

Speculations on evolution of the terrestrial lithosphere–asthenosphere system—Plumes and plates

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Abstract

In the early Earth, accretionary impact heating, including collision with a large, Mars-sized object, decay of short-lived radioisotopes, and (after an initial thermal run-up) continuous segregation of the liquid Fe–Ni core resulted in extensive the melting of the silicate mantle and in the formation of a near-surface magma mush ocean. Progressive, continuous degassing and chemical–gravitational differentiation of the crust–mantle system accompanied this Hadean stage, and has gradually lessened during the subsequent cooling of the planet. Mantle and core overturn was vigorous in the Hadean Earth, reflecting deep-seated chemical heterogeneities and concentrations of primordial heat. Hot, bottom-up mantle convection, including voluminous plume ascent, efficiently rid the planet of much thermal energy, but gradually decreased in importance with the passage of time. Formation of lithospheric scum began when planetary surface temperatures fell below those of basalt and peridotite solidi. Thickening and broadening of lithospheric plates are inferred from the post-Hadean rock record. Developmental stages of mantle circulation included: (a) 4.5–4.4 Ga, early, chaotic magma ocean circulation involving an incipient or pre-plate regime; (b) 4.4–2.7 Ga, growth of small micro-oceanic and microcontinental platelets, all returned to the mantle prior to 4.0 Ga, but increasing in size and progressively suturing sialic crust-capped lithospheric amalgams at and near the surface over time; (c) 2.7–1.0 Ga, assembly of cratons surmounting larger, supercontinental plates; and (d) 1.0 Ga–present, modern, laminar-flowing asthenospheric cells capped by gigantic, Wilson-cycle lithospheric plates. Restriction of komatiitic lavas to the Archean, and of ophiolite complexes±alkaline igneous rocks, high-pressure and ultrahigh-pressure metamorphic terranes to progressively younger Proterozoic–Phanerozoic orogenic belts supports the idea that planetary thermal relaxation promoted the increasingly negative buoyancy of cooler oceanic lithosphere. The thickening of oceanic plates enhanced the gravitational instability and the consequent overturn of the outer Earth as cold, top-down oceanic mantle convection. The scales and dynamics of deep-seated asthenospheric circulation, and of lithospheric foundering+shallow asthenospheric return flow evidently have evolved gradually over geologic time in response to the progressive cooling of the Earth.

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1. Earth's origin in the solar system

Ours is the only planet in the solar system that now exhibits plate tectonic processes (Stevenson, 2003). U/Pb isotopic dating of lunar minerals and meteorites (Patterson, 1956; Papanastassiou and Wasserburg, 1971) indicates an age of ~4.55 Ga. Theoretical considerations suggest that the solar system condensed from a locally dense cloud of interstellar gas that spun up to form a rotating disc (Safronov, 1972; Cameron, 1978; Taylor,

1992) around the T Tauri proto-Sun (e.g., Hutchison et al., 2001). Accretionary sweep-up of planetesimals in the outer regions of the disc gave rise to the planetary bodies orbiting the Sun (Shoemaker, 1984). Although initial arriving condensates may have been cold, impact heating due to increasing mass and gravitational attraction would have resulted in a progressively hotter planet as it grew (Wetherill, 1976; Kaula, 1979). This primordial phase of planet formation likely was completed by ~4.55 Ga. The concentrations of long-lived radioactive isotopes (e.g., U, Th, ⁴⁰K) and the initial presence of short-lived isotopes (e.g., ²⁶Al) further heated the young, heterogeneous Earth (Lee, 1970). Because reduced iron–nickel alloys melt at lower

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temperatures than mantle silicates at elevated pressures, a dense iron-rich liquid would have separated and descended to the center of the hot Earth, coalescing as a molten Fe+Ni metal core overlain by a semi-solid Mg+Si-rich mantle (Elsasser, 1963; Stevenson, 1983). Continuous infall of the core released vast amounts of heat, causing even more complete segregation of the liquid core and the refractory Mg-rich silicate mantle (Birch, 1965; Flaser and Birch, 1973), and in partial fusion of the latter. Very early impact of a Mars-sized object would have supplied additional heat while ejecting mostly mantle material to a lunar orbit (Wetherill, 1988, 1990; Cameron and Benz, 1991; Halliday, 2004). Judging by petrochemical and isotopic data for the Moon and meteorites, the terrestrial magma mush ocean must have been generated before ~4.4 Ga (Boyett and Carlson, 2005; Kleine et al., 2005). It would have been a near-surface, several-hundred-km-thick layer because the Earth's adiabatic gradient has a larger dP/dT value than does the mantle solidus P – T curve (and geothermal gradient), promoting convection (Stolper et al., 1981; Herzberg, 1987; Abe, 1997; Davies, 1999).

By ~4.4 Ga, the magma ocean had essentially solidified, and at least very thin lithospheric platelets existed, as necessary to account for the western Australian 4.2–4.4 Ga detrital zircons derived from granitic rocks (Compston and Pigeon, 1986; Wilde et al., 2001; Watson and Harrison, 2005). This is so because the onset of melting for felsic bulk-rock compositions necessary for zircon growth lies at lower temperatures than the solidi of basalts and peridotites (see Fig. 1). The oldest preserved terrestrial rocks are the 4.03 Ga granitic orthogneisses from NW Canada (Bowring and Williams, 1999). Hence for a sialic protocrust to have crystallized, shallow levels of oceanic

crust and uppermost mantle material must have been well below their fusion temperatures. The presence of water-laid metasedimentary rocks in western Greenland dictates that portions of the Earth's surface had cooled below the boiling point of H_2O by ~3.9 Ga (Moorbath et al., 1972; Moorbath, 1984), and possibly much earlier (Wilde et al., 2001; Mojzsis et al., 2001). Thus, at least thin segments of the crust+mantle lithosphere and perhaps a liquid water ocean bounded the outer skin of the planet several hundred million years prior to formation of the surviving Archean rock record (Sleep et al., 2001; Harrison et al., 2005). Such Hadean (pre-rock record) crustal materials were efficiently recycled back into the mantle (Amelin et al., 1999; Boyett and Carlson, 2005).

