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**Invited Review**

# Geological Evidence for the Operation of Plate Tectonics throughout the Archean: Records from Archean Paleo-Plate Boundaries

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**ABSTRACT: Plate tectonics describes the horizontal motion of rigid lithospheric plates away from mid-oceanic ridges and parallel to transforms, towards deep-sea trenches, where the oceanic lithosphere is subducted into the mantle. This process is the surface expression of modern-day heat loss from Earth. One of the biggest questions in Geosciences today is "when did plate tectonics begin on Earth" with a wide range of theories based on an equally diverse set of constraints from geology, geochemistry, numerical modeling, or pure speculation. In this contribution, we turn the coin over and ask "when was the last appearance in the geological record for which there is proof that plate tectonics did not operate on the planet as it does today". We apply the laws of uniformitarianism to the rock record to ask how far back in time is the geologic record consistent with presently-operating kinematics of plate motion, before which some other mechanisms of planetary heat loss may have been in operation. Some have suggested that evidence shows that there was no plate tectonics before 800 Ma ago, others sometime before 1.8–2.7 Ga, or before 2.7 Ga. Still others recognize evidence for plate tectonics as early as 3.0 Ga, 3.3–3.5 Ga, the age of the oldest rocks, or in the Hadean before 4.3 Ga. A key undiscussed question is: why is there such a diversity of opinion about the age at which plate tectonics can be shown to not have operated, and what criteria are the different research groups using to define plate tectonics, and to recognize evidence of plate tectonics in very old rocks? Here, we present and evaluate data from the rock record, constrained by relevant geochemical-isotopic data, and conclude that the evidence shows indubitably that plate tectonics has been operating at least since the formation of the oldest rocks, albeit with some differences in processes, compositions, and products in earlier times of higher heat generation and mantle temperature, weaker oceanic lithosphere, hotter subduction zones caused by more slab-melt generation, and under different biological and atmospheric conditions.**

**KEY WORDS: Archean, tectonics, ophiolite, OPS (oceanic plate stratigraphy), orogeny.**

## 1 GEOLOGIC SIGNATURES OF PLATE TECTONICS

Plate tectonics is recognized through documentation of plate boundary processes including sea-floor spreading, transform faulting, and sinking of oceanic crust and lithosphere at deep sea trenches where oceanic slabs plunge beneath linear chains of magmatic arcs (Wilson, 1965). These discoveries demonstrate creation of new lithosphere at mid-ocean ridges, its lateral transport along strike-slip transform faults, and recycling of this juvenile lithosphere to the mantle at trenches marking subduction zones (Fig. 1a). These processes leave distinctive rock records, documenting the operation of plate tectonics in old terrains (Dewey and Bird, 1970). In the modern tectonic regime, small fragments of oceanic lithosphere are occasionally scraped off subducting oceanic plates, along with their overlying sediments (oceanic plate stratigraphy or OPS; Kusky et al., 2013b), to be accreted above the trenches at convergent margins, forming remnants of oceanic lithosphere known as ophiolites (Coleman, 2012). Some continental material is eroded and deposited in the trenches, or scraped off above the Benioff zones, and brought back to the mantle (von Huene and Scholl, 1993). When these materials reach depths of 100–120 km volatiles including water are released, partially melting the overlying mantle wedge, creating buoyant magmas that rise to create island- or continental-margin arc magmas, with a distinctive geochemical signature reflecting this specific history. In the modern plate mosaic, belts of strongly-deformed OPS with low temperature/high pressure metamorphism form an accretionary prism between the trench

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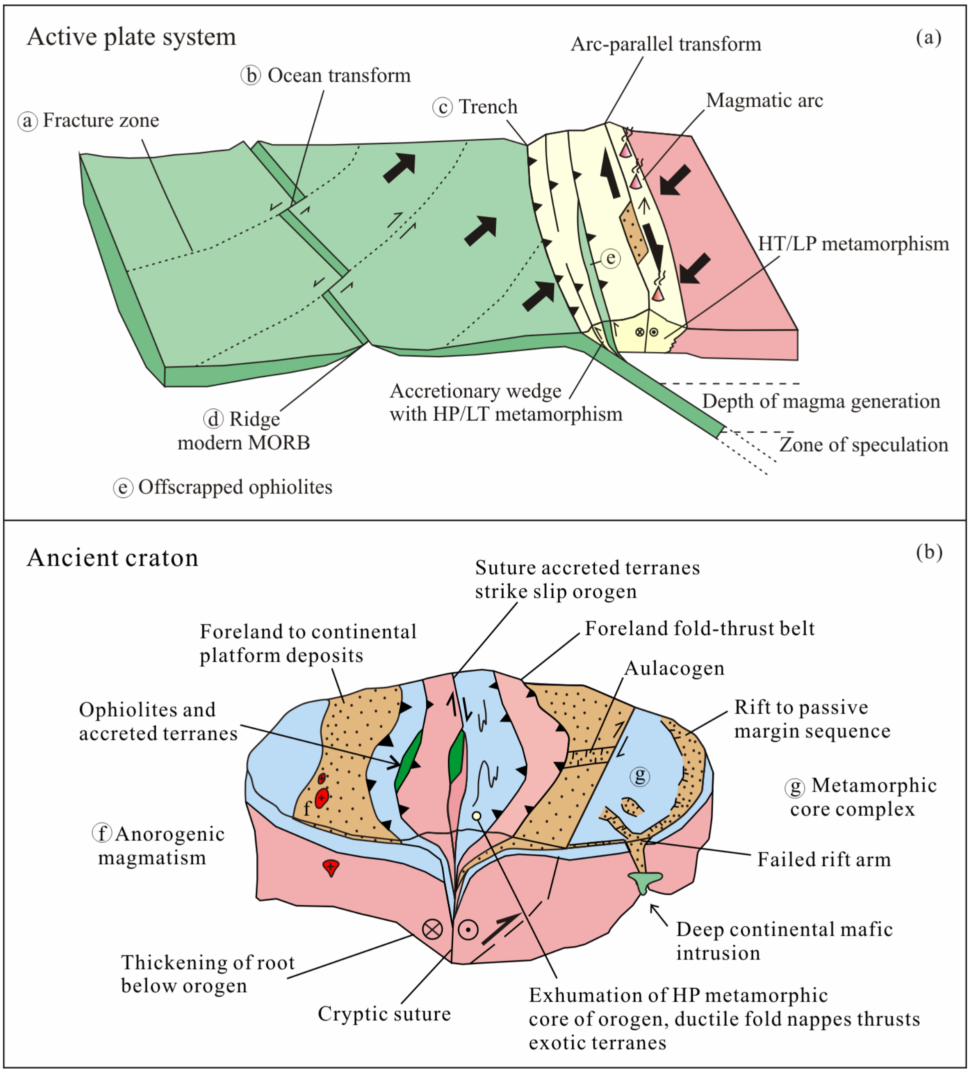


