Archean Subduction: Fact or Fiction?

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Abstract

Subduction drives plate tectonics and builds continental crust, and as such is one of the most important processes for shaping the present-day Earth. Here we review both theory and observations for the viability and style of Archean subduction. High Archean mantle temperature gave low mantle viscosity and affected plate strength and plate buoyancy. This resulted in slower or intermittent subduction, either of which resulted in Earth cooling profiles that fit available data. Some geological observations are interpreted as subduction related, including an "arc" signature in various igneous rocks (suggesting burial of surface material to depths of 50–100 km), structural thrust belts and dipping seismic reflectors, and high-pressure–low-temperature and low-pressure–high-temperature paired metamorphic belts. Combined geodynamical and geochemical evidence suggests that subduction operated in the Archean, although not, as often assumed, as shallow flat subduction. Instead, subduction was more episodic in nature, with more intermittent plate motion than in the Phanerozoic.

1. IMPORTANCE OF UNDERSTANDING EARLY SUBDUCTION PROCESSES

Plate tectonics: planet-wide mantle convection with moving surface plates, which mostly deform at their boundaries by seafloor spreading, subduction, and transform faults

Geodynamical: relating to the thermomechanical physics of Earth's large-scale tectonic processes

Subduction: the process of two converging plates, in which one of the lithospheric plates slides below another into the mantle

Over the past few billion years (Ga), plate tectonics has sculpted Earth, with bi-modal distribution of continental and oceanic lithosphere, mountain ranges, and deep-sea trenches. It also plays a dominant role in the thermal evolution of Earth because mid-ocean ridges release the bulk of Earth's internal heat (Jaupart et al. 2007).

Geodynamically, the subduction process plays a key role in this. It forms the main engine for plate tectonics, with the gravitational pull from dense slabs providing approximately 90% of the driving force (Forsyth & Uyeda 1975). Furthermore, the complex rheological structure of subduction zones allows the decoupling of plates at the surface (Tackley 1998, Gerya et al. 2008) and might have been partly responsible for the longevity of continental lithosphere through the formation of mobile belts (Lenardic et al. 2003).

From a geochemical point of view, subduction arguably has been and still is the most important process in building the compositional structure of Earth. Today, subduction zones are the primary sites of continental crust formation (Davidson & Arculus 2006). Furthermore, subduction provides the main pathway for surface material (sediments, volatiles) to (re)enter the deep Earth (e.g., Rüpke et al. 2004) and seems to be the only mechanism capable of doing so. Subduction zones also form the primary sites for mineral mining (Bierlein et al. 2009).

The debate about Archean tectonics revolves around the existence, or lack thereof, of a form of plate tectonics on the early Earth. Was plate tectonics operating then? If so, how different was it from that of the modern environment? Did subduction always have its prominent role? These questions are addressed in this review. We present geodynamical arguments on Archean plate subduction, followed by observational evidence and a discussion about potential changes in subduction style. We end with a description of future research directions.

2. SUBDUCTION MODELS

The operation of plate tectonics critically depends on Earth's interior temperature. Temperature directly controls the strength, density, and composition of rocks and might have had significant effects on the viability and style of the subduction process. Below, a brief review of modern subduction forces and Earth's thermal evolution provides the background for a discussion about (*a*) the potential secular changes in subduction physics, (*b*) the viability of early subduction, and (*c*) how subduction dynamics links back to Earth's thermal evolution.

2.1. Driving and Resisting Forces for Subduction Today

Any large-scale motion of the solid Earth is ultimately driven by gravity acting on lateral density variations. Subducting plates are denser than ambient mantle material and therefore sink into the mantle. This major driving force for plate tectonics and mantle convection is balanced by resisting forces arising from material deformation (plate bending, mantle drag) and the phase-change buoyancy effect (**Figure 1*a***).

For ongoing subduction with a deep slab, the main driving forces are thermal slab pull, given by $F_{SP} \sim 3.3 \times 10^{13}$ Newtons per meter (N m^{-1}), and the uplifted phase change at ~400 km of olivine to spinel, given by $F_{OS} \sim 1.6 \times 10^{13}$ N m^{-1} (Turcotte & Schubert 2002). Other, smaller driving forces can be significant: Negative buoyancy of eclogite in the upper mantle (with excess density $\Delta\rho \sim 100$ kg m^{-3}) is often ignored but amounts to $\sim 5 \times 10^{12}$ N m^{-1} for an oceanic crust thickness $d_{cr} = 7$ km throughout the upper mantle. Ridge push $F_{RP} \sim 3.9 \times 10^{12}$ N m^{-1} (Turcotte & Schubert 2002) arises from horizontal pressure gradients under aging oceanic lithosphere.

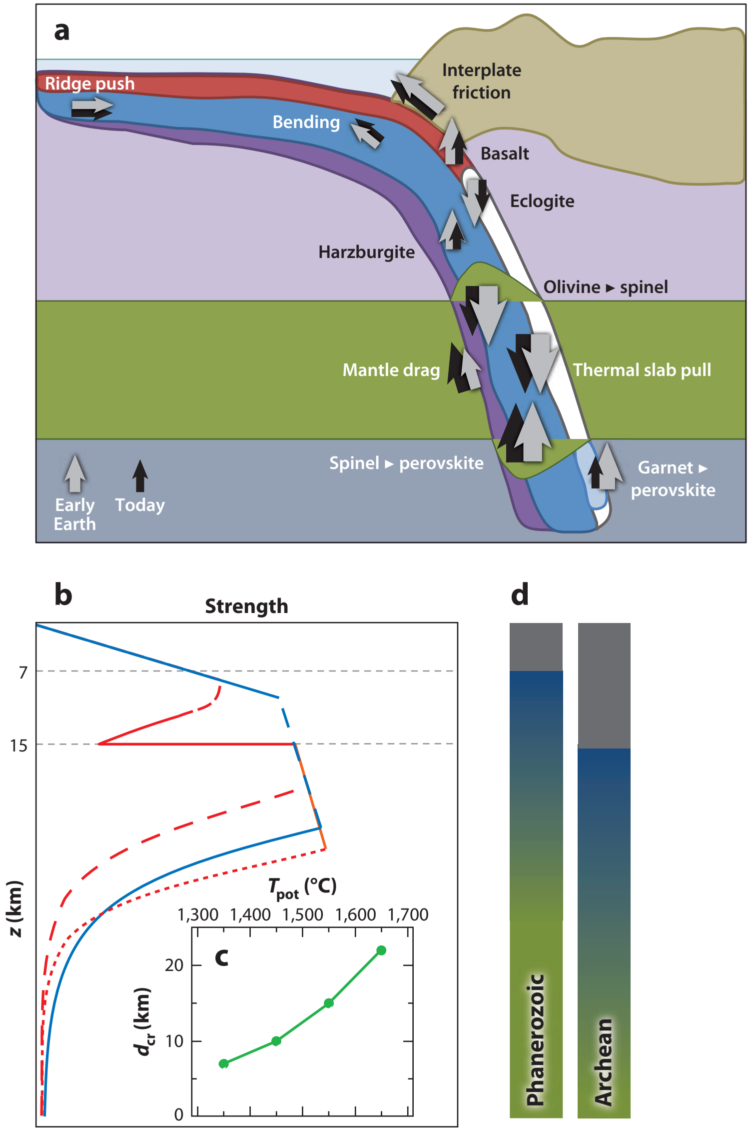

Figure 1

(*a*) Principal driving and resisting forces for subducting lithosphere, schematically indicated by arrow size and direction in black (today) and gray (early Earth). Red/white, blue, and purple lithosphere layers indicate oceanic crust, depleted mantle lithosphere, and undepleted mantle lithosphere, respectively. (*b*) Schematic strength profile of modern mantle (*blue*, after Kohlstedt et al. 1995) and hotter mantle (*red*) oceanic lithosphere with (*dotted*) or without (*dashed*) a significant effect from dehydration strengthening. (*c*) Oceanic crustal thickness d_{cr} plotted against mantle potential temperature T_{pot} (after van Thienen et al. 2004). (*d*) Schematic compositional buildup of oceanic lithosphere: mafic crust (*gray*) overlies depleted, harzburgitic (*blue*) mantle lithosphere, which overlies fertile, lherzolitic (*green*) mantle lithosphere.

