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Growth of continental crust in intra-oceanic and continental-margin arc systems: Analogs for Archean systems

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Abstract Earth's continental crust has grown and been recycled throughout geologic history along convergent plate margins. The main locus of continental crustal growth is in intra-oceanic and continental-margin arc systems in Archean time. In arc systems, oceanic lithosphere is subducted to the deeper mantle, and together with its overlying sedimentary sequence is in some cases off-scraped to form accretionary prisms. Fluids are released from the subducting slab to chemically react with the mantle wedge, forming mafic-ultramafic metasomatites, whose partial melting generates mafic melts that rise up to form arcs. In intra-oceanic arcs, they produce dominantly basaltic lavas, with a mid-crust that includes variably-developed vertically-walled intermediate plutons and higher-level dikes and sills. In continental-margin arcs, different petrogenetic processes cause assimilation and fractionation of basaltic magmas, partial melting/reworking of juvenile basaltic rocks, and mixing of mantle- and crust-derived melts, so they produce andesitic calc-alkaline melts but still have a mid-crust dominated by vertically-walled felsic plutons, which form 3-D dome-and-basin structures, akin to those in some Archean terranes such as parts of the Pilbara and Zimbabwe cratons. Notably, the continental crust of Archean times is dominated by tonalite-trondhjemite-granodiorite (TTG) plutons, similar to that of the mid-crust of these arc systems, suggesting that early continental crust may have formed largely by the amalgamation of multiple arc systems. The patterns of magmatism, in terms of petrogenesis, rock types, duration of magmatic and accretionary events, and the spatial scales of deformation and magmatism have remained essentially the same throughout geological history, demonstrating that plate tectonic processes characterized by subduction and arc magmatism have been in operation at least as long as recorded by the preserved geologic record, since the Eoarchean. However, the early Earth was dominated by accretionary orogens and oceanic arcs, that gradually grew thicker through multiple accretion events to form early continental-margin arcs by 3.5–3.2 Ga, and accretionary orogens. Slab melting and warmer metamorphism was more common in Archean arc systems due to higher mantle temperatures. These early arcs were further amalgamated into large emergent continents by ~3.2–3.0 Ga, allowing large-scale processes such as lithospheric rifting and continental collisions, and the start of the supercontinent cycle. Further work should apply the null hypothesis, that plate tectonics explains the geologic record, to test for differences in the style of plate tectonics and magmatism through time, based on the fundamental difference in planetary heat production and the evolution of rotational dynamics of the Earth-Sun-Moon system.

Keywords Oceanic arc, Continental arc, Accretionary orogen, Archean, Crustal growth

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1. Introduction

Earth is postulated to have evolved from an early syn-post accretionary magma ocean stage some 4.5 billion years ago (Solomatov, 2015), through massive disruption by the Moon-forming impact of Theia and the last great impact of the planetesimal Moneta by 4.36 Ga (Genda et al., 2017; Benner et al., 2019), to having a surface with evolved felsic crust, plate tectonics, life, water, and a climate-ocean system that has been relatively stable for billions of years (Sleep, 2000; Harrison, 2020). How did our planet grow this crust, and obtain a stable balance between crustal and mantle systems, temperature, water, and life, while the planet's interior cooled, orbital dynamics caused day length to increase, solar luminosity changed, and the solid Earth-atmosphere-ocean system evolved (e.g., Kusky and Vanyo, 1991; Korenaga, 2016, 2017, 2018, 2021a, 2021b; Klatt et al., 2021; Hoffmeister et al., 2022)?

The origin of the continental crust, mostly in the Precambrian, is an issue of great debate, and one of the fundamental outstanding problems in the Earth Sciences (National Academies, 2020). Most data suggests that continental crust grew from subduction-related accretion of oceanic material and extraction of mafic melts from the mantle in oceanic and continental-margin arc systems (see reviews in Rudnick, 1995; Rudnick and Gao, 2003; Gazel et al., 2015; Kusky et al., 2018, 2021; Windley et al., 2021; Zheng, 2021; Sotiriou et al., 2022), with modification of the primitive melts by subsequent MASH (melting, assimilation, storage, homogenization) processes (Ducea et al., 2015a, 2021a, 2021b) and possible delamination of dense arc crustal roots (Bird, 1978; Saleeby et al., 2003). These processes produced felsic magmas in continental arcs that approach the average composition of the continental crust (Rudnick and Gao, 2003). This involves crustal growth at three types of magmatic arc: (1) oceanic arcs, also termed as island arcs or intra-oceanic arcs, which are dominated by basaltic rocks above oceanic-oceanic subduction zones; (2) continental-margin arcs, which are characterized by the occurrence of andesitic rocks above oceanic-continental subduction zones; (3) oceanic or continental-margin accretionary orogens, which are produced by accretion of oceanic arcs to other oceanic arcs or continental-margins. Together with multiple accretion and subduction events, these processes gradually built continents from the oceans, with continents becoming large enough and widely emergent sometime around 3.2–3.0 Ga (Windley, 1995; Windley et al., 2021; Wang X L et al., 2022), starting the supercontinent cycle. Other ideas are that arc magmatism is not as important as many models suggest, and that magmas generated during slab melting, collision, and slab breakoff may be more important for generating the bulk of continental crust (Defant and Drummond, 1990; Gazel et al., 2015; Hildebrand et al., 2018).

Models for the formation of the earliest preserved continents include those above, plume-related plateaus forming thick early continents (e.g., Van Kranendonk et al., 2015) or modified plate tectonic models, where oceanic crust in early times was exceptionally thick because of the higher degrees of mantle melting at higher temperatures (Sleep and Windley, 1982; Kusky, 1998), and was imbricated in subduction stacks causing fluid-fluxed partial melting at depth, generating felsic melts that formed the tonalite-trondhjemite-granodiorite (TTG) suite (de Wit et al., 1992; de Wit and Hart, 1993).

Zheng (2021) further notes that additional thickening and partial melting of thrust-stacked oceanic crust at convergent margins could contribute to the formation of continental crust through TTG magmatism. The high mantle temperature in the Archean would potentially form oceanic crust as thick as 30–40 km, which if thickened during collision could reach thicknesses of 60–70 km, similar to the continental crust along continental collision zones such as the Kohistan of Pakistan (Jagoutz et al., 2007; Dhuime et al., 2009), Himalaya (Dewey et al., 1988; Yin and Harrison, 2000), and Gangdese in southern Tibet (e.g., Zhu et al., 2019; Xu et al., 2019). It is possible that the collisional thickening of oceanic crust would be much more common in the Archean than in the Phanerozoic, resulting in the partial melting of the thrust-stacked wedges, and the formation of TTGs, and younger melts with different compositions after thinning of the lithospheric mantle along convergent plate margins (Li et al., 2021; Wang X L et al., 2022).

