurements of mantle-derived igneous rocks and fine-grained clastic sediments (assumed to be adequately representative of continental crust). The results imply that the evolution of depleted mantle has been roughly linear, with no major discontinuities over the course of geologic time. This is different from other determinations of depleting mantle evolution, which show nonlinear behavior. The Sm/Nd evolution lines for continental crust and depleted mantle intersect between 3.8 – 4.0 Ga, which may indicate that the onset of continental growth was later than 4.5 Ga. Albarède then described his mathematical model, the results of which imply that time-integrated crustal additions from the mantle are about 1.8 –2.5 km³/a, whereas crustal subtractions by sediment recycling are about 0.6 –1.5 km³/a. This results in a net time-integrated crustal growth rate

of about 1 km³/a, which is similar to present-day rates determined, for example, by Reymer and Schubert.

In the next talk, Steve Goldstein presented evidence that the Rb-Sr and Sm-Nd isotopic systems are decoupled in crust-mantle evolution. REE (including Sm and Nd) reside principally in silicates, and are resistant to mobilization by weathering and metamorphism. In contrast, Rb and Sr are easily fractionated by crustal processes and reside in carbonates as well as in silicates. As a result, continental Sr, but not Nd, can be recycled into the mantle by exchange of seawater with basalt at spreading ridges and by subduction of carbonates associated with ridge processes. These effects result in mean Rb-Sr ages of the continental crust and of the upper mantle that are too young. Crustal growth curves based largely on Rb-Sr data, such as that of Hurley and Rand, are therefore incorrect.

The final paper of the morning session was given by R. Ellam, who addressed the question of whether present-day subduction zone magmatism produces material of average continental crust composition, which perhaps most would agree is andesitic. Ellam argued that modern andesitic to dacitic rocks in Andean-type settings are produced by plagioclase fractionation of mantle-derived basalts, leaving a complementary residue with low Rb/Sr and a positive Eu anomaly. This residue must be removed, for example by delamination, if the average crust produced in these settings is andesitic. Ellam argued against this, pointing out the absence of evidence for such a signature in the mantle. Either the average crust is not andesitic, a conclusion Ellam was not entirely comfortable with, or other crust-forming processes must be sought. One possibility is that during the Archean, direct-slab melting of basaltic or eclogitic oceanic crust produced felsic melts, which together with about 65% mafic material, yielded an average crust of andesitic composition.

Jan Veizer opened the discussion by pointing out that Goldstein's 400 Ma regression of Sr isotopic data from Paleozoic sediments, and also the estimate of about 20% of upper mantle Sr having been derived from continental crust, are consistent with the Veizer and Jansen recycling model. He also disagreed with Albarède's interpretation of Nd isotopic data from sedimentary rocks. Veizer stated that the Nd data could also be explained by a stationary sedimentary mass since 2.5 Ga, with 90% sediment recycling and 10% exchange. Albarède agreed that some cannibalization of sediments likely took place, but this does not imply complete recycling. Veizer also expressed concern about the paucity of Nd isotopic data near 2 Ga, and that discontinuities in Nd isotopic evolution could not be ruled out. Albarède conceded that more data is needed for the period 1.5 –3.5 Ga, but he felt that there was no evidence for discontinuous Nd evolution. Goldstein commented that a different picture might emerge if Albarade had plotted all Nd data available in the literature. Veizer commented that the rate of sediment recycling is one or two orders of magnitude faster than other crustal processes such as metamorphism or melting. Albarède agreed, but pointed out that sediment recycling is a process that takes place within the crust, and although it is uncertain how well stirred the continental crust is, it is more important to consider exchange between large-scale reservoirs such as the crust and mantle. Julius Dasch pointed out that

in addition to extensive Sr exchange at ocean ridges, large amounts of Rb can be shown to be taken up by oceanic crust during hydrothermal alteration. He also commented that the Sm/Nd isotopic system may not be as immobile as Goldstein inferred. Goldstein agreed that the Sm-Nd system is not completely immobile, but expressed doubt that these processes have an effect on the whole system, and Dasch concurred. Albarède suggested that the behavior of Rb during alteration of oceanic crust was not completely understood. Dasch disagreed, citing Stan Hart's work of 20 years ago, which shows that K, Rb, and Cs increase by factors of between 5 and 50 during hydrothermal alteration. Albarède commented that extraction of Rb from oceanic crust in black smokers is on the same order of magnitude as Rb influx during hydrothermal alteration, and that the Rb budget in the total system was quite uncertain.

Stephen Moorbath asked Rob Ellam to speculate about why the primary magmatic additions to the crust in Andean settings are basaltic today but were more andesitic in the Archean. Ellam suggested that in Archean subduction zones a basaltic or eclogitic source was melted rather than a peridotitic source as in modern-day examples. Moorbath asked if Ellam felt there were any primary andesites forming today, for example, in intraoceanic environments. Ellam responded that his conclusions were based strictly on Andean-type environments where Sr isotope evidence could not support a primary origin for the andesitic rocks. Albarède asked Ellam how confident he was that the measured Rb concentrations in Archean rocks represent true magmatic values. Ellam replied that, if anything, Rb would have been removed from the rocks by weathering, and Rb/Sr ratios would correspondingly decrease. Goldstein commented that he would expect Rb to increase under such conditions. Moorbath stated that Archean rocks from Zimbabwe he measured show tolerable Rb-Sr isochrons. Albarède expressed confidence in such a result for granitoids, but not for mafic rocks, and Moorbath concurred. Ellam commented that his arguments do not rely heavily on data from Archean mafic rocks. Bill McDonough stated that Taylor and McLennan's model favored formation of more mafic crust in the Archean and more andesitic crust during the Phanerozoic, and he asked Ellam or Albarède to comment on this in terms of the proposed constancy of Sm/Nd ratio in crust and (by implication) mantle reservoirs as a function of time. Albarède replied that the Nd isotope data base prohibits a major change in Sm/Nd during the course of Earth history but cannot rule out subtle changes. Gerry Schubert mentioned DePaolo's recent work, which claims that there has been an increase in Sm and Nd concentration in average crust with time. In a comment directed at Ellam, Nick Arndt pointed out that in the Belingwe greenstone belt of Zimbabwe, K concentrations, K/Sr ratios, and presumably Rb/Sr ratios increase in progressively more Fe-rich rocks, but the increase is greater than can be accounted for by fractional crystallization, implying some secondary process. Ellam replied that fractional crystallization effects could have produced some of the scatter on the diagram of Zimbabwe data he showed, but his point was that Sr was behaving as an incompatible element in these rocks, suggesting that plagioclase was not playing a role, and hence the high Rb/Sr ratios were not produced by crustal

processes. Arndt pointed out that similar trends are observed in single differentiated units where fractional crystallization is the major process, which suggests some secondary increase in the K content. John Tarney commented that tonalitic intrusives of the Andes, which crystallized at depth by virtue of their water content, have not undergone the loss of Sr characteristic of surficial andesitic volcanics, and therefore may not have experienced removal of plagioclase components. Roberta Rudnick asked if Tarney was implying that the tonalites of the Andes were mantle-derived rocks; and he responded affirmatively. Rudnick stated that this was incompatible with experimental data, and Tarney clarified his view by stating that they were mantle-derived in the sense of being produced in a subduction environment. Chairman Schubert then adjourned the meeting for lunch.

SESSION FOUR: PROCESSES OF CRUSTAL GROWTH AND DESTRUCTION

Jan Veizer opened this session with his talk on the solid earth as a recycling system. Using an analogy of human population dynamics, he distinguished *internal* recycling, such as sediment cannibalization or resetting of isotopic ages, from recycling via an *external* reservoir, such as subduction of oceanic crust. The concept of residence time relates only to the external component of recycling, whereas "mean age" relates to both internal and external recycling. Mean age, therefore, is always shorter than residence time. Veizer defended his views on recycling with numerous examples of geological processes, and then cited numerous observations in support of his family of crustal growth curves (see cover figure), including the secular changes in the Sr isotopic composition of seawater, the increase in K/Na of granitoids, and the REE patterns of mature sediments.

Steve Galer then discussed isotopic and geochemical evidence against recycling of continental crust into the mantle. Element ratios such as Sm/Nd, Th/Sc, and U/Pb in sedimentary masses have remained relatively constant throughout Earth history, and this can only be reconciled with steady-state recycling models if new crustal materials added from the mantle have had similar ratios. Such recycling models would also require shorter processing times for U, Th, and Pb through the mantle than are geodynamically reasonable. Models favoring subduction of pelagic sediments as the only recycling mechanism fail to account for the Pb isotopic signature of the mantle. Recycling of *bulk* crust with Pb isotopic compositions similar to those expected for *primitive* mantle would be permissible with available data, but there appear to be no plausible tectonic mechanisms to carry this out.

In discussion, Simon Klemperer asked Galer if any of his models would allow the 0.6 km³/yr of recycling suggested by Schubert. Galer replied that an upper limit for recycling would be about 1 km³/yr. Subduction of pelagic sediments at such rates would completely dominate the Pb signature of the upper mantle in about 0.5 Ga, and this is unlikely. Gerry Schubert asked if Galer's numerical modeling was dependent on the size of the mantle reservoir under consideration. Galer responded that he considered both upper mantle and whole mantle

reservoirs; neither case would allow substantial recycling. Francis Albarède then made several comments: (1) the Pb isotopic composition of the continental crust and depleted mantle were not substantially different for about the first 3 Ga of Earth history, (2) we do not really know how much Pb is stored in the lower crust, (3) we do not really understand how Pb is fractionated during extraction of continental crust from the mantle, and (4) the time constants for the Sm/Nd and other isotopic systems might be much longer than Galer assumed, perhaps on the order of the age of the Earth. In response to Albarède's first point, Galer reiterated that subduction of pelagic sediments would return far too much Pb into the mantle in too short a time to be compatible with available data. Chairman Veizer cut off the discussion at this point.

The next presentation was given by John Tarney (for Andy Saunders). He discussed oceanic basalt geochemistry as a way to monitor the signature of recycled crustal materials in their sources. Some ocean-island basalts (OIB), such as the St. Helena-type, show chemical and isotopic features consistent with a source containing recycled *oceanic* crust, but with no sedimentary component. The "DUPAL" OIB source could contain a very small (< 1%) component derived from subducted pelagic sediments, but is also dominated by recycled oceanic crustal material. Tarney argued that the complement to present-day depleted mantle consists of these components of OIB sources as well as the continental crust.

Nick Arndt then presented an interesting mechanism for recycling of lower continental material back into the mantle. Picritic magmas, possibly parental to voluminous continental volcanics such as the Karoo and Deccan, became trapped at the Moho, where they interacted with and become contaminated by lower crustal materials. Upon crystallization, the magmas differentiated into lower ultramafic cumulate zones and upper gabbroic-anorthositic zones. The ultramafic cumulates will be denser than underlying mantle and sink, carrying lower crustal components as trapped liquid, as xenoliths or rafts, and as constituents of cumulate minerals. This model provides a potentially significant crust-mantle differentiation mechanism, and may also represent a contributing factor in crustal recycling, possibly important in producing some OIB reservoirs.

In discussion, Lew Ashwal pointed out that Arndt's model is essentially the same as one proposed by Emslie to account for massif-type anorthosites, except in Emslie's model buoyant plagioclase-rich mushes rise to the surface. Ashwal stated that there is no evidence from xenolith populations, for example, that anorthosite is a substantial constituent of the present-day lower crust. Arndt cited the work of Roberta Rudnick, which shows that many lower crustal xenoliths are Al-rich, and have positive Eu anomalies. Paul Taylor then showed slides illustrating a Proterozoic layered anorthosite-norite-troctolite intrusion at Flakstadoy (Lofoten Islands, Norway), which may represent support for Ardnt's model. The mafic rocks of the intrusion have been severely contaminated with Ar^hean high-grade, presumably lower crustal materials. Some of these rocks have been recrystallized to eclogite facies mineralogy, and hence may have sufficient density to sink, as suggested in Arndt's model.

Taylor then argued that lower crustal materials could be expected to have low Rb/Sr, and that contamination by this type of material would result in a low, rather than a high initial Sr ratio signature, as Arndt indicated. Ardnt concurred with Taylor's interpretation of the Norwegian layered intrusion. Regarding the Sr isotopic question, Arndt stated that he was influenced by Ben-Othman's study of lower crustal granulites that have high initial Sr ratios despite their low Rb/Sr. Taylor commented that this was produced by Rb depletion long after those rocks formed and should be considered the exception rather than the rule. Steve Goldstein commented that a lower crustal contaminant would still be consistent with the elevated Sr isotopic composition of OIB compared to MORB. Taylor disagreed, pointing out that 2.7–2.8 Ga Scourian granulites have present-day Sr isotopic compositions indistinguishable from those of MORB. Francis Albarède commented that the type of magmatic activity discussed by Arndt should be much more abundant in oceanic settings, and wondered which sorts of settings these would be represented by. Paul Weiblen suggested that the parental magmas of MORB may also be picritic, but Albarède stated that this was still an unresloved issue. Paul Toft commented that the occurrence of banded garnet anorthosite xenoliths in a Liberian kimberlite supports Arndt's model. In a comment directed at Veizer, Toft wondered if the Earth-systems approach to crustal growth modeling predicts that the present-day Earth has already produced as much continental crust as is ever going to be generated. Veizer replied that this would be true for the present steady-state, but that steady-states and systems can evolve. Returning to the subject of Arndt's talk, Al Kröner commented that it would be difficult for contaminated cumulates to sink through rigid lithosphere. Arndt replied that even though the lithosphere is rigid, it is not undeformable. Stephen Moorbath expressed his expectation that contaminated ultramafic cumulates would simply remain in the lower crust rather than sinking into the mantle. Arndt reiterated that these would be gravitationally unstable and would indeed tend to sink. At this point Chairman Veizer suggested that the meeting be adjourned for a tea break.

The session then continued with Stuart McCallum's talk on the petrology and geochemistry of the Stillwater Complex, an Archean (2.7 Ga) layered mafic intrusion in the Beartooth Mountains of Montana. McCallum discussed his efforts to reconstruct the compositions of possible parental magmas and thereby place some constraints on the composition and history of their mantle source regions. A high-Mg andesite or boninite magma best matches the crystallization sequences and mineral compositions of Stillwater cumulates, and represents either a primary magma composition or a secondary magma formed, for example, by assimilation of crustal material by a very Mg-rich melt such as komatiite. Isotopic data do not support the extensive amounts of assimilation required by the komatiite parent hypothesis, and McCallum argued that the Stillwater magma was generated from a mantle source that had been enriched by recycling and homogenization of older crustal material over a large area.

Jan Veizer then discussed mineralization through geologic time in terms of crustal evolution and recycling. Using a large data

base of mineral deposits grouped according to age, tonnage, and genetic type, Veizer described his attempts to predict the probabilities of survival of ore deposits in the framework of plate tectonics, assuming recycling was working. He argued for a decreasing role of recycling over geologic time, and described five overlapping stages of metallogenic evolution of the Earth: greenstone belts, cratonization, rifting, stable craton, and continental dispersal.

In the ensuing discussion, Maureen Wilks suggested that Veizer should take into account the evolution of the atmospshere and related biological effects in addition to tectonic environment in considering metallogenesis. Veizer replied that he did not have sufficient time to discuss this in his talk and argued that biological effects were minimal. Wilks disagreed, stating that algae, for example, had a profound effect on the Archean atmosphere. Veizer stated the reverse point of view, arguing that biological and atmospheric evolution must be considered from a systems standpoint. Steve Goldstein asked Veizer about the occurrences of ancient detrital uraninite, which some cite as evidence for a reducing atmosphere. Veizer argued that during the Archean, atmospheric oxygen was stripped by processes involving Fe oxidation; in the early Proterozoic, the buffering mechanism switched to C and S oxidation, and the steady-state level of atmospheric oxygen increased. Al Kröner then commented that some of Veizer's compilations of ore deposit types versus age may be misleading or incomplete, and that, in general, their distributions do not reflect the tectonic setting in which these rocks formed. Veizer replied that he did not have time to discuss the statistics of mineral deposit distributions. Stephen Moorbath asked McCallum to elaborate on his conclusion of crustal recycling, and in particular, why he favored incorporation of an Archean crustal component into the mantle source of the Stillwater Complex, rather than as an assimilant into the parental magma. McCallum agreed that more appropriate terminology would be that the magma was derived from "a mixed protolith" rather than from a mantle source subjected to "crustal recycling."

The final talk of the day was given by Eric Nelson, who described a model of batholithic construction in Andean arcs and its applicability to possibly similar environments in the past. Age and compositional data from the Patagonian batholith of southern Chile show a long history of magmatism in any given area (total age range is 15–157 Ma), but different regions appear to have different magmatic starting ages. Furthermore, mafic rocks seem to be the oldest components of any given region. Nelson outlined an assembly line model involving semicontinuous magmatism and uplift, which has implications for other terranes: uplift rates will be proportional to observed ranges in age, and total uplift will be proportional to the age of the oldest pluton in any given area. Nelson suggested that misleading results would be obtained if only small areas of similar terranes in the Archean were available for study.

In discussion, Bill Collins suggested a simple field test for Nelson's model: in any given terrane, the older granitoids should have been emplaced at the higher levels, because if the terrane

is undergoing concomitant uplift, then subsequent plutons will tend to be emplaced at relatively deeper levels. Nelson replied that such a scenar io would result in a correlation between age and metamorphic grade. Since this is not observed, he concludes that only a portion of the crustal section is undergoing uplift, that plutons essentially "pump up" the crust and are emplaced at relatively similar depths. Stephen Moorbath noted that in the Patagonian example, initial Sr ratios decrease with time, as the plutons became more felsic. Nelson commented that in the Peruvian Andes the situation is reversed, probably because of the presence there of an old, thick crustal root. Urs Schärer asked Nelson if each initial Sr ratio he showed represented an individual isochron. Nelson replied that most values were single whole-rock samples whose ages were assumed from Ar-Ar biotite or hornblende data. Schärer wondered, then, how Nelson could be sure there was no Rb loss or enrichment. Nelson acknowledged this as a problem. Schärer stated further that the age distribution Nelson discussed could possibly be an artifact of different blocking temperatures for the different isotopic systems used (U-Pb, Rb-Sr, Ar-Ar). Nelson replied that this also might be a problem, but he felt that with the large range of ages, this effect might be minimal. Chairman Veizer then adjourned the session.

Session Four continued the next morning with John Dewey's talk on the tectonics of crustal growth and evolution. He described the multiplicity of mechanisms whereby continental crust is generated, redistributed, and modified in *modern-day* settings. Dewey feels that the Precambrian Earth operated much the same way as today, although there has likely been some secular evolution in these mechanisms, and it will be difficult to sort this out in terms of volume of crust generated versus time. He then described the complexities of tectonic settings such as subduction zones, rifts, continent-continent collision zones, and terranes subject to extensive basaltic underplating. Dewey reiterated a point made earlier by Kevin Burke, namely, that the complexity of modern-day settings such as the Caribbean or Southeast Asia should not be forgotten when attempting to understand the rock record in the distant past.

Al Kröner asked Dewey why the long, linear orogenic belts common in Phanerozoic and Proterozoic terranes seem to be absent in the Archean. Dewey argued for differences in rates rather than mechanisms. Higher Archean heat production might lead to faster plate motions, perhaps as much as a factor of six times present rates. Bill Compston asked whether Dewey felt it sensible to consider the concept of an "average" crust, from a geochemical standpoint, considering the complexities he described. Dewey replied that average crust would be quite difficult to determine, and that uniform, simple views of the crust are dangerous. Paul Weiblen commented that studies of Archean stratigraphy, albeit severely fragmented, might be valuable in determining tectonic environments. Dewey replied that Archean sediments offer the best hope in this regard.

Eric Nelson then discussed the effects of ridge-trench collision of crustal growth. In the southern Andes, the Chile rise is colliding with the Peru-Chile trench, producing numerous effects,

including emplacement of an ophiolite, intrusion of small silicic plutons, increased seismicity, uplift of Quaternary deposits, formation of major faults and hot springs, and establishment of a pull-apart basin on the continental shelf. Nelson speculated that ridge-trench collision may have been much more important during the Archean, because of faster spreading and longer ridge length then.

Gerry Schubert opened the discussion by asking Nelson if there was an age progression of volcanism in the southern Andes. Nelson replied negatively, but stated that much of the terrane south of the Chile margin triple junction is poorly dated. Al Kröner asked if the granitoid rocks Nelson described were S-type granites. Nelson replied that they were not, instead they resemble arc-type magmas. In a comment directed at Dewey, Urs Schärer remarked that the deformation in the Northern Himalayas may be different from the mid-Jurassic deformational event, and hence may not be related to the disappearance of the Tethys. Dewey acknowledged the evidence for earlier deformational events, but clarified that the deformed rocks he was referring to are Paleocene-Oligocene redbeds that occur all across the Tibetan Plateau. Dallas Abbott asked Nelson if there were any silicic plutons south of the Chile margin triple junction. Nelson replied that these have not yet been identified, although there are other ophiolites there that were emplaced into the forearc during northward migration of the triple junction. Al Kröner asked Nelson how he accounted for the unusual chemistry of the ophiolitic rocks he discussed. Nelson stated that among the two possibilities for ophiolite emplacement (obduction or emplacement into pull-apart basins in the forearc), the latter provided an opportunity for crustal contamination to explain the geochemistry. Kröner wondered whether the 3 – 4 Ma ophiolite age represented the time of emplacement or the time of generation, and Nelson replied that it represented the age of the volcanic rocks in the sequence. Nelson then suggested that we should consider the possibility that in the Archean, convergent plate margins were dominated by direct-slab melting, low-angle subduction, and lithospheric underplating. He felt that this might gradually evolve into the present-day situation, where convergent margins are dominated by hydrous melting of the mantle wedge. Steve Goldstein asked about the basis for direct-slab melting in the Archean. Nelson referred to a recent paper by H. Martin (Geology, 14, 753), who distinguishes trace element geochemistry of Archean versus younger tonalitic rocks. Roberta Rudnick pointed out that on this basis, the presence or absence of garnet as a residual phase could be distinguished. It was suggested that in the hotter Archean, melting of subducting slabs took place before deyhdration, leaving residual garnet, but that in the modern situation, colder downgoing slabs dehydrate before melting can occur.

The next talk in the session was given by Mike Thomas, who discussed the use of regional gravity studies in understanding Precambrian crustal growth in North America. He discussed paired gravity anomalies and how they could be interpreted as sutures even through the extensive cover of Phanerozoic platform sediments. The geometry of these anomalies can be used to estimate paleosubduction directions, and thereby, a large-scale continental growth pattern can be constructed. For North America, the results indicate outward growth from the Archean Superior and Wyoming Provinces.

In the discussion that followed, Lew Ashwal stated in a comment directed at John Dewey as well as Thomas, that Tibetan-style collision was unlikely to result in major anorthosite as a refractory residual constituent of the lower crust. Rather, anorthosites are magmatic cumulates that are nearly always emplaced in the shallow crust. Thomas replied that all that is required from the gravity data is that the lower crust have a relatively high density, and that there were several mechanisms to account for this. Dewey agreed that anorthosite was an unlikely constituent of the lower crust, but argued that many lines of evidence point to high lower crustal densities. Al Kröner commented that at a recent conference, Klaus Schulz presented convincing evidence for a mid-Proterozoic ophiolite in Minnesota. Urs Schärer asked Thomas whether the gravity anomaly at the Grenville Front represents a feature associated with the Grenville orogeny or a Mid-Proterozoic suture. Thomas replied that this feature certainly represents a Mid-Proterozoic event, possibly a 1650 Ma suture, but could not have been caused by a Grenvillian suture. Schärer wondered further if the Grenvillian orogeny may have enhanced the gravity anomaly. Thomas was noncommittal, but felt that Grenville effects would likely have modified the anomaly. Stephen Moorbath asked Thomas if he could make the important distinction between juvenile and reworked crust. Thomas replied that this was done on the basis of published geological and isotopic work rather than gravity signatures. Moorbath wondered how much of the North American continent has been covered in detail. Thomas stated that much more isotopic work needs to be done.

Nick Arndt then described his studies of the Trans-Hudson orogen of Saskachewan in which he and colleagues could distinguish isotopically an area of reworked Archean basement. Bill Collins commented that we must be careful not to overinterpret the sort of data Thomas presented. He was uncertain as to whether the cratons had actually formed as a result of the collisions now marked by sutures, or whether the sutures simply mark the sites where previously formed cratons have collided. Kröner pointed out that if an age progession could be documented, this would constitute evidence for cratonic growth by this mechanism. Paul Morgan commented that there are inherent biases in large-scale gravity interpretations. The two largest gravity anomalies in North America were not interpreted as sutures because they were known from previous work to represent something else. Dewey pointed out that these were not paired anomalies, but Morgan stated that the Snake River Plain anomaly was indeed paired, and the Mid-Continent Rift could be interpreted as a double-paired anomaly. Thomas admitted that other criteria in addition to the gravity anomalies were used to identify sutures. Simon Klemperer stated that some sutures, such as the southern boundary between the Superior and southern provinces do not yield paired gravity anomalies. Thomas agreed. Several participants engaged in a discussion of the construction of the Canadian Shield, in terms of progressive younging of accreted blocks.

Clive Neal then presented two combined talks, dealing with evidence for underplating of subducted oceanic crust in two different areas. Geochemical and isotopic data for eclogite xenoliths in kimberlites emplaced into the Kaapvaal Craton (South Africa) allow them to be divided into three types, the protoliths of which were mantle cumulates, spilitized oceanic basalts, and cumulate gabbro. An Sm-Nd "errorchron" of these eclogites gives an age of about 2.4 Ga. Neal argued from these data that the Kaapvaal Craton was underplated by subducted oceanic crust about 2.4 Ga ago, and that this process has continued through time, representing a major mechanism of crustal growth. Neal also interpreted clinopyroxene megacrysts and host alnöites from Malaita (Solomon Islands, South Pacific) in terms of a model involving assimilation of seawater altered basalt (SWAB) by primary alkali basalt under the Ontong-Java Plateau. The SWAB component may also represent underplating of subducted oceanic crust.

In the discussion, Jan Veizer asked Neal why he thought the carbon in diamond inclusions in the eclogites is biogenic in origin. Neal stated that their carbon isotopic compositions supported such a conclusion, but he did not know the actual values. Nick Arndt commented that subducted eclogite should be quite dense, and would not remain in the lithosphere underlying the Kaapvaal Craton. Neal agreed that this was a problem. Stu McCallum asked if the ilmenite and orthopyroxene required by Neal's fractionation model for the Malaitan alnoite were observed as megacrysts. Neal replied that there were bronzite, ilmenite, zircon, phlogopite, and garnet megacrysts, in addition to clinopyroxene. McCallum asked about the evidence for interpreting one of Neal's eclogite groups as being derived from oceanic gabbroic protoliths. Neal replied that these have positive Eu anomalies, requiring some plagioclase fractionation. Paul Toft asked if Neal had looked at the inclusion suite in the diamonds. Neal had not. Toft then commented that it would be difficult to subduct oceanic crust into relatively rigid depleted harzburgite lithosphere expected to be present below the graphite-diamond transition. Neal and colleagues consider that the oceanic crust in this case is subducted to the lithosphere-asthenosphere boundary. Bill McDonough asked if Neal had measured the isotopic compositions of megacrysts considered to be important in his fractionation model. Neal replied that he had measured the phlogopite and garnet megacrysts, and the results are consistent with his model. Several participants expressed concern that the Sm/Nd ratios of the "gabbroic" eclogite xenoliths were vastly different from most gabbroic rocks. Referring to Neal's 2.4 Ga eclogite "isochron," Al Kröner pointed out the lack of any accretionary event of that age in the Kaapvaal Craton. Roberta Rudnick asked if the large range in Na concentration among clinopyroxenes in the eclogites required them to have formed at a variety of depths. Neal replied that the rocks have been substantially modified since their formation, and that it is not possible to make pressure estimates based on these xenoliths.

Steve Goldstein asked Neal how the whole-rock Nd isotopic data points were determined for his 2.4 Ga eclogite "isochron." Neal evidently calculated them from mineral separate data and modal analyses. Goldstein and others wondered about the

uncertainties in this procedure, and Neal replied that they were "large." Stu McCallum asked if Neal's megacrysts could have been subjected to interaction with metasomatic fluid. Neal allowed this as a possibility, but stated that if such fluid came from the host kimberlite, then there would have been insufficient time for contamination.

The session continued with Roberta Rudnick's talk on the nature of the lower crust. She described lower crustal granulite xenolith suites, which, in general, sample more mafic lithologies than are exposed at the surface in major granulite terranes. This argues for basaltic underplating as a major crust-forming process, and tectonic underthrusting can account for the minor volumes of supracrustal rocks in xenolith suites. It is important to determine xenolith ages to properly assess crust-forming processes, but unfortunately this is difficult unless zircon-bearing rocks can be found. In the McBride Province of North Queensland (NE Australia), zircons in felsic xenoliths from <3 Ma alkali basalts range in age between 200–300 Ma, although a few have ages up to 1570 Ma. These ages correlate with events in the surface terrane, suggesting that surface exposures are reasonably representative of events throughout the crust. If lower crustal underplating is important, the lower crust may have a *younger* mean age then the upper crust, and our estimates of growth rates may be biased.

Francis Albarède asked Rudnick what evidence there was that the xenoliths were actually derived from the lower crust. Rudnick responded that all xenoliths show decompression features, indicating their rapid ascent from a hot environment. Furthermore, zircon ages of about 300 Ma combined with thermobarometry indicating pressures of about 10 kbar, require the xenoliths to have been in the lower crust at that time, since there has been no erosion in the area since then. Albarède did not seem convinced that broad-scale conclusions about the composition of the lower crust could be made on the basis of such few samples. Lew Ashwal then asked Rudnick to comment on the viability of Ross Taylor's andesite model for the average crust, considering the evidence she presented for major basaltic underplating. Rudnick replied that both she and Taylor would now favor a more basaltic bulk crust composition, at least during post-Archean times. Ashwal then asked Rudnick to comment on the mechanism by which the upper crust was produced. She favored major fractionation events. Ashwal then asked about the middle crust, which Rudnick had not discussed. Rudnick chose not to speculate. Kröner asked about the granitic melt inclusions in zircons, which he stated should represent an upper or middle rather than a lower crustal phenomenon. Rudnick replied that granitic melts were certainly being formed at about 300 Ma, as evidenced by the presence of felsic granulite xenoliths.

Steve Goldstein asked if Rudnick was able to quantify the amount of underplating. Rudnick stated that this would be almost impossible from xenolith studies, but perhaps a geophysical approach would be fruitful. Stephen Moorbath asked Rudnick to clarify the relationships between the various events she described. Simon Klemperer asked Rudnick to speculate on the nature of the lower crust prior to the

underplating events. She stated that all that could be said was that it had evolved isotopic characteristics, but this material is not represented in the xenolith suite. Klemperer then remarked that the xenolith suite is therefore biased. Rudnick suggested that the dominantly mafic Chudleigh suite is less representative of the lower crust than the McBride suite, which contains a wide variety of lithologies. Mordechai Stein then showed Sm-Nd isotopic data for similar mafic granulite xenoliths from Israel. Their ages are equivalent to upper crustal granites, and he suggested that that both were produced during an underplating event. Bill McDonough cautioned that independent evidence (i.e., geochemistry) must confirm that any given xenolith population is genetically related before isochron relationships can be attempted. Bill Compston asked if the 300 Ma underplating event inferred by Rudnick's data could account for the more felsic lithologies. Rudnick replied affirmatively, and stated that mafic underplating was an excellent heat source for crustal melting. Bill Collins asked Rudnick about the granitoids she mentioned. She stated that they were I-type calc-alkaline granitoids, interpreted as subduction-related magmatism. Chairman Kröner adjourned the session for lunch.

The session continued with Paul Weiblen's talk on the Mid-Proterozoic magmatic rocks of the Midcontinent Rift System (U.S.). Anorthositic rocks of the Duluth Complex show a variety of textural and compositional features consistent with their having been emplaced as mushes. The lack of complementary ultramafic cumulates suggests a model in which basaltic magmas were emplaced at depths of 20 km or more, where they fractionated mafic minerals that were left behind as plagioclase-rich mushes ascended to shallow depths. The associated Keweenawan lavas have chemical features suggesting a genetic relationship to the anorthositic suite. Weiblen suggests that the type of basaltic underplating proposed for this rift setting has been common throughout geologic time and represents a significant mechanism of crustal growth.

The last talk of the workshop was given by Dave Howell, who discussed the budget of Earth's oceanic sediment masses in terms of crustal growth and recycling. Based on estimates of the volume of oceanic sediments (91×10^6 km^3) and the average age of oceanic crust (55 Ma), Howell computes a continental denudation rate of 1.65 km^3/yr. This crudely balances estimated crustal production rates of about 1 km^3/yr, but the efficiency of sediment loss via subduction, for example, must be considered. Howell argued, on the basis of earthquake focal solutions, imagery of subduction zones, and plate kinematic reconstructions that little, if any, sediment was lost in this way. This yields a present-day crustal growth rate of about 1 km^3/yr. Howell then discussed the volume of continents to 1.5 Ga ago, assuming constant continental thickness and freeboard, and a constant hydrosphere volume. Howell concludes that ocean ridge length was a factor of about 1.75 greater 1.5 Ga ago, but a major uncertainty is the average spreading rate in the past.

In discussion, Howell was asked to comment on Gerry Schubert's consideration of freeboard, which requires at least some crustal recycling. Howell replied that constant freeboard could be accounted for during the past 2 Ga either by increasing

total ocean ridge length from 56,000 to 91,000 km or by increasing the average spreading rate from 5 cm/yr to 7 -8 cm/yr. Intuitively, one would expect faster spreading, considering the higher Precambrian heat flux, but this may be limited by smaller trench-pull forces due to the relatively younger and hotter oceanic crust presented to subduction zones then. Steve Goldstein asked Howell to comment on Paul Taylor's statements regarding the continents being about 0.5 km higher in the Archean. Howell replied that this was not required if ridge length of spreading rate could be higher then. Dallas Abbott disagreed, pointing out that this depends on how the Earth's heat loss was different in the past. She stated that for the pre-2.5 Ga Earth, in which continents may have been smaller and more numerous, freeboard arguments might not hold, unless the volume of the oceans was vastly different. Howell felt this could be accounted for with long ridge length and/or faster spreading, but Abbott stated that this would be inconsistent with models of Archean heat loss. Chairman John Tarney ended the discussion at this point.

GENERAL DISCUSSION

Lew Ashwal then chaired a general discussion session among the remaining workshop participants. He started by listing three possible areas of consensus reached at the workshop.

1. The Hurley-Rand crustal growth curve is probably displaced too far to the right (see cover figure).

2. Subduction of pelagic sediments has probably not been a significant process in recycling of the crust, and unless other recycling mechanisms exist, the Fyfe and Armstrong curves are unlikely. He noted, however, the consipicuous absence of these scientists from the workshop, and Steve Goldstein noted that this would not be a consensus if Francis Albarède had not already left the meeting.

3. There is a semantic problem in interpreting the starting point of crustal growth curves, related to the origin, identity, and fate of Earth's "primordial" crust.

Ashwal then listed several problem areas that could focus future research, prefacing his remarks with one person's suggestion for 10^5 additional Nd isotopic analyses, and his own feeling that more super-high resolution ion microprobes (SHRIMPs) were needed. Problem areas identified were as follows.

1. The start time for crustal growth. This relates to the quest for the holy grail of the oldest rocks on Earth. Can we expect to find rocks older than 3.8 Ga or zircons older than 4.2 Ga?

2. Recycling of Earth's primordial crust. If the Earth had such a crust, was it completely destroyed, or do we have some small hope of finding a remnant of it?

3. Recycling of Earth's continental crust after 3.8 Ga. If sediment subduction can reliably be eliminated, are there other recycling mechanisms?

4. Ancient heat loss and crust-forming mechanisms. Were they different from the present-day situation in the past?

5. Temperature of the Archean mantle. Was it different in the past?

6. Selective preservation. Is the Precambrian rock record representative?

Bill McDonough opened the discussion with a comment related to the start time of crustal generation. He suggested that there may be a story hidden in the chemistry of zircons (and their inclusions) older than 3.9 Ga that might tell us something about the primordial crust. Ashwal wondered about possible shock effects in these ancient zircons, which were produced at a time in Earth history when we would expect violent impacts to be occurring. Nick Arndt suggested that perhaps only the unshocked zircons survived. Bill Compston pointed out that lunar zircons are not obviously shocked. Ashwal stated that the lunar zircons come from multiply-shocked breccias, and that possibly the source rock for the Australian detrital zircons was similarly shocked. Compston stated that he was convinced that their probable sources were clasts of shocked rocks in other lithologies, and that possibly the paucity of intact rock sequences older than 3.8 Ga was caused by the tail-end of a heavy meteorite bombardment phase of Earth history. Paul Weiblen wondered if such a mechanism might be responsible for producing continental nucleii. Proterozoic sutures are identifiable, but for older terranes, some other crustal growth mechanism might be applicable. Bill Collins suggested that there may be mechanisms other than island arc accretion for assembling cratons, and that we should study in detail the tectonic, metamorphic, and structural history of individual blocks, relate this to the granitoid-forming processes, and compare these features between blocks in order to understand how cratons grow.

Ron Seeger returned the discussion to breccia, stating that this texture could be destroyed by metamorphic processes. Paul Warren agreed with this, and pointed out that lunar granulitic breccias are commonly coarse grained, and some are indistinguishable from so-called "pristine" rocks. Warren suggested that on the Earth, metamorphic effects might render brecciated rocks unrecognizable. Ashwal asked if this means we should not look for any record of the end of a heavy bombardment phase of Earth history, despite the fact that we know it likely occurred. Warren and others suggested that there might be a geochemical record of this. Ashwal asked why we do not see any siderophile element enrichments in ancient rocks. Stephen Moorbath said that when the Isua iron formation was first discovered, it was thought to be an iron meteorite. Warren pointed out that terrestrial impactites do not generally show enrichments in siderophiles— these elements are thought to be dispersed into the stratosphere. Paul Weiblen commented that petrographers working on Archean rocks might, in general, overlook subtle shock effects. Moorbath said that it might be worthwhile to search for impact effects in some Archean volcano-sedimentary terranes, but admitted that the chances of success would be slim. Paul Weiblen stated that the terrestrial counterpart of the lunar regolith should look very different from the products of mass wasting processes. Clive Neal suggested that we look at gold-bearing breccias in South Africa for impact effects. Jan Veizer raised the semantic question of how this primordial crust relates to growth of continental crust as we familiar with from the existing rock record. Moorbath stated that it was important not to confuse this with "continental" crust. Veizer then stated that any primordial crust or regolith would be continuously churned about in sedimentary cycles,

and although some material would be lost, most would remain. If felsic, zircon-rich crust had once been present, we should, therefore, expect to see some trace of it remaining. John Tarney wondered if this implies that the earliest terrestrial crust was basaltic. Veizer said that possibly it was. Paul Toft suggested that if the oldest crust was indeed basaltic, then perhaps we should be looking at eclogite xenoliths from kimberlites. These may have been carried to great depths by subduction or other processes and are known to contain ancient diamonds. Dallas Abbott agreed that old rocks might have been deeply buried, citing the example from modern subduction zones. Paul Taylor commented that if basaltic underplating is an important process, as many workshop participants have shown, then this produces the opposite effect, and younger rocks should be expected to be found in the deeper parts of the crust. John Tarney wondered whether underaccretion was more important than overaccretion as a crustal growth mechanism. He argued that in Archean terranes, hydrous tonalitic-trondhjemitic magmas were emplaced at the base of the crust, and that if this process continued, the rocks exposed at the surface would all have been through a high-grade metamorphism. Nick Arndt pointed out that this interpretation relies on a number of assumptions, including depth of generation, water content, and tectonic environment.

Simon Klemperer then gave a short impromptu presentation about the results of deep seismic reflection profiling, which shows that the lower crust is prominently layered, in many continental areas, regardless of the age of the surface rocks. The seismic Moho is commonly shallower than the petrological Moho, leading to the question of the nature and origin of this prominent reflector in the deep crust. The lower crust is much less well defined in Phanerozoic and Proterozoic accreted terranes, suggesting possible differences in types of lower crusts. Bill Collins confirmed that in Australia, Archean and Post-Archean terranes have very different lower crustal seismic signatures.

Jan Veizer then commented that even if a pre-3.8 Ga sedimentary mass was recycled very efficiently, it would still compose about 0.1% of the present sedimentary rock record. Nick Arndt then showed a viewgraph of a crustal growth curve (see cover figure) illustrating the concept of episodic or punctuated crustal growth. This is based on abundant isotopic data in support of a major period of juvenile crust generation at about 1.8 Ga in North America and elsewhere. Tarney pointed out that in the interval between 2.6 and 1.8 Ga, a huge volume of mafic dikes was produced, but little crust was generated. He wondered what that meant about the heat loss mechanisms during that period. Moorbath argued that crust was being produced in other areas at that time. Arndt stated that isotopic data for well-studied terranes does not support this, but possibly much crust of that age exists in Asia or elsewhere. Goldstein pointed out that the growth curve Arndt showed was based only on data for northern Europe and North America. Steve Galer then stated that periodicity in crustal growth should be reflected as deflections in Nd data for the depleting mantle curve. Maureen Wilks argued that as more isotopic work is done, the gaps between apparent periods of high crustal growth are filled in, as seems to be the case in North America. She stated further

that continental growth was more-or-less continuous, but occurred in different areas at different times. Galer suggested that the most effective way of dealing with this problem is to investigate the integrated effects of crustal growth on the mantle. Arndt argued that our isotopic data base for the mantle is insufficient to resolve whether the crust grew episodically or continuously. Bill Collins then offered support for Arndt's view, stating that fundamentally different granitoid types were produced at different stages in Earth history. Several participants suggested that K-rich granitoids represent the products of remelting of older, K-poor granitoids. Collins argued that Nd data clearly demonstrate that the bulk of 1.8 Ga granitoids represent juvenile crust. Paul Taylor pointed out that K-rich granitoids were being produced as far back as we can see geologically. Moorbath added that tonalites and trondjhemites also have been forming over the full range of geologic time, and no differences could be detected between ones of different ages. Tarney disagreed, stating that chemical differences did indeed exist. Moorbath said the proportion of K-rich granites was higher in younger times, but this was due to a recycling effect.

Steve Goldstein raised the issue of the sedimentary mass Nd data. He feels that if sediment recycling is not a viable process, then it is difficult to produce a sedimentary mass that is not older than the continental crust. Given a mean sedimentary mass age of 2.0 Ga, combined with Rudnick's evidence for major basaltic underplating, then the mean age of exposed continental crust should be much younger than 2.1 –2.2 Ga. Paul Morgan raised a question about interpretations of sedimentary Nd data. He accepts that the isotopes cannot be recycled, but wonders if somehow these could be decoupled from other components of the sedimentary mass which are recycled. Steve Galer pointed out that the constant of the Nd/Si ratio in the sedimentary rock record precludes major recycling of sediments into the mantle. Jan Veizer commented that there must have been an evolution of the cyclicity of processes, in terms or types and rates. He argued that pulses such as CADS must exist, but their frequency may have been different at different periods of Earth history. Dave Howell wondered if crustal growth has been constant, and the effect of sedimentary overturn without recycling was more rapid in the past, whether this would not produce a mean sedimentary mass age of 2.0 Ga. Goldstein stated that a uniform growth curve would lie to the right of most currently proposed curves. He further argued that if underplating was a major process, with progressively younger rocks occurring at deeper levels, then the Ga mean age of the crust must be significantly younger than the 2.0 Ga value derived from sediments.

Simon Klemperer encouraged the geochemists to construct more concrete models that could be tested geophysically, for example, with seismic or gravity measurements. Stephen Moorbath pointed out, quite correctly, that it is difficult to get geophysicists to attend meetings such as this. Lew Ashwal pointed out that it was difficult enough for geochemists to sit through four days of geochemical presentations. Paul Toft suggested that one fruitful area of future research might be to examine more closely the residual mantle material, which is the ultimate parent material for the crust. Dave Howell stated

that the wind-blown material used as crustal averages do, in fact, represent a statistically valid sampling of the crust. Even if there has been an important underplating effect in crustal growth, these regions would be included in such samples, considering the rapid uplift that takes place in certain areas. Bill Collins disagreed, stating that most granulite terranes consist of supracrustal rocks that have been buried and later uplifted. He wondered if perhaps the lower crust resists being exposed at the surface due to its high density. Howell stated that in many places, mantle peridotites could be sampled, so that lower crust should also be available at the Earth's surface. Simon Klemperer voiced support for Collins' remark, stating that to his knowledge, there are no exposures corresponding to the layered lower crust he sees seismically. Roberta Rudnick expressed doubt that any exposed granulite terrane is representative of true lower crust. These may have once been at or near lower crustal depths, but may not have resided there for very long. Howell pointed out that enormous amounts of underplating would be required for lower crustal rocks to be exposed at the surface. In any case, deep crust is exposed by detachment faulting rather than underplating. The final comment was made by Steve Goldstein, who pointed out that the Nd evidence from mantle-derived rocks as old as 3.8 Ga, which have positive epsilon Nd values and therefore imply a substantial older crustal reservoir, is apparently at odds with the 2.0 Ga mean age of crust derived from sediments. Lew Ashwal then thanked all of the participants, and adjourned the workshop at 3:45 p.m.

ABSTRACTS

CONSTRAINTS ON CONTINENTAL ACCRETION FROM SEDIMENTATION

Dallas Abbott, Lamont-Doherty Geol. Obs., Palisades, NY 10964

Portions of the present day continents are flooded and have marine sedimentation. Most of these flooded regions are near the ocean-continent boundary and are of grossly similar widths on different continents. Thus, smaller continents have a higher proportion of flooded area as a proportion of their total surface area (Fig 1). A recent compilation of total continental surface area versus percentage of flooding demonstrates this (1).

We have developed a model of percentage of continental flooding which utilizes round continents and a constant width of the zone of flooding. This model produces a reasonably good fit to the percentage of flooding on the present day continents (Fig. 2). All present day continents with a total surface area of less than 6.27×10^6 km^2 are 99-100% flooded (1). It follows that, given reasonably constant freeboard (2,3), the small Archaean continental fragments were nearly 100% flooded. We can use this fact to develop models of the percentage of flooding of the early continents. These models in turn, predict the deposition of certain proportions of marine/nonmarine sediment types on the early continents.

Our model makes several assumptions. The first is that global ridge length is directly proportional to the average (over continental megacycles) of the number of continental plates. Thus if global ridge length is doubled, this doubles the probability that a given continental fragment will be rifted into two fragments. The second assumption is that the total heat production of the earth and global ridge length are related in some simple way. For the purposes of this first model, we assume that global ridge length is approximately equal to global trench length (as is true at present), but other ratios of ridge length to trench length are possible (4).

The model results, using the continental accretion model of (3), predict a large change in the percentage of flooded continental crust, from 100% at 3.8-4.5 b.y. to 62% at 2.8 b.y. to 31% at the present time (Fig. 3). These predicted changes are consistent with observed changes in sedimentation and extrusive volcanism through geologic time (5)(Fig. 4). Continental extrusives did not become abundant until ~ 2.6 b.y. Evaporites did not become abundant until ~2.3 b.y. Conversely, secondary quartzites are very common in the early rock record, as might be expected for continents with a very high percentage of coastal surface area.

These secular changes in sedimentation and extrusive volcanism could thus be related to greater size of individual continental blocks. The rock types which become more abundant later in Earth history may require a larger size of continental mass. This hypothesis could be tested by looking at the relative proportions of sedimentary rock types on Phanerozoic continents as a function of continental size. If the proportion of sediment types is different on different size continents, this approach could be extended to Proterozoic and Archaean sediments.

Our model further predicts the average size of continental blocks at each time in early history (Fig 3). If the distribution of plate sizes is Poissonian, this model can also be used to predict the maximum possible size of entire plates and thus of continents. This provides a method of testing different possible ratios of ridge length to trench length, by comparing the predicted maximum craton size and the craton size distribution to the observed sizes of (reconstructed) Archaean cratons.

REFERENCES (1). Cogley, J.G., Continental margins and the extent and number of continents, Rev. Geophys. Space Phys., 22, 101-122, 1984. (2). Abbott, D.H., Archaean plate tectonics revisited 2. Paleo-sealevel changes, continental area, oceanic heat loss and the area-age distribution of the ocean basins, Tectonics, 3, 709-722, 1984. (3). Reymer, A., and G.S. Schubert, Phanerozoic addition rates to the continental crust and crustal growth, Tectonics, 3, 63-77, 1984.

Abbott, D.H.

(4). Hargraves, R.B., Faster spreading or greater ridge length in the Archaean?, Geology, 14, 750-752, 1986. (5). Ronov, A.B., Probable changes in the composition of seawater during the course of geological time, Sedimentology, 10, 25-43, 1968.

Figure 1. Percentage of continental flooding calculated for idealized, round continents

SUBMERGED

EMERGENT

100 % FLOODED   **66 % FLOODED**
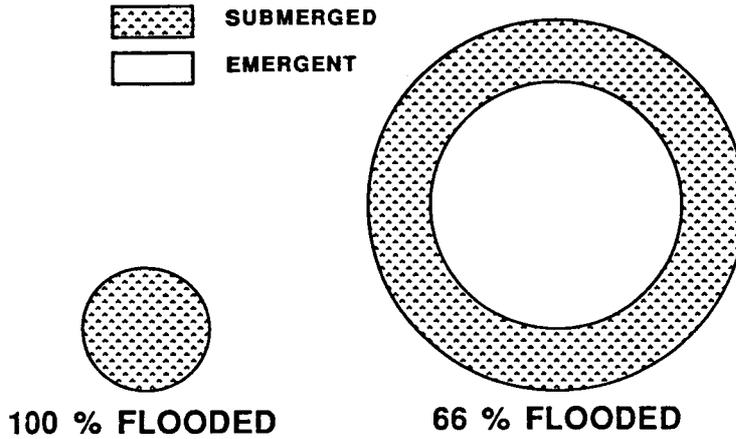
RADIAL WIDTH OF SUBMERGED ZONE
IS THE SAME FOR BOTH CONTINENTS

Figure 2. Squares: percentage of continental flooding of the present day continents from (1). Line with dots: Model calculations of the percentage of flooding.

CONTINENTAL FLOODING MODEL

New Zealand

Antarctica

Cen. Am.

FLOODING,%

SQRT AREA
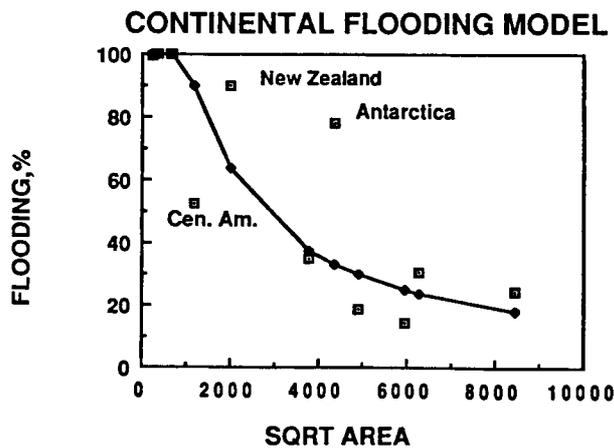
with characteristic times relative to Nd exchange between either mantle and crust (τ) or between depleted and primitive mantle (θ) being labelled with *in* or *out* for input and output respectively. Additional mass balance relationships are:

$$\tau_c^{out} = \tau_{dm}^{in} \cdot f / (1 - f) \qquad \tau_c^{in} = \tau_{dm}^{out} \cdot f / (1 - f)$$

$$\theta_{dm}^{in} = \theta_{pm}^{out} \cdot c \, (1 - f) / (f - c) \qquad \qquad (3)$$

where f is the ratio $Nd_c / (Nd_c + Nd_{dm})$ and c the ratio $Nd_c / (Nd_c + Nd_{dm} + Nd_{pm})$. The rate of crustal growth may be appreciated through the residence time τ_c of Nd in the continental crust which is such as:

$$(\tau_c)^{-1} = (\tau_c^{in})^{-1} - (\tau_c^{out})^{-1}$$

with:

$$(\tau_c^{in})^{-1} = \lambda \, \frac{x'_c - x_c}{y_{dm} - y_c} \qquad \qquad (5)$$

$$(\tau_c^{out})^{-1} = \frac{(1 - f)}{f} \left[\lambda \, \frac{x_{dm} - x'_{dm}}{y_{dm} - y_c} - \frac{f}{\theta_{dm}^{in}} \right] \qquad \qquad (6)$$

In Equations 5 and 6, the prime refers to the apparent Sm/Nd ratio inferred from the secular evolution curve of Nd isotopes and the values $x'_{dm} = 0.21$ and $x'_c = 0.17$ were computed in [6].

The principal unknown of the problem is the rate of mantle differentiation. If one assumes that all the mantle below the 670 km discontinuity is still pristine, the time constant θ_{pm}^{out} of the differentiation process should not be too far from 11 Ga. Zindler and Hart [7] calculates the present-day fraction c of primitive Nd stored in the crust as being 0.18, whereas a mean f value of 0.45 seems a reasonable estimate [6]. Altogether, these values suggest that θ_{dm}^{in} is about 4 Ga, which permits further evaluation of the time constants of crustal processes.

The crustal residence time τ_c of Nd is about 4-10 Ga and dominated by the τ_c^{in} term, whence crustal growth is inferred to be far from steady state. For a modern continental volume of 7.6×10^9 km^3, the present-day addition of mantle-derived material to the crust takes place at a rate of 2.0 ± 0.4 km^3/a, whereas the return flow of sediment to the mantle is loosely constrained but with a best estimate in the range 0.6-1.5 km^3/a. The uncertainties on these figures are currently under investigation. Their extrapolation back in time is also much uncertain. However, this simple model turns out to provide some indication of sediment entrainment at subduction zone: if the amount of oceanic sediments were to be a steady quantity on the scale of their mean travel time on the seafloor (60 Ma), 500 m of sediment riding over 3.5 km^2 arrive yearly at subduction zones. As much as 2/3 of these 1.75 km^3/a could be transferred

irreversibly to the mantle. Whether sediment recycling at a rate roughly equivalent to 3 permil of the depleted-mantle volume per Ga should be unambiguously detected through the Ce/Pb ratios [5] or lead isotopes [4] is still unclear.

Much of the model output depends on the average isotopic and elemental properties of the depleted mantle but particularly of the continental crust. The importance of the lower crust in the Sm/Nd mass balance is probably crucial but, up to now, insufficiently documented.

References: [1] DePaolo (1983) Geophys. Res. Letters 10, 705-708 [2] Reymer and Schubert (1984) Tectonics 3, 63-77 [3] Patchett and Chauvel (1984) Geophys. Res. Letters 11, 151-153 [4] Galer and O'Nions (1985) Nature 316, 778-782 [5] Newsom et al. (1986) Earth Planet. Sci. Letters 80, 299-313 [6] Albarède and Brouxel (1987) Earth Planet. Sci. Letters 82, 25-35 [7] Zindler and Hart (1986) Ann. Rev. Earth Planet. Sci. 14, 493-570.

Figure 1
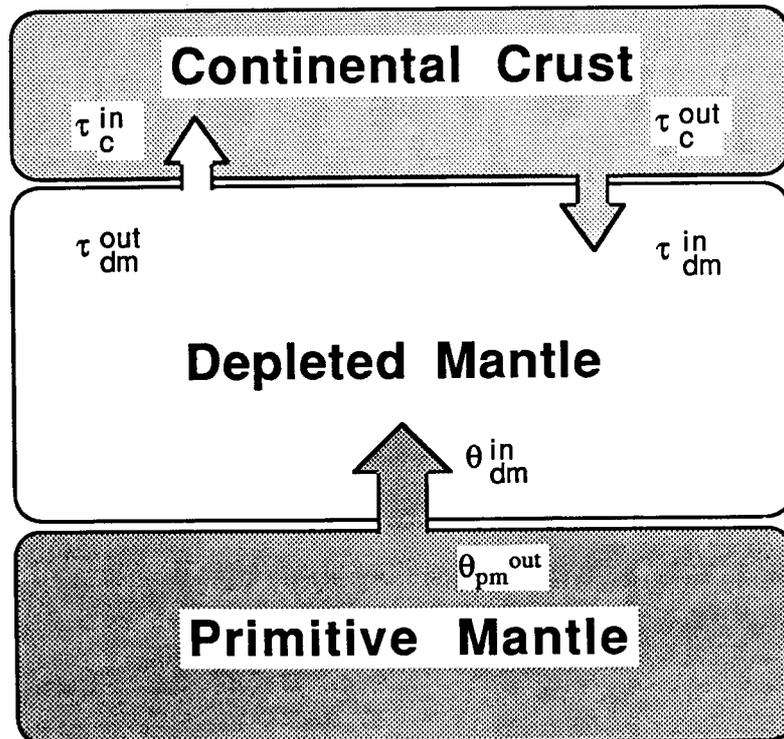
RECYCLING OF LOWER CONTINENTAL CRUST THROUGH FOUNDERING OF CUMULATES FROM CONTAMINATED MAFIC INTRUSIONS

Nicholas T. Arndt and Steven L. Goldstein (Max-Planck-Institut für Chemie, Postfach 3060, D-6500 Mainz, West Germany)

Recent studies (1) have suggested that the composition of depleted mantle has been buffered throughout Earth history by addition of enriched material. Sediments and other rocks from the upper crust can form only a small proportion of this addition because their Pb isotopic compositions (2) and Nb/U, Ce/Pb (3) and Lu/Hf ratios (4) are inappropriate. Attention has therefore focussed on entrainment of material from undepleted lower mantle (5,6). Recycling of lower crust has been considered unlikely because no generally acceptable mechanism has been proposed to return this material to the mantle. Below I outline a mechanism that involves entrapment of picritic magmas at the crust-mantle interface, contamination of these magmas with crustal rocks, crystallization of the magmas to form Fe-rich, relatively dense ultramafic cumulates, and return of these cumulates to the mantle.

Almost all stable cratons have been the site of eruption or intrusion of voluminous mafic magmas, either in the form of continental flood basalts or as large sills and dyke swarms. These magmas are relatively Fe-rich and therefore too evolved to have been equilibrium with the mantle material, and various authors (e.g. 7,8) have proposed that the parental magmas were picritic. Cox (7) suggests that relatively dense picrite was trapped at the base of the crust and differentiated there to yield less-dense evolved magmas (that erupt as basalts), and olivine, pyroxene and plagioclase cumulates. The volume of cumulate material plus that of magma trapped in the crust probably greatly exceeds that of erupted magma.

