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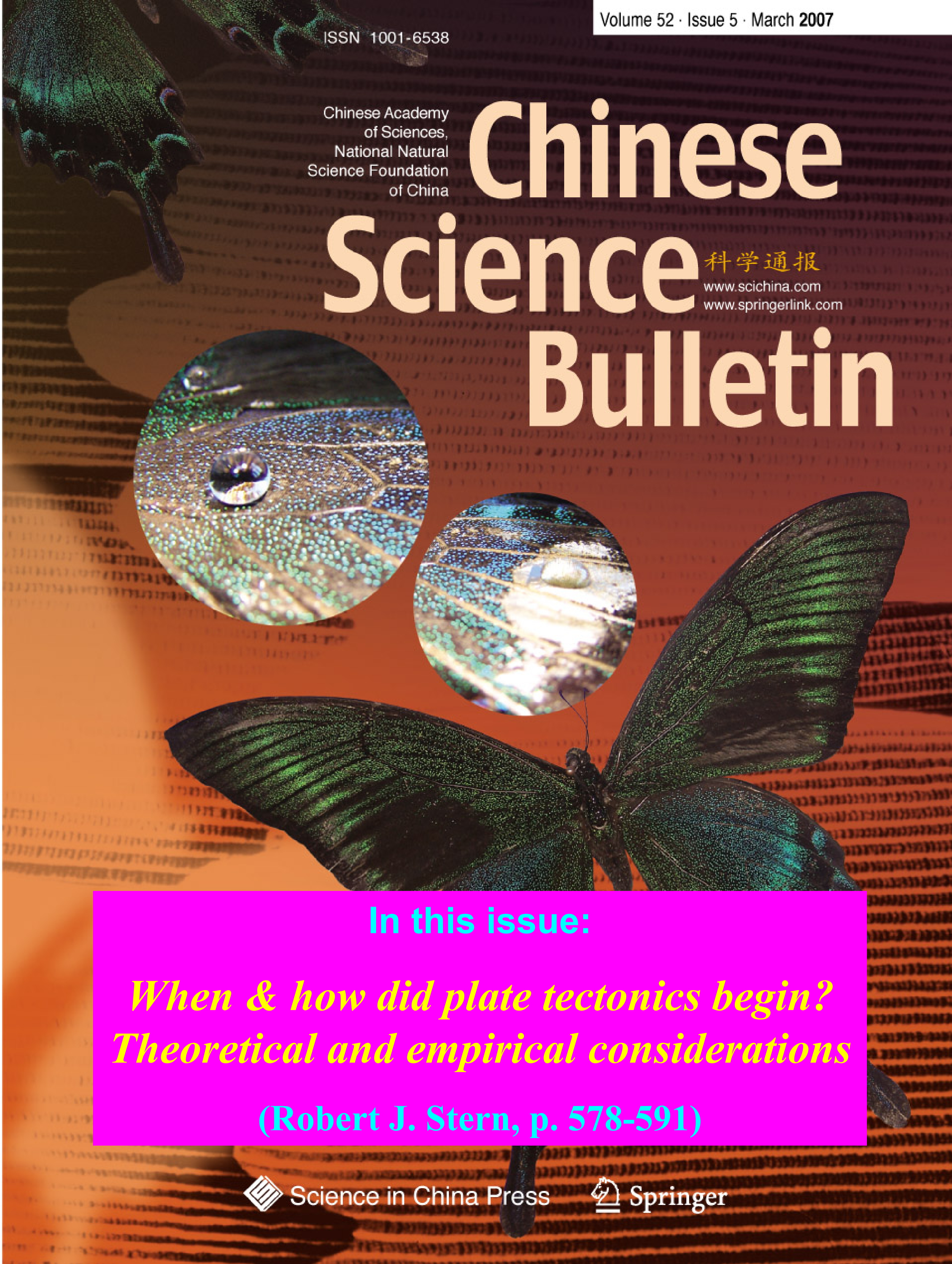

**In this issue:**

***When & how did plate tectonics begin? Theoretical and empirical considerations***

**(Robert J. Stern, p. 578-591)**

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## Editor's note

Robert (Bob) J. Stern is a Professor of Geosciences at the University of Texas at Dallas. He received a BSc in Geology with honors from the University of California at Davis in 1974, and a PhD in Earth Sciences from the University of California at San Diego (Scripps Institution of Oceanography) in 1979. He held a postdoctoral fellowship in the Department of Terrestrial Magnetism at the Carnegie Institution of Washington (1979—1982) before he joined the University of Texas at Dallas in 1982. There he was Department Head of Geosciences Department for 8 years (1997—2005). He has also held honorary positions at Stanford University (Blaustein Fellow, 2005) and California Institute of Technology (Tectonics Observatory Fellow, 2006). Bob is best described as a "traditional" geologist with many deep interests in the broad field Earth Science. Particularly he is an expert at using remote sensing, petrology, and isotope geochemistry as tools to study important tectonic problems on local, regional and global scales. He has published over 140 research papers in leading journals such as *Nature*, *Science*, *Geology*, *Earth and Planetary Science Letters*, *Journal of Geophysical Research*, *Journal of Petrology*, *Contributions to Petrology and Mineralogy*, *Chemical Geology*, *Precambrian Research* etc.

Bob focuses his research primarily on how the continental crust forms today and how it did in Earth's history. His current research projects in Egypt, Arabia and Ethiopia aim to unlock how the process of oceanic lithosphere subduction in Neoproterozoic time generated thickened arc crust, the proto-continental crust, over a few hundred million years. Also over the past decades, Bob has been studying how continental crust is actually produced today above subduction zones of the western Pacific, specifically focusing on the Mariana arc. This is accomplished by monitoring the compositional variations in time and space of volcanic rocks collected by submersibles, surface ship dredging and deep-sea drilling. He is also one of the major driving forces behind NSF-funded MARGINS program in the US to investigate petrological, geochemical and geophysical consequences of subduction-zone factories. With the concept of "plate tectonics being subduction tectonics" in mind, I myself developed a deep interest in the BIG problem of subduction initiation (*Journal of Petrology*, 2003, 44, 851—866). However, Bob is much more ambitious, and asks when plate tectonics actually began on Earth (*Geology*, 2005, 33, 557—560). Fond of observations, full of a wide range of knowledge, and equipped with an analytical mind, Bob is a creative thinker on how the Earth works. Together with Condie and Kröner, Bob organized a GSA Penrose Conference, *When did plate tectonics begin on Earth? Theoretical and empirical constraints*, held in Lander, Wyoming, USA (June 13th to 18th, 2006), which is considered by many as an unprecedented gathering on one of the most fundamental Earth problems.

I am delighted that Bob Stern accepted the invitation to contribute an article to *Chinese Science Bulletin* on this important subject. He starts with an objective overview on relevant aspects backed up by observations and physically plausible arguments, followed by discussion on geologic criteria for recognizing plate tectonics before he presents his personally preferred idea. Contrary to many views that plate tectonics began on Earth very early, in the Archean or perhaps, even in the Hadean, about ~100 Ma after Earth accretion, Bob argues for a progression of tectonic styles from Archean-type tectonics to something similar to plate tectonics at ~1.9 Ga, but the present-day plate tectonics did not begin until the Earth became cold enough to allow establishment of self-sustainining subduction in Neoproterozoic time. The author does, however, emphasize that to fully address when and how plate tectonics began requires a truly multidisciplinary effort. There is no doubt that debates will certainly facilitate a progressively improved understanding of both *when* and *how*. It is my hope that this paper will offer an impetus that encourages our Chinese scientists, in particular the younger generations, to participate in this exciting debate.