On planets of the inner solar system, the presence of geomorphic features that reflect relatively recent surficial reworking due to tectonism, weathering and/or erosion, versus the retention of ancient impact features is a function of planetary mass (Head and Solomon, 1981). Large planets such as the Earth and, to lesser extents Venus and Mars, exhibit the effects of physiographic modifications, whereas little evidence of post-accretion surface reworking exists on Mercury and the Moon. Small bodies, and those close to the Sun have insufficient gravitational attraction to retain their original complement of volatiles, which escape into space, whereas those far removed from the Sun are too frigid to allow the formation of a water ocean.

The Earth, however, is of the appropriate size and distance to retain its volatiles effectively, sequestering H_2O chiefly in a globe-encircling ocean, and CO_2 combined with Ca±Mg precipitated largely as carbonate-rich strata. Polar gas molecules such as H_2O are highly soluble in silicate melts, and drastically lower the fusion temperatures of rocks, as apparent from Fig. 1. Aqueous solution enhances the planetary capacity for solid–liquid–vapor fractionation–differentiation, and by lowering solidus temperatures, decreases the temperature of the lithosphere–asthenosphere transition at any particular depth. Partial fusion is chiefly responsible for thermally created density instabilities (melts) in the mantle. Upwelling mantle and diapiric mush magmas act as vehicles for the surfaceward transfer of buried planetary heat. Such advective transport is manifested especially by mantle overturn and by rising plumes. Coupled with return flow, this type of mantle circulation represented an important mechanism for the differential transport and tectonic reworking of sections of the low-density rocky scum early in the Earth's evolution. As planetary thermal relaxation took place, a thicker, chemically buoyant continental lithosphere gradually became stabilized in the near-surface realm. In contrast, the growth of thicker, cooler, denser oceanic crust-capped lithosphere resulted in enhanced gravitational instability of such surficial plates, another driver of convective overturn in the suboceanic upper mantle.

Mantle circulation and lithosphere–asthenosphere couplings are evidently correlated in a cause and effect relationship. But just how and when did plate tectonic processes commence, early in Earth history (Kröner, 1981; Ernst, 1983; Sleep, 1992; Parman et al., 2001; Smithies et al., 2003, 2005; Sleep, 2005; Condie, 2005) or late (Davies, 1992; Hamilton, 1998, 2003; Stern, 2005)? And how did lithosphere–asthenosphere interactions change over time?

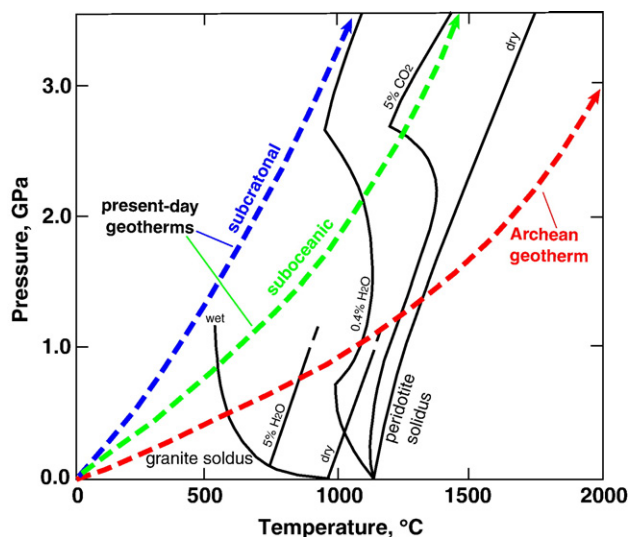


Fig. 1. Experimentally determined solidi for the onset of melting of undepleted peridotite, anhydrous and in the presence of minor H_2O and of CO_2 , and of granite, anhydrous and in the presence of minor and excess H_2O (Luth et al., 1964; Ito and Kennedy, 1967; Lambert and Wyllie, 1972; Wyllie, 1979, 1981; Presnall and Gudfinnsson, 2005). The H_2O -saturated solidus of basalt lies ~200 °C above that of wet granite (Hill and Boettcher, 1970; Poli and Schmidt, 1995; Liu et al., 1996). Present-day intraplate suboceanic and subcratonic geothermal gradients (Clark and Ringwood, 1964), and a 3.6 Ga geothermal gradient assuming about three times the modern radiogenic heat production and a stratified mantle (Lambert, 1976; Richter, 1988; Pollack, 1997) are illustrated.

2. Terrestrial heat budget and mantle circulation

Thermal mass and energy transfer in a gravitational field are mainly responsible for chemical and density differentiation of the Earth (Davies, 1999). As noted above, heat production reflecting the primordial complement of radioactive isotopes, meteorite impacts, and infall of the Fe+Ni core resulted in greatly elevated temperatures of the planet immediately after accretion. Wanke et al. (1984) estimated the relative global heating since ~4.5 Ga due to these processes as: accretion of planetesimals > Fe+Ni core formation > radioactive decay of K, Th, and U. The Archean geothermal gradient must have been significantly higher than the modern one (Fig. 1). Our planet has lost this heat through several mechanisms, chiefly by radiation (important immediately after impact of a Mars-sized body), convection, and conduction. Thermally driven gravitational instabilities transport heat by body flow (advection), and are the most efficient processes for conveying thermal energy from the deep interior to the Earth's surface (McKenzie and Weiss, 1975). Sleep (2000) described three types of planetary heat transfer, from the most to the least efficient: (i) magma oceans; (ii) plate (and plume) tectonics; and (iii) stagnant lid convection. The first, most energetic mode attended the Hadean stage of terrestrial evolution, and is not considered further here. The third, least energetic mode, characterizes modern Venus and Mars, but has not yet begun on Earth. Much of the modern, Wilson-cycle terrestrial heat dissipation occurs due to convective overturn in the suboceanic lithosphere–asthenosphere system. In contrast, the cratonic lithosphere acts as an insulating lid, leading to episodic break-up and continental drift (Gurnis, 1988; Lowman and Jarvis, 1996). On rifting and subduction, low-density lithosphere can be entrained back into the mantle due to viscous drag, but Lenardic et al. (2003) showed that a high brittle yield strength, combined with chemical buoyancy and a high-viscosity crustal root allows long-term, but imperfect stabilization of continental lithosphere relative to the denser oceanic lithosphere (Irifune, 1993) which is continuously recycled back into the upper mantle.