Southern Africa has been the site of continental volcanism and intrusion of large mafic-ultramafic complexes for the past 3 Ga. The total thickness of erupted lavas in the Pongola, Witwatersrand, Ventersdorp, Transvaal and Karoo sequences is between 5 and 10 km, and at least as much material must have been trapped in or at the base of the crust. Addition of at least 10-20 km of crust is indicated. Another estimate comes from McKenzie (9), who calculated that during the past 1.5 Ga, magma input to the base of the crust from hot-spot activity should amount to 75% of continental crust thickness. Present crustal thickness in southern Africa is about 34 km (10), and it probably was similar 3 Ga ago at the time of Pongola eruption (11,12), which raises the question of what happened to the added >20 km of mafic material. Possible explanations include crustal extension to eliminate over-thickening, or intrusion of magmas only in regions of previous crustal stretching and thinning. But a third possibility also exists; namely that the added mafic material was removed, in part from the top by erosion but mainly from the base through the following process. Olivine and pyroxene that crystallize from

trapped picritic liquids are more Fe-rich (Mg# = 0.8 to 0.6; (7)), and therefore more dense (3.4 to 3.6 g cm^{-3}), than the same minerals in mantle harzburgite (Mg# = 0.9 and p = 3.2 g cm^{-3}). When ol + px cumulates cool to ambient temperatures they become denser than the underlying mantle. Plagioclase-bearing cumulates are more bouyant. It follows that the lower, denser ultramafic cumulate portions of trapped sub-crustal sills might become detached from the crust and sink into the underlying lithosphere. We thus have a possible differentiation mechanism through which basalt and perhaps anorthositic cumulates are added to the crust and ultramafic cumulates are returned to the mantle.

However, not only ultramafic material is returned, because mafic magmas intruded into the crust commonly become contaminated with partially melted wall rocks (13). Ultramafic cumulates formed in a contaminated intrusion would contain crustal material: (a) as magma trapped between cumulate grains (up to 30% without eliminating negative bouyancy), and (b) as rafts of crustal material between cumulate sheets, and (c) as a constitutent of the cumulus minerals. Initially the Fe-rich cumulates would build up a layer immediately below the Moho, but this gravitationally unstable situation of dense cumulates overlying less-dense residual harzburgite could eventually trigger delamination of the lithosphere (14) and return of foundered material to the underlying convecting mantle.

A important aspect of this model is the composition the recycled material which is quite different from that of upper crust or sediments: (a) the contaminant is from the lower crust and has relatively low U/Pb and Rb/Sr but high Th/U, Ce/Pb, Nd/Pb and perhaps Nb/U; and (b) elements more compatible with the cumulus minerals would be concentrated in the returning material (e.g. clinopyroxene has more Nd than Pb: $D_{Nd}^{cpx-liq}=0.1$; $D_{Pb}^{cpx-liq}=0.01$ (15,16)), and ratios such as Nd/Pb or Ce/Pb would be further elevated.

This process might explain several hitherto enigmatic aspects of crust-mantle evolution. These include: (a) a mechanism for producing relatively felsic continental crust from mantle-derived magmas; (b) a means of deriving the characteristic Ce/Pb and Nb/U of MORB and OIB, which are higher than in primitive mantle or upper continental crust (3); (c) a source of the material needed to buffer the composition of depleted mantle (1), which must have non-radiogenic Nd and Hf, Sr more radiogenic than MORB, $^{207}Pb/^{206}Pb$ lower and $^{208}Pb/^{206}Pb$ higher than in upper continental crust; and (d) a source of mantle heterogenieties with chemical and isotopic characteristics similar to those of certain OIB, such as Kerguelen, Tristan and Gough, which have low $^{143}Nd/^{144}$, $^{87}Sr/^{86}Sr$ higher than MORB, moderate $^{206}Pb/^{204}Pb$ but high $^{207}Pb/^{206}Pb$ and $^{208}Pb/^{206}Pb$ (17).

References: (1) DePaolo D.J. (1983) Geophys. Res. Lett. 10, p.705-708; (2) Galer S.J.G, Goldstein, S.L. and O'Nions, R.K. (1986) Lunar Planet. Sci. XVII, p.247-248; (3) Hofmann A.W., Jochum K.P., Seufert M. and White W.M. (1986) Earth Planet. Sci. Lett. 79, p.33-45; (4) Patchett P.J. and Chauvel C.(1984) Geophys. Res. Lett. 11, p.151-153; (6) Galer S.J.G. and O'Nions R.K. (1985) Nature 316, 778-782; (7) Cox K.G. (1980) J. Petrol. 21, p.629-650; (8) Morse S.A. (1982) Amer. Mineral. 67, p.1087-1100; (9) McKenzie D. (1984) Nature 307, p.616-617; (10) Gane P.G., Atkins A.R., Sellschop J.P.F. and Seligman P. (1956) Bull. Seismo Soc. Amer. 46 p.293-316; (11) England P. and Bickle M.J. (1984) J. Geol. 92, p.353-367; (12) Nisbet E.G. (1984) Can. J. Earth Sci. 21, p.1426-1441; (13) Manhes G, Allegre C.J., Dupre B, and Hamelin B. (1980) Earth Planet. Sci. Lett. 47, p.370-382; (14) McKenzie D. and O'Nions R.K. (1983) Nature 301, p.229-231; (15) Ray G.L., Shimizu R. and Hart S.R. (1983) Geochim. Cosmoschim. Acta 47, p.2131-2140; (16) Watson E.B. Othman, D.B. Luck, J.M. and Hofmann A.W. (1987) Chem. Geol. (in press);(17) White W.M. (1985) Geology 13, p.115-118.

Growth of Continental Crust: clues from Nd isotopes and Nb-Th relationships in mantle-derived magmas

Arndt, N.T., Chauvel, C., Jochum, K.-P., Gruau, G. and Hofmann, A.W.
Max-Planck-Institut fuer Chemie, Postfach 3060, D-6500 Mainz, West Germany.

Nd isotopic data and Nb-Th relationships in Archean and modern komatiites and basalts provide information about the composition and evolution of the upper mantle, and so doing constrain interpretations of the timing of continental growth.

Nd isotopes: Komatiites and basalts from the 3.4 Ga Barberton greenstone belt, South Africa, display a narrow range of ϵNd values that cluster around zero, but slightly older (3.5 Ga) volcanics from the Pilbara Block, Western Australia, have a range from 0 to ±5. In 2.7 Ga areas (Abitibi Belt, Canada; Belingwe Belt, Zimbabwe; Kambalda area, Australia) display a wider range of ϵNd values from -2 to +8. The lower values probably come from magma contaminated with older continental crust; the higher values are three isolated analyses from the Kambalda areas. If these are excluded, we can conclude that the mantle source of these rocks had ϵNd from +1 to +5. The picture provided is of a mantle that was (a) depleted in parts earlier than 3.5 Ga ago, and (b) markedly heterogeneous in terms of Nd isotopic compositions. If this depletion can be equated with continent formation, a record is provided of continent growth that started before 3.5 Ga and continued through the Archean.

Nb/Th: Hofmann et al. (1) have shown that Nb/U is modern MORB and OIB is remarkably constant (~ 50), higher than that of primitive mantle (~ 30) and lower than that of continental crust (~ 10). Hofmann et al. proposed that these values resulted from a sequence of events: (a) growth of continental crust leaving the upper mantle variably enriched in Nb, (b) homogenization of the mantle, and (c) separation of MORB and OIB reservoirs. Isotopic compositions of the oceanic basalts indicate that their sources segregated more than ~ 1.7 Ga ago, and it was proposed that the homogenization event took place soon after major continent growth at the end of the Archean.

Uranium is a mobile element and Nb/U values in altered komatiites are unreliable, but useful information comes from Nb-Th-La relationships. Fig. 1 shows that Archean volcanics (black dots) plot distinctly below the modern volcanic trend and close the primitive mantle value. However, this result cannot be interpreted unequivocally as evidence of a more primitive Archean mantle because two Tertiary komatiites and a basalt from Gorgona Island, Columbia (black squares) plot together with the Archean rocks: The low Nb/Th of both groups of rocks may stem from a peculiarity of komatiite formation rather than a feature of one of the Archean upper mantle. One possibility is that Gorgona komatiites formed by high-degree melting of an OIB-type source. If so, we are left with two possible interpretations of the Archean data: (a) they also come from an OIB type source and magmas from N-MORB type Archean mantle are not present in greenstone belt sample suites. This mantle is likely more depleted than that which gave the greenstone volcanics: inferences on continent growth based on greenstone belt Nd data would therefore have to be revised; (b) the Archean volcanic source was indeed primitive, supporting Hofmann et al. ideas about mantle evolution and continent growth. Available data do not enable us to decide which interpretation is correct.

References: (1) Hofmann, A.W., Jochum, K.P., Seufert, M. and White, W.M. (1986). Earth Planet. Sci. Lett. 79, p.33-45.
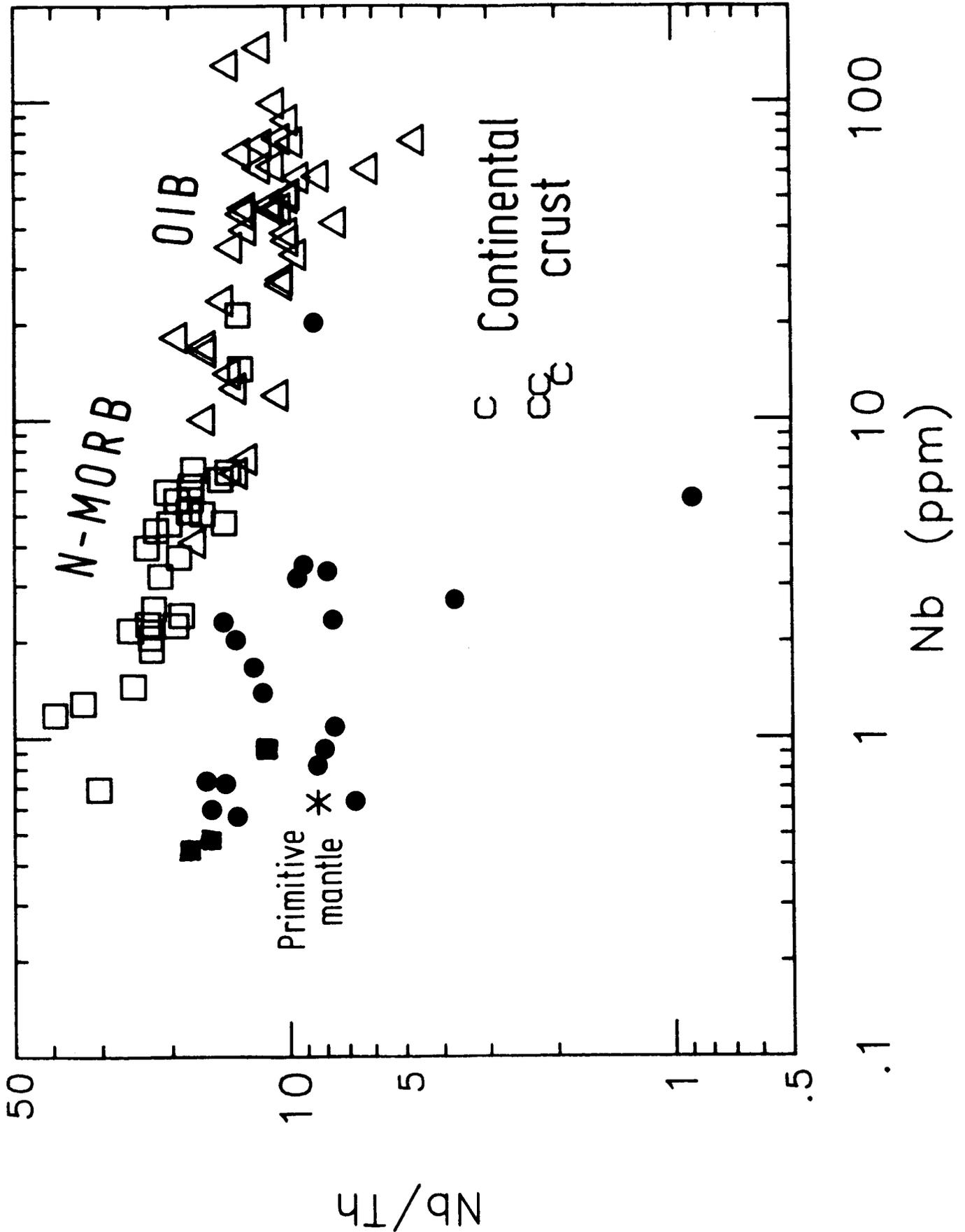
PLUTON EMPLACEMENT AND MAGMATIC ARC CONSTRUCTION: A MODEL FROM THE PATAGONIAN BATHOLITH

Robert Bruce, Eric Nelson, & Stephen Weaver, Colorado School of Mines
Dept. of Geology & Geological Engineering, Golden, CO 80401

Field relations, geochronology, and Sr isotope data from the Patagonian batholith are integrated into an arc construction model that relates crustal growth at convergent margins to pluton emplacement and differential uplift. The Patagonian batholith in the southern Chilean Andes represents the eroded roots of a subduction-related magmatic arc developed along the Andean convergent margin between late Jurassic and Tertiary time. Crystallization ages range from 157-15 Ma and record semi-continuous magmatic activity during this period (Fig. 1). In three areas studied in detail (generally <10,000 sq. km) a progression of mafic to felsic rocks is consistently observed: excluding dikes and contaminated peraluminous rocks, mafic plutons (norite, gabbro, hornblende diorite) intruded first and were followed by progressively more felsic plutons (quartz diorite, tonalite, granodiorite, and granite). This progression is confirmed by Ar/Ar, Rb/Sr, and U-Pb geochronology in three study areas. Different areas have distinct ranges in age. Such relationships can be interpreted to represent diachronous magmatism along the arc. However, a pattern of decreasing initial $^{87}Sr/^{86}Sr$ with time for samples from three widely distributed areas (Fig. 2) suggests that a long-term crustal evolution occurred throughout the magmatic arc, and that magmas of all the observed types (mafic to felsic) were generated and emplaced through the recorded history of the batholith. As an alternative to diachronous magmatism, we present a model of arc-crustal growth that relates the age-lithology relationship to pluton emplacement depth and uplift history.

The model assumes semi-continuous uplift of middle to upper crustal levels in the thermally- weakened arc, resulting from protracted intrusion of low-density magmas at these levels. Within an uplifting crustal block, deeper plutons will take longer to reach the surface than shallow plutons. Thus, as exposed today in a coherent block, older plutons must have been intruded at deeper levels than younger plutons. Furthermore, because the batholith shows a consistent age-lithology relationship (old mafic to young felsic progression), it can be concluded that the mafic plutons were intruded at a deeper level than felsic plutons. The model thus suggests that older (mafic) plutons were uplifted to the intrusion depth of younger (felsic) plutons.

The apparent age of a batholith, or of individual areas within a batholith, is to a large degree controlled by the uplift history of the arc. The model has several corollaries. 1) During the period of recorded magmatism in the area, time-integrated uplift rate is inversely

proportional to the range in ages. Thus, an area with a wide range in ages experienced a relatively slow uplift rate during the recorded magmatic period. 2) Total uplift since the earliest recorded magmatism is inversely proportional to the age of the oldest pluton. Thus an area with older plutonism will have undergone less uplift compared to an area with relatively younger plutonism. These corollaries hold only when comparing areas with comparable lithologic variation.

Similar age-lithology relationships have been reported from batholiths in Peru, California (Sierra Nevada), Idaho, and SE Alaska, suggesting that our model may be applicable in other magmatic arcs. However, the age-lithology patterns produced by the proposed processes could be obscured by other geologic events (e.g., structural or metamorphic overprints). Also, data must be available from a large area and covering a large time span for the time-integrated effects of the processes to be observed in the age-lithology patterns.
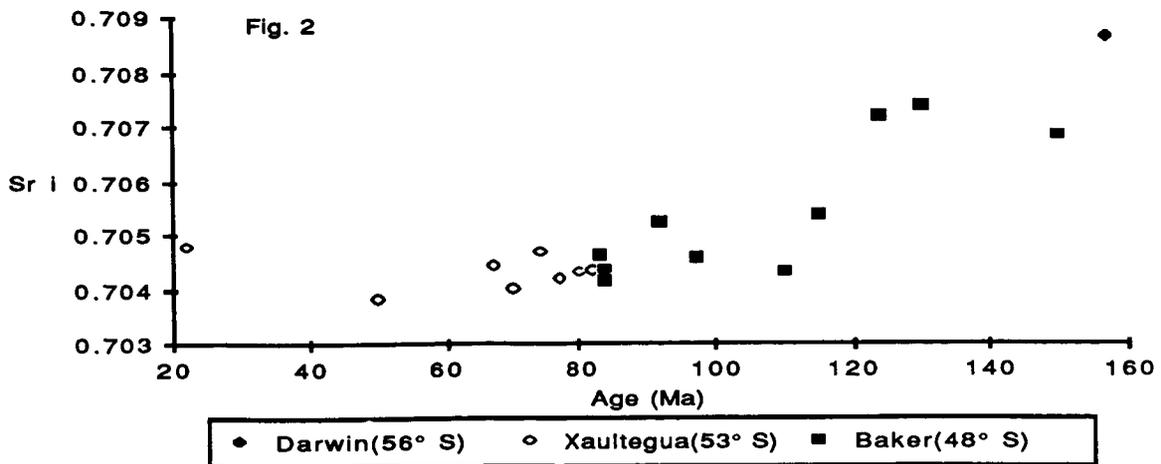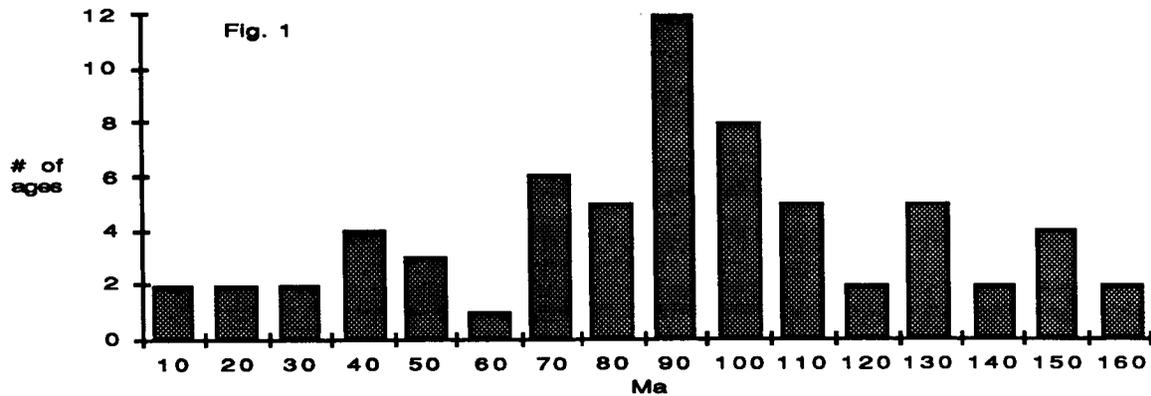
N88-30207

AN ILLUSTRATION OF THE COMPLEXITY OF CONTINENT FORMATION

Kevin Burke, Lunar & Planetary Institute, 3303 Nasa Road 1, Houston, TX 77058

Although a great variety of processes are involved in making and modifying continents, the formation of island arcs on the ocean floor is widely regarded as a common way of initiating continental growth (e.g.,1). Growth of the Antillean island arc was initiated in the Pacific (apparently by nucleation on fracture zones) at about 130 Ma (fig.1) and the processes by which that arc has evolved to its present state have been worked out in rather more detail than has proved possible for older island arcs, such as those represented within the Greenstone-granodiorite terrains of the Canadian Superior Province (2,3). It seems probable that the processes involved in incorporating arc material into continents are unlikely to be less complex than those that are affecting the Antillean arc.

Polarity of subduction along the Antillean arc reversed after the arc had collided with an oceanic plateau. As that plateau left the Pacific and entered the Atlantic, parts of the arc collided with both North and South America and marginal basins developed within the arc forming the Yucatan and Grenada basins. Continuing motion of the Caribbean Plate eastward with respect to North and South America is being accommodated by substantial strike-slip motion and the formation of a third marginal basin in the Cayman trough. Nine fragments, all exceeding a few thousand square kilometers in area and juxtaposed by strike-slip motion of up to several hundred kilometers have been identified in the Greater Antillean sector of the arc. Parts of the Antillean arc abutting both North and South America have been thrust on top of those continents, but the Aves-Swell-Lesser Antillean sector of the arc still lies within the ocean.

Without the detailed controls of plate kinematics and high resolution (>5Ma) Mesozoic and Cenozoic stratigraphy, it would not have been possible to establish as much as has been about the Antillean arc. Because of the absence of this kind of information for ancient times, it seems likely that only in exceptional cases will it be possible to make detailed statements about how continental evolution has progressed in ancient terrains. In general, we can expect to recognize only rock-types (e.g. granodiorites, MORB basalts, volcaniclastics), gross structural relations (e.g. intrusive cont acts, thrust faults) and isotopic characteristics as guides to ancient continental evolution.

REFERENCES
1. Dewey, J.F. and B.F. Windley, Growth and differentiation of the continental crust. Phil.Trans.R.Lond. 301, 189-206, 1981
2. Pindell, J.L., Dewey, J.F., Cande, S.C., Pitman, W.C. III, Rowley, D.B. and LaBrecque, J. Plate Kinematic Framework for Models of Caribbean Evolution. In Press. Mesozoic and Cenozoic Plate Reconstructions, Tectonophysics.
3. Card, K. D. Tectonic Setting and Evolution of Late Archean Greenstone Belts of Superior Province, Canada, pp. 74-76. Tectonic Evolution of Greenstone Belts. LPI Technical Report 86-10.

K. Burke
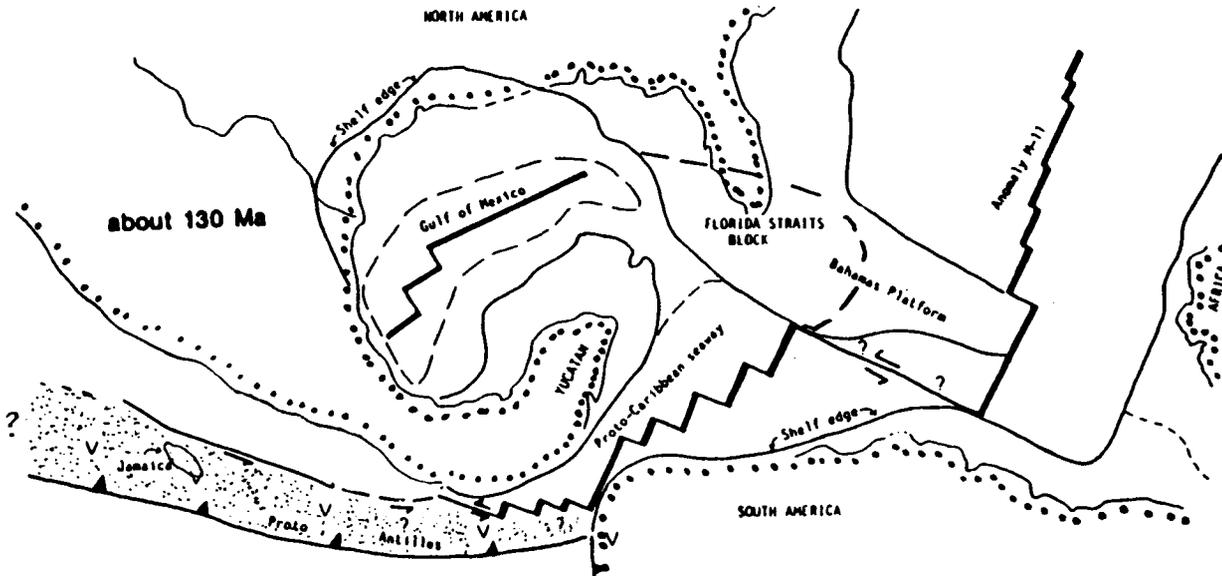
Figure (1) (Modified from Ref. 2). The Antillean arc formed near the beginning of the Cretaceous in the Pacific Ocean. Subduction was at that time toward the Americas.
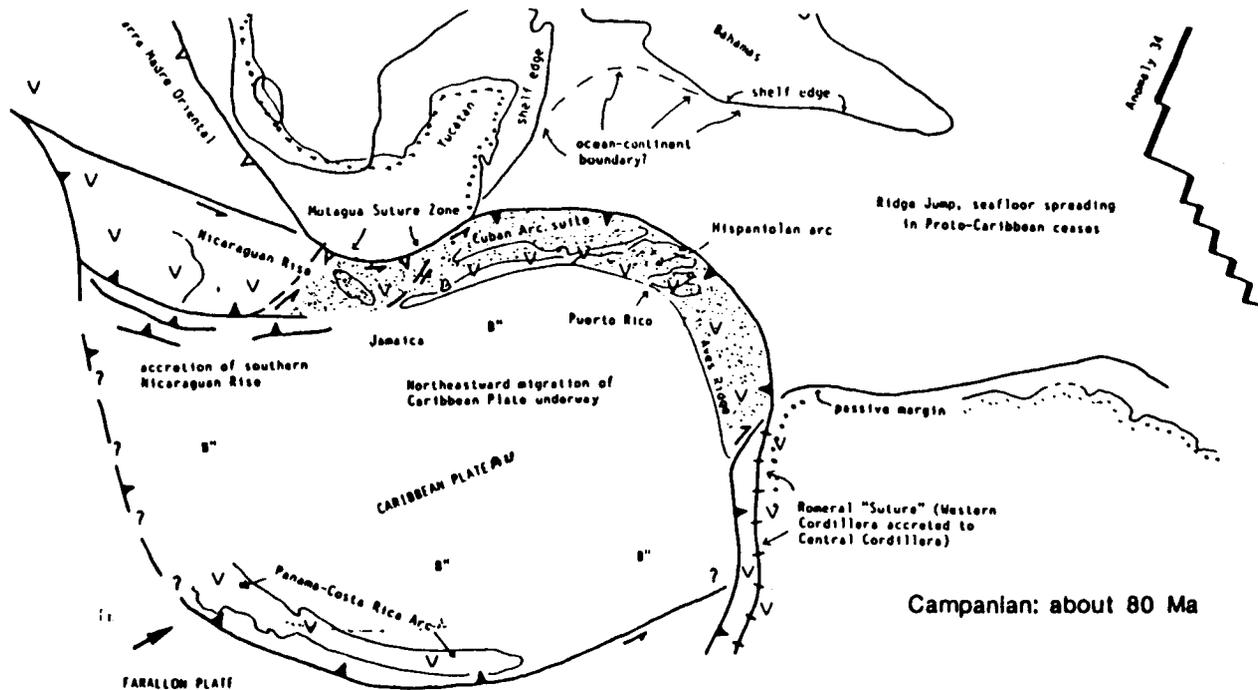
Figure (2) (Modified from Ref. 2). The Antillean Arc collided with an oceanic plateau at about 80 Ma and the polarity of subduction reversed. As the plateau entered the Atlantic collisions of the Arc with North and South America began.
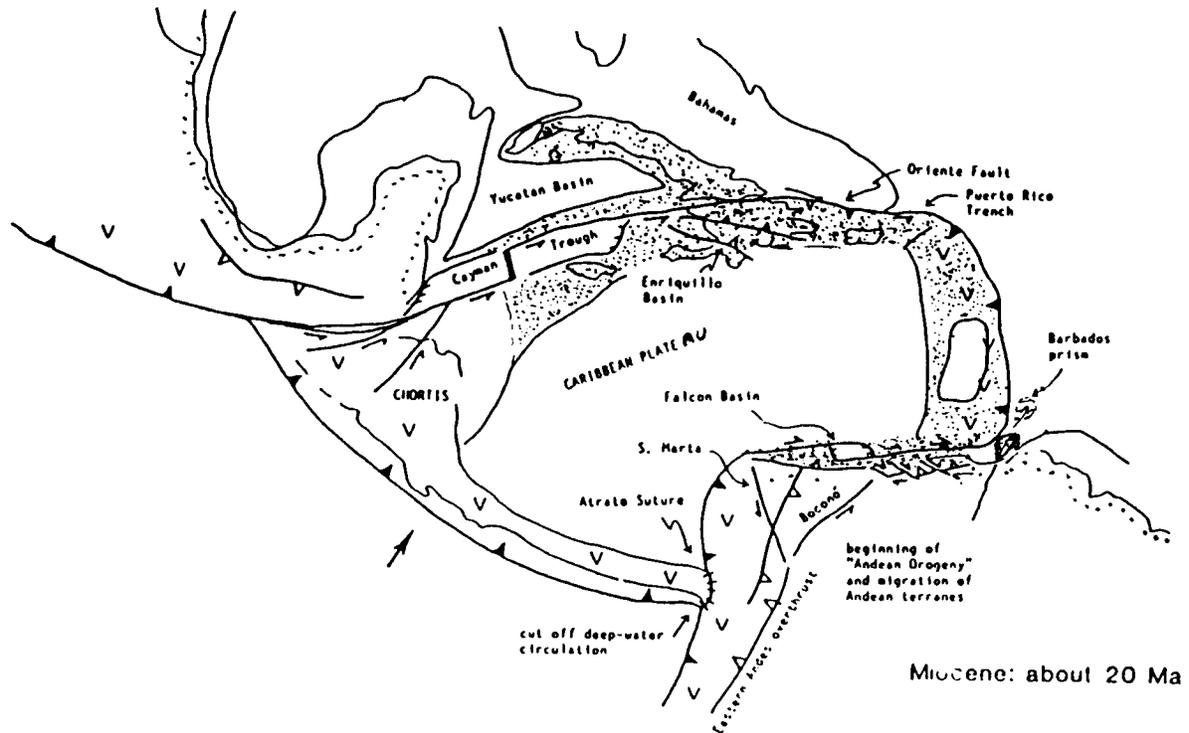
Figure (3) (Modified from Ref. 2). By 20 Ma the Antillean Arc was broken into three fragments. The Greater Antillean fragment, adjacent to North America was sliced into pieces by strike-slip faults and contained two marginal basins, the Yucatan and Cayman troughs. The Venezuelan arc fragment, adjacent to South America was similarly sliced-up by strike-slip faults. These parts of the arc had been thrust respectively onto North and South America. The Lesser Antilles-Aves Swell fragment on the eastern margin of the Caribbean alone remained active. It had been split by the Grenada trough marginal basin and was about to attempt to subduct the Barbados prism, a submarine fan of the then Orinoco River delta.

N88-30208
THE KERALA KHONDALITE BELT OF SOUTHERN INDIA:
AN ENSIALIC MOBILE BELT

Thomas Chacko, University of North Carolina, Chapel Hill, NC 27514
(present address: University of Chicago, Chicago, IL 60637);
G.R. Ravindra Kumar, Centre for Earth Science Studies, Trivandrum, India
695 010; James K. Meen and John J.W. Rogers, University of North
Carolina, Chapel Hill, NC 27514.

The Proterozoic (?) Kerala Khondalite Belt (KKB) is a large expanse of high-grade supracrustal rocks which makes up the southernmost segment of the vast southern Indian granulite terrain. The KKB is bounded on the south by the Nagercoil massive charnockite (1) and on the north by the charnockite massifs of the Cardamon Hills. Similar rock types and field relations have been described by Cooray (2) in the South-Western Belt of Sri Lanka and it is likely that this region is a southward extension of the KKB.

The major lithologies of the KKB, in decreasing order of abundance, are: 1) garnet-biotite±orthopyroxene±graphite gneisses (which includes a Na-rich and a K-rich group); 2) khondalites (graphite-sillimanite-garnet-biotite±cordierite gneisses); 3) cordierite gneisses (garnet-biotite-cordierite±orthopyroxene gneisses); along with lesser amounts of calc-silicates, basic granulites and quartzites. These lithologies are interlayered both on an outcrop scale and as mappable units on a larger scale, which strongly suggests that they represent a sequence of supracrustal rocks that was subsequently metamorphosed to the granulite facies.

The protoliths of the khondalites (sensu stricto) and cordierite gneisses are likely to have been pelitic and semi-pelitic sediments. Similarly, the quartzites and calc-silicates are also almost certainly of metasedimentary origin; possibly representing sandstones and calcareous sandstones, respectively. The basic granulites which are basaltic in composition may have been conformable lava flows within the supracrustal sequence.

The protolith of garnet-biotite gneisses is more problematic but several lines of evidence suggest that the gneisses are of metasedimentary rather than of metavolcanic origin: 1) almost all are

peraluminous; 2) many of the gneisses are graphite-bearing; and 3) zircons from these lithologies show a large degree of rounding. It is suggested, therefore, that the garnet-biotite gneisses were originally immature sediments. Compared to typical graywackes, sodic gneisses have lower ferromagnesian contents, and potassic gneisses have higher K_2O/Na_2O ratios. The geochemical features of the gneisses are more characteristic of arkoses than graywackes (3).

REE patterns for the KKB metasediments are typically LREE enriched with significant negative europium anomalies, indicating a "granitic" source region. Similarly, low Ni contents and low MgO/FeO and Ni/V ratios also suggest a relatively sialic provenance. Cr contents, on the other hand, are higher than those of most post-Archean fine-grained sedimentary rocks (4), suggesting either a secondary enrichment process or a Cr-enriched sialic source. A source for the KKB sediments may have been the charnockite massifs (or their premetamorphic equivalent) which bound the supracrustal belt. The massifs are predominantly sialic but have relatively high Cr contents.

It is possible to interpret the arkose-pelite lithologic association of the KKB sediments in terms of either foreland basin or cratonic rift basin models. However, the occurrence of rocks of basaltic composition within the sequence suggests that the rift model is more appropriate. Subsequent closure of the basin, along with the associated deformation and granulite-facies metamorphism (5, 6, 7, 8), was accomplished without the intrusion of significant volumes of calc-alkaline igneous rocks. Thus the orogen was largely ensialic in character and probably did not involve the subduction of large volumes of oceanic crust. Many of the features of the KKB are accounted for by the ensialic mobile belt model of Kröner (9). This model suggests that many early Proterozoic mobile belts were formed by aborted continental rifting followed by intracrustal compression.

Rogers (10) concluded that the Dravidian Shield of southern India, of which the KKB is a part, was a coherent crustal block by 2,500 m.y. Although extensive geochronologic information is not yet available, it is reasonably clear that the KKB is a post-Archean supracrustal sequence which was metamorphosed sometime after 2,100 m.y. (6, 11). The present study suggests that the KKB was formed by intracrustal differentiation

within the Dravidian Shield rather than by the addition of new crust or the accretion of exotic crust.

References

1. Narayanaswamy, S. and Lakshmi, P. (1967) Jour. Geol. Soc. India, v. 8, p. 38-50.

2. Cooray, P.G. (1962) Quart. Jour. Geol. Soc. London, v. 118, p. 239-266.

3. Pettijohn, F.J. (1963) U.S.G.S. Prof. Pap. 440-S, 21 pp.

4. Taylor, S.R. and McLennan, S.M. (1985) The Continental Crust: Its Composition and Evolution, Blackwell, 312 pp.

5. Ravindra Kumar, G.R., Srikantappa, C. and Hansen, E.C. (1985) Nature, v. 313, p. 207-209.

6. Srikantappa, C., Raith, M. and Speiring, B. (1985) Jour. Geol. Soc. India, v. 26, p. 849-872.

7. Ravindra Kumar, G.R. and Chacko, T (1986) Jour. Geol. Soc. India., v. 28, p. 277-288.

8. Chacko, T., Ravindra Kumar, G.R. and Newton, R.C. (1987) Jour. Geol., in press.

9. Kröner, A. (1981) in Precambrian Plate Tectonics, Elsevier, p. 57-83.

10. Rogers, J.J.W. (1986) Jour. Geol., v. 94, p. 129-144.

11. Chacko, T. (1987) Unpubl. Ph.D. thesis, University of North Carolina, 191 pp.

THE TIMING AND MECHANISM OF DEPLETION IN LEWISIAN GRANULITES.

A.S. Cohen, Department of Earth Sciences, University of Cambridge, Cambridge CB2 3EQ, U.K.; R.K. O'Nions, Department of Earth Sciences, University of Cambridge, Cambridge CB2 3EQ, U.K.: M.J. O'Hara, Department of Geology, University College of Wales, Aberystwyth SY23 3DB, U.K.

It is well documented that some granulite facies rocks show severe depletion of heat-producing and related LIL elements, and this has given rise to much debate concerning the mechanism responsible for the depletion. The presence of CO_2-rich fluid inclusions frequently found in granulites has led some authors to suggest that depletion is the result of pervasive flooding by externally derived CO_2-rich fluids, while others have appealed to dehydration and partial melting as the principal mechanism for the removal of the heat-producing elements.

Granulite facies rocks which are well preserved in the Lewisian of N.W. Scotland can offer some insight into this problem. Moorbath et al. (1) in a classic paper demonstrated that much of the complex, predominantly 'grey gneiss' of broadly tonalitic composition, had suffered severe U and Th depletion early in its history. Chapman and Moorbath (2) showed that U, Th loss and/or Pb homogenisation had ceased 2.68 Ga ago, while Hamilton et al. (3) demonstrated that protolith formation may have occurred as much as 2.92±0.05 Ga ago.

A detailed isotopic study has been made of an ultrabasic-basic sequence sampled across an outcrop width of 3 metres. The whole-rock Pb isotope results, which include the analysis of an acid gneiss collected some 10 metres from the ultrabasic sample, show that little or no depletion occurred in those lithologies with the highest solidus temperatures (the ultrabasics). This is in marked contrast to the severe depletion observed in the basic and acid gneisses, and suggests strongly that partial melting may have been the principal control on U, Th movement and on the overall depletion of the Scourian complex.

Sm-Nd and Pb-Pb mineral isochrons from a peridotite, a transition gneiss and a basic gneiss are highly precise and internally consistent, and show that the granulite facies mineralogy was not stabilised until $\geq$ 250 Ma after the cessation of U, Th and Pb movement at 2.68 Ga. These data also suggest that they record final mineral crystallisation and not isotopic cooling ages.

In summary, the following conclusions are drawn;

1. The absence of depletion in ultrabasic lithologies with the highest solidus temperatures suggests that loss of U and Th occurred principally as a result of partial melting.

2. The granulite facies mineralogy now preserved was established $\geq$ 250 Ma after cessation of U, Th and Pb movement.

3. Any surviving fluid inclusions in these minerals post-date depletion by $\geq$ 250 Ma and thus will not represent the fluid phase present at the time of depletion. In other terrains it will be necessary to demonstrate consanguinity of fluid inclusion formation and depletion in order to appeal to a genetic connection.

4. Geothermometers based on cation exchange between granulite minerals do not in the case of the Scourian define the temperature at which depletion occurred.

REFERENCES: (1) S. Moorbath et al. (1969) EPSL 6 245-256. (2) H.J. Chapman and S. Moorbath (1977) Nature 268 41-42. (3) P.J. Hamilton et al. (1979) Nature 277 25-28.

GENESIS OF ARCHEAN GRANITOIDS IN THE PILBARA BLOCK AND IMPLICATIONS FOR CRUSTAL GROWTH.

W.J. Collins, School of Earth Sciences, Macquarie University, NSW 2109 Australia

MA 756999

The Mount Edgar Batholith in the Archean granite-greenstone terrane of the Pilbara Block, Western Australia, is a large ovoid domed structure that is surrounded by the lowermost stratigraphic units of the greenstone succession, the Pilbara Supergroup. The batholith is composite, consisting of variably deformed granite plutons which can be grouped geochemically into seven major suites. Excluding the younger, "tin-granite" of the Moolyella Suite, all plutons were emplaced at ~3300 Ma. These plutons intruded the greenstone belt, dated at 3450-3550 Ma (1,2), and a gneiss complex, comprising 45% of the batholith. Field and isotopic evidence from the complex indicate that 3300 Ma orthogneisses have formed virtually in situ from an older ~3450 Ma banded gneiss sequence that, lithologically, closely resembles coeval amphibolite-grade felsic volcanics of the greenstone belt. Metamorphism of the felsic volcanics, which can be traced into type Duffer Formation, is a result of late-stage diapirism that juxtaposed the gneiss complex and greenstone belt. The gneiss complex resembles other Archean high-grade gneiss belts and represents the deep-level equivalent of the Pilbara granite-greenstone terrane. This is consistent with structural and seismic evidence which indicates that the large batholith domes of the Pilbara, comprising 60% surface area, merge at shallow depths (3) and felsic-intermediate composition rocks extend into the lower crust (4,5). In contrast, post-Archean lower crust is typically mafic (6,7) and in modern magmatic arcs, granitoid comprises less than 40% areal extent.

Geochemically, granitoids of the Mount Edgar Batholith are much more evolved than modern, Phanerozoic granitoids. With two exceptions from two hundred, all samples contain over 65% SiO_2 and Na_2O contents <u>decrease</u> from 5wt% with increasing SiO_2, whilst K_2O rapidly increases. The granitoids lack restitic mineral phases, do not contain mafic microgranitoid xenoliths (enclaves) and are not associated with gabbroic complexes; they are considered to have crystallised from melts approximating the average bulk composition of the batholith (70.6% SiO_2). These silicic melts cannot be derived from fractional crystallisation, mixing or assimilation of basaltic magmas, nor is their derivation by partial melting of basalts or andesites likely. Partial melting of tonalitic/dacitic precursors is the preferred mechanism based on published experimental data, mass balance calculations and trace element modelling.

Other batholiths of the Pilbara Block are geochemically similar to Mount Edgar and probably were derived from similar tonalitic/dacitic sources. These sources are considered to have formed at 3450-3500 Ma by 30% partial melting of thick (>60 km) basaltic crust during underplating by hot mafic to ultramafic (komatiitic) magmas, producing a granitoid dominated crust. Founding of the eclogitic residue, refractory from the partial melt event, and release of mantle-derived fluids 150-200 Ma after granitoid-crust formation resulted in recycling of the crust, forming the abundant trondhjemite-granodiorite batholiths of the Pilbara Block and leaving 20% residue of amphibolite in the lower crust.

Collins, W.J.

50

Recycling of a tonalitic/dacitic dominated crust accounts for the evolved silicic nature of Pilbara granitoids, their lack of any mafic linneages, the abundance of granitoid relative to younger terranes and provides an internally consistent model to remove the hypothesised eclogitic substrate under the Pilbara crust (e.g. 8). Crustal evolution from primordial oceanic material to highly evolved, thick relatively stable continental crust encompassed no more than 200 Ma and was a natural consequence of high heat flow in the Archean.

References:
(1) Pidgeon R.T. (1978) Earth Planet Sci. Letts. 37, 421-428.
(2) Hamilton et al. (1981) Spec. Publ. Geol. Soc. Aust. 7, 187-192.
(3) Hickman A.H. (1981) Spec. Publ. Geol. Soc. Aust. 7, 57-70
(4) Drummond B.J. (1981) BMR Journ. Geol. Geophys. 6, 123-135.
(5) Drummond B.J. (1982) Rev. Brasil. Geo. Sci. 12, 113-120.
(6) Griffin W.L. and O'Reilly S.Y. (1987) in Mantle Xenoliths P.H. Nixon Ed. (in press).
(7) Drummond B.J. and Collins C.D.N. (1986) Earth Planet Sci Letts. 79, 361-372.
(8) Sun S.S. (1984) In Archaean Geochemistry A. Kroner et al. Eds pp 25-47.

ROLE OF ZIRCON IN TRACING CRUSTAL GROWTH AND RECYCLING

W. Compston, I.S. Williams, R.A. Armstrong[*], J.C. Claoué-Long, P.D. Kinny, J.J. Foster, *Res.School Earth Sci., A.N.U.,* A. Kröner, *J.Gutenberg Univ. Mainz,* R.T. Pidgeon, *Curtin Inst. Tech. W.A.,* J.S. Myers, *Geological Survey W.A.* *also B.P.I. Univ. Witwatersrand S. Africa.*

Precise geochronology: The correct emplacement age of an igneous rock is a simple, direct measure of the age of crust formation, as well as a prerequisite to the use of Nd and Sr isotopic compositions of bulk rock samples for protolith modelling. It is registered more reliably and more precisely by concordant or near-concordant U-Pb ages in single zircon crystals than by any other material (Table1).

Table 1: **Comparison of Isukasia ages.**

		(Ma)
Schist, volc.	(Rb-Sr)	3660±60
BIF	(Pb-Pb)	3710±70
Garben., volc.	(Sm-Nd)	3750±40
Single zircons	(therm. ion.)	3769±10
Bulk zircons	(therm. ion.)	3813±18
Single zircons	(ion probe)	3807±2

It is a mistake to rely on Sm-Nd whole-rock isochrons, which can be mixing lines (Table 2), and bulk samples of altered rocks are open to later movement of Rb and Sr too often for confidence. Multigrain zircon samples can be unreliable also if they are composite in origin.

Table 2: **Kambalda greenstone ages** (Ma)

Metabasalt	(Rb-Sr)	2610±30
Interflow sediment	(Zircon) [1]	≤2692±4
Later granite	(Zircon) [2]	>2662±6
Metabasalt	(Pb-Pb)	2720±105
		2726±78
Metabasalt,u'mafics	(Sm-Nd)	3262±44
		3230±120

Precise geochronology is an essential prerequisite for tectonic reconstructions of crustal growth. Zircon age studies of volcanics and adjacent gneisses in South Africa have recently clarified the long-standing problem of their relative ages (Table 3).

Table 3: Kaapvaal single zircon ages.
Hooggenoeg volcanics, BGB [3] 3444±8Ma
Theespruit Fmn.,BGB [3]   <3536±6 Ma
3433±4<Ancient Gneiss Complex[4]<3644±4

Xenocrysts as material tracers: The ages of zircon xenocrysts can be used to trace magma sources. 'Accidental' xenocrysts provide samples of hidden older crust, while 'restitic' zircons (when identified) constitute direct, visible relics of the magma protolith.

Their use as material tracers is less dependent on geochemical modelling than use of bulk-rock isotopic compositions. Sr and Nd isotopes trace earlier 'continental crust' defined as having high Rb/Sr and low Sm/Nd, i.e. *felsic upper crust* that is already well-evolved geochemically through intracrust processing of primitive mantle-derived magmas. On the other hand, magmatic zircons occur in several types of mafic rock and associated felsic differentiates, so that xenocrysts originating from such rocks register their former existence and accurately specify their ages in favourable cases.
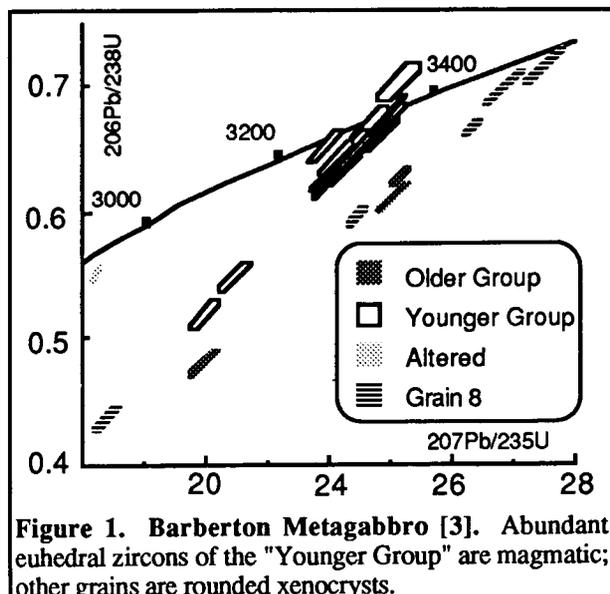

Figure 1. Barberton Metagabbro [3]. Abundant euhedral zircons of the "Younger Group" are magmatic; other grains are rounded xenocrysts.

Xenocryst identification: Xenocrysts must be distinguished from magmatic or later zircons, which can only be done efficiently by intragrain dating by ion microprobe. Bulk samples of zircons frequently obscure composite sources through averaging and they can generate misleading Pb loss chords (Fig. 1).

Zircon morphology, internal structure and the distribution of zircon ages within populations are also relevant to identifying xenocrysts (Fig. 2), but the separation of zircon types becomes progressively more difficult as original magmatic plutons become transformed into gneisses and one or more metamorphic zircon components are introduced.
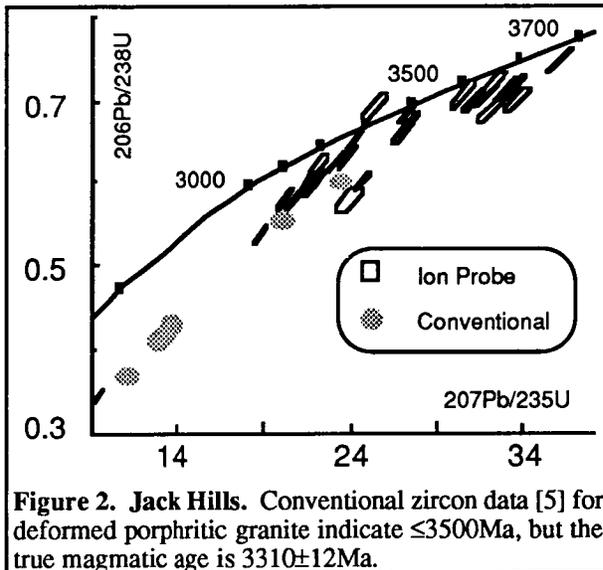
Figure 2. Jack Hills. Conventional zircon data [5] for deformed porphritic granite indicate ≤3500Ma, but the true magmatic age is 3310±12Ma.

Zircon solubility in magmas using the approach of Watson and Harrison is a valuable pointer to zircon inheritence if original magma temperatures can be estimated and provided equilibrium was approached. Magmatic zircons are expected to form the principal population in a magma originally undersaturated in zircon .

Felsic rocks or mafic rocks? The U and Th contents of zircons do not uniquely identify their sources as felsic or mafic. Extremes of both U contents and Th/U can be found in either type (Table 4). Arguments can be made in support of both felsic and mafic source rocks for the oldest-known terrestrial zircons (4.1 to 4.3 Ga) but proof positive is lacking.

Table 4. **U contents of zircons in mafic rocks**

	U/ppm
Kimberlites	3-235
Anorthosite (Adirondacks)	33-400
Granophyre (Ercall) [6]	60-1890
Andesite	87-7800
Gabbro (Ontario)	580-880
Norite (Sudbury)	590-2270
Granophyre (Tasmania) [7]	2000-7000

Preservation of original ages. The age record of zircons will be largely lost if recrystallization occurs, but instances are known of localities within grain interiors having much older ages than the rest of the grain with no visible boundary between old and young. Recrystallization proceeds as a kinetic front, with the unrecrystallized zircon completely preserving its original age, in sharp contrast to that of the immediately adjacent recrystallized zircons [8]. No diffusion profiles have been observed, only linear mixing lines generated by spatial overlap of the sampled area between original and recrystallized zircon during analysis.

Field evidence in disconnected exposures of Archaean gneiss in West Australia and Greenland sometimes fails to predict the measured zircon ages, signifying unsuspected complexities in field interpretation, or far more frequent and complete recrystallization of zircon than studies of high-grade metamorphic rocks have indicated. Hydrothermal alteration may grow a dominant population of new zircon, and may be the principal mechanism of age loss in zircon.

The oldest terrestrial rock. Work on early Archaean gneisses and gabbroic anorthosites in Western Australia has not yet detected the source of the 4.1 to 4.3 Ga zircons found as detrital grains in Archaean quartzites, nor found any xenocrysts of such age in igneous plutons. New work is being directed at the distribution of these oldest zircons within the supracrustal succession, and within clasts of older quartzites from the Jack Hills conglomerate.

1. Claoué-Long *et al.* Econ. Geol. *(submitted).*
2. Compston *et al.* Ceochim. Cosmochim.

Acta *(submitted)*.
3. Armstrong *et al.* E.P.S.L. *(submitted)*.
4. Kröner *et al.* E.P.S.L. *(submitted)*.
5. Pidgeon *et al.* Aust J Earth Sci. *(submitted)*.
6. Compston *et al.* J. Geol. Soc. Lond. (submitted).
7. Hergt *et al.* (in prep.).
8. Black *et al.* Cont. Min. Pet. (1986).

N88-30212

CHRONOLOGY OF EARLY LUNAR CRUST

E.J. Dasch, NRC/NASA JSC, Houston, TX 77058 and Oregon State University, Corvallis, OR 97331; L.E. Nyquist, SN4/NASA JSC, Houston, TX 77058; and G. Ryder, Lunar and Planetary Institute, 3303 NASA Road One, Houston, TX 77058.

Although the time of formation of the early lunar rocks is critical to an understanding of their petrogenesis, few well documented samples of pristine plutonic lunar rocks (or PPLRs) have been dated unequivocally (1). Isotopic information on several of these rocks and their minerals do not form systematic arrays on isochron or other graphs. Commonly, the more aberrant data are excluded from the final regressions or other calculations, but the selection criteria, though perhaps reasonable for a given rock, is not standardized or systematic. The most common problem for many Rb/Sr and Sm/Nd internal (mineral) or whole rock ages is significant departure of one or several points from the best-fit isochrons. In addition, the several dating techniques have, in some cases, resulted in markedly different "ages" for the same rock or even the same sample.

Nyquist (1) compiled a histogram of the available age data on PPLRs. (The system for recognizing PPLRs, devised by Morgan et al. (2), Warren and Wasson (3), and Norman and Ryder (4), among other workers, is based on low siderophile abundances, cumulate mineralogy and textures, and, in some cases, on "primitive" chemical characteristics such as unevolved trace element patterns (that is, no KREEP component). This select group of rocks constitutes a very reduced population of lunar samples.) These ages, all >3.9 Ga (26 numbers) were determined by $^{39}Ar/^{40}Ar$, Sm/Nd, and Rb/Sr techniques. Of special interest are the oldest apparent ages of crystallization. Of the three techniques, the oldest ages as well as the oldest average ages were Rb/Sr ages. Three of the six Rb/Sr ages plot near 4.5 Ga. The seven Sm/Nd ages range from 4.2-4.5 Ga, averaging 4.3 Ga. $^{39}Ar/^{40}Ar$ ages are distinctly younger, with a peak at 3.9 Ga, but with three ages of 4.4 Ga, and some evidence of older events in the higher temperature releases of some of the more complex spectra. The $^{39}Ar/^{40}Ar$ ages mainly can be explained by at least partial resetting of crystallization ages by impacting, especially by basin-creating events such as Imbrium. There is no apparent explanation as to why Rb/Sr ages should be greater than Sm/Nd ages, in some cases on the same sample.

AGES OF MAJOR LUNAR ROCK TYPES

Anorthosites: The very low measured $^{87}Sr/^{86}Sr$ ratios of selected anorthosites (e.g. 5) leads most workers to believe that they are very early lunar differentiates, perhaps from a magma ocean (even though $^{39}Ar/^{40}Ar$ dates are apparently younger). Recently, Lugmair (6; pers. comm.) has reported a Sm/Nd mineral isochron age of 4.44 + 0.02 Ga for ferroan anorthosite 60025. The date is important not only for its precision but also for its significantly "young" age, relative to the age of the moon. The age implies that, if the global magma ocean hypothesis is valid, the early anorthositic crust did not form until 120 Ma after the accretion of the moon. This result is in possible conflict with apparently older ages for possibly derivative or younger rocks (eg. 4.52+0.10 Ga for anorthositic norite 15455,228 (7); note that the uncertainty envelopes barely overlap).

Norites, troctolites, and dunites: These members of the Mg-suite of rocks are recognized as important constituents of the early crust, but their ages and chemical characteristics have been clouded by one or more kinds of natural or analytical open system behavior. Mineral isochron data selected by Nyquist (1) indicate crystallization ages from 4.1-4.5 Ga; although most of these dates have rather large uncertainties, several appear to be reasonably precisely dated, and, of these, a few have the same age as the accretion age of the moon. These problems, along with new data for Apollo 15 norites, have been recently discussed by Dasch et al. (8).

KREEP: The petrogenesis of these apparently widespread, enigmatic rocks is very imperfectly known. Most of the isochron ages cluster around 3.9 Ga, with model ages near 4.3 Ga (eg. 9).

Mare basalts: Although mare basalts mainly crystallized between 3.1 Ga ago and the end of widespread, intense bombardment about 3.9 Ga, recent work on Apollo 14 mare basalt fragments has uncovered a series of seven clasts with crystallization ages from 3.96 to 4.33 Ga (10). The existence of mare basalt volcanism during this period thus has somewhat complicated our understanding of early crustal evolution of the moon.

$^{87}Rb/^{86}Sr$ RATIOS FOR LUNAR ROCKS AND THE UNDIFFERENTIATED MOON.

A critical parameter in understanding the evolutionary history of lunar rocks is the $^{87}Rb/^{86}Sr$ ratio of the undifferentiated or whole moon. Ratios for derivative lunar rocks can be compared, and the effects of partial melting, fractional crystallization, and mixing or contamination can, in principle, be evaluated. The NASA/JSC laboratory has used a value of 0.05 for this ratio for the past 15 years. There are, at present, at least six lines of evidence supporting this value; two are the result of recent studies:

1. In a plot of Sm/Eu ratios vs. Rb/Sr ratios of carefully selected rocks, Nyquist et al. (11) derived the value of about 0.05 for the $^{87}Rb/^{86}Sr$ ratios from the intercept of the Rb/Sr curve with the curve for chondrites, assuming no Eu anomaly.

2. Time of crystallization vs. initial $^{87}Sr/^{86}Sr$ ratios (T-I Plots) of Apollo 12 basalts indicate 0.05 as a reasonable number for the bulk moon (12, 13).

3. Modeling of a variety of lunar processes by the JSC group during the period 1978-1987 have successfully used this value.

4. A plot of initial Sr isotopic composition vs. ε_{Nd} results in a value, within experimental error, of 0.05 (13).

5. A T-I plot for a selected group of Apollo 14 mare basalts also suggests a ratio very near 0.05 (8).

6. Careful isotopic work on pristine anorthosite 60025 is consistent with this value (6; pers. comm.).

ANALYSIS OF EARLIEST LUNAR AGES

Many of the pristine plutonic lunar rocks yield ages with analytical uncertainties too large for an unequivocal assignation of dates for the formation of the lunar crust. The main problem appears to be the intense comminution and shock metamorphism that these rocks have undergone as a result of meteoritic bombardment and the consequent but difficult to quantify redistribution of nuclides. Additional contributing problems are partial or continued equilibration of some components in the subsolidus state owing to elevated temperatures deep in the lunar crust, and inmixing of exotic components, such as other rock types, that cannot completely be excluded during sample preparation. Resulting age uncertainties preclude an unequivocal chronology for the fundamental problems--the validity of the magmasphere/magma ocean hypotheses, the age of the most ancient anorthosites relative to the time or times of intrusion of the Mg-suite of rocks, and the delimiting of the period of lunar crustal formation.