(Yaoling Niu, Executive Editor, Department of Earth Sciences, Durham University, UK)

# When and how did plate tectonics begin? Theoretical and empirical considerations

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**Plate tectonics is the horizontal motion of Earth's thermal boundary layer (lithosphere) over the convecting mantle (asthenosphere) and is mostly driven by lithosphere sinking in subduction zones. Plate tectonics is an outstanding example of a self organizing, far from equilibrium complex system (SOFFECS), driven by the negative buoyancy of the thermal boundary layer and controlled by dissipation in the bending lithosphere and viscous mantle. Plate tectonics is an unusual way for a silicate planet to lose heat, as it exists on only one of the large five silicate bodies in the inner solar system. It is not known when this mode of tectonic activity and heat loss began on Earth. All silicate planets probably experienced a short-lived magma ocean stage. After this solidified, stagnant lid behavior is the common mode of planetary heat loss, with interior heat being lost by delamination and "hot spot" volcanism and shallow intrusions. Decompression melting in the hotter early Earth generated a different lithosphere than today, with thicker oceanic crust and thinner mantle lithosphere; such lithosphere would take much longer than at present to become negatively buoyant, suggesting that plate tectonics on the early Earth occurred sporadically if at all. Plate tectonics became sustainable (the modern style) when Earth cooled sufficiently that decompression melting beneath spreading ridges made thin oceanic crust, allowing oceanic lithosphere to become negatively buoyant after a few tens of millions of years. Ultimately the question of when plate tectonics began must be answered by information retrieved from the geologic record. Criteria for the operation of plate tectonics includes ophiolites, blueschist and ultra-high pressure metamorphic belts, eclogites, passive margins, transform faults, paleomagnetic demonstration of different motions of different cratons, and the presence of diagnostic geochemical and isotopic indicators in igneous rocks. This record must be interpreted individually; I interpret the record to indicate a progression of tectonic styles from active Archean tectonics and magmatism to something similar to plate tectonics at ~1.9 Ga to sustained, modern style plate tectonics with deep subduction——and powerful slab pull——beginning in Neoproterozoic time.**

Plate tectonics, subduction, Precambrian, geodynamics

In June 2006, a Geological Society of America Penrose conference in Lander, Wyoming, USA, convened to discuss when plate tectonics began. Here is how the meeting was described: "*Earth is the only planet with plate tectonics, and it is controversial why and when this began. Some argue that plate tectonics already operated in Archean time, whereas others argue for a much later beginning. Numerical experiments show that Earth's thermal boundary layer, the lithosphere, controls mantle convection. We now understand that the sinking of dense lithosphere in subduction zones is responsible for plate motions and seafloor spreading and is the principal way Earth cools. The hotter early Earth may have had a weaker and less dense lithosphere and so had a different style of mantle convection. This theoretical background is essential to provide context for empirical (field and lab analyses) results. Demonstrating that plate tectonics operated at any given time requires evidence for subduction and independent plate motions and rotations.*

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*Understanding when and why plate tectonics began is one of the most important unresolved problems in understanding Earth; progress toward resolving this requires a creative and interdisciplinary effort. Theoretical considerations include Earth's thermal history, what powers the plates, and how Earth's progressive cooling may have affected this evolution. Empirical considerations include the results of a wide range of field- and laboratory-based measurements and analyses that serve as proxies for plate motion or operation of subduction zones and ocean ridges, including paleomagnetism, isotopic and trace element compositions of igneous rocks, styles of deformation, the temporal distribution of hallmarks of subduction (ophiolites, blueschists, eclogites, and diamond- and coesite-bearing ultrahigh-pressure assemblages) and secular distribution of associations diagnostic of plate tectonics such as passive margin sequences and arc rock assemblages.*"

Sixty-one geoscientists from 14 nations shared a wide range of opinions regarding when plate tectonics began; further details about the meeting can be found at refs. [1, 2], and meeting presentations can be found at http://utdallas.edu/~rjstern/PlateTectonicsStart/presentations.htm. My conclusion that the modern style of plate tectonics began late in Earth history, ~1 Ga or younger[3] is a minority opinion. Most geoscientists at the Penrose conference concluded that plate tectonics began much earlier[1,4], but there was general agreement that Earth's tectonic style must have evolved as the planet cooled. The most pertinent considerations for understanding when plate tectonics began are explored below. First the modern plate tectonic/subduction system is described. Then the unusual nature of Earth's tectonic style relative to those of its neighbors is noted, and the implications of this explored. Finally, some of the geologic evidence of plate tectonic processes are presented and discussed.

## 1 What is plate tectonics? What makes the plates move?

Plate tectonics describes how lithospheric sectors move and interact across Earth's nearly spherical surface[5]. Plate motions are described using absolute or relative reference frames with a rotation pole and angular velocity. Plate tectonics is a powerful but purely kinematic description; the theory was formulated without knowing what forces drive the plates. Early on, the driving force was mistakenly thought to result from convection of the asthenosphere, the weak mantle beneath the lithosphere (Figure 1(b)). This is a deeply embedded idea that is still taught to Introductory Geology classes. One popular textbook says that an important force for plate motions is "Convective motion in the asthenosphere applies drag to the base of the plate". This consideration, if true, leads to the reasonable conclusion that the ancient Earth, being hotter, with a lower viscosity and rapidly mixing asthenosphere, would have more or faster plates. In contrast, if our present understanding—that the plates are mainly powered by sinking of the oceanic lithosphere in subduction zones—is correct, then plate tectonics could not have happened until the Earth cooled sufficiently to allow gravitationally unstable oceanic lithosphere to form and dominate the ocean basins.

The hypothesis that asthenospheric convection powers plate motions is not supported by modern geodynamic considerations (outlined below) or from observations of relationships between asthenospheric and plate motions. First, the asthenosphere is weakly coupled to the lithosphere and has little effect on its motions[6]. The asthenosphere generally does not flow parallel to the overlying plate. Beneath the interior of an oceanic plate, asthenosphere moves towards the ridge[7], whereas around the plate margins the asthenosphere flows into or away from the basin, depending whether the ocean is growing (Atlantic-type) or shrinking (Pacific-type)[8]. Some workers infer an "eastward drift" of asthenosphere[9] but east-moving plates do not move faster than west-moving plates. There may be some basal drag affecting the thick lithospheric keels of continents[10−12], but this flow impedes as well as encourages the motions of continent-bearing plates. The most spectacular example of how independent are plate motions and asthenospheric flow is the Australia-Antarctic Discordance (AAD), where Indian and Pacific asthenospheric mantle domains meet and sink into the mantle beneath the mid-ocean ridge separating the Australian and Antarctic plates[13].

It is now clear that the plates drive themselves as they organize overall mantle convection[14,15]. Plate motions ultimately result from the negative buoyancy of the lithosphere subsiding from ridges (ridge push) towards a trench and sinking in subduction zones (Figure 1(a))[6]. Lithosphere is buoyant when it forms by spreading at a mid-ocean ridge, where it consists only of basaltic crust.