Convection in both the Earth's core and mantle is driven by two phenomena, bottom-up and top-down thermally generated gravitational instabilities. In the mantle, this is manifested by hot, buoyant overturn (including plumes), and cold, dense, sinking slabs, respectively (Davies, 1999; Anderson, 2001). To varying degrees over geologic time, these processes undoubtedly have driven mantle circulation. The presence of deep-seated heat concentrations and hot, ultra-thin, surficial platelets in the Hadean Earth suggests that widespread, vigorous mantle overturn and plume ascent from the base of the high-temperature anomalies mainly powered the convection (Lambert, 1981; Davies, 1993; Condie, 1994, 2005). Certainly by Phanerozoic time however, the thermal asperities residing in the deep mantle had lessened, whereas due to longer residence times at the surface and thus greater thermal relaxation, progressively thicker, cooler, oceanic lithospheric plates would have become more negatively buoyant, resulting in foundering, and driving mantle convection from the top of the planetary surface downward (Anderson, 1987; Conrad and Lithgow-Bertelloni,

2002). In contrast, the largely craton-capped plates would be chemically buoyant, and would remain stabilized in the near-surface environment.

Today most of the terrestrial heat loss takes place in and adjacent to oceanic spreading centers, whereas older portions of ocean basins and continental platforms are typified by lower thermal fluxes (Sclater et al., 1980). Substantial energy transfer occurs as descending plates warm in the hotter mantle, but slab heating at depth and cooling of the adjacent mantle merely results in internal temperature modification, not planetary heat loss. The early Earth probably was more heterogeneous in terms of thermal energy generation and temperature anomalies, and overall was characterized by substantially higher global heat flow. A hotter early Earth is demonstrated by the abundance of Archean komatiites and their scarcity thereafter. Depending on their H₂O contents, low-viscosity komatiitic lavas typical of Archean greenstone belts were fully molten at temperatures in the range of 1300–1600 °C (Green, 1975; Nisbet et al., 1977; Green, 1981; Abbott et al., 1994; Grove and Parman, 2004), reflecting large degrees of partial melting of the upwelling mantle. Such ultramafic liquids could only have been generated in oceanic and rifted continental realms of very high thermal flux. In contrast, thick, coherent crustal sections of high-grade granulitic gneiss characterize portions of Archean cratons (Lambert, 1976; Boyd et al., 1985; Artemieva and Mooney, 2001; Sleep, 2005), demonstrating that some parts of the continental lithosphere were typified by less extreme geothermal gradients.

In Hadean time, mantle advection likely was dominated by turbulent-to-poorly-organized circulating asthenospheric cells (Condie, 2005) and by plume-propelled heat transfer (see Abbott et al., 1994; Tackley, 2000), bringing vast amounts of thermal energy to the planetary surface, and from there lost to space. No rock record has survived this inferred chaotic period of Earth history. As excess heat asperities in deep-seated asthenospheric mantle declined, and as the peripheral skin of the planet cooled below the peridotite and basalt solidi, Archean production of mantle plumes and small, rapidly overturning asthenospheric cells would have allowed the continued efficient transfer of heat from the mantle toward the surface. Propelled chiefly by bottom up mantle flow, a thin, discontinuous rind of lithosphere would have capped these circulating systems, and platelet tectonics would be initiated. Compared with the present-day planet, lowered rigidity in a hotter Archean mantle probably resulted in much smaller scale convection cells typified by lower aspect ratios (Turcotte and Schubert, 1982). The horizontal dimensions of these cells probably were controlled by the high-temperature, exceedingly thin lithospheric lids (platelets), and by the rapid ascent of columns of hot, buoyant asthenosphere, promoting the formation of diminutive, more-or-less cylindrical convection cells.

Thus, in contrast to the modern terrestrial configuration involving relatively few magmatic hotspots fed by deep-seated mantle source domains, and instead dominated by the differential motions of large-scale plates and their underlying laminar convection cells (Anderson, 2005), the Hadean, and early Archean Earth evidently was characterized by oceanic regions underlain by many small, rapidly flowing upper mantle cells, with immature island arcs and continental debris swept into

cooler eddies above regions of mantle stagnation or regional down welling (Baer, 1981a,b; Boak and Dymek, 1982; Ernst, 1983; Richter, 1984). This scenario qualitatively accounts for the observed Archean geologic record, provided that the dimensions of asthenospheric convection cells are mirrored in the structures and sizes of the preserved continental crustal units and their mantle lithospheric underpinnings. The disaggregated, small-scale fabrics of ancient granite–greenstone belts and spatially associated tonalitic gneiss terranes (Goodwin, 1981a,b; Percival et al., 1994) have gradually given way over time to the remarkably long, continuous, paired metamorphic and volcano–plutonic belts of the Phanerozoic Era (Miyashiro, 1961, 1972). The familiar convergent plate petrotectonic features and subduction-mediated accretionary complexes of present-day continental crust are well-documented in Proterozoic orogenic belts (e.g., Windley, 1995; Lucas et al., 1996; Korsman et al., 1999; Cook et al., 1999; Condie, 2005).