Mid-ocean ridge basalt (MORB): the most common of the mafic lavas, formed by melting of the water-poor upper mantle

Mg#: Mg/(Mg+Fe); iron (Fe) is more incompatible than magnesium (Mg), so that the Mg# reflects the degree of melting

Resisting forces arise from friction and buoyancy (Billen 2008). The main resistance comes from mantle drag surrounding the plate and slab (Davies 1999, Billen 2008). Plate-bending force estimates vary considerably (e.g., Conrad & Hager 1999, Capitanio et al. 2007), from $\sim 1 \times 10^{12}$ N m^{-1} (as calculated from Davies 2009) up to ~25% of the total subduction energy dissipation (Di Giuseppe et al. 2008). Buoyant basaltic oceanic crust is stable until approximately 40 km depth, where it transforms into denser eclogite under equilibrium conditions (Hacker et al. 2003), resulting in a modest resisting force of $<2 \times 10^{12}$ N m^{-1}. However, meta-stability of coarser-grained gabbro or lawsonite-amphibolite, on the basis of laboratory data complications and seismic observations, is likely to occur up to 60–150 km depth (see Hacker et al. 2003 for a discussion), which would significantly amplify this buoyancy force. The density of the depleted mantle lithosphere is slightly lower than fertile mantle under the same pressure-temperature conditions, and this situation results in a buoyancy force of a few tera-Newtons per meter (1 tera-Newton $= 10^{12}$ N). Phase change of spinel+garnet to perovskite+magnesiowüstite at ~660–750 km leads to slab stagnation and layered mantle convection and causes a significant resistance to subduction. Although comparable with (but opposite to) the olivine-to-spinel transition force, this phase change has a complex net effect on subduction near the surface because slabs bend and deform in the transition zone; therefore, it does not simply push slabs upward.

2.2. Consequences of a Hotter Early Earth

A hotter mantle would have had a profound influence of the dynamics of the subduction process. Here, the evidence for increased mantle temperature in the past is reviewed, and the consequences for the buoyancy and strength of the subducting plates are illustrated.

2.2.1. Evidence for a hotter mantle. The earliest Earth formed by accretion of solar system material; collision of planetesimals, including a giant Moon-forming impact; and differentiation into core and mantle. All these processes released enough energy to melt the entire Earth. Relevant here is that Earth cooled down from very hot conditions to a threshold temperature (~200 K hotter than today, according to Labrosse & Jaupart 2007) at which rigid plates became first viable. From the paleo-Archean [i.e., less than 4 billion years ago (<4 Gya)], the geological record constrains mantle temperatures. Because mantle temperature varies vertically by ~ 0.3 K km^{-1} due to adiabatic (de)compression, it is convenient to use the mantle potential temperature T_{pot}, defined as the mantle temperature extrapolated adiabatically to the surface. We have several indirect techniques to measure Archean T_{pot}. Pressure-temperature data from Archean high-grade terranes suggest a continental geotherm that is quite similar to today's (England & Bickle 1984). A more commonly used indicator for past mantle temperatures is the liquidus temperature of primitive, basaltic melts, such as mid-ocean ridge basalts (MORBs) from ophiolites and greenstone belts, which formed by decompression melting of rising mantle material and percolated to the surface to form the oceanic crust. A hotter mantle starts melting deeper and to a larger degree (McKenzie & Bickle 1988, van Thienen et al. 2004), and the (finite amount of) iron in the melt will be diluted with more magnesium. Therefore, MgO content and Mg# of primitive melts are indicative of mantle temperature. However, volatiles (such as water) significantly affect melting temperatures, so knowledge or assumptions about the volatile content of the Archean mantle are important. On the one hand, a dry Archean mantle may have been 200–300 K hotter (Nisbet et al. 1993, Abbott et al. 1994a), with even hotter mantle plumes forming the very-high-magnesium komatiites; on the other hand, melting under wetter (subduction?) conditions required an only 100-K warmer mantle (Grove & Parman 2004). Data from Herzberg et al. (2010) suggest mid-to-late Archean mantle temperatures in excess of today at $\Delta T_{pot} = 150$–250 K. Therefore, field evidence suggests an average mantle cooling rate since the mid-Archean of approximately 100–300 K over 3 Ga or ~30–100 K Ga^{-1}, probably somewhat lower than today's cooling rate of ~50–120 K Ga^{-1} (Jaupart et al. 2007). Below, we discuss how surface heat flow and cooling rates are directly linked to plate-tectonic vigor, thereby providing independent, additional constraints.

2.2.2. Plate buoyancy. Mantle temperature and crustal thickness are empirically related (van Thienen et al. 2004) (**Figure 1**). Melting and mantle depletion start at ~45 km under modern spreading ridges but start ~35 km deeper for every 100°C in mantle temperature increase (Vlaar et al. 1994). Crustal density ρ_c (kg m^{-3}) increases with the extrusion temperature T_{ex} (°C) (which relates linearly to MgO content) as given by van Thienen et al. (2004):

$$\rho_c = 1500 + 1.925 T_{ex} - 5.153 \times 10^{-4} T_{ex}^2 \tag{1}$$

The density of the depleted mantle residual after the melt extraction depends approximately linearly on the depletion F, with a density decrease of 0.726 kg m^{-3} per depletion percent (Schutt & Lesher 2006), although estimates range from 3 to 0 kg m^{-3} per depletion percent. These layer thicknesses and densities can be combined to approximate the neutral-buoyancy thermal age of oceanic lithosphere, for which thermal and compositional buoyancy cancel out. At present, this is approximately 25 million years (25 Ma) and increases to ~75 Ma for $\Delta T_{pot} = 100$ K (Davies 1992, van Thienen et al. 2004). However, buoyant basaltic crust can transform into eclogite, which is denser than ambient mantle material. This transformation, which occurs only after burial to at least 40 km, i.e., after subduction, essentially makes oceanic lithosphere of all ages negatively buoyant.

The above buoyancy calculations assume that the lithosphere stays coherent. Modern continental subduction (e.g., India or Arabia under Eurasia) suggests that perhaps the buoyant continental upper crust delaminates from the subducting slab, which then becomes negatively buoyant, and can continue to subduct (e.g., Capitanio et al. 2010). A similar mechanism has been suggested for buoyant Archean oceanic lithosphere (Davies 1992).

2.2.3. Material strength. Modern plate tectonics has plates that are strong enough to remain coherent for a geologically significant time but weak enough to bend into a subduction zone, and it has slabs that are mostly coherent. Rocks deform in a brittle, elastic, or viscous way (or a combination of those), with viscous deformation dominant in warm areas away from the surface (**Figure 1**). Viscous deformation of mantle or crustal material is commonly described by a set of deformation rates (or strain rates) $\dot{\varepsilon}_i$ (Hirth & Kohlstedt 2003, Korenaga & Karato 2008):

$$\dot{\varepsilon}_i = A\sigma^n C_{OH}^r \exp(\alpha\varphi) \exp\left(-\frac{E^* + pV^*}{RT}\right), \tag{2}$$

with stress σ, stress exponent n, material water content C_{OH}, melt fraction ϕ and related constant α, activation energy E^* and volume V^*, universal gas constant R, and absolute temperature T. Different rocks, materials, and phases have different parameters and different strain rates for the same T, p, C_{OH}, and σ. Furthermore, multiple deformation mechanisms i are operating simultaneously, with diffusion creep and dislocation creep the most relevant ones for mantle and lithosphere deformation. The total deformation is the sum of those $\dot{\varepsilon}_{tot} = \sum \dot{\varepsilon}_i$, and the effective material viscosity is calculated as $\eta_{eff} = \frac{\sigma}{\dot{\varepsilon}_{tot}}$. In cold, deforming areas, stresses might exceed the material's brittle or plastic strength σ_y, which results in either brittle failure or low-temperature plasticity (Kohlstedt et al. 1995), reducing the material's effective viscosity $\eta_{eff} \leq \frac{\sigma_y}{2\dot{\varepsilon}_{tot}}$. See Billen (2008) for a review.

Mantle temperature has several, partly counteracting effects on plate and slab strength, each embedded in Equation 2. First, as with most natural materials, rock strength depends exponentially on ambient temperature directly through E^*, typically $\sim$375 kJ mol^{-1} for diffusion creep and 510–520 kJ mol^{-1} for dislocation creep, each $\pm$50–75 kJ mol^{-1} (Karato & Wu 1993, Hirth & Kohlstedt 2003). As a rule of thumb, every $\Delta T_{\mathrm{pot}} = 100$ K weakens rock by one order of magnitude. Second, weaker asthenosphere indirectly enhances development of sublithospheric small-scale convection, which reduces lithospheric thickness [thus explaining modern seafloor flattening; see, e.g., van Hunen et al. (2005) and references therein] and therefore plate strength.