**Figure 1.** Active plate tectonic system and ancient craton. (a) Active plate system. Note how the oceanic lithosphere moves away from the oceanic spreading centers, parallel to transforms, then descends back to the mantle in subduction zones. Water and other volatiles released from the subducting slab at 110 km aid the partial melting of the mantle wedge, creating arc magmas. (b) Ancient craton. When the ocean basin between two continents is consumed and the continents collide, a diagnostic suite of orogenic structures is produced. Note the orogenic core with high-grade metamorphic rocks, accreted terranes and ophiolites, grading outward to foreland fold-thrust belts, foreland basins, then remnant continental platform deposits. Note also on the right side of diagram, the older continent has been rifted, removing an unknown portion of the craton, and leaving a triple-junction-rift-passive margin sequence. Bottom of figures is approximately the Moho.

and the magmatic arc that has high temperature/low pressure metamorphism (Brown and Johnson, 2018; Brown, 2006). This association forms exclusively at island and Andean arcs, and is recognized as one of the hallmark signatures of plate tectonics (Fig. 1a).

Lateral motion of plates brings island arcs and continents into collision, where arc terranes may be added to active continental margins, or continents collide to form orogenic belts characterized by internal high-grade metamorphism, succeeded outwards by accreted terranes, belts of far-travelled nappes, fold-thrust belts, then low-grade foreland basins (Fig. 1b). This distinctive tectonic zonation in orogenic belts forms one of the most-convincing hallmarks of the operation of plate tectonics in the geologic record (Kusky et al., 2016; Şengör et al., 2014; Fritz et al., 2013; Hildebrand, 2013; Windley, 1993; Dewey, 1977; Dewey and Bird, 1970; Wilson, 1968, 1965; Collett, 1927).

## 2 GEOLOGICAL EVIDENCE FOR THE OPERATION OF PLATE TECTONICS IN THE ARCHEAN

The best way to recognize plate tectonics in very old rocks is to systematically document from Archean terranes the same associations of structures, sedimentary and igneous rocks, and metamorphic patterns that are produced by plate tectonics at plate boundaries in young Phanerozoic orogens (Fig. 1b). In this section, we document such plate boundary zone associations from Archean rocks, producing clear evidence for the operation of plate tectonics on Earth throughout the Archean.

Archean extensional plate boundaries are preserved in rare cases as failed rifts in continents, such as the circa 3.1 Ga Pongola structure in South Africa (Gold, 2006; Burke et al., 1985). Successful extensional plate boundaries form oceanic spreading centers, where new oceanic lithosphere is created by mantle upwelling, partial melting, and cooling beneath the world's oceans.

Since no ocean basins older than 200 Ma old are preserved today on Earth, we must seek records of what ancient ocean crust looked like from remnants of oceans that closed, where slices of these oceans were offscraped at paleo-convergent plate boundaries, becoming emplaced as ophiolites on top of continental margins, or within accretionary prisms (Figs. 2a, 2b). Early work (Coleman, 2012; Casey et al., 1981; Anonymous, 1972) suggested that most oceanic lithosphere had a similar structure, grading down from deep-sea sediments to pillow lavas, then a sheeted dike complex, into high-level isotropic gabbros into layered gabbros, then into an ultramafic cumulate section, and finally into depleted mantle rocks including harzburgite, and more rarely lherzolite (e.g., Coleman, 2012). Recent studies have revealed much greater diversity in ophiolites and modern ocean crust on the sea floor (Fig. 2c), leading to new definitions of how to recognize ancient oceanic crust (Coleman, 2012; Dilek and Furnes, 2011; Kusky et al., 2011) from magma-poor types, to magma-rich types, and those associated with subduction zones (forearc and backarc), and those with ocean spreading centers (Furnes et al., 2015). These can be simplified to having a depleted mantle section from which melts were removed to form the overlying crustal section, and should include some plutonic rocks such as gabbros, and volcanics such as pillow basalts. The lavas have the distinct chemical signature of MORB's (mid-ocean ridge basalts), a hallmark of oceanic crust with variations dependent on specific tectonic setting (Fig. 2).

Ocean plate stratigraphy (OPS) is the "sequence of sedimentary and volcanic rocks deposited on oceanic crust substratum from the time it forms at a spreading center, to the time it is incorporated into an accretionary prism at a convergent margin (Kusky et al., 2013b; Kusky and Bradley, 1999; Wakita, 1997; Bradley and Kusky, 1992)". Figure 2a shows the temporal development of typical OPS on the oceanic substratum, as it is transported across the ocean basin in Fig. 2b, to be incoporated into the accretionary wedge where it is imbricated and strongly deformed during accretion. Typical OPS grades upwards from the pillow lava section of the oceanic crust, to deep-sea chert or limestone, banded iron formation, to shales and mudstones, into distal then proximal turbidites (sandstone and shale), then into conglomerates or olistostromes (Fig. 2a). The uppermost turbidites to conglomerates are derived from the erosion of nearby or far-distant continental blocks and transported by axial currents along the trench. When the full sequence enters a trench, it is strongly deformed and repeated