3. Crust–mantle differentiation

Modern ocean basins are floored by young basaltic crust no older than ~0.2 Ga. Derived by the decompression partial melting of rising asthenosphere beneath- and consolidated at-spreading centers, oceanic crust and its depleted mantle underpinnings cool, recrystallize, become partially hydrated, and continuously return to the deep upper mantle. Probably such mafic–ultramafic lithosphere was also carried back into the Archean mantle by return flow (i.e., subduction) required to accommodate the new oceanic lithosphere generated by abundant, actively ascending mantle overturn and/or plumes. The ultimate fate of oceanic lithosphere is still a matter of debate, but in the modern Earth, devolatilization and partial melting of it and/or parts of the overlying mantle wedge have produced the calc-alkaline magmas that typify island arcs and active continental margins (Green and Ringwood, 1967; Martin, 1986; Rapp et al., 1995; Tatsumi, 2000; Stern, 2002). Such intermediate igneous material is less refractory and lighter than the mantle, so buoyantly rises and accumulates near the surface. Along with off-scraped accretionary sediments, docked micro-continental terranes, and tectonic scraps of oceanic crust, calc-alkaline plutonic and volcanic rocks have accumulated over time and make up the continental crust. Some calc-alkaline materials are themselves recycled, but intermediate to felsic igneous rocks represent most of the juvenile additions to the sialic crust; other igneous, metamorphic, and sedimentary lithologies mainly represent secondary reworking of the primary calc-alkaline suite (Ernst, 1999; Tatsumi, 2005).

Hadean–Early Archean asthenospheric circulation evidently was so vigorous that most, but not quite all (Compston and Pigeon, 1986), felsic materials were returned to the mantle due to viscous drag (Condie, 1980; Hargraves, 1981; Lenardic et al., 2003; Rogers and Santosh, 2004; Condie, 2005). Isotopic evidence suggests that, to a first approximation, upper mantle homogenization continued until about 3.6–3.8 Ga, that the average age of separation from the mantle of the now-preserved continental crust is 1.8–2.8 Ga, and that much of the terrestrial sialic crust (~70% or more) was generated by Early Proterozoic

time (Veizer, 1976; Jacobson and Wasserburg, 1981; DePaolo, 1981; Nelson and DePaolo, 1985; Taylor and McLennan, 1985; Turcotte and Kellogg, 1986; Condie, 1998). So although calc-alkaline arcs may have been produced by primitive Hadean subduction, the preservation of continental crust remained nil until the Early Archean due to near-total reincorporation into the mantle. By ~3.8 Ga, the rate of return to the mantle appears to have fallen below the rate of production of calc-alkaline material, hence gradually increasing amounts of sial were preserved, reflecting vigorous but decelerating asthenospheric convection. This process gradually removed LIL+incompatible elements from the residual mantle and, as energy input to the thermally driven gravitational instability waned, the rapid generation of both oceanic and continental crust from progressively more depleted mantle declined toward the present low and very low generation rates, respectively. Episodic to semicontinuous net additions of sialic material to the continental crust seem to have reached a maximum in the Late Archean and Early Proterozoic, and have declined ever since (O’Nions et al., 1980; Lambert, 1980; Nelson and DePaolo, 1984; Reymer and Schubert, 1986; McCulloch and Bennett, 1994).

4. Evolution of the lithosphere–asthenosphere system

Certain lithotectonic associations readily correlated with modern-style plate tectonic processes, such as blueschists and complete ophiolite stratigraphic sections, appear to be uncommon from all but the Late Proterozoic and younger rock record (Ernst, 1983; Rogers and Santosh, 2004; Stern, 2005; Maruyama and Liou, 2005). This begs the question: Did subduction and other plate tectonic processes characterize the early Earth? Advective heat transfer certainly has attended Earth evolution since infall of the Fe+Ni core and solidification of the magma mush ocean, yet no rocks have survived from Hadean time. Nevertheless, a peripheral solid rind and small, thin platelets surmounted by granitoids must have bounded the terrestrial surface beginning at least by 4.4–4.2 Ga, judging from detrital zircons of those ages. The hydrosphere was extant by no later than 3.9 Ga as required by the existence of metasedimentary strata, and may have been present much earlier (Wilde et al., 2001; Mojzsis et al., 2001). Thus, solid lithosphere clearly was present prior to the preservation of the earliest terrestrial rocks. Moreover, because the present-day Earth’s interior exhibits asthenospheric convection, it must have been overturning even more vigorously during the higher temperature geologic past, dominated by bottom-up (hot, buoyant) and to a lesser extent by top-down (cold, negatively buoyant) gravitational instabilities.

In the early Earth, extremely rapid, and initially poorly organized asthenospheric convection cells would have driven soft, weak platelets against and beneath one another, but due to elevated planetary temperatures and thinness of the slabs combined with a very small magnitude lithosphere–asthenosphere density inversion, platelets would not have been deeply subducted before heat conducted in from the adjacent mantle and rising temperatures would eliminate any negative buoyancy. This dynamic thermal condition accounts for the profound lack of high-pressure (HP) rocks such as blueschist–eclogite

belts and ultrahigh-pressure (UHP) collisional complexes in all but the youngest Proterozoic terranes. In Phanerozoic Wilson-cycle subduction zones, thick lithospheric plates carry relatively cool rocks to considerable depths (= high-pressure/low-temperature trajectories), whereas attending Archean subduction, small, warm platelets were rapidly heated to mantle temperatures (= low-pressure/high-temperature trajectories). Modern alkalic mafic igneous rocks such as ocean–island basalts form through very small degrees of partial fusion of the deep upper mantle, and are virtually lacking from Archean crust because the Earth's ancient geothermal gradient was too high. Instead, reflecting this elevated geothermal gradient, larger degrees of partial melting and komatiitic lavas were produced at relatively shallow mantle depths. Ductile micro-platelets also would be consistent with the small scale of preserved individual Archean granite–greenstone belts as oceanic crust was sutured against primitive island arcs (e.g., Percival et al., 1994; Mueller et al., 2002; Wyman et al., 2002). Some greenstone belts may represent oceanic plateaus derived by subsea effusion of asthenospheric plume heads. Archean tonalitic batholiths

might have been generated at relatively shallow depths by fractional crystallization of basaltic magma (i.e., Bowen, 1928; Sisson et al., 2004), through partial fusion of eclogite or mafic amphibolite (Green and Ringwood, 1967; Holloway and Burnham, 1972; Boettcher, 1973; Pertermann and Hirschmann, 2003; Dufek and Bergantz, 2005), or possibly by incipient melting of hydrous, fertile upper mantle peridotite (Kushiro et al., 1968; Mysen and Boettcher, 1975; Arculus, 1981).