Other sites of new additions to the continental crust include plume-related large igneous provinces, which are associated with large outpourings of mantle-derived magmas, influencing other planetary systems and life on Earth (Ernst et al., 2021), but these lavas are overwhelmingly basaltic and do not match the bulk composition of the continental crust. Still other models suggest that the continental crust formed by some other plume-type processes on early Earth that was in a stagnant lid mode, by mechanisms that are unknown on the present Earth (Kröner, 1984; Johnson et al., 2014; Bédard, 2018). In this paper, we review the data and models for crustal growth by subduction-related arc magmatism through time, for which there are geological examples on Earth, using predominantly modern examples and moderate comparison with Archean terranes.

2. Growth of continental crust in Intra-Oceanic Arc Systems (IOAS)

2.1 Components of intra-oceanic arc systems

Studies of Phanerozoic accretionary-collisional orogens have led to the suggestion that they are modern analogues to Archean orogens (Kusky, 1989; Şengör and Natal'in, 1996;

Kusky and Polat, 1999; Cawood et al., 2009; Kusky et al., 2018, 2021a; Windley et al., 2021) (Figure 1). Where an oceanic plate subducts beneath another, an oceanic arc (equivalent to island arc or intra-oceanic arc) forms, generally with or without small accretionary prisms, whereas when the oceanic plate subducts beneath a continent or microcontinent, a continental-margin arc forms, with or without small to large accretionary prisms, and with or without subduction erosion (von Huene and Scholl, 1991; Straub et al., 2020). When subduction brings arcs together, they collide, resulting in the growth of an accretionary or "Turkic-type" orogen (Şengör et al., 1993, 2021, 2022; Şengör, 1996; Cawood et al., 2009; Xiao et al., 2020). There is a continuum of crustal thicknesses and compositional progression from a dominantly mafic crust in thin oceanic arcs to more felsic and thicker crust in arcs built in continental-margin accretionary orogens. Multiple processes are involved in thickening and differentiating arc crust from oceanic to continental-margin arcs, including magmatic inflation by multiple generations of intrusions, partial melting and differentiation, thickening by shortening (thrusting, bulk strain), under-thrusting during accretion of exotic terranes and collisions, and perhaps in some cases, loss of dense arc cumulates by delamination. The merits of these processes in different cases are discussed below.

The presence or absence of accretionary prisms (and filled or unfilled trenches) is largely dependent on the proximity to sources of sediment such as rivers or glaciers (and thus to the ambient climate), and therefore most intra-oceanic arc systems have under-filled trenches, and narrow to non-existent accretionary wedges (Scholl and von Huene, 2007; Zheng, 2021). This is likely to have been true in the early Archean as well, when there were no large continental erosional sources (Windley et al., 2021), but frequent arc/arc collisions that would have provided abundant immature detritus, and led to crustal thickening during the collisions. Intra-oceanic arc systems (IOAS, or OAS for short) were the most important setting of crustal growth in the Phanerozoic and Proterozoic (Şengör et al., 1993, 2021; Stern, 2010), and most-likely were also in the Archean (Kusky and Polat, 1999). To paraphrase Stern (2010) "*IOASs are the fundamental building blocks that are assembled over geological time into orogens*, *cratons*, *and continents*, *and continental crust is basically a mosaic of nests of island arcs*, *welded and melded together*."

Intra-oceanic arc systems (Figure 2) are mostly submergent with only chains of small islands sticking above sea level. In the present plate mosaic, most intra-oceanic arcs are located in the western Pacific (e.g., Tonga, Izu-Bonin-Mariana, or IBM; Figure 1), where they are subducting old dense oceanic lithosphere with significant slab rollback, so these arcs tend to be under extension and have narrow fore-arc accretionary wedges, if any (Figure 2). They are divisible into four main zones, trench, fore-arc, magmatic arc, and backarc, and are typically about 200–400 km wide, 15–35 km thick, and hundreds to thousands of km long (Stern, 2010). The magmatic axis of any arc is located generally at a minimum of 150 km from the trench, since geometrically it simply takes that distance to take a thick, flat-lying plate of oceanic lithosphere with a finite strength and flexural rigidity, and bend it down into the asthenosphere to a depth of ~110 km where it can release enough fluids to cause partial melting in the mantle wedge (e.g., Zheng et al., 2016, 2022; Zheng, 2021) to generate the mafic arc magmas (Figure 2), whereas any coeval intermediate-felsic magmatism may be sourced in the crust.

Many intra-oceanic arcs preserve primitive arc magmatic systems or fore-arc ophiolites, and show high degrees of melting in the first 5–10 Myr of their history (Reagan et al., 2010, 2013, 2017; Maunder et al., 2020), related to subduction initiation. A standard subduction initiation sequence has been established, as a magmatic sequence changing from typical mid-ocean ridge basalts (MORB) through boninite to island arc tholeiite, typically over a short time of a few to tens of millions of years (Reagan et al., 2010, 2013, 2017; Stern et al., 2012; Stern and Gerya, 2018; Maunder et al., 2020). Subduction initiation sequences have now been recognized in many forearc ophiolites in orogens of all ages, from active (Bonin arc, Reagan et al., 2017), to Phanerozoic (Himalaya; Dai et al., 2021), to Archean (Ning et al., 2020; Zhong et al., 2021).

In more mature intra-oceanic arcs (Figure 1) the magmatism is concentrated in the magmatic axis. Extension in mature arcs in many cases leads to the formation of backarc basins with sea-floor spreading (Figure 2). The upper volcanic section of most IOAS consists of mafic volcanic rocks (mostly basalt), with a more felsic (tonalitic) middle crust (Figure 3) typically including a series of smaller steep-walled plutonic complexes. Fore-arc mantle is typically composed of serpentinized harzburgite-pyroxenite beneath the arc, and lherzolitic-harzburgitic beneath the back arc (Stern, 2010).

Since most intra-oceanic arcs today are not associated with extensive accretionary wedges, and initiate above retreating trenches soon after subduction initiation, they are the site of generation of fore-arc crust consisting of basalt, boninite-picrite, diabase, gabbro, and serpentinized harzburgite (Stern, 2002, 2004; Whattam and Stern, 2011; Dai et al., 2021; Ning et al., 2020). This sequence precedes the magmatism associated with the main island arc tholeiite (IAT) trends in the arc. It is thought that most ophiolites in orogenic belts initially form in forearc settings, since crust generated there is easily obducted during collisional orogenesis (Shervais, 2001; Stern et al., 2012; Stern and Gerya, 2018; Ning et al., 2020). Subduction initiation magmatism (MORB-boninite-picrite-IAT) is short-lived (Maunder et al., 2020), and by about 5–10 Myr after subduction initiation forearcs generally become cold and amagmatic, but with

