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Plate tectonics: What, where, why, and when?

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A B S T R A C T

The theory of plate tectonics is widely accepted by scientists and provides a robust framework with which to describe and predict the behavior of Earth's rigid outer shell – the lithosphere – in space and time. Expressions of plate tectonic interactions at the Earth's surface also provide critical insight into the machinations of our planet's inaccessible interior, and allow postulation about the geological characteristics of other rocky bodies in our solar system and beyond. Formalization of this paradigm occurred at a landmark Penrose conference in 1969, representing the culmination of centuries of study, and our understanding of the *"what"*, *"where"*, *"why"*, and *"when"* of plate tectonics on Earth has continued to improve since. In this Centennial review, we summarize the major discoveries that have been made in these fields and present a modern-day holistic model for the geodynamic evolution of the Earth that best accommodates key lines of evidence for its changes over time. Plate tectonics probably began at a global scale during the Mesoarchean (c. 2.9–3.0 Ga), with firm evidence for subduction in older geological terranes accounted for by isolated plate tectonic 'microcells' that initiated at the heads of mantle plumes. Such early subduction likely operated at shallow angles and was short-lived, owing to the buoyancy and low rigidity of hotter oceanic lithosphere. A transitional period during the Neoarchean and Paleoproterozoic/Mesoproterozoic was characterized by continued secular cooling of the Earth's mantle, which reduced the buoyancy of oceanic lithosphere and increased its strength, allowing the angle of subduction at convergent plate margins to gradually steepen. The appearance of rocks during the Neoproterozoic (c. 0.8–0.9 Ga) diagnostic of subduction do not mark the onset of plate tectonics, but simply record the beginning of modern-style cold, deep, and steep subduction that is an end-member state of an earlier, hotter, mobile lid regime.

Contents

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1. Introduction

The formulation and eventual acceptance of the theory of plate tectonics in the late 1960s was a monumental turning point for science, which has forever changed the way that we think about the Earth and other extraterrestrial rocky bodies. Amongst other key criteria, the operation of plate tectonics is thought to be a necessary condition for the emergence of complex life (Stern, 2016) and hence the ongoing search for habitable planets outside of our solar system is now deeply entwined with understanding how, when, and why this unusual geodynamic regime initiated on Earth. These questions have been investigated by countless authors and many reviews have been written on the topic in recent years (e.g. Condie and Kröner, 2008; Shirey et al., 2008; Hawkesworth et al., 2010; Korenaga, 2013; Palin et al., 2020 and others); however, there remains much contention. The defining feature of plate tectonics is independent horizontal motion of lithospheric plates across the Earth's surface, which is enabled by sea floor spreading at divergent plate boundaries (Le Pichon, 1968), by strike-slip faulting at transform plate boundaries (Woodcock, 1986), and by one-sided subduction into the mantle at convergent plate boundaries (St-Onge et al., 2013; Parsons et al., 2020; Zhang et al., 2020). As such, reported discrepancies concerning the timing of onset of plate tectonics are intrinsically linked to the different strengths and weaknesses of evidence supporting operation of the Wilson Cycle (e.g. Li et al., 2018; Wan et al., 2020) or independent plate motion and rotation (e.g. Brenner et al., 2020).

In this Centennial review, we summarize the major contributions that have been made to key aspects of this debate since acceptance of the plate tectonic paradigm in the late 1960s, focusing on four fundamental discussion points: *what* defines a plate tectonic regime, *where* does plate tectonics operate, *why* does it occur, and *when* did it begin on Earth? Major unanswered questions that remain in this field of study are then outlined, alongside opportunities that we propose as valuable future research directions. We also provide references to more comprehensive works on each topic where more detailed discussion can be found.

2. Birth of a paradigm

Plate tectonics has been accepted by most scientists since the late 1960s as a reliable description of how the Earth's lithosphere 'behaves'; however, the inception of this paradigm began many years earlier (Romm, 1994). First-order observations of similar shapes of coastlines either side of the Atlantic Ocean have been noted and theorized upon since the late 16th century by explorers such as Sir Francis Bacon. In his 1620 work *Novum Orgaum*, he noted "both the New World [South America] and the Old World [Africa] are broad and extended towards the north, narrow and pointed towards the south", though Bacon made no inference of both having been joined together in the past. Later papers published in the 18th and 19th century by various philosophers and naturalists, including Theodor Christoph Lilienthal and Alexander von Humboldt, continued to document geometric and geologic similarities along each coastline, but attributed their current separation to a Biblical catastrophe (cf. Kearey et al., 2009).

Fundamental discoveries by renowned geologists James Hutton and Charles Lyell in the 18th and 19th centuries marked a transition in scientific thought from "catastrophism", where geological change occurs due to highly energetic events happening suddenly and unpredictably, to "uniformitarianism", where change takes place by lower-energy events occurring gradually over time (Gould, 1965). The concept of uniformitarianism, often encapsulated by the maxim "*the present is the key to the past*", forced the subsidiary implication that the Earth was extremely old, conflicting with estimates of ~20–200 Myr made at the time by Lord Kelvin (cf. Burchfield, 1990). Uniformitarianistic principles were first applied to the idea of "drifting" landmasses by Frank Taylor, an American physicist, in 1910, who presented a hypothesis resembling what is now referred to as continental drift (cf. Le Grand, 1988). In Taylor's model, formerly polar continents were driven laterally towards the equator, creating an equatorial bulge around the Earth and colliding to form approximately east–west trending mountain ranges (e.g. the Alpine–Himalayan orogenic belt). A continent was also suggested to have broken apart to form the Atlantic Ocean. While conceptually close to the truth, Taylor incorrectly suggested that the gravitational pull of the Moon (i.e. tidal forces) was responsible for the continental migrations, which led to his overall hypothesis being discounted by his peers.

In 1912, Alfred Wegener – a German meteorologist – proposed a similar model of horizontal continental motion and expanded on Taylor's ideas by documenting several independent sets of older, "pre-drift" geologic data, supporting the idea that some were previously connected (Wegener, 1912). The most compelling of these arguments involved the continuity of geological structures (e.g. the Cape Fold Belt), stratigraphic sequences, and fossil fauna and flora across the modern-day continental shorelines of South America and Africa (Wright, 1968; Piper et al., 1973). Further evidence was provided by documentation of the current distribution of Permian–Carboniferous glacial deposits and associated striations, which show more sensible orientations and distribution patterns if continents were re-assembled with South Africa centered on the south pole (Opdyke, 1962). Wegener termed this continental assembly Pangaea – literally meaning "all the Earth" – which is now understood to have later broken apart into two supercontinents: Laurasia in the north (North America, Greenland, Europe, and Asia) and Gondwana in the south (South America, Antarctica, Africa, Madagascar, India, and Australasia) (e.g. Olsen, 1997). These continental masses were separated by the Tethys Ocean – the proto-Mediterranean Sea – and surrounded by Panthalassa – the proto-Pacific Ocean (Arias, 2008). Unfortunately, Wegener's ideas were initially rejected by many European and North American geologists, as they required discarding the existing scientific orthodoxy of a static Earth, and due to his theory being based on multidisciplinary data in fields of study that he was not an expert. Small faults were used by prominent scientists at the time to reject the broader-scale hypothesis outright, and a critical limitation was Wegener's inability to provide a plausible mechanism for continental motion (cf. Kearey et al., 2009). Soon afterwards, Holmes (1928) proposed that convection currents in the mantle powered by the heat of radioactive decay may have dragged continents across the Earth's surface, though it is known today that this force has minimal influence on lithospheric plate motion (see Section 5). Nonetheless, this idea, which emerged nearly 40 years before formalization of the theory of plate tectonics, planted the seed for deciphering mechanisms that could explain the wealth of observational data supporting a mobile Earth surface.

Developments in the field of paleomagnetism and radiometric dating during the 1940s and 1950s revealed that many continental igneous rocks preserve magnetic pole positions and orientations that differ from the present day (Keevil, 1941; Holmes and Smales, 1948; Collinson and Runcorn, 1960). Two competing interpretations can be drawn from these data: (1) the Earth's magnetic poles remained static over time, but the continents wandered; or (2) the continents remained fixed as magnetic poles migrated across the Earth's surface. The latter interpretation would be acceptable for data obtained from a single supercontinent, such as Pangea, but cannot account for several discrete landmasses identifying multiple poles in different places at the same time in Earth history, unless the ancient magnetic field was not bipolar. These data thus provided further support for the notion that landmasses may have moved great distances across the Earth's surface over time (Cox and Doell, 1960).

Mapping of the ocean floor during and after World War II revealed a semi-continuous "mid-ocean ridge" (MOR) system more than 65,000 km long that stood tall above the adjacent abyssal plains (e.g. Ewing and Heezen, 1956). In 1962, marine geophysicist Harry Hess studied these maps and developed his seminal theory of sea floor spreading, suggesting that new ocean crust was created at MOR systems and spread out laterally, pushing the continents apart (Hess, 1962). In this model, new oceanic crust formed from upwelling and cooling of magma at ridges, divided in two, and each half moved laterally away from the ridge. Hess hypothesized that sea floor spreading would thus be driven by thermal convection cells in the mantle, and old, cold crust must be destroyed elsewhere on the Earth so that the planet's surface area remained constant. Continued mapping of the oceans ultimately revealed vast bathymetric depressions situated at some ocean margins that were associated with intense volcanic and seismic activity (e.g. Jongsma, 1977). These phenomena were concluded to be consistent with features expected from subduction of oceanic lithosphere at convergent plate boundaries. Further and final support for the sea floor spreading hypothesis came from the discovery of "magnetic anomalies" retained within seafloor basalt, which formed roughly parallel to a central MOR and were symmetrical on either side (Vine and Matthews, 1963). The recognition of transform faults that connect linear belts of tectonic activity (Wilson, 1965) allowed the Earth's surface to be divided into a complex mosaic of seven major and several smaller plates that rearrange continuously like a jigsaw puzzle. Geometrical relationships defined between plates moving across a spherical planetary surface (e.g. McKenzie and Parker, 1967) and more information derived from seismic observations about their behavior following subduction into the mantle (Coney and Reynolds, 1977) refined these geophysical models of oceanic lithosphere formation, evolution, and destruction.

Formalization and widespread acceptance of the plate tectonic paradigm is often agreed to have occurred in 1969 at the Geological Society of America Penrose Conference, Pacific Grove, California, entitled "*The Meaning of the New Global Tectonics for Magmatism, Sedimentation, and Metamorphism in Orogenic Belts*". Many prominent geoscientists outlined observations and interpretations at the meeting and published seminal papers soon afterwards that supported plate tectonics having operated on Earth for many millions of years (Dewey and Bird, 1970; Kay et al., 1970; Minear and Toksöz, 1970; Oxburgh and Turcotte, 1970). Notably, the broad-scale synthesis presented at that meeting has changed surprisingly little since (Le Pichon, 2019); but what was the geological orthodoxy beforehand and how did interpretations of Earth evolution differ? The pre-plate tectonics 'static' model of the Earth interpreted all tectonic features as having formed essentially by vertical movements at specific locations – so-called "geosynclinal theory". The fundamental concepts of this theory were first outlined by geologist James Hall at his Presidential address made to the Geological Society of America in 1857 (cf. Knopf, 1960). In this model, geosynclines were geographically fixed domains of deep subsidence where sediments accumulated and were eventually buried deeply enough for metamorphism and partial melting to occur at their bases. The morphology of a mountain belt thus corresponded to the original location of greatest sediment accumulation in the geosyncline (i.e. the deepest part of the trough). Sub-types of these geosynclines were classified based on whether volcanic rocks were present in the succession: if so, these were called eugeosynclines, and if not, they were called miogeosynclines (Bond and Kominz, 1988). As such, in the context of the plate tectonic paradigm, miogeosynclines would represent basins forming along the passive margin of a continent, which typically contain clastic and biogenic sedimentary rocks (sandstone, limestone, and shale), and eugeosynclines would represent accretionary or collisional orogens containing deformed and metamorphosed sedimentary and volcanic sequences (Shimron, 1980; Palin et al., 2013; Sepidbar et al., 2019).

Many mechanisms were suggested to drive the formation and evolution of geosynclines, but most prominent was 'gravitational sliding', which invoked isostatic warping of sedimentary piles and minor thrusting of different strata along low-angle fault systems (Krebs and Wachendorf, 1973). Alternatively, some scientists supported the idea of a contracting Earth (cf. Dott, 1997), which assumed that our planet formed in a fully molten state and has since been cooling and contracting. Shrinkage of the Earth's outer shell would have caused lateral compressional forces that folded (or crinkled) geosynclinal sedimentary sequences upwards to produce orogenic belts. While both hypotheses involve minor components of local horizontal motion, it is important to note that geosynclinal theory personified the idea of a static (immobile) Earth surface and so struggled to explain many common geological structures and phenomena that are prevalent on Earth today. By contrast, the theory of plate tectonics provides a unified explanation of all the Earth's major surface features and has revealed unprecedented linkages between many fields of study (Condie, 2015; Palin et al., 2020). We explore some of these phenomena in the sections below.