The least equivocal age data on PPLRs and other lunar rocks, however, suggest that:

(1) the lunar anorthositic crust may not have formed for about 120 Ma after the primary accretion of the moon at 4.56 Ga;

(2) at least some members of the Mg-suite of rocks (norites, troctolites, dunites) crystallized very soon after lunar accretion--within a very few 100s of Ma after 4.56 Ga--and cannot presently be distinguished chronologically from anorthositic crust formation;

(3) the onset of mare basalt formation began much earlier (about 4.33 Ga) than has been assumed and was in process before the most intense period of bombardment (about 3.9-4.0 Ga ago); and,

(4) recent evidence supports a value of about 0.05 for the $^{87}Rb/^{86}Sr$ ratio of the whole (or undifferentiated) moon.

COMPARISON OF EARLIEST LUNAR AND TERRESTRIAL EVENTS

With respect to bulk composition and earliest crustal formation, much more is known about the moon than about earth (see also Ryder and Dasch, this volume). Evidence of earth's oldest crust, about 3.8 Ga, or, arguably, about 4.28 Ga (14), still is about 280 Ma younger than its accretion age, and these rocks and minerals represent evolved rather than primitive crust. During the first 280-760 Ma of its existence, earth's much larger and more active mantle evidently obliterated existing traces of crust. Moon's earliest record suggests, however, that earth's early crustal evolution was probably petrologically complex and rapid.

REFERENCES: [1] Nyquist, L.E.(1981) in Magmatic Processes of Early Planetary Crusts, LPI Workshop, 71-77. [2] Morgan, P.H., et al.(1972) PLSC-3, 1361-1376. [3] Warren, P.H. and Wasson J.T. (1979) PLSC-10, 2051-2083. [4] Norman, M.D. and Ryder, G. (1980) PLSC-11, 317-331. [5] Papanastassiou, D.A. and Wasserburg, G.J. (1969) EPSL 5, 361-376. [6] Lugmair, G.W. (1987) LPSC-18, 584,585. [7] Nyquist, L.E. et al. (1979) in The Lunar Highlands Crust, LPI Conf., 122-124. [8] Dasch, E.J., et al. (1987) LPSC-18, 219,220. [9] McKay, G.A., et al. (1978) PLSC-9, 661-687. [10] Dasch, E.J., et al. (1987) LPSC-17, 150.151. [11] Nyquist, L.E., et al. (1973) PLSC-4, 1823-1846. [12] Nyquist, L.E., et al. (1979) PLSC-10, 77-114. [13] Nyquist, L.E., et al. (1981) EPSL-55, 335-355. [14] Compston, W., and Pidgeon, R.T. (1986) Nature, 321, 766-769.

CONSTRUCTION OF MASS - AGE CURVES FOR THE CONTINENTAL CRUST: AN EMPIRICAL MODEL AND AN EXAMPLE FROM THE WESTERN U.S.

D. J. DePaolo, G.S. Schubert, Ann Linn, Department of Earth and Space Sciences, UCLA, Los Angeles, California 90024   CD 146017

To elucidate the growth history of the continental crust it is necessary to know the present distribution of crust of different ages. We refer to this as a mass-age function. This information represents a starting point rather than the final answer to the problem because there is ample evidence that continental crust is not permanent. The establishment of mass-age curves for the continents is possible, but even if there were a complete data set for surface rocks there are uncertainties due to the lack of direct knowledge about the extent of underplating and injection of the lower crust by mantle-derived material that is not in evidence at the surface. These issues become central when one attempts to construct a quantitative curve.

We have at UCLA collected enough data on the southwestern quadrant of the U.S. to make a reasonable attempt at constructing a mass-age curve for this part of North America. The area we have studied represents about one per cent of the total mass of the continents, but it contains rocks ranging in age from early Archean to Tertiary, and provides an example of the various considerations that must go into determining the mass-age function.

At the heart of the determination of crustal age are the Sm-Nd isotopic systematics. Application of this technique requires the use of an empirical mantle evolution curve for the $^{143}Nd/^{144}Nd$ ratio. For our modelling we have used a simple straight-line evolution represented by $\varepsilon_{Nd}(T) = 8.6 - 1.91T$ where T is age in Gyr. Unlike some previous work, however, we do not use the model age T_{DM} as determined from the measured Sm/Nd ratios of individual rocks, but rather we use the *initial* ε_{Nd} values of igneous and metamorphic rocks and average crustal ε_{Nd} vs. T evolution curves to connect the initial (T, ε_{Nd}) points with the mantle evolution curve. Average crustal values of the Sm/Nd ratio vary with crustal age and can be computed to a good approximation from the expression $^{147}Sm/^{144}Nd (T) = 0.15 - 0.016T$, where T is the age of mantle derivation.

Another consideration in determining age is the mixing of materials of different ages during various crust-forming processes and the fact that the final isotopic compositions are weighted by Nd content rather than mass. We recognize two types of mixing in our model. In one case there is mixing between mantle-derived magmas and preexisting crustal material of known age. This generally occurs in areas where cratons are intruded by later mantle-derived magmas. In these cases the mixing calculation is straightforward except that the Nd concentrations of
the endmembers must be assumed. We assume the Nd concentration for the mantle component of such mixtures to be 60 per cent of the concentration in the crustal component. A second type of mixing occurs in continental margin magmatic arcs where the old crustal material is mostly sedimentary and is itself a mixture of crustal components of different ages. In this case we have several alternatives available. We have chosen to take the one with the least assumptions, which is to assume that the age distribution in the crustal material is uniform. It can be shown that in general the derived age of the mixture is not strongly dependent on the age distribution in the crustal component

In areas where young (Mesozoic) plutons intrude Precambrian basement, we can determine the mixing proportions in the partially mantle-derived plutons, but we need to estimate what fraction of the crust these represent. In this case we have assumed that the fraction of outcrop area of intrusive rocks is equal to the volume fraction in the crust. This procedure does not work for the Late Tertiary

where there is extensive volcanism. The effects of the Tertiary magmatism are a remaining uncertainty.

These procedures are then used to determine the mass-age curve for the crust. The crustal thickness is also used for this calculation although we find that an assumption of uniform crustal thickness would not change the distribution of ages significantly. The mean thickness of the crust (ie. depth to Moho) is not age dependent for the study area, although the range of observed thicknesses is larger for younger crust.

This exercise produces a semi-quantitative mass-age curve and helps identify the issues that must be resolved in order to make it quantitative. It also shows that about 200 data points are required to produce a reasonable representation of this area. One can estimate that 20,000 data points would be necessary to complete a global mass-age curve. To advance beyond this treatment requires a more detailed knowledge of processes in the lower crust associated with major magmatic events affecting cratons after their initial stabilization.

DOES SUBDUCTION ZONE MAGMATISM PRODUCE AVERAGE CONTINENTAL CRUST?

R. M. Ellam, Department of Earth Sciences, The University of Oxford, Parks Road, Oxford, OX1 3PR, U.K. C. J. Hawkesworth, Department of Earth Sciences, The Open University, Walton Hall, Milton Keynes, MK7 6AA, U.K.

Estimates of the average composition of the continental crust suggest that it is intermediate (1,2), and by invoking processes similar to those seen at Recent destructive plate margins, the andesite model was able to account for this average composition. However, there is now increasing evidence to suggest that andesites, far from being primary mantle melts, are actually derived by intracrustal fractional crystallisation of basaltic parents (3). Thus the net flux from mantle to crust at Recent subduction zones is basaltic, and consequently the average composition of newly formed continental crust today is somewhat more mafic than crustal estimates would suggest.
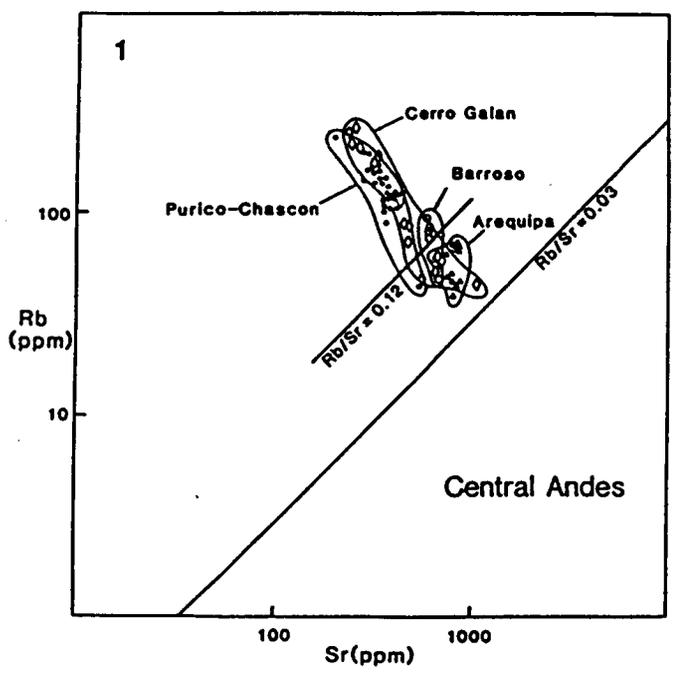
The problem may be simply illustrated on diagrams of Rb versus Sr content (figs. 1-3). Modern andesites (fig. 1) have high Rb/Sr ratios similar to those inferred from radiogenic isotope studies for the continental crust. However, it is striking that evolution towards this high Rb/Sr character is accompanied by decreasing Sr contents. This implies that fractional crystallisation of plagioclase played a dominant role in the evolution of the andesites. Thus, if andesites are produced by intracrustal processes, the flux from mantle to crust is best represented by their basaltic parents. However, subduction-related basalts from both intraoceanic arc and continental margin environments (fig. 2) have low Rb/Sr ratios similar to the bulk earth (0.03). Hence, not only is new crust mafic, but it also has low Rb/Sr, such that Recent crustal growth is apparently not responsible for large-scale fractionation of Rb/Sr between mantle and crust.
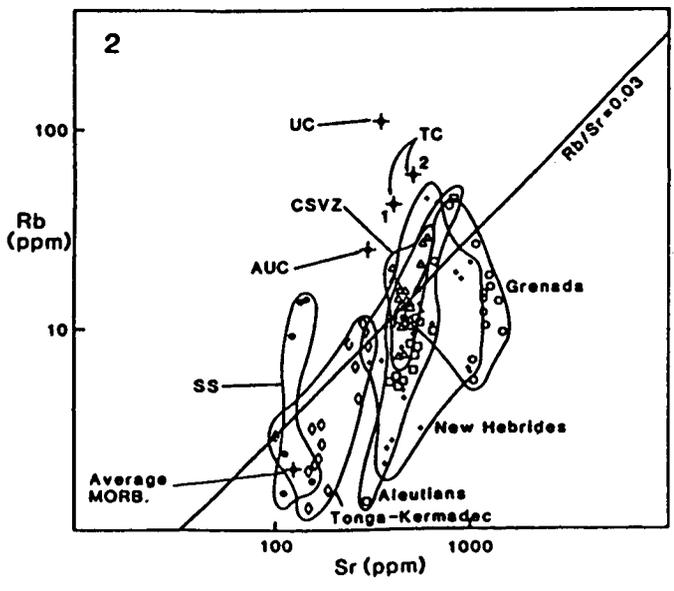
An implication of this petrogenetic model for andesites is that Recent continental crust will be differentiated into upper crust composed of andesites and dacites and a complimentary lower crust made up of the mafic cumulates and restites. A possible mechanism by which to generate an average crust of andesitic composition despite a net mantle to crust flux of basalt is to recycle the mafic cumulates and restite of the lower crust back into the mantle.

However, calculations reveal that the cumulate is likely to be gabbroic with up to 50% plagioclase (4,5). Not only is this lower crust likely to be considerably less dense than the underlying upper mantle, unless perhaps it is rapidly subjected to eclogite facies metamorphism, but it will also possess a positive Eu anomaly. If lower crust is recycled into the upper mantle, it is perhaps surprising that its predicted high Eu/Eu* signature has not been subsequently sampled by mantle-derived magmas.
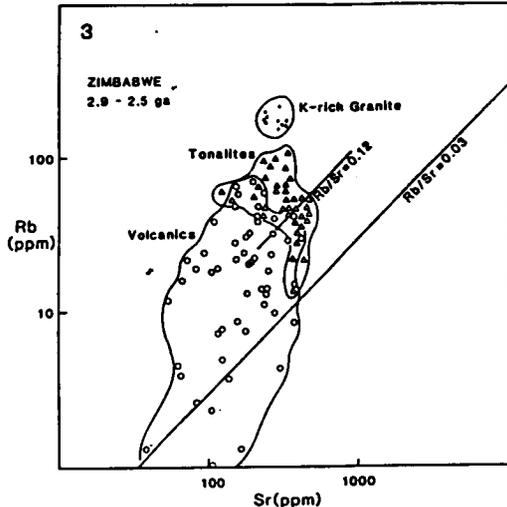
In contrast, Archaean volcanic and intrusive rocks from Zimbabwe which have been shown to represent newly formed continental crust between 2.9 and 2.5 Ga (6,7) show a quite different trend on the Rb versus Sr diagram (fig. 3). High Rb/Sr ratios, comparable with crustal estimates do occur, but these are accompanied by increasing Sr content, an indication that plagioclase did not play a major role in the generation of these high Rb/Sr ratios. Thus Archaean felsic crust may be consistent with derivation directly from mantle depths. However, a peridotite source is implausible, and a basaltic or amphibolitic source seems more likely.

It would appear that the mechanism of continent formation has changed with time. The present composition of new crust is likely to be basaltic and has low Rb/Sr. In contrast newly generated Archaean crust seems to be more akin to preferred average values for the bulk crust, i.e., high SiO_2 and high Rb/Sr. We conclude that these characteristics of the continental crust are housed within Arcaean rocks which represent a quite different mode of crust formation involving the fusion of pre-existing basaltic material at mantle depths. There is at least circumstantial evidence to suggest that crust-forming processes changed with time (8), and further evidence from lower crustal xenoliths (9) suggests that the mafic cumulates predicted by the model of Recent crustal growth may be largely restricted to areas underlain by post-Arcaean crust.

Finally, an important corollary of this model is that unless there is major recycling of lower crust back into the upper mantle, continued growth of crust at subduction zones will result in a more mafic bulk continental crust, and the degree of Rb/Sr fractionation between the mantle and crust will slowly diminish with time. Thus we are forced to revise our idea of subduction zones as sites of major mantle-crust fractionation and propose that they in fact represent regions where the chemical differences between the continental crust and the upper mantle are gradually reduced.

SUBDUCTION ZONE MAGMATISM
Ellam, R. M. and Hawkesworth, C. J.

61

REFERENCES

1. Taylor, S.R. & McLennan, S.M. (1981) *Phil. Trans. Roy. Soc. Lond.* A301, 381-399

2. Weaver, B.L. & Tarney, J. (1984) *Nature* 310, 575-577

3. Grove, T.L. & Kinzler, R.J. (1986) *Ann. Rev. Earth Planet. Sci.* 14, 417-454

4. Grove, T.L. & Baker, M.B. (1984) *J. Geophys. Res.* 89, 3253-3274

5. Ellam, R.M. (1986) Ph.D. thesis, Open University 210p

6. Hawkesworth, C.J. *et al.* (1975) *Earth Planet. Sci. Lett.* 25, 251-262

7. Hawkesworth, C.J. *et al.* (1979) *Earth Planet. Sci. Lett.* 43, 285-297

8. Taylor, S.R. & McLennan, S.M. (1985) *The Continental Crust: its composition and evolution*

9. Rogers, N.W & Hawkesworth, C.J. (1982) *Nature* 299, 409-413

Pb ISOTOPE CONSTAINTS ON THE EXTENT OF CRUSTAL RECYCLING INTO A STEADY STATE MANTLE;

S.J.G. Galer[1], S.L. Goldstein[2] and R.K. O'Nions[3]. [1]Scripps Instit. of Oceanography, La Jolla, CA. 92093, USA; [2]Max-Planck-Instit. für Chemie, Mainz, FRG; [3]Dept. of Earth Sciences, Univ. of Cambridge, UK.

Introduction: Recycling the continental crust back into the depleted mantle (DM) has been proposed by many authors, *e.g.* [1-4], with the mechanism for effecting this recycling taken to be subduction of pelagic sediment which preferentially samples the upper crust. It has not been generally recognised, however, that steady state concentrations will be approached for elements in the crust and the DM [5] and realised for models involving 'no-growth' of the continental crust with time [3,4]. Some of the ramifications of such steady states are examined for the Th-U-Pb system given the short residence time inferred for Pb in the DM [6].

Steady states: As a consequence of cycling material between the crust and the DM the *mean elemental storage ages* [5] $\langle T_S \rangle_{j(i)}$ of element i in the reservoirs j of the continental crust $(j=2)$ and the depleted mantle $(j=1)$ reach a steady state such that:

$$n_i = C_{2(i)} M_2 / \langle T_S \rangle_{2(i)} = C_{1(i)} M_1 / \langle T_S \rangle_{1(i)} \qquad (1)$$

where $C_{j(i)}$ is the concentration of i in reservoir j, M_j the mass of j and n_i the mass of i exchanged per unit time. From (1) it follows that:

$$n_{i1}/n_{i2} = [C_{j(i1)}/C_{j(i2)}]/[\langle T_S \rangle_{j(i1)}/\langle T_S \rangle_{j(i2)}] \qquad (2)$$

for two elements $i1$ and $i2$. This is quite general and must apply for each of the pairs Th-Pb, U-Pb, and Th-U of the Th-U-Pb system. It is only for bulk crustal recycling that $n_{i1}/n_{i2} = C_{2(i1)}/C_{2(i2)}$ and hence material additions from the DM are the same composition as the crust itself. In this instance, $\langle T_S \rangle_{2(i)}$ becomes identical for all elements in reservoir $j=2$ and equal to its overall 'mean age' of recycling into the DM. Thus for element fractionations at sites of egress and ingress from the DM, the elemental ratios of the additions n_{i1}/n_{i2} are not the same as those in the crust or mantle. In such circumstances systematic changes are expected to be observed in the element ratios of newly-formed crust and as it becomes gradually processed and injected back into the DM.

Because element ratios in the sedimentary mass such as Th/Sc [7], Sm/Nd [8,9] and U/Pb [10] have remained nearly constant through time, it follows that any crustal recycling involving the sedimentary mass cannot have resulted in significant inter-element fractionation at sites where it enters the DM otherwise secular changes would be observed in the element ratios of sediments and as inferred from ore deposit Pb. This immediately poses difficulties for recycling models involving subduction of pelagic sediment due to the substantial element fractionation likely to occur at destructive margins; that is unless a mechanism can be found for returning this material in bulk to the DM (*e.g.* [11]).

The relative inventories of Th, U, Pb and other highly incompatible elements in the crust and DM provide additional constraints on their mean storage ages in the DM using equation (1). Because the Sm/Nd ratio is not fractionated greatly within the crust, the Nd isotopic model age for the sedimentary mass of ~2.0 Ga [8,9] is likely to be a reasonable estimate of the Nd storage age in the crust, *cf.* [4]. For bulk crustal recycling this storage

age will also apply for other elements besides Nd such as Pb.
Taking a Pb content for the continental crust as a whole of 10 ppm
[5,7,12] and 0.025 ppm in the MORB source [6] together with a
crustal mass of 0.0053 of the silicate earth and $\langle T_S \rangle_{2(Pb)}$=2.0 Ga,
yield $\langle T_S \rangle_{1(Pb)}$ of ~0.94 Ga and ~0.27 Ga if the DM comprises the
entire mantle or just the upper mantle (0.272 of the silicate
earth) respectively. These values should be directly comparable
with the Pb residence times in the DM of ~0.6 Ga [6] (for a bulk
earth Th/U ratio K=3.9). The crustal inventory of highly incom-
patible elements therefore provides a strong contraint on the
timescales of chemical processing of the DM. The timescale to mix
and disperse DM materials should be less than this processing time
[6]; *i.e.* <0.27 Ga if the DM is confined to the upper mantle. For
the preferential recycling of upper crust or pelagic sediment with
~15 ppm Pb [7,12] into the DM this timescale would be further
shortened by a factor of ~1.5. This appears to require a DM reser-
voir that is chemically processed every ~10^8 a if comprising the
upper mantle or once in <10^9 a for the whole mantle. Neither these
nor intermediate alternatives are consistent with the longer
timescales deduced for mixing the upper or whole mantle from
numerical studies [13,14].

Mean isotopic ages: It is a property of all crust-mantle
models that derive the crust from a fixed mass portion of the
mantle with or without crustal recycling, that the mean ages of
parent-daughter fractionation (or *mean isotopic ages* [5]) of the
crust and mantle reservoirs are identical. This must hold for each
of the three Pb isotope systems. The mean isotopic age $\langle T_I \rangle_j$ is
defined with respect to the bulk system evolution as:

$$\langle T_I \rangle_j = [1/\lambda].\ln\{1 + (\alpha_j - \alpha_S)/(\mu_j - \mu_S)\} \qquad (3)$$

where λ is the decay constant, α_j is the isotope ratio in reser-
voir j, μ_j its parent isotope-daughter isotope ratio and S refers
to the bulk crust-DM system; α_S is given by:

$$\alpha_S = [\Sigma\ C_{j(Pb)}M_j\alpha_j]/[\Sigma\ C_{j(Pb)}M_j] \qquad (4)$$

One of the effects of recycling crust is to strongly decouple the
measured and time-integrated parent-daughter ratios in both the
crust and the DM. Equation (3) also stipulates that the
$^{207}Pb-^{206}Pb$ ages of the crust and DM reservoirs be identical, or:

$$[\alpha_1' - \alpha_S']/[\alpha_1 - \alpha_S] = [\alpha_2' - \alpha_S']/[\alpha_2 - \alpha_S] \qquad (5)$$

where ' denotes the ^{207}Pb system. If most of the Pb is con-
centrated in the continental crust then $\alpha_S \approx \alpha_2$ and it can be shown
that the Pb-Pb age is equal to the storage age $\langle T_S \rangle_{1(Pb)}$ in the
DM. Thus any component part of the continental crust with a sub-
stantially greater inventory of Pb than the DM should on recycling
define a Pb-Pb age of ~0.6 Ga with 'average' DM Pb in accordance
with the Pb residence time in the DM for K=3.9 [6].

Taking pelagic sediment Pb isotope ratios as typical of the
upper crust [12] and those of average MORB [12,15] yields a Pb-Pb
age of ~4.2 Ga, with permissible values likely to be >3 Ga. This
result appears to limit the quantity of material that may be added
to the DM in the form of pelagic sediment or average upper crust
[6]. Similarly, the estimated Pb isotopic compostion of the lower
crust [12] and DM yield an age of ~2.7 Ga, which also limits the
incorporation of solely lower crustal materials into the DM.

The overall Pb isotopic composition of any recycled crustal
component has to lie close to the geochron at μ~8.2 in order to

fulfil the need for a Pb-Pb age of ~0.6 Ga with DM Pb. This in turn engenders that it be somehow representative of the entire crust. Indeed, recycled crust is required to have a Pb isotopic composition more similar to primitive mantle than the separate components of the continental crust itself [5]. Aside from the difficulty in finding a suitable portion of bulk continental crust there remains the question of the exact mechanism by which this material might be injected into the DM. One possibility, for example, would be direct additions in continental collision zones [16]. Because pelagic sediment preferentially samples the upper crust it cannot contribute the bulk of this material, effectively ruling out subduction as a prime means of adding crustal Pb to the DM. Hence any pelagic sediment that is subducted has to be quantitatively small or else effectively purged of Pb by destructive margin magmatism [6].

Summary: Recycling of continental crust back into the DM has been advocated by numerous authors, *e.g.* [1-4]. However, recycling of crust leads to the development of steady states both in that portion of continental crust recycled and in the DM, which poses constaints on the portions of crust that may be involved in recycling and on the rate at which the DM must be chemically processed to achieve such steady states. In considerations of the 'no-growth' hypothesis for the continental crust the following points are noted: (1) the uniform composition of the sedimentary mass with time is only consistent with recycling material from the crust in bulk without incurring element fractionations; (2) the present inventories of Th, U and Pb in the continents require that 'processing' of the DM occurs on a shorter timescale than is consistent with current numerical studies of mantle convection; (3) the short storage age inferred for Pb in the DM [6] does not allow recycling of specific portions of the crust such as the upper or lower crust alone, but permits recycling of bulk crust that isotopically resembles primitive mantle for Pb; (4) subduction of pelagic sediment itself is not a viable mechanism for returning crust to the mantle. These and other considerations [5,6,12,17,18] constrain the manner and amount of permissible crustal recycling and fail to support it as important in crustal or mantle evolution.

References: [1] Russell R.D. & Birney D.J., 1974, *Phys. Earth Planet. Inter.*, *8*, 158-166. [2] Zartman R.E. & Doe B.R., 1981, *Tectonophys.*, *75*, 163-179. [3] Armstrong R.L., 1981, *Phil. Trans. R. Soc. A*, *301*, 443-472. [4] DePaolo D.J., 1983, *Geophys. Res. Lett.*, *10*, 705-708. [5] Galer S.J.G., Goldstein, S.L. & O'Nions R.K., 1986, *Lunar Planet. Sci. XVII*, 247-248; 1987, *Chem. Geol.* (submitted). [6] Galer S.J.G. & O'Nions R.K., 1985, *Nature*, *316*, 778-782. [7] Taylor S.R. & McLennan S.M., 1985, *The Continental Crust: its composition and evolution*, Blackwell. [8] Goldstein S.L. *et al.*, 1984, *Earth Planet. Sci. Lett.*, *70*, 221-236. [9] Miller R.G. *et al.*, 1986, *Chem. Geol.*, *57*, 87-99. [10] Albarède F. & Juteau M., 1984, *Geochim. Cosmochim. Acta*, *48*, 207-212. [11] Hilde T.W.C., 1983, *Tectonophys.*, *99*, 381-397. [12] Newsom H.E. *et al.*, 1986, *Earth Planet. Sci. Lett.*, *80*, 299-313. [13] Hoffman N.R.A. & McKenzie D.P., 1985, *Geophys. J. R. astron. Soc.*, *82*, 163-206. [14] Gurnis M. & Davies G.F., 1986, *J. Geophys. Res.*, *91*, 6375-6395. [15] Ito E. *et al.*, 1987, *Chem. Geol.* (in press). [16] Molnar P. & Gray D., 1979, *Geology*, *7*, 58-62. [17] Hofmann A.W. *et al.*, 1986, *Earth Planet. Sci. Lett.*, *79*, 33-45. [18] Patchett P.J. *et al.*, 1984, *Earth Planet. Sci. Lett.*, *69*, 361-379.

DECOUPLED Rb-Sr AND Sm-Nd ISOTOPIC EVOLUTION OF THE CONTINENTAL CRUST.

(S.L. Goldstein, Max-Planck-Institut für Chemie, Saarstrasse 23, Postfach 3060, D-6500 Mainz, FRG.)

Introduction: Sm-Nd isotopic studies of clastic sedimentary rocks have been used to constrain the growth and evolution of the continental crust through determination of "crustal residence" ("Tcr") ages (1), which are estimates of the average time that the Nd in a sediment has been resident in the continental crust. The method is useful because the geochemical behavior of Sm and Nd is relatively simple: they reside primarily in silicates; the ratio of Sm/Nd in granitoid and sedimentary rocks is about 50 percent lower than depleted mantle; and this ratio is resistant to the effects of intra-crustal processes such as weathering and metamorphism. In contrast, overall Rb-Sr isotopic systematics of continental rocks are more obscure because Rb/Sr ratios and Sr isotopic compositions are highly variable, the ratio of Rb/Sr is easily fractionated by intra-crustal processes, and carbonates as well as silicates are an important reservoir of crustal Sr. Nonetheless, an evaluation of Nd and Sr isotopes in sedimentary rocks indicates that the two isotope systems are decoupled. Rb-Sr is affected to a much greater degree than Sm-Nd by exchange between the crust and mantle, which has important implications for the isotopic evolution of the mantle and the continents.

Evidence for Decoupling of Nd and Sr: The Sm-Nd isotopic systematics of sedimentary rocks provide a basis for evaluation of Rb-Sr in continental crust. It has been shown that Sm/Nd and Nd isotopic compositions of sediments representative of large areas of the continental crust vary within a small range that corresponds to Tcr ages of about 1.7 ± 0.35 Ga (2,3). Sediments deposited during the Phanerozoic have Tcr ages of about 1.9 ± 0.4 Ga (1,4-7), and do not show a clear trend toward significantly younger Tcr ages with time. From this it can be inferred that the Phanerozoic has not been an era of large continental growth, although the data allow for about a 10% addition of juvenile material. Because the Tcr age of a sediment represents only the average residence ages of the sources, the relationship of these ages to the mean age of the continental crust must be inferred using independent criteria.

Evaluations of the mass of preserved sediments vs. depositional age indicate exponential decrease from the present day, half of the sedimentary mass has been deposited within the last 250-600 Ma, and the source of sediments is primarily (up to 90%) pre-existing sediments (8,9). Therefore the sedimentary mass behaves as a nearly closed system. These observations and literature data on sediments indicate that the mean Tcr age of the whole sedimentary mass is about 2.0 Ga (3,10).

Rb-Sr isotopic systematics of clastic sediments contrast markedly with Sm-Nd in that the Sr isotopic compositions *have remained nearly constant at the time of deposition* since the late Proterozoic, with 87Sr/86Sr = 0.71-0.72 (6,11-13). This is the same range as observed in particulates carried by major rivers draining orogenic belts at present (14). However, in these rocks Rb/Sr are typically >1. If the sedimentary mass behaved as a

nearly closed system, then the high Rb/Sr should be reflected in the Sr isotopic compositions of newly deposited sediments: these should increase to be >0.73-0.74 within 500 Ma of the time the system became closed.

The observations of small Phanerozoic crustal additions, a nearly closed sedimentary system, and intra-sedimentary recycling on a fast time scale are consistent with the Nd isotope data on Phanerozoic sediments, which show more negative ϵ(Nd) at the time of deposition toward the present. These cannot be easily reconciled with the constancy exhibited by initial Sr isotopic compositions. The Nd and Sr isotopes appear to be decoupled, and a process that has little effect on Nd buffers the Sr isotopes.

The Sr isotopes in clastic sediments can be easily buffered by the small influx of juvenile material to the sedimentary mass. Sr is lost during weathering due to its solubility in water. Clastic sediments typically have Sr/Nd ~ 3, while volcanogenic detritus has Sr/Nd ~ 40. Small additions thus have a large effect on Sr isotopes, while Nd behaves as a nearly closed system.

Effects on the Isotopic Evolution of the Mantle and Crust: It appears that significant recycling of upper crustal material to the mantle has not been an important process in crust-mantle evolution (15-18), and therefore significant amounts of upper crustal Nd and Pb have not returned to the mantle. However, continental Sr can be recycled to the mantle through exchange of seawater Sr with basalt at spreading ridges (19), and subduction of carbonate (20) associated with spreading ridges.

Hydrothermal exchange processes at spreading ridges may have important effects on the Sr isotopic evolution of the crust and mantle. Approximately 10^{10} moles of seawater Sr exchanges with Sr from MORB each year (19), resulting in a loss of 0.6×10^7 moles of 87Sr from the continents. The crustal production rate of 87Sr is 3×10^7 moles if Rb $\approx$ 32 ppm (21), and thus the net amount of 87Sr lost to the mantle is about 20% of that produced by Rb decay in the crust. Over time this recycling process inhibits the growth of radiogenic 87Sr in the crust and as a result the "mean isotopic age" (18) of Sr in the continental crust is lower than the true age, by 20% if the present exchange rate were typical through Earth history. Exchange also affects the Sr isotopic evolution of the upper mantle. For the simple case of an average seawater 87Sr/86Sr $\approx$ 0.705 (estimated from ref 24), constant seafloor spreading rates, and Sr exchange similar to present rates, this process would account for 20-30% of the increase of 87Sr/86Sr of the upper mantle since 4.0 Ga.

The integrated effect of continent to mantle recycling of Sr on the isotopic evolution of the mantle and continents depends upon seafloor spreading rates, the Sr isotopic composition of seawater, and the growth rate of the continents through time. For the continents, the increase in Sr isotopic composition is inhibited by the addition of juvenile material from the mantle and recycling of continental Sr to the mantle. If seafloor spreading was faster in the past, and/or if carbonates are

subducted along with oceanic crust, both the effect of decreasing the apparent mean Rb-Sr age of the continents, and the contribution of continental Sr to the increase of the $87Sr/86Sr$ of the upper mantle would be greater than the above estimates. Upper limits to the effect of the hydrothermal exchange on the mantle and crust can be estimated: (a) if seafloor spreading rates are proportional to the square of the rate of heat production in the Earth, then exchange could account for the total increase of $87Sr/86Sr$ in the upper mantle; (b) if the crustal growth rate has been uniform since ~4.0 Ga, and carbonate subduction is not important, then the $87Sr/86Sr$ of the continents might not be increasing in the present day.

Conclusions: The isotopic systematics of Nd and Sr are apparently decoupled in the continental crust due to preferential recycling of continental Sr to the mantle. An effect of this process is that the "mean isotopic age" of Sr in the continental crust is younger than Nd. Recycling of continental Sr into the mantle has contributed significantly to the increase of the Sr isotopic composition of the upper mantle over Earth history, although it is neither the sole nor (probably) the major contributor of radiogenic Sr to the upper mantle.

References: (1) O'Nions, R.K. et al. 1983, Earth Planet. Sci. Lett. 63, 229-240. (2) Taylor, S.R. et al. 1984, Geochim. Cosmochim. Acta 47, 1897-1905. (3) Goldstein, S.L. et al. 1984, Earth Planet. Sci. Lett. 70, 221-236. (4) Miller, R.G. & O'Nions, R.K. 1984, Earth Planet. Sci. Lett. 68, 459-470. (5) Michard, A. et al. 1985, Geochim. Cosmochim. acta 49, 601-610. (6) Davies, G. et al. 1985, Earth Planet. Sci. Lett. 75, 1-12. (7) Andre, L. et al. 1986, Chem. Geol. 57, 101-115. (8) Garrels, R.M. & MacKenzie, F.T. 1971, *Evol. of the Sed. Rocks*, W.W. Norton & Co. 394 pp. (9) Veizer, J. & Jansen, S.L. 1979, J. Geol. 87, 341-370. (10) Miller, R.G. et al. 1986, Chem. Geol. 57, 87-99. (11) Gebauer, D. & Grünenfelder, M. 1974, Contrib. Mineral. Petrol. 47, 113-130. (12) Bofinger, V.M. et al. 1970, Geochim. Cosmochim. acta 34, 443-445. (13) Cordani, U.G. et al. 1978, SEPM Stud. in Geol. 6, 93-117. (14) Goldstein, S.L. unpub. data. (15) Patchett, P.J. et al. 1984, Earth Planet. Sci. Lett. 79, 361-375. (16) Hofmann et al. 1986, Earth Planet. Sci. Lett. 79, 33-45. (17) Newsom H.E. et al. 1986, Earth Planet. Sci. Lett. 80, 299-313. (18) Galer, S.J.G. et al. 1986, Lunar Planet. Sci. XVII, 247-248. (19) Albarede, F. et al. 1981, Earth Planet. Sci. Lett. 55, 229-236. (20) Mattey, D.P. 1987, Terra Cognita 7, 31-38. (21) Taylor, S.R. & McLennan, S.M. 1985, *The Cont. Crust: its Compos. & Evol.*, Blackwell Sci. Pub., 312 pp. (22) Veizer, J. & Compston, W. 1976, Geochim. Cosmochim. acta 40, 905-914.

CRUSTAL EVOLUTION AND THE ECLOGITE TO GRANULITE PHASE TRANSITION IN XENOLITHS FROM THE WEST AFRICAN CRATON

S.E. Haggerty, D.V. Hills, P.B. Toft. Dept. of Geology, Univ. of Mass., Amherst, Massachusetts, 01003, U.S.A.

The Man Shield forms the southern segment of the West African Craton and is divided into three Precambrian age provinces (Fig 1): Leonean ($\approx$3.0 Ga), Liberian ($\approx$2.7 Ga); and Eburnian ($\approx$2.0 Ga). The Leonean and Liberian provinces are dominated by granitic gneisses and migmatites with sporadic occurrences of charnokites, granulites, amphibolites and ferruginous quartzites. The Eburnian province is distinctive in the abundance of greenstones, metagraywackes, and synkinematic granites. These provinces are fault bounded against mafic and pelitic foldbelts (1) of Pan African age ($\approx$550 Ma). The Liberian province is of particular interest because lower crustal and upper mantle lithologies have been sampled by diamondiferous kimberlites (90-120 Ma) (2-6), and the region offers an unusual opportunity: (a) to establish a stratigraphy to depths of approximately 200 km; (b) to determine a xenolith paleogeotherm; (c) to test the validity of crust-mantle discontinuities in relation to petrology and geochemistry; and (d) to evaluate continental growth and accretionary mechanisms during the early Archean. This investigation has concentrated on suites of eclogites from the Koidu diatreme, Sierra Leone, and on granulites from the Sample Creek kimberlite pipe in Liberia, localities that are 110 km apart, astride 11°W.

Approximately 100 xenoliths were selected for bulk chemical, petrographic, electron microbeam and magnetic analyses. Two sub-populations of eclogites from Koidu are apparent, distinguished according to Mg content. The higher MgO group (16-20 wt%) approaches komatiitic basalt, and the lower MgO population (6-13 wt%) has close chemical affinities to alkaline hawaiites. Granulite facies xenoliths from Sample Creek are compositionally similar to high alumina alkaline basalt and to estimates of lower crustal compositions typical of cratonic shield regions (7); garnet anorthosites (>90 modal % An_{85}) and high Mg eclogites are also present. The overall AFM distribution is similar to the calc-alkaline trend but at lower total iron (Fig 2). Anorthosites and high Mg eclogites fall into two distinct fields, whereas the lower Mg eclogites and the granulites link these two fields and overlap each other. The largest specimen in the collection (SCC-5 from Sample Creek), $\approx$25cm per side, is banded on a 1-5cm scale into plagioclase, and pyroxene + garnet layers. Three sub-samples of this xenolith were selected to represent plagioclase dominant, pyroxene + garnet-rich, and intermediate lithologies. Analyses fall respectively into the fields of Sample Creek anorthosites, low Mg Koidu eclogites, and the overlap group of Koidu eclogites and Sample Creek granulites, providing further evidence for a geochemical continuity between these suites.

There are no deep seismic data for the Man Shield to constrain a Moho. Nevertheless, the granulites and low Mg eclogites can be arrayed by consideration of chemistry and specific gravity (Fig 3). For these xenoliths, SG is directly proportional to FeO+MgO and inversely proportional to alkalies and to SiO_2.

Although these relationships are more pronounced for the granulites than for the eclogites the mole ratio of $(Na_2O+K_2O+SiO_2)/(FeO+MgO)$ for both suites is taken to represent the concentration of crustal over mantle components. Samples that plot to the right and left of a continental lower crustal average composition (7) have more highly crustal and more mantle-like affinities, respectively. All eclogites but one plot to the left of this line and no granulites are denser than 3.3; granulites and eclogites overlap between SG 3.0 and 3.3. Seismic P-wave velocities estimated from SG (8) show a range from 6.6 to 8.7 km/sec, spanning typical crustal and mantle values, with a transitional region between 40 and 70 km (Fig 3).

Garnet-Cpx geothermometry (9) has been applied by assuming a range of pressures, and P-T gradients for 30 xenoliths have been obtained (Fig 4). The best constrained xenolith is a diamond + graphite eclogite, followed by diamondiferous eclogites, and eclogites having graphite, coesite, or kyanite as accessory minerals. The upper and lower bounds respectively, for eclogites and granulites, are based on estimates (10,11) for this transition. A xenolith paleogeotherm (Fig 4) is defined, bracketed by model cratonic 40 and rift 90 mW/m² geotherms (12).

The ages of exposed rocks in the Man Shield and the presence of diamonds of ultramafic affinity suggests that a geochemically depleted subcratonic lithosphere, rooted to the crust, was in place at 3.5-2.7 Ga (13,14). Chemical data for the xenolith suites may be interpreted as differentiated Archean magmas that failed to erupt. In conclusion: (a) the granulites, eclogites and anorthosites appear to be geochemically, geophysically and petrologically related; (b) this continuum may represent the fractionation and segregation of felsic components from the mantle to an early Archean crust; (c) the paleogeotherm links the present day low heat flow to a much earlier episode of high heat flow characteristic of crustal regional metamorphism; (d) the crust-mantle boundary is at least in part a gradual but interlaminated geochemical, mineralogical and seismic transition; (e) the timing and style of granulite and eclogite formation in the Man Shield was between 3.5 and 2.7 Ga ago probably in a non-subduction related volcanic environment.

REFERENCES
(1) Hurley P.M. et al., (1971) Geol. Soc. Amer. Bull., 82, 3,483; (2) Haggerty S.E. (1982) J.G.R., 87, 10,811; (3) Haggerty S.E. (1983) N. Jb. Min. Mh., 8, 375; (4) Haggerty S.E. and Tompkins L.A. (1983) Nature, 303, 295; (5) Tompkins L.A. and Haggerty S.E. (1984) Proc. 4th I.K.C., 83; (6) Haggerty S.E. and Toft P.B. (1985) Science, 229, 647; (7) Taylor S.R. and McLennan S.M. (1985) Continental Crust: Comp. & Evol. 73-95, Blockwell; (8) Hall J. (1986) 51-62 In: Dawson et al., Geol. Soc. Spec. Pub. 24; (9) Ellis D.J. and Green D.H. (1979) C.M.P., 71, 13; (10) Ito K. and Kennedy G.C. (1970) Min. Soc. Amer. Spec. Pap., 3, 77; (11) Furlong K.P. and Fountain D.M. (1986) J.G.R., 91, 8,285; (12) Pollack H.N. and Chapman D.S. (1977) Tect. Phys., 38, 279; (13) Haggerty S.E. (1986) Nature, 320, 34; (14) Boyd F.R. and Gurney J.J. (1986) Science, 232, 472.
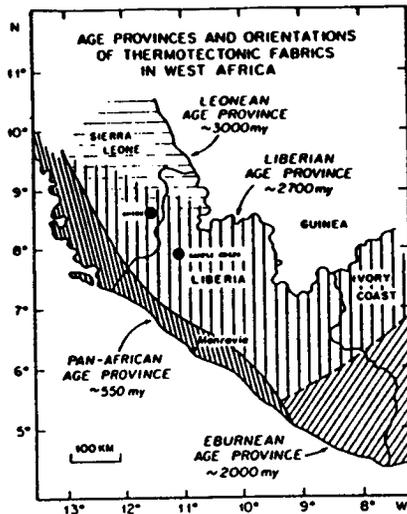
Fig 1
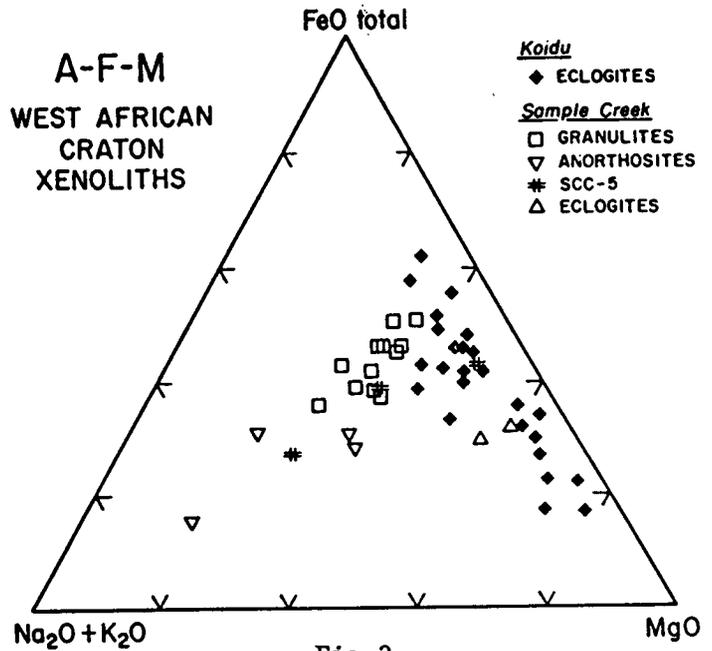
Fig 2
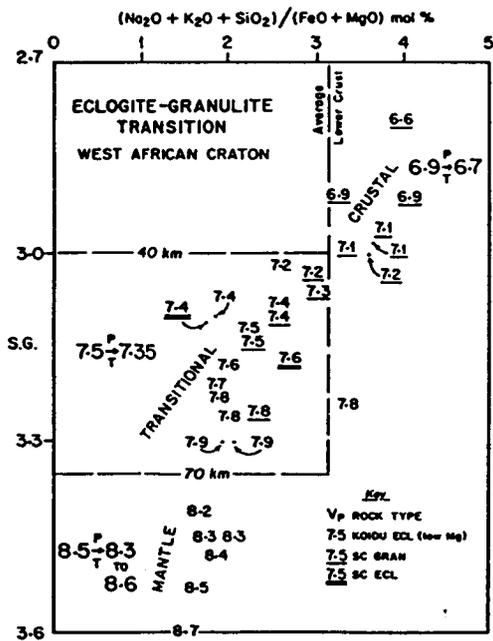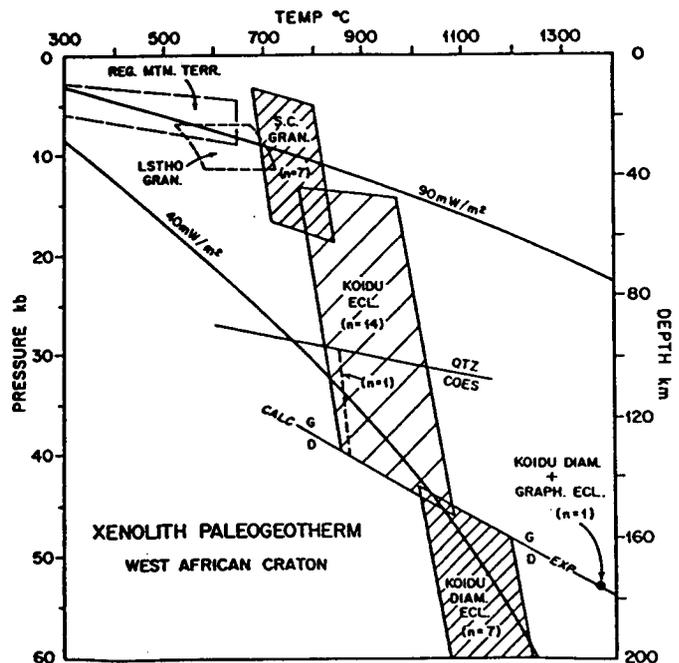
A-F-M

WEST AFRICAN
CRATON
XENOLITHS

FeO total

Koidu
◆ ECLOGITES

Sample Creek
□ GRANULITES
▽ ANORTHOSITES
✳ SCC-5
△ ECLOGITES

Na₂O + K₂O

MgO

Fig 3

ECLOGITE-GRANULITE
TRANSITION
WEST AFRICAN CRATON

(Na₂O + K₂O + SiO₂)/(FeO + MgO) mol %

Fig 4

XENOLITH PALEOGEOTHERM
WEST AFRICAN CRATON

TEMP °C

METAMORPHIC P-T PATHS AND PRECAMBRIAN CRUSTAL GROWTH IN EAST ANTARCTICA.

S.L. HARLEY, EARTH SCIENCES, OXFORD, ENGLAND, OX1 3PR.

An extensive crustal history from ca. 3750 Ma (1) (or even 3950 Ma) to 540 Ma (2) is preserved in granulites from Enderby Land and Prydz Bay, East Antarctica (fig 1). Independent studies of basement gneisses and intrusive dyke suites demonstrate that a substantial volume of Archaean crust some 35-40 km thick, the Napier Complex, was stabilised subsequent to major tectonic events around 3100-2900 Ma b.p. (3) until underplating by Proterozoic crust at 960 Ma-1000 Ma (3) caused en-masse partial uplift of the terrane. Granulites of the Napier Complex were resident in the mid- to lower-crust for some 2000 million years prior to late Proterozoic collision.
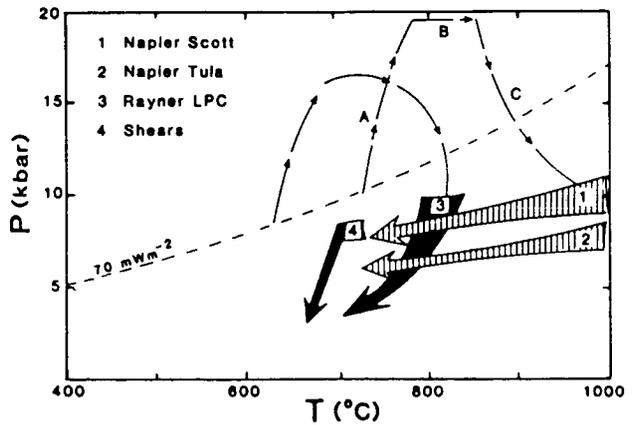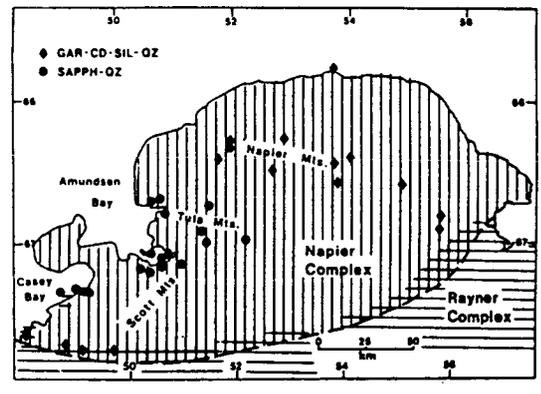
Fig 1 : Locality map of Archaean (Napier) and Late Proterozoic (Rayner) granulites, with areas of reworking.

Fig 2 : Archaean (1 and 2) and Proterozoic (3 and 4) P-T paths, with proposed earlier segments. A - early collision; B - late collision heating; C - syn-extension.

The Napier Complex contains granulites metamorphosed at 950-1000°C and 7-11 kilobars (3) (fig 2). Evidence in support of this extremely high metamorphic grade include (a) regional occurrence of the high-T mineral assemblages sapphirine + quartz (3,4,5,6,7), osumilite + garnet (3,6,8), hypersthene + sillimanite (3,4,9), spinel + quartz (3,4), and intermediate pigeonite (3,10,11) (b) high alumina contents (10-13 wt % Al_2O_3) of initial orthopyroxenes coexisting with garnet or silliminate in metapelites (10,12) (c) Fe-Mg exchange geothermometry of garnet - pyroxene pairs (12). This metamorphism affected a diverse assemblage of rocktypes including supracrustals (aluminous pelites, quartz sandstones, ironstones, evaporites, chemical sediments, volcanics) and orthogneisses which in age range from ca. 3070 Ma to 3750 Ma (3). Orthogneisses include a low-Y, HREE depleted suite suggested to be derived from deep-seated remelting of mafic sources and a non-depleted suite consistent with the remelting of earlier felsic crustal rocks (13). Examples of both types were emplaced at the time of the main deformation-metamorphic events. It is likely that magmatism was significant in redistribution of the internal heat budget at 3100 Ma, leading to similar high-temperatures over a 9-14 km depth range in the lower crust (10,11).

Near-isobaric cooling P-T histories (fig 2) are typical of
the Napier Complex at all exposed crustal levels (7,9,12) (11-5
kbar initial). These paths include an element of decompression
(1-2 kbars), suggesting the crust was slightly overthick at the
commencement of cooling (ca. 40-45 km), but are dominated by
cooling from 950°C to 650-700°C (12). Important evidence in
support of near-isobaric cooling include (a) secondary garnet
coronas and rims in mafic, felsic, and pelitic granulites
(9,12,14,15) (b) geothermobarometry of secondary assemblages
and rims (12,14), and (c) secondary hypersthene + sillimanite
coronas on sapphirine.

The P-T paths for the Napier Complex, coupled with the
intense deformation history and associated magmatism at 3100-
2900 Ma (3), suggest a deep extensional setting for the
metamorphism. It is considered that these granulites were
equilibrated following extension of crust previously thickened
in a collisional orogeny. Extension in this model is the last
phase in the development of a diffuse, wide, orogenic belt, as
typified by the Tibetian Plateau. Magmatism is a natural
consequence of such a collision-extension scenario, involving
both remelting of extant crustal material and accretion plus
remelting of new material derived from upwelling asthenosphere.
It is expected that the collision-extension process lasted no
longer than 50-80 Ma (16), after which the Napier Complex was
stabilised.

The continued stability of the Napier Complex after 3000 Ma
is documented by studies of metamorphic overprints on these
granulites and observations on Proterozoic tholeiitic dyke
swarms (17). P-T data pertaining to a regional deformation
event at 2460 Ma (2) indicate continued conditions of 6-9 kbars
and 650-700°C some 500 Ma after the major crustal stabilisation
events (2,9,12). Experimental and petrological studies of
phenocryst assemblages in dykes aged 2400 Ma, ca. 1800 Ma and
1100 Ma furthermore demonstrate crystallisation at 8 kbar (17).
Thus, the granulites of the Napier Complex remained deeply
buried until at least 1100 Ma.

The Rayner Complex (4), to the south of and adjacent to the
Napier, is part of a late Proterozoic granulite complex
(1000-960 Ma) consisting mainly of mid- to late-Proterozoic
orthogneisses and metasediments (18). Reworking of the Archaean
craton is confined to its immediate margin and in localised
shear zones cutting the Archaean gneisses. Granulites from the
Rayner Complex and associated belts in East Antarctica preserve
reaction textures and zoning data which indicate near-isothermal
decompression P-T-t histories (fig 2) subsequent to the major
deformations (18;19). These histories are consistent with
continental collision, probably involving a recently extended
and hot margin or magmatic arc. Related retrograde shear zones
in the Napier Complex also show near-isothermal decompression
(fig 2), from 8 kbar at 700°C to 3-4 kbar at 650°C (20,21).
Thus, granulites which had resided within 10 km of the base of
the crust since ca. 3000 Ma were decompressed through some 14-17
kms at ca. 1000 Ma without pervasive deformation. This

metamorphic imprint is interpreted to reflect the underplating of the Napier Complex, over a large region (10,000 km^2), by at least 15 km thickness of younger (Proterozoic) crust.

Two very important implications of these crustal histories for the interpretation and measurement of crustal growth rates are:

(1) isobaric cooling paths can demonstrate that a stable, normal, thickness of crust was present in a region by the conclusion of a major tectonic event - no matter what has happened subsequently in later, unrelated, events. A volume of continental crust equivalent to the area of the Napier Complex (approx. 50,000 km^2) x 35 km was present by 3000 Ma.

(2) underplating may play an important role in the exhumation of old granulite terranes, without leading to any pervasive over-printing, magmatism, or isotopic signature. The volume of underplated crust can often only be estimated through the use of metamorphic P-T path relationships, and yet must be considered in calculations of the growth rate of continental crust.

1. Black L.P., Williams I.S. and Compston W. (1987) Int. Conf. on Isotope Geochemistry and Geochronology (Cambridge).
2. Black L.P., James P.R. and Harley S.L. (1983) J. Metam Geol 1, p. 277-303.
3. Harley S.L. and Black L.P.(1987) Spec Publ. Geol Soc Lond 28.
4. Sheraton J.W., Offe L.O., Tingey R.J. and Ellis D.J. (1980) J. Geol Soc Aust, 27, p 305-317.
5. Grew E.W. (1980) Amer, Mineralogist 65, p 821-836.
6. Ellis D.J., Sheraton J.W., England R.N. and Dallwitz (1980) Contribs Mineral Petrol 72, p. 123-143.
7. Ellis D.J. (1980) Contrib Mineral, Petrol, 74 p 201-210.
8. Grew E.S. (1982) Amer. Mineralogist 67, p 762-787.
9. Harley S.L. (1983) Antarctic Geoscience p 24- 30.
10. Harley S.L. (1987) J. Metam Geol 5, (in press).
11. Sandiford M.A. and Powell R. (1986) Amer. Mineralogist 71, p 9 46-954.
12. Harley S.L. (1985a) J. Petrology 26, p 819-856.
13. Sheraton J.W. and Black L.P. (1983) Lithos 16, p. 273-296.
14. Ellis D.J. and Green D.H.(1985) J. Petrology, 26, p.633-662.
15. Sandiford M.A. (1985) J. Metam Geol. 3, p 155-179.
16. England P.C.(1987) Tectonic Settings of Metamorphism, p 3-22.
17. Kuehner S. (1986) Unpubl PhD Thesis Univ Tasmania.
18. Black, L.P., Harley S.L., Sun S.S. and McCulloch M.T. (1987) J. Metam Geol 5, p. 1-26.
19. Ellis D.J. (1983) Antarctic Geoscience p 20-24.
20. Sandiford M.A. (1985b) J. Struct. Geol 7, p 477-488.
21. Harley S.L. (1985b) J. Metam Geol 3, p 179-200.

74

156 032

EARLY INTENSE CRATERING: EFFECTS ON GROWTH OF EARTH'S CRUST

William K. Hartmann, Planetary Science Institute, Tucson, AZ 85719

Observations of crater densities in lunar regions of various ages show that the cratering rate during the first 500 my of lunar history was $\geq 10^3$ x the present rate (1). Because of O isotope ratios and other evidence, we believe the moon formed close to Earth, and the same early intense cratering affected Earth, too. Probably it was a solar-system-wide phenomenon, resulting from sweep-up of the planetesimals at the end of planet formation.

In all probability, the cratering rate in the first 10-100 my averaged vastly higher. The mass flux needed to accumulate Earth and the moon within the permitted formation interval of 10-100 my is of the order $2(10^9)$ x the present rate (2). This is entirely understandable from accretion models, which indicate a very rapid accretion of planetesimals (3) and a longer, declining rate of sweep-up of the remaining planetesimals after the planets gained nearly their present mass (4).