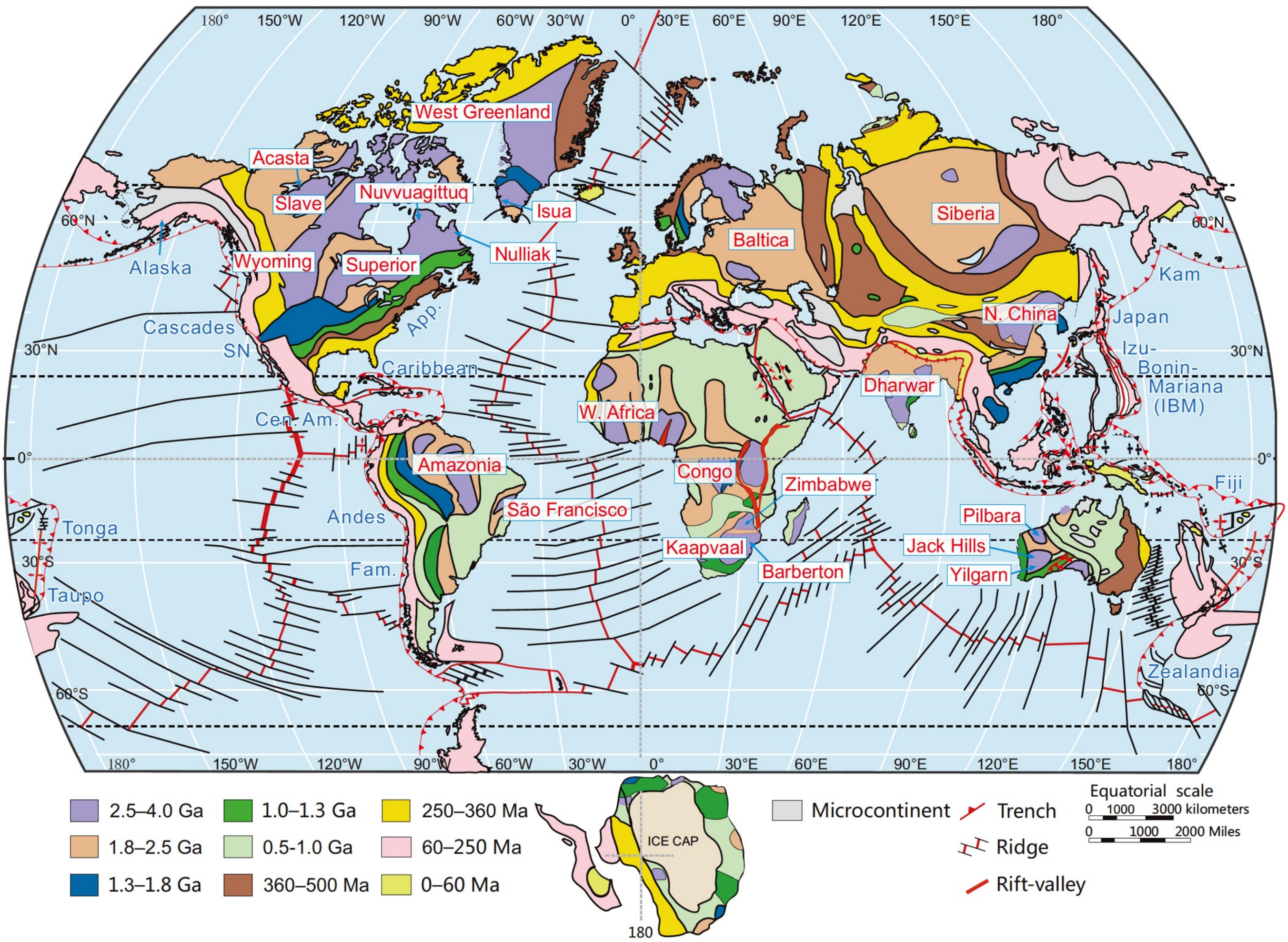

Figure 1 Plate tectonic map of the world (modified from Windley et al., 2021) showing main cratons and Archean areas labeled in red and intra-oceanic and continental-margin arcs labeled in blue. The main Archean cratons are labeled in red, while the extant and Phanerozoic arcs are labeled in blue. Abbreviations as follows: App., Appalachians, including the Ordovician Taconic Orogen; Cen. Am, Central American arc; Fam, Famatinian (and Pampean) peri-Gondwanan arcs. Map compiled from numerous sources, including: UNESCO, 1976; Liou et al., 1990; Goodwin, 1996; Condie, 1997; Windley, 1995; Maruyama et al., 1996; Kusky and Polat, 1999; Maruyama et al., 2007; Utsunomiya et al., 2007; Sawada et al., 2018.

deformation concentrated in the outer forearc, where serpentinized mantle peridotite may be exposed in the outer trench wall. The inner forearc may host a forearc basin (Figure 2), with sediments (0 to a few km thick) derived from the arc, or erosion of the accretionary wedge, if one has developed and is uplifted above sea-level. Volcanism along the arc axis is dominated by vesicular and porphyritic fractionated basalts, along with lesser high-Mg andesites and rhyodacites and shoshonites, which are underlain by a tonalitic plutonic middle crust. Lavas along the arc axis may be tholeiitic, high-alumina or calc-alkaline, but backarc basin basalts are dominantly tholeiitic (Arculus, 2003). In some cases, felsic volcanics also erupt in the arc, typically forming large caldera complexes (e.g., Finney et al., 2008). Many active IOAS have backarc basins formed by sea-floor spreading in which crustal sequences and structures are similar to those of mid-ocean ridges (Grevemeyer et al., 2021). This may be initiated by splitting the arc (leaving a remnant arc), or be localized in the initial fore-arc, or behind the active magmatic front. Backarc basins, forearc basins, and the flanks of emergent volcanoes may be sites of sedimentation, but unless the arc is near a sediment source such as a continent or a colliding arc, the trench is typically underfilled or empty with subduction erosion at the trench (e.g., Straub et al., 2020). Backarcs typically have the thickest volcanogenic, arc-derived sedimentary sequence, including volcanic ash near the arc, and may be up to ~1 km thick (Dorobek, 2008; Balázs et al., 2016). Since these basins are above the CCCD (calcium carbonate compensation depth), they may also contain carbonates or other biogenic sediments, if climatic and biologic production conditions are permissive (Balázs et al., 2016).