Oceanic crust alone is less dense than asthenosphere, but the mantle part of lithosphere is colder and therefore denser than asthenosphere. Oceanic lithosphere thickens by conductive cooling of the upper mantle, which progressively increases bulk lithosphere density (crust plus mantle) as it ages (Figure 2). This results in a buoyancy crossover, when oceanic lithosphere becomes denser than the underlying asthenosphere. For the modern Earth, this "buoyancy crossover time" happens relatively quickly and today lithosphere is denser than asthenosphere after 20—40 million years[16,17].

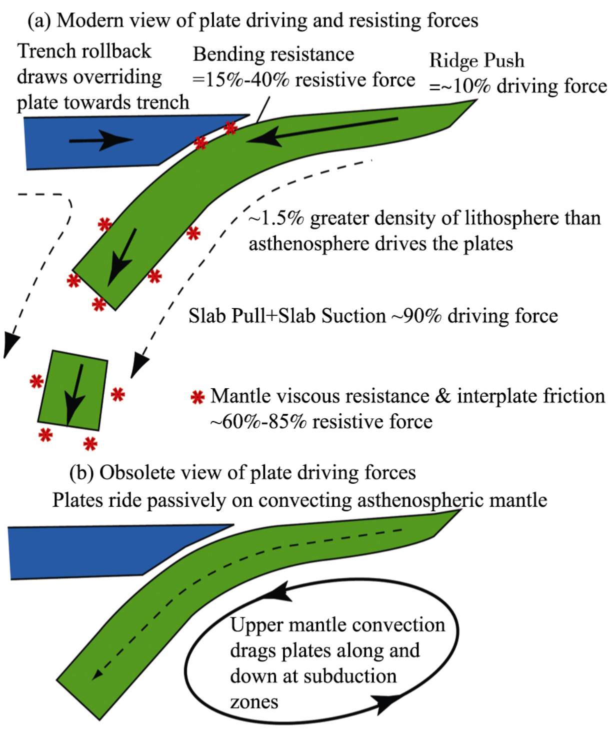


**Figure 1** Summary diagram of plate-driving forces. (a) Modern view of plate-driving forces; (b) obsolete view of plate-driving forces. In spite of the broad acceptance of (a) by the geodynamic community, (b) is considered by many geologists to be important. See text for further discussion.

The negative buoyancy of aging oceanic lithosphere provides the potential energy that powers plate tectonics. This potential energy is partitioned between ridge push and slab pull and suction. Ridge push provides about 10% of the driving force, and descent of lithosphere in subduction zones provides the rest; because of this, Earth's present tectonic style is more aptly described as "Subduction Tectonics"[3]. Modern plate tectonics is characterized by very deep subduction. Tomographic images suggest that some subducted slabs can be traced with some confidence down to 1100—1300 km, and with less confidence down to about 1700 km but it is not clear yet whether such slabs sink all the way to the core-mantle boundary[18]. Sinking of the lithosphere in subduction zones controls plate motions in two ways, by directly pulling on the plate (slab pull) and by entraining the surrounding mantle to descend with it (slab suction). Even detached subducted slabs help drive the plates, by entraining mantle that sinks with the slab (Figure 1(a)); this is known as "slab suction". Observed plate motions are best predicted if slab pull and slab suction forces each account for about half of the subduction-related driving force[19,20]. The plates do not move parallel to the dip of the Wadati-Benioff zone, but also have a vertical component that results in trench roll-back, as required by shrinking of the Pacific Ocean[21].

The sinking of lithosphere in subduction zones is resisted by dissipation forces in the lithosphere and by viscous resistance of the asthenosphere (Figure 1(a)). Lithospheric resistance includes bending as plates descend into subduction zones (~40% of total dissipation[22]), faulting, and sliding resistance deformation at the outer trench swell[23]. Asthenospheric viscosity and resistance increases with mantle depth. Physical equilibrium requires a balance of driving and resisting forces, resulting in a "plate tectonic speed limit", which for the Cenozoic is approximated by the northward movement of India at 170 mm/a during Paleocene time, ~60 Ma.

In summary, there is a strong consensus among geodynamicists that plate tectonics and subduction drive mantle convection, encapsulated as "Top-down Tectonics"[24]. This consensus warrants re-examining when in the history of the cooling Earth these forces were sufficiently great to overcome the resisting forces and establish plate tectonics. First, we briefly examine the other silicate planets to stress the point that plate tectonics is unusual in the space dimension, and so may also be unusual in the time dimension.

## 2 Plate tectonics is an unusual way for a silicate planet to cool

The geoscientific subdiscipline of tectonics is concerned with understanding deformation of the outer conductive shell—the lithosphere—of Earth. This deformation occurs at all scales, forming mountain chains and mid-ocean ridges as well as deforming mineral grains. Ultimately, tectonic style reflects how the planet loses its internal heat, and plate tectonics is only one mode of

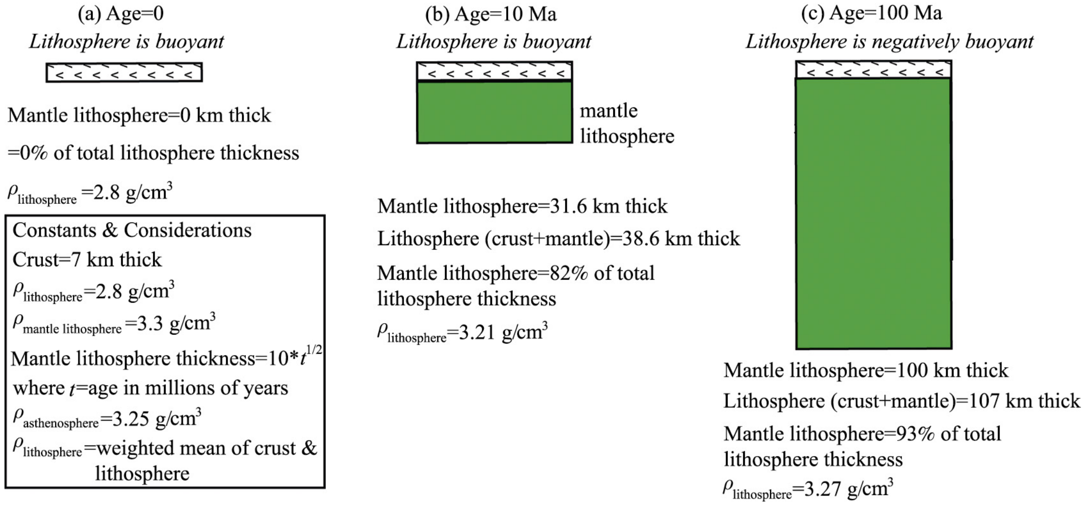


**Figure 2** Development of negative buoyancy with increasing age of modern oceanic lithosphere.

planetary cooling. Of the 5 largest silicate bodies that we know (Mercury, Venus, Earth, Moon, and Mars), only Earth has subduction zones and plate tectonics[25,26]. Because tectonic styles fundamentally express cooling mode and larger bodies retain heat longer than small bodies, it may not be surprising that the largest silicate planet—Earth—has a different tectonic style from other smaller silicate bodies. Nevertheless, the unusual cooling style of the largest silicate planet warrants closer examination.