Third, higher temperature also has various rheological consequences through increased melt production. Partial melting dehydrates and therefore strengthens rocks (Equation 2). Hirth & Kohlstedt (2003) report a 100-fold increase in strength during the wet-to-dry transition, although Korenaga & Karato (2008) point out significant uncertainty regarding this finding. A 100-fold increase probably significantly overestimates the actual increase during mid-ocean ridge melting because water concentrations in upper-mantle material away from subduction zones are much smaller than in the water-saturated conditions for which "wet" experiments are done (Hirth & Kohlstedt 2003, Hirschmann et al. 2009). Crustal rock strength is lower than in mantle material by several orders of magnitude (Kohlstedt et al. 1995, Ranalli 1995), so thick Archean oceanic crust reduces plate strength. Finally, partial melts in the rocks significantly reduce material strength, and Hirth & Kohlstedt (2003) report a factor-of-10 reduction in strength for 5% melt.

2.3. Models for the Viability and Vigor of Archean Subduction

Early work on the viability or vigor of Archean plate tectonics emerged from two contrasting models: (*a*) Mantle convective vigor increases with increasing mantle temperature, and, because plates form part of that convective cycle, plate tectonics might have been faster, too (e.g., Christensen 1985); and (*b*) plates were too buoyant to subduct, and Archean subduction may not have been viable at all (Nisbet & Fowler 1983). Since then, these models have been further elaborated.

2.3.1. Effect of plate buoyancy on subduction. Thick Archean oceanic crust had an older neutral-buoyancy age, before which spontaneous subduction was not viable (Davies 1992, van Thienen et al. 2004). However, subductibility involves more than just the concept of neutral buoyancy. Even buoyant plates subduct, as shown by subducting young oceanic plates and mid-ocean ridges today. This is partly because older parts of the plate pull down younger parts, but, more importantly, buoyant basaltic oceanic crust transforms into dense eclogite at depth and makes any slab negatively buoyant to provide the aforementioned slab pull. Even buoyancy from considerably thicker basaltic oceanic crust near the surface would be small compared with that from slab pull. Therefore, for ongoing, deep subduction, crustal eclogitization at $\sim$40 km effectively eliminates the buoyancy problem (Korenaga 2006, van Hunen & van den Berg 2008). However, crustal buoyancy could still have been important for two reasons. First, in the absence of slab pull (e.g., at the birth of a new subduction zone or after recent slab break-off), crustal buoyancy becomes a dominant force and could prohibit subduction (re)establishment (van Hunen & van den Berg 2008). Second, the coarse-grained, gabbroic part of oceanic crust can remain meta-stable to a larger depth and might not transform into denser eclogitic phases until 60–150 km (Hacker et al. 2003). This might play a key role in modern shallow flat subduction (van Hunen et al. 2002).

Archean crustal settling has also been proposed to solve this buoyancy problem (Davies 2006, 2008). Today, subducted oceanic crust eventually mixes back into the mantle by convective stirring (Allègre & Turcotte 1986), which allows it to remelt at the next passage through a melting zone. However, perhaps eclogitic MORBs were too dense for the weaker Archean mantle to be convectively entrained and thus sank to the core-mantle boundary. This would leave the Archean

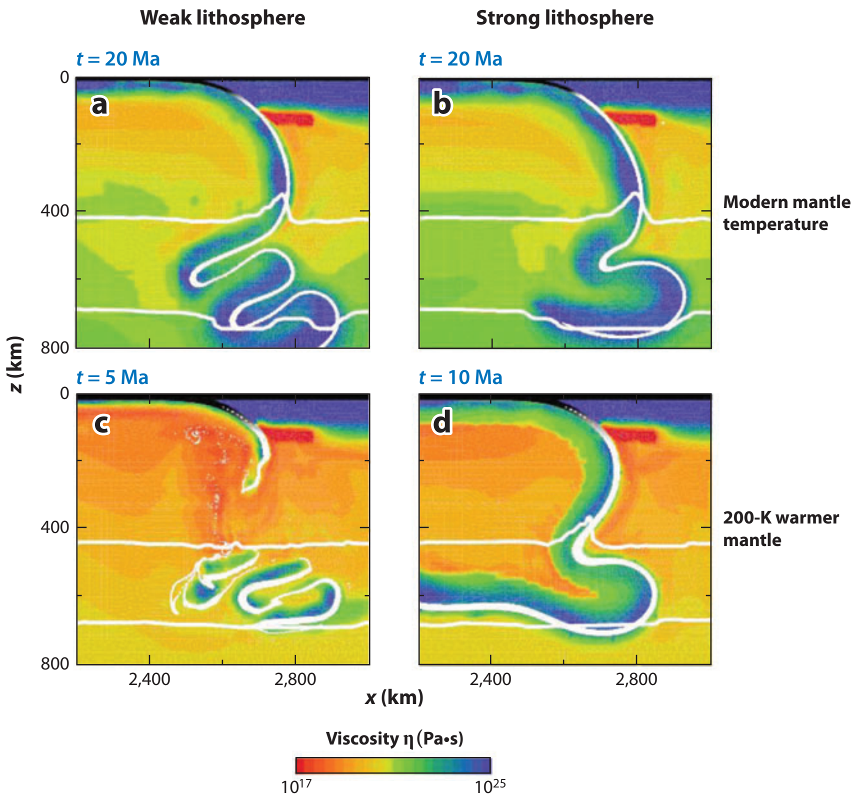

Figure 2.

Numerical model results for Archean subduction dynamics. (*a*) $\Delta T_{pot} = 0$ K, $d_{cr} = 7$ km, no dehydration strengthening. (*b*) As in panel *a*, but with a 100× stronger depleted mantle lithosphere, which increases the effective plate thickness. (*c*) $\Delta T_{pot} = 200$ K, $d_{cr} = 7$ km, no dehydration strengthening. The subducting plate is too weak to sustain continuous subduction, resulting in frequent slab break-off and intermittent subduction with a typical million-year (Ma) subduction time interval. (*d*) As in panel *c*, but for a 100× stronger depleted mantle lithosphere; no slab break-off occurs. For model details and parameters, see figures 3 and 8 in van Hunen & van den Berg (2008). Abbreviations: d_{cr}, oceanic crustal thickness; T_{pot}, mantle potential temperature.

mantle more depleted and would result in d_{cr} comparable with today's, despite higher mantle temperatures. However, the more fertile early Archean mantle (Condie 2005a) argues against this model. Furthermore, whether crust would settle in a vigorously convecting, hotter mantle has been debated (Brandenburg & van Keken 2007).

2.3.2. Effect of plate and mantle strength on subduction. Enhanced, deep melting at the ridge could have resulted in a thicker, strong Archean depleted mantle lithosphere (Korenaga 2006) that perhaps required more energy to bend and could have resulted in slower subduction. Korenaga (2006) illustrates how this fits with the longer time interval between older supercontinents. However, as discussed above and further elaborated below, various plate-weakening mechanisms might compensate—or overcompensate—for this dehydration strengthening (**Figure 2**) (van Hunen & van den Berg 2008). Furthermore, lithosphere is rheologically layered, with a strong core between weaker layers (Kohlstedt et al. 1995); this layered structure would reduce bending energy dissipation (Capitanio et al. 2007). Finally, plate radii will likely adjust to plate-bending forces, such that strong plates would have increased bending radii up to 500 km; in this case, bending forces remain small and do not significantly slow down subduction (Davies 2009).

Recently developed, fully dynamic early-Earth subduction models explicitly account for mantle and plate deformation (van Hunen & van den Berg 2008, Sizova et al. 2010). These models evaluate the combined effect of changing plate buoyancy and rheology on the viability of the subduction process in a hotter Earth (**Figure 2**). Unlike the models described in Korenaga (2006), these show that hotter mantle subduction with thick, strong mantle lithosphere and thick, weak oceanic crust is faster than modern subduction. Models without such large dehydration strengthening result in much weaker slabs that do not have the strength to stay coherent. Consequently, frequent slab break-off occurs (**Figure 2**), resulting in a more episodic style of Archean subduction, with typical subduction time intervals of a few Ma (van Hunen & van den Berg 2008). Break-off is possibly enhanced by the larger tensile stress between the thicker buoyant basaltic oceanic crust near the surface and its dense eclogitic counterpart at depth. Such spontaneous slab break-off contrasts with its modern analog, which is almost exclusively associated with continental collision (e.g., Wortel & Spakman 2000, van Hunen & Allen 2011). Time-averaged subduction velocities (van Hunen & van den Berg 2008) illustrate how plate speed is limited by mantle friction for $\Delta T_{pot} < 150$ K (consistent with Davies 1999). For $\Delta T_{pot} > 150$ K, frequent slab break-off—and therefore slab strength rather than mantle strength—limits average subduction velocity. These results correspond qualitatively with Sleep (2000), who maintains that plate tectonics operates best in a given mantle temperature range, and with Sizova et al. (2010), who recognize changing subduction regimes from no-subduction regimes ($\Delta T_{pot} > 200$–250 K), to presubduction regimes without modern one-sided subduction due to weak plates, to modern-style subduction regimes for $\Delta T_{pot} < 175$–160 K.