3. What?

What defines a plate tectonic regime and how does this differ from other possible geodynamic scenarios? Multidisciplinary study of rocky bodies in our solar system – including planets, moons, and asteroids of various sizes – shows that a wide range of tectonic regimes may occur at their surfaces (Watters, 2010), and these may transition between states with time as the body cools (Fig. 1). Following established conventions, we emphasize that 'plate' is the colloquial term for a discrete mass of lithosphere (Barrell, 1914), which may be entirely oceanic, entirely continental, or have components of both. The lithosphere – or lid – on a rocky body may be distinguished from its underlying asthenosphere in several ways. For example, a thermal definition can be used based on whether the dominant mode of heat flow is by conduction (lithosphere) or convection (asthenosphere) (Chapman and Pollack, 1977). Alternatively, from a rheological perspective, the lithosphere acts in a rigid manner, whereas the underlying asthenosphere is weaker and able to flow over geological timescales (Walcott, 1970; Doglioni et al., 2011). The behavior of the lithosphere divides geodynamic scenarios into two end members: stagnant and mobile.

Stagnant lid regimes are characterized by significantly lower horizontal surface (lid) velocities compared to internal (asthenospheric mantle) velocities, which differ by around two to three orders of magnitude (Weller and Lenardic, 2018). Many forms of stagnant lid regime are theorized to occur on rocky planets during their lifecycles, all of which allow mass and energy exchange between the surface and interior, but with limited (if any) horizontal displacement (Solomatov and Moresi, 1997; Piccolo et al., 2019, 2020). Recent conceptual models consider that the early Earth was an unstable stagnant lid planet with unsubductable lithosphere, and that mantle overturns were triggered by inefficient coiling of the stagnant lid (Bédard, 2018). As such, stagnant lid regimes may be considered analogous in many respects with the pre-plate tectonic orthodoxy of geosynclinal theory, where almost all tectonic activity occurs due to vertical motion. By contrast, mobile lid regimes are characterized by substantial horizontal motion of lithospheric plates with respect to the underlying asthenosphere (Cawood et al., 2006), which typically have relative velocity ratios of 0.8–1.8 (Weller and Lenardic, 2018). Plate tectonics, as it occurs on Earth today, is the only known form of a mobile lid tectonics in the rocky bodies in our solar system (Poirier, 1982; Head et al., 2002; Wade et al., 2017; Stern et al., 2018), although others can be speculated upon. Mass and energy exchange between a planet's surface and interior is relatively easy in a mobile lid geodynamic regime, with subduction of oceanic and/or continental lithosphere at convergent plate margins continuously transporting volatiles and solid rock into the Earth's interior (Poli and Schmidt, 2002; Rüpke et al., 2004; Weller et al., 2016; Cao et al., 2019; Lamont et al., 2020), and return processes generating new crust at arcs (Hawkesworth et al., 1997; Collins et al., 2016; Li et al., 2020) and divergent spreading centers (Spiegelman and McKenzie, 1987; Sinton and Detrick, 1992; Morgan et al., 1994).

Given the vast amount of observational data that now exist for planets, satellites, and smaller bodies (e.g. Ceres) in our solar system, we are learning more and more about the rich variety of geological

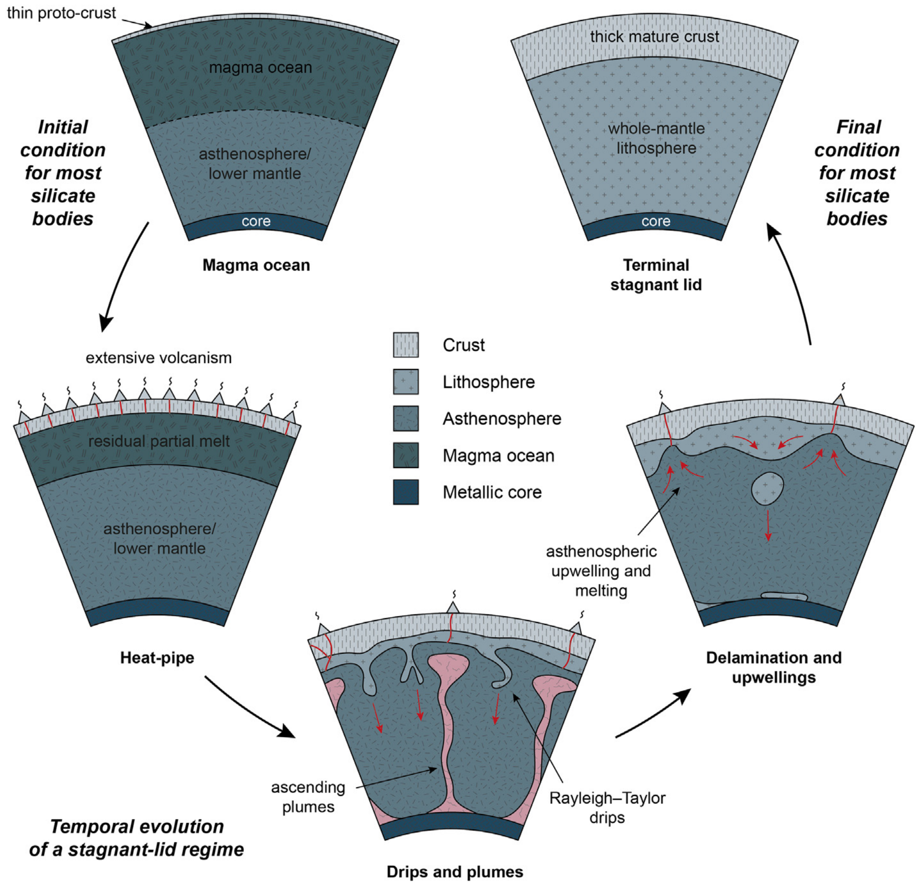

Fig. 1. Temporal evolution of various forms of stagnant-lid tectonic regimes on silicate bodies, such as the Earth. Layer thicknesses are diagrammatic and not shown to scale. Direction of arrows represents a schematic birth-to-death evolution. Modified after Palin et al. (2020).

features that may form on the surfaces of rocky or icy bodies (e.g. Stern et al., 2015). A conceptual tectono-magmatic evolution of a rocky planet over time is shown in Fig. 1 (after Palin et al., 2020). Critically, plate tectonics (i.e. a mobile lid tectonic regime) does not feature in this generic evolution, as it is an 'unexpected' geodynamic state that is thought to require many independent factors to be favorable, such as the presence of surface water (Korenaga, 2020). The initial condition for all rocky planets and satellites that can internally differentiate is that of a magma ocean near to the body's surface above a solid lower mantle and metallic core (Weyer et al., 2005; Elkins-Tanton, 2012). The thickness of this magma ocean depends on body radius; for instance, small bodies with low gravity, such as the Moon, will experience a smaller increase in pressure with depth (dP/dz), and so the peridotite solidus is reached at much greater depth than larger bodies with higher dP/dz, such as Mars (Elkins-Tanton, 2008). Integrated petrological and thermal models of the very early Earth suggest a fully molten magma ocean to shallow depths (~20–30 km) situated above a partially molten crystal-rich mush that extended to a depth of ~300 km (Abe, 1997; Elkins-Tanton, 2012). Complete solidification of this terrestrial magma ocean/mush likely occurred within 1–10 Myr (e.g. Monteux et al., 2016), although this timescale on other planets depends strongly on body size, which controls the surface area to volume ratio (SA/V). For example, Earth has SA/V ~ 4.6×10^{-4}, although the hypothesized magma ocean on 4 Vesta – the second largest body in the asteroid belt, but with a radius just ~9% of Earth's and SA/V ~ 1.1×10^{-2} – is thought to have solidified completely in just hundreds of thousands of years (Neumann et al., 2014). Thus, larger rocky bodies are expected to remain geologically active over significantly longer timescales than smaller rocky bodies.

Crystallization of a primitive terrestrial magma ocean would have proceeded by expulsion of melts towards the Earth's surface, where they may either extrude as volcanic lava flows or solidify during ascent, forming plutons (Mole et al., 2014; Rozel et al., 2017; Piccolo et al., 2020). The earliest stage of the evolution of a stagnant lid geodynamic regime is expected to be dominated by volcanism onto a relatively thin and hot primordial crust that thickens with time. This scenario, with volcanism dominating over plutonism, has been suggested for the Hadean Earth and has been dubbed heat-pipe tectonics (Fig. 1: Moore and Webb, 2013). Continuous eruption of lava and thus repeated burial of older flows causes this primitive crust to thicken, which makes it increasingly more difficult for ascending melts to reach the surface (Malviya et al., 2006; O'Neil and Carlson, 2017). Thus, over time, volcanism becomes subsidiary to intrusive magmatism. Old mafic lavas that are buried during continued igneous activity and crustal thickening will undergo metamorphic transformation to amphibolite and granulite at pressures exceeding ~6 kbar, with garnet stabilizing at lower crustal conditions (>12 kbar; Raase et al., 1986; Palin et al., 2016a). If sufficiently hydrated, these metabasalts will partially melt, and experimental and petrological modeling has shown that they should produce magmas of tonalite–trondhjemite–granodiorite (TTG) composition (Moyen and Stevens, 2006; Martin et al., 2014; Palin et al., 2016b). These felsic melts rise towards the surface of the Earth and may either stall and form plutons in the lower, middle, or upper crust, or erupt onto the surface as lavas and pyroclastic materials. All Archean terranes contain abundant TTG plutons (or metamorphosed versions thereof – gray gneisses), which are thought to represent Earth's first stable continental crust (Martin, 1993; Moyen and Martin, 2012; White et al., 2017), and importantly, as discussed in Section 6, likely did not form via subduction (e.g. Martin et al., 2014; Palin et al., 2016b).

Continued thickening of a mafic crust produces high-density eclogite at >20 kbar (>60 km), which is gravitationally unstable compared to underlying peridotite (Ito and Kennedy, 1971; Aoki and Takahashi, 2004). On the hotter Archean Earth, these lower crustal portions are predicted to ductilely deform and "drip" into the underlying mantle via short-wavelength, density-driven downwellings (van Thienen et al., 2004; Fischer and Gerya, 2016). Mantle plume activity is expected to accompany this regime, with new crust forming via enhanced magmatic activity over regions of upwelling (e.g. Piccolo et al., 2019). This stagnant lid environment dominated by intrusive magmatism into a thick crust instead of the repeated extrusion of lavas is often referred to as a drip-and-plume geodynamic regime, or colloquially "plutonic squishy lid" (e.g. Lourenço et al., 2020). Cooling of the mantle and thickening of newly formed lithosphere is shown via two- and three-dimensional thermo-mechanical models of the Archean Earth to increase the spacing between mantle plumes and to inhibit localized drip-like density inversions (Fischer and Gerya, 2016; Piccolo et al., 2020). Then, high-density eclogite and underlying depleted mantle terminally sink into the asthenosphere via broad-wavelength and large-volume delaminations (e.g. Kay and Kay, 1993; Zegers and van Keken, 2001; Foley et al., 2003; Piccolo et al., 2019). Convective upwellings in the asthenospheric mantle drive decompression melting and continued formation of new mafic crust, which may be buried and melted to form new felsic TTGs in a cyclical process that continually builds new continents (Kamber et al., 2002; Moyen and Martin, 2012; Palin et al., 2016b; Wiemer et al., 2018).

All differentiated planetary bodies – whether they exhibit a stagnant or mobile lid regime while geologically active – will ultimately evolve towards having a global, thick crust and whole-mantle lithosphere as their terminal state (Fig. 1). This is an inevitable result of secular cooling of a planet's interior causing the Rayleigh number to fall below the threshold at which convection is effective, such that any remaining internal heat can only be lost via conduction (Jarvis, 1984; Bunge et al., 1997). In this terminal stagnant lid state, the body geologically 'dies' and is expected to exhibit minimal if any tectonic activity at is surface, although continued cooling and planetary contraction may induce localized deformation or cause reactivation of pre-existing lines of weakness (e.g. Watters et al., 2012; Watters et al., 2016; Valantinas and Schultz, 2020). This tectonic mode is expressed today in the solar system by Mercury and the Earth's Moon (Hauck II et al., 2004).

4. Where?

The question of *where* plate tectonics operates is not straightforward to answer, despite earlier statements that Earth is the only known planet to exhibit this style of mobile lid regime. If plate tectonics is one of many intermediate steps in the ever-changing lifecycle of a silicate body (Fig. 3), there is every possibility that another planet – in our solar system or beyond – may have transitioned into this regime at some point in time and has since transitioned out. Such an argument could theoretically be made for Venus, which experienced a global resurfacing event at c. 300 Ma (Strom et al., 1994); the cause of which remains unknown. Can we confidently ascertain the geodynamic regime(s) that came before if all surface evidence has been erased? Even if relics of Venus' ancient past remain, we will likely not discover them for many years. Focused mapping and detailed laboratory investigation of rocks on the Earth's surface have been conducted for decades, although many points of debate about relatively simple questions remain concerning the evolution of tectonics on our planet. How long, then, might it take to map out, sample, analyze, and interpret the vast geological richness of the Venusian surface in order to come to a somewhat complete understanding of its geological past? While we present this as simply a rhetorical question, it raises the key philosophical issue that '*where*' may be just as readily phrased as '*when*' if we are not discussing the Earth.