Planets lose heat even as radioactive decay in their interiors produces heat. Interior heat is lost by conduction and advection (shallow intrusions, lava flows, and hydrothermal activity). Every year the Earth is estimated to lose $4.2\times10^{13}$ W of heat: 32 TW is conducted through the surface thermal boundary layer (lithosphere) and about 10 TW may be lost by hydrothermal activity at mid-ocean ridges[27]. Tectonically active planets like Earth lose more heat than they produce and thus progressively cool, but it is also possible for a planetary interior to heat up, particularly if it has a thick conductive lid and high concentrations of radioactive elements deep in the planet.

Early in their histories, all the planets were partially to largely molten (magma ocean stage), but these were brief beginning episodes[28]. The magma ocean stage was the inevitable result of tremendous early heat sources: planetary accretion, greater abundance of radioactive U, Th, and K, presence of now-extinct radionuclides (especially $^{26}$Al), continued bolide impacts, core formation, and perhaps the Sun's T-tauri phase[27−29]. A magma ocean is the most effective way for such a hot planet to lose heat, but once this ocean solidifies and a conductive lid forms, the body must cool differently. Stagnant lid—a single lithospheric plate encompassing the entire planet—seems to be the dominant mode heat loss for the silicate planets, but there are many variants. We especially do not understand the magmatically active stagnant lid mode, where lithosphere nevertheless must sink—perhaps by delamination—into the deep interior, in order to compensate for magma moving to the surface[30,31]. The reasons for stagnant lid behavior vary: sometimes this reflects buoyant lithosphere, sometimes because the lithosphere is too strong, sometimes because the underlying asthenosphere is too viscous, and sometimes because the planet is too small and cold. Given that stagnant lid tectonics is the dominant mode of silicate planets today, it seems likely that Earth also experienced one or more stagnant lid episodes during its 4.5 Ga history. Stagnant lid tectonics on Earth was almost certainly accompanied by abundant tectonic and magmatic activity, either as a prelude to or interlude between plate tectonic episodes.

The mode of heat loss and thus tectonic style may change multiple times over the life of a cooling planet[25]. Ultimately, it is a planet's fate to cool by conduction alone; when this happens, the planet becomes tectonically and magmatically inactive, either dead or hibernating. This is the present situation of Mercury and our moon. These bodies lose heat only by conduction through the surface, which is a very inefficient way to cool. These small bodies have thick, strong stagnant lids that are very stable.

It is possible for the interior of a stagnant lid planet to heat up, if the conductive lid is thick enough and the abundance of radioactive elements is great enough. The mantle potential temperature ($T_p$)—the temperature of adiabatically decompressed mantle—will increase in this case, possibly to the melting point. In this scenario, an inactive stagnant lid planet may become magmatically active again; this may be the cause of periodic resurfacing events, as inferred for Venus, or for the formation of lunar mare. Stagnant lid behavior is thus a very stable mode of heat loss, characterizing dead (Moon, Mercury) as well as active (Mars, Venus) planets.

In general, heat loss through a stagnant lid increases with increasing $T_p$, because the conductive lid thins with hotter interior temperatures. As the lithosphere thins, conductive cooling is supplemented by convective heat loss due to periodic advective breakthroughs such as hot spot/heat pipe volcanism or lithospheric delamination. Both delamination and hot spot volcanism are favored by higher $T_p$ for a stagnant lid planet, and larger, warmer planets like Venus have unstable stagnant lids. The Martian stagnant lid is moderately unstable, disrupted by the large Tharsis volcanoes and the Valles Marineris rift. Linear magnetic anomalies[32] and the presence of rocks similar to andesites[33] in the ancient southern highlands of Mars suggest that plate tectonics may have operated in the ancient past, but no more. Earth's twin, Venus, has a stagnant lid that is very unstable. It is tectonically and magmatically much more active than Mars, with periodic and spectacular resurfacing events[34]. Venusian resurfacing events may mark times of large-scale lithospheric failure, when volcanism is stimulated by sinking of portions of the lower lithosphere. Similar lithospheric failure on Mars probably caused that planet's hemispheric discontinuity, with a densely cratered, ancient lithosphere preserved in the south and a much younger lithosphere in the north[35].

These tectonic possibilities are explored in Figure 3, modified after ref. [25] to show some of the modes of planetary heat loss discussed above as a function of Urey ratio (relative importance of heat production to heat loss) and $T_p$. Earth's Urey ratio is estimated to range from 0.16[36] to 0.65—0.85[37], with most researchers accepting a value of ~0.4[38]. These curves show only some simple possibilities, intended to demonstrate that plate tectonics can only occur when appropriate mantle

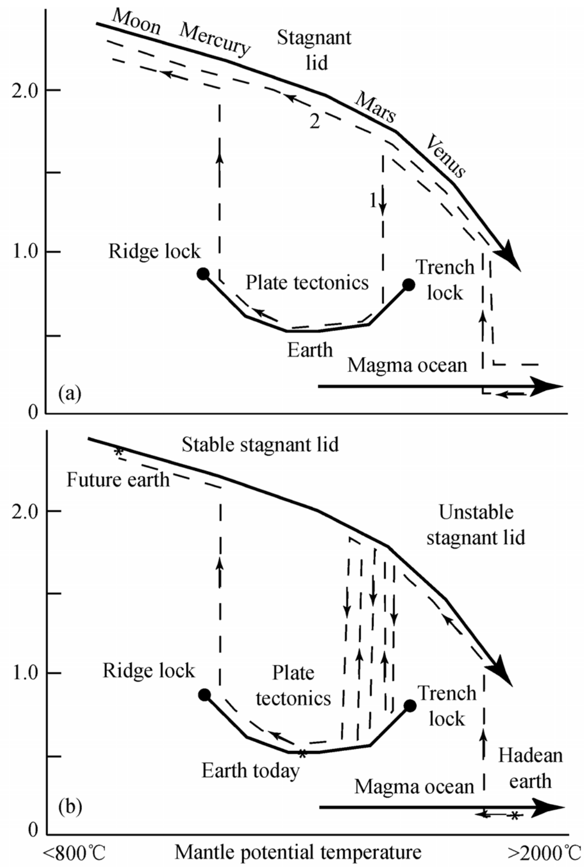


**Figure 3** Diagram showing evolution of cooling silicate bodies through three modes of planetary heat loss—magma ocean, plate tectonics, and stagnant lid—as a function of Urey Ratio (heat production/heat loss) and $T_p$ (mantle potential temperature) modified from ref. [25]. (a) Stagnant lid encompasses a wide range of magmatic and tectonic styles for planets with only one lithospheric plate, from that of essentially 'dead' bodies (Mercury, Earth's moon) to magmatically and tectonically active planets such as Venus; this is simplified as stable (cold) and unstable (hot) stagnant lids in (b). Dashed line shows two possible thermal histories for silicate planets, one with four tectonic mode transitions, from magma ocean to stagnant lid to plate tectonics to stagnant lid, the other remaining as stagnant lid throughout planetary history. (b) Same as (a) but showing relative thermal states and tectonic styles of the terrestrial planets and moon. Dashed line illustrates thermal evolution of a planet with a prolonged transition from unstable stagnant lid to plate tectonics, perhaps like the Proterozoic Earth.

thermal conditions exist. We can be confident that magma ocean and stagnant lid tectonic scenarios exist, although we are only starting to investigate the range of stagnant lid behaviors. These seem to range from very strong, stable lids to unstable lids that are affected by intense magmatism and deformation. The three fundamental tectonic styles—magma ocean, stagnant lid, and plate tectonics—are further complicated by interior processes that are independent of tectonic style, including mantle plumes, large igneous provinces, and lithospheric delamination[39]. It is not clear where some proposed pre-plate tectonic deformation styles fit in this diagram, such as eclogitic "drip" tectonics[40] or catalytic crustal delamination—mantle refertilization recycling[41]; these may be manifestations of a very unstable stagnant lid rather than proto-plate tectonics.

