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The First 800 Million Years of Earth's History

J. V. Smith

Phil. Trans. R. Soc. Lond. A 1981 **301**, 401-422
doi: 10.1098/rsta.1981.0120

Phil. Trans. R. Soc. Lond. A **301**, 401–422 (1981)
Printed in Great Britain

The first 800 million years of Earth's history

By J. V. Smith, F.R.S.
Department of the Geophysical Sciences, University of Chicago, Chicago, Illinois 60637, *U.S.A.*

The Earth grew initially from hot planetesimals of reduced material, and began to differentiate early into a metal-rich core, a silicate-rich mantle and a volcanic surface. Accretion ended with volatile-rich oxidized material, which did not interact with the core because of a mantle barrier. Quenching of magmas at the surface and foundering of cool material was effective in cooling the Earth by 1600 J per gram of magma. Magmas transported radioactive elements and volatiles to the outer 50 km. The lower mantle became stranded below the critical adiabat for melting. The core may have been fully molten.

Radiogenic and non-radiogenic heat sources produced a surface heat flow that decreased from *ca.* 1.3×10^{14} W at 4.45 Ga B.P. to the present value of 4×10^{13} W. In spite of uncertainties in the production and transport of heat, it seems safe to conclude that intense volcanism occurred throughout the pre-Archaean era before survival of crust. From phase-equilibrium data for peridotites and calc-alkaline rocks, it is deduced that a range of ultrabasic to basic magmas was produced from a peridotitic upper mantle, and that calc-alkaline magmas were produced during foundering of crust. An anorthositic crust was not formed.

From observations of lunar basins, it is concluded that a minimum of *ca.* 500–1000 impact basins were formed on Earth in *ca.* 100–200 Ma before 3.95 Ga B.P. Plausible calculations suggest that *ca.* 20 times as many basins were formed in the preceding *ca.* 300 Ma. Impact debris would be cooled by water in wet crustal rocks and oceans.

It is assumed that after 4.0 Ga B.P. impacts would have only temporary local effects on the crust, and emphasis is placed on upper-mantle convection as the main driving force for geochemical transport. Sedimentation and metamorphism were important factors during foundering of crust. 'Oceanic' and 'continental' regions developed above up-welling and down-welling segments of convection cells, and these regions resembled greenstone belts and high-grade regions of the Archaean era. As heat flow declined, *polygonal tectonics* was replaced by *symmetrical linear tectonics* during the Archaean era, and then by *asymmetrical linear tectonics* during the Proterozoic era.

Introduction

Because the Earth should have been accreting near the time of crystallization of most meteorites (*ca.* 4.55 Ga B.P.), and because the earliest surviving crust dates back only to *ca.* 3.8 Ga B.P., there is no direct geological information on the first 750 Ma of Earth history. Consequently this paper is based on controversial inferences drawn from the Moon, other planets and meteorites, coupled with backward extrapolation from surviving terrestrial rocks, especially those of Archaean age. Readers are warned that there is no consensus on the origin of the Earth and Moon, and that a new model for heterogeneous accretion (Smith 1979) differs from other models, including a model for homogeneous accretion (Ringwood 1979). The present paper incorporates material from articles in Windley (1975), and has some but not all ideas in common with those in Shaw (1980), Lambert (1980*a*) and Tarling (1978). Space restrictions forbid referencing to many hundreds of relevant papers, and readers are therefore referred to Smith (1979) and a comparative review of planetary crusts from a lunar perspective (Smith 1980). Papers on the role of impacts on crustal evolution (Frey 1980; Grieve 1980) appeared after preparation of the manuscript.

Accretion

Cosmochemical and cosmophysical evidence

(*a*) Because of (i) survival of asteroids ranging up to 1000 km across, (ii) need for bodies over 100 km across to explain the mineralogy and petrology of some iron meteorites and achondrites, (iii) presence of huge impact basins in the crusts of Mercury, Moon, Mars and perhaps Venus, and (iv) dynamical theory, it is assumed that small bodies collided to produce a decreasing number of planetesimals of increasing mass culminating in a single planet in each growth zone (Smith 1979, pp. 11–19).

(*b*) Because many meteorites crystallized near 4.55 Ga B.P., and the Moon developed distinct crustal and subcrustal reservoirs of radiogenic elements by at least 4.4 Ga B.P. (Carlson & Lugmair 1979), the Moon must have accreted most of its mass ($>99.9\%$) within *ca.* 100 Ma.

(*c*) Because capture of a fully grown Moon would have destroyed distinct crustal and subcrustal reservoirs, the Moon must have been in Earth orbit by *ca.* 4.4 Ga B.P.; backward extrapolation from tidal-energy models that the Moon was at the Roche limit at *ca.* 3 Ga B.P. are misleading.

(*d*) All calculations of the temperature profile of the Earth depend on controversial assumptions, even when constrained by chemical, mineralogical and petrological observations. Strong heat sources were needed to melt (i) parent bodies of some meteorites, by 4.55 Ga B.P., (ii) at least half of the Moon, not later than 4.4 Ga B.P., and (iii) probably all of Mercury, not later than the time of formation of impact basins. Because gravitational accretion energy of *cold* bodies, even when coupled with radiogenic heat, is insufficient to produce such extensive early melting, it is assumed that early planet-forming bodies were hot (>1000 K). Because of evidence for excess ^{26}Mg in some meteorites, an obvious candidate is radiogenic heat from ^{26}Al, whose half life is 0.7 Ma: electrodynamic induction heating is also possible. If even parent bodies of some meteorites melted, it is plausible that the Earth underwent crystal–liquid differentiation as it grew, thereby causing continual transfer of radioactive elements to the surface by volcanism (Smith 1979, p. 78). Paradoxically, such early continued differentiation of a hot growing Earth can produce a cooler final Earth than late differentiation of an initially cold Earth. Core formation is continuous. For accretion over 10^8 a, with intense volcanic activity transferring radiogenic heat and gravitational infall energy continuously to the surface for radiation to space, most of the Earth can remain crystalline. The theoretical temperature of 4000 K at 500 km depth (Kaula 1979*a*) is misleading because melting would have transferred heat to the surface, leaving peridotite near the solidus temperature of *ca.* 2500 K (Kaula 1979*b*). Most energy from sinking of Fe,FeS-rich diapirs should enter the core (Shaw 1978), which might become fully molten. Convection through a near-solid mantle would transfer heat upwards. Although Jakosky & Ahrens (1979) ignored the initial relative velocity between planetesimals and the Earth, and the upward transfer of heat, their conclusion that the Earth's surface remained cold appears correct: indeed solar radiation provides 10^4 times more heat than radioactivity in the present Earth.

Because xenoliths from the upper mantle contain substantial amounts of noble siderophile elements (e.g. Au, Ir), and ^{3}He is still being released from the mantle (Smith 1979, p. 29), the

Earth did not volatilize, at least during accretion of the last one-sixth of its mass, and convection in the mantle was insufficient to allow equilibration of the outer part of the Earth with the metallic core.

(*e*) Because of the need to find sources of meteorites ranging from highly reduced enstatite meteorites to oxidized carbonaceous meteorites, the highly speculative assumption was made (Smith 1979, p. 77) that planetesimals ranged from reduced types near the Sun to oxidized types from the asteroid zone outwards; indeed oxygen isotopic ratios allow assembly of both the Earth and Moon from a mixture of highly reduced and moderately reduced meteorites, coupled with a small fraction of oxidized meteorites.

To explain roughly equal fractions of CO_2 on Earth and Venus and essential absence of CO_2 on Mercury and the Moon, it was assumed that planetesimals rich in CO_2 and other volatiles were sling-shot into the inner Solar System by the giant planets, at velocities high enough (20–40 km/s) that only the two larger planets, with escape velocity near 10 km/s, were able to retain gaseous impact debris; the smaller planets, Mercury and Moon, with low escape velocity could not do so (Smith 1979, p. 77). Mars retained some volatiles because the ratio of impact velocity to escape velocity was intermediate. Unfortunately it is difficult to model high-energy collisions (see, for example, O'Keefe & Ahrens 1977), but no other explanation has appeared so far of the apparent correlation between planetary mass and volatile content. Currently there seems to be no escape from the need for late accretion of volatile-rich debris onto a largely crystalline silicate mantle.

(*f*) Although an intense solar wind is needed to dissipate almost all the H and He expected in the inner solar nebula, time and duration are difficult to define theoretically. Perhaps hot planetesimals lost most of their gaseous elements before accreting into the inner planets. Because Venus contains little water even though it should have captured nearly equal masses of H_2O and CO_2 from late oxidized planetesimals, most of the primordial H is presumed lost through the atmosphere and ionosphere, while heavier CO_2 was retained. For the Earth, loss of two-thirds of the primordial H is speculative (Smith 1977). Freed oxygen would be sufficient to convert about 5–10 km thickness of FeO to Fe_3O_4; of course, some oxygen might be combined with C and S.