This situation has two related effects on formation of crusts of Earth and other planets. First, as pointed out by Safronov (5) and Hartmann and Davis (6), the 2nd, 3rd, ... largest bodies accompanying the planets may have been large enough, relative to the planets themselves, to have dramatic effects, including disruption of large volumes of the planet, including mantle and crust if they had formed. Hartmann and Davis (6) attributed the moon's origin to such a giant, disruptive impact. Recent modelling (7,8) shows that large regions of any pre-existing crust and upper mantle could be vaporized and/or ejected as finely disseminated and heated dust, with consequent geochemical effects. Giant impacts would be stocastic events, dotted through the first 10^8 years or so, until the required large bodies collided with planets or were ejected from the solar system.

The second effect involves the more continual rain of impactors that were smaller (but large by present day standards). These would form a continuum with the "giant impacts." (The nature of the impactor size distribution was such that the smaller the body, the more of them.) Ringwood (9) discussed the possible production of a silicate atmospheres devolatized from the crust. Frey (10) proposed that large impacts punched holes in the early crust, piling up "continental" crustal ejected debris in other areas and exposing hot mantle areas where convection was enhanced; this could have abetted proto-continent formation. Hartmann noted that magma ocean evolution must be modelled in the presence of this process, which competes with crust formation by continually breaking up and redistributing the early, solid, anorthositic surface (11). Also, the impact rate at the close of planet forming period was high enough that impacts comparable to the proposed K-T boundary event happened on roughly a monthly-to-yearly basis (12).

Figure 1 (adapted from Fig. 12) illustrates some of these points. Curve "t=0" shows the approximate impact rate shortly after the close of

planet formation; curve "t=500 m.y." shows the rate 500 m.y. after planet
formation. The curves are based on the above results. The actual impact
rate declined approximately exponentially with time from the first to the
second curve. As can be seen, during this interval there was opportunity
for a few giant impacts, many impacts large enough to create basins
comparable to the moon's Orientale and Imbrium basins (10^3-km-scale in
diameter of disrupted zone), and thousands of smaller-scale craters.

Depending on the time-scale of crustal formation and evolution on
Earth and other planets, models of proto-crustal evolution should take
into account the possible competitive influence of repeated impact
cratering, which would disrupt any hypothetically homogeneous
proto-crustal layer, creating thick and thin spots, and affecting cooling
timescales and global- or continental-scale topographic/structural/
tectonic features.

While the geographic expression of these effects may be long
vanished on Earth because of tectonic and erosive effects, they may be
still visible on Mars, where large impact basins may be detected, and
where relatively young volcanics dominant one hemisphere, while a much
older crustal surface dominates the other hemisphere. (13)
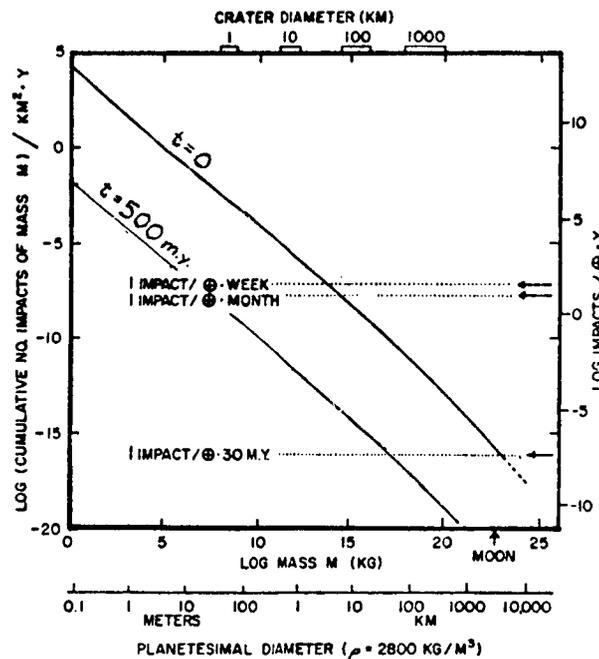
FIG. 1. Estimated impact cratering rate on early Earth at close of
planet formation and 500 m.y. later. These rates are on the order of 10^9
and 10^3 times the present rate, respectively. (Adapted from ref. 12).

REFERENCES

1. Hartmann, W.K. (1972). Astrophys. Space Sci. **12**, 48-64.
2. Hartmann, W.K. (1980). Icarus **44**, 441-453
3. Greenberg, R., Wacker, J., Hartmann, W.K., and Chapman, C. (1978). Icarus **35**, 1-26.
4. Wetherill, G.W. (1977). Proc. Lunar Sci. Conf. 8th, 1-16.
5. Safronov, V. (1966). Sov. Astron. AJ **9**, 987.
6. Hartmann, W.K. and Davis, D.R. (1975). Icarus **24**, 504-515.
7. Benz, W., Slattery, W., Cameron, A. (1986). In Origin of the Moon (Houston: Lunar & Planetary Inst.).
8. Melosh, H. and Sonett, C.P. (1986). In Origin of the Moon (Houston: Lunar & Planetary Inst.).
9. Ringwood, A.E. (1979). Origin of the Earth and Moon. (Springer-Verlag, New York), p. 295.
10. Frey, H. (1977). Icarus **32**, 235.
11. Hartmann, W.K. (1980). In Proc. Conf. Lunar Highlands Crust (Houston: Lunar & Planetary Institute).
12. Hartmann, W.K. (1986). In Origin of the Moon (Houston: Lunar & Planetary Inst.).
13. Wilhelms, D.E. and Squyres, S.W. (1984). Nature **309**, 138-140.

THE GROWTH OF CONTINENTS AND SOME CONSEQUENCES SINCE 1.5 Ga

David G. Howell, U.S. Geological Survey, Menlo Park, Ca. 94025

The growth history of continents remains a controversial topic. The subject is particularly confused because continents contain recycled crustal rock in addition to first generation mantle derivatives. The proliferation of Sm/Nd data is helping to allay this problem as these data provide the timing when the host rock hatched from the mantle, regardless of all subsequent crustal processes (e.g. weathering or melting and recrystallization). Furthermore, an assemblage of such data from a cratonal area can also provide the percentages of rock that differentiated from the mantle at varying times. As an example, Patchett and Arndt (1986) indicate that the Churchill Province of the Canadian shield underwent a tectono-thermal event 1.9-1.7 Ga, yet 95 percent of the rock hatched from the mantle before 2.7 Ga. This is contrasted with the mid-continent region of the United States where again the rocks indicate a thermal event 1.9-1.7 Ga, but in this instance the crystallization represents a first cycling out of the mantle. The body of data presently available suggest that 50 percent of the modern volume of continents formed by 2.5 Ga, 80 percent by 1.5 Ga (Nelson and DePaolo, 1985). The last 20 percent of the continental crustal volume has been added since the mid-Proterozoic, yielding an average net rate of 1 km^3/y.

A rate of 1 km^3/y for continental growth appears reasonable based on estimates of the Mesozoic and Cenozoic budget for continental denudation, island arc and oceanic island volcanism, and the efficacy of accretion and underplating within subduction zones. Reymer and Schubert (1984) have calculated a global average production of approximately 1.35 km^3/y for newly formed volcanic-crustal rock. The total volume of sediment on the world's ocean crust (91 x $10^6 km^3$, exclusive of the Arctic ocean deposits and after subtracting volumes due to volcanic ash and porosity) divided by the average age of the crust (55 Ma) indicates a effective denudation rate of 1.65 km^3/y. In order to determine if continents are growing, one must factor in the efficiency of subduction accretion, i.e., how much of the material riding on an ocean plate is accreted versus the amount recycled back into the mantle. Earthquake focal solutions coupled with a growing number of well imaged subduction zones, in concert with accurate plate kinematic reconstructions, are now permitting an assessment of both the total volume of material entering a subduction zone and the volume of material that has become accreted. A preliminary appraisal of data from the Aleutian and Lesser Antilles arc systems indicates an accretion efficiency between 75 and 100 percent. Therefore, on a global basis, the gross yearly accretion should fall in the range between 2.25 and 3.0 km^3, and of this, a net of 0.6 to 1.35 km^3/y would constitute new crustal mass.

The current volume of continental crust is ca. 7.6 x $10^9 km^3$. If 20 percent of this is younger than 1.5 Ga then the mid Proterozoic volume was 6.1 x $10^9 km^3$. Inferring that the surface area of the earth has remained constant and also assuming constancy for both the thickness of continents (based on the fundamental strength of quartz) and the volume of sea water (based on the chemical constancy for at least the last 1 Ga, nonetheless, the volume constancy remains very much an assumption), one can predict changes in continental freeboard. However, stratigraphic and paleogeographic considerations imply that the freeboard has not changed! A possible solution to this dilemma, while maintaining the assumptions above, is to increase the length of the oceanic spreading ridges in order to displace enough sea water to maintain a constant freeboard. From the dimensions above it follows that the area of ocean crust would increase from its modern area of 3.1 x 10^8 to 3.5 x $10^8 km^2$, with a consequent reduction in the height of the sea water column by about 500 m. Therefore, to maintain a constant freeboard, we must

displace a volume of sea water equal to $1.75 \times 10^8 km^3$ (0.5 km x $3.5 \times 10^8 km^2$).

Dividing the modern area of the ocean crust by the modern ridge length of 5.6×10^4km provides an average ocean width, and when this is divided by 55 Ma, an average ocean crustal spreading rate of 5 cm/yr is calculated. With this spreading rate, an average cross-sectional area (4200 km^2) can be determined by using the depth versus age curves of Sclater and others (1971). This area divided into the additional volume of the ocean basin that presumably existed 1.5 Ga ($1.75 \times 10^8 km^3$) indicates that an additional 4.2×10^4km of ridge is required to maintain a constant freeboard, i.e., ridge length would have been 1.75 times as long as the modern ridge length. Obviously, added to the list of assumptions is the requirement that the average spreading rates in the Proterozoic were the same as they have been for the past 100 Ma, and preliminary paleomagnetic data support this assertion (McWilliams, oral comm. 1987).

The various parameters calculated above allow an additional determination. Hargraves (1986) provides a simple equation that relates heat flux from the ocean crust to its area and the ridge length:

$$Q(tm) = 17.3\ A^{2/3}\ L^{1/3},$$ where A and L are in km, and the answer is in watts.

From this, one calculates a heat flux for the mid-Proterozoic of 39.5×10^{12} watts, which is 1.36 times the modern heat flux. This determination is reasonable given the nature of radioactive decay, but this author is not able to make an indepth evaluation of this parameter.

References:

Hargraves, R.B., 1986, Geology, v. 14, no. 9, p. 750-753.
Nelson, B.K. and D.J. DePaolo, 1985, Geol. Soc. Amer. Bull., v. 96, p. 746-754.
Patchett, J.P. and N.T. Arndt, 1986, Earth and Plan. Sci. Ltrs., v. 78, p. 329-338.
Reymer, A. and G. Schubert, 1984, Tectonics, v. 3, no. 1, p. 63-77.

THE EARLY ARCHAEAN CRUSTAL HISTORY OF WEST GREENLAND AS RECORDED BY DETRITAL ZIRCONS

P.D. Kinny[1], W. Compston[1] and V.R. McGregor[2]

1. Research School of Earth Sciences, Australian National University, Canberra.
2. Atammik, 3912 Sukkertoppen, Greenland.

Continents generate felsic rocks, and felsic rocks crystallize with accessory zircon. Other primary rock-types which can be zircon-bearing include gabbros, anorthosites and even kimberlites, but in general only felsic rocks contain zircons in abundance. The presence of abundant detrital zircons in clastic sedimentary rocks is therefore an indication that continental crust is (or was) nearby, while the isotopic clocks within each individual zircon grain record the ages of the principal components from the crust. Employing the ion probe for rapid and precise isotopic analysis of individual zircon grains (1) we are thus able to obtain information about the Earth's early crust, **whether or not that crust still exists or is presently exposed on the Earth's surface.**

Mid-Archaean sediments from the high-grade gneiss terrain of Western Australia's Yilgarn Block have yielded zircons with U–Pb ages over 4100 Ma; currently the Earth's oldest-known minerals (2,3). These sediments were deposited *ca* 3100 Ma ago, over one billion years after the time of formation of the old zircons. The high relative abundance of old zircons in the Jack Hills sequence (12%, 3) implies a fertile, voluminous source which we interpret as felsic continental crust, especially considering that the zircons possibly were reworked from older sediments. If significantly older clastic sediments were still present in the region, they presumably would provide a more comprehensive record of felsic components from this early crust.

We have begun to apply a similar approach to identify early crustal components in the Archaean craton of southern Greenland. Here, the earliest recognised supracrustal components: the Isua belt and the Akilia association, are entrained within and intruded by the Amîtsoq gneisses, all of which are cut by Ameralik dykes. The Malene supracrustals postdate the dykes and are intruded by the mid-late Archaean Nûk gneisses. The Isua sequence has been dated precisely at 3807 ± 2 Ma, through U–Pb analyses of igneous zircon from deformed clastic rocks of volcanic origin (4). The Akilia rocks are widely considered to be contemporaneous, despite younger zircon ages of *ca* 3600 Ma obtained from samples of fine-grained deformed metasediments (5). Either these zircons have been isotopically reset or they do not include a significant detrital component, or the detrital component is masked by younger high-U grains. Either way, the true age of the Akilias is unknown, and awaits the discovery of igneous zircon in volcanic or intrusive phases such as have been used to date Isua. Since it now appears that an Amîtsoq gneiss dated at 3820 Ma (6) probably is representative of a wider exposure of tonalitic Amîtsoq gneisses with Akilia inclusions on the islands south of the mouth of the Ameralik fjord, some of these Akilias probably are older than the Isua supracrustal belt (situated 100 km inland at Isukasia).

Using the ion probe we have analysed a suite of zircons from a banded fuchsite quartzite exposure atop the divide between Ameralik and Godthåbsfjord, 15 km northeast of the Qôrqut hotel. Close inspection of the outcrop (immediately after the Oxford meeting) showed that the unit, which attains a maximum thickness of 30m, lies conformably between a sheet of Malene gabbroic amphibolite and a homogeneous unit of Nûk *(s.l.)* gneiss (Nutman and Friend, *pers. comm.*). The zircons, which give early Archaean ages, appear in thin section to be concentrated in heavy-mineral rich bands, indicating that this unit probably is a true meta-sandstone rather than a unit of recrystallized mylonite or of vein quartz (*c.f.* 7). The zircon population includes

numerous grains with fine igneous-type zoning and apatite inclusions, and many with distinctly pitted surfaces implying a probable detrital origin. Remarkably, these textures have been preserved despite metamorphism in the late Archaean to amphibolite facies (*ca* 2800 Ma). Uranium contents range from 20 to 1100 ppm. ^{207}Pb/^{206}Pb ages (45 analyses) range from 2870 to 3890 Ma. ^{207}Pb/^{206}Pb and U content are anti-correlated only for spots with over 300 ppm U, reflecting the effects of Pb loss from these areas during the late Archaean metamorphism. The bulk of the low-U areas appear to have been relatively undisturbed at that time. Most of the low-U analyses are still near-concordant (Fig. 1), and rather than there being a continuum of ^{207}Pb/^{206}Pb ages, there are obvious peaks and gaps in the age spectrum (Fig's 1,2). The three apparently oldest grains were each analysed in two places, and in every case the more discordant spot has a lower ^{207}Pb/^{206}Pb, so minor Pb loss has occurred at some stage in the past from parts of all grains. The principal age population, representing 45% of all analysed grains, has ^{207}Pb/^{206}Pb ages ranging from 3640 to 3690 Ma, corresponding to the intrusion ages of the Amîtsoq augen gneisses of outer Ameralik, and of the majority of Amîtsoq gneisses at Isukasia (*c.f.* 8). Three grains which are distinctively low in U (20 – 35 ppm) give a very much older minimum age of 3774 ± 36 Ma. Three others are still older at 3800 ± 13 Ma, comparable to the age of zircons from the Isua supracrustal belt (4). One grain (#21) gives a minimum age of 3837 ± 31 Ma, and the oldest two grains (#3,#16) give 3883 ± 14 Ma (all 2σ). The discrete groupings of low-U age populations argues against major early Pb loss. All grains in the quartzite appear to be detrital, and the low-U parts of those grains preserve ^{207}Pb/^{206}Pb ages close to their original crystallization ages.

The age of deposition of the quartzite certainly is younger than *ca* 3630 Ma, the age of the youngest detrital zircons. The population of zircon ages is precisely what would be expected from a clastic Malene sequence deposited on or adjacent to a continental basement of Amîtsoq gneisses, prior to the time of exposure (and hence probably formation) of Nûk gneisses. There are no Nûk-age zircons present. The age spectrum resembles most closely the known early Archaean components of the outer Godthåbsfjord/Ameralik terrain plus some additional older phases (Fig. 2). The oldest detrital zircons in the quartzite are older than any known exposed rocks in the region and comparable to the oldest zircon cores in the Uivak gneisses of Labrador (9, L.Schiøtte, *unpubl.*). Apparently, the oldest felsic crust exposed within the range of provenance of the quartzite at the time of deposition was formed *ca* 3900 Ma ago. This may indeed represent the first felsic crust in the region, although a more confident conclusion awaits the analysis of detrital zircons from the earlier Isua and Akilia sequences. Other remnant continental nucleii apparently diffentiated from the mantle at earlier times (2,3,10). The popular notion of a major crust-forming episode at 3800 Ma, as has been adopted for many crustal growth models, is no longer defensible.

(1) Compston W., Williams I.S. and Meyer C. (1984) J. Geophys. Res. 89 Suppl. B p. 525–534.
(2) Froude D.O., Ireland T.R., Kinny P.D., Williams I.S., Compston W., Williams I.R. and Myers J.S. (1983) Nature 304 p. 616–618.
(3) Compston W. and Pidgeon R.T. (1986) Nature 321 p. 766–769.
(4) Compston W., Kinny P.D., Williams I.S. and Foster J.J. (1986) Earth Planet. Sci. Lett. 80 p. 71–81.
(5) Baadsgaard H., Nutman A.P., Bridgwater D., Rosing M., McGregor V.R. and Allaart J.H. (1984) Earth Planet Sci. Lett. 68 p. 221–228.
(6) Kinny P.D. (1986) Earth Planet. Sci. Lett. 79 p. 337–347.
(7) Friend C.R.L., Nutman A.P. and McGregor V.R. (1987) J. Geol. Soc., London 144 p. 369–376.
(8) Baadsgaard H., Nutman A.P. and Bridgwater D. (1986) Geochim. Cosmochim. Acta 50 p. 2173–2183.

(9) Collerson K.D. (1983) <u>LPI Technical Report 83-03</u> p. 28–33.
(10) Black L.P., Williams I.S. and Compston W. (1986) <u>Contrib. Mineral. Petrol. 94</u> p. 427–437.

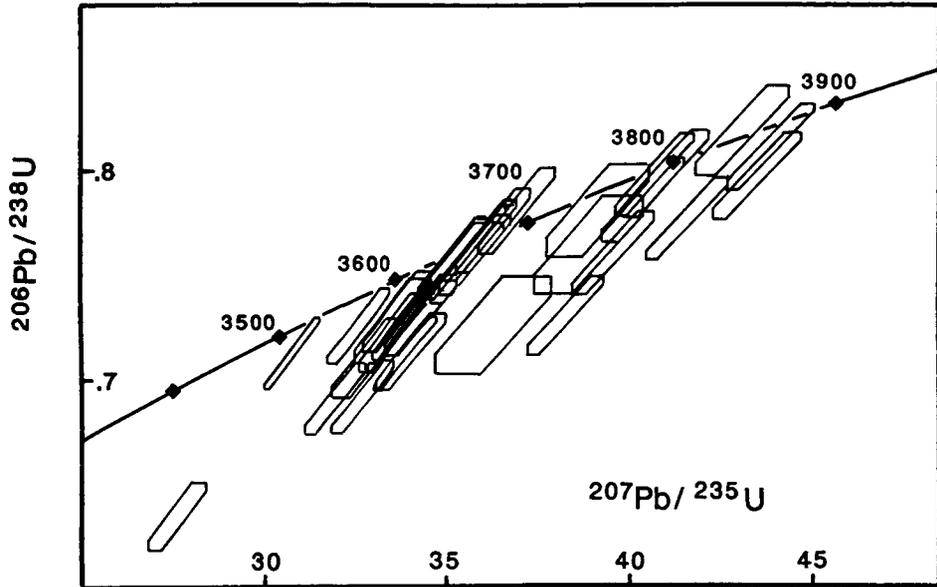Figure 1 : Concordia diagram showing ion probe U–Pb analyses of low-U parts of zircons from banded fuchsite-bearing quartzite, 15km northeast of the Qôrqut hotel, West Greenland. 1σ error boxes. Ages in Ma.
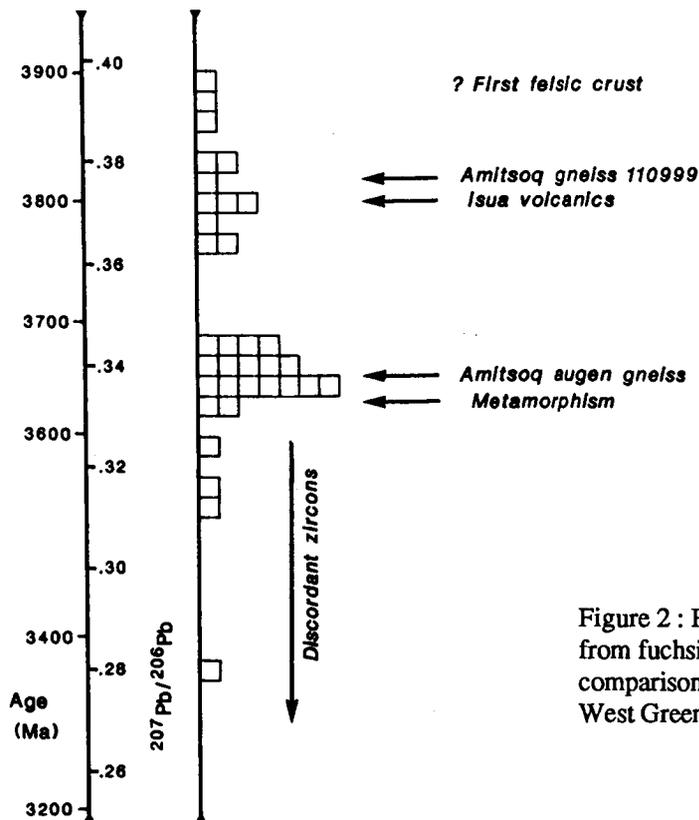
Figure 2 : Histogram of $^{207}Pb/^{206}Pb$ ages of zircons from fuchsitic quartzite, near Qôrqut. Shown for comparison are known ages of early Archaean crust in West Greenland (4,6, unpublished ion probe data).

DEEP SEISMIC REFLECTION PROFILING AND CONTINENTAL GROWTH CURVES

Simon L. Klemperer, British Institutions' Reflection Profiling Syndicate (BIRPS),
Bullard Laboratories, Madingley Rise, Madingley Road, Cambridge, CB3 0EZ, U.K.

Deep seismic reflection profiling over large areas of western Europe, North America and Australia shows that the lower crust is layered on a vertical scale of a few tens to a few hundreds of metres and is seismically quite distinct from the upper crust and upper mantle. The layering is detected beneath many different terranes of varying crustal type and basement age, from Archaean to Cenozoic, and is often apparently younger than the tectonic or metamorphic age of the exposed upper crust. Continental growth curves estimated from upper crustal measurements alone may underestimate Phanerozoic crustal additions, and mean crustal ages similarly estimated may be too high.

The reflective lower crust

Reflection profiling is the highest resolution technique currently available for the study of the lower continental crust. Objects only a hundred metres thick, a couple of kilometres in diameter and a few percent different in density and seismic velocity from the surrounding medium are detectable at the Moho. Bodies of these dimensions and up to at least twenty kilometres in length are commonly detected in the lower crust but only rarely in the upper crust or the upper mantle. A reductionist synthesis of the global results of crustal reflection profiling, shown in fig. 1 (after [1]), emphasises the pronounced variation in reflectivity pattern with depth. Beneath an upper crust that is non-reflective apart from sedimentary reflections and occasional dipping fault-plane reflections is the strongly reflective lower crust and the normally unreflective upper mantle. Though not ubiquitously applicable, and though certainly only one possible approach to the understanding of deep profiles (for more complex subdivisions of crustal reflective types see e.g. [2,3]), many profiles from North America, western Europe and Australia [4,5] show this characteristic pattern. (Other continents are as yet largely unstudied.)
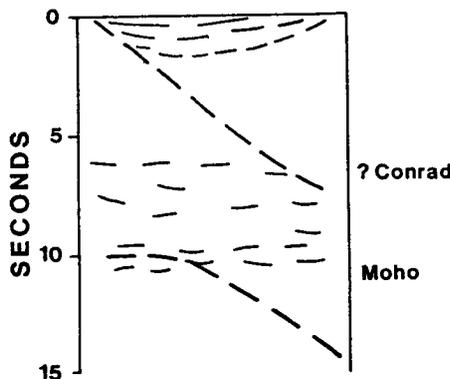
Three examples of the 10,000 km of data collected by BIRPS in the shallow epicontinental seas around Britain (fig. 2) are shown (figs. 3, 4 and 5). These data are from the Archaean Lewisian province of northwest Scotland [6] (fig. 3); from the North Sea in the area of Caledonide Early Palaeozoic continental collision and terrane accretion [7] (fig. 4); and from the Late Palaeozoic Hercynian orogen in the English Channel [8] (fig. 5). In all three cases the reflective lower crust is clearly distinct from the upper crust and from the upper mantle, and the transition between these zones is fairly sharp. Even in the Lewisian, where granulites once metamorphosed at pressures typical of the lower crust (e.g. [9]) are now exposed at the surface and comprise the upper crust, the present-day lower crust is far more reflective than the present-day upper crust. Clearly the Lewisian granulites (and, by the same test, some other granulite terranes e.g. [10]) are not typical of in situ lower crust, though whether by virtue of their original genesis or as a result of changes produced by late-stage uplift or deformation, is uncertain.

Despite the variable origins and complex history of the basement to Britain, the crustal reflection data do not show marked differences across terrane boundaries.

A TYPICAL BIRP
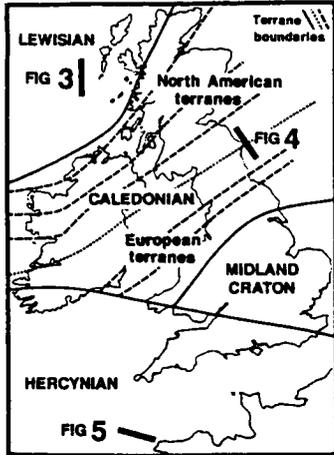
Figure 2   **Figure 3**   **Figure 4**   **Figure 5**

LEWISIAN   CALEDONIAN   HERCYNIAN
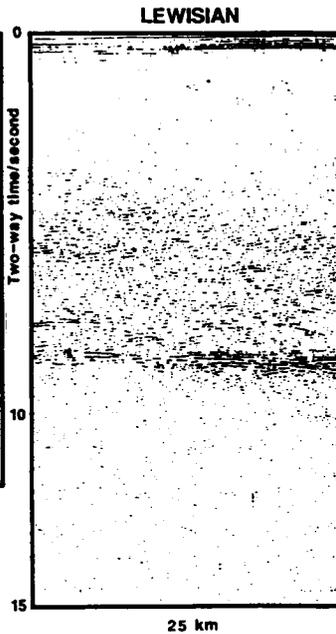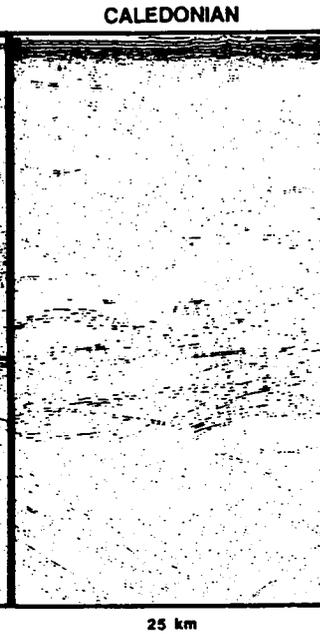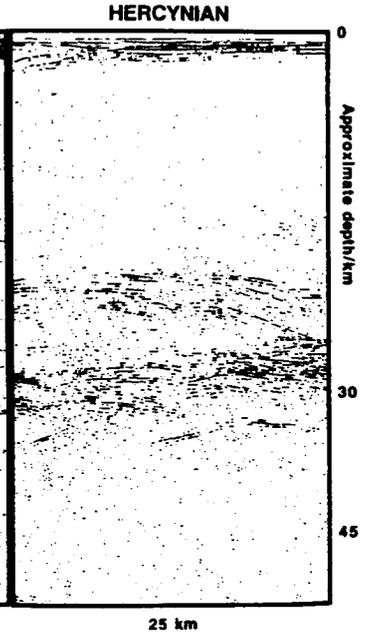
No gross, systematic variation in reflectivity patterns has been noted that can be
related to the Precambrian and Palaeozoic age structure of Britain. The existence of
similar reflectivity patterns in adjacent but disparate terranes strongly suggests
that the reflectors are younger than the age of accretion and younger than the age
of the exposed basement. The reflectors in the lower crust may have formed over
wide areas in a single event [11], or the reflectivity pattern might represent a
process that occurs in extended or multiple periods, at least since the Palaeozoic
European orogenies, and possibly continues at the present. There are at least two
separate pieces of evidence for continuous modification of the lower crust. First,
the evidence of reflection and refraction profiles that crustal roots no longer exist
beneath the Caledonides and other Palaeozoic mountain belts implies that lower
crustal creep or flow has destroyed these roots [12]. This argument suggests that
the whole of the lower crust, rather than the reflectors in particular, is undergoing
continual modification. Second, suggestions that the depth [13] and the thickness
[14] of the reflective lower crust are correlated with present-day heat flow imply
equilibration of crustal reflective structure with tectonothermal activity on time-
scales of only a few tens of million years.

Significance of lower crustal reflectors for crustal growth curves

Though the relative youth of the reflective lower crust may be accepted, it
remains an open question whether the reflectors represent crustal growth or crustal
reworking. Hypotheses to explain the lower crustal reflections include mafic sills
or layered igneous intrusions [15]; extensional or compressional strain-banding or
mylonite zones developed in the 'ductile' lower crust [16]; and laminae of hydrated
rocks [17] or open fluid-filled cracks [18]. New igneous intrusions would clearly
represent crustal growth, whereas strain-related or fluid-related reflectors would
represent tectonic or metamorphic crustal reworking episodes. Though not directly
indicative of crustal growth, such tectonothermal events in the lower crust might
well be accompanied by intrusion from the mantle even if these intrusions are not
directly detected as reflectors.

Although the nature of the reflective lower crust is still controversial, these
reflectors are important since they represent a very significant volume of material.
If, very crudely, 50% of the crust shows strong seismic layering in its lower half,

and if the reflectors correspond to half the volume of the lower crust, then one-eighth of all continental crust ($\simeq 10^9$ km^3) may have been strongly reworked (if the layering is tectonic or metamorphic) or extracted from the mantle (if intrusive) at ages much younger than those of the exposed upper crust.

An example: Basin & Range province, western U.S.A.

Though the reflection database is not yet sufficiently extensive nor sufficiently understood to allow more than generalisations on a global scale, in at least one area, the Basin and Range province, seismic reflection data strongly suggest crustal growth. COCORP reflection profiling has been used to show that the reflectors at the Moho and at 3 to 6 km above the Moho are younger (based on usual geological criteria of cross-cutting relationships) than Palaeozoic or Mesozoic accretionary episodes [19]. These reflectors probably represent Neogene syn-extensional intrusive rift pillows produced by mantle melting. Assuming a conductive geotherm, the temperature at the Moho during the Neogene has been close to the basalt solidus [20]. The Moho is nearly flat even beneath regions with highly variable upper crustal extension, implying either rapid lower-crustal flow or, more probably, in view of the extensive coaeval volcanism, crustal additions of 5 to 10 km from the mantle to restore the crust to constant thickness [21]. These direct estimates of crustal additions are in agreement with calculations of the volume of mantle melt generated by extension of the lithosphere [22]. Though there may have been Neogene crustal additions of 15% to 30% ($\simeq 0.5$ to 1.0×10^7 km^3 in 20 Ma, or 0.25 to 0.5 km^3.a^{-1} in this area alone), Precambrian basement is present beneath the eastern Basin and Range province and this area is conventionally regarded as being of $>$ 1.7 Ga crustal age [23].

Conclusions

The Law of Superposition is wrong on a gross scale: the crust often becomes younger with increasing depth. Ages of exposed rocks may be overestimates of mean crustal age, and unless crustal recycling by delamination is an efficient and widespread process Phanerozoic crustal growth may be faster than the often quoted value of 1 km^3.a^{-1}.

References

[1] Matthews D.H., Cheadle M.J., in: Reflection seismology: A global perspective, M. Barazangi, L. Brown, eds., Am. Geophys. Un. Geodynamics Series 13 5-19, 1986. [2] Allmendinger R.W., Nelson K.D., Potter C.J., Barazangi M., Brown L.D., Oliver J., Geology, 15 304-310, 1987. [3] McGeary S., Geophys. J. Roy. Astr. Soc., 89 231-238, 1987. [4] Barazangi M., Brown L., eds., Reflection seismology: a global perspective, and Reflection seismology: the continental crust, Am. Geophys. Un. Geodynamics Series 13 and 14, 1986. [5] Matthews D.H., Smith C.A., eds., Deep seismic reflection profiling of the continental lithosphere, Geophys. J. Roy. Astr. Soc., 89 1-447, 1987. [6] Brewer J.A., Smythe D.K., Tectonics, 5 171-194, 1986. [7] Klemperer S.L., Matthews D.H., Geology, 15 195-198, 1987. [8] BIRPS, ECORS, Jl. Geol. Soc. Lond. 143 45-52, 1986. [9] Weaver B.L., Tarney J., Earth Planet. Sci. Letts., 55 171-180, 1981. [10] Klemperer S.L., Brown L.D., Oliver J.E., Ando C.J., Czuchra B.L., Kaufman S., Can. J. Earth Sci., 22 141-153, 1985. [11] Bois C., Cazes M., Hirn A., Matte P., Mascle A., Montadert L., Pinet B., Geophys. J. Astr. Soc., 89 297-304, 1987. [12] Meissner R., Wever T., Flüh E.R., Annal. Geophys., 5B 357-364, 1987. [13] Klemperer S.L., J. Geophys., 61 1-11, 1987. [14] Wever Th., Trappe H., Meissner R., Annales Geophysicae, 5B 255-266, 1987. [15] Meissner R., Geophysical Surveys 1 195-216, 1973. [16] Phinney R.A., Jurdy D.M., Geophysics, 44 1637-160, 1979. [17] Hall J., in: Reflection seismology: The continental crust, M. Barazangi, L. Brown eds.: Am. Geophys. Un. Geodynamics Series 14 223-231, 1986. [18] Matthews D.H., in: The nature of the lower continental crust, J.B. Dawson, D.A. Carswell, J. Hall, K.H. Wedepohl, eds. Spec. Publ. Geol. Soc. Lond. 24 11-22, 1986. [19] Klemperer S.L., Hauge T.A., Hauser E.C., Oliver J.E., Potter C.J., Geol. Soc. Am. Bull. 97 603-618, 1986. [20] Lachenbruch A.H., Sass J.H., in The Earth's Crust, ed. J.G. Heacock, Am. Geophys. Monog. 20 626-675, 1977. [21] Gans P.B., Tectonics, 6 1-12, 1987. [22] McKenzie D.P., Bickle M.J., The volume and composition of melt generated by extension of the continental crust, J. Petrol., submitted. [23] Farmer G.L., DePaolo D.J., J. Geophys. Res., 88 3379-3401, 1983.

GROWTH OF EARLY ARCHAEAN CRUST IN THE ANCIENT GNEISS COMPLEX OF SWAZILAND AND ADJACENT BARBERTON GREENSTONE BELT, SOUTHERN AFRICA

A. KRÖNER (University of Mainz, FRG), W. COMPSTON (RSES, Australian National University, Canberra), A. TEGTMEYER (Finnigan-MAT, Bremen, FRG), C. MILISENDA (University of Mainz) and T.C. LIEW (Max-Planck-Institut für Chemie, Mainz)

Models for crustal growth in early Archaean terrains center largely on the relationships between 3 distinct rock assemblages: (a) tonalitic-trondhjemitic gneisses, often polydeformed, at high metamorphic grade and associated with mafic dykes and shallow-water supracrustal sequences, (b) greenstone belts with mafic-ultramafic to felsic volcanics and clastic sediments, and (c) high-level TTG and granite plutons that intrude the greenstone belt rocks, often resulting in intricate contact relationships. These 3 assemblages are well exposed in the SE Kaapvaal craton of southern Africa, and despite many detailed studies their tectonic setting and geodynamic relationships are still ambiguous (1). We present new geochronological and isotopic data for rocks of the Ancient Gneiss Complex (AGC) of Swaziland and the adjacent Barberton Greenstone Belt (BGB) that clarify many of the controversial questions and that enable us to document a more complex history of early Archaean crustal growth in the Kaapvaal craton than previously assumed.

The mafic to felsic metavolcanics of the Onverwacht Group in the BGB have so far been considered to be the oldest rocks in the region on account of Sm-Nd whole-rock isochrons with ages of 3530 ± 50 Ma(2) and 3560 ± 240 Ma(3). Whole-rock Pb-Pb and Sm-Nd data on cogenetic mafic to ultramafic rocks alone, however, suggest a younger age of ≈3450 Ma (4,5), and this is supported by our single zircon 207Pb/206Pb ages derived from direct grain evaporation (6): Detrital zircons from a metaquartzite underlying the Onverwacht volcanics in NW Swaziland yield tightly clustered ages with a mean of 3456 ± 4 Ma while an acid volcanic flow from the Komati Formation has a zircon age of 3438 ± 2 Ma. These data bracket the Onverwacht mafic volcanic activity between 3438 and 3456 Ma and suggest that the earlier Sm-Nd ages are too old, most likely as a result of contamination with older continental crust and/or combination of genetically unrelated samples in the dated material.

Detrital zircons from a sample of metagreywacke in the lowermost stratigraphic unit of the Fig Tree Group have suprisingly uniform U-Pb ages (SHRIMP data) that combine to 3453 ± 9 Ma (7) and suggest an extremely uniform sialic source region to the SE of the original depository, perhaps identical to that sampled by the Onverwacht metaquartzite. It is also possible, however, that these zircons are derived from weathering of Onverwacht acidic volcanic material, but the greywacke geochemistry favours a TTG source.

Granitoid pebbles and gneiss clasts from a conglomerate in the Moodies Group overlying the Fig Tree strata display geochemical features of highly evolved continental crust (8), and SHRIMP U-Pb zircon ages between 3570 ± 6 Ma and 3518 ± 11 Ma have been reported (7). Because of the small errors in age these data conclusively demonstrate the existence of evolved sialic crust in the source region of the Moodies sediments prior to the generation of the Onverwacht volcanics, and both the ages and geochemical signature rule out derivation of these granitoids from Onverwacht-type volcanic precursors. The entire BGS seqence is deformed by the Kaapvalley and Stentor TTG plutons with U-Pb zircon ages of 3229 ± 5 and $3347+67/-60$ Ma respectively (9). The Moodies Group must therefore be older than these ages, and deposition and early deformation in the BGB took place within a period of less than 100 Ma.

Published age and isotopic data for rocks of the AGC are in the range 3417±34 Ma to 3555±111 Ma (10,11) and, combined with the apparent primitive isotopic systematics (ϵNd(T)$\approx$0, 87Sr/86Sr initial ratio $\approx$0.700), this was interpreted to mean that the AGC either formed from mantle-derived melts with very short crustal residence times at about the same time as the BGB volcanics or that at least some of the TTG rocks in the AGC were derived from deep crustal melting of BGB mafic volcanics (10,12).

We have found considerable compositional, structural, metamorphic and geo-chronologic heterogeneity in the AGC that no longer justifies this complex to be considered as a simple early Archaean gneiss terrain. The oldest rocks appear to be strongly flattened tonalitic orthogneisses which are in faulted contact with BGB strata in NW Swaziland and reveal a complex history as docu-mented by single grain and grain-domain U-Pb data (13). The earliest component at 3644±4 Ma, that we interpret as the crystallization age, is preserved in zircon cores that are surrounded by younger growth dated at 3504±6 Ma and 3433±8 Ma. Some of this growth can be related to new formation of distinctly 'flat' zircon grains that are aligned within the foliation of the gneiss, and we relate this growth to the metamorphic/structural event that transformed the original tonalite pluton into an grey gneiss. A still younger age of $\approx$3160 Ma is recorded in some zircons and is equivalent to a further magmatic/metamor-phic event during which leucocratic orthogneisses formed in the AGC of NW Swaziland that now occur only a few km N of the above tonalitic gneiss.

In NE and central Swaziland banded tonalitic gneisses similar to that discussed above yielded single-grain zircon ages of 3450-3556 Ma and have ϵNd(T) around 0. We suggest that crust of this type was already widespread by the time when the Barberton greenstones formed and was sampled by the Fig Tree and Moodies sediments.

In SW Swaziland TTG gneisses of the AGC contain a large infolded greenstone belt remnant known as Dwalile sequence and consisting of tightly folded upper amphibolite grade metaquartzites, BIF, calc-silicate gneisses, amphibolites and serpentinites that were once komatiite lavas. Detrital zircons from the metaquartzites have single-grain ages between 3543±31 Ma and 3566±23 Ma, while whole-rock Sm-Nd data scatter significantly and do not provide precise age in-formation. The metaquartzites, together with metamorphic garnet, define an array indicating a metamorphic event at 3450±264 Ma ago while Nd isotopic sys-tematics and geochemistry for the metavolcanics indicate variable contamina-tion with the interbedded sedimentary material and/or older sialic crust. Negative ϵNd(3.5Ga) values between -0.3 and -3.7 for the metaquartzites and the zircon ages suggest that these rocks originate from a $\approx$3.55 Ga old granitoid source terrain that is itself derived from melting of still older crust. Our age data are in accord with structural observations (14) that imply deformation of AGC tonalitic gneisses prior to Dwalile deposition, and we infer from this that the greenstones were formed near or on tonalitic crust.

Also in central Swaziland the AGC contains the infolded Mahamba and Mhkondo sequences of shallow-water metasediments with metaquartzites, pelitic gneisses and BIF that are now in granulite grade (T=700-900°C, P=6-7.5 kb) and were metamorphosed at a crustal depth of about 25 km. Detrital zircons from these rocks have near-concordant ages between 3485±10 and $\approx$3400 Ma, and the two sequences may have formed from erosion of earlier AGC gneisses, perhaps at the same time when the Barberton greenstones accumulated farther NW. The granulite event remains undated but must have occured prior to 3.3 Ga ago (15).

The combined data from the various terrains and sequences within the AGC and the BGB reveal a complex history of growth and deformation for this early Archaean crustal segment between 3.64 and 3.0 Ga ago that is schematically

shown in Table 1. The magmatic and tectonic evolution of the AGC was closely related to that of the BGB and suggests that early Archaean greenstone and gneiss terrains formed in similar settings but at different crustal levels. We are unable to reconstruct the precise tectonic regime for each phase of evolution, but it seems certain that neither the Dwalile nor the Barberton greenstones evolved in an entirely oceanic environment as shown by their association with metaquartzites, and there seems little doubt that at least part of the granitoid terrain of the AGC is older than the greenstone sequences.

Table 1. Evolution of early Archaean crust in the southeastern Kaapvaal craton

Age	Ancient Gneiss Complex	Barberton Greenstone Belt
≈3000	Intrusion of high-level granites	Intrusion of high-level granites
≈3200	Intrusion of TTG suite and granites, strong deformation	Intrusion of granodiorite
≈3300	Intrusion of granodiorite	Intrusion of TTG diapiric plutons deformation
≈3350	Granulite metamorphism, deformation, thrusting, uplift	Deposition of Moodies Group
≈3400	Deposition of shallow-water Mahamba and Mkhondo sequences	Deposition of Fig Tree Group deformation
3430-3440	High-grade metamorphism, thrusting and uplift	Deposition of Onverwacht Group TTG intrusion, deformation
≈3500	Deposition of Dwalile greenstone sequence	
		?
3450-3556	Tonalite intrusion, metamorphism, deformation, intracrustal melting, potassic granite emplacement	
3644±4	Tonalite intrusion	

References: (1) Jahn, B.M. (in press), in Geologic evolution of the Barberton greenstone belt, Geol. Soc. America Mem. (2) Hamilton, P.J. et al. (1979), Nature, 279, 298-300. (3) Jahn, B.M. et al. (1982) Contrib. Mineral. Petrol., 80, 25-40. (4) Brévart, O. et al. (1986) Earth Planet Sci. Lett., 77, 293-302. (5) Chauvel, C. et al. (in press) Earth Planet. Sci. Lett. (6) Kober, B. (1986) Contrib. Mineral. Petrol., 93, 482-490. (7) Kröner, A. and Compston, W. (in press) Precambrian Res. (8) Reimer, T.O. et al. (1985), Precambrian Res., 29, 383-404. (9) Tegtmeyer A. and Kröner, A. (1987) Precambrian Res., 34 (in press) (10) Carlson, W. et al. (1983), Nature, 305, 701-704. (11) Barton, J.M. et al. (1980), Nature, 283, 756-758. (12) Anhaeusser, C.R. (1975) Ann. Rev. Earth Planet. Sci., 3, 31-53. (13) Compston, W. and Kröner, A. (in press), Earth Planet. Sci. Lett. (14) Jackson, M.P.A. (1984), in Precambrian tectonics illustrated, E. Schweizerbart, Stuttgart, 1-18. (15) Milisenda, C. and Kröner, A. (1986), Fortschr. Miner., Beih.1, 118.

DIAPIRS OF THE MEDITERRANEAN RIDGE: THE TECTONIC REGIME OF AN
INCIPIENT ACCRETED TERRANE. Y. Mart, National Institute of Oceanography,
Haifa 31080, Israel.

From Oman to Calabria, through the Anatolian and the Iranian mountain
chains, across Cyprus and the eastern Mediterranean, a regional tectonic
feature occurs, characterized by mountainous terranes, isoclinal folds,
overthrust and underthrust faults, nappes, abundant ophiolites and diapirs.
On land these structures include the Oman, Zagros, Taurus and Troodos
mountain ranges, whereas in the eastern Mediterranean Sea, the tectonic
regime led to the evolution of the Mediterranean ridge and the Hellenic
trenches.

The Mediterranean ridge is the predominant structure in the eastern
Mediterranean basin.   The ridge forms arcuate series of uplifted diapiric
features that are located in the Ionian Sea from Calabria to Cyprus, and are
separated from the Aegean plate by a series of deep, arcuate trenches, which
are the upper part of a complex subduction zone, plunging northeastwards.
The ridge and the trenches are the products of the active tectonic collision
between Africa and western Eurasia.   The ridge is considered to be a
sedimentary accretional wedge that accumulated due to the subduction of the
underlying Tethyan oceanic crust and is controlled by collisional
compressional stresses.   There is ground to presume that in the future the
tectonic collision processes are likely to merge Crete with the
Mediterranean ridge, leading to the evolution of a mountain chain that would
evolve from the present eastern Mediterranean ridge and would accrete to
southern Europe (1).

The occurrence of diapirs in the Mediterranean ridge stems mostly from
the massive deposition of salt and gypsum in the Mediterranean basin during
the late Miocene. The diapiric emplacement of the evaporitic sequence is
not obvious, because the mobilization of the salt beds and the initiation of
the diapiric upward flow are constrained by the relatively shallow thickness
of the Plio-Pleistocene sedimentary overburden and by the low heat flow that
prevails in the eastern Mediterranean. The diapirs consist also of early
Cretaceous shales as well as other gravitationally metastable strata (2)
which are less mobile than salt. Furthermore, the diapiric ascent should
have been impeded by the apparent compressional tectonic regime in the ridge
that would have led to thrust faulting, where the fault planes were tightly
packed and could hardly be considered as a preferred environment for the
emplacement of shallow diapirs. The numerous diapirs of the Mediterranean
ridge indicate therefore that the tectonic regime on the ridge is not
compressional.

Studies of subduction trenches and their surroundings show that shallow
ridges occur seaward of the trenches in many places. Series of rifts
develop at the crest of these ridges and migrate toward the subduction zone.
The coexistence of rifts, which are extensional structural features,
adjacent to trenches, which are regions of compressional tectonic regime,
occurs because as the oceanic crust approaches the proximity of the trench,
it is uplifted slightly prior to its subduction. This uplift leads to the
development of the the rifts near the oceanic trenches. It is presumed that
the plate subduction under the Hellenic trenches could have been associated

with the crustal uplift underneath the Mediterranean ridge, and that the occurrence of diapirs in the Ridge could possibly be considered extensional, developed as secondary extensional features in a region of compressional tectonic regime.

The collisional motion between the African and the Eurasian plates would further enhance accretion of sediments in the Mediterranean ridge, which would attain subaerial exposure, and eventually would become a mountain range accreted to southern Europe (1). The numerous diapirs of salt and shales that occur in the ridge would be commonplace features in the future accreted terrane, indicating an intermediate extensional phase in the tectonic history of the development of crustal growth due to terranes accretion.

References

1. Le Pichon, X. (1983) In: K.J. Hsu (Editor), Mountain Building Processes. Academic Press, London. pp. 201-211.

2. Ryan, W.B.F., Kastens, K.A. and Cita, M.B. (1982) Tectonophysics, 86, pp. 213-242.

THE TECTONIC SETTING OF THE SEYCHELLES, MASCARENE AND AMIRANTE PLATEAUS IN THE WESTERN EQUATORIAL INDIAN OCEAN.

Y. Mart, National Institute of Oceanography, Haifa 31080, Israel.

A system of marine plateaus occurs in the western equatorial Indian Ocean, forming an arcuate series of wide and shallow banks with small islands in places. In spite of the geographical proximity and the morphological similarity of the Seychelles Bank, Amirante Arc and Mascarene Plateau, their geological structure and origin contrast sharply. The Seychelles Bank forms a plateau founded on Precambrian granite basement, except for the western section of the Bank, where middle Eocene syenite, diorite and microgranite build the lithological foundations. The Seychelles Bank builds the northwestern part of the Mascarene Plateau, but the remainder of that Plateau, from Saya de Malha Bank to Mauritius Island, is founded on Paleocene basalts, and late Cretaceous basalts prevail in the Amirante Arc, located SW of the Seychelles.

The oceanic basins that surround the Seychelles - Amirante region are of various ages and reflect a complex seafloor spreading pattern. The Somali Basin to the northwest is of late Jurassic - early Cretaceous age and is associated with the separation of Madagascar from Africa. The Mascarene Basin to the southwest is of late Cretaceous to Paleocene age, and it was formed due to the break-up of India from Madagascar. The evolution of the northwestern Indian Ocean Basin started during the Paleocene, and the oceanic basin separating the Seychelles - Mascarene Plateau from the Chagos Plateau, and has been actively spreading since the latest Eocene.

The structural analysis of the Seychelles - Amirante - Mascarene region reflects the tectonic evolution of the western equatorial Indian Ocean. The development of the Seychelles started with the separation of the Seychelles - India block from Madagascar in the late Cretaceous, which formed the Mascarene Basin. During this stage, the Seychelles was an integral part of India, but the two separated in the early Paleocene, when the evolution of the NW Indian Ocean Basin started. There is evidence suggesting that the Seychelles - India break-up was associated with very intensive volcanic activity which is probably the origin of the Mascarene and Chagos oceanic plateaus and the Deccan Traps of India. Thus the NW Indian Ocean spreading center was located between the Seychelles, Mascarene and Chagos marine plateaus to the SW and India to the NE from the Paleocene to the late Eocene.

It is suggested that due to the seafloor spreading during this tectonic stage, the Seychelles continental block drifted southwestwards to collide with the oceanic crust of the Mascarene Basin, forming an elongated folded structure at first, and subsequently a subduction zone. This crustal subduction caused the evolution of the Amirante Arc and Trough, and led to the intrusion of syenite and microgranite into the Seychelles block. A subsequent jump of the NW Indian Ocean spreading center during the late Eocene started to separate the Chagos Plateau from the Mascarene Plateau, and consequently reduced the tectonic compression between the Seychelles Bank and the Mascarene Basin, thus terminating the last major tectonic event that affected the Seychelles region.

The morphological similarity, the lithological variability and the different origin of the Seychelles Bank, the Mascarene Plateau and the Amirante Arc emphasize the significant convergent effects of various plate tectonic processes on the development of marine plateaus.

EVIDENCE FOR CRUSTAL RECYCLING DURING THE ARCHEAN: THE PARENTAL
MAGMAS OF THE STILLWATER COMPLEX, <u>I. S. McCallum</u>, Department of Geological
Sciences, University of Washington, Seattle, WA 98195

Those few instances where Archean mafic and ultramafic igneous rocks are
preserved in their original state assume an unusual importance since they
provide a constraint on the composition and mineralogy of the Archean mantle
source regions and the extent to which these source regions represent prim-
ordial mantle, depleted or enriched mantle, or mantle containing a recycled
crustal component. The Stillwater Igneous Complex is a large Archean mafic/
ultramafic intrusion in which the original minerals, textures, and structures
are preserved. The complex is located along the northwestern margin of the
Beartooth Mountains which form part of the Wyoming Archean Province. Its
crystallization age, based on a Sm-Nd internal isochron, is 2.7 Ga (1). The
Stillwater magma was intruded at a depth of ≈ 10 km into a thick sequence of
metasedimentary rocks broadly similar to modern continental shelf/slope
sediments. The absence of internal deformation of the cumulate layering in
the complex suggests a relatively stable tectonic environment during emplace-
ment. A Proterozoic (1800-1600 Ma) low grade regional metamorphism has resul-
ted in the development of greenschist facies mineral assemblages along fract-
ures marking channelways of water infiltration. A more subtle effect of this
reheating is revealed in disturbances in Rb-Sr isotopic systematics due, most
likely, to the localized redistribution of Rb (1). The intrusion is cut by a
series of diabase dikes which have had no obvious effect on the original cum-
ulus minerals. Laramide uplift and mild deformation have tilted the complex
to the extent that the layering is now near vertical.

In order to evaluate the composition and history of the Archean mantle
from which the Stillwater magmas were generated, it is necessary to determine
the compositions of the parental magmas. This has proven to be a difficult
task. Classical methods, such as analysing chilled margin samples or sum-
mation of weighted averages of cumulate units, are of little value in the
case of large complexes which have a history of multiple injection of differ-
ent magmas coupled with marginal contamination. Recent studies (2,3) have
shown that a promising approach to this problem is to determine the composi-
tions of coeval dikes and sills. Those exposed at the margins of Stillwater
and Bushveld complexes have somewhat unusual compositions. The magmas that
apparently formed the Ultramafic series in both complexes were high in MgO
(10-15%) and SiO_2 (52-58%), compositionally similar to modern boninites.

In this study, the computer program SILMIN (4) was used (a) to model the
crystallization history of the most likely parental magmas among the dike/
sill set, (b) to determine the composition of hypothetical magmas that are
consistent with the observed cumulate sequences, mineral compositions (inclu-
ding trace elements (5)), and relative mineral proportions, and (c) to con-
strain the composition and history of the mantle source regions. The program,
which models equilibrium crystallization, crystal fractionation, magma mixing
and solid phase assimilation in magmatic systems, incorporates thermodynamic
models for the solid and liquid phase solution properties and utilizes Gibbs
free energy minimization techniques to describe stable multi-phase equilibria
in both closed and open systems. Basic conditions in all calculations were:
fractional crystallization, oxygen fugacities on the QFM buffer, and a
pressure of 300 MPa.

The constraints imposed by the cumulate sequence are summarized in figs.
1 and 2. The U-series magmas were dominant during the crystallization of the
Ultramafic series and Lower Banded series, while the B magmas were dominant

during the crystallization of the Middle Banded series. As shown in fig. 3, one of the samples from the mafic norite dike suite, sample CC2-813 (6), shows good agreement between the calculated crystallization sequence and .mineral compositions and those observed in the U magma cumulates. However, mineral proportions are significantly different, particularly ol/opx ratios. In the case of the B magma, dikes of the diabase group (6) provide a reasonable "calculated versus observed" match with the B magma cumulates.

Results for a hypothetical magma (H-2), the composition of which was adjusted to reproduce Ultramafic series crystallization sequences, mineral compositions, and mineral proportions is shown in figure 4. In terms of major elements, H-2 is quite similar to boninites. However, trace element abundances and H_2O contents are significantly different. At least two possibilities exist for the genesis of this magma: (a) It is a secondary magma produced by the assimilation of crustal material by a MgO-rich magma such as komatiite, as suggested by Longhi et al. (2). (b) It is a primary magma formed by the partial melting of a mantle source of anomalous composition. To test the komatiite parent hypothesis, a SILMIN run was made in which a primary komatiitic magma (SA 3062 from Arndt (7)) assimilated crustal granodiorite while fractionating under isenthalpic conditions. The stoichiometry of the assimilation reaction and the Ma/Mc ratio is calculated at all stages. The results are shown in figure 5. It is clear that this AFC process is capable of producing high MgO, high SiO_2 magmas, e.g., B' in figure 6, which are similar in major elements (with the exception of K_2O) and trace elements to those calculated for the parental magma of the Stillwater Complex. However, it is clear that extensive assimilation is required (Ma/Mc ≃ 1).

Isotopic data provide an additional test of the AFC model. Oxygen isotopic ratios (8), measured on whole rocks and plagioclase separates from a representative suite of samples, indicate that the intrusion has retained a mantle magmatic signature. The bulk melt value of 5.9 per mille severely limits the amount of crustal contamination that could have occurred. Initial whole rock $^{143}Nd/^{144}Nd$ ratios are constant throughout the complex (1) with ϵ_{Nd} = -2 at 2.7 Ga. Initial Pb isotopic ratios measured on plagioclase separates are also constant throughout the complex with the exception of a small variation in $^{207}Pb/^{204}Pb$ in Basal series samples, due to localized contamination with country rocks (9). In contrast, Sr isotopic data (pers comm., B. Stewart, UCLA) show a significant, but largely unsystematic, variation throughout the complex (Table 1). The variations in initial $^{87}Sr/^{86}Sr$ do not correlate with lithologic changes interpreted to be due to injection of different magmas. The initial values for whole rocks show a wider scatter (ϵ_{Sr} = -2.0 to +22.2) than do those of plagioclase separated from the same samples (ϵ_{Sr} = +8.5 to + 19.6) lending credence to the suggestion that Rb may have been redistributed during the Proterozoic heating event (1). Sr isotopic ratios of Basal series samples also show the effect of localized contamination during the early stages of chamber development. The picture that emerges from the isotopic and trace element data is that of a homogeneous primary magma from an enriched mantle source, repeatedly injected into a crustal magma chamber with minor contamination in its earliest stages of development. The large amounts of crustal assimilation required in the AFC model are not consistent with the isotopic data. Partial melting of a homogeneous mantle source of harzburgitic or pyroxenitic composition formed by recycling older crustal material into the mantle is, at present, the model which provides the best explanation of the available data.