Plate tectonics is a very effective mode of planetary heat loss, as reflected by estimates of Earth's Urey ratio discussed above. This is because spreading at mid-ocean ridges efficiently delivers interior heat to the surface at the same time that subduction zones bring cold surface materials to cool the interior. Nevertheless, plate tectonics may only be favored for relatively short intervals in planetary evolution, where it happens at all, because it is such an effective mode of planetary cooling. In the future, Earth's plate tectonic regime will be shut down by ridge lock, when the mantle becomes too cool to melt by adiabatic decompression and the mid-ocean ridges will no longer act as plate boundaries. Sometime in the past, plate tectonics could not occur because of trench lock. Trench lock occurred because decompression of Earth's hotter mantle generated thick oceanic crust so that oceanic lithosphere was too buoyant to subduct (Figure 3). We do not know how long conditions favoring plate tectonics might have existed on Earth, but these considerations deem it very unlikely that these should persist for the life of the planet.

## 3 Plate tectonics as a self-organizing far from equilibrium complex system

Plate tectonics is a spectacular example of a self-organizing far from equilibrium complex system (SOFFECS). The SOFFECS concept[42] considers that self organization requires an open physical or chemical system, a large steady outside source of matter or energy, non-linear interconnectedness of system components, dissipation, and a mechanism for exporting entropy. Prigogine et al.[42] used convection in a pan of water heated on a stove as an example of self-organization, but it is actually organized from above, by surface tension. Anderson[43] noted that plate tectonics also satisfies these requirements and thus is an SOFFECS. Plate tectonics is now Earth's way to export entropy, it is an open system that draws on the great store of interior energy, and dissipation occurs when the plate bends in subduction zones and is viscously resisted by the mantle (Figure 1). Plate tectonics is a very non-linear response to the peculiarities of how magnesiasilicate solids accommodate and respond to heat loss. Because of the tremendous difference between the mantle potential temperature and that of Earth's exterior, our planet is an inherently unstable, far from equilibrium system. In this sense, plate tectonics is by far the largest SOFFECS on this planet.

The SOFFECS concept encourages us to resist an overly strict application of the principal of uniformitarianism—"The present is the key to the past". Blind application of this principle impedes objective reconstruction of the history and mystery of a truly dynamic Earth. Unique events and episodes like the Cretaceous/Tertiary impact and extinction and the Neoproterozoic "Snowball Earth" demonstrate the limitations of uniformitarianism. Instead we should appreciate that plate tectonics is a special feature of an unusual planet. We know that plate tectonics is an important way that Earth is unique among the planets, and the present uniqueness in space supports the notion that plate tectonics is also likely to be limited temporally. The realization that plate tectonics is an SOFFECS emphasizes this point: Because SOFFECS reflects a very disequilibrium system, they start and stop, reorganizing rapidly and unpredictably. It seems inevitable that the gigantic Earth SOFFECS known as plate tectonics experienced multiple phases before evolving to what we see today.

## 4 Physical requirements for beginning plate tectonics

Given that the negative buoyancy of old oceanic lithosphere drives plate motions today, we must understand how lithospheric density has changed as Earth cooled over the past 4.5 Ga if we are to understand when plate tectonics began. Plate tectonics could not have started until a significant fraction of the lithosphere became gravitationally unstable. Lithospheric instability is necessary but not sufficient, because plate tectonics also requires that lithosphere be weak enough to rupture, founder, and bend, yet strong enough to remain coherent from spreading ridge into the subduction zone.

Oceanic lithosphere density is largely determined by its age (Figure 2). Oceanic crustal thickness and composition (and thus crustal density) is controlled by $T_p$ because this determines how much melting accompanies a given amount of mantle upwelling[44]. The mantle potential temperature for the Archean Earth was significantly higher than today[45], perhaps by 300—500℃. Seafloor spreading and decompression mantle melting of the hotter Archean Earth would generate thicker oceanic crust[46] and this would result in more buoyant lithosphere. It should have taken much longer for this lithosphere to become negatively buoyant than it does at present, making it very difficult to start subduction.

Another question concerns when plate tectonics can be sustained. Because conductive cooling and thickening is a function of age, oceanic lithosphere must ultimately become gravitationally unstable relative to asthenosphere, although for a hotter early Earth this may have taken many tens or even hundreds of millions of years, compared to 20—40 million years today. If the buoyancy crossover time was much greater than it is today, negatively buoyant lithosphere should have been removed by subduction much more rapidly than it could have been produced, causing subduction to stop or become episodic. Today the mean age of oceanic lithosphere is ~100 Ma[47], much older than the buoyancy crossover time of 20—40 Ma, so that there is an inexhaustible supply of negatively buoyant lithosphere and subduction can be sustained indefinitely. Earlier in Earth history, the crossover time was surely greater so that subduction would have quickly exhausted the supply of negatively buoyant oceanic lithosphere. Subduction in this situation would not have been sustainable; it may have occurred episodically, after a protracted period of oceanic lithospheric cooling, or it may not have happened at all.

Considerations of lithospheric density alone cannot resolve when plate tectonics began. The lithosphere must also be ruptured so that asthenosphere can rise above it, allowing the lithosphere to sink. The great strength of the lithosphere, which increases with age and thus gravitational instability, is a fundamental problem for all models of subduction initiation[48]. Many researchers conclude on this basis that subduction zones form where the lithosphere is already ruptured, such as at transform faults or fracture zones[49,50]. The transition from stagnant lid to plate tectonics also requires lithospheric rupture to nucleate subduction, but it is not clear what this would have been. Perhaps a meteorite impact or major episode of delamination caused the initial rupture that ultimately evolved into the first subduction zone.

Water is another important consideration. The fact that most active plate boundaries on Earth are underwater may be why we have plate tectonics and other planets do not. Water weakens rocks, lowers the melting point and lowers the strength of the lithosphere and the viscosity of the mantle[51,52]. Water also is needed to make serpentine which may be key for weakening lithospheric mantle. Certainly the abundance of water on Earth favors plate tectonics but only if lithosphere is gravitationally unstable.