New Earth model

These arguments led to a new Earth model (figure 1), which incorporates some features from earlier models but rejects others. When the Earth was five-sixths grown at approximately 50 Ma after condensation began in the solar nebula (figure 1 *b*), it had captured most planetesimals in its growth zone. Progressive melting had produced (i) a liquid core rich in Fe and siderophile elements, (ii) a silicate mantle, composed largely of reduced material, and (iii) an unstable surface zone composed of volcanic rocks containing most of the radioactive elements. The silicate mantle had a metal/silicon ratio not much greater than unity, and a low ratio of Fe^{2+}/Mg. Its outer part was composed largely of pyroxenes, olivine and garnet, and the inner part of high-pressure equivalents.

The Moon was nearly fully grown, and was close to but outside the Roche stability limit (*ca.* Earth radii). Some debris from planetesimals impacting the Moon was lost to Earth (Hodges 1979), and little volatile material was retained. The Earth retained all impacting material expect for some light gases. Incoming planetesimals were more oxidized than earlier ones, and the metal/silicon ratio was greater than unity because Fe^{2+} increased the olivine/pyroxene ratio.

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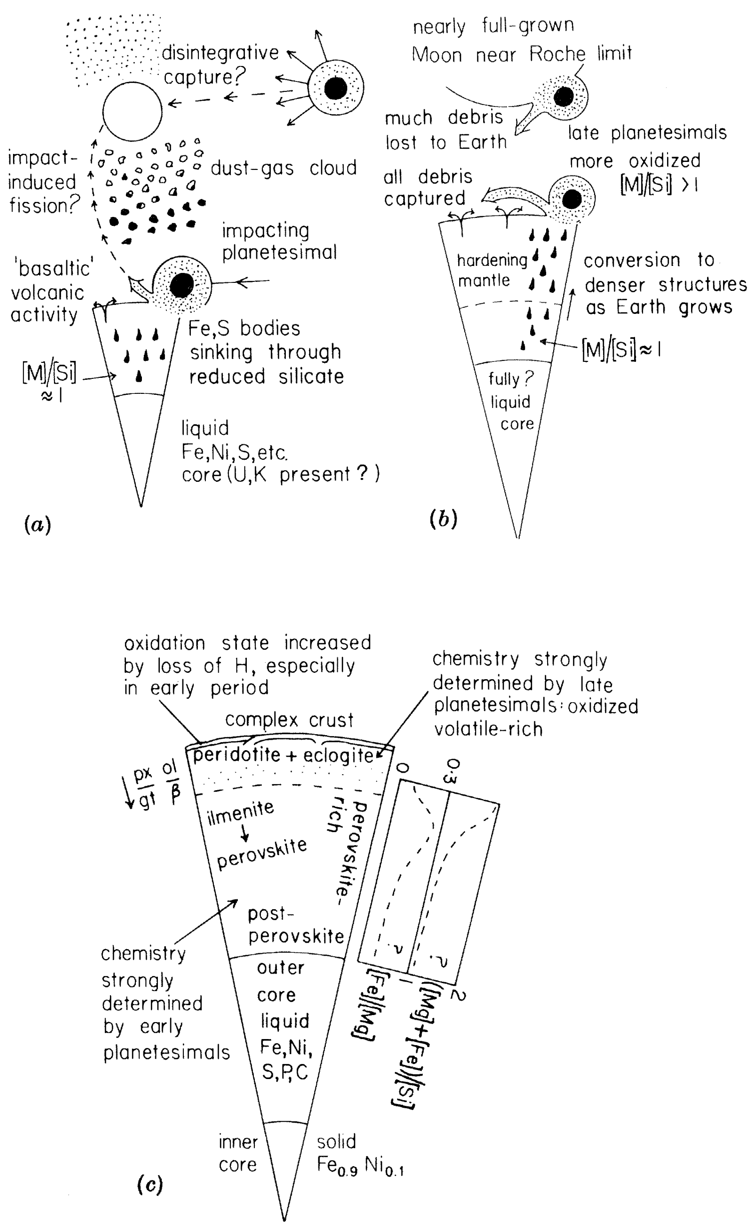

FIGURE 1. Scheme for heterogeneous accretion of the Earth and Moon. (*a*) The infant Earth is capturing planetesimals from the middle of its growth zone, and differentiation is occurring into a liquid Fe,Ni,S-rich core and a silicate mantle. Volcanic activity transfers heat to the surface, and Fe,S-rich bodies are sinking through silicate. (*b*) The adolescent Earth is capturing planetesimals that originally formed at the outer parts of the growth zone, with a preference for the Mars side over the Venus side. These planetesimals are more oxidized on average, and have a greater ratio of octahedral cations to Si. The lower mantle does not equilibrate completely with new material. The Earth retains all volatile-rich impact debris, but the Moon loses most. (*c*) The present Earth has a complex crust overlying an upper mantle composed of peridotites and eclogites. With increasing pressure, olivine inverts to β-phase and spinel, and clinopyroxene becomes incorporated into garnet. Below 670 km depth, the lower mantle consists of high-pressure minerals, of which the perovskite variety of $(Mg,Fe)SiO_3$ may be dominant. Most of the core remains liquid, but part has crystallized into the inner core. From Smith (1979, fig. 12).

Some 50 Ma later, the Earth reached essentially its present mass. Sinking FeS-rich diapirs kept the core molten. Growth of the core throughout accretion of the Earth obviates late catastrophic core formation.

The present Earth (figure 1*c*) has a solid inner core, whose crystallization from the liquid outer core may provide energy to drive the magnetic dynamo. The mantle is almost entirely crystalline, and is split into a lower mantle of high-pressure minerals, including the perovskite variety of (Mg,Fe, etc.) SiO_3, and an upper mantle of peridotite and eclogite rocks. The crust is extremely complex, and is rich in magmatophile elements.

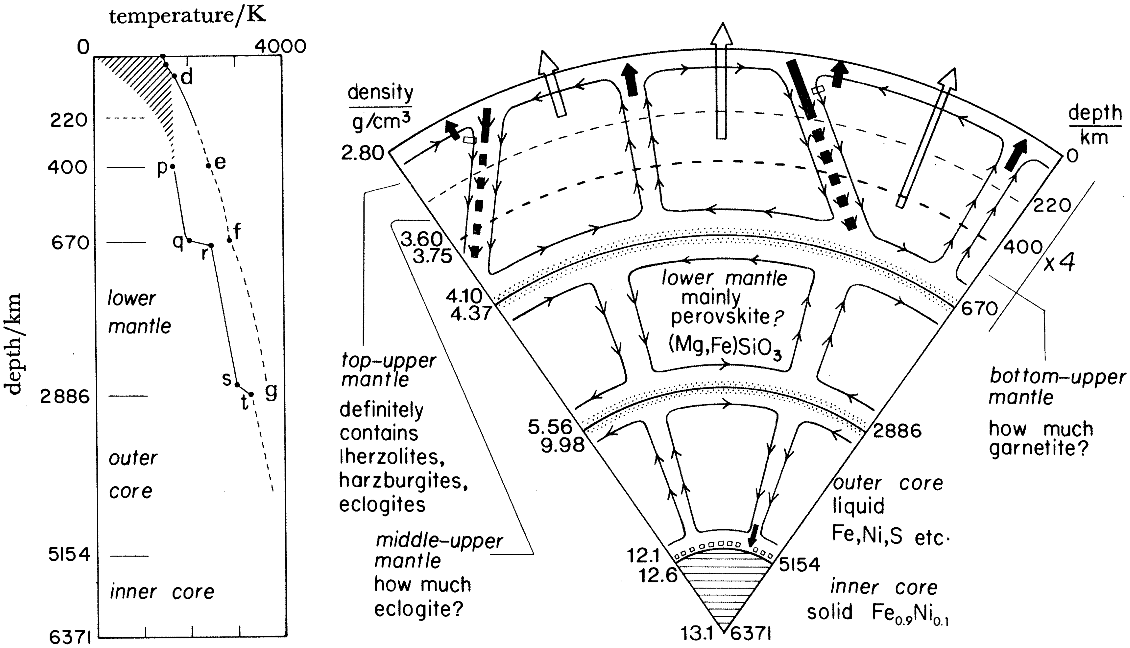

FIGURE 2. Convection systems, thermal boundary zones and temperature profile for the present Earth. Positions and shapes of convection systems were chosen more for artistic than scientific reasons. The inner core grows by sinking of metal crystals from the outer core. Thermal boundary zones (dotted) separate the lower mantle from the outer core and upper mantle. The upper mantle, whose thickness is magnified fourfold, is split into top, middle and bottom sections, with suggested boundaries at *ca.* 220 and 400 km depth, following Anderson (1979). Formalized convection systems pass through the two boundaries. Magmas reach the surface from the interior of convection cells (cf. Tatsumoto 1978; open arrows), and from upwellings (mid-ocean-ridge basalts) and mantle wedges (both as solid arrows). Subduction is illustrated by filled rectangles, and interaction with a mantle wedge by open rectangles. The temperature profile from Jeanloz & Richter (1979, fig. 6*c*) has segments pq and rs controlled by adiabatically driven convection, and segments qr and st by thermal conduction. The shaded region in the diagram on the left covers most of the range between hot mantle rising under ocean ridges and cold slabs sliding under continental margins. The segmented curve def is a hypothetical extrapolation of the solidus for arid peridotite. A lower mantle composed mainly of $MgSiO_3$–perovskite might begin to melt near the purely hypothetical curve fg. A discontinuity might occur at f if the bulk composition changes at 670 km depth.