There is often a strong across-arc zonation in the style of arc magmatism (Figure 3). Forearc magmatism is typically the first to form and includes boninitic melts as described above, whereas the magmatic axis is dominated by tholeiitic-

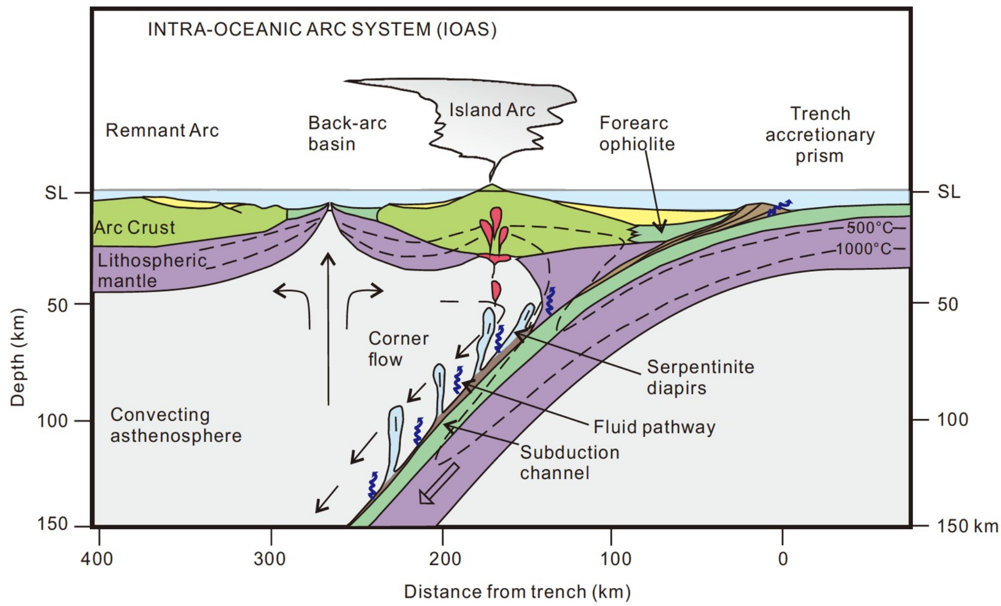

Figure 2 Simplified geology of intra-oceanic arc systems. The term intra-oceanic arc system refers to the entire trench, accretionary prism, forearc, arc, back-arc, crust and mantle components, whereas the term oceanic arc generally refers to only the magmatic portion of the system, as shown in Figure 3. Modified after Stern (2010).

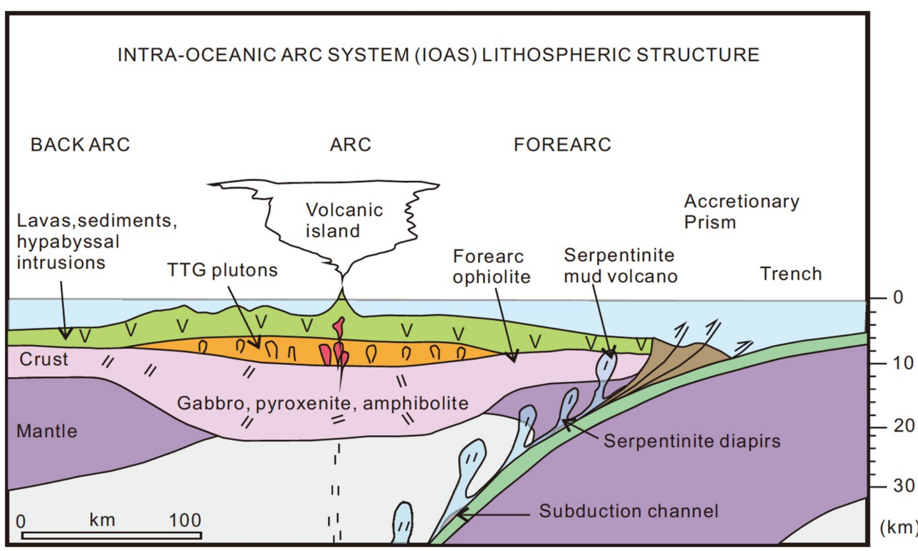

Figure 3 Simple cross-section of an intra-oceanic arc, based on seismic studies of the Izu-Bonin-Mariana (IBM, Figure 1) arc (Suyehiro et al., 1996). Note the mid-crustal layer of felsic intrusions consisting mostly of tonalite to granodiorite. Modified from Stern (2010), after DeBari et al. (1999) and various other sources. The magmatic arc is only part of the larger intra-oceanic arc system, as shown in Figure 2.

and high-Al basalts along with calc-alkaline andesites (Kelemen et al., 2003). Backarc basin basalts (BABB) are often dominated by arc-like calc-alkaline lavas in the early stage and MORB-type tholeiites in the late stage, with a range of lavas from MORB-like to arc-like but are more hydrous than MORB (Pearce and Stern, 2006). Compositionally arc magmas tend to be (but are not always) Si-, LILE (K, U, Sr, Pb) and H_2O-enriched, resulting from much higher degrees of melting than typical MORB- or ocean island basalts (OIB) basalts (up to 70% more melting). Trace elements that are not fluid-mobile (e.g., Nb, Zr, Ti, HREE) tend to be depleted. These elemental patterns reflect high degrees of melting in a water-rich environment beneath the arc (and above the subducting plate), which clearly distinguishes arc magmas from MORB and OIB lavas (Stern, 2002; Zheng et al., 2016; Zheng, 2019). This is also why arc magmas tend to be coarsely plagioclase- and olivine-porphyritic, and have high concentrations of LILE such as Rb, Sr, Ba, U, K, and Pb. The flat HREE patterns show that there is no garnet in the region where the melt last equilibrated with the peridotite, at depths of 15–80 km, although melting may have initially occurred much deeper, from 80 to 160 km (see Straub et al., 2014;

Jagoutz and Kelemen, 2015).

Other processes may affect and alter arc magmas once they rise into the crust. They may assimilate surrounding crust, fractionate, heavy mafic/ultramafic cumulates may settle or possibly even delaminate, and the arc may tend to become more felsic with time (e.g., Behn and Kelemen, 2006; Tatsumi et al., 2008). Early arc magmas tend to be boninites and tholeiites reflecting high-degrees of melting, and a subduction initiation sequence of MORB-boninite-IAT tholeiite has been identified in a number of nascent forearcs and obducted ophiolites (Reagan et al., 2010, 2013, 2017; Whattam and Stern, 2011; Dai et al., 2021; Ning et al., 2020). These early arc magmas may evolve into later calc-alkaline and shoshonitic lavas as the arc matures and the crust becomes thicker, while other variables such as the age, thickness, temperature and angle of the subducting plate, and upper plate stresses related to slab rollback, may also play a role in magma composition. For instance, Eocene flat subduction of an oceanic plateau beneath Fiji (Figure 1) caused hydrous partial melting of the slab, generating tonalitic melts now exposed in Viti Levu (Bonner et al., 2020). Subduction of very young hot oceanic crust, or episodes of ridge subduction (as in south Japan (Figure 1), and shown to be similar to Archean equivalents), have been suggested to produce adakitic magmas (Defant and Drummond, 1990; Kusky et al., 2004), but these may also be formed by sediment subduction and melting (Ribeiro et al., 2016; Zeng et al., 2016).