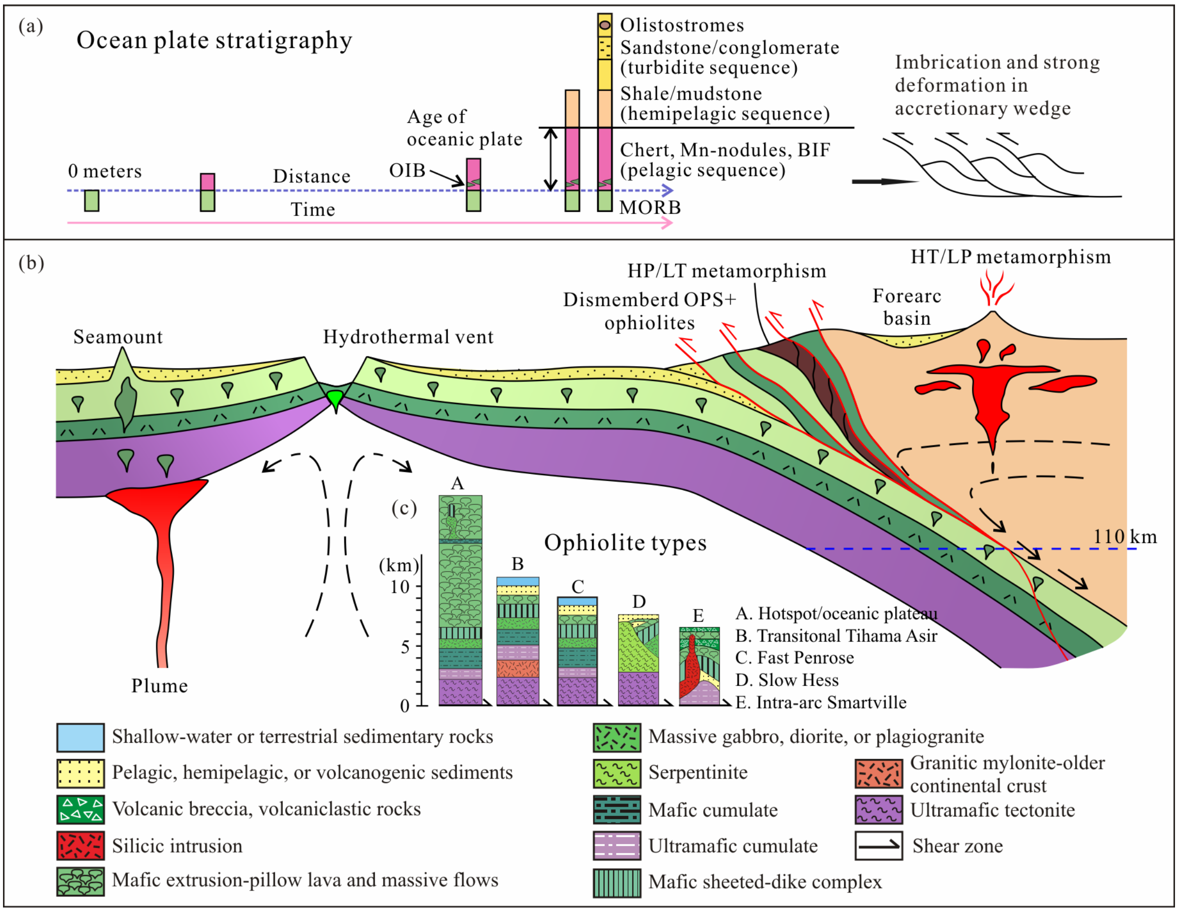


**Figure 2.** Cross section of an ocean basin, showing spreading center, and accretionary orogen above a subduction zone. Panel a shows the development of ocean plate stratigraphy, and its structural disruption as it enters the trench. Panel b shows the oceanic spreading center where oceanic lithosphere is produced, and moves laterally away towards trenches in a ridge-centered reference frame. On the left, the young oceanic lithosphere is shown intruded by off-axis OIB or plume type magmas. On the right, the oceanic plate moves slowly towards the trench, slowing accumulating its OPS, which becomes imbricated, offscraped, and underplated by thrust faults and strongly deformed as it enters the trench. Panel c below the main figure shows different types of oceanic lithosphere that develop depending on the balance between spreading rate and magma supply.

by numerous thrust faults, or deformed so strongly that it forms a tectonic mélange. OPS has now been recognized in orogenic belts of all ages (Fig. 3), ranging from sequences being offscraped at convergent plate boundaries today, throughout the Phanerozoic, Proterozoic, and through the Archean all the way back to the world's oldest preserved rocks in the circa 4.0 Ga Nulliak greenstone belt in the Saglek Block of Labrador (Komiya et al., 2015). In the Nulliak Belt (Fig. 3), MORB-basalts are overlain by a sequence of meta-carbonates, chert, pelites, and clastic rocks (Komiya et al., 2015). This is positive evidence for the lateral motion of oceanic plates away from ridges, accumulating the oceanic sedimentary sequence, and being offscraped and added to the overriding arc or continental plate at paleo-convergent plate boundaries, and of the progressive accumulation of pelagic oceanic sediments, and of the final deposition of clastic, continental- or arc-derived clastic sediments. The whole package is partly offscraped by thrusting in the trench and added to an overriding arc or continental plate at a paleo-convergent plate boundary. Well-documented examples have now been recorded in accretionary orogenic belts of all ages, and thus OPS provides the first-order evidence of the operation of plate tectonics throughout Earth history.

Many Archean greenstone belts have long been known to contain fragments of arcs and ophiolites, but this has been debated (Kusky, 2004; de Wit and Ashwal, 1997), based *inter-alia* on the false preconception that ophiolites must follow the 1973 Penrose definition. Nevertheless, understanding of these rocks has improved considerably in the last two decades so that now many different types of ophiolites re recognized ranging from magma-poor to magma-rich types (Fig. 2c), and from those generated in mid-ocean ridges to those created in supra-subduction settings (Furnes et al., 2014; Kusky et al., 2011). Based on new models of ophiolite structure, some of the best-documented Archean ophiolites include portions of the 2.5 Ga Shangyin-Zunhua Belt of North China (Kusky and Li, 2010; Kusky et al., 2001), the 3.3 Ga Barberton Belt (de Wit et al., 2018), and the 3.8 Ga Isua Belt (Fig. 4).

The 2.5 Ga Shangyin ophiolite is one of the most complete Archean ophiolites (Kusky and Zhai, 2012), containing a well-exposed Moho between mantle harzburgites, a 1–2 km thick mantle transition zone with interlayered harzburgite, mafic and ultramafic cumulates, gabbros, and overlying gabbros and basalts, with local dike complexes (Kusky and Li, 2010). Northern parts of the sequence are extensively intruded by gabbroic, trondhjemitic, and tonalitic sills, and cut by numerous diabase dike swarms. This igneous stratigraphy (Fig. 4) takes into account the fact that the ophiolite and surrounding area has been affected by Mesozoic intrusions (Kusky and Zhai, 2012; Kusky and Li, 2010). The upper section contains silicic fine-grained sedimentary and umber deposits and BIF's, and the whole sequence is dismembered, metamorphosed to amphibolite facies, and intruded by several generations of younger magmas (Fig. 4).