2.4. Subduction and Consequences for Earth's Thermal Evolution

Therefore, mantle temperature influences subduction dynamics, but mantle temperature, in return, is also influenced by subduction dynamics through its effect on surface heat flow. Useful estimates of the secular cooling of Earth can be made using the following simplified heat balance (e.g., Christensen 1985, Sleep 2000, Korenaga 2006):

$$C\frac{dT_{pot}(t)}{dt} = H(t) - Q(t), \tag{3}$$

where C is the total heat capacity of Earth, $H(t)$ is the radiogenic heat production, and $Q(t)$ is the total surface heat flux. Today, Earth releases approximately 41–46 terawatts (TW) of heat (Jaupart et al. 2007), of which >60% comes from oceanic lithosphere. Both $H(t)$ and $Q(t)$ are reduced by the production of continental crust heat, which immediately escapes Earth and does not provide significant heating of Earth's interior. This leaves a present-day heat flow of $Q_0 \approx 36$ TW. The present amount of radiogenic heat production H_0 is a significant fraction of that (expressed as the Urey ratio, which is defined as $Ur = H_0/Q_0$), whereas the remaining heat flux cools Earth. Therefore, H_0 is essential to calculate a thermal history for Earth but, unfortunately, is subject to considerable debate: The present ratios of the four main heat-producing systems (^{238}U, ^{235}U, ^{232}Th, and ^{40}K) and their half-lives are known (e.g., Turcotte & Schubert 2002), but absolute quantities are not. Cosmochemical and petrological constraints suggest that $Ur \approx 0.21$–0.49 (Jaupart et al. 2007). $H(t)$ is calibrated with $H_0 = Ur\, Q_0$, with Ur as the only significant uncertainty.

The relevance of solving Equation 3 becomes clear with the need to define $Q(t)$. Because Q_0 is largely controlled by today's plate tectonics (Jaupart et al. 2007), $Q(t)$ requires information

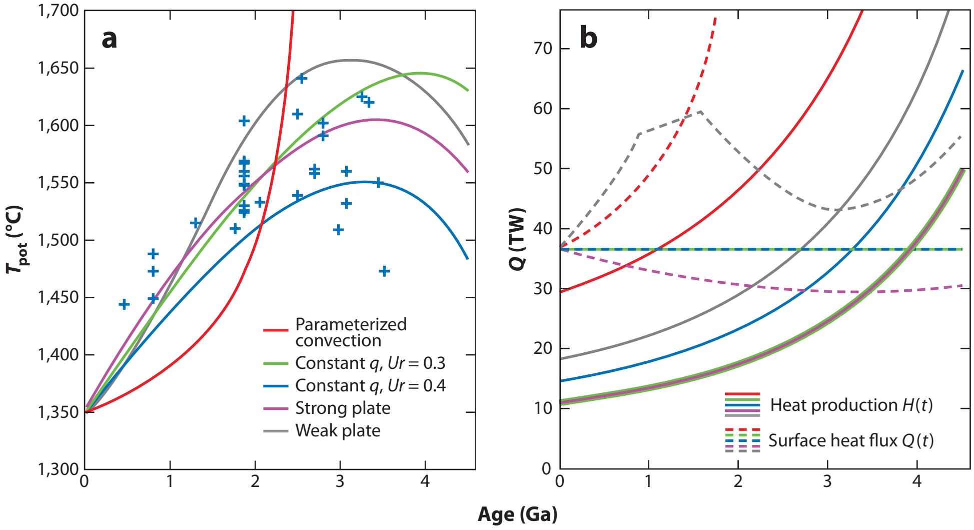

Figure 3

(*a*) Modeled thermal evolution for Earth using different conceptual models: a parameterized convection model (using $Ur = 0.8$ and equations 9 and 10 and table 1 from Davies 2009); two constant heat-flow models (using $Ur = 0.3$ and $Ur = 0.4$); a strong-plate model (using $Ur = 0.3$, after Korenaga 2006, and equations 6–13 and table 1 from Davies 2009, with the correction to substitute $d = d_p$ only on the right side of his equation 6); and a weak-plate model based on dynamical subduction modeling (using $Ur = 0.5$ and the default subduction velocity from van Hunen & van den Berg 2008, figure 9). For all models, Equation 3 is integrated through time numerically for given C, $H(t)$, and $Q(t)$ with $Q_0 = 36$ TW, equivalent to 160 K Ga^{-1} cooling for $Ur = 0$ (with a mass $M = 6 \times 10^{24}$ kg and average heat capacity $K =$ 1,200 J kg^{-1}, which give $C = 7.2 \times 10^{27}$ J K^{-1}). Data points are petrological T_{pot} estimates for nonarc lavas (excluding komatiites) from Herzberg et al. (2010). (*b*) $Q(t)$ (*dashed line*) and $H(t)$ (*solid line*) for each model in panel *a*. Abbreviations: C, total heat capacity of Earth; $H(t)$, radiogenic heat production; $Q(t)$, total surface heat flux; T_{pot}, mantle potential temperature; TW, terawatt; Ur, Urey ratio (defined as $Ur = H_0/Q_0$).

about the (strongly debated) past (plate-)tectonic vigor. This debate essentially boils down to the question: What limits plate-tectonic vigor? For today, a model in which plate motion is governed by mantle energy dissipation in the mantle works well (e.g., Davies 1999), but extrapolating this concept to Earth's past may result in unsatisfactory thermal history results. Below are the main models, and their typical thermal evolutions are shown in **Figure 3**:

1) Parameterized convection model: In this model, plate-tectonic vigor and surface heat flow are controlled by energy dissipation from mantle convection (e.g., Christensen 1985, Davies 2009). A hot mantle gives faster convection and fast cooling, requiring a hotter mantle to start with. This positive feedback may result in unrealistically high Archean mantle temperatures. Using $Ur \geq 0.8$ (see **Figure 3**) or assuming that Q_0 is larger than the geologically recent $Q(t)$ could resolve this problem (Davies 2009). Furthermore, including Earth layering (core-mantle or upper/lower-mantle layered convection) delays cooling (Davies 2008, 2009) and results in more modest mantle temperatures throughout most of Earth history. Nonetheless, cooling curves have a poor fit to the petrological estimates of past mantle temperatures (Herzberg et al. 2010). Therefore, perhaps early plate-tectonic vigor was not, as it is today, controlled by mantle resistance—a concept that is intrinsic in all models described below.
2) Constant heat-flow model: In absence of past heat-flow data, the basic assumption here is that $Q(t) = Q_0 = 36$ TW is constant, perhaps with $Q(t)$ fluctuating on timescales much

shorter than any typical thermal adjustment time for Earth (Labrosse & Jaupart 2007). If the underlying physical mechanism is not specified, plate-tectonic vigor and surface heat flow are assumed to be independent of mantle temperature. Despite the absence of a physical basis for this constant heat-flow assumption, **Figure 3** shows a good fit and available potential temperature data for reasonable *Ur* values.

3) Strong-plate model: If Archean plates were stronger (as discussed above), plate velocity was perhaps limited by energy dissipation from plate bending at subduction (Korenaga 2006) with slower subduction and less surface heat flow. Cooling curves from this model fit well with *Ur* estimates and T_{pot} data (Herzberg et al. 2010; see also **Figure 3**). However, as discussed above, whether Archean plates were compositionally stronger than today is debated.
4) Weak-plate model: If plates were weaker in a hotter mantle (as discussed above), spontaneous break-off of slabs may have hampered ongoing subduction, reducing the average subduction velocity for $\Delta T_{pot} > 100$–200 K (van Hunen & van den Berg 2008). Rapid recent cooling requires a relatively high *Ur* at 0.4–0.5, but T_{pot} remains considerably below 1,700°C and fits the available data well.

Plate tectonics is not necessarily the only mechanism capable of cooling Earth throughout most of its history; extensive flood volcanism in a stagnant-lid setting, for example, is a viable alternative (van Thienen et al. 2004). All discussed models have significant uncertainties from poorly known model parameters. Rheological uncertainties, in particular, form the basis for these different models. Furthermore, constant continental surface area is implicit in all models, despite significant debate.

3. FIELD CONSTRAINTS FOR SUBDUCTION

Modern plate tectonics is evidenced by relative displacement of rigid blocks, separated by plate boundaries that exhibit the most geological activity. From a geologist's perspective, occurrence of plate tectonics can therefore be demonstrated by the existence of (former) plate boundaries. Divergent plate boundaries are difficult to identify because relicts of mid-ocean ridges tend to be subducted, but subduction produces long-lived geological markers that effectively shape large portions of the continental crust.