Many forms of tectonic activity have been documented elsewhere in our solar system, of which two unique cases occur on the Galilean satellites – the four largest moons of Jupiter. The innermost satellite, Io, is thought to currently exhibit heat-pipe tectonics (Turcotte, 1989; Spencer et al., 2020), as expressed by hundreds of active volcanoes widely distributed across its surface (Spencer et al., 2007). Remote sensing suggests that Io is internally differentiated, and its bulk density indicates the presence of a metallic core, thick silicate mantle, and relatively thin crust (Anderson et al., 1996, 2001). Extensive volcanism requires the existence of a global source of magma at depth below its surface, which is thought by many to be sustained by tidal heating of its solid interior (Hamilton et al., 2013), although some workers suggest that it has a global magma ocean (Khurana et al., 2011). Spectral analyses of eruptions imply very MgO-rich lavas of picritic or komatiitic composition, equivalent to those predicted to form during decompression melting of a hotter Archean terrestrial mantle (Williams et al., 2000) or at the head of a mantle plume (Arndt et al., 1997). As such, Io may represent an analogue for the very early Earth, albeit at a much smaller scale.

By contrast, Europa, the smallest of the Galilean moons, has recently been reported to exhibit a form of mobile-lid behavior that closely resembles plate tectonics on Earth (Kattenhorn and Prockter, 2014), though with some subtle differences. Europa contains a small metallic core, a thick rocky mantle, and a subsurface liquid-water ocean (~80–100 km) immediately beneath a solid H_2O-ice crust (~10–30 km) (Anderson et al., 1998; Kuskov and Kronrod, 2005). Evidence for active cryo-tectonics is provided by the extremely low crater densities across Europa's surface, which implies a very young mean age and so a mechanism for continuous recycling (Bierhaus et al., 2005). Dilational bands with surface features offset symmetrically on either side thus resemble terrestrial MOR spreading zones and provide evidence of new ice generation. Conservation of surface area and 'tectonic' reconstructions of ice-crust plates were interpreted by Kattenhorn and Prockter (2014) to support transport of surface material into the interior of Europa's ice shell along a linear domain, taken to be an analogue of a convergent plate boundary on Earth. Interestingly, active cryo-volcanism has been inferred on the 'overriding' ice crust, thus providing further support for a brittle, mobile, plate-like shell of H_2O-ice situated above a warmer, convecting layer (Sparks et al., 2017). Thus, despite most studies focusing on our neighbor rocky planets that have similar physical and chemical properties to Earth, Europa may instead be the first extraterrestrial solar system discovered to exhibit features closely resembling mobile lid tectonics.

Our neighboring rocky planets Venus and Mars both show a wide variety of geological features on their surfaces, most of which are expressions of various forms of stagnant lid tectonics (Fig. 1; Solomatov and Moresi, 1996; Reese et al., 1998), although others have been debated to represent evidence for plate tectonic-like behavior. Abundant >1000-km-diameter shield volcanoes on Venus (Ernst and Desnoyers, 2004) and lava flows spatially resembling terrestrial flood basalts (Lancaster et al., 1995) both indicate extensive subsurface mantle plume activity, which is common in a "drip-and-plume" geodynamic regime. However, transform faults (Ford and Pettengill, 1992; Koenig and Aydin, 1998), linear MOR-like features (Head and Crumpler, 1987), and asymmetric and curved trench-like depressions (Sandwell and Schubert, 1992) observed in radar maps morphologically resemble surface structures associated with the three types of terrestrial plate margin. Transient spikes in sulfur dioxide contents in the Venusian atmosphere (Esposito, 1984; Marcq et al., 2013) suggest that several regions of the surface are currently volcanically active, indicating that the interior is hot enough to maintain planetary-scale geological activity. Even more importantly, crater counting suggests that the entire Venusian surface is younger than c. 300 Ma (Strom et al., 1994), indicating that all such features formed in the same geodynamic environment, lending support to hypotheses for the early Earth that plate tectonic-like features make form locally within a larger-scale stagnant lid regime (Nimmo and McKenzie, 1998).

Mars exhibits recent (<40 Ma), localized volcanism (Hartmann et al., 1999; Schumacher and Breuer, 2007) and sporadic seismic activity (Anderson et al., 1977; Banerdt et al., 2020), although most planet-wide geological activity is thought to have ceased at c. 3 Ga (Carr and Head III, 2010). Older terranes show clear evidence for active tectonics, metamorphism, and magmatism, with one of the most curious features of the planet being its pronounced hemispheric dichotomy (Andrews-Hanna et al., 2008): here, the northern hemisphere is comprised of low-elevation plains and thin (~32 km) mafic crust, whereas the southern highlands are high-elevation and the crust is much thicker (~58 km) (Wieczorek and Zuber, 2004). The hemispheric boundary has been studied in detail and has revealed much geomorphological evidence for the flow of liquid water, which has led many researchers to suggest that the northern lowlands may once have been covered by a vast ocean (Baker, 1979; DiAchille and Hynek, 2010; Oehler and Allen, 2012; Wade et al., 2017). Given the importance of surface water for stabilizing subduction of oceanic lithosphere, this observation has spurned many investigations into whether Mars has ever exhibited a mobile lid tectonic regime (Sleep, 1994).

Two major geological features on Mars lend support to this hypothesis. First, mafic rocks of the southern highlands preserve elongate linear remnant magnetic anomalies that have alternating polarities (Connerney et al., 2005) and so superficially resemble the magnetic stripes that form on the ocean floor via sea floor spreading on Earth (Vine and Matthews, 1963). This hypothesis is weakened somewhat by the lack of a geometrical 'spreading center' and that their widths are an order of magnitude greater than the stripes observed on Earth (Connerney et al., 1999); however, in rebuttal, it can be argued that these may record similar plate tectonic-like behavior at a much faster rate than on Earth – thus producing thick stripes instead of thin stripes – or that magnetic-field reversal rates occurred over a much longer timescale on Mars. Nonetheless, these features may also be accounted for by non-plate tectonic processes, such as the episodic intrusion of dikes (Nimmo, 2000). The other major geological feature on Mars that shows cursory resemblance to plate tectonic features on Earth is the Valles Marineris trough system, which Yin (2012) suggested is a >2000-km-long and 50-km-wide strike-slip fault zone. Purported evidence for offset comes from displaced impact craters on either flank of the valley, although the trough system has been alternatively argued to have formed due to catastrophic flooding or graben-like crustal collapse due to movement of subsurface magma (Schultz, 1998; Andrews-Hanna, 2012a, 2012b, 2012c). As such, current opinion within the geological and planetary science communities is that Mars did not ever exhibit subduction. Nonetheless, both Mars and Venus hold much promise for understanding the geological processes that operate in stagnant lid tectonic regimes and continued exploration of both planets in the future will return a wealth of new data that may also shed light on the evolution of the early Earth.

A final note to be made concerning '*where*' plate tectonics may operate must mention planets that lie outside of our own solar system – exoplanets. There have been many technological advances in the past 50 years that have substantially improved our ability to locate and quantify physical (e.g. mass and radius), orbital (e.g. period, semi-major axis), and/or geochemical (e.g. atmospheric composition) properties of exoplanets (Seager and Deming, 2010; Marcy et al., 2005). Some such criteria may be used to argue for the operation of plate tectonics; for example, monitoring of an exoplanet's atmosphere may allow detection of sudden spikes of sulphate aerosols injected into it by large explosive volcanic eruptions (Misra et al., 2015). While not diagnostic of plate tectonics, observations on Earth show that explosive volcanism is most commonly related with silica- and gas-rich magmas that form above subduction zones (Eichelberger et al., 1986), as opposed to more silica-poor and effusive volcanism that occurs in large igneous provinces caused by mantle plume activity (White and McKenzie, 1995). As a consequence, both 2D and 3D thermo-mechanical modeling has been applied to rocky exoplanets of various mass–radius relationships to determine the likelihood of convection and/or surface plate motion (O'Neill and Lenardic, 2007; Noack and Breuer, 2014), and thermodynamic modelling of exoplanet compositions has been used to predict their interior mineralogy (Wagner et al., 2011; Dorn et al., 2015; Unterborn et al., 2016; Foley and Smye, 2018; Putirka and Rarick, 2019). Such studies predict that plate tectonics may be inevitable on super-Earths (Valencia et al., 2007), which are defined by having a mass ~2–10 times that of Earth. Other exogenic factors that likely control whether an exoplanet develops mobile lid behavior include its initial internal temperature (Noack and Breuer, 2014), the degree of solar insolation (Van Summeren et al., 2011), and the presence of surface water (Korenaga, 2011).

5. Why?

The question of *why* Earth exhibits plate tectonics is surprisingly well understood in terms of broad-scale geodynamics, although there is still much fine detail to resolve. Indeed, soon after formulation of the plate tectonic paradigm, several studies were dedicated to understanding what drives plate motion, approaching the issue from both observational and theoretical/modeling perspectives. In a landmark study, Forsyth and Uyeda (1975) compared key physical properties of Earth's major tectonic plates and identified certain variables that showed strong positive and negative relationships, while others showed no correlation. The main conclusion to emerge from that study was that the forces acting on the downgoing slab control the velocity of oceanic plates and are an order of magnitude stronger than any other 'edge' or 'body' force. Thus, the sinking of dense oceanic lithosphere into the underlying mantle at convergent plate boundaries appears to be the main driving force for horizontal surface motion (e.g. Carlson et al., 1983; Conrad and Lithgow-Bertelloni, 2004; Coltice et al., 2019). This gravitational edge force – slab pull (F_{SP}) – dominates, although pushing apart newly formed oceanic lithosphere at mid-ocean ridges (ridge push: F_{RP}) also contributes. In addition, convection in the asthenospheric mantle makes a small contribution to driving plate motion by frictionally dragging the underside of the lithospheric mantle (basal drag: F_{BD}), and iceberg-like lithospheric roots that hang down into the asthenosphere may be pushed along by this 'mantle wind' (Kaban et al., 2015). The absolute magnitudes of these competing forces and their relative importance through time has been further refined and quantified by geodynamic modeling; for example, young and hot oceanic lithosphere is more buoyant than old and cold oceanic lithosphere (Afonso et al., 2007; Weller et al., 2019), such that the magnitude of F_{SP} may evolve as a subduction zone matures (Conrad and Lithgow-Bertelloni, 2002), and higher temperatures within the Archean mantle (below) would have reduced mantle viscosity and thus absolute values of F_{BD} (Artemieva and Mooney, 2002).

Convection is a fundamental characteristic of the mantle and facilitates cooling of the Earth over time (e.g. Hanks and Anderson, 1969; Davies, 1993; DeLandro-Clarke and Jarvis, 1997; Korenaga, 2003; Labrosse and Jaupart, 2007). As temperature changes both horizontally and vertically through the Earth's mantle, and the absolute depths of the Moho and lithosphere–asthenosphere boundary vary according to tectonic setting (Karato and Karki, 2001; Anderson, 2000; Profeta et al., 2015), discussion of secular cooling of the Earth and other rocky bodies in our solar system requires use of a common reference frame. As defined by McKenzie and Bickle (1988), the mantle potential temperature (T_P) is the adiabatic extrapolation of a mantle geotherm to a planet's surface in any given geological environment; for example, the mantle T_P above a mantle plume would be higher than the mantle T_P for a divergent plate margin (mid-ocean spreading ridge). The T_P for ambient mantle reflects interplay between heat lost due to convection in the asthenosphere and/or conduction through the lithosphere, and heat gained due to radioactive decay of heat-producing elements in the mantle and conductive heating from the core (e.g. Korenaga, 2011). Importantly, because the efficiency of each of these parameters varies with time, ambient mantle T_P must also have changed simultaneously since formation of the Earth (Fig. 2), alongside the ratio of internal heat generation in the mantle compared to mantle heat flux, called the convective Urey ratio (Ur) (Korenaga, 2008a, 2008b).

The magnitude and rate of change of mantle T_P can be constrained in numerous ways, and so allows estimation of the value of Ur at different points in geological time. The Phanerozoic value of Ur is estimated to be 0.23 ± 0.15, and thermal models that extrapolate it backwards through the Proterozoic, Archean, and Hadean eons (Korenaga, 2008a, 2008b) produce a concave-upwards mantle T_P curve that has a maximum value at c. 2.8–3.2 Ga between ~1675 °C (Ur = 0.23) and ~1575 °C (Ur = 0.38) (Fig. 2). As today's ambient mantle T_P is ~1350 °C (Herzberg et al., 2010), these thermal models predict cooling of ~75–100 °C/Gyr, although this intrinsically relies on geochemical assumptions of the Earth, such as it having chondritic concentrations of radiogenic heat-producing elements (Leitch and Yuen, 1989). This thermal modeling exercise has been supported in recent years by analytical petrology. For example, the chemistry of unaltered mantle-derived magmas is an excellent recorder of the physical conditions present in their source region at the time of extraction from their residue (e.g. Cone et al., 2020), including temperature and pressure. This was exploited by Herzberg et al. (2010), who calculated the liquidus temperatures for a small dataset of non-arc basalts of various ages, which fall roughly between curves for Ur of 0.23 and 0.38, implying Archean upper-mantle T_P values around 1500–1650 °C (Fig. 2). Similar calculations for komatiites require liquidus temperatures up to 1800 °C, consistent with independent evidence that such lavas form due to mantle plume activity (Campbell et al., 1989). It should be noted that other studies have applied similar petrological analysis to larger datasets (e.g. Condie et al., 2016; Ganne and Feng, 2017) and concluded that ambient Archean mantle T_P outside periods of supercontinent formation was colder than estimates provided by Herzberg et al. (2010); potentially as low as ~1350–1500 °C (Fig. 2). This would define a less pronounced secular cooling rate of ~30–50 °C/Gyr. While an absolute difference of ~150–200 °C in predicted mantle T_P seems small, it has profound implications for the viability of subduction during the Archean, and so the operation or not of global plate tectonics (e.g. Gerya, 2014; Piccolo et al., 2019).