In Barberton, seven major thrust sheets each with distinctive tectonic histories have been delineated (de Wit et al., 2018; de Wit, 2004), each resembling parts of different types of ophiolites formed in modern oceanic backarc-like settings. New field, geochemical and geochronological data (Grosch and Slama, 2017) indicate that the primitive Kromberg massive and pillowed mafic lavas, gabbros, and ultramafic cumulate sequence and overlying cherts are a fragment of an Archean ophiolite, and were not

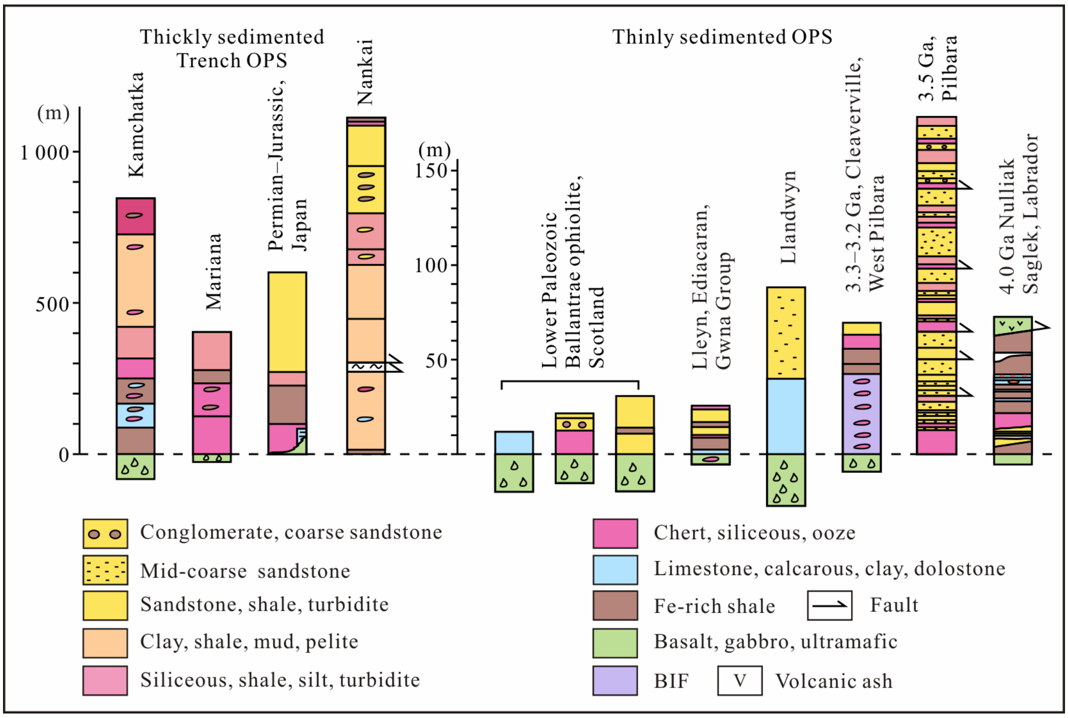


**Figure 3.** Examples of OPS from modern environments, and back through time to 4.0 billion years ago, showing remarkable similarity, demonstrating the operation of the lateral movement of oceanic plates for at least 4 billion years. Data for sections from the following sources: Kamchatka (Kersting, 1995); Mariana (Plank et al., 2000); Japan (Wakita, 2012); Nankai (Shipoard Scientific Party, 2000); Ballantrae (Sawaki et al., 2010); Lleyn (Maruyama et al., 2010); Cleaverville (Kato et al., 1998); 3.5 Ga Pilbara (Kato and Nakamura, 2003); Saglek (Komiya et al., 2017, 2015).