Fossil subduction markers broadly fall into three main groups: (*a*) geochemical markers from igneous rocks, suggesting formation in a subduction environment; (*b*) structural markers, such as accreted exotic terranes or thrusts, which juxtapose rocks of different origin; and (*c*) metamorphic markers, characterized by paired high-pressure and high-temperature metamorphic belts. The Archean rock record has no unambiguous evidence for any of these, but many features of Archean geology have been interpreted as a variation on a modern scenario. A key issue here is how much Archean subduction is expected to depart from its modern counterpart.

3.1. A Geochemical "Arc" Signature in Archean Continental Crust

The existence of a geochemical "arc" signature in igneous rocks is commonly used as evidence for Archean subduction. In the following paragraphs, we describe the geochemical components of the so-called "arc" signature and discuss whether or not it really implies a subduction-related (continental or oceanic arc) environment.

3.1.1. Nature and origin of the arc signature in modern rocks. The term "arc" signature is colloquially used to describe common geochemical properties (mostly trace elements) in modern island or continental arcs. The key defining element is the decoupling between large-ion lithophile

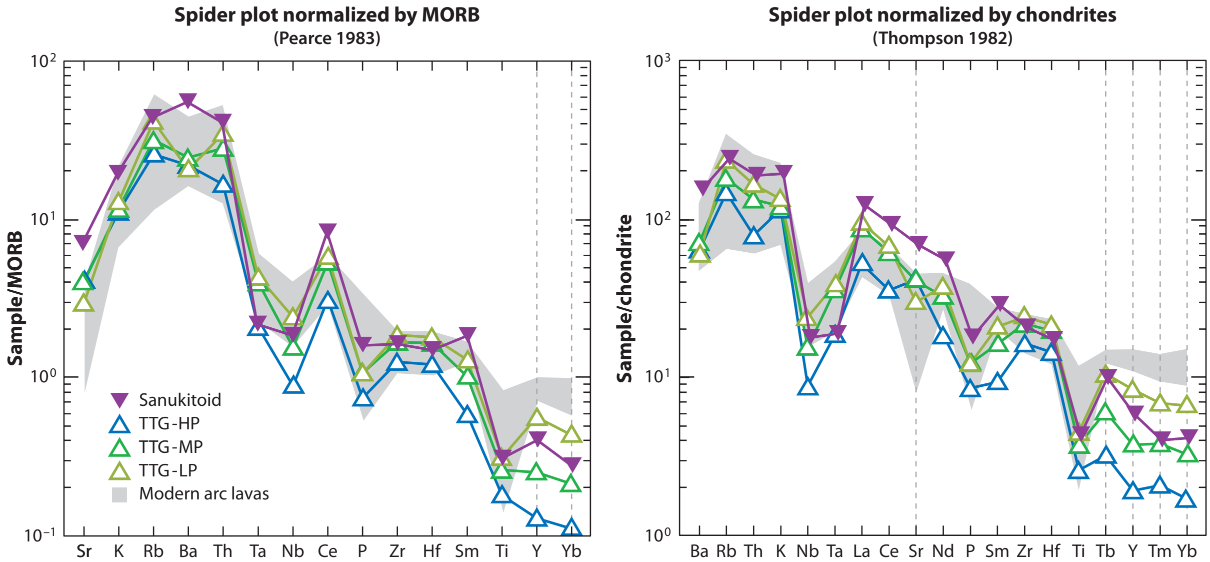

Figure 4

Average composition of low-pressure (LP), medium-pressure (MP), and high-pressure (HP) tonalites-trondhjemites-granodiorites (TTGs) (Moyen 2011) and sanukitoids (Martin et al. 2010) compared with modern arc rocks (*gray field*; data from GEOROC, **http://georoc.mpch-mainz.gwdg.de/**). Abbreviation: MORB, mid-ocean ridge basalt.

elements (LILE) and high-field-strength elements (HFSE). Relative to MORBs, arc rocks have comparable HFSE but higher LILE contents (Pearce 2008). For example, low contents of niobium (Nb) and tantalum (Ta), two HFSE, result in a pronounced negative Nb-Ta anomaly and corresponding low Nb/Th ratios (typically <5) for arc rocks (Rollinson 1993) in a chondrite- or MORB-normalized diagram (see **Figure 4**).

The arc signature is defined primarily on the basis of trace elements. All igneous rocks, from basalts to rhyolites, can show this character, but they cannot form through one unique process. Hereafter, we discuss separately the origins of arc signatures in mafic rocks (which can be generated by partial melting of the peridotitic mantle) and the origins of arc signatures in intermediate to felsic rocks (which cannot be primary mantle melts).

Mafic arc basalts and andesites are enriched in LILE but not in HFSE, compared with MORBs (Pearce & Peate 1995), suggesting enrichment of the mantle source by a LILE-rich, HFSE-poor component. Their generally accepted model involves fluid-fluxed melting of the upper mantle (e.g., Tatsumi & Eggins 1995, Tatsumi 2005), with the fluid-mobile LILE transported from slab to mantle wedge by fluids. Other subduction-related processes can generate LILE/HFSE decoupling for mantle-derived rocks, such as transport by silicate melts rather than fluids or by mechanical mixing of sediments, although the latter processes generate a less pronounced LILE/HFSE decoupling (e.g., niobium-enriched basalts in Sajona et al. 1996; low-silica adakites in Martin et al. 2005).

An arc signature can be obtained in situations that differ from genuine subduction. Continental crust is made of rocks bearing an arc-like composition (Taylor & McLennan 1985, Rudnick & Fontain 1995, Davidson & Arculus 2006). Therefore, mixing between "nonarc" material and continental crust yields an apparent arc signature irrespective of the site where it takes place; such a signature could be originated by (*a*) continental matter buried in the mantle, by subduction or delamination; (*b*) contamination of a nonenriched basalt by the continental crust through which

Large-ion lithophile elements (LILE): large cations that have small positive charges (e.g., Cs, Rb, K, Ba, Pb, and Sr) and that are fluid mobile at >500°C

High-field-strength elements (HFSE): small cations that have large positive charges (e.g., Zr, Hf, Ti, Nb and Ta, and Sc) and that are typically both immobile during low-temperature alteration, and not fluid-mobile at high temperature

Tonalite-trondhjemite-granodiorite (TTG): the most common component of Archean crust, this group of sodic granitoids are broadly calc-alkaline and have trace elements broadly similar to arc rocks

it ascends (Arndt & Jenner 1986, Barley 1986, Green et al. 2000); or (*c*) the mantle itself being a nondepleted peridotite. It is commonly accepted that present-day depleted upper mantle (the depleted MORB mantle, or DMM, in, e.g., Hofmann 1988) is the result of the extraction of ∼2% continental crust (Hofmann 1988) from a hypothetical, chondritic primitive mantle (McDonough & Sun 1995). If, before significant continental crust extraction, Archean mantle was still partly primitive, as suggested by the systematic variation of Archean basalts (Condie 2005a), then all melts from this source would show some features of the continental crust—i.e., they would have weak arc characteristics.

Felsic arc rocks cannot be (directly) generated by partial melting of the mantle, and other sources must be envisaged. Possible models include:

- Differentiation of arc basalts: Intracrustal processes such as fractional crystallization or mixing with preexisting crust cannot significantly affect LILE/HFSE ratios. Consequently, an arc signature cannot (easily) be generated from the differentiation of a "nonarc" basalt.
- Melting of amphibolites (e.g., Foley et al. 2002): Low-pressure melting (i.e., within the crust) has similar problems. Namely, only amphibole can fractionate LILE from HFSE, such that it is not possible to generate a pronounced arc signature without an "arc-like" source. At high pressures, however, with rutile stable during melting (review in Moyen & Stevens 2006), fractionation of HFSE can perhaps become efficient enough to bring the melt composition into the "arc" domain, as proposed for high-pressure tonalites-trondhjemites-granodiorites (TTGs) (Moyen 2011).
- Inherited arc signature: Any rock generated from an "arc-like" source within the crust will retain an arc signature, even without (active) subduction. For example, all granites generated from arc crust will be "arc-like," even if they form after the cessation of subduction. Considering the overall arc signature of the continental crust (Rudnick & Fontain 1995), this implies that a vast majority of granitoids will have some arc characteristics.

3.1.2. Archean rocks with an arc signature. A large proportion of Archean igneous rocks show some enrichment of LILE over HFSE. Many Archean assemblages have therefore been interpreted as "arc-related" or "arc-influenced."

Archean greenstone belts are dominated by mafic and ultramafic lavas that form thick volcanic piles. For those rocks, the GEOROC database reveals that (*a*) only very uncommon Archean rocks plot in the fields of modern nonarc rocks (Condie 2005a), (*b*) ∼40% of Archean mafic rocks have a thorium/niobium ratio greater than 0.2 and plot in the field of modern arc rocks, and (*c*) the remaining ∼55–60% of Archean mafic rocks plot in between the fields of modern arc basalts and MORBs.