The structure and composition of oceanic lithosphere created at divergent spreading centers is controlled by mantle T_P (McKenzie and Bickle, 1988) and so the viability of subduction initiation and the transition from any form of stagnant lid tectonics to a stable, global form of mobile lid tectonics is fundamentally linked to the thermal history of the Earth. Petrological modeling of melting in the mantle and construction of new oceanic lithosphere at divergent plate margins has been performed by many workers. For detailed reviews of these processes, the reader is referred to Langmuir et al. (1992), Kinzler (1997), and Asimow et al. (2004); however, in relation to secular change and the initiation of plate tectonics on Earth, it is sufficient to note that a cold present-day mantle T_P produces a thin and low-MgO oceanic crust, whereas a hotter Archean mantle T_P produces a thicker and high-MgO oceanic crust (Ziaja et al., 2014). The petrophysical and geodynamical implications of such a secular change in oceanic crust composition are profound, impacting the lithologies that form in descending slabs (Palin and White, 2016; Palin and Dyck, 2018), and so their density and material strength (McNutt and Menard, 1982; Weller et al., 2019). Comparative thermo-mechanical modeling of Phanerozoic and Archean oceanic lithosphere has suggested that a hotter Archean mantle reduced the buoyancy contrast between oceanic lithosphere and underlying asthenosphere (e.g. Van Hunen and Moyen, 2012). Generation of a relatively thick and strong (more depleted) mantle lithosphere and relatively thick and weak (hotter) oceanic crust in the Archean would have produced mechanically weak subducting slabs that experienced frequent losses of coherency (Van Hunen and van den Berg, 2008), thus breaking apart at shallow depths and developing an episodic style of Archean subduction, with a typical duration of a few Myr (Moyen and Van Hunen, 2012). Similar petrological calculations of density variation according to metamorphic phase transformations also predict that the thicker Archean oceanic lithosphere was primed to subduct (e.g. Weller et al., 2019), although likely not at steep angles (>10°) that characterize most Phanerozoic convergent margins (Syracuse et al., 2010).

The issue of identifying how and why subduction could initiate on the early Earth is separate from constraining the petrological and geodynamic conditions that are needed for it to become self-sustainable. Numerical modeling shows that one-sided subduction consisting of a downgoing slab and an overlying arc requires a low-strength zone to form at the plate interface (Hassani et al., 1997; Tagawa et al., 2007) with an effective coefficient of friction <0.1. Dry rocks are unable to achieve this condition, indicating that aqueous fluids must be present at convergent plate margins to 'lubricate' plate motion (Gerya et al., 2008). This is readily achieved on Earth where liquid water

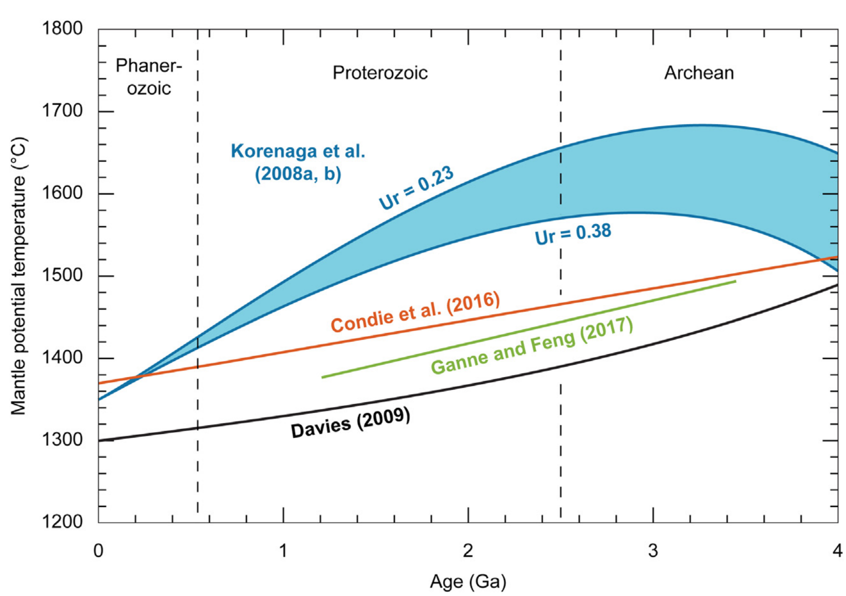

Fig. 2. Proposed variation in ambient mantle potential temperature since c. 4 Ga based on various petrological and geodynamic proxies and models. Present-day mantle potential temperature is 1350 ± 50 °C. Ur = convective Urey ratio. See main text for discussion of data and data sources.

has been present on the planet's surface since c. 4.4. Ga (Wilde et al., 2001; Maruyama et al., 2013) and may be transported to various depths within subduction zones as pore fluids in sediments and sedimentary rocks (e.g. mudstone; You et al., 1996), structurally bound water in hydrous minerals, such as chlorite and amphibole, in hydrothermally altered and metamorphosed oceanic crust (Katayama et al., 2006; Palin et al., 2014; Hernández-Uribe and Palin, 2019a), and serpentine in metasomatized mantle lithosphere (Hyndman and Peacock, 2003; Ranero et al., 2003; Coltorti and Grégoire, 2008). Metamorphism during burial of these lithologies causes dehydration and pulse-like release of H_2O, CO_2, and other volatile species (e.g. halogens) at fore-arc and sub-arc depths (Poli and Schmidt, 1995; van Keken et al., 2011; Hernández-Uribe and Palin, 2019b). Transport of water into the deep mantle may be achieved by its incorporation into nominally anhydrous minerals, such as olivine and clinopyroxene (Karato, 2003). The ability for a planet to acquire and retain surface water over geological timescales thus appears to be a critical factor for determining the viability of plate tectonics (Regenauer-Lieb et al., 2001; Lécuyer, 2013; Wade et al., 2017) and should be considered alongside other important astronomical factors when predicting the habitability of planets outside of our solar system.

6. When?

When did plate tectonics initiate on Earth? Given that independent plate motion must be facilitated by a global network of plate boundaries, evidence for isolated occurrences of subduction at any point in time is not enough to justify the operation of this planet-wide geodynamic regime. This sobering fact is undoubtedly the reason for such contention in the literature, where the wide range of interpretations of the timing of onset of plate tectonics presented in Fig. 3 stems from the reliability of different lines of evidence for satisfying this 'global' criterion. Additional opaqueness comes from the likely interpretation that the initiation of subduction occurred over an extended period, and so it is unreasonable to assign a well-defined age to this onset, as is often the case in the literature. In a recent review of secular change, Palin et al. (2020) divided indicators of subduction preserved in the geological record into three groups – petrological, tectonic, and geochemical/isotopic. Although not data in the purest sense of the word, the results of thermo-mechanical (geodynamic) and/or petrological modeling can also be used to interpret the timing of subduction initiation by comparing simulation output to real-world observations. We follow this scheme again here by briefly outlining evidence for plate tectonic processes for each category, and the strengths and weaknesses of each.

6.1. Petrological evidence

Petrological evidence for plate tectonics comprises rocks that form only in convergent plate margin settings, including those belonging to the downgoing slab and the overlying arc. If these lithologies are discovered in the rock record and can be reliably dated using geochronology, they would represent firm evidence of subduction having operated at that point in Earth history, although this need not have been at a global scale. Associated evidence of horizontal plate motion may be inferred from rocks that are diagnostic of oceanic spreading ridges – for example, mid-ocean ridge basalt (MORB) and associated sheeted dike complexes – although these features alone do not require subduction to be operating elsewhere on a planet, as lithosphere may be readily destroyed in non-plate boundary settings to preserve surface area, as shown in Fig. 1.

A key group of petrological indicators used to identify subduction is high-pressure/low-temperature (HP/LT) metamorphic rocks, such as blueschist and jadeitite (e.g. Stern, 2005; Stern et al., 2013). These lithologies form exclusively in subduction zones along geothermal gradients of ~150–440 °C/GPa (Fig. 4; Ernst, 1988; Sorensen et al., 2006; Palin and White, 2016) due to metamorphism of hydrated basalt and metasomatism of the mantle wedge just above the slab interface, respectively. Blueschists and jadeitite also often occur as exotic blocks in serpentinite-bearing mélanges, confirming a subduction zone environment of formation (Tsujimori and Harlow, 2012). Eclogite is often considered within this HP/LT category of rocks that are diagnostic of subduction zone metamorphism, although deeply buried mafic roots

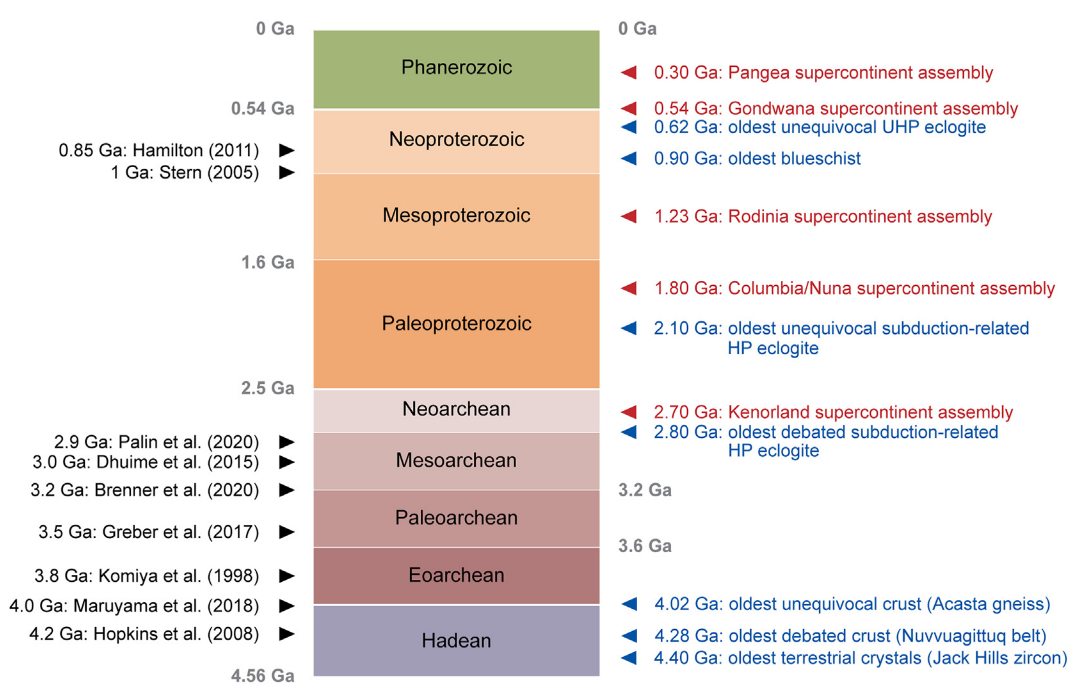

Fig. 3. Representative set of proposed ages for the onset of plate tectonics on Earth. Selected global-scale tectonic events and petrological milestones are included for reference. See text for discussion of key features. Ages are from Komiya et al. (1999), Nutman et al. (2002), Van Hunen et al. (2004), Hopkins et al. (2010), Greber et al. (2017), Ernst (2018), Maruyama et al. (2018), and Palin et al. (2020).

of overthickened continental crust – such as in the Pamir region of the Himalayan orogen (Hacker et al., 2005) – show that some eclogite can form by extreme crustal deformation and thickening (Austrheim, 1991). Nonetheless, eclogite with either MORB geochemistry or containing the minerals lawsonite and/or glaucophane is diagnostic of metamorphism in subduction zones (Becker et al., 2000; Palin and Dyck, 2018). The polymorphic transition of quartz to coesite defines the boundary between HP and UHP metamorphism, which occurs at ~ 26 kbar at 500 °C and ~28 kbar at 900 °C kbar (Chopin, 1984; Liou et al., 2004). As such, coesite-bearing eclogite represents exceptionally deep burial and exhumation of crustal materials from a depth of at least 100 km in the mantle, which is difficult to explain without invoking steep subduction of oceanic lithosphere (e.g. Jahn et al., 2001).