References: (1) DePaolo and Wasserburg (1979) GCA, 43, 999-1008; (2) Longhi et al. (1983) JGR, 88, B53-B69; (3) Sharpe and Hulbert (1985) Econ. Geol., 80, 849-871; (4) Ghiorso (1985) CMP, 90, 107-120; (5) Lambert and Simmons (1987) AJS, 287, 1-32; (6) Helz (1985) Montana Bur. Mines Pub. 92, 97-117; (7) Arndt (1986) J. Petrol., 27, 279-301; (8) Dunn (1986) J. Petrol., 27, 987-997; (9) Czamanske et al. (1986) EOS, 67, 1251; (10) McCallum et al. (1980) AJS, 280A, 59-87.
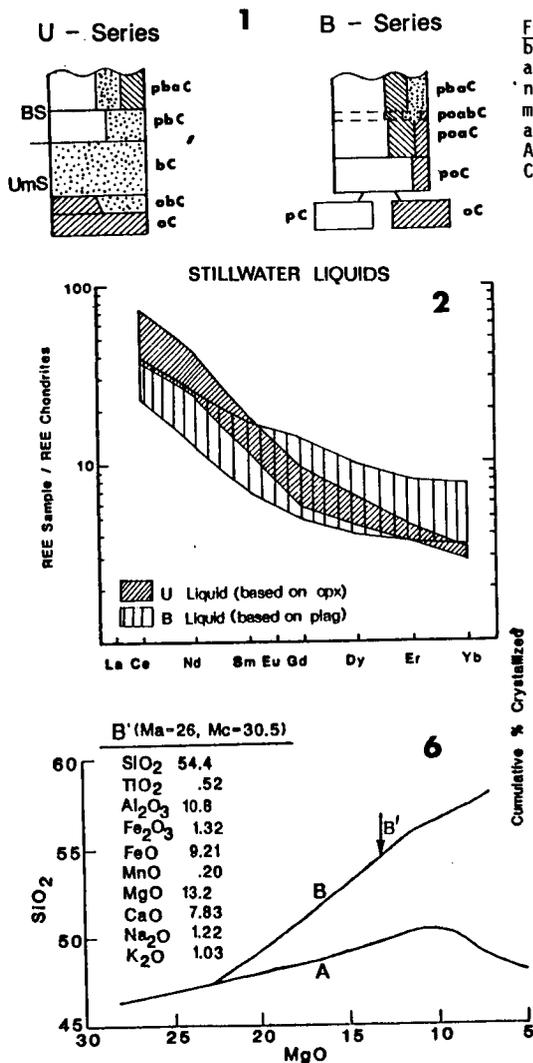
Fig. 1. Cumulate sequences from U- and B-series magmas (o-olivine, b-bronzite, p-plagioclase, a-augite). Fig. 2. Calculated REE of U and B magmas (5). Fig. 3. Composition and crystallization of mafic norite dike. Fig. 4. Composition and crystallization of hypothetical magma H-2. Fig. 5. Composition and crystallization of komatiite (A) and contaminated komatiite (B). Fig. 6. SiO_2 and MgO variation for A and B paths from figure 5. Composition of magma at B' is shown. Compare with composition H-2.
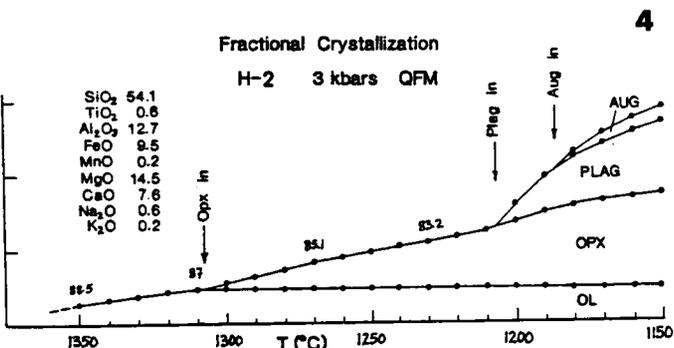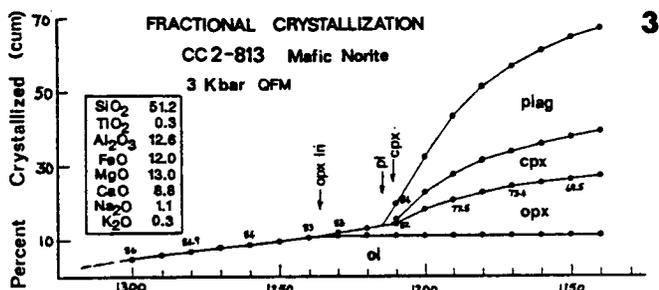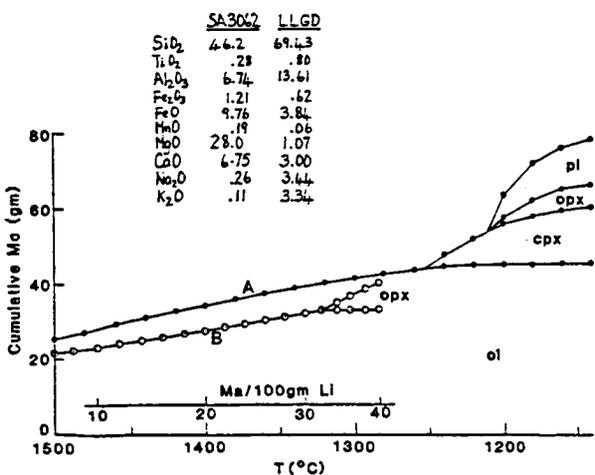
Table 1. Initial isotopic ratios (data obtained by B. Stewart, UCLA)

Sample	Zone*	$^{87}Sr/^{86}Sr(T)$	$\epsilon_{Sr}(T)$
50 PL	GN-III	.70259±3	19.6±.5
31 PL	AN-I	.70214(3)	13.1(.5)
18 WR	OB-III	.70203(3)	11.6(.5)
67 WR	OB-I	.70278(3)	22.2(.5)
67 PL	OB-I	.70238(2)	16.6(.3)
2 WR	N-I	.70108(3)	-2.0(.5)
2 PL	N-I	.70182(3)	8.5(.4)
70 PL	UMS	.70229(3)	15.3(.5)
46 WR	BS	.70272(3)	21.4(.3)
46 PL	BS	.70297(3)	25.0(.4)

*Terminology of McCallum et al. (10)

CHANGING STYLES OF CRUSTAL GROWTH IN SOUTHERN AFRICA: CONSTRAINTS FROM GEOCHEMICAL AND Sr-Nd ISOTOPE STUDIES IN ARCHAEAN TO PAN AFRICAN TERRAINS

F. McDermott, C.J. Hawkesworth and N.B.W. Harris, Dept. of Earth Sciences, The Open University, Milton Keynes, MK7 6AA, U.K.

A comprehensive Nd isotope study of crustal rocks from Southern Africa defines crustal growth rates since about 3.5 Ga. Sediment, granite and crustal xenolith samples from the Archaean Kaapvall and Zimbabwe cratons (3.6 - 2.6 Ga) and the Limpopo, Kalahari, Namaqua and Damara mobile belts (3.5 - 0.5 Ga) have been analysed to constrain the balance between newly formed and reworked crust in successive orogenic episodes. The inferred cumulative growth curve for this segment of Gondwanaland is compared with that obtained from composite samples of Australian shales[1].

Surface samples and crustal xenoliths from the Kaapvaal and Zimbabwe cratons yield Nd model ages (DM) in the range 3.7 - 2.8 Ga[2] while samples from the Limpopo mobile belt which separates the cratons have model Nd ages of 3.2 - 2.5 Ga[3]. Thus, the Archaean cratons and the Limpopo and Kalahari mobile belts which together constitute about 50% of the present day crust of southern Africa had formed prior to 2.5 Ga. Samples from the Namaqua mobile belt yield model Nd ages of 2.3 - 1.5 Ga[4,5,6]. Metasediments and granitoids from the Pan-African Damara belt yield model Nd ages in the range 2.4 - 1.2 Ga[7,8] suggesting that little significant crustal growth occurred during this Pan-African orogeny.

A plot of time vs. ε Nd clearly demonstrates that the last orogenic event (the Pan-African Damara orogen) incorporated relatively more material with low $^{143}Nd/^{144}Nd$ ratios than did the older events (Fig. 1).
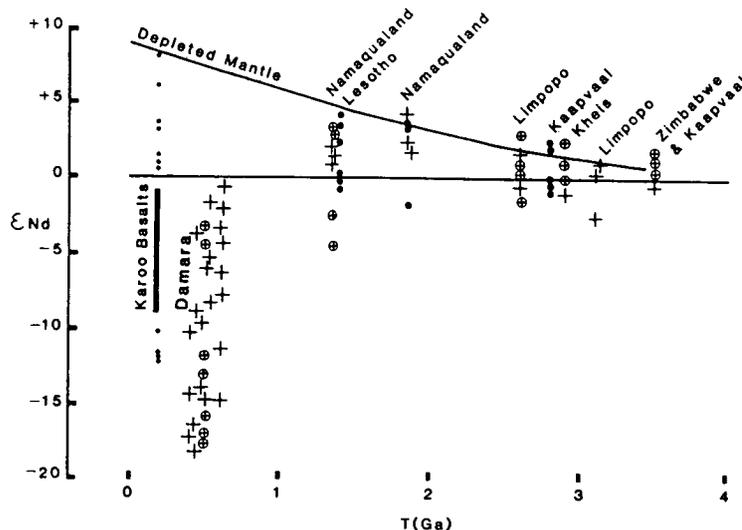
Fig. 1 ε_{Nd} vs. time plot for crustal rocks from Southern Africa. The depleted mantle evolution curve is taken from Allegre and Rousseau[1].

CHANGING STYLES OF CRUSTAL GROWTH IN SOUTHERN AFRICA...........
McDermott, F., Hawkesworth, C.J., Harris, N.B.W.

96

Variations in ε Nd with time is quantitatively related to the rate at which new crust is generated using a simple model[1]. This model corrects for preferential erosion of younger source terrains by incorporating an erosion co-efficient (K) which relates the mean age of the continental crust at any time to the mean age of eroded sediment. The model assumes that the erosion co-efficient is constant throughout geological time, and that orogenic events occurred at intervals of 0.5 Ga since 3.5 Ga ,and allows calculation of both the mean crustal age at any time and the rate at which new crust is generated.

Possible growth rates for the limiting conditions of K=2 (slow erosion) and K=6 (rapid erosion) are shown in Fig. 2. It is noted that the inferred growth curves from this study are significantly different from those inferred for Gondwanaland from Australian shale data[1]. Firstly, the growth rate implied from the Australian shale data is lower during the Proterozoic and secondly it indicates that ~ 20% of the crust formed during the last 1.0 Ga, in contrast to the <10% as implied from this study. Terrains older than 1.4 Ga are dominated by new crustal material, whereas in younger areas intracrustal reworking appears to have been the dominant process.
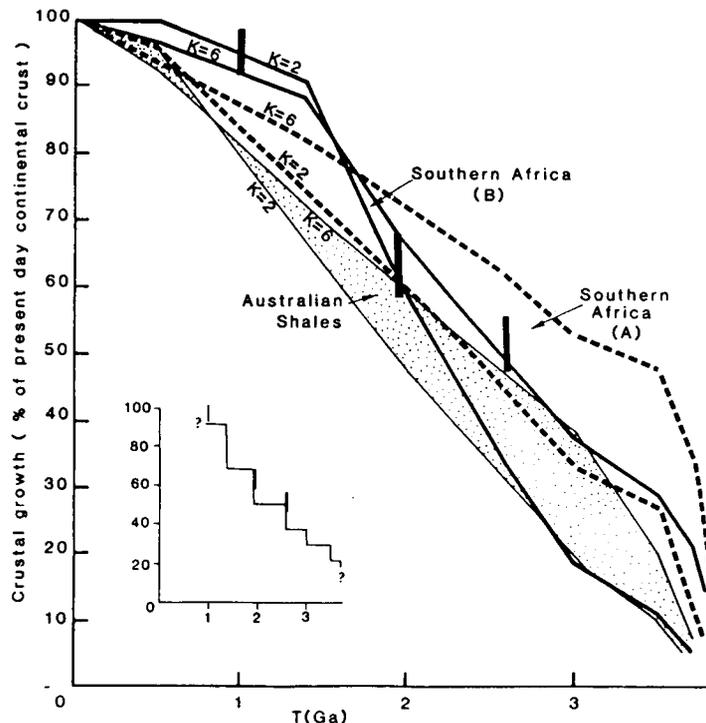
Fig. 2 Integrated crustal growth rates inferred from Australian shale data (Allegre and Rousseau[1]) and Southern Africa (this study) for erosion co-efficients K = 2 and K = 6. Vertical bars indicate empirical estimates for crustal growth in southern Africa based on surface area estimates.

CHANGING STYLES OF CRUSTAL GROWTH IN SOUTHERN AFRICA...........
McDermott, F., Hawkesworth, C.J., Harris, N.B.W.

97

In order to investigate rates of crustal reworking a detailed Sr and Nd isotope study was undertaken on the sediments and granitoids of the Pan-African Damara belt of Namibia. The granites are characterised by high initial $^{87}Sr/^{86}Sr$ ratios (> 0.710) and all granites intruded in the period 650 - 460 Ma have model Nd ages in excess of 1.0 Ga, suggesting that Damara magmatism was not associatied with significant crustal growth. Detailed isotope studies on metasediments from a range of stratigraphic levels have revealed a trend of increasing $^{87}Sr/^{86}Sr$, accompanied by decreasing $^{143}Nd/^{144}Nd$ with depth. The oldest Damara metasediments (Nosib Group) were deposited about 900 Ma ago and have model Nd ages in the range 1.7 - 2.9 Ga, whereas younger molasse-type sediments deposited in the period 650 - 540 Ma have model Nd ages in the range 1.4 - 1.7 Ga, suggesting that sediments were derived from successively younger source terrains with time.

Intracrustal reworking increases Rb/Sr ratios relative to Sm/Nd so that calculated model Sr ages are lower than model Nd ages in Damara metasediments. Furthermore, the ratio of model ages $(T^{Nd}_{DM}/T^{Sr}_{DM})$ may be used to assess the amount of intracrustal reworking of sediment source terrains. The stratigraphically oldest sediments (Nosib Group) have highest model age ratios (2.1 - 2.7) and so record greater degrees of source terrain intracrustal reworking than younger sediments which yield model age ratios in the range 1.3 - 2.2. Average rates of intracrustal fractionation of Rb/Sr have been calculated by averaging the inferred increase in Rb/Sr over the total crustal residence age of a sample i.e. the T^{Nd}_{DM} age. Calculated rates of source terrain intracrustal reworking are highest from the stratigraphically younger sediments (Kuiseb Formation and Nama Group) where Rb/Sr ratios have increased by a factor of about 1.2 - 1.7 per Ga. In contrast, Rb/Sr ratios have increased at a rate of 1.0 - 1.2 per Ga in the Nosib Group source terrains.The apparent increase in the rate of intracrustal reworking with time reflects a decrease in the amount of time which elapsed between crust formation and the onset of intracrustal reworking. This time interval is estimated to be about 1.0 Ga for the Nosib Group source terrains and about 400 Ma for those of the Kuiseb Formation and Nama Group.

References

1   Allegre, C.J. and Rousseau, D. (1984). Earth Planet. Sci. Lett., 67, 19-34.
2   Carlson, R.W., Hunter, O.R. and Barker, F. (1983). Nature, 305, 701-703.
3   Harris, N.B.W., Hawkesworth, C.J., van Calsteren, P. and McDermott, F. Earth Planet. Sci. Lett. (in press).
4   Cornell, D.N., Hawkesworth, C.J., van Calsteren, P. and Scott, W.D. (1986). Trans. Geol. Soc. S. Afr., 89, 17-28.
5   van Calsteren, P., Harris, N.B.W., Hawkesworth, C.J., Menzies, M.A. and Rogers, N.W. (1986). Geol. Soc. Spec. Publ., 24, 351-362.
6   Rogers, N.W. and Hawkesworth, C.J. (1982). Nature, 299, 409-413.
7   Hawkesworth, C.J., Kramers, J.D. and Miller, R.McG. (1981). Nature, 289, 278-282.
8   McDermott, F. (1987). Unpubl. Ph.D. Thesis, Open Univ.

STRUCTURE OF SOUTHWESTERN MICHIPICOTEN GREENSTONE BELT, ONTARIO: EVIDENCE FOR ARCHEAN ACCRETION? George E. McGill and Catherine H. Shrady, Dept. of Geology and Geography, Univ. of Massachusetts, Amherst, MA 01003.

In addition to petrological and geochemical considerations, the dominant tectonic style during the Archean must be determined if we are to understand how early crust evolved. Our understanding of modern plate tectonics is heavily dependent on the areal distribution of geophysical anomalies, volcanism, and topographic features along plate boundaries. Rocks and structures characteristic of specific environments within the modern plate-tectonic framework are rarely, if ever, used as criteria for these environments. In contrast, reconstruction of past tectonic frameworks depends entirely on preserved rocks and structures. The Archean presents particular problems due to the possibility that important deep crustal and upper mantle processes were different (1). Only by using all types of evidence can we be confident that the tectonic style in the Archean was similar (or not similar) to that in the Phanerozoic. Our study of the Michipicoten Greenstone Belt in Ontario has two primary goals: 1) to decipher the structural evolution of the belt through very detailed field mapping and structural analysis, and 2) to provide a well understood kinematic framework for radiometric dating, and for studies of the geochemistry, petrology, and sedimentology of the rocks. The southwestern part of the belt is admirably suited to our purposes because access is relatively easy, rock exposures are superb, and enough rocks with younging indicators are present to facilitate structural interpretation.

The Michipicoten Greenstone Belt extends for about 150 km ENE from the northeastern angle of Lake Superior (Fig. 1). The eastern limit of the belt is the high-grade Kapuskasing zone, which separates the Michipicoten from the Abitibi Greenstone Belt (2). We are mapping key areas in the southwestern part of the belt at a scale of 400' = 1", tying these areas together with 1 mile = 4" mapping or by use of recent Ontario Geological Survey preliminary maps by Sage and others (3). All of our mapping to date (4) has been within Chabanel Township, and most of it is in or adjacent to the large fume kill downwind from the sintering plant in Wawa where outcrops are very abundant and easily located.
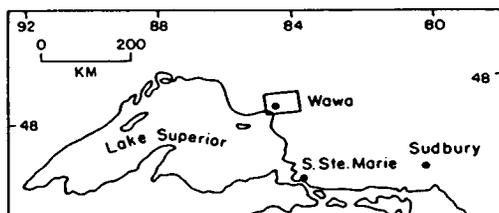
Fig. 1. Location of Michipicoten Greenstone Belt

The rocks of our area may be divided into the traditional lithologic types (3): mafic-intermediate volcanics, intermediate-felsic volcanics, clastic sediments, and chemical sediments (including iron formation). In the southwestern portion of the Michipicoten Greenstone Belt there is an interior (northern) terrane consisting mostly of intermediate to mafic volcanics. These rocks dip north to northeast at intermediate angles, and are overturned (south and southwest younging). Along the southern margin of the belt is a thick sequence of volcanic rocks with minor iron formation that youngs

northward and has steep north or south dips. Between these dominantly volcanic terranes is an extensive belt of clastic sedimentary and pyroclastic rocks (Fig. 2). Much of our detailed mapping has been concentrated in this belt because of the abundance of good indicators of younging direction, and because cleavages generally are better developed in the metasedimentary rocks than they are in the metavolcanic rocks.

In the eastern part of the area we have mapped there is a single, abrupt reversal of younging direction within the sedimentary belt that is interpreted to be a fault (or a faulted isoclinal fold). To the west, the width of the sedimentary belt increases (Fig. 2), in part due to increasing thicknesses of individual lithologic units, but also we believe due to fault imbrication (imbrication is very difficult to prove in the absence of both fossils and laterally continuous stratigraphic sequences). Within the widest part of the sedimentary belt we have been able to define and map lithologic "packages" of turbidite, graywacke, arkose, conglomerate, and pyroclastic rocks, most of which are bounded by contacts that we believe are faults, based on the presence of one or more suggestive characteristics (stretched pebbles in conglomerate, local spaced cleavages, gossans, mafic sills, networks of quartz veins, narrow linear topographic depressions, truncated layering, or abrupt tops reversals). In addition, the sedimentary belt appears to be separated from the northern volcanic terrane by a fault or faulted unconformity that we have been able to trace across the entire width of Chabanel Township.

Tentatively, we interpret the mapped relationships as representing a thrust-imbricated sequence of volcanic and sedimentary rocks. Because of the faulting, the rocks in the northern (interior) volcanic terrane may be exotic with respect to the volcanic rocks along the southern margin of the belt near Wawa; hence simple, single-basin models for the depositional environment are suspect. So far, we have found no microscopic indicators of shear sense in specimens collected along or near the suspected thrust faults, but the X-axes of stretched clasts plunge at moderate angles to the ENE. If the stretching is related to the thrusting, then the transport direction is oblique to the present-day trend of the faults, and also not perpendicular to the long axis of the greenstone belt as a whole.

Evidence that these faults are among the oldest structures in the belt is rather good. Mafic sills are locally abundant in the metasedimentary rocks, and some of these cross the traces of faults with no evidence of shearing and no apparent displacement. Others appear to terminate against faults. Although cleavage is commonly not well developed in these sills, there are a few exposures of mafic sill rocks carrying an early cleavage that is approximately parallel to bedding in the enclosing metasediments. This is the oldest of three or four regionally important cleavages and is associated with a pervasive flattening. The X-axes of stretched clasts also lie in this cleavage. Locally, there is evidence that the rocks were not completely lithified at the time this early cleavage was formed. The overall impression is that sill intrusion and bedding-related faulting and cleavage were all early and roughly coeval.

We thus infer an early episode of thrusting and extensive overturning, probably occurring before the rocks were completely lithified (which may explain the scarcity of microscopic shear-sense indicators). A complex younger history of intrusion, cleavage formation, folding, and faulting has obscured the evidence for this tectonically important early deformation. The early deformation, the nature of the sediments (5), and the geochemistry of

the lavas (6) are all consistent with an accretionary wedge or forearc basin environment for deposition and early deformation of volcanics and sediments. It has been stated (6) that the abundance of felsic pyroclastics and epiclastic sediments indicates eruption through sialic crust. However, we have been unable to find any clasts that are not of volcanic and plutonic rocks found within the greenstone belt itself, so there is no unequivocal evidence for a nearby continental source during deposition. It appears as if at least the early history of the Michipicoten Greenstone Belt, from deposition of the sediments and volcanics through early deformation, is compatible with an essentially modern plate-tectonic style.

References

(1) Richter, F.M., Earth Planet. Sci. Letts., 73, 350-360, 1985.
(2) Percival, J.A., & Card, K.D., Geology, 11, 323-326, 1983.
(3) Sage, R.P., et al., Ontario Geol. Surv. Prelim. Maps P.2439, P.2440, P.2441, P.2442, P.2568, 1982; P.2607, 1984; and P.2636, 1985.
(4) McGill, G.E., & Shrady, C.H., Jour. Geophys. Res., 91, E281-E289, 1986.
(5) Leggett, J.K., ed., Trench-forearc geology: sedimentation and tectonics on modern and ancient active plate margins, Geol. Soc. London, 1982.
(6) Sylvester, P.J., Attoh, K., and Schulz, K.J., Canadian Jour. Earth Sci., in press.
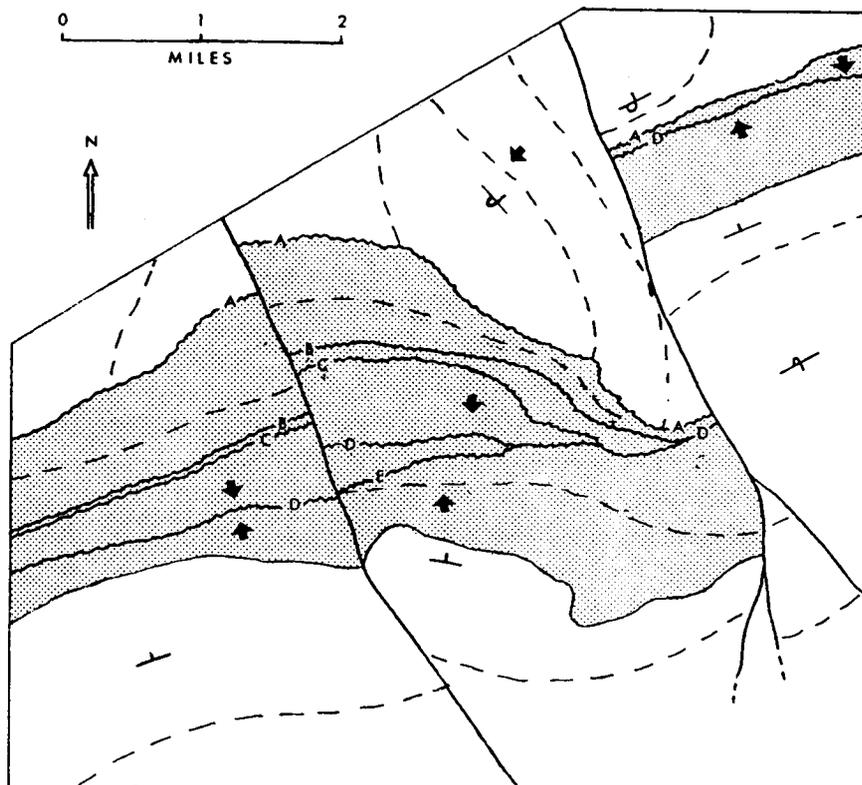
Fig. 2. Inferred thrust faults (A-E) in Chabanel Township. Shaded area is dominantly metasediments and pyroclastics; unshaded areas are dominantly volcanics. Arrows indicate younging direction. Dashed lines trace general trends of bedding and flow boundaries.

MID TO LATE PROTEROZOIC MAGMATISM WITHIN NORTHEASTERN NORTH AMERICA AND ITS IMPLICATIONS FOR THE GROWTH OF THE CONTINENTAL CRUST, J.M. McLelland, Department of Geology, Colgate University, Hamilton, New York 13346

Recent studies of the mangerite-charnockite-alaskite suite exposed in the Adirondack Highlands strongly suggest that these rocks were emplaced under anorogenic, or mildly extensional, conditions[1,2]. The characteristic signatures of the rocks are high $(FeO/FeO+MgO)$ and (FeO/Fe_2O_3); mildly alkaline to sub-alkaline and metaluminous trends; high Ga/Al_2O_3; and within-plate concentrations of Nb vs Y, Ta vs Yb, Rb vs (Y+Nb) and Rb vs. (Yb+Ta). Evolved members of the series are low in CaO and MgO and high in alkalies and halogens. All of these properties are consistent with anorogenic magmatism comprising acidic crustal melts and mantle derived mafic additions to the crust.

Major and minor element trends, as well as field evidence, strongly suggest that the anorogenic acidic suite is coeval, but not comagmatic, with closely associated anorthositic massifs. Present outcrop configurations are consistent with the evolution of the acidic and anorthositic rocks in zoned, bimodal magma chambers cored by the mafic constituents and overlain by explosive, caldera-type volcanism. A similar setting is believed to apply to anorthosite-charnockite complexes throughout the Grenville Province and across the Grenville Tectonic Front in the southern Nain Province[3].

Ages directly determined for anorthosite massifs in the Grenville and Nain Provinces range from ∿1600 Ma (Harp Lake, Mealy Mts.)[4] to 1079 Ma (St. Urbain)[4], with the Sept Isle (540 Ma)[5] representing a rare, and small, lower extreme. Age determinations on associated acidic orthogneisses tends to confirm these ages and indicate that the emplacement of the anorthosite-charnockite suite extended over ∿500 Ma in the Grenville and adjacent Nain, Provinces.

Excluding the easternmost section of the Grenville Province with the apparently allochthonous Mealy Mts.[6], the area the anorthosite bearing terrain is ∿500,000 km². Of this sum, anorthosite massifs themselves account for ∿70,000 km² or 12% of the area. Assuming that these mantle-derived contributions to the crust were emplaced over an interval of ∿500 Ma, we find the average rate of crustal growth to be ∿.025% per 10^6 years. This is close to the rate of constant crustal addition proposed by Hurley and Rand[7]. If anorthosite emplacement was focused into the interval 1400-1100 Ma then the growth rate would increase to ∿.04% per 10^6 years which is closer to the estimates of Nelson and DePaolo[8] for rapid production of continental crust during the mid to late Proterozoic.

102

(1) McLelland, J., 1986, Geology, v. 14, p. 229-233.
(2) Whitney, P., 1986, Geol. Soc. Am. Prog. Abst., v. 18, p. 788.
(3) Emslie, R., 1978, Precambr. Res., v. 7, p. 61-98.
(4) Ashwal, L. and Wooden, J., 1984, Nature, v. 306, p. 679-680.
(5) Higgins, M. and Doig, R., 1981, Can. Jour. Earth. Sci., v. 18, p. 561-573.
(6) Rivers, T., and Chown, E., 1986, Geol. Assoc. Canada Spec. Pap. 31, p. 31-51.
(7) Hurley, P. and Rand, J., 1969, Science, v. 164, p. 1229-1242.
(8) Nelson, B. and DePaolo, D., 1985, Geol. Soc. Am. Bull., v. 96, p. 746-754.

THERMAL MODELS PERTAINING TO CONTINENTAL GROWTH; Paul Morgan,
Geology Department, Box 6030, Northern Arizona University, Flagstaff, AZ
86011, USA, and Lew Ashwal, Lunar and Planetary Institute, 3303 NASA Road One,
Houston, TX 77058, USA.

Thermal models are important to understanding continental growth as the
genesis, stabilization and possible recycling of continental crust are closely
related to the tectonic processes of the earth which are driven primarily by
heat. The thermal energy budget of the earth has been slowly decreasing since
core formation, and thus the energy driving the terrestrial tectonic engine
has been decreasing. We use this fundamental observation to develop a logic
tree defining the options for continental growth throughout earth history.

Primordial Crust. It is highly likely that the final stages of earth
accretion were accompanied by deep zones of melt near the earth surface and
that fractional crystallization in these melts formed a primordial crust. It
is unlikely, however, that this crust would have been sufficiently
differentiated to produce Archean continental crust. Multiple melting and
differentiation of this primordial crust and the underlying primitive mantle
could have contributed to the Archean granitoid crust. Models of a Hadean
terrestrial magma ocean suggest that this crust may have been 10 to 15 km in
thickness and iron-rich leucodiorite to leucogabbronorite in composition (1).
It is likely that these low density felsic rocks were either remelted or
provided nuclei for early Archean crust.

Impact Phase. Impact modification of the primordial crust up to about
the time of formation of the oldest surviving terrestrial rocks seems
inevitable by analogy with the lunar history, but apart from isolated younger
impact structures, there is no evidence of this phase of earth evolution in
the rocks. This lack of evidence suggests relatively rapid reworking of the
early crust either through weathering/resurfacing, igneous/metamorphic
activity, or recycling. It is possible, even probable that igneous and
tectonic activity during this phase may have been localized and/or triggered
by impacts, and thus the evidence of impact activity quickly destroyed.

Early Archean - Oldest Surviving Continental Crust. The coincidence
between the ages of the probable decline of late heavy bombardment on the
earth (approx. 3.9 Ga) and the oldest surviving continental rocks (approx. 3.8
Ga), and the total lack of evidence for impact events in these rocks suggests
that impact energy prevented stabilization of continental crust prior to about
3.8 Ga. Older detrital zircon ages (approx. 4.2 Ga, ref. 2) indicate that
older crust existed, but was reworked. Thus, there is evidence for
pre-Archean crust (continental?) and some form of early crustal reworking.
The fundamental questions are how much continental crust existed at the start
of the Archean, and how fast was new continental crust being added?

Most models for the generation of continental crust invoke the remelting
of basaltic crust, probably with excess water, and possibly with some mixing
with existing continental derived material, to form intermediate composition
melts (andesite, ref. 3). In a modern plate-tectonic setting this process is
associated with subduction and arc-volcanism. The rate of creation (and
associated subduction) of oceanic crust is closely related to the present
global heat loss (e.g., 4), and thus presumably the addition of new arc-
related continental crust is related to the global heat budget. If a similar
tectonic regime prevailed in the Archean, higher global heat loss in the
Archean (5) would have resulted in a higher rate of addition of continental
crust through arc-magmatism. In a strict "uniformitarianism" model of
continental crustal generation, the rate of production of new continental

THERMAL MODELS AND CONTINENTAL GROWTH
Morgan, P. and Ashwal, L.D.

104

crust is expected to have decreased steadily with time. Models with minor variations on "uniformitarianism" such as faster spreading ridges, more spreading ridges, and/or thicker oceanic crust are not expected to change this basic result.

Lower rates of Archean continental crust generation are only consistent with higher global heat loss in the Archean if the global tectonic regime was fundamentally different from plate tectonics. Alternative models for Archean tectonics have been suggested (e.g., ref. 6), and Taylor and McLennan (7) argue that chemical differentiation of the Archean crust was related to melting at mantle depths, in sharp contrast with the intracrustal processes responsible for differentiation of the post-Archean crust. Only the thermal implications of these deductions are considered here. If Archean crust was not generated in association with arc-magmatism, then this suggests that plate tectonics may not have been operating in its present form or differentiation though arc-magmatism was insignificant, perhaps in association with the subduction of more buoyant oceanic lithosphere. Morgan and Phillips (8) have argued that high global heat loss could be efficiently lost without plate tectonics in a hot-spot tectonics systems, perhaps similar to modern venusian tectonics.

Growth of Continental Crust. Arguments presented above pertain only to the rate of addition of new continental crust, and recycling and reworking of this crust must also be included in consideration of the net rate of continental growth. Continental crust may be recycled into the mantle through subduction (e.g., ref. 8), and it can be argued that more rapid crustal generation in the Archean associated with higher global heat loss was balanced by more rapid crustal recycling. However, until the efficiency of modern crustal recycling is established, it is difficult to constrain the rates of recycling under hotter thermal regimes.

The composition of surviving Archean crust is statistically low in the heat producing elements, U, Th and K, relative to younger crust and there is a relatively sudden appearance of radiogenically "normal" crust around the Archean-Proterozoic boundary (2.8 - 2.3 Ga). Intrinsic crustal heat generation can have a significant effect on the geotherm, and in a hot Archean global thermal regime it is possible that high heat production crust may have been selectively reworked (9). Lithospheric strength is controlled primarily by crustal thickness and the geotherm, and in a "hot' thermal regime, i.e., with Moho temperatures around 800 degC, all strength is concentrated in the upper crust, and is sensitive to upper crustal heat generation. As the Moho temperature cools below about 800 degC, significant strength is gained by the uppermost mantle, and lithospheric strength becomes less sensitive to upper crustal heat generation. Thus, a possible thermal explanation for the transition in U, Th and K contents between surviving Archean and Proterozoic crusts is a gradual global decrease in mean Moho temperature, reducing the sensitivity of lithospheric strength to upper crustal heat generation, and reducing selective reworking of the crust.

Ancient Continental Geotherms. Whatever the global tectonic process in the Archean, higher global heat loss implies that the mean lithospheric thickness was less than at present and that the geotherm was higher on average. Higher mantle temperatures have been inferred from the occurrence of Archean komatiitic lavas. In contrast, relatively low Archean geothermal gradients beneath some surviving blocks of Archean crust have been deduced from the presence of diamonds of Archean age (e.g., 10). Some regions of thick, cool lithosphere therefore appear to have been stable in a hot global thermal regime.

THERMAL MODELS AND CONTINENTAL GROWTH
Morgan, P. and Ashwal, L.D.

Two basic parameters control the stable thickness of the continental thermal boundary layer (lithosphere), the heat production within the layer, and the heat input to its base (11). The thermal boundary layer can be thickened, and perhaps stabilized by decreasing the heat input to its base. By increasing lithospheric mantle heat production so that the lithospheric geotherm becomes asymptotic to the asthenospheric geotherm, heat input through the base of the lithosphere is minimized and geotherms representing a thick lithosphere with a hot asthenosphere can be generated. Perhaps a feature of continental stabilization is the creation of such a geotherm through enrichment of the lithospheric mantle in U, Th and K associated with lithospheric shortening and/or metasomatism. Xenolith data indicate that such enrichment may have occurred under some continental regions (12), and lead isotope data indicate heterogeneities that may have been caused by long-term enrichment of portions of the upper mantle in U and Th (13).

Concluding Remarks. Thermal models place important constraints upon the growth and stabilization of continental crust. Secular decrease in global heat flow suggests a decrease in the rate of continental crust generation in "uniformitarianism" models, although the net increase or decrease in this crust is controlled also by coupled recycling. Non-smooth continental growth curves are only predicted by non-uniform models of crustal generation and/or by time-dependent models of crustal recycling or reworking.

The constraints implied by these thermal models can be represented in the form of a logic tree which may be pruned by constraints from other models or observations.

References: (1) Hofmeister, A.M. (1983) J. Geophys. Res., 88, 4963-4983. (2) Froude, D.O., T.R. Ireland, P.D. Kinny, I.S. Williams, W. Compston, T.R. Williams and J.S. Meyers (1983) Nature, 304, 616-618. (3) Arth, J.G., and G.N. Hanson (1972) Contrib. Mineral. Petrol., 37, 161-174. (4) Sclater, J.G., C. Jaupert and D. Galson (1980) Rev. Geophys. Space Phys., 18, 269-311. (5) Wasserburg, G.J., G.J.F. MacDonald, F. Hoyle and W.A. Fowler (1964) Science, 143, 465-467; McKenzie, D.P., and N. Weiss (1975) Geophys. J. R. Astro. Soc., 43, 425-451. (6) Hargraves, R.B. (1981) in A. Kroner, Precambrian Plate Tectonics, Elsevier, New York, 21-56; Lambert, R. St. J. (1981) in A. Kroner, Precambrian Plate Tectonics, Elsevier, New York, 453-467. (7) Taylor, S.R., and S.M. McLennan (1985) The Continental Crust: Its Composition and Evolution, Blackwell, Oxford, 312 pp. (8) Fyfe, W.S. (1978) Chem. Geol., 23, 89-114; McKenzie, D. and R.K O'Nions (1983) Nature, 301, 229-231. (9) Morgan, P. (1985) J. Geophys. Res., 90, C561-C570. (10) Richter, F.M. (1985) Earth Planet. Sci. Lett., 73, 350-360. (11) Morgan, P. (1984) Phy. Chem. Earth, 15, 107-193; Morgan, P., and J.H. Sass (1984) J. Geodynamics, 1, 143-166. (12) Jordan, T.H. (1975) Rev. Geophys. Space Phys., 13, 1-12. (13) Basaltic Volcanism Study Project (1981) Basaltic Volcanism on the Terrestrial Planets, Pergamon Press, New York, 1286 pp. (see Ch 7).

The Tectonic Setting of Archean Anorthosites; D. A. Morrison and W. C. Phinney, Johnson Space Center, houston, Texas and D. E. Maczuga, LEMSCO, Houston.

Archean anorthosites, anorthositic gabbros and related rocks form intrusive complexes in every Archean craton. These complexes are characterized by the presence of equidimensional plagioclase megacrysts of typically uniform and calcic (An80-90) composition. Many, if not all, such complexes developed in greenstone belt terrains and appear to be genetically linked to contemporary volcanics. The petrogenesis of Archean anorthosites appears to argue for development in oceanic settings and implies intensive magmatic activity.

The parent liquid of Archean anorthosites appears to have been tholeiitic with depleted light rare earths. FeO abundances in cumulate plagioclase from Archean anorthosites representing a variety of settings suggests that the parent liquid had a relatively high (10 - 12%) iron content. Plagioclase megacryst-bearing tholeiitic pillow basalts, flows, sills and dikes, spatially associated with anorthosites in the Bird River area of Manitoba and the Bad Vermilion Lake complex of Ontario for example, have rare earth concentrations and major element abundances which suggest that they may represent a parent liquid. Experimental crystallization of basalts from the Bird River area show that such rocks may indeed represent a parent liquid for Archean anorthosites because plagioclase of the appropriate An content is produced over a limited temperature range. However, only small amounts of plagioclase are produced and the ratio of cumulate to melt may be as great as one part cumulate to 20 parts melt. Experiments also indicate that plagioclase in the An range appropriate for Archean anorthosites is on the liquidus for only a few degrees before reaching a cotectic and that cotectic proportions are approximately 70% plagioclase 30% mafic. Significant volumes of many Archean anorthosites are gabbroic anorthosites close to cotectic proportions. Archean anorthosites therefore are cumulates from a relatively small, essentially isothermal, crystallization interval of a tholeiitic magma. Consequently, production and assembly of the volume of cumulates represented by Archean anorthosites requires much larger volumes of parent liquid. Because the accumulation process is close to isothermal, the cumulates must be extracted from the liquid, the chamber replenished and extraction repeated many times to form the cumulates. This hypothesis, coupled with the observation that at least some Archean anorthosites are linked to supracrustal sequences, suggests that the anorthosites mark the sites of intensive volcanism during their accumulation. The most appropriate setting is an oceanic environment as is suggested by the volcanics. Whether or not Archean anorthosites represent relatively minor intracontinental rifts or fully developed ocean basins is indeterminate at present.

THE PETROGENESIS OF "OCEANIC KIMBERLITES" AND INCLUDED MANTLE MEGACRYSTS: THE MALAITAN ALNOITE.

Clive R. NEAL, Dept. of Geological Sciences, University of Tennessee, Knoxville, TN 37996-1410, USA.

The study of unambiguous sub-oceanic mantle has been facilitated by the occurrence of anomalous kimberlite-type intrusives (with peridotite xenoliths) on Malaita in the Solomon Islands [1,2]. These "pseudo-kimberlites" were termed **alnoites** by [3], and are basically mica lamprophyres with melilite in the groundmass [4]. Alnoitic magmas were explosively intruded into the Ontong Java Plateau (OJP) 34 Ma ago [5]. The OJP is a vastly overthickened portion of the Pacific plate (up to 42km; [6]) which now abuts the Indo-Australian plate. Malaita is considered to be the obducted leading edge of the OJP [7].

Conspicuous among the mantle xenoliths at Malaita is a large megacryst suite, similar to those found in continental kimberlites, but with additional augite megacrysts. The cpx megacrysts form a continuous array [Ca/(Ca+Mg) = 0.507-0.345; Mg# = 65.4-85.7: see Fig. 1] from augite to subcalcic diopsides; clinopyroxene-ilmenite intergrowths are intermediate. As the cpx megacrysts display the widest range of compositions (relative to other megacryst phases), only these will be considered in this trace element study. A direct origin for these megacrysts from the host alnoite is questioned by [8]. They demonstrated that cpx phenocrysts from the alnoite do not fall on the trend defined by the megacrysts. Furthermore, Stuckless and Irving [9] demonstrated that megacrysts in alkali basalts from SE Australia have <u>less</u> radiogenic $^{87}Sr/^{86}Sr$ ratios than the host volcanic. Similar characteristics documented from South Africa [10,11], India [12], and the U.S.A. [13], would appear to negate a phenocrystal origin for megacrysts.

RESULTS: The large, continuous cpx compositional range, relative to corresponding phases in kimberlite and alkali basalt, has afforded a unique opportunity to study megacryst-host relationships in detail. Six augite and six subcalcic diopside megacrysts, as well as four alnoite samples, were analysed for $^{87}Sr/^{86}Sr$, $^{143}Nd/^{144}Nd$, and the REE. The cpx megacrysts have similar REE profiles, although CRN198 (subcalcic diopside) is depleted in total REE, and the augites have slightly lower La/Yb ratios (Fig. 2). Calculated <u>equilibrium liquids</u>, using published cpx/liquid partition coefficients [14,15], demonstrate that the cpx megacrysts have not crystallized from an alnoitic liquid. This parental liquid resembles an alkali basalt for the augites, whereas the subcalcic diopsides crystallized from a liquid with a steeper REE profile, between alkali basalt and alnoite.

FIG. 1

Sr and Nd isotope ratios become more radiogenic from the augite ($\simeq$0.70341 and $\simeq$0.512756) to the subcalcic diopside megacrysts ($\simeq$0.70384 and $\simeq$0.512785) to the host alnoite ($\simeq$0.70445 and $\simeq$0.512810). Oxygen isotope results from the megacrysts, exhibit a general decrease from the augites ($\simeq$+6.6) to the subcalcic diopsides ($\simeq$+6.1).

DISCUSSION: The isotope and REE data would seem to indicate no direct relationship is possible between the cpx megacrysts and the host alnoite. Indeed, the isotopic disparity between the augite and subcalcic diopsides would preclude any suggestion that they formed from a common parental magma. However, the large, continuous compositional array exhibited by major elements in the cpx megacryst suite (Fig. 1) would argue the opposite case. This array has been interpreted as representing evolution paths of fractionating, parental megacryst magmas, which evolve to more Ca- and Fe-rich compositions [2]. The proposed model for megacryst and alnoite petrogenesis requires that the megacrysts fractionated from a single, parental proto-alnoite magma, while assimilating an isotopically distinct component (an AFC process). This model requires that the augite megacrysts fractionated first, as they have the lowest Sr and Nd isotope ratios, and the flattest REE profile of the calculated equilibrium liquid. Subcalcic diopsides crystallized later because they have: 1) Sr and Nd ratios intermediate between the augite megacrysts and the alnoite; 2) calculated equilibrium liquids with REE profiles which are steeper than those for the augites, but are not alnoitic. Such an interpretation would require the parental proto-alnoite magma to become depleted in Fe and Ca, and enriched in Mg as megacryst fractionation continued.

The effect of megacryst fractionation on bulk magma composition has been studied by Schulze [16,17]. He concluded that >8% ilmenite crystallization in any megacryst suite would deplete the remaining magma in Fe relative to Mg, if olivine was not a major phase. Abundant ilmenite megacrysts are found at Malaita. A decrease in Ca/(Ca+Mg) ratio will be witnessed if: 1) cpx is an abundant liquidus phase; 2) little olivine is fractionated; 3) olivine, or Mg-rich rocks are assimilated [18]. No olivine, and only relatively minor bronzite megacrysts have been found at Malaita. Therefore, at Malaita we consider that it is plausible that augite megacrysts precipitated before the subcalcic diopsides.

The nature of the assimilated component must allow the increase of both Sr and Nd isotope ratios. The only geologically feasible component able to satisfy these conditions is seawater altered basalt which is considered to underplate the OJP. Assimilation of such a Mg-rich component would not only increase Sr and Nd isotope ratios, but would decrease the Ca/(Ca+Mg) ratio of the assimilating magma [18]. As the alnoite contains up to 25 wt% MgO, it is considered to represent the residual, Mg enriched liquid after AFC. An AFC projection [19] has been calculated (Fig. 3) between a proto-alnoite liquid (represented by the augites) and a seawater-altered basalt component [20]. Average megacryst isotope compositions are used. Oxygen isotope data support this model in that there is a decrease in $\delta^{18}O$ from augites to subcalcic diopsides, consistent with assimilation of hydrothermally altered oceanic crust.

FIG. 2

Subcalcic Diopsides
Augites   Alnoites

CRN198
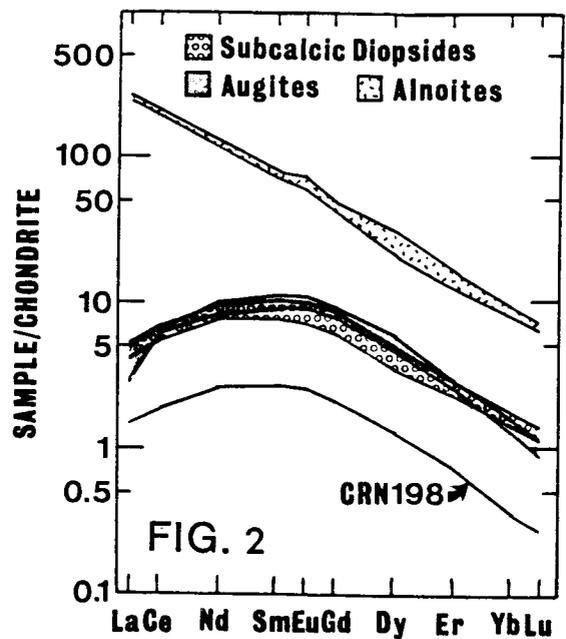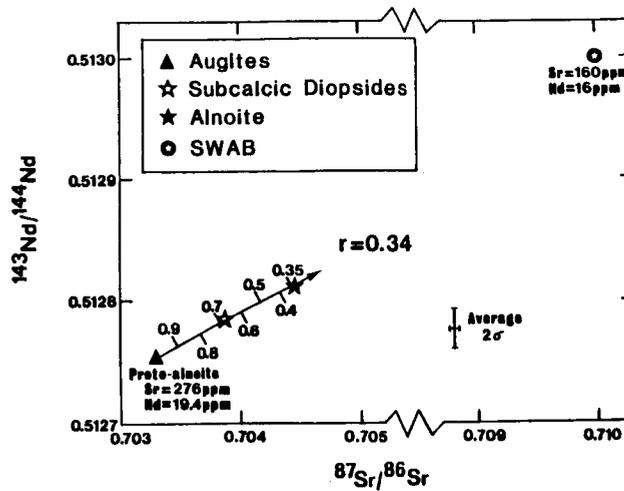
FIG. 3

CONCLUSIONS

Initial diapiric upwelling beneath the OJP produced the proto-alnoite magma. After impingement on the rigid lithosphere, megacryst fractionation occurred, with augites precipitating first, representing the parental magma. Seawater-altered oceanic crust, which underplates the OJP, was assimilated by the proto-alnoite magma during megacrysts fractionation (AFC).

REFERENCES 1) Nixon & Coleman, 1978, Bull. Australia Soc. Explor. Geophys. 9, 103-107;   2) Nixon & Boyd, 1979, In: The Mantle Sample: Inclusions in Kimberlites and other Volcanics, 400-423;   3) Allen & Deans, 1965, Min. Mag. 34, 16-34;   4) Rosenbusch, H., 1887, Mikroskopische Physiographie, 805.   5) Davis, 1977, Second Int'l. Kimb. Conf.; 6) Coleman, 1976, Tech. Bull. 2, (IDOE workshop, Suva, 1975), 140-234;   7) Coleman & Kroenke, 1981, Geo-Marine Lett. 1, 129-134; 8) Delaney et al., 1979, Contr. Min Pet. 70, 209-218;   9) Stuckless & Irving, 1976, GCA 40, 209-213;   10) Kramers et al., 1981, Nature 291, 53-56;   11) Jones, 1984, PhD thesis, Univ. Leeds, UK;   12) Paul, 1979, GCA 43, 389-394; 13) Menzies et al., 1985, Jour. Pet. 26, 663-693;   14) Schnetzler & Philpotts, 1970, GCA 34, 331-340;   15) Grutzeck et al., 1974, GRL 1, 273-275;   16) Schulze, 1982, EOS 63, 464;   17) Schulze, 1984, 3rd Int'l. Kimb. Conf. 2, 97-108;   18) Kelleman, 1986, Jour. Geol., 94, 829-843;   19) DePaolo, 1981, EPSL 53, 189-202;   20) McCulloch, et al., 1981, JGR 86, 2721-2735.

RIDGE-TRENCH COLLISION IN ARCHEAN AND POST-ARCHEAN CRUSTAL GROWTH: EVIDENCE FROM SOUTHERN CHILE

E. P. Nelson, Colorado School of Mines, Golden, CO 80401 and
R. D. Forsythe, Rutgers Univ., New Brunswick, NJ 08903

The growth of continental crust at convergent plate margins involves both continuous and episodic processes. Ridge-trench collision is one episodic process that can cause significant magmatic and tectonic effects on convergent plate margins. Because the sites of ridge collision (ridge-trench triple junctions) generally migrate along convergent plate boundaries, the effects of ridge collision will be highly diachronous in Andean-type orogenic belts and may not be adequately recognized in the geologic record.

Ridge collision involves mechanical and thermal changes caused when a buoyant, bathymetrically-high, and thermally-active spreading ridge approaches, collides with, and potentially subducts below a trench. Both 'near-field' and 'far-field' effects of ridge collision have been proposed (1-4). Such effects will vary depending on the triple junction geometry (various angles between ridge and trench and various convergence vectors are possible). Near-field effects are tectonic and magmatic features developed in the immediate region of the triple junction. Proposed near-field effects include anomalous heat (and related metamorphism) in the arc and forearc, uplift followed by subsidence of the arc massif, a reduction or hiatus in arc magmatism, anomalous forearc magmatism, ophiolite obduction, and tectonic erosion. Far-field, or more regional consequences of ridge collision, include: changes in the thermal structure of the continental lithosphere, systematic changes in arc magmatism, changes in the continental stress regime resulting in intraplate and plate-margin deformation (e.g., Basin and Range-type and California borderland-type tectonic disruption), marginal basin formation (e.g., Gulf of California), and regional uplift and/or subsidence.

The Chile margin triple junction (CMTJ, 46°S), where the actively spreading Chile rise is colliding with the sediment-filled Peru-Chile trench, is geometrically and kinematically the simplest modern example of ridge collision. Spreading ridge segments are approximately parallel to the trench, and convergence vectors approximately normal. Magmatic and tectonic features in the Taitao region of the Andean forearc believed related to the history of ridge collision over the last 3-6 Ma (1, 2) include: (a) the 3-4 Ma Taitao ophiolite exposed <15 km from the trench, (b) silicic, epizonal, 3.0-5.5 Ma plutons exposed near the ophiolite, (c) hot springs (otherwise rare in southern Chile), (d) the Golfo de Penas, a pull-apart basin on the continental shelf near the CMTJ, (e) a major margin-parallel fault (Liquiñe-Ofqui fault) that runs along the Quaternary volcanic arc and that appears to terminate to the south in the Golfo de Penas in a zone of normal and strike-slip fault splays, (f) active microseismicity in the region, and (g) uplifted Quaternary deposits. Far-field consequences of ridge collision include reduction in seismicity and volcanism, chemical changes in arc magmatism, elevated geothermal gradient in the Andean backarc (5), and possibly regional uplift.

RIDGE-TRENCH COLLISION AND CRUSTAL GROWTH
NELSON, E. P. AND FORSYTHE, R. D.

111

The south Chile margin illustrates the importance of the ridge-collision tectonic setting in crustal evolution at convergent margins. Similarities between ridge-collision features in southern Chile and features of Archean greenstone belts raise the question of the importance of ridge collision in Archean crustal growth. Like Archean greenstone belts, the Taitao ophiolite is characterized by a lower section of mafic-ultramafic rocks overlain by an upper section of basaltic to rhyolitic volcanics of mixed calc-alkaline and MORB character intercalated with locally-derived volcaniclastic sedimentary rocks. Also like greenstone belts, the forearc in the Taitao region is intruded by granitic rocks geochemically similar to the nearby coeval volcanics.

Archean plate tectonic processes were probably different than today (e.g., 6-8); these differences may have affected the nature and importance of ridge collision during Archean crustal growth. Proposed differences include: higher heat flow, increased spreading rates, greater ridge length, smaller plates, and thicker oceanic crust. In conclusion, we suggest that smaller plates, greater ridge length, and/or faster spreading all point to the likelihood that ridge collision played a greater role in crustal growth and development of the greenstone-granite terranes during the Archean. However, the effects of modern ridge collision, and the processes involved, are not well enough known to develop specific models for Archean ridge collision.

References:
1. DeLong S. E. and Fox P. (1977) Maurice Ewing Series, V. 1, p. 221-228
2. Barker, P. F. (1982) J. Geol. Soc., p. 787-802.
3. Forsythe R. D. and Nelson E. P. (1985) Tectonics, p. 23-27.
4. Forsythe R. D. et al. (1986) Geology, p. 23-27.
5. Douglas B. J., Saul S. L. and Stern C. R. (1987) J. Geol., p. 241-254.
6. Sleep N. H. and Windley B. W. (1982) J. Geol., 363-379.
7. Hargraves R. B. (1986) Geology, 750-752.
8. Abott D. H. and Hoffman S. E. (1984) Tectonics, p. 429-448.

REAPPRAISAL OF THE GEOLOGY OF KANGIMUT SAMMISOQ, AMERALIK FJORD, SOUTHERN WEST GREENLAND: CRUSTAL STRUCTURE AND INTERPRETATION OF ISOTOPIC DATA.

Allen P. Nutman[1], Clark R.L. Friend[2] and Vic R. McGregor[3]; [1] Department of Earth Sciences, Memorial University of Newfoundland, St. John's, Newfoundland, Canada, [2] Department of Geology, Oxford Polytechnic, Oxford, England, [3] Atammik, 3912 Sukkertoppen, Greenland.

In Godthåbsfjord (G on Fig. 1), in the North Atlantic Archean Craton, early Archean Amitsoq gneisses are found in association with the mid to late Archean Nuk (s.l.) gneisses (1,2,3). The Amitsoq gneisses contain the Ameralik dykes, now subconcordant strips of amphibolite. The Nuk and Amitsoq gneisses have amphibolite facies polymetamorphic assemblages. To the southeast of Godthåbsfjord, there is an extensive region affected by granulite facies metamorphism at 2800 Ma (4,5).

Early models of crustal evolution for Godthåbsfjord and the region to the southeast suggested that the early Archean Amitsoq gneisses were reworked during a 3000 - 2800 Ma crustal accretion - differentiation event, marked by injection of the voluminous Nuk (s.l.) gneisses (e.g. 6). This event was regarded to have culminated in a single regional metamorphic peak at 2800 Ma which outlasted all significant ductile deformation, and was marked by amphibolite facies assemblages in Godthåbsfjord and granulite facies assemblages to the south (e.g. 7,8).