Samples of the mantle beneath backarc basins (Figure 2) (Ohara, 2006) show host lherzolite-harzburgite cut by dunite dikes and porous channels that form by the interaction of melt and pyroxene. In contrast, IOAS forearc mantle typically consists of extremely depleted harzburgite. Fore-arc spinels (chromite), if unaltered, may preserve very significant petrogenetic information (Dick and Bullen, 1984; Kusky and Glass, 2007; Huang et al., 2021), and have high Cr-number (Cr+Al)>0.4, which can be a great aid in interpreting the tectonic setting of ancient forearc peridotites (e.g., Kusky et al., 2007, 2021b; Huang et al., 2017; Ning et al., 2020). Understanding the nature of the mantle beneath the arc axis is hindered by the thickness of the crust in this region, but samples can be obtained from xenoliths (Arai and Ishimaru (2007) and are typically spinel peridotites without garnet that show high degrees of mantle metasomatism, forming secondary orthopyroxene replacing olivine. Seismic studies (Takahashi et al., 2008) suggest that significant amounts of pyroxenite and amphibolite may also be present.

Another component of oceanic arc and supra-subduction ophiolite development and evolution through time is the formation of layered calcic anorthosite-gabbro-ultramafic complexes, which form in the magma chamber of modern oceanic arcs as evidenced from exposed sections (e.g., Davidson and Arculus, 2005; Jagoutz and Kelemen, 2015; Sotiriou and Polat, 2020), cognate xenoliths (Arculus and Wills, 1980; Conrad and Kay, 1984) and seismic sections (Kiddle et al., 2010; Sotiriou and Polat, 2020). The lower magma chamber or "root' of the Cretaceous (greenschist-low amphibolite facies) chromite-layered Kohistan island-arc in Pakistan (Khan et al., 1989; Jagoutz et al., 2007; Dhuime et al., 2009; Petterson, 2010; Bilqees et al., 2016) was separated and partially subducted farther down the Himalayan Yarlung-Tsangpo subduction zone, metamorphosed, and exhumed as the Tora Tigga (amphibolite facies-Jan and Tahirkheli, 1990) and Jijal (granulite and eclogite facies) complexes (Ringuette et al., 1999). All these rocks, their lithologies, geochemistry and structures serve as ideal modern analogues for comparable Archean layered calcic anorthosite-gabbro-ultramafic complexes (mostly hornblende-bearing) that typically occur as remnant thrust-bound strips in amphibolite-granulite TTG orthogneisses in, for example, West Greenland and South India. The type examples are the 2973±28 Ma Fiskenaesset Complex in Greenland (Myers, 1985; Polat et al., 2009, 2010, 2018) - for a Paleozoic analogue in New Zealand, see Gibson and Ireland (1999); and the 2588±4 Ma Sittampundi Complex in India (Subramaniam, 1956; Sajeev et al., 2009; Mohan et al., 2013). An important confirmatory link between the modern and Archean arc-generated complexes is provided by many studies of experimental petrology (e.g., Claeson and Meurer, 2004; Müntener and Ulmer, 2006) that have demonstrated that the fractional crystallization and cumulate history of hydrous basaltic arc-type magmas give rise to the amphibole-bearing, cumulate, high-Fe chromite-layered, high-An plagioclase anorthosite-gabbro-ultramafic complexes.

2.2 The crustal composition paradox

One of the conundrums of plate tectonics is that most continental crust is *inferred* to be extracted from the mantle in juvenile oceanic arc systems, which have a dominantly mafic composition, but the average composition of the continental crust is estimated to be andesitic (Rudnick and Gao, 2003; Polat, 2012). To account for this discrepancy, it is widely *assumed* that the lower ultramafic arc cumulates delaminate and sink back to the mantle (Rudnick, 1995), despite the fact that no foundered arc roots or "drips" have ever been conclusively imaged seismically, and it is difficult to reach eclogitic conditions of a sufficiently thick mafic layer in the crust necessary for the density increase to promote the hypothesized foundering (Hildebrand and Whalen, 2017; Hildebrand et al., 2018; Wang and Kusky, 2019). In any case, when arcs collide, either with each other or with continental fragments, they are structurally thickened by thrusting and internal strain, and undergo another episode of anatexis to form the present-day crust (e.g., Kelemen, 1995; Kusky and Polat, 1999; Polat, 2012), with contributions from partially melting the accreted crust and from the mantle, and from slab failure melts generated after the subducting slab breaks off and all arc magmatism has ceased (Hildebrand and Whalen, 2014a, 2014b; Fu et al., 2019). However, despite this late-stage differentiation through anatexis, most of the continental crust preserves original trace element signatures indicative of formation in primitive oceanic arc systems (Hawkesworth and Kemp, 2006).

For Archean examples, Nagel et al. (2012) and Adam et al. (2012) and Sotiriou et al. (2022) have all shown that the geochemical characteristics of Eoarchean TTG's are consistent with melting of subduction-derived oceanic arc crust, including original tholeiitic basalts and boninites from deep arc crust. Nagel et al. (2012) demonstrated that Eoarchean TTG's from the 3.8 to 3.7 Ga Itsaq Complex in SW Greenland (Figure 1) have inherited their REE and HFSE trace element characteristics from spatially associated arc lavas. Adam et al. (2012) showed that boninites from the ~3.96 Ga Nuvvaugittuq greenstone belt of Quebec, when experimentally melted in the lab, produce tonalitic melts similar to the enclosing 3.66 Ga TTG rocks. Thus, in the Archean as primitive arcs collided with each other and became thickened by thrusting (e.g., Kusky and Polat, 1999; Windley et al, 2021), the arc-related basalts were partially melted to produce the TTG's, perhaps with a smaller component derived from slab melting (Polat, 2012). These arc-related magmas were also intruded by late- to post-deformation slab failure magmas, and thus were gradually able to build a progressively more differentiated continental crust (Kusky et al., 2021). This crust includes tonalitic mid-crust associated with mafic-locally intermediate volcanic rocks, and individual accreted arc segments separated by sutures defined by thin sedimentary sequences with slices of dismembered fore-arc crust forming Archean ophiolites (de Wit and Ashwal, 1997; Kusky, 2004). The combination of structural thickening and multiple episodes of plutonism typically results in the formation of dome-and-keel structures at mid-crustal depths (Kusky et al., 2021). When the size of the plates increased sufficiently through mantle cooling and the associated increasing size of convection cells, continents began to emerge towards the end of the Archean, starting the phase of modern-style Wilson Cycles and supercontinents (Windley et al., 2021), thus completing the first two steps of the orogen-craton-orogen cycle (Kusky et al., 2007, 2018).