## 5 Geologic criteria for recognizing plate tectonics

The theoretical considerations outlined above motivate critical examination of the geologic record for evidence of the earliest preserved record of plate tectonics. To do this, we must first determine what are the most reliable indications of plate tectonic activity, especially those that are likely to be preserved. We must recognize that some evidence has been removed by surficial erosion and by tectonic erosion at subduction zones[53], or obliterated as a result of deformation or metamorphism. The geologic record is surely incomplete, and we can never know how incomplete it is. Nevertheless, constraints about when plate tectonics began are at least partially preserved in the geologic record. Given these uncertainties, any indications for plate tectonics beginnings must be regarded as a minimum constraint; all we can say is that plate tectonics began no later than the age of the oldest reliable evidence.

We should also be conservative in establishing criteria for the operation of plate tectonics, avoiding especially circular reasoning. Care should be taken not to confuse observations with interpretations (e.g., the observation is pillowed basalts with Nb depletions, the interpretation is formation of a back-arc basin). The mere existence of continental crust, igneous and metamorphic rocks, and deformation of a certain age should not be taken to indicate that these must have been produced by plate tectonic processes. We know that Venus and Mars do not have plate tectonics, yet they experience a lot of igneous activity and deformation, as should have the pre-plate tectonic Earth. Presumably the pre-plate tectonic Earth was dominated by unstable stagnant lid tectonics and magmatism dominated by hotspot magmatism and delamination of the lower crust and mantle lithosphere[39,41,54].

In the following discussion, some of the most robust evidences for the operation of plate/subduction tectonics are presented and discussed: ophiolites, blueschist and ultra-high pressure metamorphic belts, eclogites, passive margins, transform faults, paleomagnetism, igneous geochemistry, and isotopes. There are different ways that these lines of geologic evidence can be considered, not only when the assemblage or property first appeared but also when it first becomes common. These lines of evidence are summarized in Figure 4 and their significance discussed below. It is perhaps not surprising that some of the lines of evidence suggest different times for the start of plate tectonics.

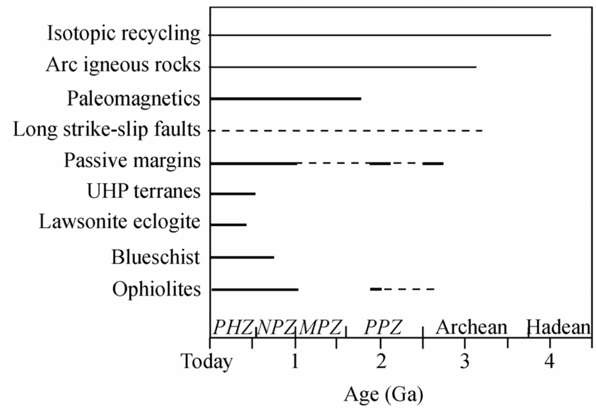


**Figure 4** Criteria for the operation of plate tectonics back through Earth History. Solid lines indicate strong indications that are well dated, thin lines represent more ambiguous indications that are well dated. Dashed lines indicate lesser degrees of confidence about the indicator and/or its timing.

(i) Ophiolites are sections of oceanic lithosphere emplaced on the continents and are reliable indicators of plate tectonic activity. Ophiolites manifest two modes of lithospheric motion expected from plate tectonics: seafloor spreading to form the crustal section and plate convergence to emplace this on the continent. Where ophiolites form is controversial; fortunately this is not important for the present discussion because all tectonic settings for ophiolites require subduction for formation and emplacement[50]. Complete ophiolites, with pelagic sediments, pillowed basalts, sheeted dikes, gabbros, and tectonized ultramafic residues (harzburgite and lherzolite) are rare. The search for the ophiolitic evidence of plate tectonics should include fragmentary ophiolites but all of the components should be present, if disrupted. Pillowed tholeiitic basalts with concave-downward REE patterns and associated with harzburgitic ultramafics containing Cr-rich spinels are a minimum requirement for inferring that a disrupted mafic-ultramafic suite is a disrupted ophiolite. Ophiolites may be readily removed by erosion, because the best preserved examples are emplaced as the uppermost unit of a nappe stack.

The 2.5 Ga Dongwanzi ophiolite of NE China was thought to be the oldest ophiolite[55], but it has been challenged[1,56]. The oldest convincing ophiolites are ~2.0 Ga Purtuniq ophiolite of Canada[57] and 1.96 Ga Jormua ophiolite of Finland[58]; it is probably significant that these are about the same age. Even these ophiolites show interesting differences from more common, Neoproterozoic and younger ophiolites, which generally show "suprasubduction-zone" (SSZ) affinities. Jormua metabasalts are enriched and have no SSZ characteristics and are interpreted to have formed in a small Red Sea-like oceanic rift.

(ii) Blueschist is a metamorphosed mafic rock containing abundant sodic amphibole, which forms as a result of high-$P$, low-$T$ metamorphism[59]. Blueschists are synonymous with "B-type" UHP terranes[59] and are characteristic of Pacific-type orogenic belts[60]. The perception that blueschist only forms in subduction zones is based on their association with ancient subduction mélanges and is confirmed by studies of active subduction zones[61−63]. The best-studied blueschists, those of the Franciscan and Sanbagawa terranes, appear to have been subducted 15 — 70 km deep before returning to the surface[60].

It has been known for a half-century that blueschists are not found in very ancient rocks[64]. It is now widely acknowledged that the oldest blueschists date from Neoproterozoic time, ca. 800 — 700 Ma. These are widely distributed, in West Africa and India[59]. Blueschist of apparent Neoproterozoic age was also reported in western China[59] but its metamorphic age waits confirmation (may have formed in the Paleozoic). Somewhat older (ca. 940 Ma) blueschist was suggested to exist in South China[65], but no further study has verified this important occurrence. It is widely recognized that pre-Neoproterozoic blueschists are unknown. This absence does not seem to be a preservation problem. Because blueschists are preserved in fossil subduction zones, they are not likely to be removed by erosion. The preservation of ~3.2 Ga high $P$/low $T$ metamorphic rocks in S. Africa[66] indicates that blueschists would be preserved somewhere on Earth if they were produced and exhumed in pre-Neoproterozoic time. One might think that the absence of blueschists from the pre-Neoproterozoic record would be widely acknowledged as important evidence for when the modern episode of subduction tectonics began, but this is not the case. Instead, many geoscientists think that the absence of pre-Neoproterozoic blueschist indicates that the pre-Neoproterozoic Earth was hotter so that blueschist was not generated in subduction zones. Surely the early Earth had a somewhat higher mantle potential temperature but this may not have affected the thermal structure of subduction zones and thus whether or not blueschist was produced. The thermal structure of subduction zones mostly reflects the age of the subducted lithosphere and the convergence rate[67], not mantle potential temperature. This controversy might be resolvable by developing robust geodynamic models for subduction zones in a hotter Earth.

(iii) Ultra-high pressure (UHP) metamorphic terranes are important indicators of ancient subduction zones. These form when continental crust is subducted to depths >100 km and return to the surface. Metamorphic assemblages in UHP terranes include coesite and/or diamond and indicate peak metamorphic conditions of ~700—900℃ and 3—4 GPa or more[68]. In contrast to blueschists, the significance of which has long been appreciated, UHP terranes have been recognized for only the last 15 years or so, and new localities will surely be found in the future. The oldest, reliably dated UHP locality is in Mali, where coesite-bearing gneiss was metamorphosed at ~620 Ma[69]. The oldest diamond-bearing UHP terrane is found in Kazakhstan, where diamond- and coesite-bearing paragneiss of the Kokchetav massif was subducted >120 km deep at ~530 Ma[70]. The first evidence for deep subduction of continental crust is thus found in late Neoproterozoic and early Cambrian rocks.