Toward the end of the Archean, small quartzofeldspathic masses and immature island arcs, produced chiefly during mantle overturn combined with igneous differentiation–segregation, apparently had been assembled into thick, buoyant continental cratons (Williams et al., 1991; Aspler and Chiarenzi, 1998; Bleeker, 2002, 2003). Emergent landmasses became increasingly common (Schubert and Reymer, 1985), hence mechanical erosion and sedimentation became increasingly important processes in the production of chemically contrasting sedimentary facies. Concomitantly, biologically mediated increasing oxygenation of the atmosphere resulted in more intense chemical alteration (weathering) and in further

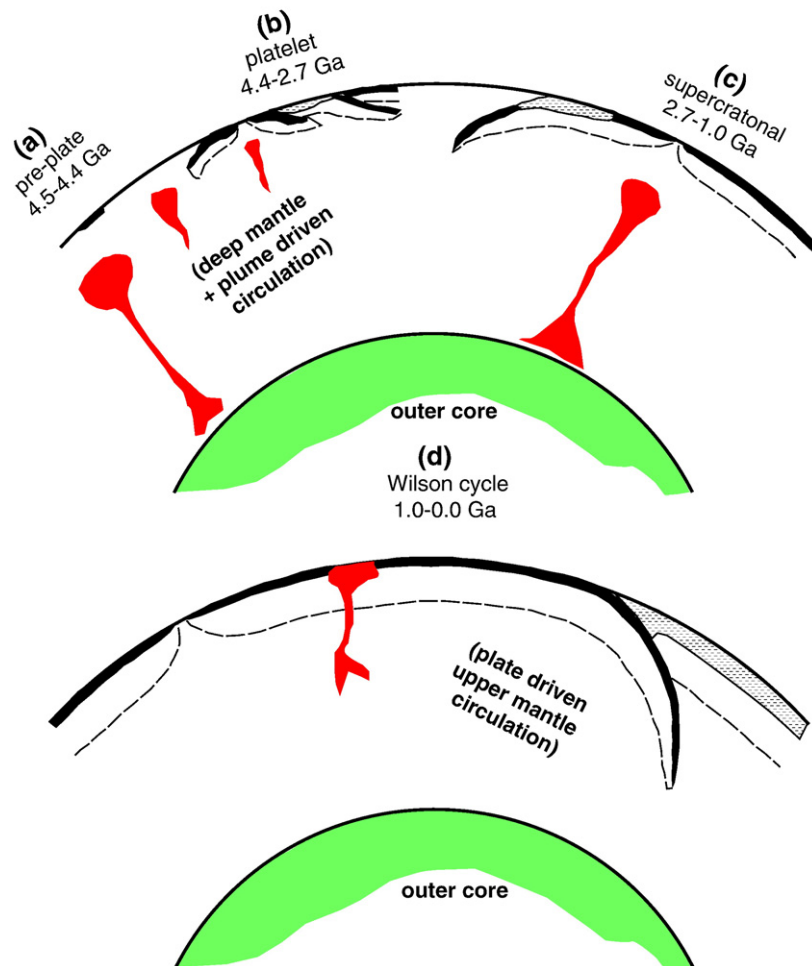


Fig. 2. Conjectured, progressive four-stage history of the terrestrial lithosphere–asthenosphere system. Oceanic crust is shown in black, continental crust in a dashed pattern. Thicknesses of the various components of the lithosphere are exaggerated for clarity. Early advective heat transport was accomplished to a large extent by mantle overturn and by plume ascent (shown in red), resulting in bottom-up control of convection. Over geologic time, thickening, enlarging oceanic lithospheric plates and chemically buoyant cratonal lids came to dominate mantle circulation and overturn through top-down control of asthenospheric convection by oceanic lithosphere. See text for discussion.

differentiation of the sedimentary section (e.g., [Kasting, 2001](#); [Melezhik et al., 2005](#)). This period was also characterized by the gradual transition from thin, hot Archean platelets, driven about by numerous small, rapidly convecting asthenospheric cells and plumes, to relatively thicker, laterally more extensive, cool plates, the differential motions of which may have been a function of mantle flow on a far grander scale, as well as more intense lithosphere foundering (i.e., subduction) ([Ernst, 1983](#)).

Four intergradational stages in the plate tectonic evolution of the Earth, and by inference, the changing scale and dimensions of asthenosphere–lithosphere interactions, are conjectured to have typified the outer portions of the planet over geologic time ([Smith, 1981](#); [Kröner, 1981](#); [Goodwin, 1981a,b](#); [Condie, 1982, 2005](#)). These are as follows: (a) A Hadean incipient lithosphere stage was attended by solidification of a near-surface magma mush ocean and the surface formation of a thin, discontinuous rocky scum, as surface temperatures fell below the solidi of peridotite and basalt. Reflecting viscous drag, turbulent-to-poorly-organized, vigorous convective overturn inserted this tenuous lithosphere back into the hotter mantle, completely recycling and obliterating the rock record prior to 4.03 Ga. (b) This condition gave way to a Hadean–

Archean stage in which the Earth's surface was covered by small, soft, hot, incipiently subductable platelets. Decompression fusion of hot upwelling asthenosphere and voluminous deep-seated plumes generated massive thicknesses of basaltic crust ([Sleep and Windley, 1982](#)), and through its shallow subduction, crustal suturing, and near-surface partial fusion, produced granite–greenstone belts and tonalitic complexes. The gradual accumulation and aggregation of these lithotectonic entities resulted in the amalgamation of island arcs and microcontinental terranes to form protocontinents. (c) Continued growth of sialic crust led gradually at the end of the Archean to a Proterozoic supercratonic stage typified by the emergence of broad continental platforms, development of freeboard and coupled sedimentary differentiation as well as to intracontinental suture zones. The biogenically mediated destruction of methane and increase in oxygen content of the atmosphere combined with occasional transpolar drift of continental assemblies resulted in the onset of terrestrial glaciations ([Holland, 1984, 1994, 2002](#); [Melezhik et al., 2005](#)). (d) Accompanying continued lateral enlargement of lithospheric plates and therefore, of ever larger asthenospheric convection cells, a Late Proterozoic–Phanerozoic stage of modern-style plate