It is assumed (figure 2) that the mantle of the present Earth is convecting in two levels, with a thermal boundary region at the density discontinuity near 670 km. Such a boundary is consistent with various geophysical evidences (McKenzie & Weiss 1975; Anderson 1979; Jeanloz & Richter 1979; Richter 1979), and makes it easier to explain the chemical properties of the upper mantle (O'Nions *et al.* 1979). The convection cells are merely schematic, and complications definitely must occur, especially with respect to minor discontinuities at *ca.* 220 and *ca.*

400 km depth. Certainly the top section of the upper mantle contains various harzburgites, lherzolites and eclogites, but the nature of the middle and bottom sections is speculative (see later). Generation and transport of magmas are shown schematically by arrows (see legend to figure 2).

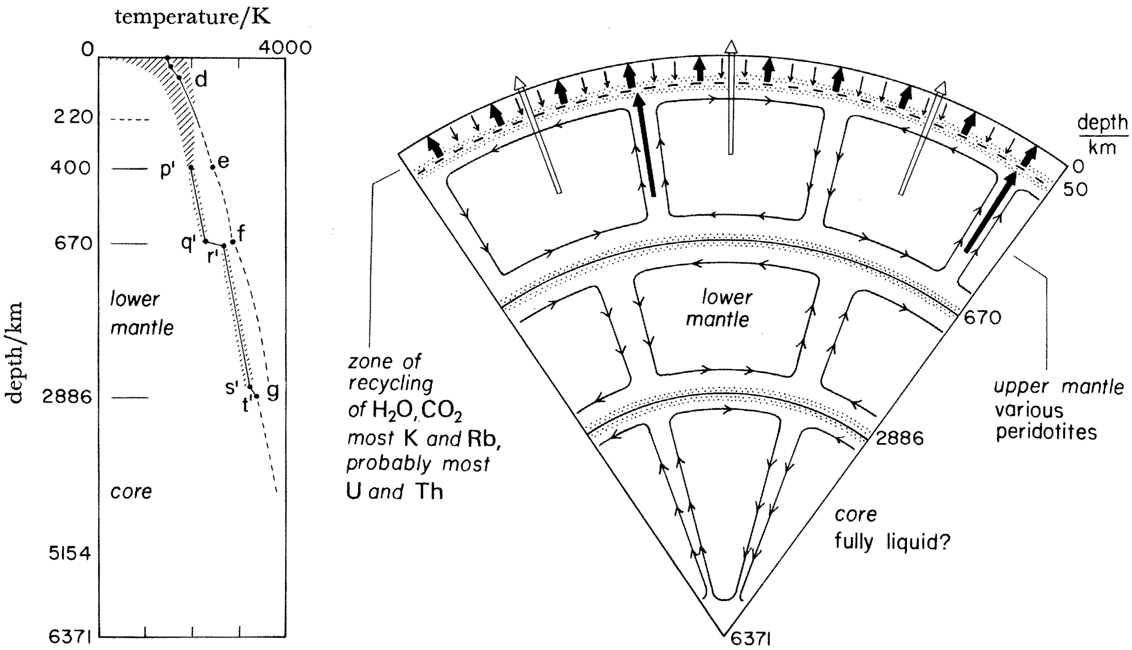

FIGURE 3. Suggested movements and temperature profile in the Earth at 4.45 Ga B.P. Compare with figure 2.

The two-level convection system of the mantle is assumed to go back to the end of accretion (Lambert 1980*a*), thereby reducing chemical interaction between materials accreted early and late. Figure 3 is a suggested convection system at 4.45 Ga B.P. Thermal boundary zones separate the lower mantle from a core (fully liquid?) and the upper mantle. Radiogenic heating and impact-induced phenomena are concentrated in the outer 50 km (see later), and intense volcanic activity (short heavy arrows) is coupled with sinking of cooled material (short light arrows), which melts again as it becomes heated. Volatiles are concentrated in the outer 50 km, and the upper mantle is composed mainly of arid peridotites. Arid is defined as being denuded in both H_2O and CO_2. Magma generation is less intense below 50 km, and is represented schematically by arrows from up-welling regions and from centres of convection cells. Plate tectonic processes have not begun.

Further discussion will follow evaluation of (i) constraints from phase-equilibrium studies, (ii) estimates of heat production and temperature profiles, and (iii) bombardment history of the Moon.

PHASE-EQUILIBRIUM CONSTRAINTS

The generation of magmas by partial melting, and the subsequent alteration and metamorphism, depend on many complex factors, and only the general factors are given here. Wyllie (1979, 1980*a*, *b*), Winkler (1974) and Ringwood (1975) give details.

Figures 4–6 summarize important constraints for the outer 700 km of the Earth. Curves involving silicates can be displaced *ca.* 100 K in response to typical chemical substitutions found in terrestrial rocks and minerals, and by *ca.* 500 K for some extreme substitutions: thus substitution of Fe for Mg and of OH for F reduces stability fields of mica and amphibole, while substitution of Cr for Al favours spinel at the expense of garnet.

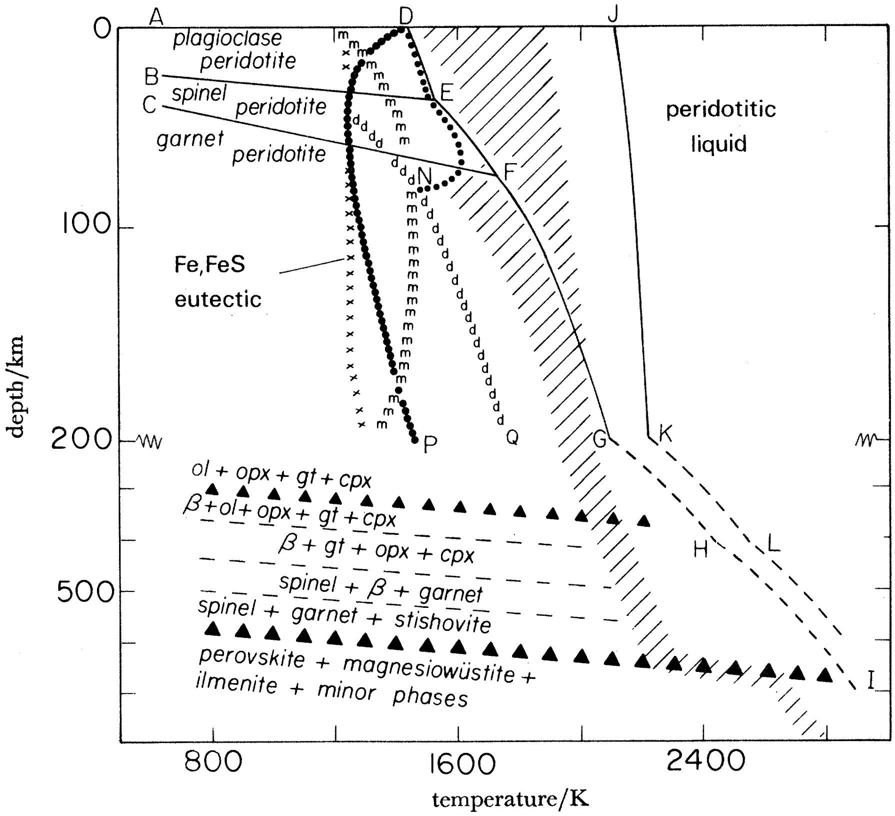

FIGURE 4. Schematic curves for phase equilibria of peridotitic compositions in outer 700 km. Mainly from Wyllie (1979) and Akaogi & Akimoto (1979). Melting of a lherzolitic composition begins at the solidus DEFGHI and is completed at the liquidus JKL. The liquid changes in general from basic to ultrabasic composition as the temperature increases. The kinks in the solidus and liquidus curves at G and K result merely from change of the depth scale at 200 km for convenience of drafting. Curves mmm and ddd show stability limits of mica and dolomite. DP and DENQ show beginning of melting of H_2O–peridotite and CO_2–peridotite. Shaded band is suggested temperature range for early Earth (figure 3).

Peridotite comprises coarse-grained rocks composed of olivine ($Mg_{1.8}Fe_{0.2}SiO_4$ approximately) and various amounts of orthopyroxene ($Mg_{0.9}Fe_{0.1}SiO_3$ approximately), garnet (usually rich in $Mg_3Al_2Si_3O_{12}$), clinopyroxene (rich in $CaMgSi_2O_6$), spinel ($(Mg,Fe)(Al,Cr)_2O_4$), plagioclase ($(Ca,Na)(Al,Si)_4O_8$), mica (rich in $KMg_3AlSi_3O_{10}OH$) and minor phases. If the silicate portion of the Earth were chemically homogeneous, plagioclase- and spinel-peridotites would occur respectively in areas ADEB and BEFC of figure 4. Garnet-peridotite would occur below CF, specifically as lherzolite composed of olivine, orthopyroxene, garnet and clinopyroxene.

Arid lherzolite (i.e. free of H_2O and CO_2) would begin to melt along DEFG. Garnet and clinopyroxene are consumed before orthopyroxene and olivine, yielding a liquid that changes from basic to ultrabasic composition as melting proceeds, until completion at JK to give a lherzolitic liquid. Peridotite composed only of olivine and orthopyroxene is called harzburgite, and is described as being depleted in basaltic component, or briefly barren. Lherzolite is fertile because it can yield basaltic liquid.