(iv) Eclogites are often thought to reflect plate tectonic processes but there are many ways to generate garnet-clinopyroxene metamorphic and igneous rocks. Lawsonite-bearing eclogites clearly manifest metamorphism in a subduction zone but these are only found in Phanerozoic terranes[71]. Medium *T*-high *P* eclogites and granulites are known from as early as Neoarchean time, but these represent significantly higher geothermal gradients than are found in modern subduction zones[72]. Figure 5 summarizes metamorphic styles through Earth history, showing that the low *T*, high *P* metamorphic style characteristic of modern subduction zones has not been observed for rocks older than Neoproterozoic.

(v) The passive margin record is a powerful constraint. Passive continental margins such as those around the

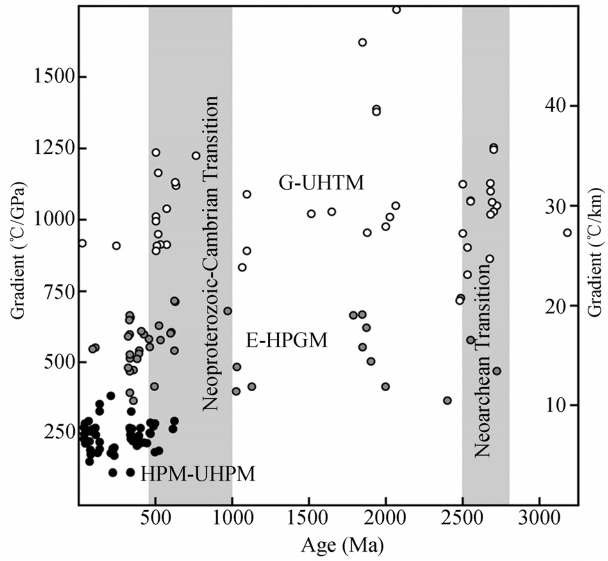


**Figure 5** Plot of apparent thermal gradients versus age of peak *P-T*, for the three main types of metamorphic belts. Open circles: Ultrahigh temperature granulite metamorphism (G-UHTM) belts; gray circles: Medium temperature eclogite-high pressure granulite metamorphism (E-HPGM) belts, and filled circles: lawsonite blueschist-eclogite metamorphism and ultrahigh pressure metamorphism (HPM-UHPM) belts. From ref. [72].

Atlantic Ocean form when continents rift apart and a long-lived ocean basin forms outboard of the continent. They take tens to hundreds of millions of years to form, allowing hundreds of meters of thermal subsidence that creates even greater accommodation space for sediments. The proximity of passive margins to continental sediment sources such as rivers and glaciers ensures that they will contain enormous thickness of sediments. These sedimentary sources are built on very different crusts: continental crust on the landward side, oceanic crust on the seaward side, and transitional crust beneath. Because passive margins are associated with thick continental crust, they are difficult to subduct entirely and so can be expected to be partly preserved in ancient suture zones. Examination of the geologic record indicates that the oldest passive margins date to ~2.7 Ga, but become common first about 2.0 Ga, again about 0.9 Ga (Dwight Bradley, pers. comm. 2006).

(vi) Transform faults—plate bounding strike-slip faults like the San Andreas fault—are one of the three fundamental types of plate boundaries, and the identification of ancient transform faults clearly indicates operation of plate tectonics. Sleep[73] argued that the development of major strike-slip faults at a given time indicated rigid plates and weak plate boundaries and thus should be taken as evidence for plate tectonics.

Well-documented Archean strike-slip faults occur in the Yilgarn and Pilbara cratons of Australia[74]. Sleep[74] argued that the Inyoka fault in southern Africa was the oldest documented strike-slip fault (~3200) Ma but others interpreted it as a thrust. Clearly this sort of analysis must be done with caution. We know that not all strike-slip faults are transforms, only those with significant length and offset. Transforms in oceanic crust can be short but those on continental crust are typically hundreds to thousands of kilometer long, with tens to hundreds of kilometer of offset. Shorter strike-slip faults with smaller offsets form in a wide range of settings which may not reflect plate tectonic activity. For example, strike-slip faulting has been inferred for Venus from radar imagery[75,76]. This is not due to plate tectonic activity but may reflect differential stresses due to downwelling mantle, which is coupled to the lithosphere to produce horizontal compressive stresses and horizontal shearing of Venus' stagnant lid.

(vii) Paleomagnetic measurements could potentially constrain when significant differential motion between crustal blocks began. However, uncertainty increases the farther back in time paleomagnetic measurements reach, and reconstructions of Precambrian continental motions are often uncertain or ambiguous[77]. These uncertainties result from the possibility that Earth's magnetic field has not always been a dipole approximating the geographic poles, the possibility of true polar wander, and the overprinting of primary magnetization by remagnetization associated with alteration and metamorphism. Cawood et al.[4] recently concluded that the existing paleomagnetic database demonstrates differential movements of continents in pre-Neoproterozoic times. They noted that comparisons of coeval paleopoles yields ambiguous longitudes and polarities, but that the paleomagnetic results nevertheless indicate that Australia and Baltica drifted independently between 1770 and 1500 Ma, requiring independent plate motions and thus plate tectonics. Other examples of differential movement of continental blocks in pre-Neoproterozoic times are given in refs. [78－80]. Clearly, careful paleomagnetic studies are important for answering the question of "When did plate tectonics begin?", but the aforementioned complications must be overcome.

(viii) The chemical composition of igneous rocks can be used to infer tectonic setting. In particular, documenting compositions of ancient igneous rocks that are like those produced today in island arcs is taken as strong evidence for the operation of subduction zones and thus plate tectonic processes. Basalts from modern intra-oceanic arc systems show distinctive enrichments or 'spikes' in fluid-mobile elements (e.g., K, Sr, Pb) relative to neighboring elements. These spikes are associated with strong relative depletions in incompatible high-field strength (HFS) cations, especially Nb and Ta[81]. These distinctive characteristics can be inferred from normalized plots of trace elements ("spider diagrams") and are also the basis of a wide range of trace element "discriminant diagrams", widely used to infer the tectonic setting of ancient igneous rocks[82, 83]. Interpreting ancient basalts using this approach is useful for infering the existence of ancient magmatic arcs that formed by subduction, and there are many papers that do this (Cawood et al.[4] offerd several examples). Kerrich et al.[84] inferred Archean examples of convergent margin associations in Archean time, focusing on boninites, Mg-andesites, and adakites.