Typically two solutions are proposed to explain the "arc" or "arc-like" trace-element signature of many Archean mafic rocks: (*a*) melting in a subduction-like environment (e.g., Polat & Kerrich 2006); and (*b*) contamination of mantle-derived lavas by preexisting continental crust (Arndt & Jenner 1986, Barley 1986, Green et al. 2000). Above we discuss a perhaps simpler solution (Condie 2005a) of deriving an arc signature from the melting of an enriched mantle source (from which no continental crust has yet been extracted), with little need for subduction or crustal contamination.

On the other hand, some uncommon Archean rock types, ranging from mafic to intermediate, are identified as similar to minor rock types of modern subduction. They include boninites (Smithies et al. 2004a) and shoshonites (Kerrich & Ludden 2000), as well as niobium-enriched basalts, magnesian andesites, and adakites (Polat & Kerrich 2004). Although not all vaguely enriched mafic rocks carry an "arc" signature, this specific subset clearly represents a distinct set of Archean rocks. Their arc signatures are much more pronounced than those of Archean basalts, suggesting some sort of mantle fertilization by a fluid or crustal component. Furthermore, these arc rocks tend to occur as packages, in which different rock types with similar characteristics are closely associated with one another in both space and time (Polat & Kerrich 2004, 2006; Smithies et al. 2004b), suggesting a process involving the burial of surface matter to 50–100-km depth.

The dominant geochemical evidence used to infer Archean subduction comes from plutonic rocks from the TTG series (Martin 1994, Moyen 2011). TTGs are similar to arc rocks [although TTGs are notably depleted in heavy rare-earth elements (HREE) and have elevated strontium contents compared with typical arc lavas], and this similarity has been widely used as evidence for a subduction-related origin. However, any environment allowing mafic rocks to melt is equally suitable, such as an oceanic plateau (Willbold et al. 2009) or a mid-oceanic ridge (Rollinson 2009). This purely comparative approach is therefore a dead end, and a deeper understanding of TTG genesis must be sought.

Heavy rare-earth elements (HREE): the lanthanides Gd to Yb; they preferentially partition into garnet (and amphibole to a lesser degree) during melting or fractionation

There is a large consensus that TTGs are generated by partial melting of a mafic source (Moyen & Stevens 2006, Moyen 2011) and that HREE depletion in TTGs reflects stability (during melting) of a phase that sequestrates HREE. Only garnet seems a realistic possibility, implying a melting pressure of >10–12 kbar (Moyen & Stevens 2006). In fact, three main groups of TTGs are identified (Moyen 2011), with compositional differences reflecting the stable assemblage during melting and therefore the depth of the source of these melts. The first is a low-pressure group formed at <12 kbar, which hardly requires subduction but can form at the base of a thick crust or a mafic plateau (Willbold et al. 2009). The second, a medium-pressure group, is more ambiguous. The third is a high-pressure group that requires melting at >18–20 kbar, which is difficult to attain without burial of basalt from the surface into the mantle. Such deep melting can therefore be taken as evidence for subduction (Moyen 2011).

Another type of granitoid, the sanukitoid suite (see review in Martin et al. 2010), occurs in the late Archean, from ~2.9 Gya onward. Sanukitoids probably result from interactions between the mantle and felsic melts (Martin et al. 2010), either TTG melts generated by the melting of buried (subducted?) basalts or melts from buried sediments. Sanukitoid formation therefore requires a mechanism to carry crustal rocks (basalt or sediments) down into the mantle.

3.1.3. The meaning of Archean arc signature. Therefore, for some rocks, the (geochemical) case for Archean subduction is perhaps not as strong as previously claimed. Mafic arc signatures can be explained by crustal contamination or a perhaps nondepleted mantle source. Some TTGs formed by low-pressure melting, not necessarily in a subduction environment. However, some other rocks (shoshonites, boninites, andesites, sanukitoids, adakites, and high-pressure TTGs) carry a more convincing arc signature. Despite their differences, all these rocks formed by deep (>50-km) melting using surface material, and it is tempting to interpret all these arc signatures as reflecting Archean subduction. However, it must be stressed that geochemistry is "blind" to the geodynamic environment. Igneous rock composition is determined by the source and the melting (pressure, temperature, fluid) conditions, irrespective of the mechanism. Subduction obviously fits all the geochemical constraints, but so does any model based on discontinuous, downgoing parcels of crustal rocks [e.g., "drips" or delamination (Bédard 2006)]. However, purely intracrustal models (e.g., those that involve melting, differentiation, or contamination of ordinary basalts) are ruled out by the requirement for (*a*) a deep (>50-km) evolution and (*b*) the presence of mantle and crustal material during generation of even primary magmas.

3.2. Fossil Subduction Zone Structures

Modern subduction generates several key structural features. Subduction is associated with thrusting. A main thrust with many subordinate thrust planes collectively forms the accretionary wedge, at scales ranging from a sedimentary accretionary wedge (affecting a fore-arc sedimentary basin) to a crustal-scale orogenic wedge (with crustal fragments decoupled from the underlying lithospheric mantle). Structurally, subduction therefore results in juxtaposition of different units (tectonic mélange, accretionary wedge, or orogenic wedge) in a fold-and-thrust belt or terrane accretion, with small (typically former arc) lithospheric terranes juxtaposed along mostly transpressive contacts with limited crustal thickening (Coney et al. 1980).

The map pattern of many, if not most, Archean cratons is strongly reminiscent of accreted terranes. In the Superior Province in Canada, for instance (Percival et al. 2002), a dozen East-West elongated terranes are juxtaposed along tectonic boundaries; each terrane (or subprovince) has a distinct age and lithology (**Figure 5**). Similar patterns are observed in neo-Archean (Yilgarn) as well as meso-Archean (West Pilbara, Kaapvaal) continents.

Shallow Archean fold-and-thrust belts, accretionary wedges, and tectonic mélanges are difficult to recognize [see, however, possible examples in Greenland (Windley & Garde 2009)]. It has been suggested that the dominant structures are best interpreted as bulk coaxial shortening of the crust (Chardon et al. 2009) and gravity-driven tectonics involving the sinking of dense greenstone belts into the soft basement (Gorman et al. 1978, Bouhallier et al. 1995, Collins et al. 1998), both of which suggest deformation of a hot, "soft" Archean lithosphere (Choukroune et al. 1995).

However, strong evidence for thrusting in the paleo-Proterozoic and Archean crusts is observed in seismic reflection profiles that do invariably show flat reflectors (Ludden & Hynes 2000, de Wit & Tinker 2004, Goleby et al. 2004, Korja & Heikkinen 2008). Most interpretations describe shallow-dipping structures, which sometimes extend into the subcontinental mantle (Calvert et al. 1995). These structures are considered as fossil thrust planes (**Figure 5**) that could be the remains of paleosubductions. Alternately, structural evidence suggests that a hot-orogens model is more appropriate for the deformation of Archean crust (Cagnard et al. 2006, Rey & Coltice 2008). In this model, crust is too weak to thicken significantly (England & Bickle 1984, Rey & Houseman 2006); instead, convergence is accommodated by bulk shortening and lateral flow, superimposed on important vertical displacements. Deformation is distributed, and lateral flow of the lower crust generates the flat fabric observed in seismic profiles (Chardon et al. 2009). A hot orogen can behave as an accretionary system, in which crustal blocks of contrasting origin are tectonically juxtaposed.

3.3. Metamorphism

The key element of modern subduction is the quick burial of cold lithosphere, resulting in cold geothermal gradients (low-*dT/dP*, or high-pressure–low-temperature, HP-LT) and metamorphism to blueschist and eclogite facies (Ernst 1988, Brown 2009). The low-*dT/dP* assemblages form a narrow belt that is associated with, and tectonically juxtaposed to, a low-pressure–high-temperature (LP-HT) metamorphic belt inboard, forming paired metamorphic belts. Continental material may be temporarily subducted (forming ultra-high-pressure, crust-derived metamorphic rocks) and subsequently exhumed. In an orogenic wedge, this yields clockwise pressure-temperature paths into granulite facies conditions, pervasive crustal melting, and retrogression of the early high-pressure assemblages, the latter of which are typically preserved only as small boudins and inclusions in migmatitic gneisses.