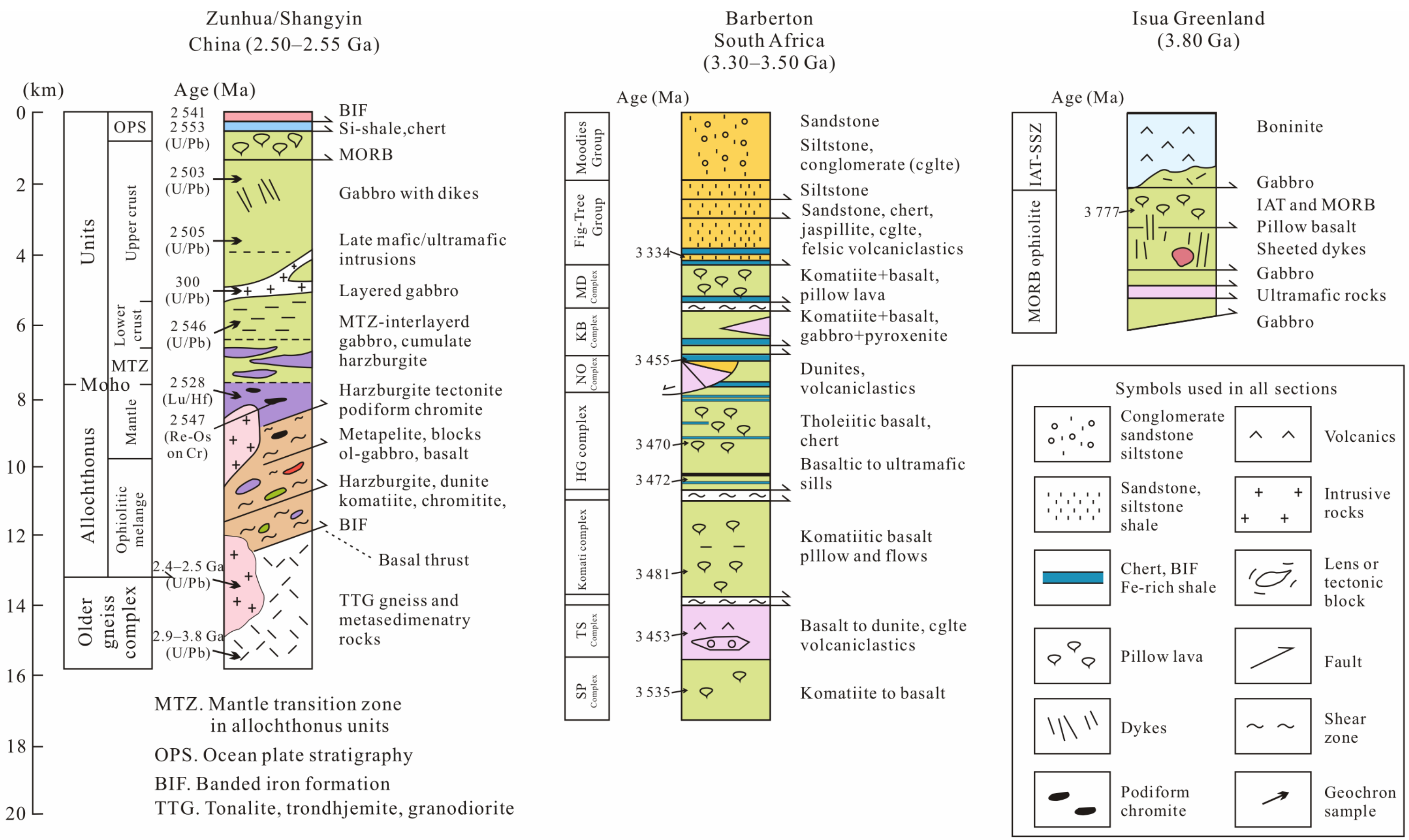


**Figure 4.** Tectonostratigraphic sections through three of the best documented Archean ophiolitic sequences, including the 2.5 Ga Shangyin ophiolite, Barberton, and Isua. Sections are summarized from de Wit et al. (2018), Kusky et al. (2016), and Furnes et al. (2015). Note the significant tectonic disruption of all examples. Abbreviations for the Barberton column include MD, Mendon Complex, KB, Kromberg Complex, NO, Noisy Complex, HG Complex, Hooggenoeg Complex, TS, Threespruit Complex, SP, Sandspruit Complex.

erupted through or deposited on older continental crust. Detrital zircons from the 3.3–3.2 Ga Fig Tree and Moodies groups reveal no evidence for older continental crust in the Barberton source area during deposition (Drabon et al., 2017).

In the circa 3.8 Ga Isua Belt of Greenland (Fig. 4), pillow lavas, possible sheeted dikes, gabbroic and ultramafic rocks, are interpreted as a small fragment of a Paleoarchean ophiolite (Furnes et al., 2007). There are hundreds if not thousands of ophiolitic fragments within Archean greenstone belts (Furnes et al., 2014; Kusky, 2004), so these three examples spanning the entire length of the Archean show that processes of sea-floor extension and magmatism were in operation in the Archean, in a manner very similar to that of today. Accordingly, there is a record of extensional oceanic plate boundaries and plate tectonics throughout the duration of Earth history (Furnes et al., 2015; Kusky et al., 2011; de Wit and Ashwal, 1997).

Large strike-slip faults form at transform plate boundaries, above zones of oblique subduction, or in association with horizontal motions reflecting lateral motions of plates. Phanerozoic examples include the Alpine fault of New Zealand (600 km long), the North Anatolian fault of Turkey (1 500 km long), and the San Andreas of California (1 200 km long). Archean cratons also contain abundant >1 000 km strike-slip faults demonstrating horizontal motion of large rigid crustal blocks in the Archean. Examples include the >1 100 km long 2.7 Ga Quetico fault, which transects the entire length of the Superior Craton (Fig. 5), and is a terrane boundary between the Quetico meta-sedimentary and Wabigoon meta-igneous provinces