However, recent detailed mapping (9) combined with U-Pb zircon dating (H. Baadsgaard, pers. comm., 1986) shows that the Færingehavn and Ameralik areas (F and A respectively, Fig. 1), consist of three terranes that were assembled at ca. 2700 Ma and were subsequently futher deformed and metamorphosed. The boundaries between the terranes are zones of strongly banded gneisses and schists, interpreted to be mylonites formed during the tectonic juxtaposition of the terranes. Structurally highest and to the south of Godthåbsfjord is the Tasiusarsuaq terrane (upper structural unit, Fig. 1), affected by 2800 Ma granulite facies metamorphism and dominated by tonalitic gneisses. This overlies the Tre Brødre terrane (middle structural unit, Fig. 1), which is dominated by the 2800 - 2750 Ma (H. Baadsgaard, pers. comm., 1986) granodioritic Ikatoq gneisses (a subdivision of the Nuk s.l. gneisses), which have never undergone granulite facies metamorphism. Structurally below the Tre Brødre terrane is the Færingehavn terrane (lower structural unit, Fig. 1), dominated by the >3600 Ma Amitsoq gneisses. The Tre Brødre and Færingehavn terranes form most of Godthåbsfjord. Following assembly of the terranes, folds and steeply-inclined shear zones developed (Fig. 1). During its tectonic juxtaposition with the other terranes and subsequent folding, 2800 Ma granulite facies assemblages of the Tasiusarsuaq terrane were patchily retrogressed under amphibolite facies conditions and the underlying Færingehavn and Tre Brødre terranes underwent amphibolite facies metamorphism (9).

There has been lengthy debate whether the region affected by 2800 Ma granulite facies metamorphism (now called the Tasiusarsuaq terrane) contains any >3600 Ma Amitsoq gneisses

Nutman et al.   Reappraisal of ↓ Kangimut sammisoq, W. Greenland

113

Figure 1. Sketch
geological map of
Kangimut sammisoq,
Ameralik fjord.
Lower structural
unit - Færingehavn
terrane; middle
stuctural unit -
Tre Brødre terrane;
upper structural
unit - Tasiusarsuaq
Terrane.

Qôrqut granite with Íkátoq gneiss rafts
  with Amîtsoq gneiss rafts

Íkátoq gneisses with anorthositic enclaves

Malene supracrustal rocks

Amîtsoq gneisses and Akilia association

Mid Archaean gneisses affected by
ca. 2800 Ma granulite facies metamorphism

flux of Pb-bearing hydrous fluids following
ca. 2700 Ma tectonic intercalation
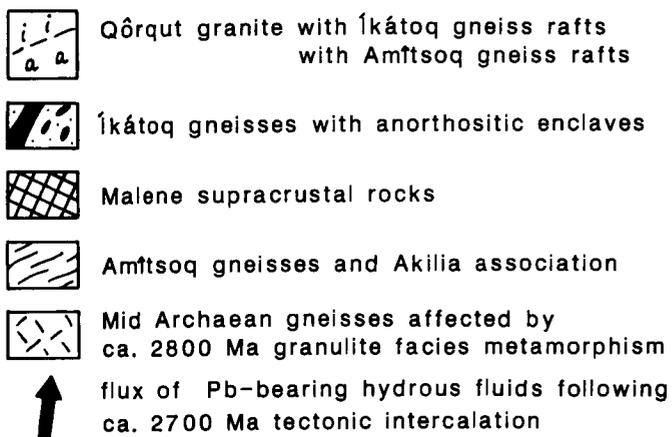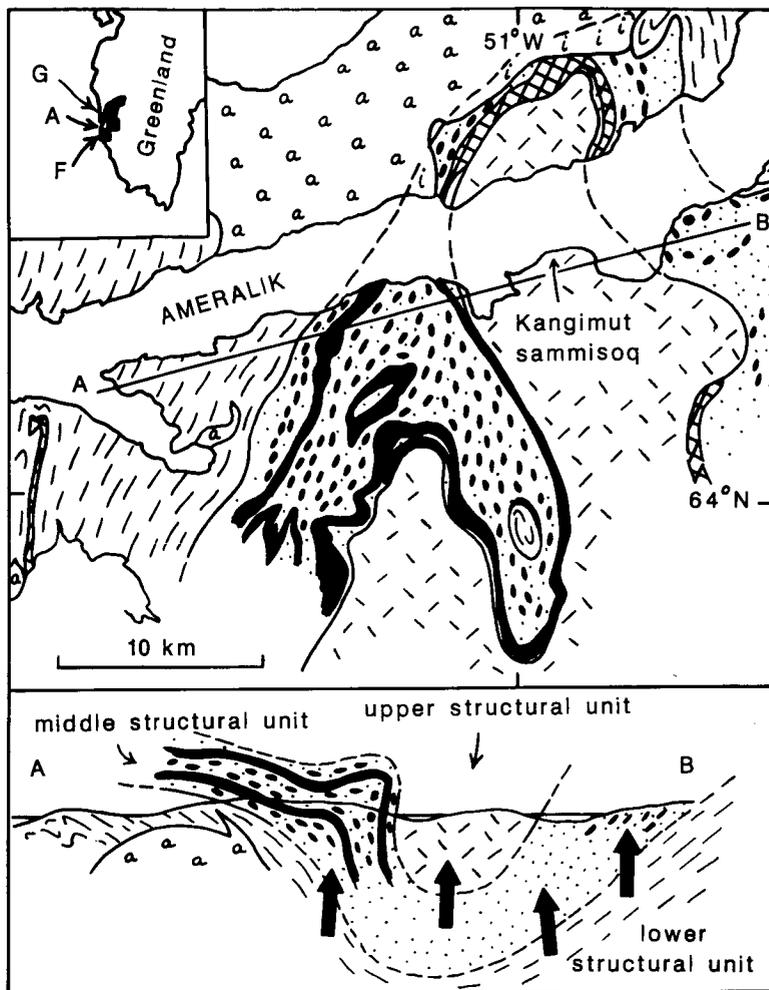
(e.g. 5, 10, 11 and references in these papers). The locality on
which this debate has been centered is Kangimut sammisoq (KS),
formerly known as Kangimut sangmissoq, Ameralik fjord (Fig. 1).
At KS the gneisses are rather inhomogeneous and contain
homogeneous, tabular amphibolite bodies, that resemble strongly
disruted Ameralik dykes. This gave rise to the suggestion that
the KS gneisses could be strongly reworked Amitsoq gneisses (5).
The KS rocks are almost totally retrogressed under amphibolite
facies facies conditions. Pb/Pb isotopic studies show that these

rocks contain a component of Pb derived from an early Archaean source such as Amitsoq gneisses (10). However, Nd, Sr and U-Pb zircon studies of them show no evidence of crustal residence prior to 3000 Ma (10 and P. Kinny, pers. comm., 1986).

One interpretation of the KS gneisses is that they are Amitsoq gneisses, and that their Amitsoq-like Pb component is more or less in situ, but has been mixed with more juvenile Pb during regional metasomatism, associated with the 2800 Ma granulite facies metamorphism (11). However, this interpretation required that these rocks had lost their whole rock Nd, Sr and zircon U-Pb isotopic signature of a >800 Ma crustal residence prior to 2800 Ma. On the basis of the current understanding of isotopic systematics of rocks affected by granulite facies metamorphism, we accept the isotopic evidence that the KS gneisses do not contain a detectable Amitsoq component. On the other hand we do support the suggestion that the KS gneisses and other similar rocks in the Tasiusarsuaq terrane are Nuk gneises whose magmatic protoliths passed through Amitsoq gneisses during their emplacement, and thus became contaminated with Amitsoq Pb (10,12). Instead we suggest another explanation of the KS gneisses, based on a better understanding of the region's structural history.

The KS gneisses are tectonically underlain by Ikatoq and Amitsoq gneisses (Fig. 1). During assembly of the three terranes and subsequent folding and shearing, the originally granulite facies KS gneisses were heterogeneously deformed, and almost totally retrogressed under amphibolite facies conditions. A gneiss from one of these shear zones contains 17 ppm Pb, whilst less deformed, but almost totally retrogressed KS gneisses have an average Pb content of 8 ppm. Hydrous fluid (containing Pb) necessary for retrogression of the granulite facies assemblages may have been derived from or passed through the underlying Ikatoq and Amitsoq gneisses (Fig. 1). Mixing of this added Pb to Pb present in the KS gneisses at the time of granulite facies metamorphism, would give rise to the scatter in the Pb/Pb isotopic plots for these rocks (10). Support for this explanation, which requires Amitsoq gneisses at depth, comes from the high I_{Sr} value (0.7060) for 2690 Ma (i.e. post assembly of the three terranes) granite sheets that cut the KS gneisses (13).

This example of Pb mobility associated with retrogression is a demonstration of the use of Pb isotopes to show the movement of and source regions of fluids during tectonic events and associated metamorphic recrystallisation. The recent structural studies (9) show that Godthåbsfjord consists of several terranes, the rocks of which are lithologically distinct, of different ages, and have under gone different metamorphic histories. Therefore, geochemical studies of gneisses from throughout the region cannot be used to monitor metasomatism associated with the 2800 Ma event which gave rise to granulite facies metamorphism solely in the Tasiusarsuaq terrane (14).

References
1) McGregor, V.R. (1973) Phil. Trans. R. Soc. Lond., A273, 343-358.
2) Moorbath, S., O'Nions, R.K., Pankhurst, R.J., Gale, N.H. & McGregor, V.R. (1972) Nature, 240, 78-82.

115

3) Baadsgaard, H. (1973) Earth Planet. Sci. Lett., 19, 22-28.

4) Pidgeon, R.T. & Kalsbeek, F. (1978) Can. J. Earth Sci., 15, 2021-2025

5) McGregor, V.R., Nutman, A.P. & Friend, C.R.L. (1986) Lunar and Planet. Inst. Tech. Rpt., 86-04, 113-139.

6) Bridgwater, D., McGregor, V.R. & Myers, J.S. (1974) Precambrian Res., 1, 179-198.

7) Wells, P.R.A. (1979) J. Petrol., 20, 187-226.

8) Coe, K. (1980) Precambrian Res., 11, 357-371.

9) Friend, C.R.L., Nutman, A.P. & McGregor, V.R. (1987) J. Geol. Soc. Lond., 144, 369-376.

10) Jones, N.W., Moorbath, S. & Taylor, P.N. (1980) Lunar and Planet. Inst. Tech. Rpt., 86-04, 59-62.

11) Collerson, K.D., McCulloch, M.T., Bridwater, D., McGregor, V.R. & Nutman, A.P. (1986) Lunar and Planet. Inst. Tech. Rpt., 86-04, 30-36.

12) Taylor, P.N., Moorbath, S., Goodwin, R. & Petrykowski, A. (1980) Geochim. Cosmochim. Acta, 44, 279-296.

13) Roberts, I.W.N. (1979) Unpublished Ph.D. thesis, University of Wales, Aberystwth.

14) Nutman, A.P., Bridwater, D. & McGregor, V.R. (1986) Lunar and Planet. Inst. Tech. Rpt., 86-04, 78-83.

ARCHEAN CRUSTAL EVOLUTION OF THE NORTHERN NORTH CHINA

CRATON; Qian Xianglin, Chen Yaping and Liu Jinzhong, Department of Geology, Peking University, Beijing 100871, People's Republic of China.

The Archean granulite facies rocks of the North China (Sino-Korean) Craton mostly occur inside the northern boundary forming an unique and spectacular granulite belt trending roughly E-W from eastern Hebei, North China in the east to Mt. Daqinshan, western Inner Mongolia in the west, ranging about 1 000 km long (1). Over the years in the middle portion of this Archean high-grade metamorphic belt a stratigrphic unconformity between the khondalite rock assemblage and the medium in composition granulite assemblage in Datong-Xinghe area is determined by the authors (Fig). It is first that the early Precambrian terranes of the Craton are differentiated into comparable stratigraphic sequences with different convincing tectonic events and therefore the tectonic evolutionary regime of the Craton and its growth in early Precambrian are significantly changed in interpretation. Prior to the authors of this paper the evolutionary stages of the Craton in early Precambrian have been recognized in terms of the consideration of the metamorphic volcano-sedimentary cyclicity with geochemical characteristics, but there is no indication of tectonic events and of stratigraphic sequences (2). However, the gneissic foliation of both rock assemblages seems to be concordant in strongly structural transposition during the highgrade metamorphism.

The banded coarse gneisses of the highly rich in aluminum and calcium khondalite suite are mainly composed of alternative in sedimentary cyclicity quartz feldspar gneiss, garnet sillimanite gneiss and hypersthene plagioclase gneiss with some layers of graphite gneiss near the bottom and thin bedded marbles at the top of the suite with a little volcanic layers intercalated. This khondalite suite in thickness of several kilometers unconformably overlies on the granulite assemblage - grey Gneiss Complex which comprises different types of granulites and banded iron formation with a large mount of tonalitic series of intusions which occupy the 50-80 % exposed area of the Gneiss Complex. The tonalitic intrusions in age of ca 2.6 Ga with coarse mafic xenoliths of different origin intruded into the granulite assemblage of the grey Gneiss Complex in age of greater than 2.8-3.0 Ga, and they were entirely intruded by the deformed and metamorphosed enclaves of fine-medium in grain size mafic dyke swarms, but those two kinds of intrusions are lacking in the widely distributed overlaying khondalite suite in which some ultramafic sills are in existence. The sedimentary rhythms of the khondalite suite is quite distinct and the younging sense of the sequence is clearly to be defined. It is worthy to mention that the near bottom succession of the suite in different localities along the exposures of the unconformity in the area (Fig) properly resemble each other in sedimentary origin of a great shallow and stable water basin. It indicates that the northern Craton was stabilized in Late Archean and a sedimentary
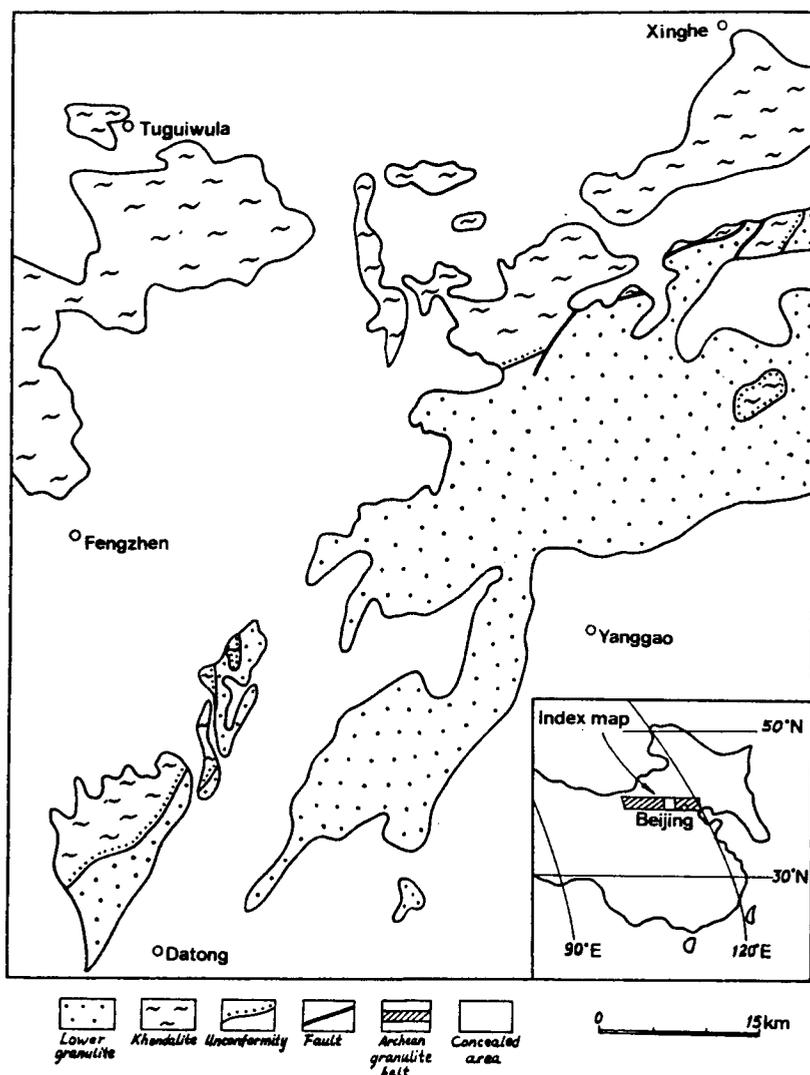
Figure, The exposed unconformity in the Archean granulite belt of the North China Craton.

cover was formed over the deformed basement.

In comparison with this platform sedimentary cover succession, the Archean granulite facies rock assemblage of the Wulashan Group westerly in Mt. Daqinshan in terms of distinct sedimentary sequence almost lacking in tonalitic series of intrusion and the Archean amphibolite facies rock assemblage of the Fuping Group southerly in the northern Mt. Taihang might be of an extended distribution of the khondalite sedimentary cover over the northern Craton. Moreover, this platform cover succession is suggested in a great possibility to have extended easterly to regions of Beijing and East Hebei, where it was eroded and the highgrade metamorphic granulite basement was exposed first on the early Proterozoic surface truncated and unconformably overlain by the early Proterozoic rifting volcano-sedimentary sequences (3,4).

ARCHEAN EVOLUTION OF THE NORTH CHINA CRATON
Qian Xianglin, Chen Yaping and Liu Jinzhong

The Archean granulites,representing the Archean lower crustal rock assemblages, was to be involved in a core complex in form of a granulite belt of an Andean type orogeny in latest Archean (5). The Archean rock assemblages in the belt had experienced polyphase metamorphism and a series of coarse sillimanite granite bodies were formed in situ by anatexis of khondalite suite along the coeval shear zones.

To the east of the Tancheng-Lujiang (Tan-Lu) fault zone the eastern part of the northern North China Craton in Liaoning and Jiling Provinces might have been a suspect terrane joined in the Cratonafter the formation of the granulite belt in latest Archean, because there is no resemblance between them in Archean crustal evolution.

References

(1) Qian Xianglin, Cui Wenyuan and Wang Shiqi, 1985. Records of Geological Research, Peking University, p20-29 (in Chinese).
(2) Chen Yuqi, Sun Dazhong and Wu Jiashan, 1984. J. Geodynamics, 1:251-277.
(3) Qian Xianglin, 1982. Collection of Structural Geology, No.2, p34-42 (in Chinese).
(4) Jahn Bor-Ming and Zhang Zongqing, 1984. Contrib. Mineral. Petrol., 85:224-243.
(5) Windley, B.F., 1981. In A. Kroner (Ed): Precambrian Plate Tectonics, Elsevier, p1-20.

GROWTH OF THE LOWER CONTINENTAL CRUST

Roberta. L. Rudnick, Research School of Earth Sciences, The Australian National Univ., G.P.O. Box 4, Canberra, A.C.T. Australia 2601

One of the largest uncertainties in crustal composition and growth models is the nature of the lower continental crust. Specifically, by what processes is it formed and modified, and when is it formed, particularly in reference to the upper crust? The main reason for this lack of information is the scarcity of lower crustal rock samples. These are restricted to two types: rocks which outcrop in granulite facies terrains and granulite facies xenoliths which are transported to the earth's surface by young volcanics.

Granulite facies terrains occur over large areas of the earth's surface and have been the subject of numerous petrologic, geochemical and isotopic studies. Of the 40 granulite terrains described in the western literature, 50% are Archean, 35% Proterozoic, 10% Paleozoic and 5% Mesozoic or Cenozoic (based on metamorphic ages). In contrast, lower crustal xenoliths are nearly all carried by young (Mesozoic to Cenozoic) basaltic or kimberlitic magmas which erupt through relatively young crust; there is only one occurrence of lower crustal xenoliths from an Archean region [1], of the remainder, 45% occur in Proterozoic crust and 55% occur in post-Precambrian crustal regions. The temporal differences between granulite facies terrains and granulite xenoliths may be one reason for the marked compositional differences between them: granulite terrains are dominated by felsic to intermediate lithologies, whereas granulite xenoliths are predominately mafic. However, a second explanation for this compositional disparity is that granulite facies terrains, by nature of their formation, are supracrustal in origin and provide little information on the lower continental crust.

Granulite Terrains

All granulite terrains contain some amount of supracrustal lithologies (i.e., metasediments), implying that at least some of these rocks formed at the earth's surface. Determining how these rocks were transported to the deep crust and then back to the surface provides important information on the tectonic settings in which granulites form. Newton and Perkins [2] suggested that the consistent equilibration pressures of 0.85 ± 0.15 GPa for massif granulites implied metamorphism of these rocks near the middle of a doubly thickened crust. However, P-T (pressure-temperature) *paths* rather than peak metamorphic conditions may allow more insight into tectonic settings. Recently described isobaric cooling paths for many granulite terrains may be interpreted to be the result of granulite formation in: (1) continental rift settings [3], (2) continental arc settings [4], (3) the lower plate of doubly thickened crust [5,6] or (4) the upper plate of doubly thickened crust if synmetamorphic intrusion occurs. Thus multiple interpretations are possible for a given P-T path and the ultimate recognition of tectonic settings for granulites may require delineation of P-T-t (where t = time) paths. If and when such tectonic constraints become available it is likely that granulite terrains will be shown to have formed in a variety of tectonic settings [7], thus some may be representative of the lower crust and others may not.

Granulite Xenoliths

Until such questions regarding the tectonic environment of formation for granulites terrains can be answered, the only *bona fide* lower crustal samples available are granulite xenoliths. The presence of high-pressure (i.e., 0.8-1.0 GPa) mineral assemblages and decompression features in many granulite xenoliths attest to their derivation from the deep crust at the time of eruption of the host.

To date, much fewer data are available on the composition of granulite xenoliths in comparison with granulite terrains. The available studies do indicate that many granulite xenolith suites are composed of mafic lithologies, although metasedimentary lithologies may be locally

important [8]. Thus basaltic underplating of the continental crust, coupled with some form of tectonic underthrusting may be important crust-forming processes and granulite facies metamorphism and partial melting are important lower crust modifying processes. Age information on lower crustal xenoliths is required in order to assess the role and significance of these processes through time. For example, are the abundant mafic lithologies found in xenoliths truly representative of the lower crust, or are they simply manifestations of the very volcanism which transported them to the surface?

Unfortunately, it can be very difficult to obtain reliable ages for lower crustal xenoliths. In order to use Rb-Sr and Sm-Nd whole rock methods to determine the age of a suite of xenoliths, the latter must first be shown to be cogenetic. This is rarely possible. Furthermore, if the magma giving rise to a cogenetic suite of xenoliths has variably assimilated country rocks, the interpretation of isochrons can be complicated by mixing trends. This may be a common phenomena for xenoliths which form by intrusion of mafic magmas into the hot lower crustal environment. Measuring internal isochrons for a given xenolith overcome many of these problems, however such dates can only represent the last metamorphic equilibration, which may not correspond to the peak metamorphism if the xenolith has cooled slowly in the deep crust. In addition, clean mineral separates may be very difficult to obtain in rocks containing several generations of metamorphic minerals. Zircon ages provide information on the metamorphic ages and possibly protolith ages. Yet conventional zircon analyses are difficult for xenoliths because of the small sample sizes and the mafic composition of many xenoliths. In addition, high-grade rocks commonly have multiple zircon generations, which may or may not be distinguished visually. Table 1 summarizes the available age information on lower crustal xenoliths.

Table 1. Dates from lower crustal xenoliths

Locality	Rock type	Method(s)	Age (Ma)	Interpretation	Reference
Partan Craig, Scotland	Felsic granulite	Sm-Nd mineral	360	Metamorphic age (time of basaltic volc)	9
	Felsic granulite	U-Pb zircon	>2000	Highly discordant	10
Lesotho, southern Africa	Mafic cumulates and felsic granulite	Sm-Nd WR	1400	Crystallization age	11
Calcutterroo, South Australia	Mafic granulites and felsic granulite	Sm-Nd and Rb-Sr WR	2170 1300	Nd Model age (felsic) Mixing age?	12 13
Eifel, West Germany	Mafic granulites	Sm-Nd mineral	175	Metamorphic age	14
Chudleigh, Queensland	Mafic cumulates	Nd-Sr	≤100	Mixing age	13
McBride, Queensland	Mafic, intermediate and felsic granulites	U-Pb zircon	250-350 300 and 1570	Metamorphic age Crystallization ages	15

The two granulite xenolith suites from north Queensland provide an interesting contrast in lower crustal evolution. Vents of the < 10 Ma old Chudleigh volcanic province are situated near a major fault boundary between Proterozoic metamorphics of the Georgetown Inlier and Paleozoic rocks of the Tasman fold belt. Crustal xenoliths are all mafic cumulates, ranging from plagioclase-rich to pyroxene-rich varieties, and equilibrated between 20 and 40 km depths [16].

GROWTH OF THE LOWER CONTINENTAL CRUST
R. L. Rudnick

Major and trace element concentrations and present day Nd and Sr isotopic compositions of the xenolith suite show excellent correlations and demonstrate the cogenetic nature of the xenolith suite. These xenoliths are samples of cumulates crystallized from a magma which underwent simultaneous assimilation and fractional crystallization within the deep crust [13]. Because of the highly variable whole rock Sm/Nd ratios of the xenoliths, the Sr-Nd isotopic correlations and the isotopic-trace element correlations degrade as the isotopic data is back-calculated to earlier times, suggesting that the xenoliths are not older than 100 Ma. Thus this suite of xenoliths appears to be related to an earlier stage of the Cenozoic volcanism.

In contrast with the Chudleigh province, the < 3 Ma xenolith-bearing alkali basalts of the McBride province, north Queensland, erupt thorough Proterozoic rocks of the Georgetown Inlier. The lower crustal xenoliths from this province have diverse lithologies, ranging from mafic through felsic compositions. The xenoliths' protoliths formed through diverse processes including underplating and crystallization of basaltic magmas, crystal accumulation from basaltic magmas, partial melt removal from more evolved rock types leaving a mafic residue, crystallization of felsic magmas and crystal accumulation from felsic magmas [17]. U-Pb ion microprobe analyses of zircons from 7 of the xenoliths show that the ages of the xenoliths correlate well with the major orogenic episodes manifest at the earth's surface [15]. All xenoliths underwent high-grade metamorphism in the late Paleozoic (~300 Ma ago), when voluminous felsic ash-flows were erupted on and high-level intrusives were emplaced into the upper crust. This event was accompanied by intrusion of mafic mantle-derived melts into the lower crust which caused anatexis of pre-existing crust. The 100 Ma span of concordant ages for these late Paleozoic zircons are probably due to slow cooling in the lower crust after the orogeny. Several xenoliths contain Proterozoic zircons with ages of ~1570 Ma, corresponding to the time of regional amphibolite facies metamorphism in the Georgetown Inlier. There are no zircon ages less than 200 Ma, suggesting that, in contrast with the Chudleigh xenoliths, none of these samples are related to the Cenozoic igneous activity.

The important conclusions arising from the forgoing xenolith studies are:

(1) The majority of mafic lower crustal xenoliths formed through cumulate processes, restitic xenoliths are rare. Thus basaltic underplating appears to be an important lower crust-forming process, which, based upon the available age determinations, has occurred during multiple episodes for any given crustal region.

(2) Formation and metamorphism of the deep crust is intimately linked to igneous activity and/or orogeny which are manifest in one form or another at the earth's surface. Therefore, estimates of crustal growth based on surface exposures is representative, although the proportion of remobilized pre-existing crust may be significantly greater at the surface than in the deep crust.

References

[1] Halliday, A., et al. (1985) Scott. J. Geol. 4: 465-491. [2] Newton, R.C. and Perkins, D., III. (1982) Am. Mineral. 67: 203-222. [3] Sandiford, M. and Powell, R. (1986) Earth Planet. Sci. Lett. 79: 151-158. [4] Wells, P.R.A. (1980) Earth Planet. Sci. Lett. 46: 253-265. [5] England, P.C. and Thompson, A.B. (1985) J. Petrol. 25: 894-928. [6] Ellis, D.J. (1987) Geology (in press). [7] Windley, B.F. (1981) J. geol. Soc. London 138: 745-751. [8] Kay, R.W. and Kay, S.M. (1981) Rev. Geophys. Space Phys. 19: 271-297. [9] van Breeman, O. and Hawkesworth, C.J. (1980) Trans. R. Soc. Edinburgh 71: 97-102. [10] Halliday, A.N., et. al (1984) Trans. R. Soc. Edinburgh 75: 71-74. [11] Rogers, N.W. and Hawkesworth, C.J. (1982) Nature 299: 409-413. [12] McCulloch, M.T. et al. (1982) Nature 300: 166-169. [13] Rudnick, R.L., et al. (1986) Geochim. Cosmochim. Acta 50: 1099-1115. [14] Stosch, H.-G. and Lugmair, G.W. (1984) Nature 311: 368-370. [15] Rudnick, R.L. and Williams, I.S. (1987) Earth Planet. Sci. Lett. (in press). [16] Rudnick, R.L. and Taylor, S.R. (1987) Spec. Publ. Geol. Soc. Australia (in press). [17] Rudnick, R.L. and Taylor, S.R. (1987) J. Geophys. Res. (submitted).

THE COMPLEXITY, DEPTH, AND RAPIDITY OF PROCESSES THAT FORMED THE LUNAR CRUST.

Graham Ryder[1] and Julius Dasch[2] [1]Lunar and Planetary Institute, 3303 NASA Road One, Houston, TX 77058. [2]NRC/NASA Johnson Space Center, Houston, TX 77058 and Oregon State University, Corvallis, OR 97331.

From its birth the Moon had a large-scale, complex magma system. The evidence is the massive differentiation of the Moon that has been partially preserved. The system may have been a magma ocean or a magmasphere; even in the former case it was superposed by smaller but also very complex magma systems. The main episode produced a plagioclase-rich crust including genuine anorthosites; it was over by about 4.35 b.y. ago, although magmatism continued. The processes of crust-building remain in serious dispute.

1). What is the lunar crust?-Prior to the space-age, the nature of the lunar crust (if it had one) was anyone's guess, and apparently none surmized correctly. Geophysical data show that the Moon has a low density crust composing about 10% of its volume, overlying an ultramafic mantle (the Earth's crust is only 1% of its volume). Apollo seismic data indicate that this crust is about 50 km thick on the nearside; the offset of the lunar center of mass from its center of figure (towards the Earth) is generally interpreted to result from a thicker (75 km?) crust on the farside. Apollo samples demonstrate that the low density phase is plagioclase; for pure anorthite a little over 20 km worth would be in the outer regions of the nearside [1]. The lunar crust is not anorthosite, although at least its upper part is anorthositic.

2). Evidence for massive differentiation.-(i) The crust is very rich in plagioclase that requires complete extraction from about 30% of its volume, or a depth of 200 km, assuming a Moon similar to the silicate portion of chondrites, which it geophysically resembles. More plagioclase, or less efficient extraction, requires a thicker melted zone. (ii) Much of the plagioclase occurs in the rock type **ferroan anorthosite**; mafic complements to this rock type have not been found, so the differentiation reached far below the depth of penetration of even large basins. (iii) Mafic rocks ((**dunites, troctolites, norites, and gabbros**) constitute a significant though not established portion of the crust. They are not easily relatable to whatever magma system produced the ferroan anorthosites, nor in most cases to each other. Orthopyroxene rather than clinopyroxene in most of these crustal rocks implies large-scale melting or previously-differentiated sources or both; the trace elements require differentiated sources or complex assimilation processes [2]. (iv) **Mare basalts** (erupted from perhaps 4.3 to about 3.0 Ga or even later) have negative Eu anomalies and other non-chondritic trace-element relative abundances. Plagioclase is rarely near their liquidii, and the trace element characteristics are believed by most to be necessarily inherited from the sources, not from complex later processes. Mare basalt sources were at depths of 200-400 km [3]. These sources are differentiated, having lost plagioclase and in some cases gained ilmenite. Their origin was certainly complex; nonetheless the effects of differentiation were ultimately felt through at least the outer half of the Moon. (v) The **trace element-rich KREEP** is a common chemical component of highland polymict rocks. Wherever found, it has a strangely uniform trace element abundance pattern [4]. It is most easily explained as the end-product of fractional crystallization, probably on a global scale or by a commonly repeated process. It was later reworked by large-degree partial melting volcanism and by impact redistribution. A very large proportion of the bulk lunar incompatible element budget is now in the crust.

3). Timing of massive differentiations.- The chronology of the early lunar crust is discussed in a companion abstract [5]. The ferroan anorthosites date near to the origin of the Moon, according to their low $^{87}Sr/^{86}Sr$ ratios; according to model ages based on BABI they cannot be much younger than 4.56 Ga. An internal Sm-Nd isochron by [6] for ferroan anorthosite 60025 corresponds to an age of $4.44+.02$ Ga; some of the more mafic rocks ("Mg-suite") apparently have even older ages. Mare basalt sources closed at about 4.35 Ga for Pb, Sm, and Sr systems, and the complementary isotopic and trace element characteristics of KREEP, the most evolved component, were formed by this time. A few mare basalt samples show that at 4.3 Ga there was a stable crust onto which lavas could flow. Thus it is evident that the main part of the crustal formation was complete by 4.35 Ga, and the system had cooled below the closure temperatures of the main non-gaseous radiogenic isotopic systems, i.e. to depths of the

order of a few hundred kilometers. After 4.35 Ga, partial melting of mantle and crustal sources continued, but not on a global scale. By 3.9 Ga, the crust had cooled enough for the Apennine Front (Imbrium ring) to remain isostatically uncompensated; nonetheless the crust appears to have been simultaneously hot enough in the same region to have melted and formed KREEP volcanism (perhaps by a combination of heat input and pressure release from the Imbrium impact). The essential point of the chronology however is that crustal and mantle Sm-Nd (and other trace element) evolution departed from chondritic early, when major crustal development took place. That departure has been preserved, and its effects influence and are demonstrated by subsequent melting events. The ferroan anorthosites appear to have formed prior to the isotopic departures from chondritic i.e. they appear to have been the earliest rocks.

4). Nature of crustal materials.- The intense cratering of the highlands up to 3.8 Ga ensures that almost all of the samples collected are polymict, commonly melted, breccias. Only rare samples can be recognized as having preserved an igneous chemistry (lacking meteoritic siderophile contamination, or admixture of polymict materials). Few of these have retained vestiges of an igneous texture. The relationship of these "pristine" rocks to the polymict rocks and highlands compositions is disputed, so the actual rock types in the crust and their relative proportions are loose parameters at present. The search for more (small) igneous samples hidden away in the Apollo collection, mainly as clasts in breccias, continues. So do the deconvolutional attempts to understand the chemistry of the polymict rocks.

Highland (crustal) rocks are almost entirely plutonic, perhaps because volcanics are the most susceptible to impact comminution. They comprise three distinct(?) suites: ferroan anorthosites, Mg-suite, and KREEP. At least the Mg-suite itself cannot be from a single magma system, but is a polyglot. Unfortunately, radiogenic isotopic data for many of the Mg-suite samples is lacking; some of that available is confusing.

(i) Ferroan anorthosites probably formed from magmas with roughly chondritic trace element patterns [6,7] very early in lunar history. They are vestigiously coarse-grained; mafic minerals (rarely more than a few %) are homogeneous within a sample, except for rare genomict samples.. Ferroan anorthosites are among the most slowly-cooled crustal rocks known in the solar system. They form a fractionated sequence (Mg' 70-40) but contain no evidence for trapped liquid, containing very low abundances of incompatible elements. Their lack of complementary mafic to ultramafic rocks (or even mineral fragments) suggests that they formed in a magma system larger than the depths of excavation of even the large lunar basins. They also appear to be of global extent [8], even making allowance for plagioclase that should not be attributed to anorthosite i.e. the plagioclase in norites, troctolites, and basalts.

(ii) Mg-suite rocks all have trace element patterns more evolved than the ferroan anorthosites, yet are more magnesian for a given plagioclase composition. A wide range of textures and mineral chemistries suggests varied cooling environments e.g. dunite 72415 has slightly zoned olivine crystals which contain higher calcium contents than the absolutely homogenous olivines in troctolite 76535. Most have Ti/REE ratios indicating an origin from evolved, probably mixed, magmas, unlike the chondritic Ti/REE of ferroan anorthosites [2]. Most have low Ca/Al ratios (orthopyroxene-dominant); some have higher ilmenite, augite, and Ti/REE. The ages of the Mg-suite samples appears to be varied, but also confusing (see companion abstract).

(iii) KREEP is rare as igneous rocks; instead its presence dominates the trace elements and radiogenic isotopes of polymict breccias. Volcanic KREEP was sampled among the Apollo 15 and 17 samples, and is 3.8, 3.9 Ga old respectively. An evolved plutonic(?) sample ("quartzmonzodiorite") produced a Pb zircon age of 4.35 Ga [9]. The essential characteristics of KREEP had been produced by 4.35 Ga; activity until 3.8 Ga moved it to the surface with little effect on its trace element patterns. Even more evolved (very tiny) samples, roughly "granite", have been found especially among the Apollo 14 samples; their varied characteristics and limited isotopic data preclude general definitive statements about their style and time of petrogenesis.

5). Processes which produced crustal materials.- Even now the processes which produced the lunar crust are disputable. They were reviewed by Warren [10]. Their elucidation depends on gaining a better understanding of the igneous rocks in the highlands, the unravelling of the chemistry of polymict rocks,

and understanding the lateral and vertical variations of rock types in the crust (e.g. [8]). Walker [11] abandoned the concept of a lunar magma ocean at all, finding serial magmatism adequate to explain crustal rock types. Longhi and Ashwal [12] suggested the mechanical (diapiric) separation of anorthosites from their mafic complements. Nonetheless, ferroan anorthosites could have been produced in a magma ocean, one which had a roughly chondritic abundance pattern of refractory incompatible elements (they could indeed have floated, unlike most Mg-suite samples). The parent had evolved from a volatile-depleted chondritic (bulk Moon) composition mainly by partial melting and fractional crystallization of olivine and some pyroxene. A very large scale system is required to account for the absence of mafic cocumulates, the amount of anorthosite in the crust, and the non-chondritic nature of mare basalt sources. A magma ocean might not be required if the Eu anomaly of mare sources can be explained by near-surface complex processes and if the crust does not have a positive Eu anomaly overall (11). Production (i.e. initial crystallization) of plagioclase crust from a magma ocean is not itself very simply explained [13]. The Mg-suite rocks require a separate set of origins, presumably later than the magma ocean, because their origin requires the presence of evolved materials, KREEP or KREEP-like, and at an early stage. They formed from multiple magmatic episodes, but the primitive Mg's of some of the Mg-suite rocks suggests large-scale melting. Possibly that a massive overturn of a density-unstable mantle following rapid crystallization of a magma ocean caused massive melting of uprising Mg-mantle as well as causing sinking of the Fe-rich, Eu-depleted sources for the later mare basalts. The residual liquid from the earliest ocean (Ur-KREEP) may have played a considerable role in influencing the chemistry of varied Mg-suite rocks. Whatever the nature of these events, they were pretty much history by 4.3 Ga, by which time KREEP and the sources of mare basalts had cooled below their closure temperatures. Later magmatic activity included possible Mg-suite plutonic magmas, and certainly remelting of KREEP to form volcanic rocks. Even polymict KREEP rocks have essentially basaltic (cotectic) compositions indicating magmatic rather than solely impact control.

6. Differences from Earth.- The Moon preserved its evidence of early differentiation, in apparent contrast with the Earth, because of the early shut-down of its magma systems, or at least major mantle convection. An important point is that on the Moon a lot of complex things happened very rapidly even though it is a rather small body (the eucrite parent body did not have such complexity). Does this mean that the early terrestrial evolution was also very complex, but that continued activity and remixing has ironed things out into a simpler system again? Are there terrestrial relics of this early complex history? Clearly the Earth's early history could not have been very similar to that of the Moon, because not so much plagioclase could have contributed to a terrestrial crust.

References. [1] Wood J. (1986) Origin of the Moon, LPI, Houston, 17. [2] Norman M. and Ryder G. (1980) PLPSC-11, 317. [3] Delano (1986) PLPSC-16, D201. [4] Warren P. and Wasson J. (1979) Rev. Geophys. Space Phys. 17, 73. [5] Dasch E.J et al (1987) this volume. [6] Lugmair G. (1987) LPS XVIII, 584 and pers comm. [7] Ryder G. (1982) Geochim. Cosmochim. Acta 46, 1591. [8] Spudis P. and Davis P. (1986) PLPSC 17-E84. [9] Compston W. et al. (1984) LPS XV, 182. [10] Warren P. (1985) Ann. Rev. Earth Planet. Sci. 13, 201. [11] Walker D. (1983) PLPSC-14, B17. [12] Longhi J. and Ashwal L. (1985) PLPSC-15, 571. [13] Morse S. (1982) PLPSC-13, A10.

CONTINENTAL CRUSTAL FORMATION AND RECYCLING : EVIDENCE FROM OCEANIC
BASALTS. A.D. Saunders, J. Tarney and M.J. Norry (Department of Geology,
University of Leicester, U.K.).

Despite the wealth of geochemical data for subduction-related magma
types, and the clear importance of such magmas in the creation of continental
crust, there is still no concensus about the relative magnitudes of crustal
creation versus crustal destruction (i.e. recycling of crust into the
mantle). The role of subducted sediment in the formation of arc magmas is now
well documented; but what proportion of sediment is taken into the deeper
mantle?

Integrated isotopic and trace element studies of magmas erupted far from
presently active subduction zones, in particular basaltic rocks erupted in
the ocean basins, are providing important information about the role of
crustal recycling (e.g. Palacz and Saunders, 1986). By identifying potential
chemical tracers, it is possible to monitor the effects of crustal recycling,
and produce models predicting the mass of material recycled into the mantle
throughout long periods of geological time.

Ocean island basalts (OIB) exhibit extremely diverse compositions, which
doubtless reflects the number of processes and components involved in the
formation of their respective sources. HIMU or St Helena-type OIB, with high
$^{206}Pb/^{204}Pb$ and low $^{87}Sr/^{86}Sr$ ratios, also have low K/Ta, Rb/Ta and Ba/Ta
ratios (lower than in most mid-ocean ridge basalts (MORB)). These data are
consistent with derivation from a source which has retained low Rb/Sr, K/Ta,
Rb/Ta and high U/Pb ratios for long periods of geological time. Such a source
may be produced by introduction of recycled <u>oceanic</u> crust into the mantle,
(Hofmann and White, 1982), perhaps as a distinct megalith layer at the
core-mantle boundary or on the 650 km discontinuity. Virtually no sedimentary
component can be detected in these basalts.

Other OIB types - including the so-called DUPAL OIB - have variably high
$^{87}Sr/^{86}Sr$, $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios. These basalts have higher
Rb/Ta, Ba/Ta and K/Ta ratios than the St Helena-type OIB, consistent with an
'extra' component of added continental crustal material (subducted
sediment?). However, the proportion of sediment required to produce the
raised K/Ta, Ba/Ta etc ratios is very small - possibly much less than one per
cent. This is because, relative to mantle abundances, sedimentary rocks
contain two to three or even four orders of magnitude higher abundances of
lithophile elements.

Pb-isotope sytematics require that the St. Helena-type OIB and the DUPAL OIB
sources remain chemically separate for long periods of time, and it has been
suggested that the DUPAL reservoir originates in the sub-continental
lithophere. We propose a model whereby the 'sedimentary component' of the
DUPAL reservoir is created in the sub-continental mantle, as a result of
subduction processes contemporaneous with the formation of the continental
crust, during Archaean times. More recent detachment of this material
(McKenzie and O'Nions, 1983) makes it available for incorporation into the
OIB source. However, sub-continental mantle is not, alone, sufficient to
produce the characteristics of the DUPAL OIB source. The high Ta and Nb
abundances of all OIB require a further component - the subducted oceanic
crust megalith layer. We suggest that plumes ascend from this megalith layer

CRUST RECYCLING: EVIDENCE FROM OCEAN BASALTS
Saunders, A.D., Tarney, J. & Norry, M.J.

126

and interact with diffuse buoyant zones of (previously subcontinental) lithospheric mantle, which now lie in the upper mantle benath the ocean basins, corresponding to geoid highs. Where ascending plumes rise direct to the surface, without interacting with the fragments of ancient lithopshere, they erupt to form St Helena-type OIB. Where they do interact, they form DUPAL OIB, the proportion of the DUPAL component depending on the mixing relationships.

The model to be presented implies that little or no sediment is recycled into the deep megalith layer. Most must be stripped off in the subduction zone, to be incorporated in arc volcanics or trapped in the sub-continental lithosphere, to be eventually returned in the DUPAL-type OIB mantle.

REFERENCES

Hofmann, A.W. and White, W.M., 1982. Mantle plumes form ancient oceanic crust. Earth planet. Sci. Lett., 57, 421-436.

McKenzie, D.P. and O'Nions, R.K. 1983. Mantle reservoirs and ocean island basalts. Nature, 296, 821-825.

Palacz, Z.A. and Saunders, A.D., 1986. Coupled trace element and isotope enrichment in the Cook-Austral-Samoa islands, southwest Pacific. Earth planet. Sci. Lett., 79, 270-280.

GEOCHRONOLOGICAL AND LEAD-ISOTOPE EVIDENCES FOR RAPID CRUST FORMATION IN MIDDLE-PROTEROZOIC TIME: THE LABRADOR EXAMPLE

Urs Schärer, Université du Québec à Montréal, C.P. 8888, Succ. A, Montréal, Québec, H3C 3P8, Canada.

Extensive U-Pb geochronological studies in the Grenville and Makkovik provinces have shown that eastern Labrador is underlain by two distinct crustal blocks (Fig. 1, see also references 1-3). The northern block is composed of Archaean basement, lower Proterozoic intrusives and volcano-sedimentary supracrustal sequences. Regional metamorphism at about 1780 Ma ranging in degree from greenschist to amphibolite facies conditions affected all supracrustal units. This metamorphism occurred during a final phase of Makkovikian Orogeny (3, 4) and reflects the youngest event recognised in the northern block. In contrast, for the southern block U-Pb data obtained on zircon, monazite and titanite demonstrate that primary magmatic activities occurred 1710 to 1630 Ma ago postdating significantly the latest events in the north. These activities lead to the emplacement of large volumes of mafic to intermediate plutonic complexes including layered gabbro-norite intrusions and tonalitic to granodioritic bodies. These rocks, which now underlie the Labrador part of the Grenville Province and the Trans-Labrador Batholith were intruded by different phases of anorogenic plutonism at 1500 Ma (pegmatites), 1420 Ma (Michael Gabbros) and 1300 Ma (syenites-granites) and were overprinted to various degrees during Grenvillian Orogeny. Metamorphism in Grenvillian time locally reached melting conditions with migmatite formation and granite emplacement. Beside melting, Grenvillian orogeny at 1030 to 960 Ma caused large-scale thrusting of crustal segments, associated with gneissification of most 1710-1630 Ma rocks and local new-growth of zircon, monazite and titanite. Despite strong Grenvillian overprint primary mineral populations could be, isolated allowing the unambiguous determination of primary crystallisation ages.

With the exception of two rocks (1710 and 1680 Ma old) all basement rocks in the southern block were formed in a very short time interval from 1660 to 1630 Ma ago. This suggests that a large portion of crust was formed in a short period of about 30 Ma. The emplacement of the plutonic complexes was accompanied by distinct phases of high-grade metamorphism causing local melting and severe mineralogical and structural overprint only a few million years after primary rock formation (1,2).
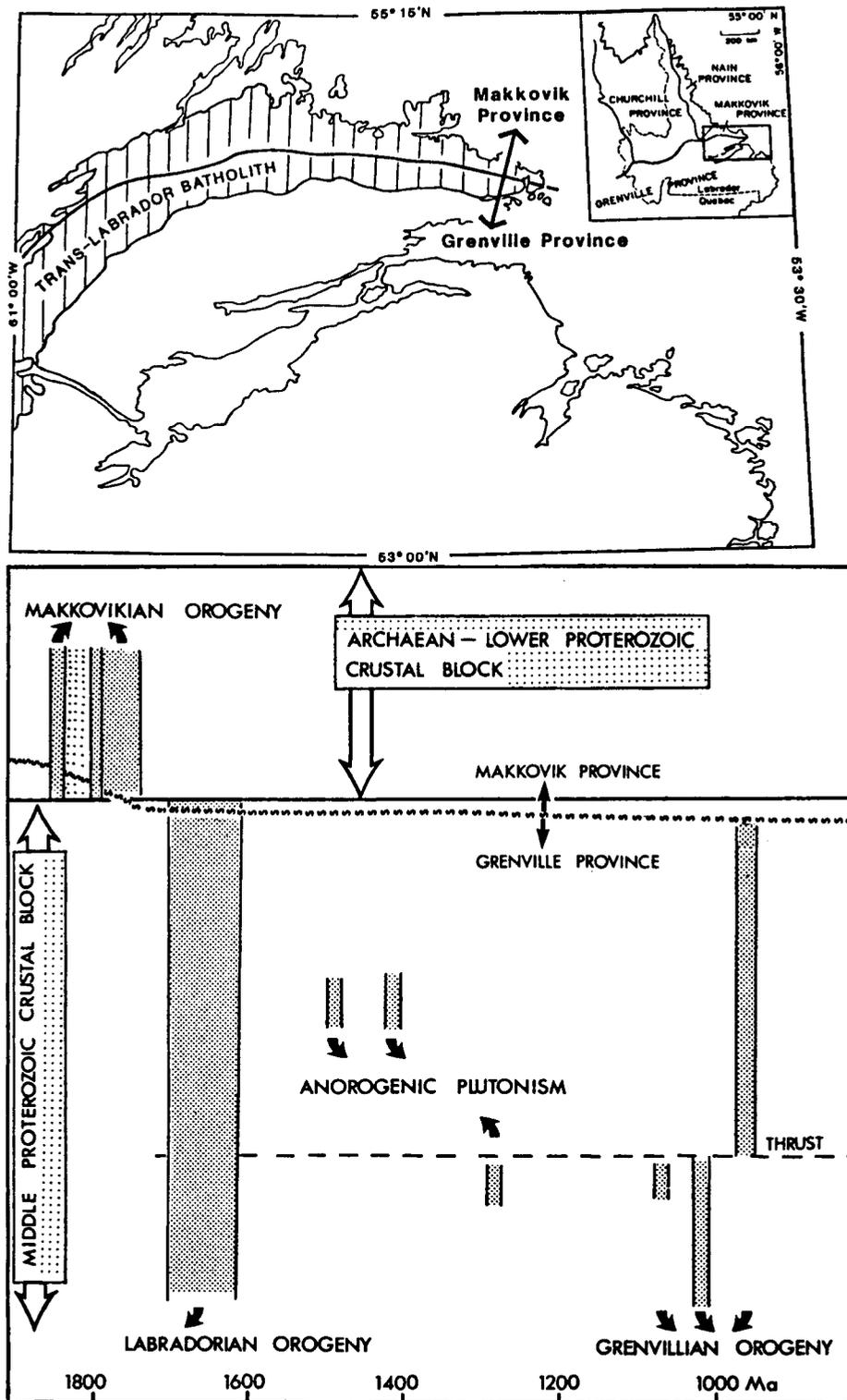
Figure 1: Tectonic setting of the study area, and schematic summary diagram showing the different magmatic and tectonic events (as distinguished from U-Pb ages, reference 1-3) plotted vs. time and tectonic setting.

The striking lack of ages older than 1710 Ma in the southern block, and the absence of inherited zircons (with one exception) suggest that significant amounts of juvenile crust was formed, and added to the northern block during Labradorian Orogeny (1,2, and 5) between 1710 and 1630 Ma ago.

In order to substantiate the juvenile character of the middle-Proterozoic crustal block, the isotopic composition of lead in leached K-feldspars from the same rocks were analysed (6). Since K-feldspars are characterised by very low uranium/lead ratios, their lead isotopic compositions still reflect the initial isotopic compositions at the time of crystallisation. These lead isotope analyses demonstrate that the 1710 to 1630 Ma crust was derived from sources with time integrated uranium/lead ratios close or slightly above the mantle reference line proposed by Zartman and Doe (7), thus laying significantly below the two-stage lead evolution curve of Stacey and Kramers (8), and far below the model curve for upper crust (7). An origin from primitive sources, most likely the mantle, is corroborated by similarities with lead isotopic compositions in sulfides from the about 1780 Ma old Komatiites in the Cape Smith belt (9). Furthermore, the lead isotope data show that the "Labradorian" crustal segment preserved its primitive isotope signature even through Grenvillian melting. This implies that the 1710-1630 Ma old crust has suffered a uranium-depletion event during or shortly after its formation. i.e. the rocks evolved with significantly lower uranium/lead ratios relative to average crust (8) or upper crust (7). This depletion in uranium can be ascribed to regional high-grade metamorphism which accompanied magmatism during Labradorian Orogeny producing granulite facies mineral assemblages in most of the rocks. The Labrador example of syn-magmatic high-grade metamorphism, causing an early depletion in uranium and other elements such as Rb and K, might serve as a model for other granulite facies terrains.

References:

(1)   Schärer U. et al. (1986) Contrib. Mineral. Petrol. 94, p. 438-452.

(2)   Schärer U. and Gower C.F. (1987) Precambrian Res. (in press).

(3)   Schärer U. et al. (1987) Can. J. Earth Sci. (submitted).

(4)   Gower C.F. and Ryan B. (1986) Geological Association of Canada, Special Paper 31, p. 175-189.

(5)   Thomas A. et al. (1986) Geological Association of Canada, Special Paper 31, p. 281-296.

(6)   Schärer U. , unpublished data

(7)   Zartman R.E. and Doe B.R. (1981) Tectonophysics 75, p. 135-162.

(8)   Stacey J.S. and Kramers J.D. (1975) Earth Planet. Sci. Lett. 26, p. 207-221.

(9)   Brévart O. (1985) Ph.D thesis, Université de Paris VII, Paris, France.

PHYSICAL PROCESSES IN THE GROWTH OF THE CONTINENTAL CRUST;

G. Schubert, Department of Earth and Space Sciences, University of California, Los Angeles, Ca 90024-1567

$CD / 46617$

Major mechanisms of crustal addition are volcanism and plutonism at plate boundaries and within plate interiors. Island arc magmatism is the primary manifestation of plate boundary processes while hotspot magmatism is the major intraplate process. Island arc magmatism has been the predominant crustal addition mechanism during the Mesozoic-Cenozoic, but hotspots have been a significant source of new crust. The major question about ancient crustal growth processes is whether island arcs have always dominated hot spots. It is important to realize that hot spots are as basic and integral a component of mantle convection as are island arcs. Hot spots are the surface manifestation of mantle plumes which originate from the unavoidable instability of the hot lower thermal boundary layers at the core-mantle interface that constitute the large-scale convection system of the plates. Some fully 3-dimensional numerical simulations of mantle convection in a spherical shell show that plumes can dominate upwelling while linear structures (sheets) dominate downwelling. That plumes are capable of transporting a great deal of heat is amply demonstrated by the Jovian moon Io; just a few hotspots on the satellite give off as much heat as the entire geothermal heat flow.

Mechanisms of crustal subtraction include sediment subduction, delamination and subduction erosion. Delamination is illustrated by numerical calculations of the instability of a cold upper boundary layer suddenly emplaced above a low-viscosity asthenosphere. The foundering of the lithosphere influences the upwelling of plumes and their interaction with the lithosphere; numerical simulations depict the phenomena.

One approach to deciding if island arc magmatism dominated ancient crustal growth is to assess the rate at which the process has operated in the recent past. We have surveyed the volumes of new material in island arcs and have correlated those volumes against the times of operation of the arcs. All the data for Mesozoic-Cenozoic arcs fall in the range 20 to 40 km^3 km^{-1} Ma^{-1}. However, according to Sm-Nd isotopic data, a number of Precambrian terrains including the Arabian-Nubian Shield, the Canadian Shield, and the West Central U.S. have grown at rates about an order of magnitude larger than recent arc addition rates. Another way to view the arithmetic is to compare total growth rates in km^3 yr^{-1} to the present cumulative arc addition rate of about 1 km^3 yr^{-1}. These ancient terrains grew at rates comparable to the Mesozoic-Cenozoic worldwide addition rate. These localized addition rates are comparable to present day global rates. The major conclusion is that formation of these crustal segments required plate margin and/or hot spot tectonics operating at rates considerably in excess of more recent rates. Further, these enhanced rates cannot simply be explained by higher heat flow in the past since heat flow was only slightly larger for the Arabian-Nubian shield at the end of the Proterozoic, for example. Temporal variability in the mantle convection engine must be responsible for the enhanced rates of crustal production.

One physical observable that has been used to constrain models of crustal growth is sealevel. Since the area available to the oceans depends on the area of the continents, continental growth or decline over long periods of time is directly measurable by secular variations in

90   PHYSICAL PROCESSES IN THE GROWTH OF THE CONTINENTAL CRUST
        Schubert, G.

132

sealevel. While there have been short term variations in sealevel, due to glaciation-deglaciation cycles, for example, sealevel has remained within ±200 m of its present value over the last 250 Myr. This observation has been termed the constancy of freeboard (the height of the continents above sealevel) and has been applied to the entire Phanerozoic and even further back in time. We have developed a simple physical model to explore the consequences of constant freeboard, or any other assumption about freeboard variation, on continental growth. Because the ocean basins deepen as the Earth's heat flow declines with time, constancy of freeboard requires a net crustal growth of about 25% or 1 km^3 yr^{-1} since the end of the Archean. A post-Archean increase in freeboard by 200 m requires continental growth of only 10%, while a decrease in freeboard by 200 m during this same period necessitates a crustal growth of 40%. Shrinkage of the continental crust since the end of the Archean is highly unlikely.

Global geoid and seafloor topography data have been used to identify and study oceanic plateaus and swells that have either continental crustal roots or anomalously thick oceanic crusts. The crustal volumes of the features have been determined from their bathymetries and the assumption of Airy isostasy. Our results provide an accurate assessment of the total volume of fragmented continental crust trapped in the seafloor and of the volume of oceanic crust likely to be incorporated into the continents. Our results can also be used to address the question of the minimum size of continental fragments necessary to avoid subduction into the mantle.