The oldest blueschist, UHP eclogite, and jadeitite on Earth are c. 0.8–0.7 Ga (West Africa, India, and western China: Maruyama et al., 1996), c. 0.63 Ga (Pan African orogenic belt, Southwestern Brazil; Liou et al., 2009), and c. 0.47 Ga (Oya-Wakasa, Japan; Nishimura and Shibata, 1989), respectively, although HP eclogite, with or without MORB geochemical signatures, occurs in several Paleoproterozoic terranes worldwide (see below: Fig. 4). The marked increase in abundance of these HP/LT rock types during the Neoproterozoic has been attributed to many factors, including a late onset of global subduction at that time (e.g. Stern, 2005). However, in light of other lines of evidence suggesting plate tectonics having begun prior to c. 0.9 Ga (Fig. 3), preservation bias likely also plays a key role in overprinting older occurrences (Whitney and Davis, 2006), a change in exhumation mechanism may have taken place during the Neoproterozoic (Agard et al., 2009; Palin et al., 2020) such that older examples were unable to return to the Earth's surface, or else the hotter Archean mantle (Fig. 2) may have increased subducted slab-top geotherms so that diagnostic low-temperature minerals, such as glaucophane and lawsonite, could not stabilize (cf. Early and Late Archean subduction zone geotherms in Martin and Moyen, 2002). A recent model, supported by observed secular changes in basalt compositions through time (Keller and Schoene, 2012; Furnes et al., 2014), is that a cooling of the mantle and an associated decrease in the maficity of oceanic crust through time gradually allowed sodic amphibole (glaucophane) and lawsonite to stabilize in Neoproterozoic (and younger) low-MgO hydrated basalt (Palin and White, 2016). In this scenario, older Paleoproterozoic and Archean high-MgO hydrated basalt would have formed actinolite and chlorite-rich assemblages at equivalent HP/LT subduction zone conditions (Palin and Dyck, 2018), which resemble greenschist-facies assemblages that occur throughout Archean greenstone belts. Thus, detailed geochemical investigation and thermobarometry are required to assess whether inconspicuous greenstone-like units in ancient terranes record hidden evidence of subduction-related HP metamorphism (e.g. François et al., 2018).

A final note on this topic must be made concerning the relevance of HP vs. UHP eclogite as an indicator for the operation of plate tectonics. Geodynamic arguments (see Section 6.4) suggest that subduction on a hotter early Earth would have occurred at shallow (<10°) angles – if it did at all – such that it may have been impossible for subducted crust to reach UHP conditions. Even if subduction operated at high angles, it is also notable that hotter Archean slabs are predicted to have been mechanically weaker than colder Phanerozoic counterparts, meaning that they would lose coherency during subduction and break apart at shallow depths before reaching the HP–UHP transition (Van Hunen and Moyen, 2012). Detached and eclogitized slab fragments that had transformed to become denser than the surrounding mantle and would terminally sink into the deep Earth (Aoki and Takahashi, 2004), achieving UHP metamorphic conditions, but never able to return to the surface for study. By contrast, detached but partially eclogitized (buoyant) fragments would ascend towards the surface, recording HP peak metamorphic conditions. Thus, the presence of coesite should therefore be viewed as sufficient, but not necessary, for identifying steep subduction during the Archean. With this in mind, it is notable that several HP eclogites occur in Paleoproterozoic terranes (Fig. 4) with ages c. 1.8 Ga to c. 2.1 Ga, including the Congo Craton, Democratic Republic of the Congo (François et al., 2018), the Nagssugtoqidian Orogen, south-east Greenland (Müller et al., 2018a, 2018b), and the Trans-Hudson Orogen, Canada (Weller and St-Onge, 2017). Although individual occurrences of HP eclogite may be viewed as potential evidence for localized subduction systems, similar lithologies forming in multiple terranes in the

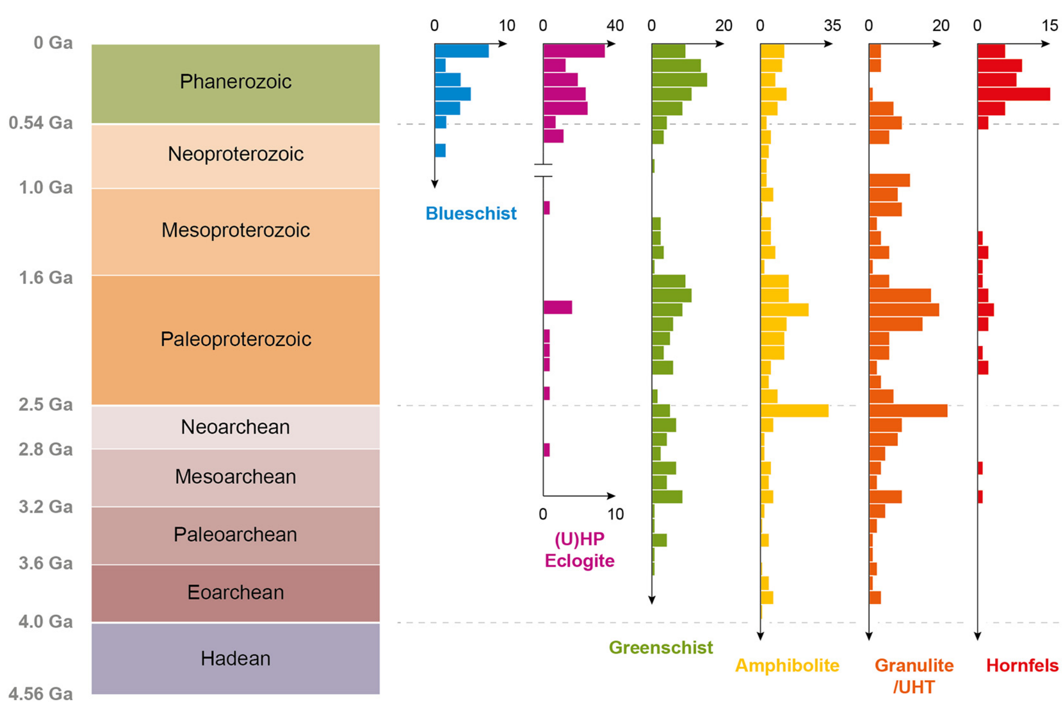

Fig. 4. Secular distribution of metamorphic lithologies of different facies in the rock record (data modified from Gard et al., 2019). Note that this dataset is not comprehensive and is intended to show general trends rather than precise proportions of metamorphic rock types of any age.

same ~300 Myr period is notably more likely to support a global network of plate boundaries having been established at that time (cf. Wan et al., 2020). The curious paucity of HP eclogite in the rock record between 1.8 and 0.8 Ga (Fig. 4) is yet to be satisfactorily explained, although coincides with a global period of tectonic quiescence – the Boring Billion (Roberts, 2013). Finally, brief discussion must be made on mafic eclogite from the Kola Peninsula, Russia (Mints et al., 2010), and Fennoscandian Shield (Dokukina et al., 2014), which are reported to have equilibrated at *P*–*T* conditions of ~16 kbar and ~750 °C at c. 2.87 Ga, and ~24 kbar and ~700 °C at c. 2.82–2.72 Ga, respectively. While these examples may be considered by some to be the oldest known HP examples, there is much debate about the age of metamorphism for these localities (Mints and Dokukina, 2020), with other researchers arguing that they formed during later regional tectonic overprinting during the Svecofennian (1.9–1.8 Ga) orogeny (e.g. Yu et al., 2017). Further research on these rocks using new and high-resolution techniques in petrochronology may help to resolve this issue.

Ophiolite complexes represent alternative petrological evidence for subduction, as they represent fragments of oceanic lithosphere that have been tectonically emplaced (obducted) onto continental crust during plate convergence (Miyashiro, 1975; Dewey, 1976). Until recently, the oldest certified ophiolite on Earth was the Purtuniq ophiolite, Cape Smith belt (c. 2.0 Ga; Scott et al., 1991, 1992), which lies within the Trans-Hudson orogen, Canada; a Proterozoic collisional orogen with many temporal and spatial similarities to the Cenozoic Himalayan orogen (St-Onge et al., 2006). However, Santosh et al. (2016) documented a Late Neoarchean ophiolite from the c. 2.5 Ga Yishui complex, North China Craton, with its lithology, petrology, and geochemistry confirming a suprasubduction zone genesis. Although Kusky et al. (2001) proposed that the Dongwanzi greenstone belt (c. 2.51 Ga), North China Craton, contains dismembered fragments of an Archean ophiolite sequence, this interpretation has been disputed by many other research groups (Zhai et al., 2002), as have sheeted dikes and associated pillow basalts in the Isua supracrustal sequence (c. 3.8 Ga), Greenland, reported by Furnes et al. (2007) and Jenner et al. (2009). Recent studies have also reported well-preserved ophiolite-like successions of Neoarchean age such as those from the Miyun Complex in the North China Craton (e.g., Santosh et al., 2020). These ancient examples are less readily accepted as obducted Archean oceanic lithosphere by the broader geoscience community due to their incompleteness, as individual components of ophiolites – such as sheeted dikes and pillow basalts – may form individually in non-subduction zone tectonic settings (Moore et al., 1982; Vanko and Laverne, 1998). However, additional petrological evidence in some localities supports the interpretation that these greenstone belts do represent metamorphosed oceanic crust (cf. Tang and Santosh, 2018). Arc-type andesite-bearing greenstone belt volcano-sedimentary successions occur in the Superior, Slave, and Yilgarn Archean cratons, among others (cf. Boily and Dion, 2002). These andesitic members are intercalated with graywacke and other volcaniclastic strata that commonly occur along modern-day continental and island arcs, such as boninite, shoshonite, and high-Mg andesite (Condie, 1989; Parman et al., 2001). When considered together as an entire volcano-sedimentary stratigraphic package, subduction-zone processes represent the most likely explanation for their genesis.

6.2. Tectonic evidence

Tectonic evidence for plate tectonics indicates independent plate motion or rotation, or else describes large-scale geological features that were created by dominantly horizontal tectonic forces. Evidence for the former is readily provided by paleomagnetism, although this technique can be challenging to apply to rocks formed on the early Earth (Van der Voo and Channell, 1980). Typically, paleomagnetic studies and the identification of apparent polar wander in Archean terranes is complicated by the lack of suitable stratigraphic sections that are horizontal, have remained undeformed, and have not been remagnetized since acquisition of their primary magnetism. Further issues arise with providing geological constraints for different sedimentary or volcanic formations, due to uncertainties associated with many isotopic dating techniques increasing with absolute age (Schoene et al., 2013). Nonetheless, several studies have managed to circumvent these issues. In particular, a recent study by Brenner et al. (2020) presented new paleomagnetic data from tholeiitic metabasalts in the East Pilbara Craton, western Australia, and demonstrated that the different paleolatitudes of the terrane documented by sequential phases of volcanic activity required plate motion of at least 2.5 cm/yr between c. 3.35 and c. 3.18 Ga. This is comparable to documented rates of continental drift on the Phanerozoic Earth (~2–10 cm/yr) and exceeds those predicted for stagnant- and sluggish-lid models (up to 2 cm/yr; Fuentes et al., 2019), suggesting the operation of Wilson Cycle-like plate motion during the Mesoarchean.

In parallel with paleomagnetism providing evidence of motion of *individual* continental blocks, larger-scale tectonic evidence of drift of *multiple* blocks is readily provided by the supercontinent cycle. A supercontinent is a vast landmass formed by accretion of most (or all) continental fragments that exist on Earth at any point in time (Rogers and Santosh, 2004). Due to the limited degree of horizontal motion associated with various forms of stagnant lid regime, evidence of supercontinent formation provides strong support for mobile lid tectonics. The first undisputed supercontinent that formed on Earth assembled at c. 2.0–1.8 Ga, termed Columbia/Nuna (Rogers and Santosh, 2002; Meert and Santosh, 2017), and was followed by Rodinia (1.2–1.1 Ga), Gondwana (0.54 Ga), and Pangea (0.30–0.25 Ga) (Rogers and Santosh, 2004). Two supercontinents are also suggested by some researchers to have formed on the Archean Earth – Ur (3.0 Ga; Mahapatro et al., 2012) and Kenorland (2.7–2.5 Ga: Aspler and Chiarenzelli, 1998) – and if true would provide strong support for a globally established network of subduction zones and operation of the Wilson Cycle at that point in time. A diverse range of geodynamic models has been proposed to account for this billion-year cyclical pattern of assembly and breakup of supercontinents (cf. Nance et al., 2014). Double-sided subduction (Maruyama et al., 2007) and/or multiple sets of subduction zones within a single oceanic basin (Santosh et al., 2009) have been proposed to promote the rapid assembly of continental fragments into supercontinents. Supercontinent dispersal is thought to be driven by mantle plumes associated with large igneous provinces and giant dike swarms, which have been temporally linked to the demise of Columbia/Nuna and Rodinia (Ernst et al., 2008).

Parallel and pseudo-linear belts that preserve low-temperature/high-pressure (LT/HP) mineral assemblages in one terrane and high-temperature/low-pressure (HT/LP) mineral assemblages in an adjacent terrane are called paired metamorphic belts (Miyashiro, 1961, 1973). These belts record convergent margin activity, where LT/HP metamorphism occurs in the subducted slab, forming blueschist- and eclogite-facies metamorphic rocks, and HT/LP metamorphism occurs in the overlying island or continental arc, forming amphibolite-facies, granulite-facies, or ultrahigh temperature (UHT) rocks (Iwamori, 2000). Such belts record the penecontemporaneous metamorphism along contrasting apparent thermal gradients – one cold and one hot – in discrete terranes that are later tectonically juxtaposed (Oxburgh and Turcotte, 1971). The classical locality for paired metamorphism is the Sanbagawa Belt, Japan (Banno and Nakajima, 1992), although similar belts have been documented in Precambrian terranes (e.g. Katz, 1974). Paired metamorphic belts therefore record many forms of arc-related activity, such as subduction and crustal shortening and thickening.