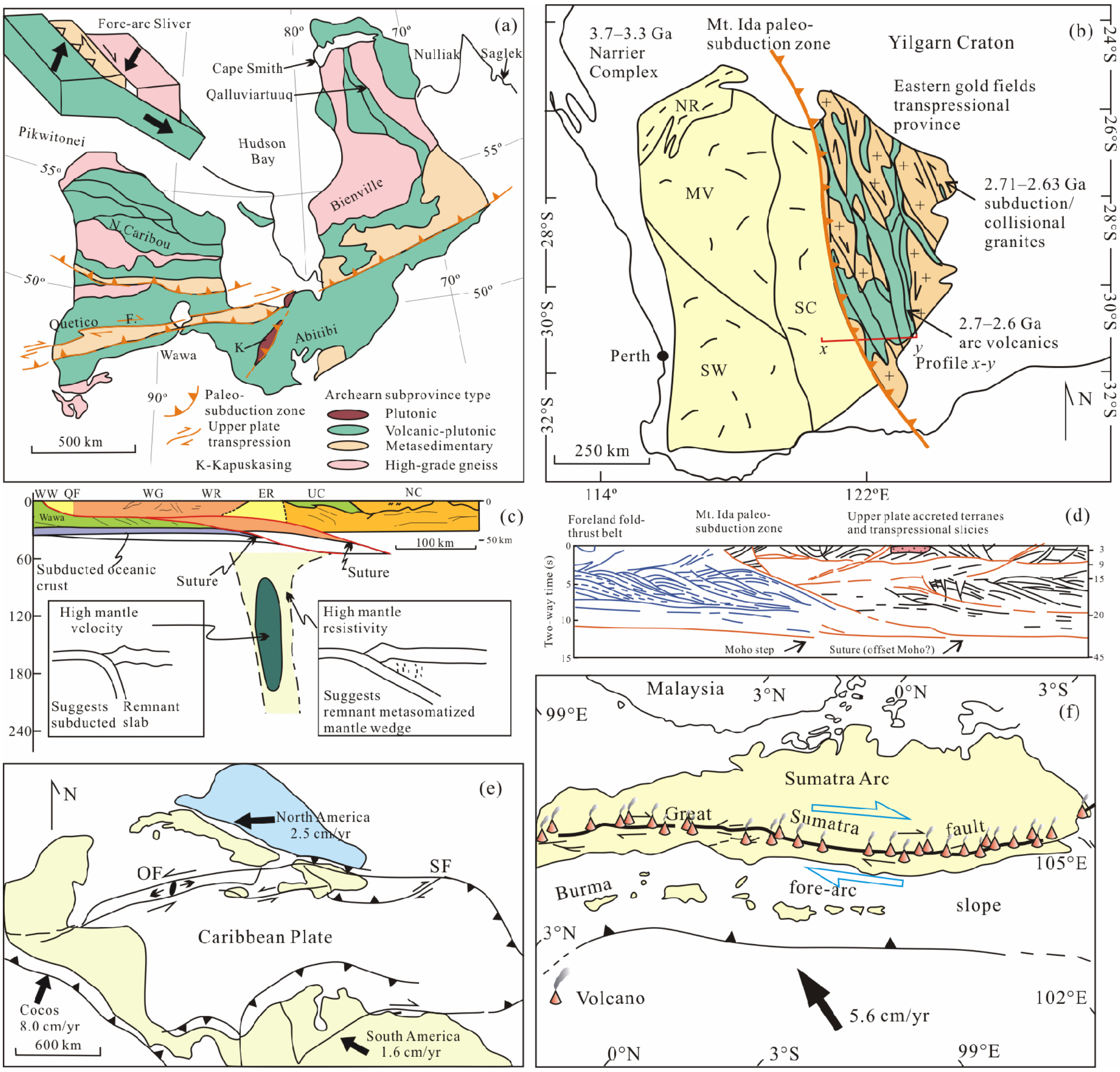


**Figure 5.** Evidence for plate tectonics in Archean terranes includes craton-scale strike-slip faults, and seismically defined paleo-subduction zones that show offsets of the Moho, remnant slabs, and zones of metasomatized mantle above the Archean paleosubduction zones (Percival et al., 2012). (a) Map of the Superior Craton (Percival et al., 2012), showing a paleosubduction zone beneath the Quetico Province (panel c), and upper-plate oblique-slip fault analogous to the Great Sumatra fault (panel f) or the Septentrional-Oriente fault (SF and OF) between the Caribbean and North American plates (panel e). (b) Map of the Yilgarn Craton, and seismic profile (panel d from Drummond et al., 2000) showing suture and offset Moho along the suture between the Eastern Goldfields Province, and western part of the craton, composed of the Southern Cross (SC), Southwest (SW), Murchison (MV) and Narrier (NR) complexes. Abbreviations in panel c as follows: WW. Wawa; QF. Quetico fault; WG. Wabigoon subprovince; WR. Winnipeg River subprovince; ER. English River subprovince; UC. Uchi subprovince; NC. North Caribou superterrane.

(Percival et al., 2012). In the Yilgarn Craton of Australia, the circa 1 100 km long Ida oblique slip fault forms the boundary between the eastern and western Yilgarn terranes (Zibra et al., 2017), and the upper plate to the east is sliced by numerous Archean strike-slip faults, forming a system similar to the dextral Sumatran fault zone (Fig. 5) that extends along the topographic axis of Sumatra and accommodates the oblique component of convergence between the Australia/Indian plate and the overriding Sunda plate (Fitch, 1972), or the Septentrional-Oriente fault (Fig. 5) that accommodates the strike-slip component of strain partitioning between the Caribbean and North American plates (Dolan and Mann, 1998). Thus, Late Archean cratons are marked by strike-slip faults with similar scales and structures as younger plate-bounding transform faults, and the geological record demonstrates lateral movement of crustal blocks by 2.7 Ga (Percival et al., 2012). Preserved parts of older, Eoarchean cratons, are much smaller than the Neoarchean cratons, but also preserve evidence of transform tectonics, on the scale of the preserved cratons. In the Australian Pilbara the 3.5–3.3 Ga Lalla Rookh and Whim Creek belts formed in pull-apart basins along major craton-scale strike-slip or transform faults analogous to the San Andreas and North Anatolian transforms (Krapez and Barley, 1987).

Orogenic belts form in zones of plate convergence, such as above subduction zones, or in wide plate boundary zones where two continents collide. Characteristics of these orogenic belts include early thrust faults indicating crustal shortening (Fig. 6). Early horizontal thrust faults and nappes are known from orogens of all ages (McClay, 2012), including the famous Cenozoic Austro-Alpine nappes, the Paleozoic Appalachian nappes, and in the Mesozoic Cordillera of western North America. Similarly, early horizontal thrust and inverted nappe structures are well-documented in the Early Archean Pilbara and in the North Atlantic cratons, and in the Mid–Late Archean Zimbabwe, Kaapvaal, Yilgarn, Slave, Superior, North China, and Brazilian cratons (Kusky et al., 2016; Cawood et al., 2009; Kusky and Vearncome, 1997). Young orogens such as the Alps and Himalaya exhibit well-defined tectonic zonations, grading from highly-deformed and metamorphosed hinterlands, including accreted arcs and other terranes, through zones of nappes, to foreland fold-thrust belts, and eventually into relatively undeformed foreland basins. In old orogens in Archean cratons, deeper crustal levels are typically exposed, and these orogens have been subjected to further later tectonic overprinting events. Despite this, clear orogenic tectonic zonations similar to those of the Alps and Appalachians have been documented from the 2.5 Ga North China (Kusky et al., 2016), 2.6 Ga Slave (Bleeker and Hall, 2007; Kusky, 1989), 2.7 Ga Brazilian (Hildebrand, 2005), 2.7 Ga Zimbabwe (Kusky, 1998), 3.5 Ga Pilbara (Hickman, 2012) and Yilgarn (Myers, 1995) cratons.

When oceanic plates are subducted, they return to the mantle to be recycled, but dehydration reactions in the subducting slabs