EVOLUTION OF THE DHARWAR CRATON - A TERRAIN OF EARLY ARCHAEAN CRUSTAL STABILITY, LONG TERM OROGENIC CYCLES AND LARGE SCALE PALAEOBIOLOGICAL ACTIVITY

R.Srinivasan and S.M. Naqvi

(National Geophysical Research Institute, Hyderabad-500007)

Traceable history of the Dharwar Craton goes back to the ~3400 m.y. old tonalitic to trondhjemitic Fundamental Gneiss whose REE composition indicates its derivation from a pre-existing basalt which apparently had very short time of crustal residence ($^{87}Sr/^{86}Sr$ (i) = 0.700 to 0.701). The Fundamental Gneiss is preserved only as rare relicts in the vast gneissic complex of the Indian Peninsula - the Peninsular Gneiss, and as pebbles in the conglomerates of the Archaean Dharwar sequence. Study of these relicts, shows evidence of a deformation episode (pre-DF_1) prior to the deposition of the high- and low-grade Dharwar supracrustal sequence.

The Dharwar supracrustal sequence which is older than 2600 m.y., provides evidence for early crustal stability. Layered igneous complexes, mature quartz arenites, shelf associations, deposition of persistent banded iron and manganese formations which serve as stratigraphic markers in the lower sections of the supracrustal sequence, may be cited as evidences for this early stability. The tectonic environment of sedimentation and magmatism in the supracrustal belts seems to have evolved from a rifted

continental margin, through a stable shelf to a geosynclinal type of setting as in some Proterozoic basins.

The supracrustal rocks show evidences of three phases of deformation (DF_1 to DF_3). A structural unity is evident amongst the high- and low-grade supracrustal sequences, the Peninsular Gneiss and the granulites of the Dharwar craton. This structural unity suggests the evolution of the Peninsular Gneiss and granulites synkinematically with the deformation of the supracrustals (the Dharwar sequence). There is a wide spread of ages from 3200 m.y. to 2600 m.y. in the Dharwar craton. The structural unity amongst rocks developed over such a wide range of ages implies long term stability of stress regimes in the Archaean lithosphere.

Geochemistry of the volcanic and sedimentary rocks of the Dharwar supracrustal belts show that (i) volcanic rocks of the stable depositional environment are LREE enriched; in contrast, those of the mobile zone show a wide range of REE patterns; (ii) the sediments in the lower most sections are richer in soda; potash and zirconium show increment upward in the stratigraphic sequence; (iii) the quartz pebble conglomerates in the early part of the sequence show detrital pyrite and uraninite indicating low partial pressures of oxygen in the atmosphere; (iv) the greywackes in the upper part of the sequence show evidence of derivation from a continental as well as intrabasinal volcanic source; some greywackes show REE patterns with negative europium anomaly, suggesting that, the provenance from which they were derived

had witnessed granodiritic crustal differentiation; (v) presence of stromatolites, $\delta^{13}C$ values for carbon in graphitic schists (ranging from -18 ‰ PDB to -36.50 ‰ PDB) and rarely preserved cyanobacterial filaments suggest palaeo-biological photosynthetic activity that influenced large scale deposition of limestones, banded iron and manganese formations in the Archaean of the Dharwar craton.

The foregoing observations indicate that, as compared to other Archaean terrains, the Archaean Dharwar craton appears to have more evolved characteristics.

N88-30242

LITHOSPHERIC EVOLUTION OF THE NORTHERN ARABIAN SHIELD: CHEMICAL AND ISOTOPIC EVIDENCE FROM BASALTS, XENOLITHS AND GRANITES

(M. Stein, Max-Planck Institut für Chemie, Postfach 3060, D-6500 Mainz, FRG, and Department of Geology, Hebrew University of Jerusalem, Israel)

The evolution of the upper-mantle and the lower crust - the continental lithosphere, in the area of Israel and Sinai was studied, using the chemical composition and the Nd-Sr isotopic systematics from mantle and crustal nodules, their host basalts and granites.

The basalts were extruded during the Phanerozoic over a stable platform that was consolidated in the late Pan-African orogeny (1). The following magmatic events were sampled:
(a) Late Triassic(?)-Liassic (subsurace basalts);
(b) Early Cretaceous (basalt flows in the Ramon cirque);
(c) Neogene-Quaternary (basalt flows and tuffs in the Golan and the Galilee).

The basalts are alkalic in composition. They are characterized by strong LREE enrichment ($(La/Yb)_n$ = 7-60) by "oceanic type" Nb/U of about 48, and by a positive Nb anomaly in the "spider diagram", hence they have characteristics similar to OIB. These observations indicate that the basalts are not contaminated by crustal material.

The basalts have ϵNd values of +3.5 to +6.0 and $^{87}Sr/^{86}Sr$ = 0.7029-0.7034. The average ϵNd = +4.9 indicates that the source had $^{147}Sm/^{144}Nd$ > CHUR for a considerably long time. However, the average Nd (+4.9) is lower than the value expected in the depleted mantle (MORB source type) in the last 200 Ma (around +10). These observations are explained by enrichment of the depleted peridotite source in LIL elements including Rb and Nd.

The peridotite nodules are hosted in Mesozoic basalt (from the Ramon Cirque) and Cenozoic basalts (from the Galilee and Golan). They are spinel-lherzolites and harzburgites in composition (Group I of Frey and Printz (2)) and they have metamorphic textures, hence they are accidental to the host basalts. The peridotites show negative correlations between MgO and CaO, Al_2O_3, Yb and positive correlations between MgO and Ni, Co. These relations are explained by partial melting processes in the upper-mantle. The peridotites are residues after 20-40% melting of "primitive-mantle sources". The melts in equilibrium with these residues cannot be alkali basalts, similar to the hosts; rather they might be komatiitic or picritic magmas evolved in the earlier history of the upper-mantle. The nodules yield a negative correlation between Mg/Si and Al/Si, and they plot among other peridotites nodules from different locations in the world along the "geochemical fractionation trend" defined by Jagoutz et.al.(3) Nodules MHZ-236 and MHZ-243 from Maale-Atzmaut are close to the "primitive" Mg/Si - Al/Si values. However, these peridotites (and also the "less primitive" on the geochemical trend) have higher Ca/Al ratios than the chondritic value.

Several nodules from the Ramon have anomalous high (and

different) Ca/Al ratios and they are also enriched in Fe. There
is a textural evidence for the existence of pyroxenite veins that
crystallized inside these peridotites. These might explain the
high Ca/Al ratio. Most of the peridotites that were studied have
$(LREE/HREE)_n > 1$. This observation is not predicted by their
residual nature and is explained by a later penetration of
metasomatic fluid enriched in LIL elements (component B of Frey
and Printz (2))

The peridotites are divided into two groups according to
their ϵNd-$^{87}Sr/^{86}Sr$ values:

I   MORB(DM) type, ϵNd = +10.1 to +9,8 $^{87}Sr/^{86}Sr$ = 0.7027-0.7030

II   Enriched ("MC") type, ϵNd = +4.1 to +5.8 $^{87}Sr/^{86}Sr$ =
0.7035-0.7029

The "DM" group evolved in earlier depletion events of the
primitive mantle. These events might correspond to the melting
that formed the residual peridotites mentioned above. CHUR model
age (T_{CHUR}) for the depleted nodule KH-300 is 2.2 Ga.

The "MC" nodules evolved from the depleted mantle (DM) via a
metasomatic enrichment in LIL elements including Rb and Nd.
(Similar to the B component above.) Depleted mantle model ages
for the "MC" nodules are 762-1015.

Whole-rock isochrons of the peridotites and mineral
isochrons of separate clinopyroxenes and orthopyroxenes indicate
heating events in the mantle sampled by these rocks. Several
heating episodes were detected:
I.   390 Ma (Silurian-Devonian) indicated by the Ramon nodules.
II.   190 Ma (Triassic-Jurassic) indicated by the Galilee nodules.
III.   >50 Ma (Cenozoic) indicated by the Golan nodules.
These events are followed by an uplift in the upper-crust.

Garnet pyroxenites and kaersutites hosted in
Upper-Cretaceous tuff in the Carmel are cumulates from mafic
magmas. They have metamorphic textures and different ϵNd values
than the host rocks, hence they are not cogenate to their hosts
and rather represent older cumulates. A 400 Ma Sm-Nd isochron age
indicates the crystallization age of these pyroxenites. Mineral
isochrons (based on clinopyroxene and garnet) correspond to
111 Ma and indicate a thermal event that is related to Cretaceous
magmatism. The kaersutites have Rb-Sr and Sm-Nd isochrons of 213
Ma and 189 Ma respectively. This ($\simeq$ 200 Ma) age is related to a
heating event in the lithosphere beneath the Carmel.

Mafic granulites hosted in alkali-basalts and tuffs from
Karnei-Hitin (Galilee) and Golan are samples from the lower
crust. Granites from the exposed Arabian Shield in Sinai
represent the upper crust. The granulites have metamorphic
textures, they have a gabbro-anorthosite composition and are
interpreted as being cumulates from mafic magmas. Sm-Nd isochron
ages indicate 700 and 620 Ma. These ages might be the
crystallization time from melts of the depleted mantle.

The granites studied are both calcalkaline and alkaline.
They have an Sm-Nd isochron age of 610 Ma and an initial of about
+4.0, and initial $^{87}Sr/^{86}Sr$ = 0.7028 confirming the juvenile

Stein M.

138

nature of the crust created during the Pan-African event. The low ϵNd-87/86 values indicate that the granites were produced from a basaltic precurser. Therefore the early crustal evolution in this area is accompanied by extensive basaltic magmatism.

There are two different stages in the evolution of the upper-mantle that are sampled in this area

I   Evolution of a depleted mantle from the primitive mantle.

II   Transition from depleted mantle to continental lithosphere (subcontinental lithosphere and continental crust)

The evolution of the continental lithosphere is characterized by introduction of mafic melts into the uppermost mantle; from these melts pyroxenites ,kaersutites and gabbro-anorthosites crystallized. The latter composed the lower crust. The granites which form the upper crust in the area were produced from basaltic precurser.

The magmatism and the metasomatism making the lithosphere are related to uprise of mantle diapirs in the uppermost mantle of the area. These diapirs heated the base of the lithosphere, eroded and replaced it with new hot material. It caused a domal uplift of the lithosphere (and the crust). The doming resulted in tensional stresses that in turn might develop "transport channels" for the basalts.

References
Bielski M. 1982 Ph.D. Thesis Hebrew University, Jerusalem
Frey F. and Printz M. 1978 E.P.S.L. 38, 129-176
Jagoutz E. et al. 1979 Proc. Lunar Planet Sci Conf 10th, 2031-2050

TRACE ELEMENT DIFFERENCES BETWEEN ARCHAEAN, PROTEROZOIC AND PHANEROZOIC CRUSTAL COMPONENTS - IMPLICATIONS FOR CRUSTAL GROWTH PROCESSES.

J. Tarney (Dept. Geology, U. Leicester, U.K.), L.E.A. Wyborn, J.W. Sheraton & D. Wyborn (Bureau of Mineral Resources, Canberra, Australia)

Critical to models for continental crust growth and recycling are the processes through which crustal growth takes place. In particular, it is important to know whether these processes have changed fundamentally with time in response to the Earth's thermal evolution, and whether the crustal compositions generated are compatible with crustal remobilisation, crustal recycling, or represent primary additions.

There are some significant and consistent differences in the major and trace element compositions of crustal components with time which have important implications for crustal growth processes. These will be illustrated with reference to Archaean rocks from a number of shield areas (Antarctica, Australia, India and the N. Atlantic craton), Proterozoic granitoids from Australia and elsewhere, Palaeozoic granitoids from Australia and Scotland, and Mesozoic - Recent granitoids from present continental margin belts. Secular changes are based on large I-type granitoids which, in any time period, make up at least 80% of all granitoids. Chemical differences are best outlined using multi-element diagrams where trace element compositions are normalised to estimated primordial mantle compositions: the patterns thus represent the integrated effects of the processes involved in generating the granite. Surprisingly some rather simple and consistent patterns emerge using this technique.

As a broad generalisation, granitoids tend to fall into two main groups. Either they are Sr undepleted and Y depleted, or they are Sr depleted and Y undepleted: volumetrically there are few examples of Sr and Y undepleted granites, and even fewer of Sr and Y enriched granites. These features relate to mineralogical control, either during magma generation or during development of the source. The Y and HREE depletion infers derivation from a source which has residual garnet, but not plagioclase, whereas the Sr depletion infers sources with residual plagioclase but not garnet. Interestingly there are other element co-variations. Sr-rich, Y-poor granitoids also tend to be Ba- and P-rich but Rb-, U-, Th- and K-poor and with a high K/Rb ratio. The reverse situation occurs with the Sr-depleted, Y-undepleted granitoids.

All granitoids have a significant negative anomaly for Nb and Ta on mantle-normalised diagrams: this is always larger for Y-depleted types. It implies that a mineral phase capable of sequestering these elements was stable at the time of granite generation or during development of their source region. Growth of a crustal reservoir markedly depleted in Nb and Ta relative to other incompatible elements must imply a complementary reservoir in the mantle which is correspondingly enriched in these elements. This complementary reservoir is that supplying ocean islands and alkaline magmas.

In time there are essentially four main types of I-type granitoid: (1) early Archaean tonalites and trondhjemites, (2) late Archaean to middle Proterozoic K-rich tonalites and granodiorites, (3) early Palaeozoic granodiorites, and (4) late Palaeozoic, Mesozoic and Cenozoic tonalites. A very high proportion of Archaean granitoids are of the Sr-undepleted Y-depleted type (but differ in also showing P-depletion). This type is also common, though not exclusively so, in post-400Ma cordilleran margin granitoids, where there is an undoubted link with subduction. Conversely granitoids of Proterozoic to early Palaeozoic age are dominantly of

TRACE ELEMENT DIFFERENCES: CRUSTAL COMPONENTS
Tarney, J., Wyborn, L., Sheraton, J. & Wyborn, D.

140

Sr-depleted Y-undepleted type. Calculated model source ages for the
Sr-undepleted Y-depleted granitoids is usually nearly coincident with their
emplacement age; whereas for the other type there is a significant crustal
residence time for the source of these granites, which can be up to 1000Ma
for some Australian Palaeozoic granites.

The results have a bearing on models for crustal evolution. The
steady-state model implies that the present volume of continental crust was
created in the Archaean, but that recycling has dominated ever since. This is
at variance with the distribution of incompatible elements in major I-type
batholiths with time. Elements such as K, Rb, Th and U suddenly peak in the
late Archaean and early Proterozoic, and then progressively (in global terms)
begin to decrease with time, with Sr-undepleted tonalites typical of
subduction zones becoming dominant in the later Phanerozoic. If crustal
recycling processes were dominant, the expectation would be that the
concentrations of these incompatible elements would increase with time. The
fact that specific granitoid compositions appear to dominate or be typical of
specific time intervals supports instead models of episodic crustal additions
through time.

The change to more K-rich granites in the late Archaean and particularly
the early- mid-Proterozoic has long been known, and has commonly been
interpreted as indicating a transition from essentially primary mafic-source
derived granitoids to those derived through remobilisation of this felsic
crust. This carried an assertion that a considerable proportion of the crust
was generated in the Archaean, and that a high proportion of granites from
then on were derived through melting of this continental crust. However there
are severe difficulties in deriving the relatively K-rich Proterozoic and
Palaeozoic I-type batholiths from pre-existing Archaean crust because (with
the exception of late Archaean anorogenic granites) the bulk of the Archaean
felsic crust is low in K,Rb,Th,U,Y and the HREE, whereas the vast majority of
Proterozoic and early Palaeozoic granitoids are enriched in these elements
and have very different element ratios. Moreover, in Australia, Rb-Sr and
Sm-Nd model source ages for this type of Palaeozoic and Proterozoic granite
are all Proterozoic or younger, confirming that they cannot be derived from
Archaean crust.

There are then significant differences in compositions of granitoid
crustal additions throughout geological time, with a particular type of
granitoid apparently dominating a particular time period (though not
necessarily being exclusive to it). This implies that the tectonic processes
giving rise to granite generation have changed in response to the Earth's
thermal evolution. The high rates of crustal generation in the Archaean are
most easily reconciled with melting of a mafic source under hydrous
conditions. Post-400Ma cordilleran granitoids have some similarities with,
but are subtly different from, Archaean granitoids, and reflect subduction
processes, with their source being in the mantle wedge. K-rich Proterozoic
and early Palaeozoic granitoids are not easily reconciled with subduction
processes, and probably reflect intracratonic melting: but not of
pre-existing crust. Their source could lie in the sub-cratonic lithosphere or
a mafic lower crust underplate. The reason for the sudden appearance of K-,
Rb-, Th- and U-rich crustal material in the late Archaean and Proterozoic
remains enigmatic. It represents a complement to crustal material low in
these elements which dominates the Archaean crust. Where was it stored? In
the sub-continental mantle?

CONTINENTAL GROWTH THROUGH TIME BY UNDERPLATING OF SUBDUCTED OCEANIC CRUST: EVIDENCE FROM KIMBERLITES IN SOUTH AFRICA AND THE SW PACIFIC

Lawrence A. TAYLOR and Clive R. NEAL, Dept. of Geol. Sci., Univ. of Tennessee, Knoxville, TN 37996.

In the dynamic model of plate tectonics, it is evident that crustal components are returned to the mantle by subduction. Chemical signatures of these subducted components have been identified in ocean island volcanics [1] and in island arc volcanics [e.g., 2]. Indeed, an origin involving a subducted protolith has been postulated for certain types of xenoliths in kimberlite, including diamonds [e.g., 3, 4]. Recent studies of eclogite xenoliths in kimberlites from southern Africa [5-7] and megacrysts from the Malaitan alnoite, Solomon Islands [8], indicate that lithospheric underplating by subducted oceanic crust has occurred in these two contrasting areas. We report the results of new eclogite studies from the Bellsbank kimberlite, South Africa, and isotopic data from the Malaitan alnoite megacryst suite. This forms the basis for discerning the role of lithospheric underplating in the growth of cratons and in the evolution of mantle-derived magma.

Eclogite Xenoliths in Kimberlite: Our investigations have centered on eclogite xenoliths in kimberlites from southern Afica. This area is dominated by the Kaapvaal and Rhodesian cratons which contain Archean crustal rocks up to 3.5AE [9]. Based upon our preliminary study last year [5-6], **three groups of eclogite xenoliths were identified** from four kimberlites (in Angola, Namibia, Lesotho, and So. Africa). These groups represent "normal" mantle eclogites, but also xenoliths which have retained a signature of **an oceanic crustal protolith.** More recently, we have discovered all three types within one kimberlite, that from Bellsbank, So. Africa [10]. These eclogite groups can be roughly correlated with the broad chemical classification of eclogites by Coleman et al. [11]. The three eclogite groups each have distinctive characteristics:

Group A - MgO-rich gts ($\cong 20\%$), low Na_2O in cpx (1.5-3.0%), moderate $\delta^{18}O$ (+5.0 to +5.8), and low $^{87}Sr/^{86}Sr$ (0.7043-0.7073), and variable ϵNd (-10 to +15); olivine and opx may be present. **These eclogites represent cumulate mantle dike rocks.**

Group B - FeO-rich gts ($\cong 20\%$), high Na_2O in cpx (3-5.5%), low ^{18}O (+2.8 to +4.1), high $^{87}Sr/^{86}Sr$ (0.7094-0.7100), and extremely high ϵNd (+120 to +250). **These features are consistent with high-temperature hydrothermal alteration of spilitized oceanic basalt.**

Group C - CaO-rich gts (14-19%), very high Na_2O in cpx (7-9%), low $\delta^{18}O$ (3.1-4.9), low $^{87}Sr/^{86}Sr$ (0.7033-0.7042), and variable ϵNd (-15 to +24) and ^{18}O (+3.1 to +4.9); plagioclase and K-feldspar may be present, and garnet and cpx have positive Eu anomalies. **These characteristics represent the recrystallization of a cumulate gabbro protolith, such as occurs in the lower portions of an ophiolite sequence.** Two of these eclogite groups (B and C) are similar to those described from the Roberts Victor kimberlite by Jagoutz et al. [12], and MacGregor and Manton [7].

A Sm-Nd "isochron" age of 2.4 AE has been determined for these eclogites. The presence of such "crustal" eclogites in several southern Africa kimberlites indicates that **underplating by subducted Archean oceanic crust provided an important contribution to the growth of the Kaapvaal craton (Fig. 1).**

FIG.1
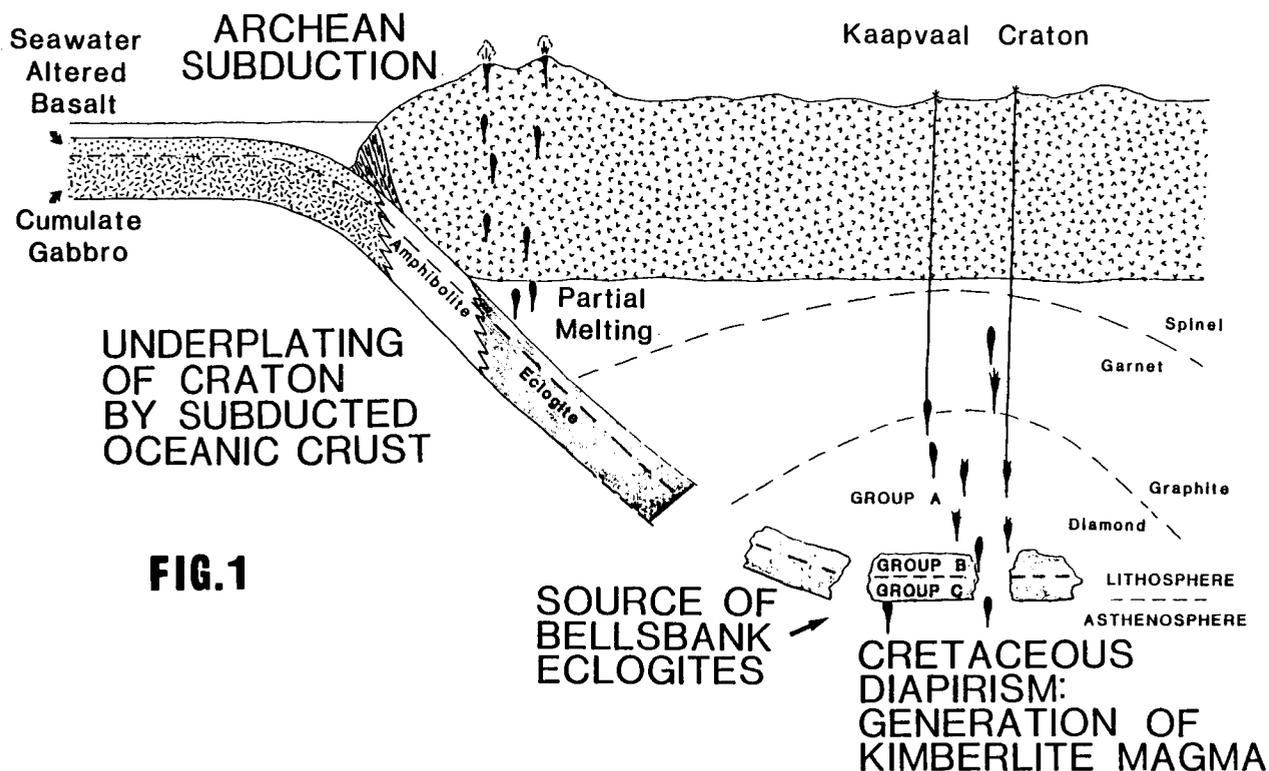
Megacrysts from the Malaitan alnoite, Solomon Islands: The Solomon Islands chain marks the boundary between the Indo-Australian and Pacific plates. The area is tectonically dominated by the Ontong Java Plateau (OJP), a vastly overthickened portion of oceanic crust (approximately 40km thick), which is now juxtaposed at the margin of the Pacific plate. The island of Malaita is the obducted leading edge of the OJP [13]. Mantle-derived alnoite magmas, containing many xenoliths, were explosively intruded 34Ma ago into the OJP and may be sampled on Malaita. Conspicuous among these is a large megacryst suite comprised of garnet, clinopyroxene, ilmenite, clinopyroxene-ilmenite intergrowths, phlogopite, and minor bronzite and zircon.

The clinopyroxene megacrysts have been studied as they form a large, continuous compositional array from augite to subcalcic diopsides with intermediate cpx-ilmenite intergrowths (see Neal, this volume). Sr and Nd isotopic ratios become more radiogeneic from the augites, to the subcalcic diopsides, to the host alnoite. $\delta^{18}O$ generally decreases from the augites to the subcalcic diopsides.

The isotopic disparity between the augites, subcalcic diopsides, and the alnoite has been explained by an AFC process
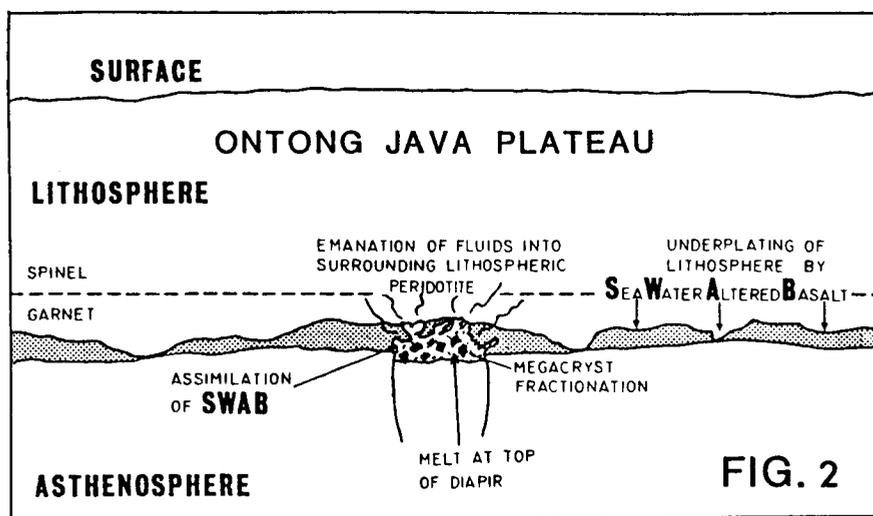
(Neal, this volume). The parental, proto-alnoite magma assimilates a subducted component of seawater altered oceanic crust while fractionating the megacryst suite. This model requires that the augite megacrysts crystallized first, having the least radiogenic Sr and Nd isotope ratios. The decrease in $\delta^{18}O$ between the augites and the subcalcic diopsides is in accordance with this model. Geophysical data suggests the OJP is **underplated by subducted derivatives of seawater altered oceanic crust** (high $^{87}Sr/^{86}Sr$ and ε Nd, low $\delta^{18}O$) which is evidenced by the presence of a basal high velocity (7.6 km/sec) layer [14] (see Fig. 2). We conclude that assimilation of this SWAB layer by the mantle-derived, diapiric proto-alnoite magma (after impingement on the rigid lithosphere) results in the isotopic evolution witnessed in the megacrysts and the resulting alnoite magma.

SURFACE

ONTONG JAVA PLATEAU

LITHOSPHERE

SPINEL

EMANATION OF FLUIDS INTO SURROUNDING LITHOSPHERIC PERIDOTITE

UNDERPLATING OF LITHOSPHERE BY SeaWaterAlteredBasalt

GARNET

ASSIMILATION OF **SWAB**

MEGACRYST FRACTIONATION

MELT AT TOP OF DIAPIR

ASTHENOSPHERE   FIG. 2

CONCLUSIONS:

The results from eclogites and megacrysts indicate that lithospheric underplating is widespread, both in geographical extent and time (the OJP is only 180my old). Two important conclusions from this work are that underplating by subducted oceanic crust has: **1) contributed significantly to the growth of the continental crust; 2) been intimately involved in the evolution of mantle-derived magmas.**

REFERENCES: 1) White & Hofmann, 1982, Nature 296, 821-825; 2) Davidson, 1985, EPSL 72, 163-174; 3) Helmstaedt & Doig, 1975, Phys. Chem. Earth 9, 95-112; 4) Schulze, 1986, Nature 319, 483-485; 5) Shervais, et al., 1986, 4th Int'l Kimberlite Conf., 326-328; 6) Shervais, et al., 1987, GSA Bull, in press; 7) MacGregor & Manton, 1986, JGR 91, 14063-14079; 8) Neal, 1987, this volume; 9) Boyd, & Gurney, 1986, Science 232, 472-477; 10) Pyle et al, 1987, LPSC XVIII, 806-807; 11) Coleman, et al., 1965, GSA Bull. 76, 483-508; 12) Jagoutz et al., 1985, The Early Earth, LPI Tech. Report 85-01, 40-41; 13) Coleman & Kroenke, 1981, Geo-Marine Lett. 1, 129-134; 14) Furomoto, 1976, Tectonophysics 34, 71-90.

THE GROWTH OF THE CONTINENTAL CRUST:
CONSTRAINTS FROM RADIOGENIC ISOTOPE GEOCHEMISTRY.

Paul N. Taylor : University of Oxford, Dept. of Earth Sciences,
Parks Road, Oxford OX1 3PR, U.K.

Most models for evolution of continental crust are expressed in the form of a diagram illustrating the net cumulative crustal mass — normalised relative to the present crustal mass — as a function of time. Thus, geochronological data inevitably play a major role in either constructing or testing crustal growth models. For all models, determining the start-time for effective crustal accretion is of vital importance. To this end, the continuing search for, and reliable characterization of, the most ancient crustal rock-units (1) remains a worthy enterprise.

The ca. 3.8 Ga Isua metavolcanics and metasediments probably still hold the best claim to represent the oldest preserved crustal rock-units yet found, and they thus provide the best age constraint presently available for the start of effective crustal accumulation. However, the discovery of very ancient (up to ca. 4.1 to 4.2 Ga) detrital zircons by Froude et al. (2) in ca. 3.35 Ga paragneisses in the Mount Narryer area of the Yilgarn Craton, Western Australia, suggests that production of some form of crust was under way at least 300 Ma before the Isua rocks were deposited, though apparently no rock record of such early crust production has survived. Further indications of an early start to crust production are seen in the Nd isotope geochemistry of early Archaean mantle-derived volcanics, including Isua, most of which show markedly positive initial $\varepsilon(Nd)$ values, evidence of extraction from source materials with a long-term history of LREE-depletion relative to a chondritic source (3,4). If the bulk earth has chondritic REE chemistry, then some complementary long-lived LREE-enriched reservoir must have existed at the time when these ancient volcanics were erupted, consistent with an early start to crust production.

Another important role for geochronology and radiogenic isotope geochemistry in constraining models of crustal evolution is to assess the status of major geological events (and provinces) as periods (and sites) either of new crust generation or of reworking of earlier formed continental crust (e.g. Refs. (5) & (6)). For this application, Sm-Nd model ages (T-CHUR, T-DM or T-CR) for LREE-enriched rock-units can be plotted in a "concordia" diagram against the corresponding Rb-Sr or Pb/Pb whole-rock isochron age data, or U-Pb zircon age data. Concordant results typify events in which new crust generation has been dominant, whereas marked discordance, with Sm-Nd model ages > Rb-Sr, Pb/Pb or U-Pb ages, is characteristic of rock-units which have been formed by reworking of ancient crustal material. Initial 87-Sr/86-Sr ratios, initial $\varepsilon(Nd)$ values, and model μ values (238-U/204-Pb ratios) provide further test criteria for this type of assessment (5). With such criteria, it may be possible to determine the level in the

Isotopic Constraints on Crustal Growth.
Paul N. Taylor.

crust at which crustal reworking has occurred (7). The distinction between new crustal additions and ancient crustal reworking is important because reworking represents only the internal reorganization of crustal material and has no effect on crustal growth rates. For constraining growth models, crust formation ages are the essential data, and in areas of crustal reworking, Sm-Nd model ages (T-DM or T-CR) will usually constitute the best estimates of these. Wrong use of crustal reworking ages as substitutes for crust formation ages tends to bias crustal growth models by under-representation of early crust-formation events.

From age characterization of major geological provinces, using the criteria outlined above, the mass (or volume) of crust surviving to the present day should be determinable as a function of crust formation age. Thence the average age of the crust can be established. If no recycling of crust into the mantle has taken place over the course of geological history, then the mass-age relationship would also constitute a full description of the crustal growth history. However, Armstrong (8) has claimed that much of the isotopic evidence bearing on the growth of the continental crust is incapable of resolving models of uniform crustal growth with minimal recycling from those in which, after an initial growth stage, rates of crust production come to be matched by rates of crustal recycling, with consequent attainment of no-growth, steady-state conditions. Thus, according to Armstrong, in the debate on growth versus no-growth, the evidence provided by radiogenic isotope geochemistry is inconclusive.

More recent developments, however, appear to set severe limitations on recycling of crust, at least by the process of sediment subduction. White (9) has argued that the good correlation of $\epsilon(Hf)$ and $\epsilon(Nd)$ in oceanic basalts could not be sustained against significant feed-back of marine sediment into the upper mantle, because of incoherent fractionation of Lu/Hf relative to Sm/Nd and related poor correlation of $\epsilon(Hf)$ and $\epsilon(Nd)$ in sediments. In addition, trace element arguments also severely restrict the amount of sediment subduction allowable: Pb/Ce and Cs/Rb ratios are more-or-less constant in oceanic basalts, and very different from the ratios in sediments. Recycling of sediment to the mantle would degrade the constancy of these ratios (9). White concludes that sediment can be subducted to depths of ca. 100 km beneath island arcs, but is not returned to the suboceanic mantle. Galer & O'Nions (10) have also argued against recycling of crustal Pb (and other highly incompatible elements) back into the mantle. Their recognition of the short residence times of Th, U & Pb in the upper mantle requires replenishment of these elements from another reservoir, but the upper continental crust is eliminated as a suitable source because of its high 207-Pb/204-Pb ratios relative to MORB. Instead, entrainment of lower mantle is the preferred replenishment mechanism.

In modelling crustal growth without recycling, valuable constraints on growth rate variations through time can be provided if variations in the average age of the continental

Isotopic Constraints on Crustal Growth.
Paul N. Taylor.

146

crust can be monitored through geological history. Goldstein et al. (11) have shown that a uniform rate of crustal growth should result in a present average age for the crust corresponding to half the start-time for crust production (i.e. half the age of the earliest-formed surviving crust). If the early history of the earth was characterized by greater rates of crust production than today (as a result of the exponential decline in radiogenic heat production though time?), then a somewhat greater mean age of the present crust should be expected.

Further to these theoretical considerations, Goldstein et al. (11) have addressed the question of the average age of the exposed continental crust by determining Sm-Nd crustal residence model ages (T-CR) for fine-grained sediment loads of many of the world's major rivers, thus sampling a very large percentage of the surface area of the continental crust. Sm-Nd isotopic compositions of these natural crustal composites show a remarkable degree of uniformity, and indicate an average age for the exposed crust of 1.70 +/- 0.35 Ga.

If this mean age is representative of the whole continental crust, including material well below the present erosion level, then within the limits imposed by uncertainties on the mean age determination, there is no clear proof of any significant departure from a uniform growth model with a start-time of ca. 3.8 Ga. In such a model, no more than ca. one third of the mass of the present crust could have been formed by the end of the Archaean: larger proportions, as have often been advocated in the past, can only be accommodated if the average age of deep crust is substantially greater than that of the exposed crust, or if significant recycling of the crust into the mantle is admitted.

References.

(1) Moorbath S, Taylor PN & Jones NW.(1986) Chem.Geol.57,63-86
(2) Froude DO, Ireland TR, Kinny PD, Williams IS, Compston W, Williams IR & Myers JS (1983) Nature 304,616-618
(3) Hamilton PJ, O'Nions RK, Bridgwater D & Nutman A. (1983) Earth Planet.Sci.Lett. 62,263-272
(4) Huang Xuan, Bi Ziwei & DePaolo DJ.(1986) Geochim.Cosmochim. Acta 50,625-631
(5) Taylor PN, Jones NW & Moorbath S.(1984) Phil.Trans.R.Soc. Lond. A310,605-625
(6) Patchett PJ & Arndt NT. (1986) Earth Planet.Sci.Lett. 78, 329-338.
(7) Moorbath S & Taylor PN. (1986) Spec.Publ.Geol.Soc.Lond. 24,211-220
(8) Armstrong RL. (1981) Phil.Trans.R.Soc.Lond. A301,443-472
(9) White WM. (1986) Terra Cognita 6,136
(10) Galer SJG & O'Nions RK (1985) Nature 316,778-782
(11) Goldstein SL, O'Nions RK & Hamilton PJ. (1984) Earth Planet.Sci.Lett. 70,221-236

CRUSTAL GROWTH IN THE ARCHEAN: THE GEOCHEMICAL EVIDENCE

Stuart Ross Taylor,
Research School of Earth Sciences, Australian National University, Canberra, Australia

DIFFERENCES BETWEEN ARCHEAN AND POST-ARCHEAN CRUSTS

There is substantial evidence for a major period of growth and stabilisation of the continental crust toward the end of the Archean [1]. The REE patterns and Sm-Nd isotopic systematics of widespread sampling (e.g. loess deposits [1]) record little change in upper crustal composition since that time. This upper crustal event is ascribed to intra-crustal melting producing K-rich granites with a signature of depletion in Eu.

Throughout the Archean, limited and isolated areas also underwent intra-crustal melting, with the formation of upper crustal rocks depleted in Eu, followed by erosion and deposition of sediments with similar geochemical signatures [2]. This process is similar to that which produced the upper crust in late and post-Archean times. Subsequent burial and metamorphism led to the production of high-grade terrains. These mini-cratonic environments are apparently only preserved in high-grade terrains and the reason for this remains conjectural.

Archean terrains comprise both greenstone belts of rather low metamorphic grade, and the high-grade terrains. The sedimentary rocks preserved in greenstone belts reveal a distinctly different REE pattern to that observed in post-Archean sedimentary sequences. The patterns are diverse, ranging from flat MORB-like to steep LREE enriched patterns, typical of Na-rich plutonic and felsic rocks. These patterns are ascribed to local provenance effects and indicate less thorough mixing or more isolated sources and hence more first-cycle sediments than is typical of post-Archean sedimentary sequences.

The majority of greenstone belt sediments display patterns intermediate between these two extremes, approximating to equal mixtures of basic and felsic parental igneous rocks. Eu anomalies are rare. This evidence is consistent with an Archean upper crust dominated by the bimodal basic-felsic suite. Although most of the studied rocks are indeed located within greenstone belts, their petrogaphy and chemistry indicate that they sample a wider provenance. A measurable depletion in Eu would result from the presence of more than 10% of typical K-rich granites, leading to the conclusion that such granites which are typical of the present upper crust were rare in the Archean upper crust being sampled by these sediments.

A CRUSTAL GROWTH MODEL

These observations have led to a model of crustal gowth in which the Archean upper continental crust is dominated by the bimodal suite, whereas the post-Archean upper crust is dominated by K-rich granites, granodiorites and associated volcanic rocks. Major intra-crustal melting events, which were preceded on times of the order of 10^8 years by massive addition of material from the mantle occur on this model over a period of 500-700 million years toward the end of the Archean. These events are considered to represent the major episode in the growth of the continental crust which was about 75% complete by about 2500 M.Y. ago.

ARCHEAN HIGH-GRADE TERRAINS

Some (eg. Kapuskasing) represent the high grade equivalent of greenstone belts, and the REE patterns have the same characteristics [2]. There is no evidence that the REE patterns are altered by high-grade metamorphism up to granulite facies unless partial melting occurs. Other high-grade terrains (eg. Limpopo) contain two distinct suites of meta-sediments [2]. One comprises the metamorphosed equivalent of greenstone belt meta-sediments, with REE patterns showing no Eu depletion. The other suite has REE patterns which are indistinguishable from those of typical Post-Archean sediments (eg. PAAS), and must be derived mainly from the weathering and erosion of K-rich granite terrains. Major element data (eg. K/Na ratios) agree with this interpretation. The granitic source area must be small scale since the distribution of meta-sediments with PAAS characteristics is localised and adjacent regions have greenstone belt REE signatures. In contrast, although on the basis of very limited sampling, meta-sediments from the Western Gneiss Terrain do not have greenstone-belt type REE patterns but have only PAAS type patterns. The extent of this type of metasediment is not known. Similar patterns have been reported from India [3],Montana-Wyoming [4,5] and Greenland [6,7,8], but in all these regions there is considerable variability, indicating local provenance and restricted outcrop of K-rich granitic source rocks.

TECTONIC SIGNIFICANCE

The early Archean (3.6-3.2 Ae) metasediments in the Limpopo and Western Gneiss Terrains were deposited in a shallow shelf environment at the margins of a small craton (Eriksson, pers comm.). REE patterns indicate that a variety of provenances was available ranging from the typical Archean bimodal suite to K-rich granitic rocks with Eu depletion. Apparently isolated sedimentary basins or environments have preserved this evidence on scales of a few hundred square km. The available data for the Archean suggests that the PAAS type sediments are uncommon in Archean sequences. Such occurrences are accordingly interpreted here as being derived from mini-cratons of limited areal extent, relative to the dominant bimodal Archean igneous suites.

The Na-rich granites of these suites, with their steep LREE enriched-HREE depleted patterns, indicative of equilibration with a garnet-containing residue, and thus of probable mantle derivation, contrast with the K-rich granites, with flatter Eu-depleted REE patterns indicative of intra-crustal melting. Apparently only local segments of the Archean crust formed thick (>30 km) crust which underwent intracrustal melting to produce K-rich granites. Following erosion and sedimentation to produce the sediments, the terrains then were metamorphosed to granulite grade, and subsequently uplifted to provide the present exposures.

Only meta-sediments from the high-grade terrains contain the evidence of the previous existence of the mini-cratons. These do not appear to have contributed sediments with PAAS REE signatures to the greensone belts. This contrasts with present-day sedimentary environmements where only sediments deposited in fore-arc basins of wholly intra-oceanic island arcs (eg. Marianas, [9]; Devonian Baldwin Formation, [10]) fail to show any contribution from a typical upper crustal PAAS REE pattern

C - 3

ARCHEAN CRUSTAL GROWTH

The evidence suggests that the growth of fully stabilised continental crust prior to about 3.0 Ae took place on a small scale. The dominant igneous activity was the production of basaltic and Na-rich felsic rocks of the bimodal suite. In restricted regions thicker crust enabled intra-crustal melting to produce K-rich Eu-depleted granite, followed by the deposition of sediments in shallow basins. Subsequent deep burial and granulite-grade metamorphism produced the presently observed high-grade terrains.

What was the original extent of these cratonic areas? At present, they comprise perhaps 50% of Archean exposed areas. The meta-sediments which they contain were deposited on stable shelves (Eriksson, pers. comm.) Preferential preservation of such cratonic areas is likely. Greenstone belts, in contrast, were deposited in tectonically active basins (modern analogues include island-arc, back-arc and foredeep basins [11,12,13]). Recycling of this material by intra-crustal processes is likely to be much more rapid than that of the cratonic areas [14, 15] The model which emerges is one of small scattered Archean cratonic areas, suggestive of small step-like increments of crustal growth, culminating in a massive pulse of continental growth in the late Archean.

REFERENCES

1. Taylor, S.R. And McLennan, S.M. 1985. The Continental Crust: Its Composition and Evolution. Blackwell, Oxford
2. Taylor, S.R., Rudnick, R.L.,McLennan, S.M. and Eriksson, K.A. 1986. Geochim. Cosmochim. Acta, 50, 2267.
3. Naqvi, S.M., Condie, K.C. and Philip, A. 1983. Precamb. Res.,22, 125.
4. Mueller, D.A., Wooden, J.L., and Bowes, D.R. 1982. Rev. Brasil Geosc., 12,215.
5. Gibbs, A.K., Montgomery, C.W., O'Day, P.A. and Erslev, E.A. 1986. Geochim. Cosmochim. Acta, 50, 2125.
6. Boak, J.L.,Dymek, R.F. and Gromet, L.P. 1982. Lunar Planet. Sci., XIII, 51.
7. Dymek, R.F., Weed, R. and Gromet, L.P. 1983, Rapp. Gronlands Geol. Unders., 112, 53.
8. McLennan, S.M.,Taylor, S.R. and Eriksson, K.A. 1984. Geochim. Cosmochim. Acta, 48, 1.
9. McLennan, S.M., McCulloch, M.T., Taylor, S.R. and Maynard, J.B., 1985. EOS, 66, 1136 (Abstr.).
10. Nance, W.B. and Taylor, S.R. 1977. Geochim. Cosmochim. Acta, 41, 225.
11. Dimroth, E., Imreh, L., Rocheleau, M. and Goulet, N. (1982) Can. J. Earth Sci., 19, 1729.
12. Tarney, J., Dalziel, I.W.D. and De Wit, M.J. 1976, in The Early History of the Earth (Ed. B.F. Windley),Wiley, London, 131.
13. Jackson, M.P.A., Eriksson,K.A. and Harris, C.W. 1987. Tectonophys. in press.
14. Veizer, J. 1984. Proc. 27th. Inter. Geol. Congr.,11,325.
15. Veizer, J. and Jansen, S.L. 1985. Jour. Geol.,93, 625.

N88-30247 A 883723 ?

GROWTH OF PLANETARY CRUSTS

STUART ROSS TAYLOR
Research School of Earth Sciences, Australian National University

The solid planets and satellites mostly have crusts which differ markedly in composition both from their interior, and from primordial solar nebula compositions. Although this was long ago understood for the Earth, the lunar samples focussed attention on wider aspects of the problem. Pre-Apollo thinking led to the view that the moon was a primitive undifferentiated object, because of its low density [1] . Although this opinion was not universally held (e.g. the mare surfaces were correctly identified as lava flows by Baldwin [2]) the surprising thing was that the samples both from the maria and the lunar highland crust were very highly differentiated, compared to estimates of primitive solar nebula values, established from the resemblance between the solar photospheric and CI abundances for the non-gaseous elements. Indeed, the highland crustal abundances were so enriched in refractory elements that models invoking the late plastering on of a refractory-rich layer appeared [3], although these were quickly superseded by magma ocean models [4]. Further exploration revealed that Mercury, Venus, Mars and many of the larger satellites have surface compositions which differ from any reasonable estimate of their bulk composition: this serves as a working definition of a planetary crust.

Such crusts may arise in two basically different ways. Firstly, they may form as a result of planetary differentiation consequent upon melting during or shortly following accretion (e.g. lunar highland crust). These may be termed "primary" crusts. "Secondary" crusts arise later in planetary history as a result of partial melting in planetary interiors. These are typically composed of basalt, the primary melt from silicate mantles (see [5] for an extensive review). Examples include the lunar maria, the terrestrial oceanic crust, the northern hemisphere of Mars, including the great volcanoes, and probably the Venusian crust. "Tertiary" crusts may arise through further melting and differentiation of the extruded material. The continental crust of the Earth may be the sole example. Growth of primary crusts occurs concomitantly with or shortly following accretion, and is completed on short time scales (10^8 year), while growth of secondary and tertiary crusts may extend over the lifetime of the planet.

Primary Crusts

The lunar highland crust is the best studied example of a primary crust. It is 60-100 km. thick on a body whose radius is only 1738 km, thus comprising 10-12% of planetary volume. It formed by flotation of feldspar in a completely dry magma ocean. Since plagioclase will sink in a magma containing >0.1% water [6] such an early feldspathic crust is not expected to form on the Earth, for example. The date of 4440 m.y. for the ferroan anorthosite 60025 [7] provides an initial date for plagioclase crystallisation and crustal formation. Complete solidification of the magma ocean occurred by about 4350 m.y. [8], so that about 90 m.y. was apparently required for this process. Planetary accretion models based on the planetesimal hypothesis require up to 100 m.y. from t_0 (4560 m.y.) to complete planetary assembly. Lunar origin via the giant impactor hypothesis involves collision with the next largest body in the hierarchy (0.1-0.2 earth mass) and requires prior core formation in the impactor. Thus it is unlikely to occur before about 4460 m.y., only 20 m.y. before the best estimate of the age of 60025. Even allowing for the large uncertainties in all these ages, it is apparent that melting of much of the moon, and development of the thick lunar highland crust proceeded hard on the heels of the formation of the satellite. Such growth rates are essentially instantaneous on a geological time scale.

Mercury

The limited information from this planet indicates a crust similar in spectral reflectance properties to the lunar highlands [9]. The presence of a sodium cloud around Mercury is perhaps consistent with a plagioclase rich crust, in which case the crust is very likely primary due to initial planetary melting. A preferred scenario to account for the high iron-silicate ratio of Mercury is that a large portion of the silicate mantle was lost during a collision with an object about 1/6 mercurian mass [10]. Such an event is likely to have both depleted the planet in volatiles (e.g. H_2O) and triggered mantle-wide melting, so that a lunar-like crust is conceivable. The observable crust is heavily cratered and must be older than 4000 m.y. by analogy with the dated lunar highland crust. The presence of lobate fault scarps of about the same age, indicative of a slight contraction (2 km) in radius, is fatal to expanding Earth hypotheses [11]

Icy Crusts

Most of the satellites of the outer planets have low densities, consistent with rock-ice mixtures. Some of the larger bodies (e.g. Europa, Callisto) appear to have water ice surfaces. The first example appears to be continually resurfaced, possibly by tidal heating from Jupiter, while Callisto has the most heavily cratered surface in the solar system. In both bodies, melting of water in the interior, probably due to collisional heating, has led to formation of an ice surface, overlying a rock-ice interior. The cratering record on Callisto indicates that these processes occur very early, and so such crusts are primary.

Early intense cratering

The formation of primary crusts proceeds in a turbulent environment due to the effects of the intense bombardment from the continuing sweep-up of planetesimals. The evidence of the ancient battered surfaces on many planets and satellites, as well as the date of the late Imbrium collision on the lunar highland crust of 3850 m.y. indicates that the bombardment continued for several hundred million years. Accordingly, primary crusts form in the teeth of this barrage, accounting for much of the complexity of the lunar highland crust. The observable surface of the lunar highland crust represents a saturation population of craters and basins and is probably no older than 4100-4200 m.y. Estimates of the flux of objects which struck the moon between 4400 and 3850 m.y. by Wilhelms [12] include about 80 basin forming events (dia.>300km) and over 10,000 craters with diameters in the range 30-300 km. It is estimated that over 200 ringed basins, with diameters >300 km formed on the Earth in the same interval [13] which probably explains the absence of identifiable rock units older than about 3800 m.y.

Mars

North of a boundary inclined at about 28° to the equator, the Martian surface consists of volcanic plains and the large volcanoes, all probably basaltic [14]. The southern hemisphere of Mars is broadly comprised of an Ancient Cratered Terrain, older than about 4000 m.y. based both on crater counting, and from the lunar analogy. Its composition is unknown, but it seems unlikely to be acidic or very different in composition from the basaltic plains. The rationale for this is that the Viking Lander XRF data at the two sites 4000 km apart were both similar, and basaltic in composition. The fine material analysed presumably represents a planetary-wide dust average (analogous to loess)[14]. The SNC meteorites likewise are derived from a basic crust [15].

Venus

The Venusian crust also appears to be dominated by basaltic lavas and the presence of extensive areas of more fractionated rocks is conjectural. The Venera 13 and 14 XRF data [16] resemble terrestrial alkali basalt and MORB respectively. The high (4%) K_2O value in the Venera 13 analysis shows that the high K (4%), U (2.2 ppm) and Th (6.5 ppm) values from Venera 8 gamma-ray data [17] do not indicate the presence of granite. The high standing regions of Aphrodite Terra and Ishtar Terra are due to tectonic rather than compositional controls on this interpretation.

Ganymede

Ganymede has two distinct crustal types, An ancient icy cratered terrain, representing a primary crust is fractured and intruded by younger grooved terrain, also of water ice, derived by later internal melting. Some minor planetary expansion has accompanied this event, due probably to a density decrease resulting from melting of high density ice in the interior [18].

Tertiary Crusts

The continental crust of the Earth, is the only known example at present [19]. It presumably owes its existence to the presence of extensive liquid water at the surface [20]. The difficulties in producing a tertiary crust are shown by its small mass. Four billion years of growth has resulted in a continental crust less than 0.5% of Earth mass. In contrast, the secondary oceanic crust is produced at a much faster rate, producing a 5 km thick basaltic crust covering most of the planet on present timescales of 200 m.y. In contrast to both these examples the primary lunar highland crust, which comprises 10-12% of the planet, was produced in about 100 m.y.

References

[1] Urey, H.C. 1959. *Jour. Geophys. Res.* **64**, 1721

[2] Baldwin, R. B. 1949. *The Face of the Moon,* Chicago Univ. Press.

[3] Gast, P.W. 1972. *Moon,* 5, 121.

[4] Taylor, S. R. and Jakes, P. 1984. *Proc. Lunar Sci. Conf.* 5,1287

[5] *Basaltic Volcanism on the Terrestrial Planets* , Lunar Planet. Inst., Houston, Texas, 1286 pp.

[6] Walker, D. and Hays, J.F. 1977. Geology, 5, 425.

[7] Lugmair, G.W. 1987. *Lunar Planet. Sci.* XVIII, 584.

[8] Meyer, C. et al. 1985. *Lunar Planet. Sci.* *XVI*, 557.

[9] Abstracts, *Conference on Mercury,* Tucson, Ariz., 1986.

[10] Cameron, A.G.W. and Benz, W. 1987. *Lunar Planet. Sci.,*XVIII, 151

[11] Taylor, S.R. 1983. *In Expanding Earth Symp.* (Univ. Tasmania),343.

[12] Wilhelms, D.E. 1985. *Lunar Planet. Sci. XVI,* 904.

[13] Grieve, R.A.F. amd Parmentier, E.M. 1985. *L.P.I. Tech. Report,* **85-01**, 23

[14] Carr, M.H. 1981. *The Surface of Mars,* Yale Univ. Press, 232 pp.

[15] McSween, H.Y. 1985. *Rev. Geophys.,23*, 391.

[16] Surkov, Y.A. 1984. *Jour. Geophys. Res.* **89**, B393.

[17 Surkov, Y.A. 1977. *Proc. Lunar Sci. Conf. 8,* 2665.

[18] Squires, S.W. 1980. *Geophys. Res. Lett.,7*, 593.

[19] Taylor, S.R. and McLennan, S.M. 1985. *The Continental Crust: Its Composition and Evolution.* Blackwells, Oxford, 312 pp.

[20] Campbell, I.H and Taylor, S.R. 1983. *Geophys. Res. Lett.,*10,1061.

GRAVITY ANOMALIES, PLATE TECTONICS AND THE LATERAL GROWTH OF
PRECAMBRIAN NORTH AMERICA; M.D. Thomas[1], R.A.F. Grieve[1] and
V.L. Sharpton[2]: 1, Geological Survey of Canada, 1 Observatory Crescent,
Ottawa, Ontario K1A 0Y3, Canada; 2, Lunar and Planetary Institute, 3303 NASA
Road 1, Houston, Texas 77058-4399, U.S.A.

The widespread gravity coverage of North America provides a picture of
the gross structural fabric of the continent via the trends of gravity
anomalies. The structural picture so obtained reveals a mosaic of gravity
trend domains, many of which correlate closely with structural provinces and
orogenic terranes. The gravity trend map, interpreted in the light of
plate-tectonic theory, thus provides a new perspective for examining the
mode of assembly and growth of North America. Suture zones,
palaeosubduction directions and, perhaps, contrasting tectonic histories may
be identified using gravity patterns.

Kay (1) in his classic dissertation on North American geosynclines
credits the idea of the "growth of continents" to Stille (2), who proposed
that continents grew from small cratons by orogenies consolidating marginal
geosynclines. Wilson (3) recognized that such peripheral growth had
commenced as early as the Archean. He suggested that an area coinciding
with Hudson Bay formed the core of the North American continent, around
which Precambrian and later geosynclines had been piled. Dewey (4)
explained the development of the Phanerozoic Appalachian-Caledonian orogen
using plate tectonics, viewing geosynclinal development in a new light.
This and similar studies in the late sixties and early seventies
revolutionized tectonic thinking. The application of plate tectonics in
studies of Phanerozoic orogens is now virtually universally accepted, as is
the idea of outward growth of Phanerozoic orogens in North America. The
peripheral Cordilleran and Appalachian orogens have both been regarded as
having grown by successive accretion of terranes through convergent plate
tectonics culminating in collision (e.g. 5 and 6, respectively). For the
Precambrian, initial acceptance of plate tectonics was less enthusiastic,
but recently plate-tectonic models have been widely used to explain the
development of Proterozoic orogenic terranes within the Canadian Shield
(7,8,9), at the margin of the Wyoming Province (10) and in buried regions of
the craton in the midcontinent area (11,12,13). Plate tectonics has been
applied also to explain the development of Archean granite-greenstone and
metasedimentary belts (14,15) and even individual Archean greenstone belts
(16).

In the Canadian Shield, geophysical studies, notably paleomagnetic
studies (17) and analysis of gravity anomalies (18,19,20), have played a
major role in fostering the acceptance of plate tectonics as a viable
mechanism to account for the tectonic history of Precambrian terranes. A
characteristic gravity signature, in the form of a paired negative-positive
anomaly, has been noted at several structural boundaries proposed to
represent collisional sutures (21). Its characteristic shape in profile and
consistent relationships to several large-scale geological features at some
boundaries have established it as a strong indicator of collision in its own
right. Thus, deeply eroded structural boundaries with little diagnostic
geological evidence of collision have been proposed as sutures largely on
the strength of their gravity signature. For example, the Thelon Front,
marking the boundary between the Slave and Churchill Provinces, is one such
suture identified using this approach, and this interpretation has been

upheld by a number of recent geological studies (e.g. 22).

The gravity signature has a consistent "polarity" with respect to the relative ages of collided terranes. The negative anomaly is positioned always within the older terrane or straddling the boundary with the younger terrane and the positive anomaly occurs within the younger terrane. This polarity pattern is compatible with a continental collision-basement reactivation model (23). The plate overriding the subduction zone is reactivated by crustal thickening, intrusion of granites at shallow levels, granulite facies metamorphism and the generation of anorthosite at depth. Attendant isotopic updating produces younger ages in the crust of the overriding plate. Tectonic and isostatic uplift of denser granulitic crust along the margin of the younger plate is believed to lead to development of the positive anomaly, and thickened/depressed crust in the older plate produces the negative anomaly.

An additional characteristic of gravity anomalies associated with Precambrian suture zones, noted in Australia (24) and more recently in the Canadian Shield (25), are the discordant trends on opposite sides of structural boundaries (=sutures). Trends parallel to the structural boundary invariably characterize the younger terrane and oblique trends characterize the older. This pattern is also consistent with the continental collision model. On collision, the passive plate margin retains its intrinsic pre-collision pattern of structural (gravity) trends, whereas the active margin (the future younger terrane) develops new trends parallel to the arc and trench. Such a model seems to be the simplest way of explaining contrasting trends in adjacent crustal blocks.