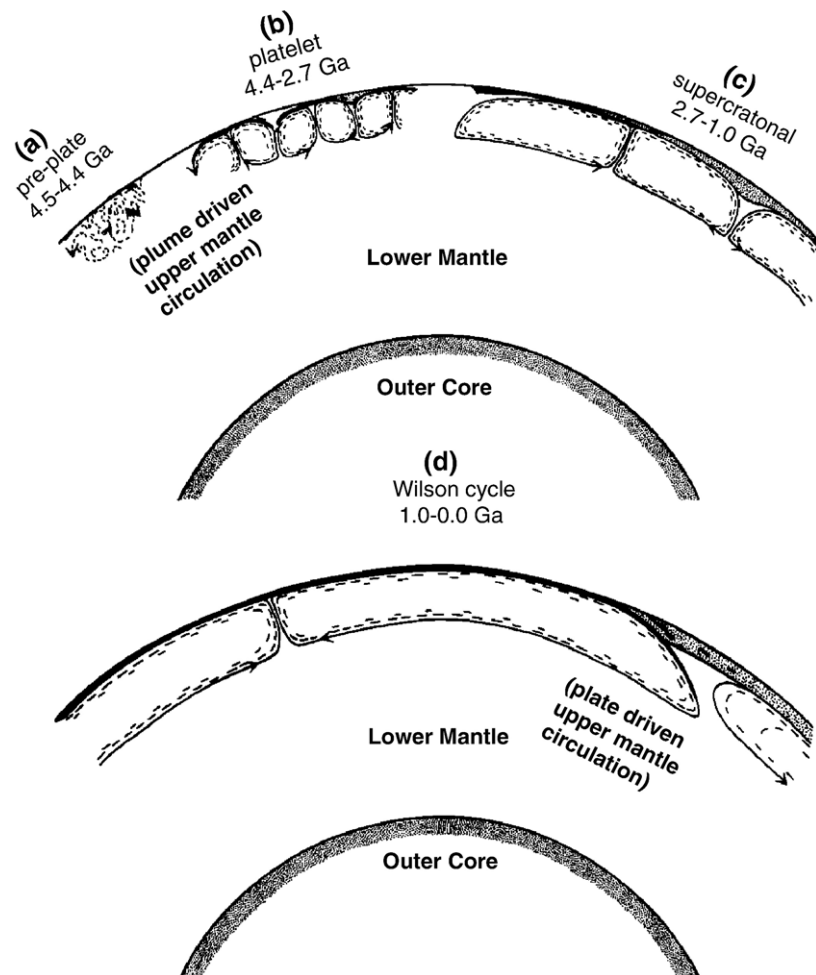


Fig. 3. Schematic pattern of asthenospheric mantle flow as illustrated in Fig. 2 (early-stage turbulent to ill-organized small laminar cells; late-stage enormous, large-aspect-ratio laminar cells), but more properly scaled (modified after [Ernst \(1983\)](#)). Oceanic crust-capped lithospheric plates are shown in black, continental crust-capped lithosphere by a stippled pattern.

tectonics arose by degrees that modified crustal development through the generation of long, linear subparallel outboard high-pressure subduction-zone complexes and inboard high-temperature volcano–plutonic calc-alkaline arcs. Inferred relationships in the upper mantle and lithosphere are illustrated schematically, emphasizing the evolution of the lithosphere in Fig. 2. Spatial relationships are more properly scaled, emphasizing the chaotic to laminar nature of flow in the asthenosphere in Fig. 3.

Reflecting terrestrial thermal evolution, early stages of the Earth apparently were dominated by vigorous ascent of hot, upwelling asthenosphere and deep-seated mantle plumes, reflecting heterogeneous mantle heat sources; with the passage of geologic time, and the lateral enlargement of progressively thickening lithosphere, convective overturn gradually became controlled chiefly by the sinking of dense, negatively buoyant oceanic plates. Asthenospheric circulation in the early Earth was rapid and chaotic, liberating vast amounts of heat to space, whereas over time, the gradual transition to larger, laminar-flowing convection cells became driven by the underflow of increasingly dense oceanic lids, and in the stabilization of chemically buoyant continental lithospheric plates. As briefly discussed below, the surviving rock record supports this speculative scenario.

5. Significance of lithotectonic complexes

5.1. Petrology and global tectonics

Ophiolite lithostratigraphic assemblages, island arc+active continental margin calc-alkaline volcano–plutonic suites, and HP+UHP metamorphic terranes reflect the operation of modern, well-recognized Wilson-cycle plate tectonic processes (Stern, 2005). Rifted, transcurrent faulted, and imbricate thrust faulted crustal complexes also attest to divergent, transform, and convergent motions of the various lithospheric plates, as do the positions and architectures of sedimentary basins, orogenic belts, and terrane amalgams in general. On a grand scale, continents and ocean basins represent first-order responses of the planet to ongoing differential plate movements and asthenospheric overturn. As sketched above, what is known regarding the Earth's thermal and petrochemical differentiation must be integrated with the ancient terrestrial rock record in order to place better constraints on the changing style of planetary evolution. The fragmentary lithologic record is retained only in the sialic crust. In general, Archean crust is typified by granite–greenstone belts, komatiitic basalts, and gray gneiss (tonalitic) terranes, but almost totally lacks mafic alkaline igneous rocks, HP blueschist–eclogite belts, and UHP metamorphic collisional complexes (Ernst, 1972; Bailey and Woolley, 2005; Stern, 2005). Fully developed ophiolite suites and multicycle platform sedimentary strata do not appear in the geologic record until near the end of Archean time, whereas the oldest HP and UHP complexes date from ~0.8 and 0.53 Ga respectively (Ernst, 1983; Scott et al., 1992; Kusky et al., 2001; Condie, 2005; Stern, 2005; Polat et al., 2005). The differences in lithologic assemblages reflect markedly contrasting Archean versus Phanerozoic thermal regimes in both crust and upper mantle, as shown in the