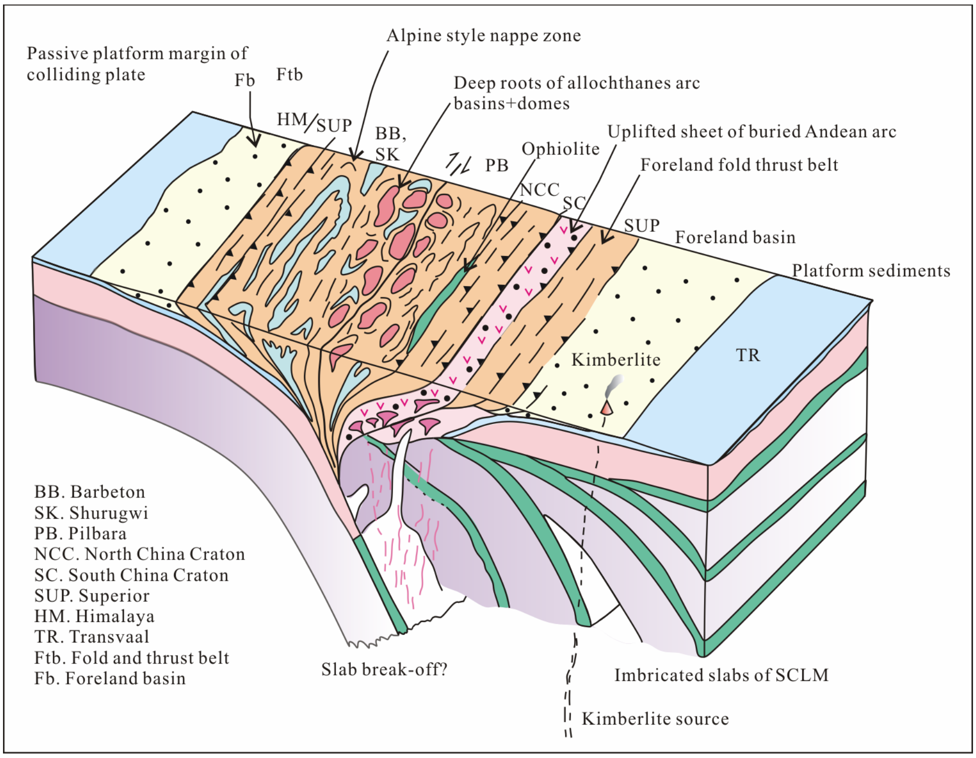


**Figure 6.** Schematic composite model of an orogen, showing the classical tectonic zonation from passive margin sequences, to foreland basins, into fold-thrust belts, obducted arcs and ophiolites, and the high metamorhic grade core of the orogen characterized by intense nappe-style folding, and zones of domal pluton emplacement. Orogenic zonations like this are found in orogens of all ages, suggesting that plate tectonics has operated since the formation of the first orogens at 4.0 Ga. Note that the different locations on the map and section correspond to real places in convergent and collision belts on Earth, including Barberton (BB), Shurugwe (SK). Pilbara (PB), North China Craton (NCC), South China (Yangtze) Craton, Superior Craton (SUP), Himalaya (HM), Transvaal (TR). Imbricated slabs of subducted ocean lithosphere for the root of the craton (based on the model of Kusky, 1993), and slab-break-off and roll back induce thinning of the SCLM and magmatism, after the model and sources cited in Kusky et al. (2014b).

and overlying sediments hydrate the overlying mantle wedge on the way down. At depths of about 110–200 km, sufficient water is released to the overlying mantle wedge to induce partial melting, and these melts rise to form a magmatic arc (Fig. 2), with a distinctive chemical signature as Island Arc Tholeiites (IAT). Rock suites that form in arcs in the present day plate mosaic have analogs with exactly the same lithologies, chemical signatures, rock associations, structural relationships, and tectonic zonations including forearcs, arcs, and backarcs throughout the Archean (Polat, 2012). Magmatic arc petrological and geochemical signatures are well-documented from the Proterozoic and Archean of Australia and Greenland going back to at least 3.1 Ga (Szilas et al., 2016; Korsch et al., 2011; Windley and Garde, 2009), and to at least 3.7 Ga in SW Greenland (Nutman et al., 2015). Thus, there is no doubt that geological processes at paleo-convergent plate boundaries were the same in the Archean as they are today, at least to a depth where the slab reaches 110 km and the release of volatiles generates arc magmatism. Although it is possible to produce "arc-like" geochemical signatures using other pressure-temperature-fluid melting conditions (von Huene and Scholl, 1993), the combination of the structural geology, sedimentology, and volcanology of the fore-arc, arc, and back-arc regions, rock types, and geochemistry strongly argues for a subduction-related origin for arc-like magmas in Archean terranes. Below the arc-magma generation depth is what we colloquially name the "zone of speculation" (Fig. 1) where Archean oceanic lithospheric slabs may have been subducted to the mantle leaving no trace, except perhaps geochemical signatures in the depleted mantle, isolated mineral xenocrysts, or stagnant slabs in the transition zone or along the core-mantle boundary.

In rare cases, samples of deeply subducted Archean oceanic lithosphere have been returned to the surface, where oceanic slabs have underplated the overlying continents forming the sub-continental lithospheric mantle (SCLM) (Kusky, 1993), and kimberlites (Fig. 6) have entrained samples of the underlying eclogite and peridotite as mantle xenoliths (Shirey and Richardson, 2011; Richardson et al., 2001). Further evidence for subduction of oceanic lithosphere in the Archean comes from geophysical data. One of the best examples is from combined deep seismic reflection, refraction and geological data across the Superior Province, that shows clearly three Archean paleo-subduction zones (Fig. 4). The first two are between the 2.7 Ga Quetico and Wawa terranes and the 3.2 Ga Winnipeg River terrane (Percival et al., 2012). These remnant slabs offset the Moho and extend to circa 300 km depth as shown by deep geophysical data (Musacchio et al., 2004; Sol et al., 2002). These data all show strong S-wave anisotropy in the remnant slabs (Musacchio et al., 2004), that is typical of oceanic lithosphere forming subducted remnants beneath Archean cratons (Percival et al., 2012; Kusky, 1993). When traced to the surface, two of these slabs coincide with greenstone belts with oceanic (MORB) pillow lavas, bordered by the metasedimentary, accretionary prism-like Quetico Domain, a clear indication of shallow level convergent margin tectonics to deep subduction in the Archean (Percival et al., 2012). The third corresponds to the boundary between the 3.0 Ga North Caribou and 3.5 Ga Hudson Bay terranes. Additional seismic surveys across the Abitibi Province to the east also show dipping reflections extending 30 km into the mantle, and are interpreted (Calvert et al., 1995) as a remnant 2.69 Ga paleo-subduction zone. Similar integration of geophysical and geological data has revealed fossil, or paleo-subduction zones in the Archean Yilgarn (Fig. 4d), North China, and Slave cratons (Kusky et al., 2014a; Kusky, 2011; Cook et al., 1999).

## 3 RECONCILING ARGUMENTS AGAINST ARCHEAN PLATE TECTONICS

Despite the abundance of geological data that support the operation of plate tectonics throughout Earth history, the apparent absence of some features have been used to suggest that plate tectonics did not start at all until at the time when these so-called diagnostic signatures of plate tectonics have been first documented (Condie, 2018; Maruyama et al., 2018; Foley et al., 2014; Kusky et al., 2013a; Dhuime et al., 2012; Næraa et al., 2012; von Hunen and Moyen, 2012; Rollinson, 2010; Harrison, 2009; Condie and Kröner, 2008; Richardson and Shirey, 2008; Stern, 2008, 2007; Brown, 2007; Smithies et al., 2007; Cawood et al., 2006; Korenaga, 2006; Moyen et al., 2006). The first is the lack of blueschist facies metamorphism in the older record (Brown and Johnson, 2018; Liou et al., 1990; Ernst, 1972). Blueschists record a cold geotherm, characteristic of modern subduction zones where the old cold subducting plates refrigerate the overlying accretionary wedges, leading to high-pressure/low-temperature metamorphic conditions (Ernst, 1973). Blueschists are extremely rare in modern orogenic belts, and many Phanerozoic orogens have none, perhaps explaining their paucity in Archean orogens. Importantly, with 200–300 ºC higher mantle temperatures in the Archean mantle (Korenaga, 2013; Abbott and Hoffman, 1984), and younger average ages of subducting slabs (Abbott and Hoffman, 1984), subduction geotherms would have been significantly warmer in Archean subduction zones, forming greenschists and amphibolites, instead of blueschists and eclogites. After collision, these would have been strongly overprinted by regional medium-pressure/temperature metamorphic conditions. Most Phanerozoic blueschists are in the circum-Pacific and Tethyan orogens that have yet to experience their "final" collisions, and subsequent overprinting by regional medium *P-T* metamorphism that can obliterate all records of previous HP conditions.