If the H_2O and CO_2 were not entirely in an atmosphere and hydrosphere, melting would begin at lower temperature (Wyllie 1979, 1980*a*, *b*). In the presence just of H_2O, potassium-free peridotite begins to melt along curve DP (touching dots), which has a minimum at *ca.* 1250 K and 2 GPa. Presence of potassium allows mica to be stable for temperatures below the curve

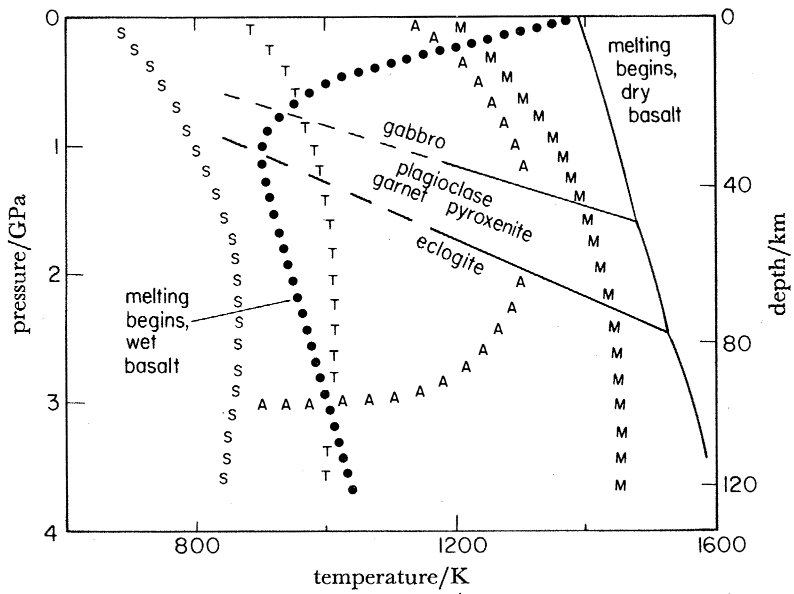

FIGURE 5. Phase relations for wet and dry basalt and upper stability limits of hydroxylated minerals; SSS, serpentine; AAA, amphibole; MMM, mica; TTT talc. From Green & Ringwood (1967). See Wyllie (1979) for many complications as the chemical composition of basalt changes.

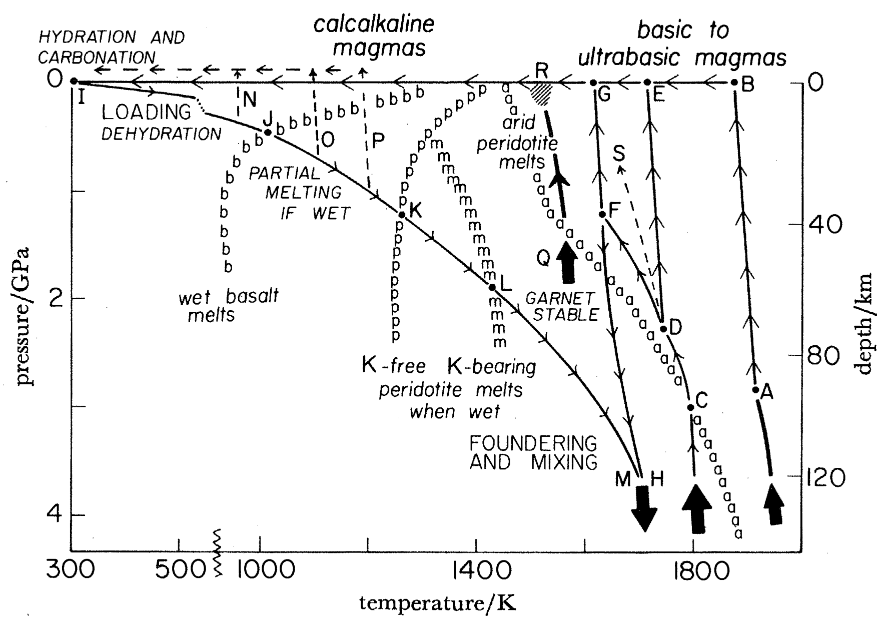

FIGURE 6. Scheme for: magma generation; loading, dehydration, and partial melting of the crust; foundering of heavy deep-crustal material, and mixing with depleted peridotite.

mmm, so long as H_2O/K is less than 2, thereby retaining OH in a solid phase and delaying the beginning of melting. For H_2O/K greater than 2, melting begins along DP. Appropriate amounts of Na, K and H_2O allow amphibole to be stable (Wyllie 1979, fig. 2). In the presence of CO_2, dolomite is stable in the triangular region of figure 4 bounded by curves ddd. At low pressure, the solubility of CO_2 (unlike H_2O) is so little that beginning of melting of CO_2-bearing peridotite closely follows that of arid peridotite (DE), and the CO_2 remains in the vapour phase. With increasing pressure, CO_2 enters liquid, and melting follows curve EN. At N, dolomite becomes stable, and at higher pressure melting follows curve NQ. In the presence of both H_2O and CO_2, complex relationships occur at pressures above N (Eggler 1978; Wyllie 1979, 1980*a*, *b*).

For simplicity, assume that during accretion of the Earth at least most H_2O and CO_2 was in the atmosphere, hydrosphere and upper part of an unstable crust. This allows most discussion of the mantle to be based on the properties of arid peridotite. With increasing pressure, all silicate minerals of garnet lherzolite transform to denser phases. Details are controversial, but the essential features (Akaogi & Akimoto 1979; Liu 1979; Mao *et al.* 1979) appear certain. Olivine begins to transform to the β-polymorph (figure 4) at about 400 km depth for the present Earth, and orthopyroxene breaks down at slightly greater depth. Clinopyroxene is absorbed by garnet with increasing pressure, and disappears completely at 500 km depth. The spinel polymorph of $Mg_{1.8}Fe_{0.2}SiO_4$ begins to replace the β-polymorph at 500 km depth, and the β-phase is gone at 600 km, where some stishovite (SiO_2 with rutile structure) appears. The major discontinuity of density at *ca.* 670 km depth is attributed to a change of bulk chemical composition together with a structural change to perovskite, $Mg_{1-y}Fe_ySiO_3$, magnesiowüstite $Mg_{1-z}Fe_zO_{1+x}$ (i.e. periclase structure type with metal vacancies), perhaps an Al-bearing ilmenite structure, and various minor phases. The density jump from 4.1 to 4.4 g/cm^3 at *ca.* 670 km is the linch pin of the concept of a thermal boundary zone separating convection systems in the upper and lower mantles (figures 2, 3).

Assume now that the Earth's crust was generated mainly by partial melting of the upper mantle, and that the lower mantle has stayed largely inert since accretion ended. The Fe–FeS eutectic at *ca.* 1250 K provides an approximate control on segregation of sulphide-rich liquid, which should sink through a silicate mantle to augment a core (figure 1*a*); see Wendlandt & Heubner (1979) for melting relations in the system Fe–S–O. Partial melting of arid peridotites in the upper mantle will produce a variety of magmas ranging from basic to ultrabasic composition, depending on the increasing proportion of melt between the solidus DEFGHI and liquidus JKL (figure 4) and the depth of melting. The basic to ultrabasic melts will be cooled at the surface, and both hydration and carbonation should occur (as well as other alteration reactions ignored here). Early lavas will be buried by later ones, and as loading occurs the crust will founder (figure 6) and undergo complex mineralogical transformations according to constraints discussed by Wyllie (1979).

A basalt that escaped hydration could enter the fields (figure 5) of gabbro, plagioclase–garnet–pyroxenite and quartz–eclogite with increasing pressure and temperature (Green & Ringwood 1967), and melting could begin at a temperature somewhat lower than for arid peridotite. Replacement of plagioclase (density *ca.* 2.7 g/cm^3) by garnet (density 3.6–4.0 g/cm^3) as the major Al-bearing mineral increases the density from gabbro (*ca.* 3.0 g/cm^3) to plagioclase–garnet–pyroxenite to quartz–eclogite (*ca.* 3.4–3.5 g/cm^3), and ultimately the absorption of pyroxene into garnet at *ca.* 400 km depth gives the rock garnetite, with density near 3.6 g/cm^3.

(These densities are adjusted to zero pressure for comparison.) Undepleted peridotite has a density near 3.40 g/cm^3 and loss of two-thirds of the CaO and Al_2O_3 to basaltic magma reduces the density to *ca.* 3.35 g/cm^3 (Jordan 1978). Thus gabbro should float on all peridotites, and eclogite should sink. Anderson (1979) proposed that sinking eclogite progressively fills the middle–upper mantle (220–410 km depth) and garnetite fills the bottom–upper mantle (410–690 km) as depleted peridotite is forced upwards into the top–upper mantle (40–220 km). This suggestion is not adopted in the present scheme, and it is assumed that most sinking eclogite is mechanically mixed with depleted peridotite to generate less-depleted peridotite in the middle and bottom zones of the upper mantle.