3. Crustal growth in continental-margin arc systems

3.1 Transition from intra-oceanic to continental-margin style arcs

As immature oceanic arcs grow, they typically experience slab rollback causing migration of the magmatic front through the forearc complex of accreted ophiolites and ocean plate stratigraphic (OPS) units (Kusky, 1989, Kusky et al., 2013; Komiya et al., 1999). In these cases, thin thrust-imbricated packages of mafic-ultramafic rocks from the oceanic crust, along with oceanic sediments (chert, shales, etc.) are stacked forming thrust-imbricated wedges, mélanges, and slices of oceanic crust (e.g., Kusky et al., 2020; Zheng, 2021). When the locus of arc magmatism moves through these structurally complex wedges, the magmas are typically intruded as sheets parallel to structures, which get deformed into folded or domal structures, as mapped in the Cretaceous Coast Ranges accretionary orogen batholith (e.g., Brown and McClelland, 2000), the Tertiary Hidaka belt of Japan (Komatsu et al., 1989), and the Mesoarchean of the Eastern Pilbara craton (Kusky et al., 2021). Later, when these already immensely complex arc systems in intra-oceanic accretionary orogens collide with each other, or accreted exotic fragments, they gradually build up a further-differentiated crustal mass by collision-induced thickening and partial melting of the crustal materials together with mantle sourced contributions (e.g., Osanai et al., 1991; Fu et al., 2019), slab failure magmatism, additional deformation, and eventually forming elongate archipelagoes (Kusky, 1989; Kusky and Polat, 1999; Polat and Kerrich, 2000; Hildebrand et al., 2018; Windley et al., 2021).

During each accretionary event of an arc-to-arc, or an arc-to-young continental-margin accretionary orogen, a subduction zone and slab must break off (a.k.a. fail) to produce a new pulse of slab failure magmas into the accretionary orogen, causing further significant differentiation towards more Si-rich crust with time (Hildebrand et al., 2018). Some numerical models suggest that slab breakoff events at shallow asthenospheric depths, with multiple out-stepping of new subduction zones resulting in growth of orogenic wedges at lithospheric depths may have been more likely in the Archean (e.g., Moyen and van Hunen, 2012; Foley, 2018), but comparable to the modern range of variation in slab behavior (Penniston-Dorland et al., 2015; Agrusta et al., 2018; Korenaga, 2011, 2018). In contrast, other numerical models suggest that higher Archean mantle temperatures actually favored negative buoyancy and deep subduction (Agrusta et al., 2018; Weller et al., 2019). In either case, these collided arcs with intervening accretionary prisms and ophiolitic slices were likely partly subaerial, forming the first protocontinents. These in turn would be loci of plate margins, because where arcs collide, the arc-forming subduction zone fails by slab breakoff and reversal of subduction polarity is typically followed by a rapid burst of slab failure magmatic intrusions (Hildebrand et al., 2018), which then leads to a new arc being built along the collision-modified margin of the collided arcs (Dewey, 1977; Draper et al., 1996; Chemenda et al., 1997; Pubellier et al., 1999; Teng et al., 2000; Clift et al., 2003; Vignaroli et al., 2008; Kusky, 2011; Kusky et al., 1987, 2016, 2018, 2020). These archipelagoes thus would represent the first emergence of continental-margin-style arcs on the planet (Windley et al., 2021).

We note that there is a continuum in crustal thicknesses, structural styles, and processes (e.g., Zheng et al., 2015) from oceanic arcs (15–35 km thick) to continental-margin arcs (~30–50 km thick), to the over-thickened continental margins (up to 70 km thick) in accretionary Andean-style orogens and continental collision zones (Zheng, 2021). While the batholith belts of the Andes arc are typically used as models for continental-margin arc systems (Hamilton, 1969; Pitcher, 1993; DeCelles et al., 2009; Ducea et al., 2015a), these are actually anomalously thick systems (Figure 4), whereas the majority of continental-margin arc systems are considerably thinner and do not become over-thickened until they are involved in arc-continent collision (Moores et al., 2002, 2006; Johnston, 2008; Hildebrand and Whalen, 2014a, 2014b). More-typical continental-margin arcs (Figure 1) form on crust with average or below-average thickness and deposit volcanic rocks in subsiding depressions, such as the Central American arc (MacKenzie et al., 2008), Cascades (Mooney and Weaver, 1989), Alaskan Peninsula (Fliedner and Klemperer, 2000; Wilson et al., 2015), Kamchatka (Levin et al., 2002a, 2002b), Tonga-Kermadec (Scholl et al., 1985), and in the Taupo volcanic zone, New Zealand (Harrison and White, 2006). Still, the processes of thickening of arcs from the oceanic to the continental-margin end members is debated, and variable in different cases, with thickening by magmatic inflation, thrusting and shortening during collisions, and by underplating of other arcs or continents during collisions (e.g., Brown and McClelland, 2000; Paterson et al., 2011; Hildebrand, 2013; Ducea et al., 2015b). In some cases, previously subducted material may also rise along the subduction channel, or in serpentinite diapirs, to underplate or "relaminate" the base of the overriding plate (Figure 5), causing additional crustal thickening, but in this case, with felsic material (Hacker et al., 2011; Keleman and Behn, 2016).

3.2 Processes and variations in continental-margin arcs

Continental-margin arcs have many broad-scale similarities to IOAS's, but also some fundamental differences (Figure 4), and are the present-day locus of the largest amount of intermediate composition magmatism on Earth. They have crustal thicknesses that vary from ~30 to 70 km, and compositions between andesite and dacite representing an average between new mafic material added from the mantle, and products of upper crustal differentiation, and/or episodes of slab failure magmatism from successive arc accretion events. Continental-margin accretionary orogens with their arcs tend to have long life-spans (including multiple separate events), from tens to hundreds of millions of years (compared to ~50–60 Myr for most IOAS; Paterson et al., 2011) and they witness frequent lateral migration of the axis of magmatism, periods of extension or contraction, periods of quiescence and magmatic flare ups, times of lithospheric thickening and postulated times of foundering of their deep roots (Ducea et al., 2015a, 2015b, 2021a, 2021b). Importantly, most continental-margin accretionary orogens do not form during a single subduction event, but represent a collage of accreted arcs formed above a succession of subducting slabs over this long interval (Figure 4).

Many of the variations in continental-margin accretionary orogens are related to the age, angle, and temperature of the subducting slab, relative velocities between over-riding and under-riding plates, and subduction of anomalous features such as seamounts, or spreading ridges (Kusky et al., 1997, 2003; Bradley et al., 2003; Horton et al., 2022). Below we describe some of the major features of continental-margin arcs, so that they can be recognized and differentiated from juvenile oceanic arcs in the geologic record, and consider their role in the growth and evolution of the continents through plate tectonics. Continental-margin arcs are similar to IOAS in that their mafic magmas are fundamentally derived from dehydration of a subducting slab driving partial melting of the mantle wedge, but they also differ from IOAS in some fundamental ways (Jagoutz and Kelemen, 2015). First, they are compositionally richer in silica reflecting the operation of additional petrogenetic processes including crustal melting and assimilation, have longer life spans, and typically experience greater variations in tectonic style than IOAS, with episodes of upper plate extension (back- or fore-arc basin formation), contraction (fold-thrust belt formation, crustal thickening), strike slip faulting, fore-arc accretion, and fore-arc erosion (Figure 4). Because of their complex evolution including successive accretion of exotic arcs, migrating magmatic fronts, periods of extension and contraction, continental-margin arcs tend to have very wide areas (hundreds of km) of thickened crust and regions influenced by processes in the continental-margin accretionary orogens containing the various accreted arcs (Figures 4 and 5), and underlying mantle wedge and/or subducting plates (Lee et al., 2007; Hildebrand, 2013; Ducea et al., 2015a).