Archean eclogites have been known for some time as inclusions in young kimberlites (Fig. 6) piercing cratons (Richardson et al., 2001), interpreted to be entrained from imbricated slabs of buoyant oceanic lithosphere with intervening trapped wedges of fertile mantle (Fig. 6), comprising the SCLM (Kusky, 1993). The reason why these eclogites remain beneath the cratons, with no known examples being exhumed during Archean continental collisions could be related to different buoyancy under slightly higher mantle temperatures, or perhaps they simply have not yet been recognized) (Ganne et al., 2011). Archean eclogites have been reported for some time from the Belomoran massif of Scandanavia (e.g., Dokukina et al., 2014), but these have been widely disputed mostly based on whether the age of the HP metamorphism is Archean or Paleoproterozoic.

When most authorities discussed the role of eclogites in subduction tectonics (e.g., Stern, 2008), they only considered low-T eclogites that are likely to occur, often with low-T blueschists, in the low-grade, upper crustal parts of orogenic belts, but they did not consider the possibility that high-T eclogites may occur in the high-T, high-grade, granulite-gneiss, deep levels of Archean orogenic belts (Fig. 6). In southern India well-authenticated eclogites and garnet websterites occur as lenses and layers up to several meters thick within garnet-rich, chromite-layered, anorthosite-gabbro-ultramafic layered complexes such as the Sittampundi Complex. Garnets in gabbros contain inclusions of omphacite, and calculated phase equilibria indicate that the peak metamorphic assemblage was garnet-omphacite-rutile-melt, which formed at 20 kbar and >1 000 ºC (Sajeev et al., 2009). The crystallization age of the anorthosite is 2 541±13 Ma, and the high-grade metamorphic age is 2 461±15 Ma (Mohan et al., 2013). The data above on ophiolites and HP rocks invalidate the speculative suggestions of Stern (2008, 2007) that plate tectonics did not "start" until the Neoproterozoic, based on the presumed lack of eclogites older than that age.

Other arguments for deep subduction in the Archean come from mineral inclusions in Archean diamonds in kimberlites. Silicate and sulfide inclusions in kimberlitic diamonds (Shirey and Richardson, 2011) older than 3.2 Ga have only peridotitic compositions, but after 3.0 Ga, eclogitic inclusions became common, interpreted to reflect the time of onset of modern style subduction. Nitrogen and carbon geochemical fingerprints of mantle-derived diamonds show that oxidized material has been subducted to the mantle since at least 3.5 Ga, and probably since 3.8 Ga (Smart et al., 2016).

The argument that no ophiolites or ophiolitic mélanges are known in terranes older than 1 Ga is incorrect. 2.5 Ga ophiolitic mélanges are well-documented in the North China Craton (Wang et al., 2016, 2013), the 2.7 Ga Slave (Kusky, 1989), and Superior provinces of Canada (Kusky and Polat, 1999), and at deeper levels would only be represented as banded gneisses in cryptic sutures. Relicts of many greenstone belts are now recognized as ophiolitic fragments in accretionary orogens extending back to 3.8 Ga (Furnes et al., 2014), and OPS can be regarded as a proxy for sea floor spreading to 4.0 Ga (Kusky et al., 2013b), demonstrating the lateral motion of oceanic lithosphere away from ridges inexorably towards trenches (Fig. 2).

## 4 SECULAR CHANGES IN PLATE TECTONIC STYLE THROUGH THE ARCHEAN, BUT STILL PLATE TECTONICS

We present many examples of extensional, transform, and convergent plate boundary structures and rock associations throughout Earth history. The styles of deformation are similar, at all scales, throughout time, and the mineralogical and geochemical components are the same in similar plate boundary settings (Keller and Schoene, 2018). It is clear from the geological record that plate tectonics, in a form similar to that of today, has operated on planet Earth since at least 4.0 Ga, the age of the oldest preserved rocks, as documented by abundant geological, geochemical, isotopic, and theoretical data (Maruyama et al., 2018; Komiya et al., 2017; Nutman et al., 2015; Korenaga, 2013; Polat, 2012; Shibuya et al., 2010; Richardson and Shirey, 2008). Despite this, there are some differences between rocks produced by plate tectonics on the early Earth, and those produced by similar processes in the modern world. The first of these relates to secular cooling of the Earth, which has produced a steady gradual change in the trace element chemistry of magmas in extensional and perhaps other settings (Keller and Schoene, 2018), differences in the thickness of oceanic crust due to higher degrees of partial melting (Foley et al., 2003; Sleep and Windley, 1982), and perhaps a dominance of more-shallow subducting young slabs with more frequent slab break-off events (Foley et al., 2003; von Huene and Scholl, 1993), on a planet with more smaller plates (Abbott and Hoffman, 1984), than in the present plate mosaic. The second main change between the early and modern Earth is the change in the biosphere and consequent chemistry of the oceans and atmosphere (Duncan and Dasgupta, 2017), with resultant changes in tectonic signatures at plate boundaries. In the Early Archean the atmosphere was much more reducing than at present—there were no extensive biogenic carbonate platforms, so carbonates formed by chemical precipitation processes. Oceans were saturated in Si, so that hydrothermal processes at mid-ocean ridges produced silica-rich chimneys and deposits (cherts, BIF's, and the magnetite-quartzite-basalt association common in some greenstone belts) instead of the black smoker mounds and associated sulfide deposits of the modern oceans (Shibuya et al., 2010).

In summary, an analysis of the rock record shows that there is no evidence that plate tectonics did not operate in a manner similar to modern style tectonics on the early Earth, 4.0 billion years ago, much as it does today. The planet lost heat then as now, by making and ageing oceanic crust, moving it laterally away from ridges along transform faults, and returning crustal material to the mantle at subduction zones to be recycled to the deep mantle. Arcs formed above subduction zones, forming more highly differentiated rocks that gradually grew in volume, melted again in later collisions, building the continents that are extant on the planet today.

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