An ultramafic lava that escaped alteration could enter the fields of plagioclase-, spinel- and garnet-peridotites during foundering back into the mantle, and it might enter the region o partial melting (figure 4). If ultramafic rocks became slightly hydrated, amphibole and/or mica might be produced within the fields outlined in figure 5; the maximum stability limit is shown, and most micas and amphiboles are stable only to lower temperature and pressure. If carbonated but not hydrated the field of dolomite-rich carbonate (figure 4) is relevant. Hydration of olivine would produce serpentinization, and foundering would result in loss of water as serpentine transformed to talc and talc in turn became unstable. Sediments might be intercalated between volcanic rocks. There are innumerable subtleties as increasing metamorphism moves rocks through the greenschist, amphibolite and eclogite facies of metamorphism.

Basic rocks that became hydrated could undergo complex reactions during foundering into the mantle, as summarized in Wyllie (1979). One possibility is that wet basalt could produce the metamorphic rock amphibolite, and melting could occur at only 900 K for 1 GPa (figure 5). For the high radiogenic heat production expected in the early crust, it is inconceivable that melting of rocks with basic composition would not occur; indeed, a wet basaltic composition should begin to melt at only 1000 K for 20 km depth and 900 K for 40 km. Melting of basaltic composition if continued to higher temperatures could produce a range of generally calc-alkaline magmas, which could crystallize at depth to a suite of K-poor granites, and near the surface to volcanic rocks of intermediate composition. K-poor granites are represented abundantly in Archaean high-grade terrains by tonalite and trondhjemite gneisses, and in recent times by batholiths in active continental margins. In the present Earth, K-rich granites are found mainly in continental regions, and are attributed to a variety of processes, including partial melting of sediments and differentiation of mantle-derived liquids. K-rich granites occur in Archaean terrains (Windley 1977), typically as late intrusive bodies, and should be expected to occur as a minor component in pre-Archaean continental crust.

This simplified account of phase equilibria is consistent with (i) primary generation of basic to ultrabasic magmas by partial melting of a peridotitic upper mantle, and (ii) secondary generation of calc-alkaline magmas in a foundering crust by partial melting of the primary magmas modified by chemical alteration and metamorphism. The degree of depletion of the upper mantle depends on the extent of partial melting and on the amount returned from the crust.

It is now necessary to look more closely at the generation of primary magmas, because latent heat of melting is a major item in the heat budget. Furthermore the chemical nature of the magma depends on the depth of melting and the nature of its ascent: thus, in the present Earth, primary magmas can be very crudely split into tholeiitic basalts, generated abundantly in

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near-surface magma chambers under mid-ocean ridges, and alkali-rich volcanic rocks of many types, generated from deep-seated 'hot spots' under both oceans and continents. Of course, hybridism and contamination cause complications, and the source regions also vary.

Consider an undepleted peridotite rising adiabatically (*ca.* 0.3 K/km) through the upper mantle to C, where melting begins (figure 6). A curved path CD is followed as specific heat (*ca.* 1.2 J/gK) is converted into latent heat of melting (*ca.* 120 J/g). Other heat sources, including radiogenic heat, are trivial for rapid ascent in an early Earth. Light magma is gravitationally unstable with respect to heavy crystalline residue, and magma might escape after only a small degree of partial melting (Walker *et al.* 1978), especially during accretion, when the Earth surface was shaken by huge projectiles. Segregation of magma from crystal–magma mush is favoured by a high concentration of melt, narrowness of the conduit, large grain size, and a high ratio of viscosity of grain to melt (Sleep 1974). If a magma is free of crystals, adiabatic ascent (DE) merely causes super-heating. If a magma retains some crystals in suspension, further melting can occur along a non-unique, curved path (e.g. DS) as specific heat is converted into latent heat. Crystalline residues after escape of magma are depleted in basaltic component, and become more refractory (not shown in figure 6). For a curved path DF, a second batch of melt may escape along FG. Details of magma composition are complex (Wyllie 1971, ch. 8), but a range of basic to ultrabasic compositions is expected in the early Earth. In general, initial melts tend to become more ultrabasic as the pressure increases, and successive partial melts tend to become less basaltic. The hotter the upper mantle the greater the depth at which melting begins during adiabatic uprise, and the greater is the tendency for production of ultrabasic magma by single-stage melting. Because the Earth is assumed to have been cooled during accretion by volcanic activity, melting is believed to have occurred mainly in the upper 100 km after accretion ended. To produce an ultrabasic magma by single-stage melting at shallow depth (< 100 km), liquid must escape from a position well above the beginning of melting of undepleted peridotite (e.g. A, figure 6). If magma escapes after only a small degree of partial melting, multi-stage melting with sequential production of basic to ultrabasic magmas should occur (Arndt 1977). The single-stage model for generation of ultrabasic magmas (Green 1975) is misleading because of neglect of the heat budget.

Komatiitic rocks (i.e. basic to ultrabasic rocks with textural evidence of quenching from a liquid) are a distinctive feature of Archaean greenstone belts (e.g. base of Barberton belt (Windley 1977, p. 25)), and are increasingly rare in younger volcanic sequences. It will be assumed that they result mostly from multi-stage melting, at less than 100 km, of peridotite rising adiabatically from deeper than 100 km, as shown schematically by open arrows in figure 3; however, the cyclic nature of the komatiites requires a complex source mechanism.

Returning to the present day, mid-ocean-ridge basalts are the most abundant volcanic rocks, and are attributed to near-surface melting of depleted peridotite rising towards the surface (figure 2). This process is represented schematically by rise of peridotite to Q (figure 6) and retention of magma during non-adiabatic ascent until the shaded region R is reached; thereupon, fractional crystallization produces tholeiitic basalts and crystalline residues (gabbros to dunites), which form an oceanic plate riding on a raft of further-depleted peridotite in the conventional plate-tectonic model.

Particularly important is continued replenishment of magma chambers under mid-ocean ridges by convective ascent of the mantle (figure 2) in contrast to local adiabatic uprise and episodic escape of magmas from deep-seated regions (open arrows in figures 2 and 3). This distinction is perhaps seen more clearly in figure 8, which will be discussed in detail later. Details of magma types in planets are being summarized in the Basaltic Volcanism Project, and general features are discussed by Walker *et al.* (1979).

Estimates of heat production

The literature on heat production in the Earth is extremely confusing because of lack of fundamental constraints on the accretion process and subsequent crystal–liquid differentiation. Nevertheless it seems certain from a combination of geochemical and geophysical arguments that a key to speculation on the development of the Earth's crust is transfer of internal heat to the surface by volcanic activity consequent upon convection in the mantle, and cooling of the mantle by foundering of the crust.

Assume that much of the gravitational energy from accretion and core formation was transferred to the surface by impacts and intense convection over 10^8 a. Each gram of liquid that cools by *ca.* 1200 K to the temperature of the Earth's surface (assumed < 373 K) loses *ca.* 120 J of latent heat and *ca.* 1500 J of specific heat. This total of about 1600 J compares with the average energy, throughout the whole Earth of *ca.* 3000 J/g from core formation (Shaw 1978), a similar value for accretional heating (Kaula 1979*a*), and *ca.* 2500 J/g to heat accreting material from an assumed arrival temperature of 1000 K to the assumed mean temperature of *ca.* 3000 K for the Earth at the end of accretion. Hence only about two episodes of cooling from melting temperature to 373 K are needed on average to dissipate the *ca.* 3500 J/g remaining from the accretional and core-formation heats. Radiogenic heat is relatively unimportant during the period of accretion, especially as partial melting should transfer much of the radiogenic elements to the outer 100 km.

Subtle complications may be of considerable geochemical importance. Partial melting during convective overturning is limited for arid peridotite by a critical adiabat with a slope near 0.3 K/km, which meets the curve for the beginning of melting at zero pressure (D, figure 4). Cool material foundering from the surface can mix with deep-seated material to produce a region below the critical adiabat. The residue from the melting of volatile-bearing peridotite can be below the critical adiabat for arid peridotite. Accretional heat should be concentrated in the outer 100 km, and should provide little heating of deep-seated regions. Sinking Fe-rich diapirs may deposit heat mainly in the core and only slightly in the silicate mantle. The depth of high-pressure inversions in crystal structures must change during accretion, along with density barriers to convection. All these factors need thorough exploration, particularly with respect to a model in which the lower mantle remains too cool for partial melting after the initial stage of partial melting during accretion. Such a model would be consistent with calculations that only one-half to one-third of the mantle has been involved in partial melting to produce the distribution of radioactive elements in the present crust and mantle (O'Nions *et al.* 1979; Jacobsen & Wasserburg 1979). Lambert (1980*b*) suggested that lower mantle convection started in the Lower Proterozoic, and that silicic differentiate from melting of the lower mantle enhanced the rate of continent formation at *ca.* 2.8 Ga B.P. This interesting suggestion is not accepted here, and heat is assumed to be transported through the lower mantle to the boundary with the upper mantle by solid-state convection without much melting and with only trivial loss of fugitive constituents (e.g. rare gases trapped inside mineral grains).