Whereas paired metamorphic belts record the tectonic and metamorphic processes that characterize ongoing subduction, the closing phase of the Wilson Cycle is documented by terminal destruction of an ocean basin leading to collision and amalgamation of multiple continental blocks and/or intervening arc systems (Wilson et al., 2019). Accretionary orogenesis typically occurs during ongoing subduction and evolves to collisional orogenesis during continental accretion (Cawood et al., 2009; Santosh et al., 2009). Accretionary orogens exhibit accretionary complexes containing MORB and deep-sea sediments belonging to the subducting oceanic plate, and medium- to high-grade metamorphic rocks and calc-alkaline/I-type batholiths belonging to the overlying continental plate (Bahlburg et al., 2009; Sepidbar et al., 2019), which are separated by a forearc basin (Hall, 2009). By contrast, collisional orogens are characterized by passive continental margin sequences (Gaetani and Garzanti, 1991), with an orogenic core of medium- to high-grade regional metamorphic rocks (Etheridge et al., 1983; Weller et al., 2013; Palin et al., 2018; Treloar et al., 2019; Kang et al., 2020). A collisional suture with remnants of oceanic components marks the zone of ocean closure (Thakur and Misra, 1984; Robertson, 2000; Palin et al., 2015; Parsons et al., 2020). Accretionary orogens often contain accretionary prisms – accumulations of material scraped off subducting oceanic lithosphere that show a downward younging of successive strata (Huang et al., 1997). The relative volume of individual units in accretionary prisms also typically decreases with age, that some portion of the accreted material is tectonically eroded and carried into the mantle via subduction erosion (Isozaki et al., 2010). This process can lead to underplating of overlying arc crust with felsic sedimentary material and allows hydration of the mantle wedge (Platt et al., 1985; Hacker et al., 2011). Accretionary prisms are sparse the geological record before c. 0.9 Ga (Hamilton, 1998), although several mélanges in Archean terranes have been interpreted as accretionary prisms, including the Schreiber-Hemlo greenstone belt (c. 2.75–2.70 Ga), Superior Province, Canada (Polat and Kerrich, 1999) and the Abitibi greenstone belt (c. 2.70 Ga), Quebec (Mueller et al., 1996).

6.3. Geochemical and isotopic evidence

Many forms of geochemical and isotopic data can be used to infer the operation of plate tectonics through geological time. The tectonic environments in which igneous rocks formed are commonly constrained using trace element ratios, particularly for mafic rocks, which can be used to identify depleted mantle (DM), enriched mantle (EM), and hydrated mantle (HM) source regions (e.g. Workman and Hart, 2005; Pearce and Stern, 2006). Basalts with these geochemical signatures are often interpreted to have formed at mid-ocean ridges, on oceanic plateaux above mantle plumes, and in arc/back-arc settings where partial melting takes place within a hydrated mantle wedge, respectively (Condie, 1985). Many researchers have applied these trace element discrimination diagrams to Archean and Proterozoic basalts in greenstone belts to determine whether these mantle reservoirs existed at various points in time (e.g. Furnes et al., 2014); however, caution is advised, as some studies have shown that these techniques are not always reliable at identifying tectonic settings on the young Earth where the mode of basalt formation can be independently verified by other geological criteria (Snow, 2006; Vermeesch, 2006; Li et al., 2015).

Greenstone belts from Archean terranes worldwide are argued by some researchers to represent obducted and metamorphosed components of Precambrian oceanic crust; as such, application of trace element discrimination techniques to their mafic components should be able to verify the geodynamic setting of protolith (basalt) formation, assuming that no significant modification of the elemental ratios involved has taken place during subsequent heating and burial. Furnes et al. (2014) used multiple incompatible element ratios (Th/Yb, Nb/Yb, V/Ti) to interpret that most basalts from Archean greenstone belts formed in convergent margin tectonic settings due to the preservation of geochemical signatures similar to Phanerozoic MORB, boninite, and island arc tholeiite. By contrast, Condie et al. (2016) suggested that modern-day tectonic settings cannot be confidently identified in rocks older than c. 2.5 Ga, as distinct EM and DM signatures only become resolvable after that time. A critical limitation to applying these geochemical techniques to define when plate tectonics began on Earth is the issue of defining what major-, minor-, and trace-element signatures basalts generated in various types of stagnant lid regime should exhibit (e.g. Hernández-Montenegro et al., 2019). Although mantle plume-related magmatism occurs on Earth today, complicating factors such as progressive depletion of the upper mantle through time and the uncertainty concerning the rate and degree of secular cooling (e.g. Ganne and Feng, 2017) introduce uncertainty into forward models of partial melt composition that depend on factors such as protolith composition, pressure and temperature conditions of melting (Weller et al., 2019; Hernández-Uribe et al., 2020a), and degree of fractional crystallization, mixing, and assimilation during magma ascent through the crust (Hastie et al., 2015). In particular, HM trace element ratios in basalt are viewed as diagnostic signatures of magma genesis at oceanic or continental arcs during the Phanerozoic, but have recently been suggested to alternatively represent intraplate mantle that has been metasomatized by assimilation of dripped or delaminated hydrous lower crust (e.g. Bédard, 2006; Fischer and Gerya, 2016; Piccolo et al., 2019).

Alongside bulk-chemical compositions, important information concerning changing geodynamics through time can be obtained from the geochemistry and isotopic signatures of individual crystals within metamorphic and igneous rocks. Diamond is of critical use for studying secular change in global geodynamics, as it is physically and chemically resistant to tectonothermal overprinting and stabilizes at pressures equivalent to ~150–180 km depth below the Earth's surface (Sung, 2000). Natural diamonds crystallize from carbon-rich solutions in the mantle and can trap minerals, fluids, or melts that occur at equivalent depths within the Earth's interior (Harte, 2010). As such, they have been used in many studies to examine how mantle 'contaminants' have evolved through time through studies of their inclusion suites and their isotopic compositions. For example, mineralogical evidence of the transition between stagnant lid and mobile lid geodynamic regimes during the Mesoarchean was provided by Shirey and Richardson (2011), who studied silicate and sulfide inclusions in diamonds from five major Archean terranes. While peridotite-like inclusion suites (harzburgite and lherzolite) occur in diamonds of all ages, eclogite-like inclusion suites (garnet plus omphacitic clinopyroxene) became dominant after c. 3 Ga. These data were interpreted to record the onset of global subduction on Earth that allowed eclogite – metamorphosed oceanic crust – and carbon-bearing fluids to be transported to subcontinental mantle depths, with the diamonds subsequently exhumed via volcanism. In an analogous fashion, stable isotope ratios in cratonic diamonds may constrain the onset and degree of crust–mantle interaction though time. For example, carbon isotopes in diamond from the Jagersfontein kimberlite, South Africa (Tappert et al., 2005), and carbon and nitrogen isotopes in diamond from the c. 3.5–3.1 Ga Kaapvaal craton, South Africa, (Smart et al., 2016) were reported to record evidence for the transport of oceanic crust and oxidized carbon-rich sediments into the mantle, presumably facilitated by subduction of oceanic lithosphere at a convergent plate margin. Oxygen and strontium isotope signatures in Archean diamonds from eclogite xenoliths exhumed from cratonic mantle in South Africa (MacGregor and Manton, 1986) are similarly thought to document subduction of hydrothermally altered oceanic crust into the mantle at c. 2.5 Ga.

6.4. Modeling

Both thermo-mechanical (numerical) and petrological (thermodynamic) modeling can be used independently or in combination to infer the likelihood of subduction at different times through Earth history (e.g. Palin et al., 2016b; Ge et al., 2018; Wiemer et al., 2018). This is best achieved by correlating the surficial imprints of key tectonic processes and/or the geochemistry and metamorphic/magmatic P–T evolution of rock types that are predicted to form in both stagnant lid and mobile lid environments with those documented in the geological record.

Thermo-mechanical modeling can be used to test how different physical variables affect evolution of the crust and mantle, and remains a highly effective method to examine the geodynamic effects of secular cooling of the Earth's mantle through time (cf. Gerya, 2014). Examination of the thermal stability of thick, mafic Archean crust has shown that eclogitization and melt-loss from its roots would have caused dripping and/or delamination of this high-density material into the underlying mantle (Fischer and Gerya, 2016; Piccolo et al., 2020; Nebel et al., 2018). It is primarily the results of such simulations that have allowed definition of the secular evolution of different forms of stagnant lid regime, as shown in Fig. 1. Such geodynamic simulations also demonstrate which petrophysical factors are necessary to initiate and sustain plate tectonics on Earth. Parameterizations have been employed that consider variations in oceanic lithospheric thickness, composition, and hydration state, and mantle T_P values for Archean, Proterozoic, and modern-day convergent margins (e.g. Gerya et al., 2008). In general, hot, thick, and highly mafic Archean oceanic slabs are not strong or dense enough to undergo steep subduction (van Hunen and Moyen, 2012), and commonly break apart when the leading-edge transforms to high-density eclogite. The well-documented importance of slab-pull forces for driving plate motion at the surface of the Earth (Conrad and Lithgow-Bertelloni, 2002) indicates that subduction likely only became self-sustainable at a point in geological time when descending slabs were strong (and cold) enough to maintain down-dip coherency, arguing against modern-day like subduction having operated on the much hotter early Earth (Foley et al., 2003; Palin et al., 2020).

By contrast with thermo-mechanical modeling, petrological modeling may be used to independently assess whether rocks exposed at the Earth's surface in Archean, Proterozoic, and modern-day terranes formed via subduction (e.g. Ge et al., 2018). This form of modeling uses equilibrium thermodynamics to predict which minerals, melts, and aqueous fluids would stabilize at various depths and temperatures within the Earth (Powell et al., 1998; White et al., 2000, 2007; Green et al., 2016; Holland et al., 2018). If geochemical mass-balance constraints can be applied, in-depth analysis of the major, minor, and trace element contents of metamorphic rocks and anatectic melts can be obtained (Spear, 1988), which can then be compared to natural lithologies preserved in different terranes worldwide (Palin et al., 2016c). Focused petrological study of Archean geodynamics has recently been facilitated by new thermodynamic descriptions of minerals and melts that may form in metabasalts (e.g. MORB, calc-alkaline basalt, ocean-island basalt), which are thought to have been precursor lithologies for generation of TTG magmas. Natural Archean TTGs and equivalent gray gneisses have historically been divided into low-pressure, medium-pressure, and high-pressure variants based on geochemical signatures that imply magma genesis in the presence or absence of plagioclase, amphibole, garnet, and/or rutile (see Moyen and Martin, 2012 for a comprehensive review). Low-pressure TTGs are thus expected to have formed from partial melting of amphibolite, medium-pressure TTGs from garnet granulite, and high-pressure TTGs from eclogite (e.g. Foley et al., 2003). Such application of petrological modeling has almost universally demonstrated that Earth's first continents did not form via subduction, as all forms of TTG melts matching natural examples may be generated in normal crustal environments (Nagel et al., 2012; Palin et al., 2016b; White et al., 2017; Ge et al., 2018; Wiemer et al., 2018; Kendrick and Yakymchuk, 2020; Laurent et al., 2020; Liu and Wei, 2020; Yakymchuk et al., 2020 and others), and calculations performed at mantle P–T conditions representative of subduction zone metamorphism shows that eclogite is highly infertile (Hernández-Uribe et al., 2020b). Whilst TTGs with appropriate major-element compositions and trace-element signatures may be generated at these high-pressure conditions (e.g. Rapp et al., 1991), the volumes produced cannot account for the proportions observed in Archean cratons. For more information about application of this petrological modeling to Archean metamorphism and TTG genesis, the reader is referred to Palin et al. (2016b) and Kendrick and Yakymchuk (2020).

7. Earth's oldest crystals and Earth's oldest crust

The geological record becomes increasingly incomplete further back in time, which is expected due to older terranes having had more opportunity to be reworked via later episodes of tectonic deformation, overprinted by thermal or regional metamorphism, or eroded away into their constituent grains. This presents many problems for geologists using the distribution of rock types through time as a tool to interpret secular changes in geodynamic processes; for example, arguments can be made about the absence of key petrological indicators of subduction, such as blueschist, prior to c. 0.8 Ga being due to lack of preservation rather than lack of formation, as they are easily retrogressed. Analogous problems arise when making interpretations from data obtained from individual outcrops of Archean age, which may not be representative of global conditions at the time of their formation. Unfortunately, this is a limitation that will likely never be circumvented unless there is an important discovery of new Archean crust in regions of the world that are currently not fully explored.