Archean metamorphism has been relatively understudied since the 1980s owing to the general consensus that Archean metamorphism is universally and homogeneously hot and uninformative. However, new internally consistent thermodynamics databases (Holland & Powell 1998) and modeling tools allow the extraction of new valuable information, including spectacular discoveries of Archean relatively low-*dT/dP* rocks. Examples of these (see also **Figure 6)** include the

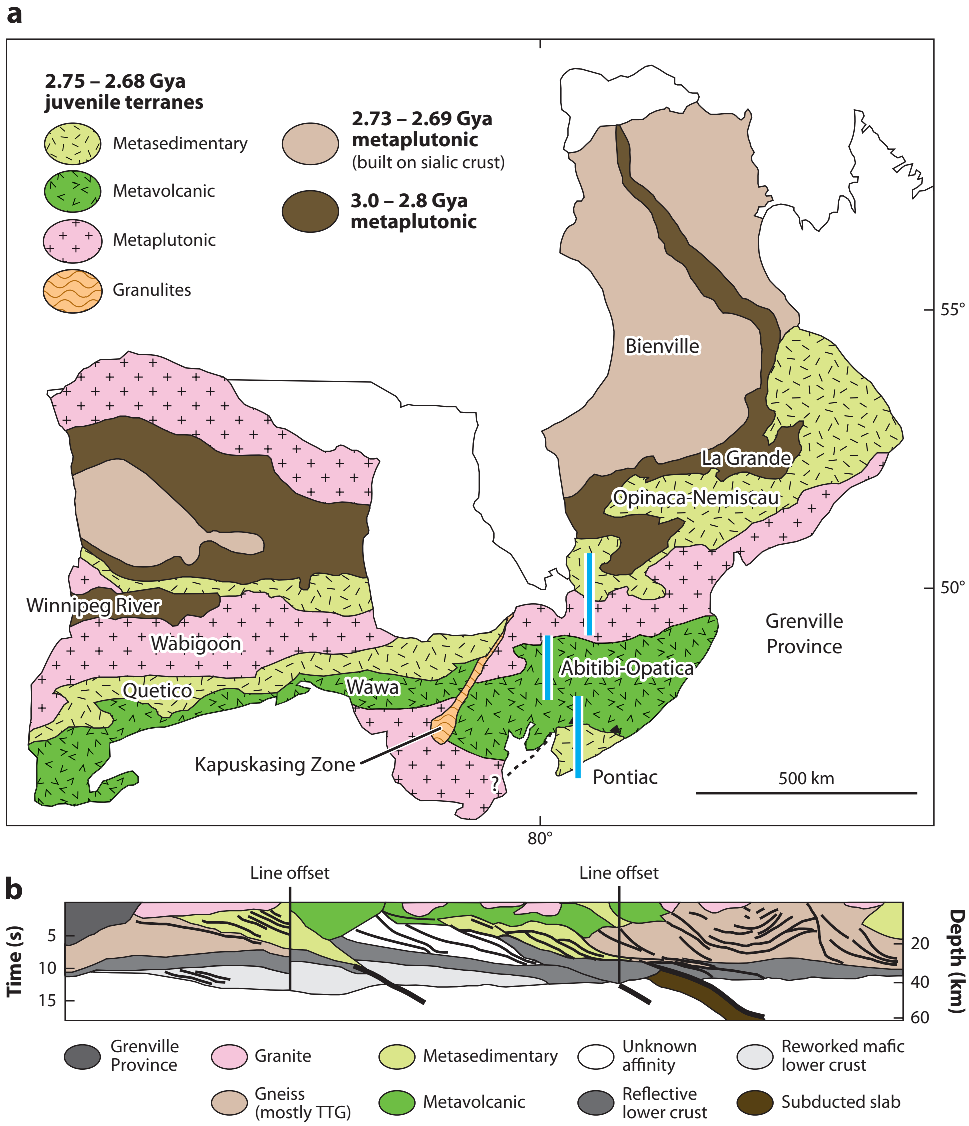

Figure 5

The Superior Province, as a potential fossil subduction zone. (*a*) Geological map of the Superior Province showing the successive terranes (adapted from Benn & Moyen 2008). The thick vertical blue lines show the approximate trace of the LITHOPROBE profile (Ludden & Hynes 2000). (*b*) The LITHOPROBE Abitibi profile (after Ludden & Hynes 2000; redrawn in Benn 2006), interpreted as an orogenic wedge with imbricated thrust planes. Abbreviation: TTG, tonalite-trondhjemite-granodiorite.

Baltic Shield, ~2.87 Gya (Volodichev et al. 2004); Barberton, ~3.2 Gya (Moyen et al. 2006, Stevens & Moyen 2007); and Bundelkhand craton, India (Saha et al. 2010a). A well-defined tectonic boundary separates the Barberton low-*dT/dP* (~12–15°C km^{-1}) and high-*dT/dP* rocks (Dziggel et al. 2006), defining a classical system of paired metamorphic belts (Stevens & Moyen 2007).

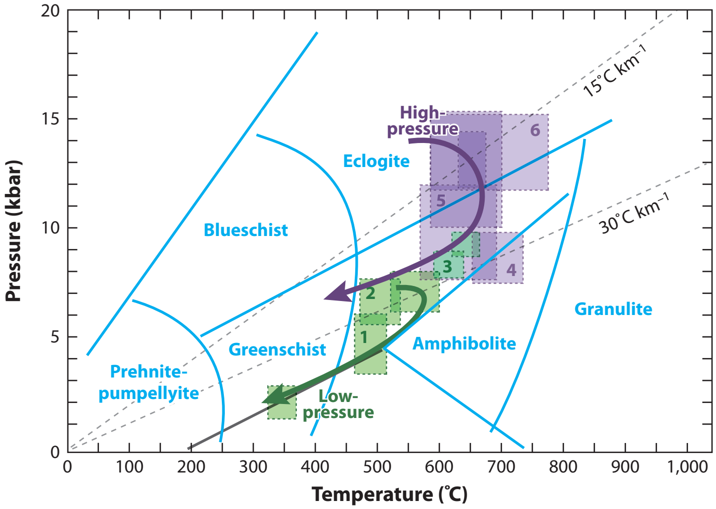

Figure 6

Metamorphic conditions in the Kaapvaal craton. 1, 3: Murchison greenstone belt, ~3.0 billion years ago (Gya) (S. Block, J.-F. Moyen, A. Zeh, M. Poujol & J. Jaguin, et al., unpublished data); 2, 4, 6: Barberton, ~3.2 Gya (Stevens & Moyen 2007, Lana et al. 2010); 5, de Kraalen, ~3.2 Gya (Saha et al. 2010b).

Archean low geotherms (~12°C km^{-1}) are higher than modern ones (5–8°C km^{-1} for blueschists) and occur only as fragments in higher-geotherm migmatitic gneisses, but this does not rule out subduction. First, modern subduction is fairly diverse, not uniformly showing well-preserved, blueschist-type low geotherms (Peacock et al. 2005); perhaps comparison with hotter modern orogenies such as the Caledonides or Hercynides is better. Second, Archean paired metamorphic belts (Brown 2006) do occur (Stevens & Moyen 2007) but are shifted toward hotter temperatures, which is perhaps unsurprising in a hotter mantle.

An alternative model to explain the absence of blueschists and eclogite in the Archean metamorphic record is related to the aforementioned weakness of the subducting slabs. In such a scenario, the resulting frequent break-off might remove the driving force for deep continental subduction and the corresponding common occurrence of ultra-high-pressure, crust-derived metamorphic rocks (van Hunen & van den Berg 2008, van Hunen & Allen 2011).

4. DISCUSSION: SECULAR CHANGE IN STYLE OF SUBDUCTION

4.1. Archean Flat Subduction?

One popular model for Archean subduction is the shallow flat subduction style (Abbott et al. 1994b), in which subducting slabs move subhorizontally below the overriding plate. Motivations for this model are both geochemical and geodynamical in nature.

Geochemically, this model hinges on the observation that typical TTGs have Mg#, chromium, and nickel contents comparable with those of experimental melts (melting of amphibolites) but lower than those of modern adakites (see Section 3.1) (Smithies 2000, Martin & Moyen 2002, Condie 2005b). It was therefore proposed that adakites form under a mantle wedge (although alternative formation models exist; see Macpherson et al. 2006) and that they acquire their higher Mg# contents by chemical interactions with peridotites as they ascend (Rapp et al. 2006, 2010). TTGs show no such evidence and would form under much shallower conditions, during flat subduction without a mantle wedge, or at the base of a thick plateau crust (Smithies 2000, Smithies et al. 2009). Mg# in TTGs could perhaps track progressive slab steepening during the Archean (Martin & Moyen 2002). However, some higher-Mg# rocks also happen to have low MgO or MgO+FeO (see Moyen 2011), and the correlation of Mg#, chromium, or nickel with SiO_2 in most rock suites makes comparing the Mg# of rocks having different SiO_2 contents meaningless. The increasing Mg# of TTGs described in Martin & Moyen (2002), in fact, reflects an expanding SiO_2 range (see their figure 2A), which probably correlates to a larger range of melting temperatures (Moyen & Stevens 2006, Moyen et al. 2010) and a geodynamical diversification for the TTG formation throughout the Archean (Moyen 2011). The geochemical case for Archean flat subduction is, therefore, not compelling.