Over-thickened continental-margin arc magmas are generally rich in slab-derived volatiles (Figure 5) and are calc-alkaline with a large range of silica contents (45–75%), and have distinctive trace element patterns (Pearce and Peate, 1995; Marschall and Schumacher, 2012; Zellmer et al., 2015; Hildebrand et al., 2018). To generate such large volumes of intermediate-felsic magmatic rocks (for instance, the various magmatic and tectonic systems comprising the Coast Mountains Batholith system of Western North America have produced 10^6 km^3 of granitic magmas in 150

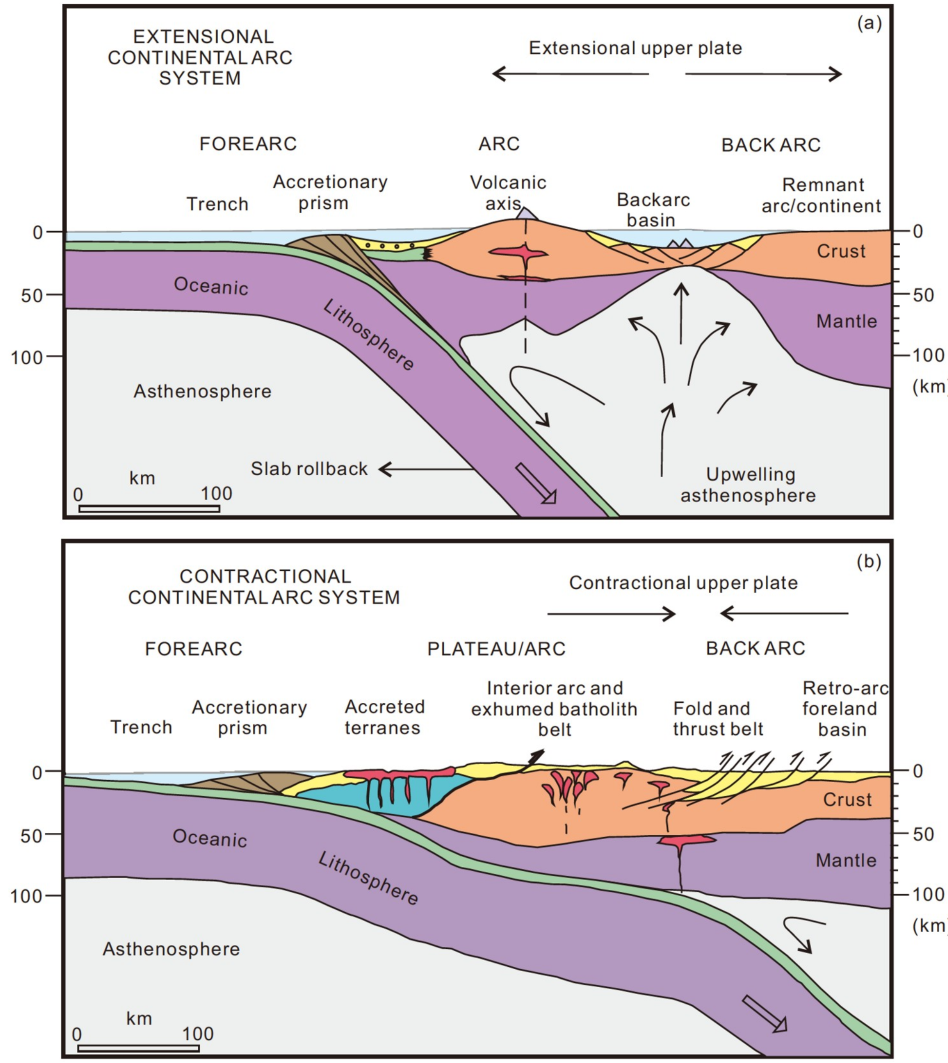

Figure 4 Sketch of two end-member continental-margins with accretionary orogens and arc systems, extensional and contractional, based in part on Ducea et al. (2015a), Pfiffner and Gonzalez (2013), and a combination of many examples from the present plate mosaic. Note the accreted terranes, simplified in this diagram, but which may include multiple terranes accreted from different-facing subduction systems to form continental-margin accretionary orogens.

Myr. Anderson, 1990; Saleeby, 1990), it is widely hypothesized that there must be residual olivine-rich cumulates or other residual masses at depth (Figure 5), but these are rarely exposed at the surface in young continental arcs creating a crustal composition paradox (Rudnick, 1995; Hildebrand et al., 2018). However, cumulates and residues such as garnet pyroxenites are exposed in some examples of deep arc crust roots (Figure 6) such as in the Talkeetna arc in Alaska (Hacker et al., 2011), the Kohistan of Pakistan (Jagoutz and Behn, 2013), the Famatinian arc in Sierra de Valle Fertil, Argentina (Otamendi et al., 2012), in the Southern Sierra Nevada, USA (Chapman, 2012), and in Fiordland New Zealand (De Paoli et al., 2009). In many cases it is thought that deep arc crustal roots have foundered, and because of their density, sunk back into the mantle (Saleeby et al., 2003; DeCelles et al., 2009; Ducea et al., 2021a, 2021b).

3.3 Crustal profiles through thickened oceanic and continental-margin arcs

Composite profiles of the crustal section of thick accreted and thickened oceanic and continental-margin arcs for Fiordland, New Zealand (Milan et al., 2017), Kohistan, Pakistan (Jagoutz and Kelemen, 2015), and the southern Sierra Nevada (Ducea et al., 2015a, 2021b), USA, are shown in Figure 6. These are characterized by an upper 2–4 km thick mafic-silicic volcanic/hypabyssal section, an intermediate 20–30 km thick batholithic section consisting of thousands of individual stocks, dikes, and sills of tonalite-granodiorite (± diorite), underlain by a deep root composed of mafic-ultramafic cumulates and residues from melting (Figure 5). The middle and upper sections of the crustal profile show dominantly vertical structures (Paterson and Fowler, 1993) (i.e., tonalitic to granodioritic domes intruding older se-