Turn now to radiogenic heat. The key factor is the recent removal of the incorrect constraint

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that surface heat flow equals radiogenic heat production. From various cosmochemical and geochemical arguments (see, for example: Taylor 1979; O'Nions *et al.* 1979), it appears that the mass fraction of U in the whole Earth lies within about $14–20 \times 10^{-9}$. Observed K/U ratios for various near-surface rocks ($0.7–2.1 \times 10^4$; O'Nions *et al.* 1978), if pertinent to subcrustal reservoirs as well, would imply about 200×10^{-6} K in the whole Earth, but various considerations (Smith 1977, pp. 354–355; O'Nions *et al.* 1979) might favour *ca.* 150×10^{-6} K. With the Wasserburg ratio of 3.8 for Th/U, the whole Earth should contain $57–76 \times 10^{-9}$ Th. In accordance with Lambert (1977), radiogenic heat should have decreased from *ca* 1.1×10^{14} W at 4.45 Ga B.P. to *ca.* 2.4×10^{13} W now (both $\pm$*ca.* 20%). These values are about half those given by O'Nions *et al.* (1978, fig. 3) for an early high estimate of U content.

To explain the present observed heat flow of 4.1×10^{13} W (Williams & von Herzen 1975), assuming of course that it is accurate, it is necessary to propose that there are other heat sources, or that transfer of radiogenic heat to the surface is delayed by sluggish convection (Daly & Richter 1978; O'Nions *et al.* 1979). Reduction of the mean temperature of the Earth (cf. Sleep 1979) by (say) 0.5×10^{-7} K/a (corresponding to 225 K over the past 4.5 Ga for a linear decrease) would yield 2.0×10^{13} W during convective and conductive transfer. Estimates of *ca.* 2.5×10^{12} W for crystallization in the core (Gubbins *et al.* 1979) and 5×10^{12} W for present heat flow from the core (Schubert *et al.* 1979) are poorly constrained. Although some delay in transfer of heat must occur, it is assumed that deep-seated, non-radiogenic heat sources constitute a substantial fraction of the present surface heat flow.

As one looks back to the early Earth, short-lived radioactivity may be ignored after 4.5 Ga B.P. (O'Nions *et al.* 1978, fig. 3), and impact heating should be trivial for the whole Earth after 4.4 Ga B.P., but should be locally important until 3.9 Ga B.P. in near-surface regions (next section). If contributions from non-radiogenic sources to the whole Earth remained constant, the total heat flow would increase about threefold from 4.1×10^{13} W to *ca.* 1.3×10^{14} W at 4.45 Ga B.P. This factor will be too low if convection were more effective in lowering the Earth's temperature in early times, but it may be difficult to find models that increase the factor higher than (say) fivefold.

It is now useful to follow Lambert (1980*a*) in comparing these estimates of the total heat flow at the Earth's surface with the observation that all areas of the present Earth with a heat flow of > 100 mW/m² display volcanism and hydrothermal action. The threefold increase of heat flow to *ca.* 1.3×10^{14} W corresponds to 250 mW/m², thereby implying volcanic activity over a substantial fraction of the Earth: indeed the present heat flow in Iceland is *ca.* 200 mW/m². Because production of radiogenic heat dropped only to one-half during the first 1.5 Ga of Earth history, it seems safe to conclude that intense volcanism occurred before survival of crust. Lambert (1980*a*) suggested that episodic formation of crust from mantle is the result of uneven transfer of heat, but this factor is probably unimportant in discussion of the first 750 Ma of Earth history.

Bombardment history of the Moon and implications for the Earth

The impact stratigraphy of the Moon (Baldwin 1963) was largely confirmed by numerous post-Apollo studies reviewed by Taylor (1975), but unfortunately there is uncertainty in the temporal variation of impacting bodies before 4 Ga B.P. There are about 40 recognizable basins ranging from younger ones of Imbrian age to older ones of Nectarian age (Howard *et al.* 1974,

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fig. 14). The younger basins are near-circular, are multi-ringed, are partly filled with mare basalt, and have a mass concentration indicative of non-isostatic adjustment of underlying crust and mantle. Older basins have irregular margins and lack mass concentrations, thereby indicating isostatic adjustment. As an example, the young Smythii basin displays three rings, of diameter 370, 640 and 850 km, with a relief of 8 km (Strain & El-Baz 1979). Its present basin volume of 2×10^6 km^3 must be less than the volume before mantle uplift and volcanic eruption into the impact crater. Calculations extrapolated from experimental craters suggest that a basin of radius 400 km would have been surrounded by a complex blanket of primary and secondary ejecta ranging from about 2 km thick at the rim to about 10^2 m thick at 1000 km from the impact centre (McGetchin *et al.* 1973). All Apollo sites should have been covered by ejecta blankets from many basins, adding up to a total of at least 10^2–10^3 m thickness. Interpretation of radiometric age data is difficult, and is complicated by the effects of numerous impacts smaller than those responsible for basins.

Crystallization ages of mare basalts date back to 3.9 Ga B.P. (Nyquist 1977; Turner 1977), and all basins must be older. A dunite with an age of *ca.* 4.6 Ga apparently survived the era of massive bombardment, but most rocks attributed to the lunar highlands show ages of 3.9 to 4.0 Ga. This clustering was attributed to a terminal cataclysm (Tera *et al.* 1974). Some lunar scientists speculated that the observable lunar basins are merely the last survivors of a huge declining population, and that the age clustering merely reflects metamorphic resetting in hot impact debris. For the Apollo 16 site, ^{39}Ar–^{40}Ar plateau ages for 43 critically selected samples (Maurer *et al.* 1978) show a cluster at 3.9 to 4.0 Ga and a weaker ill defined cluster (or tail?) at 4.1 to 4.2 Ga. On the basis of the impact model of McGetchin *et al.* (1973), the Apollo 16 region should have obtained about half of its debris from the Nectaris basin and only one-fifth from younger basins of Imbrian age: however, local cratering can change these amounts at a specific location of small area. Turner's (1979) model of multiple reworking of impact debris allows the possibility that the age cluster at 3.9 to 4.0 Ga results from a spike in the impact resetting rate. Interpretation of Apollo 16 ages as the genuine products of impacts in basins of two distinct ages would yield a time difference of *ca.* 0.2 Ga. This would allow for substantial thickening of the lunar lithosphere, as is needed to explain the tectonic difference between the irregular and ringed maria. Nyquist (1977), from a review of Rb–Sr ages, concluded that the Imbrium and Serenitatis basins (the latter preceding Nectaris stratigraphically) were formed at 3.9 and 4.0 Ga B.P. respectively, but it might be wise to hold open the possibility that the Nectaris basin dates back to 4.1 to 4.2 Ga B.P. Hertogen *et al.* (1977) attributed over half of the Apollo 16 breccias to chemical groups assigned to Nectaris and to unknown pre-Serenitatis basins. Currently it seems impossible to obtain a definitive answer, and, for the present purpose, the only reliable conclusion is that the Moon displays at least 40 impact basins, and that their age distribution is uncertain, but must have spanned enough time to allow a lithosphere to thicken.

The next stage in the argument is to ask whether the Earth was being bombarded by the same population of basin-forming projectiles as was the Moon. If lunar basins were produced by capture of moonlets as the Moon receded, the Earth might have collected only debris that escaped the Moon. If the Moon was not in Earth orbit when the lunar basins were formed, the Earth need not have undergone the same bombardment. Fortunately it seems safe to attribute the lunar basins to projectiles coming from outside the Earth–Moon system, and to deduce that the Earth and Moon were simultaneously undergoing bombardment. Mercury and Mars, and

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perhaps Venus, contain impact basins. The asteroid belt does not contain enough mass to explain the relative velocities of the asteroids, and the giant planets should have perturbed bodies into the inner Solar System. The lunar basins have a random spatial distribution indicative of bodies not in Earth orbit. Capture of the Moon by the Earth is theoretically estimated to be so catastrophic that extensive destruction of lunar surface morphology would have occurred: hence capture is not possible after formation of basins and early craters.

The next stage of the argument uses Safronov-type theory of accretion of planetesimals to provide plausible extrapolation backwards in time. A mean relative velocity of *ca.* 8 km/s is expected for the last large bodies surviving after accretion of planets in the inner Solar System, perhaps with an uncertainty of ± 2 km/s depending on the mechanics of collision processes. Monte-Carlo calculations (Wetherill 1977) yield a flux of Earth-impacting bodies that decays with a half life of 17 Ma at 4.45 Ga B.P. to 190 Ma at 4.0 Ga B.P. to 900 Ma at 1.5 Ga B.P. The estimated value of the initial half life decreases as the mean relative velocity is reduced. The long-lived tail results from orbital capture and subsequent loss by Mars, and the bodies would suffer increasing risk of destruction by impacting with other bodies. If 40 basins were formed on the Moon in a period of *ca.* 200 Ma from 3.95 to 4.15 Ga B.P., Wetherill's calculation implies that about 20 times as many basins were formed in the preceding 300 Ma from 4.45 to 4.15 Ga B.P. If this implication is accepted, most lunar basins were obliterated by volcanic activity and by burial from later impact debris. Another possibility is that the inner solar system was bombarded by planetesimals sling-shot by Jupiter. Again it is necessary that the impacts would decrease with time, and it is assumed that Wetherill's calculations provide a qualitative guide even though the starting assumptions must be quite different.