Paired gravity anomalies and patterns of contrasting gravity trends have been sought and recognized in Archean areas of the Canadian Shield in an attempt to identify some of the older and smaller protocontinental building blocks. In the western Superior Province, a paired gravity signature with characteristics that match closely those associated with Proterozoic sutures is recognized along the boundary between the Sachigo-Uchi and English River Subprovinces. The polarity of palaeosubduction inferred from the gravity pair and from trends either side of the proposed suture zone is southward. In the eastern Superior Province, contrasting gravity patterns indicate another major junction in the Ungava Peninsula. Different aeromagnetic trends suggest a possible suture in the Slave Province, although this has a less clear gravity expression.

The large gravity data bases for Canada and the U.S.A. provide almost complete coverage of the North American continent and have been used to produce a horizontal gravity gradient map. The trends of these gradients outline the major structural fabric of the continent. The North American continent has been subdivided into several first-order gravity domains on the basis of orientation, magnitude, continuity and pattern of gradient features (25). Gravity domains defined on this basis correlate closely with such features as Precambrian structural provinces and Phanerozoic orogens and demonstrate the utility of gravity domains as indicators of discrete large-scale structural units. Using the simple criterion of parallel and oblique trends at domain boundaries as an indicator of palaeosubduction direction, it is possible to establish a pattern of growth for the North American continent, including the roughly 40% of the continent buried by relatively undeformed and thin Phanerozoic platform deposits. This pattern is dominated by outward growth from the Superior and Wyoming Provinces (including the proposed extension of the latter into part of the Churchill

Province (26)). The Slave Province formed another nucleus of growth, although this is not strongly expressed in the gravity patterns.

The isotopic ages of a series of roughly east-west trending, buried Proterozoic crustal provinces in the south-central U.S.A. decrease southward from the Archean core of the continent (11,13,27), consistent with the outward growth pattern proposed on the basis of gravity patterns. Prominent gravity signatures are not, however, associated with these boundaries. It has been suggested that these crustal provinces were produced by a series of marginal basin closures and Andean-type orogenies associated with a northward-dipping subduction zone that migrated southward and accreted roughly 1300 km of sialic crust to the south of the Wyoming Province in the interval between 1800 and 1100 Ma ago (11). Models proposed for the Shield to the north (8) involve the collision of Archean protocontinents and major reactivation of the protocontinent overriding the subduction zone. In this scenario, new additions to the continental crust are restricted to granite belts and calc-alkaline volcanics produced by arc-type magmatism, obducted oceanic-type crust and sediments derived from these various lithologies. These additions, along with crustal thickening, metamorphism and dynamic activity in the collision zone, which juxtaposes crust of different character, produce mass anomalies which are reflected in the characteristic paired gravity anomalies. Conversely, the crust in the south-central U.S.A. is regarded to be a juvenile addition of continental crust that does not involve a contribution from older Archean crust (11). The lack of prominent gravity signatures associated with these terranes may be a reflection of the accretion process proposed for this region. It is viewed as proceeding in a relatively uniform fashion by lateral migration of a single subduction system (11); such a mechanism may not have produced the large density contrasts in the crust that are typical of collisional orogens.

References

(1) Kay M. (1951) Geol. Soc. Am. Mem. 48, 143 p. (2) Stille H. (1934) Res. and Progress, 1, p. 9-14. (3) Wilson J.T. (1948) Trans. Am. Geophys. Un., 29, p. 691-726. (4) Dewey J.F. (1969) Nature, 222, p. 124-129. (5) Monger J.W.H. et al. (1982) Geology, 10, p. 70-75. (6) Williams H. and Hatcher R.D. Jr. (1983) in Geol. Soc. Am. Mem. 158, p. 33-53. (7) Hoffman P.F. (1980) in Geol. Assoc. Can. Spec. Pap. 20, p. 523-549. (8) Gibb R.A. et al. (1983) Precam. Res., 19, p. 349-384. (9) Wardle R.J. et al. (1986) in Geol. Assoc. Can. Spec. Pap. 31, p. 13-29. (10) Karlstrom K.E. and Houston R.S. (1984) Precam. Res., 25, p. 415-446. (11) Condie K.C. (1982) Geology, 10, p. 37-42. (12) Sims P.K. and Peterman Z.E. (1986) Geology, 14, p. 488-491. (13) Bickford M.E. et al. (1986) Geology, 14, p. 492-496. (14) Langford F.F. and Morin J.A. (1976) Am. Jour. Sci., 276, p. 1023-1034. (15) Blackburn C.E. et al. (1985) in Geol. Assoc. Can. Spec. Pap. 28, p. 89-116. (16) Folinsbee R.E. et al. (1968) Am. Geophys.Un. Geophys. Monog. 12, p. 441-448. (17) Irving E. et al. (1974) J. Geophy. Res., 79, p. 5491-5502. (18) Gibb R.A. and Halliday D.W. (1974) Earth Physics Branch, E.M.R., Ottawa, Gravity Map Series 124-131, 36 p. (19) Thomas M.D. and Tanner J.G. (1975) Nature, 256, p. 392-394. (20) Kearey P. (1976) Earth Plan. Sci. Lett., 28, p. 371-378. (21) Gibb R.A. and Thomas M.D. (1976) Nature, 262, p. 199-200. (22) Hoffman P.F. (1987) Submitted to Geology. (23) Dewey J.F. and Burke K.C.A. (1973) Jour. Geol., 81, p. 683-692. (24) Wellman P. (1976) J. Geol. Soc. Austral., 23, p. 11-14. (25) Thomas M.D. et al. (1987) Submitted to Nature. (26) Thomas M.D. et al. (1987) Geology in press. (27) Anderson J.L. (1983) in Geol. Soc. Am. Mem. 161, p. 133-154.

SOLID EARTH AS A RECYCLING SYSTEM. Jan Veizer, Institute of Earth Sciences, Hebrew University, Jerusalem.

If plotted on mass (volume, area, tonnage) vs. time diagrams, geologic entities- for example, continental and oceanic crust, sediments, and mineral resources- display an exponential (power law) relationship, with entity per unit time increasing toward the present. This relationship is consistent with the concept of recycling and can be simulated mathematically. The authors' approach is based on the plate tectonic theory and considers area-age or mass-age distributions of crystalline basement and sediments for major global tectonic realms. Each tectonic realm is characterized by a specific lifespan, which is an inverse function of its recycling rate (b). The estimated average half-area or half-mass ages (τ_{50}) are the following: basins of active margins ~ 30 Ma, oceanic sediments ~ 40 Ma, oceanic crust ~ 55 Ma, basins of passive margins ~ 80 Ma, immature orogenic belts ~ 100 Ma, mature orogenic belts ~ 380 Ma, and platforms ~ 380 Ma. The corresponding parameters for continental crust are 690 Ma for K/Ar, and ~ 1200 Ma for Rb/Sr and U-Th/Pb dating pairs. For Sm/Nd τ_{50} is $\lesssim$ 1800 Ma, suggesting either that continental crust was growing during geologic history and/or its recycling via mantle (or lower crust) was more vigorous prior to 2500 Ma ago. The estimate maximal survival (or oblivion) ages for these tectonic realms are ~ 3-3.5 times longer than their τ_{50}. Tectonic diversity preserved in the geologic record is therefore a function of time, with oceanic tectonic realms, because of their rapid recycling, underrepresented in the rocks older than ~ 300 Ma. Sm/Nd isotopic systematics of sediments suggests that, for a near steady-state post-Archean sedimentary mass, recycling is ~ 90 $\pm$ 5% cannibalistic. This yields an estimated upper limit on crust-mantle exchange via sediment subduction of ~ 1.1 $\pm$ 0.5 x 10 g a^{-1} (~ 0.5 $\pm$ 0.2 km^3 a^{-1}); considerably less than demanded by isotopic constraints. The discrepancy may indicate the existence of additional loci, such as orogenic belts, for significant crust-mantle interaction.

MINERALIZATION THROUGH GEOLOGIC TIME: EVOLUTION
OF CONTINENTAL CRUST;

Jan Veizer, Derry Laboratory, Department of Geology, University of Ottawa,
Ottawa, Ontario, Canada, K1N 6N5 and Ottawa-Carleton Geoscience Centre;
Peter Laznicka, Department of Geological Sciences, University of Manitoba,
Winnipeg, Manitoba, Canada, R3T 2N2; and
S.L. Jansen, Department of Computer Science, Carleton University, Ottawa,
Ontario, Canada, K1S 5B6.

In analogy to living systems, geologic entities (e.g. rocks, mineral
deposits, tectonic realms and domains) are involved in the processes of
perpetual generation and destruction ("birth/death" cycles). This results
in time distribution patterns akin to age structures in living populations
and the systematics is amenable to treatment by the concepts of population
dynamics. Utilizing this theoretical approach, we predict the survivorship
patterns for major realms of the plate tectonic system, for their
constituent rocks, and for the entombed mineral resources. The present
inventory encompasses global economic accumulations of metals (Pb, Zn, Cu,
Au, Mn, Fe, Cr, Ni, U, Al, and Sn) by geologic age. The deposits of these
metals have been assigned to nine broad genetic categories, with an attempt
to relate each category to tectonic setting within the framework of the
global plate tectonics. The discussed categories are the following: (1)
magmatogene deposits associated with mafic-ultramafic rocks, (2) magmatogene
associated with acid differentiates, (3) volcanic, (4) volcano-sedimentary,
(5) hydrothermal-epigenetic, (6) detrital-sedimentary, (7) chemical-
sedimentary, (8) weathering crusts, and (9) high-grade metamorphosed and
metamorphogenic ores. The total cumulative tonnage for all categories and
metals decreases with increasing geologic age, but genetic categories within
each commodity show a consistent age pattern. In general, the rates of
tonnage decrease with age are in the following succession: weathering crusts
> detrital-sedimentary > hydrothermal > chemical-sedimentary > volcano-
sedimentary > ultramafic > metamorphic. This progression reflects a
decreasing role of recycling, and an enhanced evolutionary component, from
weathering crusts to metamorphic genetic categories of deposits. Simulation
of metallogenesis during geologic history, based on the concepts of recycl-
ing for host tectonic realms, enables differentiation of steady-state
features of mineralization from its non-uniformitarian evolutionary
component. The pattern generated by this approach is designed to maximize
the steady-state interpretation and the surviving mineralization epochs
probably understate the extent of evolution. Based on this minimalistic
approach, the overall metallogenic evolution of the Earth, can be divided
into five partially overlapping stages. These are: (1) The greenstone belt
stage, dominant in the Archean and petering out ~1.8 Ga ago, with gold
mineralization, Algoma type iron ores, and massive sulfides; (2) The
cratonization stage, with the peak in the Late Archean and Early
Proterozoic, and containing the paleoplacer type Au and U deposits, the bulk
of iron (Superior type) reserves, and related Mn deposits; (3) The rifting
stage, at ~1.8 ± 0.3 Ga ago, typified by mafic-ultramafic dykes and layered
complexes, and related mineralization of Cr, Ni, Cu, Au, Fe as well as
frequent base metal deposits, hydrothermal U, and volcano-sedimentary Mn;
(4) The stable craton phase, ~1.7 to 0.9 (or 0.6) Ga ago, with common
alkalic volcanism and plutonism but a dirth of mafic-intermediate volcano-
plutonic associations and metallic ores. The significant ores have been of
exogenic type confined to the cratonic sedimentary cover (unconformity U,
chemical sedimentary Mn and Zambian type Cu); (5) The Phanerozoic stage of

continental dispersal, characterized by varied and frequent mineralization, particularly of hydrothermal type. The fundamental control of mineralization appears to have been the tectonic evolution of continental crust, with the stages of aggregation and cratonization (Archean-Early Proterozoic), subsequent rifting (~1.8 ± 0.3 Ga ago), thickening expressed in relative stability and intertia (Middle and Late Proterozoic), and fragmentation and dispersal of the existing (super)continent(s) during the Phanerozoic. The secular progression of exogenic deposits of U, Fe, Mn and Cu is consistent with the increasing oxygenation of the atmosphere-hydrosphere system in the course of terrestrial evolution. However, contrary to the prevailing dogma, it is not clear whether this was a consequence of biological innovations or is a reflection of the evolving supply-demand redox balances for the complementary abiotic exo- and endogenic cycles.

N88-30251   ⅀⨍O-46
/56064

GROWTH OF THE CONTINENTAL CRUST: A PLANETARY-MANTLE PERSPECTIVE

Paul H. Warren   CD 146017

Institute of Geophysics, University of California, Los Angeles, CA 90024

The lack of Earth rocks older than about 3.8 Ga is frequently interpreted as evidence that the Earth formed little or no subduction-resistant continental crust during the first 700 My of its history. Such models obviously imply that the pre-3.8 Ga Earth was covered entirely or almost entirely by smoothly subducting oceanic crust (or else by no crust at all). On the other hand, the thermal regime of the early Earth probably tended to cause the oceanic crust at this time to be comparatively thin and comparatively mafic [e.g., Arndt, 1983]. The present Earth is covered by about 59% oceanic crust, averaging about 7 km in thickness, and 41% continental crust, averaging roughly 40 km in thickness. Thus continentless-early-Earth models would seem to imply a total mass of crust <<1/3 that of the present Earth. From a planetary perspective, it seems more likely that the Earth produced a substantial volume of continental crust well before 4.0 Ga, albeit little or none of this early continental material survives — for excellent reasons: (1) Continental crust is subduction-resistant, not subduction-proof. The rate of subduction of continental crust must have decayed exponentially from an early rate that would have been orders of magnitude higher than the modern rate. (2) By a simple analogy with the lunar crust, any crust extant prior to about 3.9 Ga was exposed to frequent, tremendously destructive megaimpacts. In fact, the rate of destruction per unit of surface area was probably far greater on the Earth than on the nearby but smaller Moon.

Isotopic studies indicate that the bulk of the Moon's crust formed <300 My after the origin of the solar system. Although the lunar magmasphere (or magma "ocean") hypothesis remains controversial, there can be no doubt that primordial heating caused massive melting long before the conventional heat sources, K, Th, and U would by themselves have engendered melting. Detailed petrologic studies tend to favor at least a thin global "ocean" of fully molten matter to account for the single most abundant lunar rock type, the ferroan anorthosites. Like most other "pristine" lunar rocks the ferroan anorthosites formed as igneous cumulates. However, the ferroan anorthosites have many peculiarities, e.g., high Fe/Mg despite paradoxically low Na/Ca, indicating an origin separate from the other cumulates (mainly norites and anorthositic troctolites), collectively referred to as the Mg-rich suite. The favored model to account for these contrasts requires the ferroan anorthosites to be products of plagioclase accumulation (flotation) over a primordial magmasphere, whereas Mg-rich rocks are products of slightly more recent intrusions, which tended to assimilate ferroan anorthosite and incompatible element-rich magmasphere residual liquid (later recycled into KREEP basalts). Even the average plagioclase content of the upper lunar crust ($\sim$ 75 wt%) is hard to explain without invoking plagioclase concentration by flotation over a magmasphere; explaining the much higher average plagioclase content of the ferroan anorthosites ($\sim$ 95 wt%) is impossible without a magmasphere. Also, virtually all mare basalts have (-)Eu anomalies, yet in most cases their source residual solids were plagioclase-free, implying that the sources were produced by an earlier episode of plagioclase depletion extending hundreds of km into the Moon. A similarly great depth for the magmasphere is also implied by the absence, within the sampled, upper-crustal portion of the Moon, of mafic cumulates complementary to the ferroan anorthosites.

One longstanding objection to the magmasphere hypothesis, that heating by accretion would not suffice to simultaneously melt a large fraction of the Moon, has been weakened by recent appreciation for the potential role of large impacts in the origin of the Moon. In fact, this objection never was cogent. Diverse igneous meteorites show that massive primordial melting also occurred even on modest-sized asteroids, for which accretional heating is negligible. Thus some other primordial heat source (^{26}Al?, electromagnetic induction?) must have been a key factor in the earliest evolution of the smaller planets. There are no nonlunar meteoritic anorthosites, however, so it seems unlikely that even the largest asteroids were ever molten to the same "magma ocean" extent as the primordial Moon.

Extrapolating from these "known" early planetary evolutions to the Earth involves tremendous scaling uncertainties. The outcome of differentiation will be governed among other factors by (a) phase equilibria, which are sensitive to pressure, which is in turn a function of planet size; (b) the effectiveness and duration of mixing by convection (not only during the primordial heating era, but also subsolidus convection long after most of the primordial melt crystallizes), which also correlates with planet size; and (c) bulk composition, especially volatile contents, which in the Moon are extraordinarily low (a shell of silicate matter hot enough to constitute a "magma ocean" would presumably soon outgas most of its volatiles, however).

The Earth's magmasphere apparently never produced an anorthositic crust commensurate in scale to that of the Moon; instead the Earth's oldest crust is roughly tonalitic, and even today the continental crust is apparently thinner than that of the Moon. However, from a mantle perspective the contrast between anorthositic norite and tonalite is a rather fine distinction. Compared to Ringwood's [1979, p. 34] "primordial model mantle" composition (derived from CI chondrites), the average composition of the modern continental crust [Taylor, 1982a] is enriched in Si by a factor of 1.18, but enriched in aluminum by a factor of 5.1. (It is also enriched in Ca by a factor of 2.4, essentially unchanged with respect to Fe, and depleted in Mg by a factor of 10.) The Moon's crust [Taylor, 1982a] is, from this perspective, remarkably similar; although Si is essentially unchanged relative to the identical "primordial" composition, Al is enriched by a factor of 6.9. Indeed, from the perspective of the underlying mantle (near-surface density = about 3.35 g cm^{-3}), the Earth's oceanic crust is (even after it has cooled) nearly as buoyant as its continental crust.

It has been briefly suggested [Warner, 1979, Taylor, 1982b] that the Earth's magmasphere failed to float an anorthositic crust because its melt was not dense enough due to a relatively high content of H_2O. However, calculations à la Bottinga and Weill [1970] indicate that any realistic combination of major element composition and H_2O content for the magmasphere melt will lead to plagioclase flotation. The failure of the Earth's magmasphere to generate a crust commensurate to that of the Moon in Al content and thickness is better explained as a consequence of the tendency for the aluminum of the Earth's magmasphere to enter garnet ± spinel, both of which are stabilized by high pressures, instead of exclusively plagioclase as in the case of the lunar magmasphere.

Hofmeister [1983] constructed the only previous quantitative model for the crystallization of a terrestrial magma ocean. Her modeling was limited to two initial compositions and one initial, comparatively shallow (120 km) depth. As argued previously by Warren [1985], it seems unlikely, barring an overwhelming, instantaneous input of heat (such as the superimpact hypothesised to have produced the Moon), that the Earth's fully molten magma "ocean" was ever much thicker than roughly 30 kbar

(i.e., roughly 100 km), because the depth of the partial melting necessary to generate the "ocean" would have been limited by the disparity in dT/dP between basaltic-peridotitic melting curves and adiabats (convection in the partially molten shell would guarantee that its dT/dP was essentially adiabatic).

Recent high-pressure experimental petrology studies [Takahashi, 1986; Ohtani et al., 1986, Scarfe and Takahashi, 1986] have made it possible to model the crystallization of a deeper magmasphere, and to make more realistic assumptions about the initial composition that a relatively shallow magmasphere would have. Results indicate that, regardless of initial magmasphere depth, garnet, and to a lesser extent aluminous high-pressure pyroxenes, act to prevent the Earth's magmasphere from ever producing more than about 45 km of crust (global average) with the Al_2O_3 content (18 wt%) estimated by Taylor [1982a] for the continental crust. This 45 km is an extremely conservative upper limit, based on discounting spinel crystallization (as tends to occur when the magma ocean is roughly 50 km deep) and assuming perfect fractional crystallization. Analogous models for the Moon result in a crust (assuming 25 wt% Al_2O_3) roughly 60 km thick, without spinel ever becoming a factor. In Earth models based purely on fractional crystallization, the overall composition of the crust (neglecting the difficult-to-model potential for plagioclase flotation) is an aluminous basalt, and tends to have an extremely low mg ratio, and an unrealistically high Ca content. Better "fits" to the composition of the continental crust are obtained when the "ocean" is assumed to receive periodic fresh injections of primitive melt (diminishing in flux as the ocean solidifies), but the degree of Si enrichment is never >>1. A similar scenario has been advocated for the Moon [Warren, 1985]. An initially aluminous-gabbroic continental crust may have evolved into an aluminous-tonalitic one by a combination of intracrustal differentiation, catalyzed by H_2O [Campbell and Taylor, 1983], plus renewal via mantle subduction and recycling. Aside from the first few My of solar system history, the relative abundance of "continental" or "highlands" crust is probably far more a function of planet size than it is a function of time.

References:

Arndt, N. T. [1983] Geology, v. 11, 372-375.

Bottinga, Y. and Weill, D. F. [1970] Am. J. Sci., v. 269, 169-182.

Campbell, I. H. and Taylor, S. R. [1983] Geophys. Res. Lett., v. 10, 1061-1064.

Hofmeister, A. M. [1983] J. Geophys. Res., v. 88, 4963-4983.

Ohtani, E., et al. [1986] Nature, v. 322, 352-353.

Ringwood, A. E. [1979] Origin of the Earth and Moon, Springer.

Scarfe, C. M. and Takahashi, E. [1986] Nature, v. 322, 354-356.

Takahashi, E. [1986] J. Geophys. Res., v. 91, 9367-9382.

Taylor, S. R. [1982a] Planetary Science: A Lunar Perspective, LPI.

Taylor, S. R. [1982b] Phys. Earth Planet. Inter., v. 29, 233-241.

Warner, J. L. [1979] In Workshop on Ancient Crusts of the Terrestrial Planets, p. 79-82, LPI.

Warren, P. H. [1985] Ann. Rev. Earth Planet. Sci., v. 13, 201-240.

N88 - 30252

CRUSTAL UNDERPLATING IN THE NORTH AMERICAN CRATON - EVIDENCE FROM THE PETROGENESIS OF ANORTHOSITIC ROCKS IN THE MIDDLE PROTEROZOIC MIDCONTINENT RIFT SYSTEM.

P. W. Weiblen and J.D. Miller Jr., Minnesota Geological Survey, 2642 University Ave., St. Paul, MN 55114, USA.

Circumstantial evidence for underplating as a significant mechanism for the growth of continental crust continues to accumulate, particluarly from considerations of thermal, seismic, and density constraints (1-4) on the lower crust and upper mantle. Cox (5) has made compelling geochemical and petrological arguments that in continental flood basalt provinces, the volume of underplated cumulates may be at least as large as the expression of the surface lavas. We present here new geological, petrological, and geochemical data on anorthositic rocks in the Duluth Complex (6) which indicates that underplating processes similar to those postulated by Cox (5) played a significant role in their origin. The interpretations may be applicable to the origin of anorthositic rocks in general.

The Duluth Complex represents an exposed section of mafic-feldspathic intrusions in the Middle Proterozoic Midcontinent Rift System (7,8). Early-formed anorthositic rocks (An_{50-78}) plagioclase > 80% in the mode) occur as an extensive cap to a younger series of layered troctolitic intrusions. The origin of the anorthositic suite of rocks has been enigmatic for the following reasons:

1) The field occurrence of the anorthositic rocks is complex. Outcrops consist of meter- to km-sized blocks of different plagioclase cumulates with complex intrusive and inclusion relationships. Although seven cumulus types can be distinguished on the basis of textures and intercumulus mineralogy (ol, cpx, ilm-mt), they do not define a stratiform sequence (7).

2) No mafic nor ultramafic units have been found that could represent their fractionated counterpart in a closed magmatic system with any conventional parent magma.

3) The plagioclase compositions of different textural types all overlap and indicate equilibration down to intermediate compositions, yet individual grains are commonly zoned over the entire range of composiitons An_{50-78}.

4) Although the anorthositic rocks formed earlier than the troctolitic rocks, their intercumulus mafic minerals crystallized from more evolved melts than the later troctolitic rocks.

Figure 1 summarizes a correlation found between the mineral compositions and textures of anorthositic rocks. Compositions of plagioclase show only a limited variation with significant variation in olivine. However, granular, interstitial olivine has a higher Fo content than poikilitic olivine. The former may be interpreted to reflect intercumulus trapping of olivine nuclei in a plagioclase mush saturated in olivine and plagioclase and the latter, trapping of olivine undersaturated melt which nucleated olivine after trapping. Figure 2 summarizes new data on DC argon plasma spectrophotometric analyses of plagioclase separates. As in figure 1, the An variation is restricted but there is a significant variation in mg. The variation of minor Fe and Mg substitution in the plagioclase provides a more accurate assessment of mg variations in the magmas in which the plagioclase equilibrated than the associated intercumulus mafic phases. The data suggest that plagioclase equilibrated with melt over a significant range of crystallization while melt compositions were changing due to fractionation of mafic phases. This could occur in systems in which plagioclase was suspended in the melt while mafic phases fractionated.

The variations found in both figures 1 and 2 could be produced in magma chambers where the anorthositic rocks solidified. These would have been shallow chambers at about 2 kilobars based on estimates of the thickness of the overlying volcanic edifice (about 7 km) (9) and the record of equilibration of metamorphosed country rocks (10,11). However, significant volumes of mafic cumulates would have resulted from the required fractionation of olivine and/or pyroxene. The base of the cumulates is exposed or has been intersected in drill core over a distance of over 200 km in the central part of the Duluth Complex and no appropriate mafic cumulates have been found (7). For this reason and others related to our growing understanding of the probable origin of compositional variations in the hypabyssal and volcanic rocks in the Midcontinent Rift we conclude that magma

chambers at lower to subcrustal depths were an essential part of the rift system.

The magmas which produced the anorthositic rock series in the Duluth Complex as well as the later troctolitic rocks are also the logical source for the associated hypabyssal rocks and lavas of the North Shore Volcanic Group (NSVG) (12). Trace element variations in the NSVG lavas such as those illustrated in figure 3 can be explained by extensive fractionation of mafic phases and limited fractionation of plagioclase from a MORB-like parent magma. Sc/Ni and V/Cr variations in the NSVG lavas require significant clinopyoxene fractionation relative to olivine and plagioclase. Mass balance calculations for major elements for fractionation models for derivation of the NSVG lava variations give similar results, e.g. derivation of 14.3 wt. % of an intermediate lava from a MORB-like parent requires fractionation of 19.6% Ol, 16.3% Cpx, 49.7 % Pl and 0.2% Cr-spinel. This is consistent with experimental data for expected cotectic proportions which would crystallize from a tholeiitic melt at 7 to 10 kilobars (13,14).

In a dynamic rift environment the four enigmatic characteristics of the anorthositic rock series, defined above, appear to us to be plausible consequences of the generation of plagioclase mushes at depth and their transfer to near surface chambers. Conditions under which oliviine and clinopryoxene would fractionate more efficiently than plagioclase due to relative density contrasts would appear possible in stable magma chambers at depth. Thus giving rise to the observed An/mg characteristics of plagioclase (point 2 and Fig. 2). Interruption of the accumulation of plagioclase under various convection and flow regimes and transfer of such accumulations to near surface chambers would be the inevitable result of instabilities resulting from episodic rifting of the deep chambers giving rise to the complex textural and intercumulus mineralogy of the anorthositic rocks (point 1). The complex zoning in intermediate composition plagioclase (point 3) could be a result of the dynamic process of stripping of melt from crystals during the upward and lateral transfer of plagioclase mushes. Finally, in the early stages of rifting, there might logically be longer periods of tectonic quiessence which would allow fractionation to proceed further before transfer of magma from lower to upper chambers; thus giving rise to the evolved nature of the early anorthositic rock suite (point 4). We conclude that fractionation of mafic phases and equilibration of plagioclase in tholeiitic magmas in lower- to sub-crustal magma chambers in a rift-related terrain is a viable model for generation of the anorthositic rock series in the Duluth Complex.

Inaddition to the evidence presented here for lower crustal magma chambers in the Midcontinent Rift and the evidence developed by Cox (5) for other plateau flood basalt provinces, Philpotts and Reichenbach (14) suggest similar models for the differentiation of Mesozoic basalts in the Hartford Basin in Connecticut. Also Phinney has suggested subcrustal chambers for the generation of Archean anorthositic suites (15). It would appear that underplating of continental crust by fractionation of tholeiitic magmas at depth is a long-standing and significant aspect of the growth of continental crust.

References:
(1)   Brown L.D. and others (1981 in Evolution of the Earth. Geodynamics series vol. 5, Am Geophys Union, p. 38,52 and pers. comm. (1986).
(2)   McKenzie D. (1984) Nature, p. 616, 618.
(3)   Herzberg C.T., Fyfe W.S., Carr, M.J. (1983) Contrib Mineral Petrol, p. 1, 5.
(4)   Furlong K.P. and Fountain D.M. (1986) J. Geophysical Research, p. 8285, 8294.
(5)   Cox K.G. (1980) Journal of Petrology, p. 629, 650.
(6)   Miller J.D. Jr. (1985) Unpubl PhD thesis, U. of Minnesota, Mpls, MN, pp. 525.
(7)   Weiblen P.W. and Morey G.B. (1980) Am. Jour. Sci., vol 280-A, p. 88-133.
(8)   Weiblen P.W. (in press) in Decade of North American Geology Series, GSA.
(9)   Green J.C. (1983) Tectonophysics, p. 413,437.
(10)   Simmons E.C. and others (1974) Journal of Petrology, p. 539, 564.
(11)   Lobatka T.C. and others (1981) American Mineralogist, p. 70, 86.
(12)   Green J.C. (1981) in Basaltic Volcanism Study Project, Basaltic Volcanism on the Terrestrial Planets, Pergamon Press, N.Y., p. 30-77.

(13)   Presnall D.C. and others (1979) Journal of Petrology, p. 3, 35.
(14)   Presnall D.C. and Hoover J.D. (1984) Contr. Min. Petrol., p. 170-174.
(15)   Philpotts A.R. and Reichenback I. (1985) GSA Bull., p. 1131,1139.
(16)   Phinney W.C. (1982) Workshop on Magmatic Processes of Early Planetary Crusts: Magma
       Oceans and Stratiform Layered Intrusions: Houston, LPI Tech. Rept. 82-10, p. 121-124.
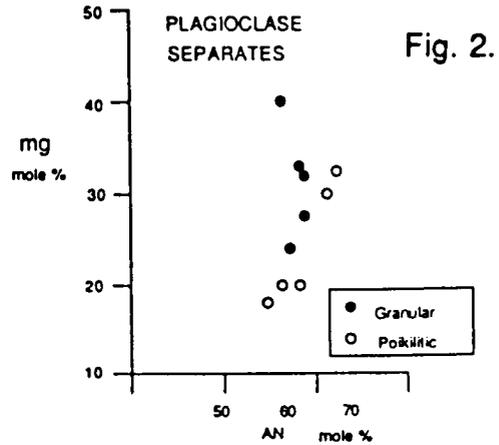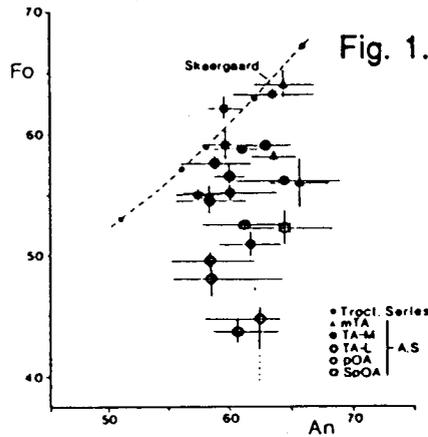
Fig. 1. An vs Fo (mole %). Averages (symbols) and ranges (error lines) of plagioclase (An) and olivine (Ol) in anorthositic and troctolitic rocks from the Duluth Complex. Data from (6). Solid symbols - granular interstitial olivine; mTA - medium grained troctolitic anorthosite; TA-M - troctolitic anorthosite, main stage; TA-L - troctolitic anorthosite, late stage; pOA - poikilitic anorthosite; SpOA - subpoikilitic anorthosite.

Fig. 2. An vs mg (MgO/(MgO+ FeO + Fe_2O_3).   Data from DC argon plasma spectrophotometric analyses of plagioclase separates from anorthositic rocks (6). Note the indication of plagioclase equilibration with more evolved interstitial melt in the poikilitic samples.
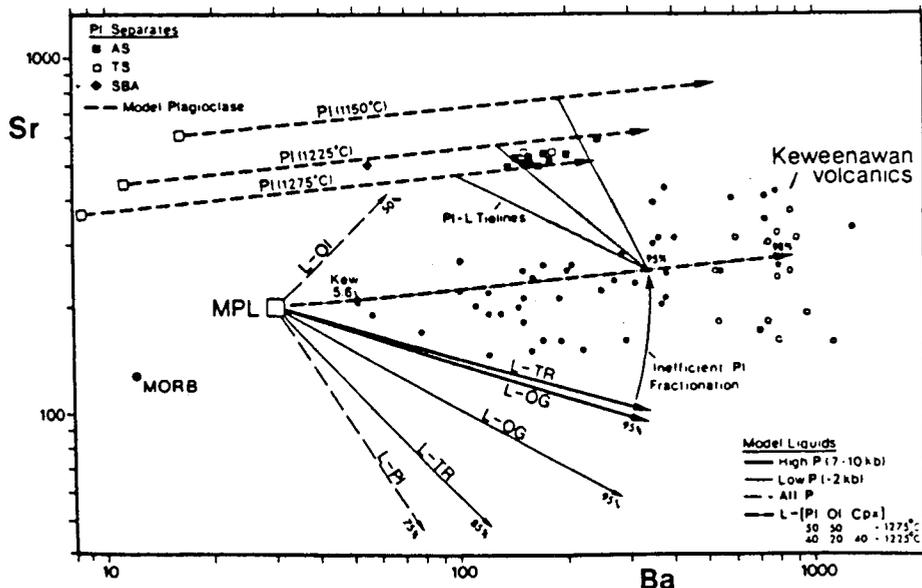

Fig. 3. Fractional crystallization modelling of Sr and Ba variations found in NSVG lavas. Data from (6). MPL - a model MORB-like primary liquid composition with the trace element enrichments found in the NSVG lavas (12). Solid lines with arrows are trajectories of MPL fractioned liquids produced by fractionation of model cotectic assemblages in %: Plagioclase (pl)-67, olivine (Ol)-30, spinel (Sp)-3 (Liquid (L) - trocotlite (Tr), high pressure (P); Pl-45, Ol-10, Cpx-45 (L-olivine gabbro (OG), low P). Dashed lines labelled Pl represent Pl compositions calculated from T dependent distribution coefficient data (6). Pl-L tie lines connect Pl compositions at different temperatures with a single lava composition. Pl separate data - AS, anorthositic Pl; TS, troctolitic Pl; SBA, Silver Bay anorthosite (6). The dot-dash trajectory cannot be produced by the fractionation of the cotectic mineral proportions. Fractionation of cotectic mafic minerals and suspension of Pl will produce the variation.

LATE-ARCHAEAN CRUSTAL GROWTH IN THE LEWISIAN COMPLEX OF NORTHWEST SCOTLAND - DIACHRONEITY IN MAGMATIC ACCRETION AND IMPLICATIONS FOR MODELS OF CRUSTAL GROWTH.

M.J. Whitehouse, Dept. of Earth Sciences, University of Oxford, Parks Road, Oxford OX1 3PR, England.

The Lewisian Complex of northwest Scotland is an Archaean basement terrane that has been reworked during the Proterozoic (1). Previous geochronological studies have established that the complex represents an entirely new crustal addition during late-Archaean times (2,3), and the difference between a Sm-Nd 'accretion' date of 2920 ± 50 Ma (3) and a Pb/Pb 'metamorphic' date of 2680 ± 70 Ma (2) has been interpreted as indicating a crustal accretion-differentiation superevent (CADS (4)) lasting 240 ± 110 Ma. A combined Rb-Sr, Pb/Pb, and Sm-Nd isotopic study reported here has been applied to the Lewisian CADS to attempt to address the problem of whether this comprised a short regional scale magmatic accretion event followed ~200 Ma later by regional high-grade metamorphism, or a long period of episodic or semi-continuous magmatic accretion prior to high-grade metamorphism.

The mainland outcrop of the Lewisian Complex consists largely of tonalitic-trondhjemitic-granodioritic gneisses (TTG) which range in metamorphic grade from amphibolite facies to pyroxene-granulite facies. Studies of LILE depletion (5,6) and metamorphic petrology (6) suggest that at the time of regional metamorphism the presently juxtaposed assemblages represented a cross-section through the middle- to lower-crust.

Three felsic gneiss suites from this crustal profile are considered here: the amphibolite facies of the northern region, the hornblende-granulite facies gneisses of the Gruinard Bay area of the central region, and the pyroxene-granulite facies gneisses of the Scourie area of the central region.

Sr- and Pb-isotopic data have confirmed that there is a prograde increase in LILE depletion and at the highest metamorphic grade Pb and Th become mobile. A 2950 ± 70 Ma Pb/Pb isochron reported from the northern region amphibolite facies gneisses (7) is now interpreted as a spurious date resulting from partial U-depletion, and in general, restriction of sample suites to TTG gneisses only does not result in useful Pb/Pb geochronological data for the metamorphic differentiation. There are however several reliable dates which suggest that this metamorphism occurred at <2700 Ma - the previously mentioned 2680 ± 60 Ma Pb/Pb date, a 2660 ± 20 Ma U-Pb zircon date

(8), and a 2650 ± 110 Ma Sm-Nd whole-rock date (9), and its regional extent is evidenced by a 2700 ± 20 Ma U-Pb zircon date from the Outer Hebrides (10).

Sm-Nd regression data considered in ε_{Nd} - t space shows first, that the Scourie gneisses have been subject to open system Sm-Nd behaviour during granulite facies metamorphism, and second, that the two lower grade suites in which Sm-Nd is assumed to be undisturbed cannot have both common age <u>and</u> source. The regression data for these latter two suites indicates high ε_{Nd}(t) values (+3.6 to +4.7), well above contemporaneous depleted mantle (11). High ε_{Nd}(t) values such as these require either an unrealistically high $^{147}Sm/^{144}Nd$ ratio in the mafic precursor, or this source to have had an unacceptably long isolation with $^{147}Sm/^{144}Nd$ greater than DM or CHUR. However, the 2σ limits on the regressions overlap with model depleted mantle, and the interpretation of lower t and ε_{Nd}(t) is favoured. Evidence that the sub-Lewisian late-Archaean mantle was light-REE depleted relative to CHUR comes from a Sm-Nd whole-rock isochron study of three layered mafic/ultramafic bodies in the Lewisian that are assumed to be mantle derived. One of these, Scouriemore, yields an errorchron date of 2670 ± 110 Ma with ε_{Nd}(t) = +1.0 ± 0.7, and granulite facies open system behaviour is indicated. The other two bodies, Achiltibuie and Drumbeg, yield respectively isochron dates of 2850 ± 95 Ma and 2910 ± 55 Ma, with ε_{Nd}(t) values of +1.9 ± 0.5 and +1.7 ± 0.3, both of which lie on contemporaneous model depleted mantle. If all three felsic gneiss suites had a common or similar source, with Sm-Nd characteristics close to model late-Archaean depleted mantle, t_{DM} model ages may be used to examine the accretion times of these suites. These t_{DM} model ages, shown in figure 1, range from 3040 to 2840 Ma for the Scourie area gneisses (terrane average ~2930 Ma), from 2880 to 2800 Ma for the Gruinard area gneisses (terrane average ~2860 Ma), and from 2850 to 2690 Ma for the northern region gneisses (terrane average ~2780 Ma). This sequence suggests a progressive increase in accretion time with increasing crustal depth. However the period of accretion from >2930 Ma to ~2700 Ma is too long to account for high-grade metamorphism at <2700 Ma, unless significant magmatic overthickening occurred shortly before this event. There is no preserved evidence for such magmatism.

A model for late-Archaean crustal growth in the Lewisian must be able to account for both the diachronous accretion sequence revealed by the t_{DM} model ages, and the high-grade metamorphism at <2700 Ma.

Two models for crustal growth during late-Archaean times are considered. The first is essentially an analogue of modern plate tectonic processes, with TTG magma produced by melting of 'oceanic' crust at a subduction zone. Diachronous accretion could reflect the development of successive marginal

basins and island arcs resulting in accretion at continental margins (12). Overthickening and high-grade metamorphism could be the result of collisions between larger composite terranes for which modern analogies exist (13). The second model is one in which subduction is absent (14); diachronous TTG production could occur by movement of mafic proto-crust over a mantle plume, or by periodic movement of the plume itself. Intra-crustal deformation could produce the necessary thickening required for high-grade metamorphism. It is concluded that detailed geochronological studies such as the one reported here cannot effectively differentiate between these two end-member models.

Figure 1
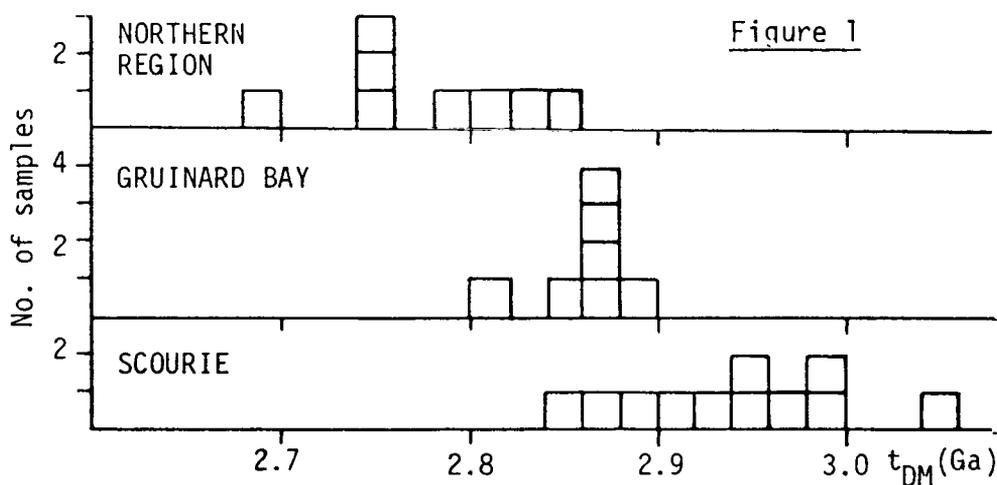
References:

1 J. Watson, 1975, in "Precambrian", Geol. Soc. Lond. spec. publ. 6, 15-29.

2 H.J. Chapman & S. Moorbath, 1977, Nature 268, 41-2.

3 P.J. Hamilton et al., 1979, Nature, 277 25-8.

4 S. Moorbath, 1977, Chem. Geol. 20, 151-87.

5 J.W. Sheraton et al., 1973, in R.G. Park & J. Tarney (eds) "The early Precambrian of Scotland and related rocks of Greenland", Univ. of Keele, pp 13-30.

6 M.B. Fowler, 1986, J. metamorphic Geol. 4, 345-59.

7 M.J. Whitehouse & S. Moorbath, Nature 319, 488-9.

8 R.T. Pidgeon & D.R. Bowes, 1972, Geol. Mag. 109, 247-58.

9 F.J. Humphries & R.A. Cliff, 1982, Nature 295, 515-7.

10 R.T. Pidgeon & M. Aftalion, 1972, Earth planet. Sci. Lett. 17, 269-74.

11 D.J. DePaolo, 1981, Nature 291, 193-6.

12 J. Tarney et al., 1977, in B.F. Windley (ed) "The early history of the Earth", Wiley, pp 131-46.

13 J.W.H. Monger et al., 1982, Geology 10, 70-5.

14 A. Kröner, 1981, Rev. Brasiliera de Geociências, 12, 15-31.

S 4013461

N88-30254

THE EFFECT OF THICKER OCEANIC CRUST IN THE ARCHAEAN ON THE
GROWTH OF CONTINENTAL CRUST THROUGH TIME; M.E. Wilks, Department of Applied
Geology, University of Strathclyde, James Weir Building, 75 Montrose Street,
Glasgow, G1 1XJ, Scotland. Now at Department of Geosciences, New Mexico
Institute of Mining & Technology, Socorro, NM 87801.

Present crustal evolution models fail to account for the generation of
the large volumes of continental crust in the required time intervals. All
Archaean plate tectonic models, whether invoking faster spreading rates,
similar to today's spreading rates, or longer ridge lengths, essentially
propose that continental crust has grown by island arc accretion due to the
subduction of oceanic crust. The petrological differences that characterize
the Archaean from later terranes result from the subduction of hotter oceanic
crust into a hotter mantle.

If the oceanic crust was appreciably thicker in the Archaean, as
geothermal models would indicate, this thicker crust is surely going to have
an effect on tectonic processes. The assumption that 20-50 km thick oceanic
crust is going to subduct in a similar mechanism as seen today needs to be
carefully reassessed. A more valid approach is to compare the possible
styles of convergence of thick oceanic crust with modern convergent zones.
The best modern analogue occurs where thick (35 km) continental crust is
colliding with thick (35 km) continental crust.

Oceanic crustal collision on the scale of the present-day Himalayan
continental collision zone may have been a frequent occurrence in the
Archaean, resulting in extensive partial melting of the hydrous underthrust
oceanic crust to produce voluminous tonalite melts, leaving a depleted
stabilised basic residuum.

Present-day island arc accretion may not have been the dominant
mechanism for the growth of the early Archaean crust. As the mantle cooled
through time, so the thickness of oceanic crust decreased with a concomitant
increase in continental crustal thickness and the dominant style of convergence
changed from that of a Himalayan oceanic crustal collision setting, to the
type of subduction now occurring at the Marianas, Scotia arcs and under the
west coast of America.

VOLCANIC CONTRIBUTION TO CRUSTAL GROWTH IN THE CENTRAL ANDES: A NEW ESTIMATE AND A DISCUSSION OF UNCERTAINTIES.

C.A. Wood, Johnson Space Center, SN4, Houston, TX 77058, USA
P. Francis, Lunar and Planetary Inst., Houston, TX 77058, USA.

Volcanism above subduction zones is a major mechanism for crustal growth (Reymer and Schubert, 1984), and compared to some other proposed processes calculation of growth rates is relatively easy given accurate volumes and ages of volcanic materials. Francis and Rundle (1976) first used this approach in a small region of the Central Andes, and extrapolated their result to the entire Central Andean arc. Their derived rate of 3 to 4.2×10^{-6} $km^3/yr/km$ of arc length is here compared with an independent estimate based upon a reconnaissance census of all major volcanoes in the Central Andes.

A NEW ESTIMATE: Using Landsat MSS images and astronaut photographs, we are cataloging every significant volcanic cone in the Central Andes (Wood et al, 1986). In the 1000 km of active arc between latitudes 18° and 27° south, approximately 500 composite conical volcanoes were constructed over the last 20 million years. Based on average cone dimensions (which yield a volume of 60 km^3/cone) a volume of 30000 km^3 of volcanic material has been erupted in the last 20 million years, for an average magma production rate of 1.5×10^{-3} km^3/yr. For the entire region the rate is thus 1.5×10^{-6} $km^3/yr/km$ of arc length. For the assumption that the volume of flat-lying ignimbrites is 75% that of the conical volcanoes, as in the small area mapped in detail by Francis and Rundle (1976), yields a total magma production rate of 2.6×10^{-6} $km^3/yr/km$. This number - and that for the cones alone - is in reasonably good agreement with the estimate by Francis and Rundle, which was based on detailed mapping of only 1 degree of latitude. However, before this coincidence lulls us into extrapolating the rate over all arcs through all time, or even accepting that it is very significant for the Andes, we must evaluate uncertainties.

INTRUSION VS EXTRUSION: How much material was intruded compared to the amount erupted? Models by Smith (1979) suggest that each ignimbrite represents only about 10% of the volume of its magma chamber. Smith suggested that the same relation holds for the andesitic conical volcanoes. Thus the magma production rates derived above may be 10 times too low. Replenishment of magma chambers and subsequent eruptions reduce the errors in estimates of production rates, but few

volcanoes are known to have had more than 2 or 3 large ignimbrite eruptions (and magma chambers would be replenished after each); simple models demonstrate that as much as 75% of the magma generated is never erupted.

NEAR VS FAR DEPOSITION: Estimates of magma production rates for volcanic arcs are based upon measurements of lava and ash in vent regions. But how much of the material from an eruption is deposited near-vent compared to that which is widely dispersed? Wide-spread distribution of tephra is common for explosive eruptions such as those which characterize Andean volcanoes, and there is a systematic increase in the percentage of total erupted material that is deposited beyond a vent area as the explosivity and volume of an eruption increase. For small eruptions such as those that build cinder cones, more than half of the erupted volume is deposited beyond the actual cone itself (Wood, 1980); for very explosive eruptions, such as Taupo (Walker, 1980), nearly 100% of the ejecta is widely dispersed and no volcanic cone is constructed at all. Quantitative evaluation of the volumetric distribution of near and far volcanic material is provided by Sigurdsson and Carey (1981) who demonstrated that 445 km^3 (out of 527 km^3 total) of material deposited from eruptions of Lesser Antilles volcanoes during the last 10^5 years is found in marine sediments far beyond their eruptive centers. Additionally, for large eruptions (volume > 0.1 km^3; Newhall and Self, 1982), significant amounts of ash enter the stratosphere and are distributed globally, thus further increasing the underestimate of magma volumes based on near vent measurements. In fact, there are no reliable estimates of erupted volumes for large explosive eruptions such as build the Central Andes. Based on the data from the Lesser Antilles, we might estimate that the observed cones in the Andes represent only 15% of the total amount of erupted material. This is an overestimate for we have ignored the ash globally distributed by the stratosphere.

NEW VS TRANSFORMED CRUSTAL MATERIAL: What is the origin of the lavas and ashes that have been discussed above? Are they new crustal materials formed of mantle rocks that have been transmogrified by various physical and chemical processes, or are they derived from pre-existing lower crust that has been revitalized by subduction related heat and volatiles? This has been a major question in petrology for decades, but various types of isotopic data now consistently indicate that oceanic sediments and granitic crustal materials are involved in petrogenesis of Andean magmas. Francis et al. (1980) have argued, however, that the source regions for Andean andesites have a Sr ratio of 0.704, but only

relatively small amounts of contamination are required to explain higher values. Ignimbrites have a stronger crustal component, and represent, at least in part, crustal recycling. Thus, we conclude that the erupted (and intruded) magmas are dominately (roughly 75%, considering relative volumes of andesites and ignimbrites) mantle-derived and largely represent a new addition to the crust.

THE KEEL: The estimates of volume and uncertainty discussed above ignore the tremendous volume (and uncertainties) of the intruded batholith or keel under the Andes. Estimates of crustal production rates *including the keel* range from 9.9 to 33 x 10^{-6} km^3/y/km arc length (Francis and Rundle, 1974; Reymer and Schubert, 1984), or 3 to 13 times more than based on the observed surface deposits.

A REVISED ESTIMATE: Based on the discussion above we suggest the following components for the magmatic crustal growth rate in the Central Andes over the last 20 my:

Observed volcanic deposits	2.6 x 10^{-6} km^3/yr/km
Magma stranded in chambers	7.8
Distant deposits	17.3
Stratospherically distributed	0.3 (??)
TOTAL	28.0
Minus reworked crust	7.0
Provisional Grand Total	21.0 x 10^{-6} km^3/yr/km

Thus, crustal growth rates derived from measures of volcanics at the surfaces of convergent plate margins are probably underestimates by a factor of about ten. The total volume of erupted and near-surface magmatic material is comparable to the volume of the keel, suggesting that previous crustal growth estimates for the Central Andes are too low.

REFERENCES: Francis, P.W., R.S. Thorpe, S. Moorbath, G.A. Kretzschmar and M. Hammill (1980) Strontium isotope evidence for crustal contamination of calc-alkaline volcanic rocks from Cerro-Galan, NW Argentina, *Earth Planet. Sci. Lett. 48*, 257-267. **Francis, P.W. and C.C. Rundle** (1976) Rates of production of the main magma types in the central Andes, *Geol. Soc. Am. Bull. 87*, 474-480. **Newhall, C.G. and S. Self** (1982) The volcanic explosivity index (VEI): An estimate of explosive magnitude for historical volcanism, *J. Geophys. Res. 87*, 1231-1238. **Reymer, A. and G. Schubert** (1984) Phanerozoic addition rates to the continental crust and crustal growth, *Tectonics 3*, 63-77. **Sigurdsson, H. and S.N. Carey** (1981) Marine tephrochronology and Quarternary explosive volcanism in the Lesser Antilles arc, in *Tephra Studies* (ed. by S. Self and R.S.J. Sparks), Reidel, Holland, 255-280. **Smith, R.L.** (1979) Ash-flow magmatism, in *Ash-Flow Tuffs* (ed. by C.C. Chapin and W.E. Elston), Geol. Soc. Am. Sp. Paper 180, 5-28. **Walker, G.P.L.** (1980) The Taupo Pumice: Product of the most powerful known (untraplinian) eruption? *J. Volc. Geothermal Res. 8*, 69-94. **Wood, C.A.** (1980) Morphometric evolution of cinder cones, *J. Volc. Geothermal Res. 7*, 387-413. **Wood, C.A., P. Francis and A. Bury** (1986) 500 volcanoes: A census of the Central Andes, *Abst, Internat. Volcan. Congress*, New Zealand, 343.

List of Attendees

Dallas Abbott
Lamont-Doherty Geological Observatory
Palisades, NY 10964

F. Albarède
CRPG-CNRS
BP 20
54501 Vandoeuvre Cedex, France

N. T. Arndt
Max-Planck-Institut
Postfach 3060
D-6500 Mainz, F. R. Germany

L. D. Ashwal
Lunar and Planetary Institute
3303 NASA Road One
Houston, TX 77058-4399

Kevin Burke
Lunar and Planetary Institute
3303 NASA Road One
Houston, TX 77058-4399

T. Chacko
University of Chicago
Department of Geological Sciences
Chicago, IL 60637

A. S. Cohen
University of Cambridge
Department of Earth Sciences
Downing Street
Cambridge CB2 3EQ, England

W. J. Collins
Macquarie University
School of Earth Sciences
North Ryde, NSW 2109, Australia

W. C. Compston
Australian National University
Department of Geophysics and Geochemistry
P.O. Box 4
Canberra, ACT 2600, Australia

J. Dasch
Oregon State University
Department of Geology
Corvallis, OR 97331

J. Dewey
University of Durham
Department of Geological Science Laboratories
South Road
Durham SH1 3LE, England

R. Ellam
University of Oxford
Department of Earth Sciences
Parks Road
Oxford, OX1 3PR, England

R. Forsythe
Rutgers University
Wright Laboratories
Department of Geological Sciences
New Brunswick, NJ 08903

S. J. G. Galer
University of California-San Diego
Scripps Institution of Oceanography
A-012
La Jolla, CA 92093

S. L. Goldstein
Max-Planck-Institut für Chemie
Saarstrasse 23
D-6500 Mainz, F. R. Germany

D. Green
University of Tasmania
Department of Geology
GPO Box 252C
Hobart 7001, Australia 23-0566

S. E. Haggerty
University of Massachusetts
Geology Department
Amherst, MA 01003

S. Harley
Department of Earth Sciences
University of Oxford
Parks Road
Oxford OX1 3PR, England

W. K. Hartmann
Planetary Science Institute
2030 E. Speedway, 201
Tucson, AZ 85719

D. G. Howell
U.S. Geological Survey
MS 999
345 Middlefield Road
Menlo Park, CA 94025

S. L. Jansen
Carleton University
Department of Computer Science
Ottawa, Ontario, K1S 5B6
Canada

P. Kinny
Australian National University
Research School of Earth Sciences
GPO Box 4
Canberra, ACT 2601, Australia

S. Klemperer
BIRPS
Bullard Laboratories
Madingley Road
Cambridge CB3 0EZ, England

A. Kröner
University of Mainz
Department of Geosciences
Postfach 3980
D-6500 Mainz, F. R. Germany

I. S. McCallum
University of Washington
Department of Geological Sciences
AJ-20
Seattle, WA 98195

F. McDermott
The Open University
Department of Earth Sciences
Walton Hall
Milton Keynes, MK7 6AA, England

W. McDonough
Research School of Earth Sciences
Australian National University
Canberra, ACT 2601, Australia

G. E. McGill
Department of Geology and Geography
University of Massachusetts
Amherst, MA 01003

J. McLelland
Department of Geology
Colgate University
Hamilton, NY 13346

S. Moorbath
Department of Geology and Mineralogy
Oxford University
Parks Road
Oxford OX1 3PR, England

P. Morgan
Geology Department
Box 6030
Northern Arizona University
Flagstaff, AZ 86011-6030

C. Neal
Department of Geological Sciences
University of Tennessee
Knoxville, TN 37996

Eric Nelson
Department of Geology
Colorado School of Mines
Golden, CO 80401

R. Rudnick
Research School of Earth Sciences
Australian National University
GPO Box 4
Canberra, ACT 2601, Australia

G. Ryder
Lunar and Planetary Institute
3303 NASA Road One
Houston, TX 77058-4399

U. Schärer
University of Quebec
Department of Earth Sciences
C.P. 8888, succ. A
Montreal, H3C 3P8, Canada

G. Schubert
University of California
Department of Earth and Space Sciences
Los Angeles, CA 90024-1567

C. R. Seeger
Western Kentucky University
Department of Geology
Bowling Green, KY 42101

M. Stein
Hebrew University
Department of Geology
Institute of Earth Sciences
Jerusalem, Israel

J. Tarney
University of Leicester
Department of Geology
Leicester LE1 7R4, England

P. N. Taylor
University of Oxford
Department of Earth Sciences
Parks Road
Oxford OX1 3PR, England

S. R. Taylor
Australian National University
Research School of Earth Sciences
GPO Box 4
Canberra, ACT 2601, Australia

M. D. Thomas
Geological Survey of Canada
1 Observatory Crescent
Ottawa, Ontario K1A 0Y3
Canada

P. B. Toft
Department of Geology
University of Massachusetts
Amherst, MA 01003

J. Veizer
Derry Laboratory
Department of Geology
University of Ottawa
Ottawa, Ontario K1N 6N5
Canada

P. Warren
University of California
Institute of Geophysics
Los Angeles, CA 90024-1567

P. Weiblen
University of Minnesota
Space Science Center
Minneapolis, MN 55455

M. J. Whitehouse
University of Oxford
Department of Earth Sciences
Parks Road
Oxford OX1 3PR, England

C. A. Wood
NASA Johnson Space Center
Code SN4
Houston, TX 77058