The search for Earth's oldest rocks is of critical importance for answering a wide range of questions related to our planet's evolution, including the nature and style of its initial tectonic regime. However, as with many aspects of such studies, fervent dispute exists with respect interpretation of data reported from different localities. Today, the oldest-known rocks on Earth are commonly accepted to occur within the Acasta Gneiss Complex, the westernmost exposure of the basement of the Slave Craton, northwest Canada (Bowring et al., 1989). This Complex contains a petrologically diverse suite of rocks ranging from metagabbro to granitic orthogneiss that mostly have metamorphic ages of c. 4.02–3.6 Ga (e.g. Stern and Bleeker, 1998; Bowring and Williams, 1999). The timing of metamorphism and melting in these rocks has been tightly constrained by U–Pb zircon geochronology (e.g. Reimink et al., 2014, 2016), which is generally considered a reliable petrochronological technique for producing high-precision ages in ancient rocks (e.g. Montgomery, 1979; Kohn et al., 2015). Further, the changing chemical systematics of zircon during metamorphism and partial melting are well studied and well understood (Lee et al., 1997; Rubatto and Hermann, 2007), such that it is straightforward to discriminate the timing of different thermal or tectonic events based on ratios of trace elements and/or REEs in different microstructural domains, such as extraction of a primitive melt from the mantle and subsequent metamorphism in an orogenic environment. However, zircon is rare in mafic and ultramafic rocks, meaning that other minerals and isotope systems are often required to date them. In a landmark study, O'Neil et al. (2008) reported a ^{146}Sm–^{142}Nd isochron age of c. 4.28 Ga from amphibolite-like mafic schist from the Nuvvuagittuq greenstone belt, Québec, Canada, making them contenders for being named as Earth's oldest crust; however, these data have proven contentious (Andreasen and Sharma, 2009), given assumptions of the initial concentration of the ^{146}Sm parent isotope on the early Earth, which is now extinct. Key aspects of the geology and tectonic interpretations for both localities are briefly summarized below, although for a more comprehensive review of the nature of Earth's first crust that encompasses many recent discoveries and tectonic models, the reader is referred to Carlson et al. (2019).

The Acasta Gneiss Complex contains a wide spectrum of lithologies, which Reimink et al. (2016) divided into four main types based on age, structure, and petrology: a layered gneiss unit composed of meter-scale tonalitic and granodioritic members; foliated, garnet-bearing amphibolite; weakly deformed metagabbro that preserves some relic igneous textures; and a dominant, massive orthogneiss with granodioritic to granitic compositions. The oldest lithological components of this suite are low-strain, mafic tonalitic gneisses of the Idiwhaa unit, which contain igneous zircons with U–Pb crystallization ages of c. 4.02 Ga (Reimink et al., 2014). These felsic gneisses were reported by Reimink et al. (2014) to be unusually Fe-rich (~9–15 wt. % FeO) and so have lower Mg-numbers (~13–18) than typical Archean gray gneisses

(~25–60, with a median of ~43; Moyen, 2011), which in turn was interpreted to record shallow-level fractional crystallization of a low-H_2O basaltic parent magma. Oxygen isotope analyses indicated the assimilation of rocks previously altered by surface water. The most likely tectonic scenario for generating such melts is an Iceland-like environment where mantle upwelling generated a thick oceanic plateau that experienced intracrustal melting, differentiation, and magma hybridization (Kröner, 1985; Reimink et al., 2014), which aligns with suggestions that the very early Earth experienced more intense plume-related magmatism than during the Proterozoic and Phanerozoic (Bédard, 2018) and that a wide variety of magmas may be produced in such intraplate environments (e.g. Hastie et al., 2010, 2016).

The Nuvvuagittuq greenstone belt, Superior Craton, is a relatively small (6 km^2) terrane, but exposes a wide variety of rock types, including felsic to intermediate orthogneiss, ultramafic and mafic sills, and metasediments (e.g. O'Neil et al., 2007). Much focus has been given in recent years to the petrology of mafic supracrustals – amphibolite-like rocks, termed the Ujaraaluk unit – that were reported by O'Neil et al. (2008) to have a ^{146}Sm–142^{Nd} whole-rock isochron Hadean age of c. 4.28 Ga (Fig. 3). Many of these mafic units show major minor, and trace element compositional similarity to metabasalt from several Early Archean terranes worldwide (Carlson et al., 2019), although those in the Ujaraaluk unit typically contain cummingtonite instead of hornblende (O'Neil et al., 2008). Based on compatible and incompatible trace element ratios, these faux-amphibolites have been interpreted to have formed from basalt derived directly from a peridotite mantle (O'Neil and Carlson, 2017) and thus may represent subsequently deformed and metamorphosed relics of Earth's oldest secondary crust.

Despite the ancient heritage of these rocks from the Nuvvuagittuq greenstone belt and Acasta Gneiss Complex, even older terrestrial materials occur in the form of detrital zircon grains within clastic metasediments in the Jack Hills area of the Archean Narryer Terrane, Western Australia (e.g. Hoskin, 2005). Early investigation of a greenschist-facies meta-conglomerate from this region by Compston and Pidgeon (1986) revealed two zircon grains with ages of 4276 ± 12 Ma, and further investigation of grains from the same locality by Wilde et al. (2001) revealed a single grain with a $^{207}Pb/^{206}Pb$ age of 4404 ± 8 Ma, the oldest ever obtained from a mineral formed on Earth. Subsequent analyses have shown that Jack Hills detrital zircons show a characteristic bimodal distribution with peaks at c. 3.3 and c. 4.1 Ga (cf. Harrison, 2009).Textural characteristics of the Jack Hills zircons, such as morphology and internal growth zoning, indicate that virtually all are derived from igneous sources (e.g. Cavosie et al., 2004); thus, the older (Hadean) age peak may be considered as a magmatic crystallization age, and the younger (Archean) age peak likely records metamorphic recrystallization and/or isotopic resetting. Although other suites of Hadean zircons exist elsewhere on Earth, including Greenland (Mojzsis and Harrison, 2002), China (Cui et al., 2013), and South America (Paquette et al., 2015), most study has concentrated on the Jack Hills materials and discussion of the information gleaned about the Hadean Earth, below, focuses on studies of this set.

Hadean zircons provide critical information about the geochemical conditions and petrology of the rocks in which they formed, and so the geodynamics of the Earth at that time. This information primarily comes in two forms: chemical and isotopic properties of the zircons themselves, and the mineralogy of inclusions within them. Several independent studies of Jack Hills zircon have reported a heavy oxygen isotope signature (Mojzsis et al., 2001; Wilde et al., 2001) that may be explained by the melt from which these grains crystallized having formed from ^{18}O-rich clay minerals. In turn, this implies that liquid water was present at the Earth's surface at c. 4.4–4.3 Ga. This conclusion is further supported by highly negative values of δ^7Li from Jack Hills zircons that reflect crystallization from a source rock that was strongly weathered (Ushikubo et al., 2008; Tang et al., 2017). Both sets of isotope data may be satisfactorily explained by weathering of pre-existing crust and formation of a clay-rich sediment at the Earth's surface, which was buried – perhaps via subduction – heated, melted, and crystallized zircon.

The existence of a primary Hadean crust that may be subject to weathering and erosion has been inferred from initial $^{176}Hf/^{177}Hf$ ratios in Jack Hills zircons, which exhibit large deviations in εHf(T) from the bulk silicate Earth (Kinny et al., 1991; Harrison et al., 2005) ranging mostly between −10 and +4 (Harrison, 2009). This has been interpreted to reflect early major differentiation of the silicate Earth (Blichert-Toft and Albarède, 2008) and primary crust formation since c. 4.5 Ga. Critically, many of these analyzed ratios cluster along a trend corresponding to a Lu–Hf value of ~0.1, which is characteristic of continental crust, and has led to many models of continental crust formation through time incorporating substantial growth immediately after the Earth's formation. The absence of this crust from the geological record is commonly attributed to destruction and re-working due to a purported intense bolide impact flux at c. 3.9 Ga, often termed the late heavy bombardment (c. 3.9 Ga; Wetherill, 1975; Gomes et al., 2005; Chapman et al., 2007). The effects of high-velocity meteorite impacts on the Earth's primitive crust have been speculated upon by many researchers, with some suggesting that impacts may have triggered subduction initiation (O'Neill et al., 2017, 2020) and induced significant fracturing, weakening, and high-temperature/short-duration metamorphism at impact sites (Byerly and Lowe, 1994; Gibson, 2002; French, 2004; Sleep and Lowe, 2014). As with many other aspects of Earth's early history, study of similar processes on our neighboring rocky planets Mars and Venus – and the Moon – may shed new light on the evolution of the Hadean Earth.

Many workers have reported the mineralogy and crystal-chemistry of inclusion suites within Jack Hills zircons (e.g., Bell et al., 2015a, 2015b, Bell et al., 2017; Cavosie et al., 2004; Caro et al., 2008; Menneken et al., 2007; Nemchin et al., 2008; Rasmussen et al., 2011), which have provided profound insight into the tectonic processes that operated at this point in Earth history. These observations have been used to support and complement suppositions made by some workers from isotope analysis that surficial crustal material was transported to pressure and temperature conditions within the Earth that allowed anatexis to take place, specifically in a subduction zone environment due to the cold *P*/*T* gradients involved (see below). Primary inclusions documented within Hadean zircons worldwide include (but are not limited to) quartz, muscovite and biotite mica, chlorite, K-feldspar and albitic plagioclase, rutile, monazite, xenotime, and even diamond (Maas et al., 1992; Trail et al., 2007; Menneken et al., 2007). A campaign-style analysis of zircons from Jack Hills by Hopkins et al. (2008) showed that quartz and muscovite comprise nearly two-thirds of all inclusions, and that these minerals often occur in close spatial association with mutual grain boundaries, implying chemical and textural equilibrium.

The common occurrence of hydrated mineral inclusions (muscovite, biotite, chlorite, and apatite) that are characteristic of peraluminous igneous rocks can be attributed on the modern-day Earth to either melting of clay-rich metasediments during regional metamorphism – as shown by syn- and post-orogenic Himalayan-type leucogranites that form in the cores of collisional mountain belts – or by production of andesite-like magmas in an island or continental arc setting (Chappell, 1999; Collins and Richards, 2008). Felsic melts may alternatively be produced as highly fractionated differentiates of originally mafic magmas, such as occur in trondhjemitic dikes that formed by hydrous partial melting of gabbro in the roof-zone of an axial magma chamber in the Semail ophiolite, Oman (e.g. Rollinson, 2008), although these occurrences are much less common than orogenic S-type magmas. Nonetheless, all these scenarios are characteristic of mobile lid geodynamic regimes with magma genesis at sites of plate convergence.

Support for the Jack Hills zircons and inclusion suites forming in a subduction zone environment has been put forward due to the results of thermobarometry performed on each. The Ti-in-zircon thermometer (Watson and Harrison, 2005) applied to zircon grains with ages c.

4.4–3.9 Ga produced a mean crystallization temperature of ~680–690 °C, which lies close to the wet melting curve for pelite and granite at *P* > 4 kbar. In addition, the Si contents of muscovite inclusions were used in combination with phase diagram-based modeling of granitic magma genesis by Hopkins et al. (2008) to constrain a mean crystallization pressure of ~6.9 kbar (at ~680–690 °C), and thus a static and linearized geothermal gradient of ~980 °C/GPa. Interpretation of what these *P*–*T* conditions mean for the geodynamics of the Hadean Earth is complex. As noted in Section 6.1, most rocks diagnostic of subduction on the Phanerozoic Earth, such as blueschist and lawsonite-bearing eclogite, form at *P*/*T* gradients <440 °C/GPa (Penniston-Dorland et al., 2015; Palin et al., 2020). Thus, while Hopkins et al. (2008) and others have interpreted that these zircons formed from melts generated in an underthrust environment – perhaps similar to a modern-day subduction zone – the *P*/*T* gradient calculated from these inclusion suites is over twice as hot as this 'upper' limit for warm subduction, at least on the Phanerozoic Earth. These data therefore may provide primary constraints on the thermal structure of Archean subduction, which is otherwise difficult to constrain in the absence of (U)HP rocks older than c. 2.8 Ga (Fig. 4).