The final stage of the argument involves estimate of the relative impact density of the Earth and Moon. For incoming bodies of infinite velocity, the impact density per area would be identical and stochastically isotropic. For slow bodies, the gravitational attraction of the Earth, with its escape velocity of *ca.* 11 km/s, acts as a convex lens, and the Earth should have a higher density than the Moon. However, the Moon should be sufficiently close to the Earth that it receives some of the focusing effect. Consequently, for incoming bodies with a pre-attraction relative velocity of *ca.* 8 km/s it is assumed that the impact density on the Earth is not greater than twice that on the Moon; hence the Earth is assumed to be bombarded by 13 to 27 times as many large projectiles as the Moon.

To summarize: at a minimum, *ca.* 500–1000 basins were formed on Earth in a period of *ca.* 100–200 Ma before 3.95 Ga B.P. Furthermore, plausible calculations suggest that this was merely the tail of a declining flux and that *ca.* 20 times as many basins were formed in the preceding 300 Ma. To put it another way, the minimum value corresponds to most areas of the Earth's surface lying within 1000 km of a basin centre, while the greater value corresponds to multiple overlap of basin positions.

Nature of the early crust

General statement

Based on the arguments already presented, the Earth at 4.45 Ga B.P. is envisaged (figure 3) to have a multi-level convection system with the upper mantle and crust nearly isolated from the lower mantle. The temperature profile is about 200 K higher than that proposed for the present Earth and convective uprise is causing magmatism in the upper mantle. A large fraction

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of the magmatophile elements are in the upper crust. The temperature at the surface is controlled largely by incoming and outgoing radiation, and most of the water is in a hydrosphere about 2 km thick while most of the CO_2 has entered chemical sediments. The crust varies in thickness up to some tens of kilometres, and is undergoing strong volcanic activity, metamorphism, sedimentation and tectonic deformation. Basin-forming impacts are frequent and cause local destruction of the crust. The principal geochemical events (figure 6) are (i) convective uprise in the upper mantle, with generation of rising basic to ultrabasic magmas (e.g. C–G; QR) and sinking residue H, (ii) cooling, hydration and carbonation at the surface (I), coupled with intercalation with sediments, (iii) dehydration and decarbonation (N) of mixed materials sinking under a load of new

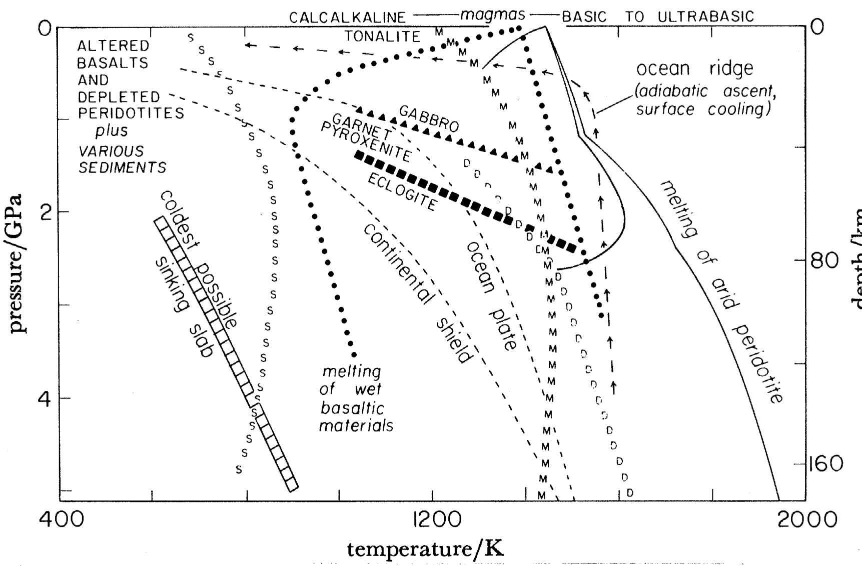

FIGURE 7. Relation between pressure–temperature curves, phase relations and rock types of present Earth in which asymmetrical plate tectonics occurs.

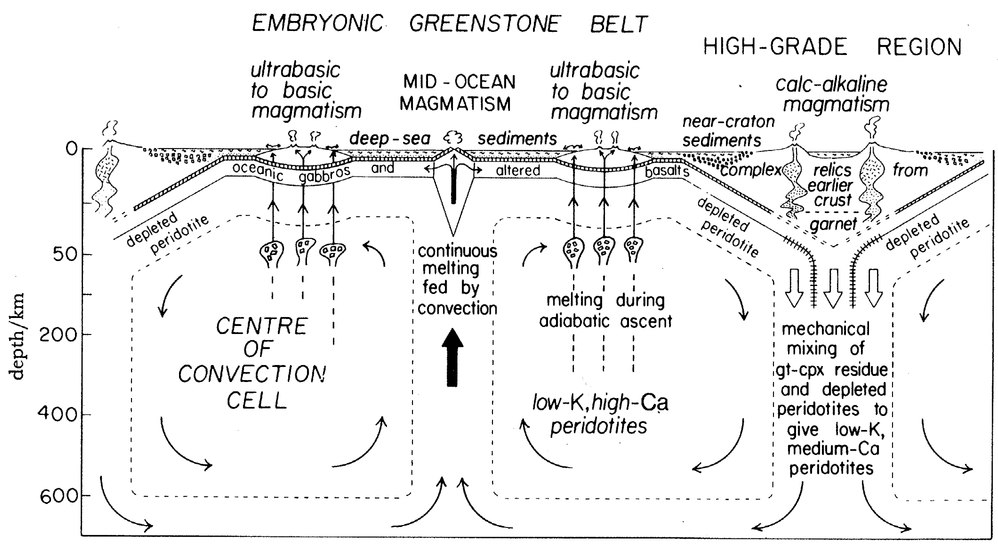

FIGURE 8. Diagram of convection and rock types for a do-it-yourself model of the pre-Archaean and early Archaean eras. It must be emphasized that the distinction between continuous melting fed by convection and melting during adiabatic ascent cannot be clear-cut, especially in the chaotic conditions during and just after cessation of accretion.

materials, (iv) partial melting of basic to ultrabasic compositions (J and K) to produce calc-alkaline magmas (e.g. O and P), which rise back to the surface, leaving a heterogeneous residue M, and (v) mixing of residues H and M to produce depleted peridotite. These geochemical events are qualitatively similar to those in the present Earth, but there are important chemical and tectonic differences.

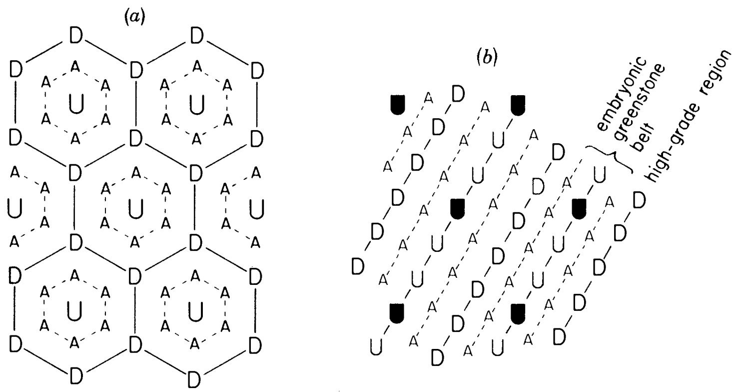

FIGURE 9. Idealized models of (*a*) polygonal and (*b*) symmetrical linear tectonics. Euler's relation will not allow hexagonal cells to cover a sphere. It is not possible to move continuously from polygonal to linear tectonics: thus only filled-in U symbols can retain position. A clear distinction between U and A regions cannot be expected.

To provide a basis for discussion, figure 7 illustrates relevant features of the volcanism–subduction cycle in a modern plate tectonics model, and figures 8 and 9 show a geochemical and tectonic model for the early Earth. In the present Earth, most volcanic activity is concentrated into local regions mainly between stable plates. In oceanic regions, adiabatic ascent at a mid-ocean ridge produces near-continuous melting in near-surface magma chambers and the resulting cooled plates of partly altered basalts, sediments and underlying cumulate rocks including gabbros and dunites ride on a conveyor belt of depleted peridotite overlying less-depleted peridotite. Continental shields have a thick base of barren metamorphic rocks, and their temperature increases more slowly with pressure at first than does that of oceanic plates, though ultimately below 200 km the two geotherms may become asymptotic to a curve pq (figure 2) close to the critical adiabat for melting of arid peridotite. As an oceanic plate slides under a continental margin, complex reactions must occur as dehydration, decarbonation and perhaps partial melting occur. Unfortunately there is great uncertainty in the temperature profile of sinking slabs, depending on assumptions about velocity, thickness, dehydration, etc., but examination of calculations in Anderson *et al.* (1978) and earlier papers suggests that a sinking slab cannot be cooler than the limit shown in figure 7 and it certainly must be cooler than an oceanic plate. Although dehydration must occur in sinking slabs, it is quite uncertain whether melting occurs in the slab and what are the causes of island-arc volcanism (see references in Saunders *et al.* (1980)). In the early Earth, it is proposed that convection in the upper mantle (figure 8) caused magmatism ranging from near-surface differentiation above regions of rapid uprise to deep-seated partial melting above the slow-moving core of a convection cell. The former produced tholeiitic-type basalts lying over cumulate rocks including gabbros and dunites, all overlying a thick layer of depleted peridotite. The latter produced volcanic constructs composed of basic to ultrabasic rocks intercalated with chemical sediments precipitated in deep water. This type of magmatism would be qualitatively similar to modern 'hot-spot' volcanism, but adiabatic uprise and partial melting would have been easier because of higher temperature and radiogenic heating than occur now. Above down-welling regions of the mantle, complex relics from earlier crust formed protocontinents. Calc-alkaline magmas were produced partly from the protocontinent and partly from the slab of oceanic rocks transported as on a conveyor belt. For various reasons, the slab heated up faster than does a modern slab and was less rigid, and became mechanically intermixed with near-craton sediments and relics of earlier crust. This mechanical mixing is of crucial importance in geochemical models because it would allow return of material from continental crust back into the mantle to be recycled to produce new oceanic crust. Undoubtedly there were tremendous temperature variations, especially near impact basins, but perhaps an average pressure–temperature curve for an early sinking slab was near IJKLM in figure 6. Part of the protocontinent and slab entered the stability field of garnet, and foundering occurred with ultimate mixing of garnet–clinopyroxene residue and depleted peridotite to give low-K, medium-Ca peridotites. For the period 4.5 to 4.0 Ga B.P., continuous foundering of crust into mantle is envisaged, assisted locally by basin-forming impacts especially for the first 100 Ma.