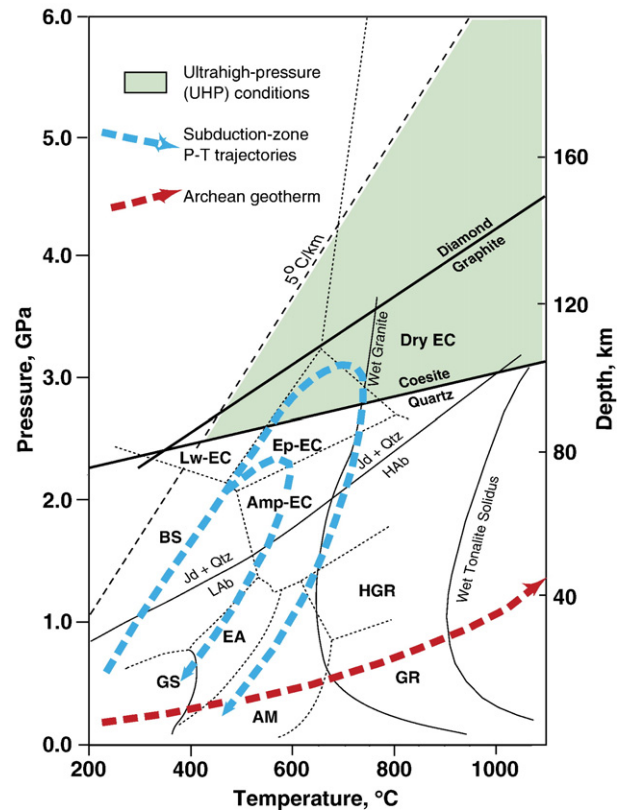


Fig. 4. Petrogenetic grid for rocks of basaltic bulk composition, modified from Liou et al. (1998) and Okamoto and Maruyama (1999). Phanerozoic P – T paths for outboard Pacific-type (HP) and Alpine-type (UHP) metamorphism are indicated, as well as a 3.6 Ga early Earth geothermal gradient. An extremely low subduction-zone geothermal gradient of 5 °C/km is also shown for reference. Literature citations to the experimental phase equilibria were provided by Liou et al. (1998). Mineral abbreviations are: Jd = jadeite; Qtz = quartz; LAb = low albite; and HAb = high albite. Metamorphic-facies abbreviations are: AM = amphibolite; Amp-EC = amphibolite–eclogite; BS = blueschist; EA = epidote amphibolite; EC = eclogite; Ep-EC = epidote–eclogite; GR = sillimanite–granulite; GS = greenschist; HGR = kyanite–granulite; Lw-EC = lawsonite–eclogite; and Px–Hf = pyroxene hornfels. Feebly recrystallized rocks of the zeolite, prehnite, pumpellyite–actinolite, and prehnite–pumpellyite metamorphic subsfacies (not illustrated) lie at lower temperatures and pressures than GS and BS.

metamorphic facies diagram of Fig. 4. But do such petrologic contrasts indicate that plate tectonic processes did not operate on the early Earth?

5.2. Two types of Late Proterozoic–Phanerozoic convergent plate junctions

Late Proterozoic and younger contractional mountain belts form near the active edges of continents or fringing island arcs. All reflect the consumption of lithospheric plates attending underflow of oceanic crust and the consequent transport and descent or accretion of spreading centers, oceanic plateaus, exotic island arcs, far-traveled sialic terranes, microcontinents, and/or salients of continental crust. Long-sustained subduction constructs a landward calc-alkaline volcanic–plutonic arc on the nonsubducted plate (DePaolo, 1981; Drummond and Defant, 1990; Hawkesworth et al., 1993; Clift et al., 2001; Hickey-Vargas et al., 2002), as well as a seaward trench complex (Ernst,

1971; Miyashiro, 1972; Frey et al., 1974; Maruyama et al., 1996). Brief episodes of underflow, such as result from the subduction of a small intervening ocean basin, also can produce an outboard subduction complex, but an inboard volcano–plutonic arc normally is only incipient formed or fails to develop at all. No two orogenic belts are identical. Many contain disaggregated or relatively complete ophiolite lithostratigraphic entities. If mafic crust is subducted prior to accretion, these underplated oceanic assemblages characteristically are accompanied by HP blueschist–eclogite metamorphic parageneses. Some microcontinental collisional terranes contain rare, scattered mineralogic relics indicating a stage of UHP recrystallization (Chopin, 1984; Smith, 1984; Sobolev and Shatsky, 1990; Coleman and Wang, 1995; Liou et al., 1998; Liu et al., 2004), but because of thorough retrograde reaction, many other collisional suture zones—which might have been subjected to similar P – T conditions—fail to exhibit any evidence of UHP metamorphism.

A broad continuum exists between two distinct end-member types of mountain building (Bally, 1981; Ernst, 2005). (i) Alpine-type orogens form during subduction of a relatively narrow belt of oceanic lithosphere lying between two sialic blocks, resulting in continental collision. They are characterized by an imbricate sequence of oceanward-verging nappes; some Alpine orogens exhibit superimposed late-stage back thrusting. Metamorphism of deeply subducted portions of such belts ranges from moderately high-pressure to UHP. Volcano–plutonic rocks of intermediate composition are rare, and those that are present typically have peraluminous, S-type bulk compositions. (ii) In contrast, Pacific-type orogens develop within and landward from long-duration oceanic subduction zones. They consist of an outboard trench–forearc complex, and an inboard magmatic arc. The subducted oceanic section recrystallized under HP conditions. Recumbent fold vergence is typically oceanward. A massive, roughly coeval calc-alkaline arc is sited landward from the trench complex on the nonsubducted plate. It consists of abundant, chiefly metaluminous, I-type intermediate volcanic rocks resting on old crust; both basement and superjacent volcanogenic units are thrown into open folds, intruded by comagmatic I-type granitoids, and metamorphosed regionally under high-temperature/low-pressure conditions. Similar to some Alpine belts, the hinterland continentward from the magmatic arc may exhibit late-stage antithetic thrusting.