A final note to be made when discussing zircons from the Jack Hills region is the occurrence and carbon isotopic signature of graphite and diamond inclusions. Ion microprobe analyses of individual and composite inclusions in zircon grains as old as c. 4.2 Ga by Nemchin et al. (2008) revealed strongly negative $\delta^{13}C$ isotope values between −5‰ and −58‰, with a median value of −31‰. These data were supported by similar analyses of graphite flakes included in c. 4.1-Ga zircon from the region by Bell et al. (2015a, 2015b), who produced $\delta^{13}C$ isotope values of −24 ± 5‰. The interpretation of these strongly negative values is contentious, as they are consistent with a biogenic origin, although not diagnostic of it. Abiotic processes that may account for light $\delta^{13}C$ signatures, such as incorporation of meteoritic materials ($\delta^{13}C$ values from +68‰ to −60‰) or carbon isotopic fractionation by diffusion are considered unlikely to produce consistently low $\delta^{13}C$

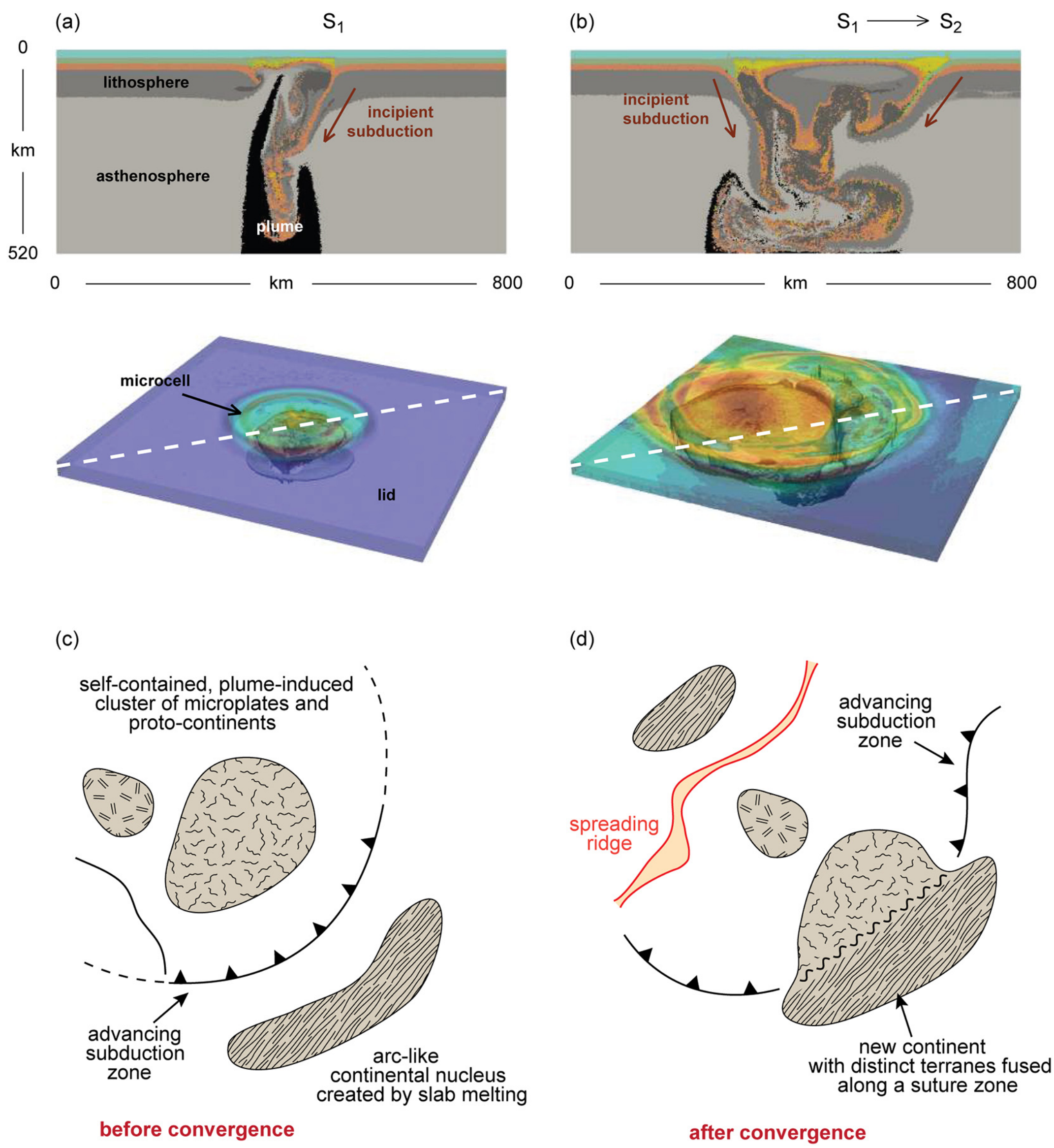

Fig. 5. Conceptual model for localized subduction initiation at the head of a mantle plume in a global stagnant lid environment. (a) Two- and three-dimensional numerical model of plume–lid interaction showing penetration and localized partial melting and destabilization of the primary mafic crust. Incipient subduction initiates at the plume head margin (S_1) and may form a semi-continuous perimeter around the microcell (modified after Piccolo et al., 2020). (b) Over time, lithospheric delamination and rollback allows the proto-subduction zone to expand outwards and the microcell grows (S_1, S_2) (c–d) Schematic plan view of plate tectonic-like features within a microcell that generate convergent plate margin interactions if several microcells interact (modified after Palin et al., 2020).

values over the range of ages of host grains analyzed in these studies (cf. Clayton, 1963; Robert and Epstein, 1982; Engel et al., 1990), giving weight to the hypothesis that primitive biological activity was taking place at the Earth's surface at this time. Nonetheless, an abiotic origin for graphite formation has recently been favored by Menneken et al. (2017), who documented CO_2 inclusions in Jack Hills zircons that also exhibited thin carbon films on the inside of inclusion walls. This close spatial relationship between graphite and CO_2 was suggested by to indicate precipitation of carbon during thermal metamorphism, and not as evidence for a terrestrial biosphere at c. 4 Ga. Regardless of the origin of the isotopically light carbon, undersaturation of carbon in the Earth's mantle requires that surficial sediments or crustal materials were transported into the interior at this time (Dasgupta and Walker, 2008). Unfortunately, these isotope data cannot directly determine whether subduction or other tectonic processes were responsible, such as dripping or delamination (Fig. 1), and thus other lines of evidence are needed to identify whether plate tectonics had begun to operate during the Hadean.

8. Summary remarks and future directions

Study of the Archean Earth has become a major sub-category of geoscience, as demonstrated by the vast number of research articles that are published on the topic each year and the frequency with which review papers are required to keep up with new developments. In many cases, the major aim of these studies is to constrain the timing of onset of plate tectonics on Earth, and historical estimates in the literature span billions of years (Fig. 3). Some workers have used mineral inclusions in Jack Hills zircon and isotopic data for felsic crust present at the Earth's surface soon after planetary formation to propose subduction having begun during the Hadean (c. 4.2–4.0 Ga: Hopkins et al., 2008), whereas most interpretations cluster around the Mesoarchean (c. 3.2–2.8 Ga: Cawood et al., 2006; van Kranendonk et al., 2007; Condie and Kröner, 2008; Tang et al., 2016; Palin et al., 2020) based on the appearance of tectonic features resembling those that characterize Phanerozoic collisional orogens, geochemical evidence for a rapid increase in primary continental crust thickness (Dhuime et al., 2015), paleomagnetic evidence for continental drift (Brenner et al., 2020), and evidence for operation of the Wilson Cycle (Shirey and Richardson, 2011). A striking gap exists between c. 2.9 Ga and c. 1 Ga, overlapping with the Boring Billion (c. 1.8–0.8 Ga), and is terminated by a separate cluster of researchers who argue for a Neoproterozoic onset (c. 1.0–0.8 Ga; Stern, 2005; Hamilton, 2011; Stern et al., 2016).

A holistic model of changing geodynamics through time should take into account many of the strong arguments for subduction having begun to operate in the deep geological past – very likely in isolated microcells that were located at the head of mantle plumes (Fig. 5a–b). The occurrence of multiple mantle plumes on the early Earth infers the existence of multiple microcells, and it is conceivable that interactions between these cells could produce plate margin-like features, such as collisional orogenesis (Fig. 5c–d), although in the absence of a global network of subduction zones. For more detailed discussion of this model, the reader is directed to Palin et al. (2020). This regime of isolated microcells with localized subduction zones eventually transitioned into a global-scale phenomenon as secular cooling of Earth's mantle allowed oceanic lithosphere to become stronger and less buoyant. The character of subduction has undoubtedly changed through time in many other ways and has impacted the diversity of its tectonic, petrological, and geochemical products preserved in the geological record. Several lines of evidence support the hypothesis that cold, deep, and steep slab subduction is a recent (<0.9 Ga) phenomenon, which is exemplified by the abrupt emergence of key rock types, such as blueschist and UHP eclogite, of that age (e.g. Fig. 4). Evidence for plate motion and continental and island arc-related activity before this point in time, such as the supercontinent cycle, can be readily accounted for by shallow subduction/underthrusting that is also documented on Earth today. A reduction in the average temperature of such shallow subduction zones accounts for secular changes in TTG composition (Martin and Moyen, 2002) – of which many require formation at pressures only achievable via subduction – and the increasing number of non-UHP eclogite with MORB-like affinity in the Proterozoic.

As the geological record becomes more fragmented with age, various forms of modeling begin to provide more in-depth insight into the likely geodynamics of the Archean (and Hadean?) Earth. Thermo-mechanical modeling argues strongly for one of many forms of stagnant lid tectonics before c. 3 Ga (Fig. 6) where the Earth was dominated by vertical plate motion and intracrustal differentiation, producing the bimodal TTG and greenstone lithological associations that are typical of Archean cratons (e.g. Piccolo et al., 2020). Poor constraints on key petrophysical properties, such as mantle T_P, obviate definitive statements concerning the style of stagnant lid regime that occurred at any point in time, although observations and interpretations made from extraterrestrial bodies in our solar system can be used to supplement the results of numerical simulations. These suggest that the Archean Earth transitioned from an environment characterized by extensive volcanism (heat-pipe world) to one characterized by intrusive magmatic activity (plutonic squishy lid), thus thickening the crust with time. Little is known about the Hadean environment, as samples are restricted to mafic supracrustal rocks in the

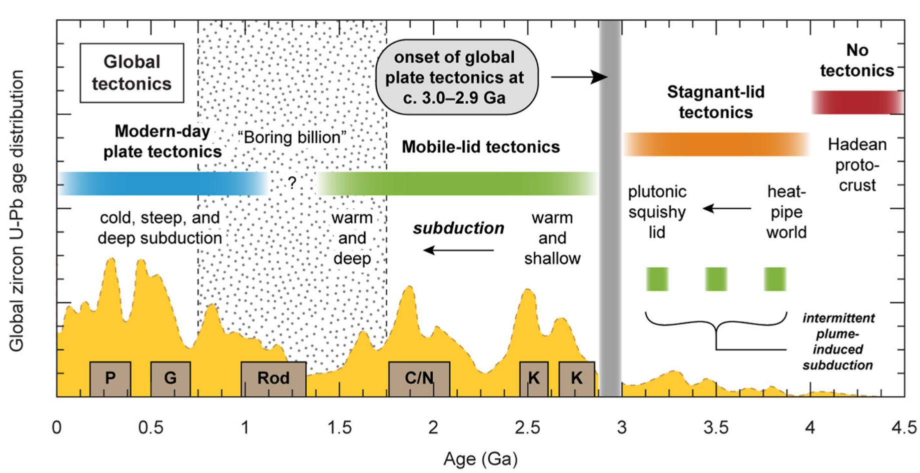

Fig. 6. Summary diagram showing the evolution of global tectonics through time as constrained by the various lines of evidence discussed in this review. Global archive of zircon U–Pb ages is from Vermeesch (2012) and brown boxes denote supercontinent events: P = Pangea; G = Gondwana; Rod = Rodinia; C/N = Columbia/Nuna; K = Kenorland. Modified after Palin et al. (2020).

Nuvvuagittuq greenstone belt (c. 4.28 Ga) and xenocrystic zircons within younger (Archean) metasedimentary rocks, such as quartzite from the Jack Hills region of western Australia (see Section 7). Nonetheless, this small suite of zircon grains has provided a wealth of information about the likely geochemical, mineralogical, and isotopic character of the crust and mantle at this enigmatic time in Earth history, as well as potentially providing evidence for an emerging biosphere less than 300 Myr after planetary accretion.

While most of the geoscience community is arriving at a consensus about these secular changes, there are several key areas of research that may yet make substantive impact on our current understanding. Firstly, continued debate concerning the magnitude of secular cooling and absolute temperatures of the Archean mantle has stymied advances in petrological and geodynamical interpretations of early Earth terranes. Many models are predicated on the supposition of a 'hot' Archean mantle, as indicated by primary magma solutions determined from non-arc basalts by Herzberg et al. (2010). However, subsequent studies have argued for a cooler mantle T_P, albeit still hotter than the present day (Fig. 4). While these differences in magnitude are somewhat small (ΔT ~ 100–300 °C), thermo-mechanical simulations suggest that they are significant enough to have dramatic consequences for the predicted forms of geodynamics and timing of onset of subduction (e.g. Piccolo et al., 2019). Developing new techniques to constrain mantle T_P and/or extracting higher-precision information from current datasets should ideally be an area of targeted research in the future. In addition, given the sensitivity of models of mantle convection to intrinsic and extrinsic parameters, the water content and so rheological properties of its mineralogical constituents must be well constrained if reliable data are to be produced. Continual advances in the ability of analytical techniques to measure smaller and smaller concentrations of trace elements and impurities in so-called nominally anhydrous minerals (NAMs) – such as hydrogen – show that mantle peridotite may in fact be a substantial reservoir of water and other volatiles within the Earth (Bell and Rossman, 1992). The physical properties of olivine, for example, are strongly influenced by its hydrogen content (Karato et al., 1986; Katayama and Karato, 2008) and uncertainty regarding the absolute concentration of H and how this varies with pressure and temperature within the Earth are yet to fully parameterized. Advances in this field have far-reaching implications for assessments of the mechanisms of heat loss within the Earth and feedbacks between mantle dynamics and surface processes.

Finally, it should always be recognized and accepted that many lines of evidence that are *compatible* with subduction do not exclude other geological processes that may transport crustal materials into the mantle, and even if subduction is accepted as the only viable mechanism responsible for forming a given geological feature of a certain age, this does not require that plate tectonics was operating at a global scale at that time. The quest for improved knowledge on these matters will continue for many decades to come; if we have learned so much over the past 50 years since the landmark Penrose conference in 1969 where plate tectonic theory was formalized, what more will we know in 50 years from now?

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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