Although figure 8 bears considerable similarity to plate tectonic models for recent times, there would be major differences. In the early pre-Archaean, narrow convection cells (Fyfe 1978; Lambert 1980*a*) might be dominant, as shown idealized in figure 9. The term polygonal tectonics is used for an idealized version. A region of up-welling U is surrounded by a down-welling boundary with neighbouring cells, and protocontinents would form at triple-junctions of down-welling D. As deep-seated heat sources for convection in the upper mantle died down in the early Archaean era, a transition to symmetrical linear tectonics might occur before transition during the Proterozoic era to the asymmetrical linear tectonics of the present day. Figure 8 might be used as a model for development of greenstone belts and high-grade regions in the early Archaean (*ca.* 3.5 Ga B.P.), as will be discussed in detail elsewhere. The volcanic constructs of ultrabasic-to-basic rocks with associated deep-sea sediments would be compressed against high-grade regions by convective motion of underlying mantle. Overthrusting would lead to thickening. Underlying basalts and gabbros would melt to give intermediate calc-alkaline magmas, which would crystallize into uprising tonalite plutons. There would be no simple pattern, and there might be surviving relics from regions with predominantly polygonal tectonics (e.g. Rhodesia; Windley 1977, fig. 2.4) and regions with predominantly symmetrical-linear tectonics (e.g. Abitibi orogen; Windley 1977, fig. 3.5). Convection cells would be narrow (*ca.* 500 km across; Lambert 1980*a*) and protocontinents would be small and would not display the abundant K-rich granites and thick sedimentary piles found in modern stable continents.

Magma ocean and anorthositic crust

Although the concepts of a Moon-wide magma ocean together with flotation of Ca-rich plagioclase to form an anorthositic crust have almost become dogmas, there are puzzling petrological and heat-budget problems that require resolution. Furthermore, there is no justification for carrying over these concepts to the Earth without a thorough analysis of all factors.

Certainly there is no evidence in Archaean rocks for a worldwide anorthositic crust. All early anorthosites occur as small bodies or layers in complexes or terrains dominated by basic rocks. The origin of young anorthosites is controversial, but they certainly formed deep in the crust, probably by crystal–liquid fractionation. The early Archaean crust of the Earth was composed mainly of a complex assemblage of basic to ultrabasic rocks undergoing various degrees of metamorphism and partial melting (Windley 1977). Both mechanical and chemical sediments occur. The pre-Archaean crustal rocks were probably similar to those found in west Greenland and other early Archaean terrains, except that impact-metamorphic rocks would be abundant during the early phase of basin formation, and ultramafic magmas should have been favoured by higher mantle temperature.

Whether there was a worldwide magma ocean at any stage of accretion is difficult to predict. Near-surface volcanism with rapid cooling of lavas in water could dissipate heat very rapidly. Impacts into water-bearing rocks, and especially into a hydrous ocean, would produce huge clouds of volatile-rich debris, which should cool more rapidly than clouds of dry debris on the Moon.

Effects of basin formation

There is no record of a basin-forming impact in Archaean rocks from North America (Weiblen & Schulz 1978). The suggestion that greenstone belts result from impacts (Green 1972) is not tenable because sediments are intercalated with volcanic rocks (Glikson 1976) and greenstone belts are considerably younger than lunar basins. Furthermore, ultramafic magmas can be generated by normal processes of melting in the mantle. Although the formation of impact basins should have major effects on magma generation and tectonic processes, convection in the mantle should be sufficiently erratic that there is no long-term memory of the positions of at least most basins. There are only a dozen or so major plates in the present Earth, and there are only a few Archaean greenstone belts and high-grade regions. If there were 1000–20000 basins, only the last dozen or so could be considered as the cause of location of crustal units. Detailed study is needed of the suggestions that impact basins were the primary causes of either continents or oceans, depending on the author (Goodwin 1976; Frey 1980; see also Grieve 1980).

As yet there are no reliable calculations on the petrological effects of basin impact in the pre-Archaean. Green's (1972) model must be modified to include the conversion of specific heat into latent heat of melting, thereby greatly reducing the amount of impact-induced melting under a basin. Furthermore, water from an ocean and from hydrous crustal rocks would serve to cool the impact debris, thus giving the effect of a wet blanket.

Unless convincing arguments are produced to the contrary, it will be assumed that impacts have only temporary local effect on the crust after 4.0 Ga B.P., and that volcanism driven by internal heat is the prime cause of destruction of crust. This volcanism is ultimately responsible for metamorphism and sedimentation, all of which are interrelated in the cycle of figure 8.

Establishment of geochemical reservoirs

The present scheme, as idealized in figure 8, can be tested by measuring the distribution of radioactive parents and daughters in early Archaean rocks, by means of model calculations like those in O'Nions *et al.* (1979) and Jacobsen & Wasserburg (1979). In particular, if crust were formed continuously and destroyed continuously from 4.45 to 3.8 Ga B.P., the earliest surviving volcanic rocks should have a radiogenic signature of a source region depleted by the amount in the transitory crust. The concept of recycling of pre-Archaean crust is not in conflict with the well established concept of irreversible formation of crust in the Archaean and later eras (Moorbath 1977). Perhaps the major problems are the uncertainties in mechanical factors involved in the foundering of crust and subsequent mixing with depleted peridotite to give less-depleted peridotite, which then becomes the source for further volcanism. Particularly important will be identification of the mineral hosts of key elements in geochemical models. Perhaps the difference in K/Rb for basalts (mid-ocean-ridge basalts *ca.* 1000; alkali-rich rocks *ca.* 200–400) will be traced in part back to partition between mica (K/Rb ≈ 300) and pyroxene and amphibole (n00–m000) as material founders into the mantle, where pyroxene remains stable and mica does not (Smith *et al.* 1979). The mantle array of Sr and Nd may result from the different stabilities of the probable mineral hosts apatite, clinopyroxene, garnet and dolomite. These factors are not peculiar to the pre-Archaean era, and will be discussed elsewhere.

Conclusions

From a combination of physical, chemical and petrological arguments, it is concluded that the Earth's surface underwent intense volcanism in the pre-Archaean era, and that the rock types were chemically similar to those found in the early Archaean era. This volcanism was driven partly by upper-mantle convection and radiogenic heating in the crust, and locally was affected by impacts. Sedimentation and metamorphism were important factors. 'Oceanic' and 'continental' regions developed above upwelling and downwelling segments of convection cells in the upper mantle, and these regions respectively resembled the greenstone belts and high-grade regions of the Archaean era. As the heat flow declined, polygonal tectonics would be replaced erratically by linear tectonics. With this model, survival of crust from 3.8 Ga B.P. is merely an accident, and discovery of even older crust would not be surprising. Basin-forming impacts would certainly have caused considerable destruction of crust up to 3.9 Ga B.P., but it seems preferable to explore the argument that the pre-Archaean crust would have foundered into the mantle anyway, by a complex combination of volcanic, metamorphic and sedimentary processes. Detailed study is needed of the implications of a multi-level convection system in the mantle: in particular, were all volatile elements (e.g. H_2O, CO_2, alkalis and halogens) swept out of the lower mantle during accretion, and did crystal–liquid processes involve high-pressure phases such as perovskite and magnesiowüstite as well as the peridotitic minerals? Because there is great uncertainty about the interaction between the crust and mantle even of the present Earth, it will not be easy to obtain a definitive conclusion about the pre-Archaean Earth. Perhaps Robert Graves in *Rocky acres* provides a fitting valedictory: 'Time has never journeyed to this lost land, . . . The first land that rose from Chaos and the Flood'.

Many discussions with A. T. Anderson, R. C. Newton, F. M. Richter, B. F. Windley and P. J. Wyllie were very helpful, but I have been forced to ignore some of their caveats. An extremely thorough review by P. J. Wyllie, and helpful comments by W. M. Kaula, R. St J. Lambert, L. O. Nicolaysen and R. K. O'Nions are highly appreciated. Apologies are expressed to all those whose papers could not be cited. Financial support from grants NASA 14-001-171 and NSF EAR77-27100 is gratefully acknowledged.

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