As is the case for other lines of inference regarding the operation of plate tectonics, care should be taken, specifically to ensure that important chemical characteristics are not the result of alteration, crustal contamination, or are inherited from the subcontinental lithosphere, as is seen for many Phanerozoic flood basalts. This approach applied to ancient rocks is particularly useful for primitive basalts, because these are likely to be less contaminated by pre-existing continental crust and any contamination that does occur is likely to be revealed by isotopic and geochronologic studies. Care must also be taken to assure that spikes of fluid-mobile elements seen on spider diagrams for ancient rocks are not the result of alteration; this concern results in reliance on immobile trace elements, especially HFS cations. The most convincing demonstration that a given suite of ancient igneous rocks formed in association with a subduction zone is to plot Th/Yb and Ba/Yb vs. Nb/Yb or Ta/Yb on the *X*-axis for igneous rocks. Arc magmas form arrays parallel to but distinctly higher than the mantle trend defined by oceanic basalts MORB and OIB[85]. These diagrams are especially useful because these elements are not redistributed unless the rocks are very altered and because vertical trends are readily interpreted to indicate crustal contamination. It is necessary to have high-quality trace element analyses in order to generate these plots, but this is now relatively common now that there are many good ICP-MS analytical facilities. It would be very useful to carry out a systematic study of igneous rocks through time using this kind of plot to see when the arc-diagnostic diagonal trends are first seen in Earth history.

(ix) Isotopic compositions are uniquely powerful for inferring times of crust-mantle differentiation and for identifying deeply recycled surficial materials. There is abundant evidence for early recycling of crust and sediments into the mantle from as far back as the early Archean, including the Pb-Pb array for MORB and OIB. Nd isotopic compositions are consistent with the formation of an early depleted reservoir in the mantle, indicating very early differentiation of the Earth's crust and mantle[86]. Isotopic studies of inclusions in diamonds from eclogites derived from the mantle beneath Archean cratons have isotopic compositions of O, Sr and Pb. These characteristics are best interpreted to reflect seafloor alteration seafloor and recycling into the mantle[87,88]. Mass-independent $^{33}$S isotope anomalies in diamonds are also strong evidence of that these materials were originally at the surface early in Earth history and were recycled into the mantle[89]. The isotopic record clearly indicates that the mantle differentiated very early and that crust and sediments were also recycled very early. The question of whether or not this requires subduction or was accomplished by another convective mode is still open.

The above exercise in identifying geologic criteria for plate tectonics is only a beginning. The list of criteria should be evaluated by interested individuals and modified accordingly. Similarly the timeline for each criteria should be updated frequently and individually. In any case, presenting the criteria and timelines in this fashion is an excellent way to move the discussion forward. Given the criteria and timelines shown in Figure 4, how are these best interpreted? Clearly the isotope record shows that there has been material moving from the Earth's surface and mixing into the mantle from very early in Earth history; this is permissive of but does not require subduction and plate tectonics. The evidence for arc-like geochemical signatures also dates from early in Earth history, strongly suggesting the presence of subduction zones and thus the operation of plate tectonics. Paleomagnetic data indicates differential plate motions at least as far back as Paleoproterozoic time, and the ophiolite record starts about this time. A very important perspective is revealed by metamorphic rocks because this reveals deep subduction. It is thus very important to note that there is no record of deep subduction prior to Neoprotoerozoic time, as indicated by the record for blueschists, UHP terranes, and lawsonite eclogites.

Another approach to the problem is to try and recognize the magmatic and tectonic fingerprints of a pre-plate tectonic Earth in the geologic record. The challenge here is the difficulty of imagining a magmatic or tectonic feature that could not have formed by plate tectonic processes. We should also think broadly about what secondary effects the start of plate tectonics would have had on the Earth System, and look for this as well as the direct evidence. A tectonic revolution of this magnitude should have an effect on the climate (perhaps triggering glacial episodes such as Snowball Earth[90]), The start of plate tectonics may also have affected Earth's moment of inertia and rotational behavior (perhaps triggering True Polar Wander[91]). Other Earth systems may also have been affected by the great tectonic revolution.

## 6 Concluding remarks

The grand question of when and how plate tectonics began offers a richly interdisciplinary avenue of solid Earth research, one that strengthens ties between many strands of the geosciences at the same time that it yields new insights into our planet's history. Because this is an investigation into deep time, the principle of uniformitarianism should be regarded as only applying to chemical and physical processes and not to indicate that SOFFECS like plate tectonics must have always existed. The exploration requires that we more fully engage our imaginations and consult a wide range of disciplines. We should look at the other silicate planets for insights as we consider more fully the range of tectonic styles that Earth might have experienced as it cooled and differentiated. It is possible that plate tectonics began very early in Earth history, perhaps as early as 100 Ma after Earth accreted[92], and has continued since then without interruption. Another possibility is that plate tectonics began relatively late in Earth history—similar to what refs. [3] and [93] conclude—and an active stagnant lid tectonic style existed prior to that time. A final possibility—and the one I prefer—is that the Earth has a more complex tectonic history than we have heretofore appreciated. Geodynamic considerations and the geologic record support an interpretation whereby a tectonic style similar in some ways to plate tectonics—especially in being able to deeply recycle surface materials to depth in the mantle and to generate arc-like magmas, but not driven by lithospheric negative buoyancy or associated with deep subduction—occurred in Archean time. The presence of ophiolites and paleomagnetic evidence for independent plate motions indicates that a short but intense episode of ‘proto” plate tectonics occurred ~1.8 to 2.0 Ga during Paleoproterozoic time — perhaps resulting from collapse of ancient oceanic basins which had finally become negatively buoyant with respect to underlying asthenosphere, but again with no evidence for deep subduction. The Paleoproterozoic episode might not have had deep subduction and was short-lived because of a long buoyancy cross-over time, whereby subduction quickly exhausted the supply of negatively buoyant oceanic lithosphere. After the Paleoproterozoic proto plate tectonics pulse ended, several hundreds of millions of years elapsed before the modern episode (modern style) of plate tectonics began in Neoproterozoic time. Establishment of self-sustaining subduction and plate tectonics late in Earth history reflected the fact that the Earth cooled sufficiently to allow a relatively short buoyancy cross-over time, so that oceanic lithosphere became negatively buoyant after a few tens of millions of years. Clearly the evidence for deep subduction and the first great ophiolite graveyards date from Neoproterozoic time.

These sorts of consideration indicate that the start of plate tectonics may have been a protracted evolution, not a step function. Furthermore, such an explanation better explains the major pulses of crustal growth inferred for the late Archean and Paleoproterozoic[94,95]. Formation of continental growth is expected to mostly occur by formation of arc crust above subduction zones[96], and such crust should grow at an approximately constant rate for as long as plate tectonics is the major crust-forming process[97]. Major increases or pulses of crustal growth are difficult to reconcile with the continuous operation of plate tectonics, but would be expected to accompany brief bursts of plate-tectonic behavior, when large volumes of water and sediment are injected into a hotter mantle and result in brief episodes of unusually vigorous igneous activity.

As we assess evidence for how and when plate tectonics began, we need to think about the likely secondary events and look for this record as well. These should include climatic and rotational (true polar wander) effects that such a tectonic revolution might trigger. I am impressed by the occurrence of major Neoproterozoic climate change and evidence for Neoproterozoic true polar wander and suspect that these are related to the beginning of modern-style plate tectonics[98,99]. The point to be stressed, however, is that the exploration of when plate tectonics began should be broadly interdisciplinary if it is to be successful. The overall effort, focused on critical examination of the Precambrian geologic record and development of refined geodynamic models, promises to lead to an improved understanding of the great solid Earth system. Such an effort may someday lead to clear answers to the questions: When—and how—did plate tectonics begin?

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