Geodynamical arguments are inspired by relatively uncommon (~10%; Gutscher et al. 2000), seismically imaged present-day shallow flat subduction. Near most flat subduction zones, buoyant features (e.g., aseismic ridges, buoyant plateaus) exist, with oceanic crust significantly thicker than normal (e.g., ~18 km at the Nazca Ridge; McGeary et al. 1985). With the proposed thick Archean oceanic crust, shallow flat subduction would perhaps be the norm (Abbott et al. 1994b). Dynamical modeling argues for three important flat subduction ingredients (van Hunen et al. 2002, 2004; Gerya et al. 2009): (*a*) buoyant features embedded in a dense subducting plate, (*b*) advancing plate motion, and (*c*) slab suction. For viable Archean shallow flat subduction, we also have to consider each of those ingredients. Today's plate tectonics is driven mostly by negative plate/slab buoyancy, and if all plates were buoyant, as suggested for the Archean and discussed above, no significant driving force for subduction would exist, resulting in no subduction rather than flat subduction (van Hunen et al. 2004). Flat subduction occurs if the overlying plate overrides the slab faster than it can sink (the Stokes velocity, which scales linearly with mantle viscosity); a weaker mantle is less likely to flatten slabs in this way (van Hunen et al. 2004). Slab suction refers to the slab dragging down mantle wedge material and creating a low-pressure wedge into which the slab is pulled (e.g., Stevenson & Turner 1977). This force scales linearly with mantle wedge viscosity and was therefore probably less important in the Archean. Thus, Archean shallow flat subduction is neither geochemically required nor geodynamically viable, and we propose to abandon this model.

4.2. Slab Penetration into the Lower Mantle

Seismic tomography shows that some slabs penetrate into the lower mantle (at ~660 km), whereas others stagnate in this transition zone (e.g., Li et al. 2008). The endothermic spinel-to-perovskite phase transition provides a buoyancy force to cold material and a significant (by a factor of 10–100) viscosity increase to enable this stagnation and can lead to time-dependent downwelling (flushing) into the lower mantle (e.g., Tackley et al. 1993, Torii & Yoshioka 2007). Slab penetration is affected by slab strength and buoyancy (Zhong & Gurnis 1995, Cizková et al. 2002) as well as trench rollback (e.g., Christensen 1996). However, exact correlation between subduction parameters and slab penetration is a primary research topic at present (Fukao et al. 2009).

Davies (1995) proposed full Archean mantle layering that would periodically collapse into catastrophic mantle overturns. This would increase upper-mantle temperature and melting, which are perhaps responsible for episodic crust formation (Condie 1998). These overturns would become more frequent and less violent over time, evolving into present-day whole-mantle convection. Archean layering was perhaps enhanced by the basalt barrier mechanism (Davies 2008): Basalt is buoyant between ~660 km and 750 km, because of its deeper transformation to its dense lower-mantle phase, and may accumulate to form a barrier to slab penetration. Crust and mantle lithosphere do not easily separate with today's strong slabs and relatively thin crust (van Keken et al. 1996), but in the Archean perhaps they did, either because crust and mantle easily segregated in the low-viscous mantle (Davies 2008) or because oceanic crust was thicker.

4.3. Episodic Subduction?

The subduction process might have been more episodic than it is today. There are indications for both long-term episodic behavior (with a typical timescale of a few hundred million years) and short-term episodicity (on timescales of a few million years).

4.3.1. Long-term episodicity. Different mechanisms could lead to long-term episodicity of plate tectonics. Large-scale mantle flushing events would drastically heat the upper mantle (perhaps by >300 K; Davies 2008) and affect subduction dynamics (as discussed above), with perhaps a temporary global shutdown of plate tectonics. Indirectly, reduced lower-mantle temperatures increased core-mantle temperature contrasts, maybe triggering more mantle plumes that perhaps resulted in unsubductable large igneous provinces (Condie 2001). Alternatively, a weaker mantle may have led to insufficient stress concentration for continuous subduction, resulting in episodic subduction on a timescale of several hundred million years (O'Neill et al. 2007). Or perhaps early plate tectonics was too inefficient to remove radiogenic heat, periodically melting the upper mantle and activating mush ocean dynamics (Sleep 2000) until the mantle cooled enough to resume plate tectonics. Such a loop would stop when plate tectonics could effectively release all radiogenically produced heat.

There is some observational support for intermittent plate-tectonic activity. The (sparse) Archean paleomagnetic record shows that apparent polar wander angular velocities varied significantly and episodically (Strik et al. 2003, O'Neill et al. 2007, Piper 2010). Estimated mantle depletion rates, taken as a proxy for plate-tectonic activity, may provide additional constraints (Silver & Behn 2008).

4.3.2. Short-term episodicity. Weak Archean slabs might not have been strong enough to support their own weight, which might have resulted in frequent slab break-off (van Hunen & van den Berg 2008). Subsequent slab (pull) removal would reduce or (temporarily) stop plate convergence, so that subduction was perhaps episodic on a million-year-timescale interval (**Figure 2**). Field observations show that a significant portion of Archean crustal rocks carry an "arc" signature. However, nonarc signatures are recognized as well, in the form of komatiites and tholeites, which are commonly associated with a mantle plume origin. In Western Abitibi (Ontario, Canada), several short-lived "subduction" events lasting 5–10 Ma are interleaved in dominant "plume" lithologies (Benn & Moyen 2008, Moyen & van Hunen 2012). Other Archean cratons such as the West Pilbara and Barberton show similarly short-period alternating of magmatism types. These short-lived arc signatures are in contrast to the >100-Ma continuous arc signature of Phanerozoic magmatism, e.g., in the Andes. These results illustrate combined geodynamical and geochemical support for Archean short-lived (several-million-years timescale) episodic subduction behavior.

4.4. The Appearance of Archean Subduction

Assuming subduction did occur in the Archean, the differences between Archean and modern Earth would translate into potential geological and geodynamical differences. The hotter crust and/or lower stresses from reduced slab pull for frequent slab break-off perhaps resulted in less crustal thickening. The metamorphic record was shifted to higher temperatures, and, in the absence of enough slab pull, high-pressure rocks were rarer and were included in large amounts of migmatitic gneisses. Common slab break-off resulted in short, discontinuous bursts of arc magmatism (as opposed to modern, continuous arc activity). Slab melting was favored over slab dehydration, thereby generating fewer andesites (related to slab dehydration) but more TTGs/adakites (related to slab melting) (Martin 1986, Defant & Drummond 1990). The Archean mantle wedge was affected more by melts than by fluids, resulting in HFSE being carried more efficiently from the slab to the arc magmas compared with the modern situation, with potential rheological effects and different subduction dynamics. With extensive mid-ocean ridge melting, basalts dominated the slab composition, with comparatively fewer terrigenous sediments (Rey & Coltice 2008). As LILE in modern arcs are primarily carried by these sediments, Archean arc basalts were less LILE enriched. If they were also richer in HFSE owing to a less depleted mantle, perhaps Archean arc rocks had a less pronounced "arc signature" compared with their modern equivalents. Therefore, the changing appearance of subduction depended on the intricate coevolution of various geodynamical, geological, and surface processes.

SUMMARY POINTS

1. A hotter Archean mantle affected plate buoyancy and strength, which, in turn, affected subduction dynamics.
2. Fitting Urey ratio estimates and thermal evolution data is easier if Archean plate-tectonic vigor was controlled by plate dynamics and not by mantle convective energy dissipation.
3. Some igneous rocks, notably high-pressure TTGs, are difficult to explain without burial of surface material to large depth and therefore carry a convincing geochemical arc signature.
4. Combined geodynamical and geochemical evidence points toward a more episodic style of protosubduction and argue against flat subduction or long-lived modern-style subduction for the Archean.

FUTURE ISSUES

1. Rheology is one of the weakest links in the chain toward our understanding of early-Earth dynamics, and future constraints are essential for a better understanding of subduction dynamics.
2. A paucity of data for the early Earth necessitates the integration of constraints from a wide range of disciplines (such as geodynamical modeling and geochemical data from igneous processes). Links between solid Earth and surface processes are beginning to emerge and might provide valuable additional constraints and insight.

DISCLOSURE STATEMENT

The authors are not aware of any affiliations, memberships, funding, or financial holdings that might be perceived as affecting the objectivity of this review.

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