5.3. *Absence of certain distinctive Wilson-cycle lithotectonic suites in ancient continental crust*

Ophiolites±alkali basalts, blueschist–eclogite belts, and UHP collisional complexes typify the outboard mafic igneous+metamorphic zones of Late Proterozoic–Phanerozoic convergent plate junctions, but are lacking from ancient sections of continental crust. Either these diagnostic plate–tectonic rock types have been completely obliterated by subsequent tectonic and/or erosional processes, or did not form during Archean time. Evidently because of higher temperature subduction-zone P – T trajectories attending the descent of thinner, warmer plates, HP and UHP complexes were not produced on the early Earth. Alternatively, the absence of these significant but vulnerable

lithologies from ancient crustal sections might reflect metamorphic overprinting or tectonic removal. However, the inboard position of a massive calc-alkaline volcano–plutonic belt is much less susceptible to post-convergence overprinting, tectonic erosion, or obliteration, hence is more likely to be preserved. Such subduction-related igneous belts have formed throughout recorded geologic time judging by the occurrence of Archean granite–greenstone belts and tonalitic terranes—the latter generated through the partial melting of mafic eclogites, metabasaltic amphibolites, fertile, hydrous peridotite protoliths, or by fractional crystallization of precursor basaltic liquids. Low-angle thrusting is characteristic of some Archean orogens, and the presence of granite–greenstone belts and voluminous gray gneiss terranes suggests that the processes of sea-floor spreading and convergent plate suturing were operating along active divergent and convergent margins, but at somewhat higher temperatures (Kröner, 1981; Condie, 1982; Ernst, 1983; Windley, 1995). Transform faulting in ancient Precambrian terranes also may reflect differential lithospheric plate motions (Sleep, 1992).

If the ophiolite petrologic suite was produced voluminously in ancient oceanic plate tectonic settings (oceanic ridge, forearc, and/or backarc) similar to lithotectonic environments characterizing the current Wilson cycle, why are such basalt–gabbro–cumulate peridotite lithostratigraphic packages largely absent from the rock record prior to ~2.0 Ga, and scarce until ~1.0 Ga? The answer may lie in the thermal structure of the Archean–Early Proterozoic Earth. Much thinner lithospheric plates, and far more numerous, vigorously rising mantle convection limbs and plumes would have been consequences of the elevated global temperatures and consequent heat flow. A high geothermal gradient means that ascending, chemically primitive asthenospheric columns must have undergone large degrees of decompression partial fusion, producing oceanic crust substantially thicker than the modern analogue (Sleep and Windley, 1982; Condie, 2005). In addition, because of the high geothermal gradient in the early Earth, the brittle–ductile transition would have resided at shallower depths in the oceanic crust than it does today. Hence it seems possible that during plate convergence and underflow, superjacent (extrusive) sections of basaltic crust could have been detached from the top of the downgoing lithosphere and inserted into the adjacent accretionary margin without involving the entire ophiolite stratigraphic sequence. The sweeping together, decapitation, and suturing of such thick oceanic crust and oceanic plateaus against island arcs would yield the ubiquitous Archean granite–greenstone belts. The small size of these amalgams may mirror the lateral dimensions of the convecting Archean asthenosphere. But by Proterozoic time, long-sustained planetary cooling resulted in lower geothermal gradients, generation of thinner oceanic crustal sections, and deepening of the brittle–ductile transition; accordingly, complete basalt–gabbro–cumulate peridotite sections of oceanic crust began to behave as integral units, and more complete lithostratigraphic entities (ophiolite complexes) were inserted into the accreting continental crust during subduction and closure of back arc basins.

It seems possible that the Pacific versus Alpine styles of Wilson-cycle convergent crust construction have less

differentiated analogues in the Archean–Early Proterozoic rock record. The present-day outboard subduction zone–inboard calc-alkaline arc versus continental collision complex shows lithotectonic similarities to ancient granite–greenstone belts versus tonalitic gneiss terranes. If so, the scales, dynamics, and thermal structures of the crust–mantle system have evolved in response to planetary cooling, but the fundamental processes of lithosphere generation and plate tectonics have not changed significantly.

6. Summary

Advective heat transfer in the upper mantle was dominated by hot, rising asthenosphere and mantle plumes, due to bottom-up drive of mantle overturn in the early Earth. By 4.4–4.2 Ga at the latest, a discontinuous rocky scum had solidified, and platelet tectonics was initiated. Reflecting elevated geothermal gradients, high degrees of decompression partial melting of ascending Archean mantle and plumes generated komatiitic lavas and thick sections of oceanic crust+oceanic plateaus. With the passage of geologic time, tectonics dominated by the differential transport of rapidly moving lithospheric platelets gradually evolved into capping plates and mantle circulation cells of greatly increased lateral dimensions. Cooling cratonic lithosphere became thicker and chemically more buoyant, reflecting crust–mantle differentiation. Oceanic crust-capped lithospheric plates also became cooler and thicker, but were characterized by increasingly negative buoyancy; such oceanic plates began to assume control over the sizes and dynamics of circulating asthenospheric cells through top-down drive of mantle overturn. The temporal restriction of intact ophiolite lithostratigraphic packages±alkali basalts to Proterozoic and younger crust, and of HP blueschist–eclogite belts+UHP collisional complexes to ~1.0 Ga and younger crust reflects the modern Wilson-cycle subduction of enormous, relatively cool oceanic lithospheric plates. The generation and suturing of convergent margin terranes has been a consequence of at least shallow subduction since incipient preservation of the terrestrial rock record beginning at 4.03 Ga, as testified to by amalgamated Archean oceanic belts+calc-alkaline arcs with collisional sialic terranes. Phanerozoic Pacific and Alpine types of convergent orogens may have early Earth analogues in granite–greenstone belts and tonalitic gneiss terranes, respectively.

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