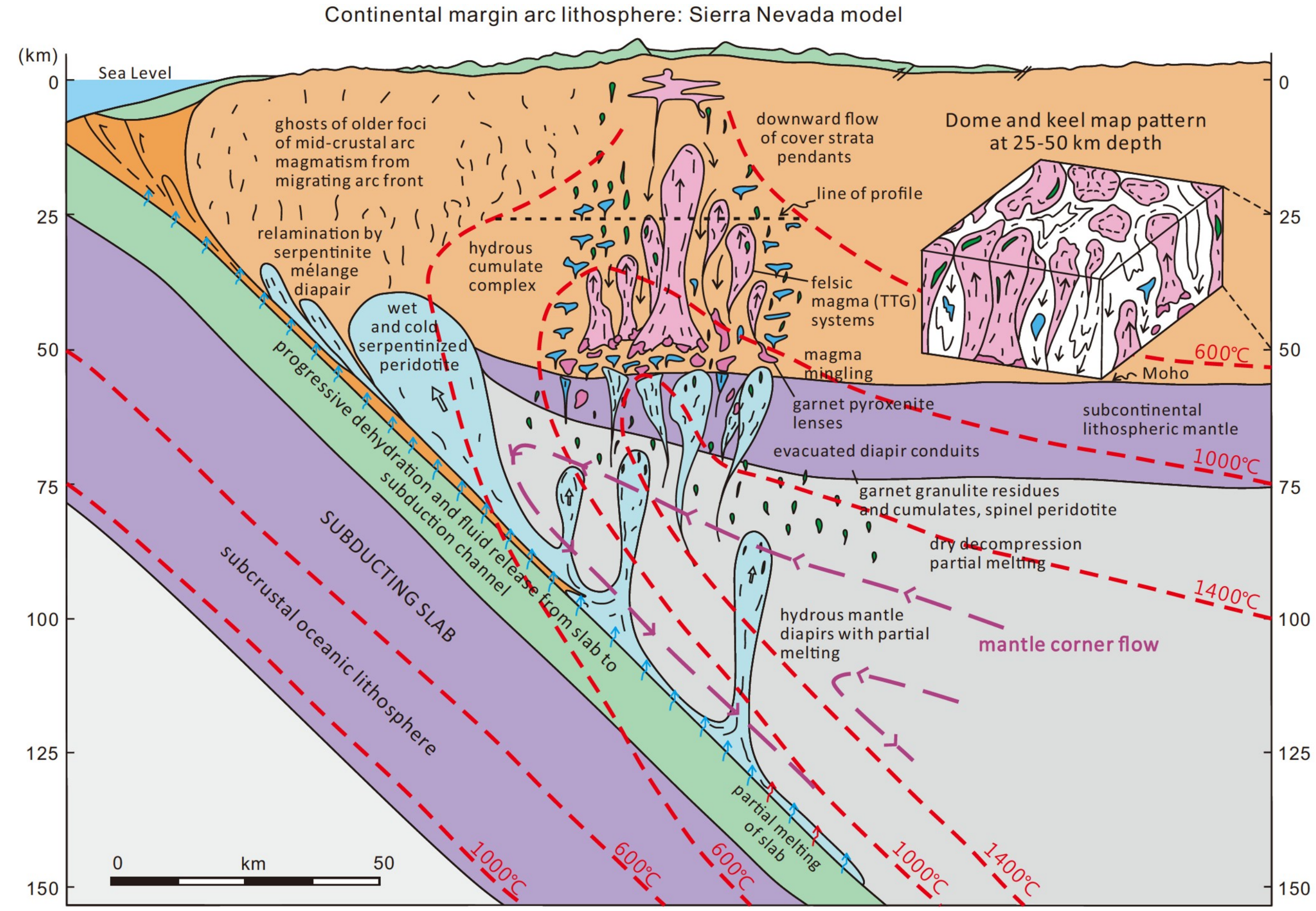

Figure 5 Model of the structural and petrological processes in a thickened continental margin arc and subduction system, based in part on the Sierra Nevada batholith (Figure 7). Model is modified slightly from Kusky et al. (2021a), developed from multiple previous studies, including most notably Saleeby et al. (2003) and Ducea et al. (2010, 2015a, 2015b), Marschall and Schumacher (2012) and Zellmer et al. (2015). Dehydration of the subducting slab and sedimentary/volcanic material in the subduction channel releases fluids which hydrate the mantle wedge and fluxes it with LILE, leading to partial melting. Serpentinite plus sediment and metabasalt diapirs likely rise from the subduction channel, bringing additional material from the slab and subducted sediments into the overlying mantle wedge, and may be the source of some arc magmas (e.g., adakites), The diapirs, plus entrained material can in some cases lead to subcrustal relamination, or incorporation of crustal xenocrysts in mantle melts (Kusky et al., 2021b). Mantle wedge corner flow continuously brings new mantle under hypersolidus flow to be metasomatized by the slab-derived fluids (e.g., Zheng, 2019). These processes gradually form a lower crust/upper mantle hydrous cumulate complex derived from the slab flux melting, with possible additions of slab or subduction channel melts under appropriate thermal conditions such as a warm slab-top geotherm. Dehydration-induced melting and remelting of this complex and even older residual material, together with slab and metasomatized wedge derived melts, may together produce large volumes of intermediate-felsic TTG magmas that rise in multiple steep-walled plutons, with the wall rocks sinking in between. This process leaves garnet pyroxenite (arclogite) residues along the Moho. The deformation conditions of the rising plutons is thought to be hot sub-solidus flow of the wall rock, with common migmatization, with some plutons rising as hypersolidus crystal liquid mushes leaving composite magmatic and tectonic foliations, depending on the depth of crystallization (generally between 15 or 20 to 50 km depth). Deformation is commonly assisted by grain boundary melts, and many chemical signatures in deep mafic granulites from arc crust roots (such as Fiordland, New Zealand (Ze on Figure 1) show a large melt flux during deformation and upward melt percolation (Daczko et al., 2016; Stuart et al., 2016, 2018). The 3-D block diagram (inset on the upper right of diagram) is drawn along a 25 km deep profile, highlighting typical dome-and-basin type patterns, such as those of the Eastern Pilbara, Zimbabwe, and some other Archean cratons (Kusky et al., 2021).

quences, forming dome-and-keel structures, with generally concordant contacts between plutons and pendants), whereas the deeper level with the cumulates and residues shows characteristic horizontal foliations, igneous layering, and deformation features formed under solidus to hot sub-solidus conditions (Figure 5). At the batholithic level, characteristic of that exposed in deeply eroded arcs (such as those in Archean cratons; Figure 5, inset) structures show a great preponderance of vertical pathways where plutons push their way upward, and the surrounding framework rocks show a return downward flow (Saleeby et al., 2003). Babeyko and Sobolev (2005) suggest that the entire lower crust beneath arcs may behave convectively with crustal overturn facilitating vertical transport of granitoid magmas upwards, and surrounding material downwards (Figure 5), forming the characteristic dome-and-keel structures of continental arcs and some Archean granite greenstone terranes (e.g., the famous domes of Pilbara; Kusky et al., 2021). The most common rocks at mid-to deep crustal levels (20–35 km) are mantle-derived melts including amphibole gabbros, horn-

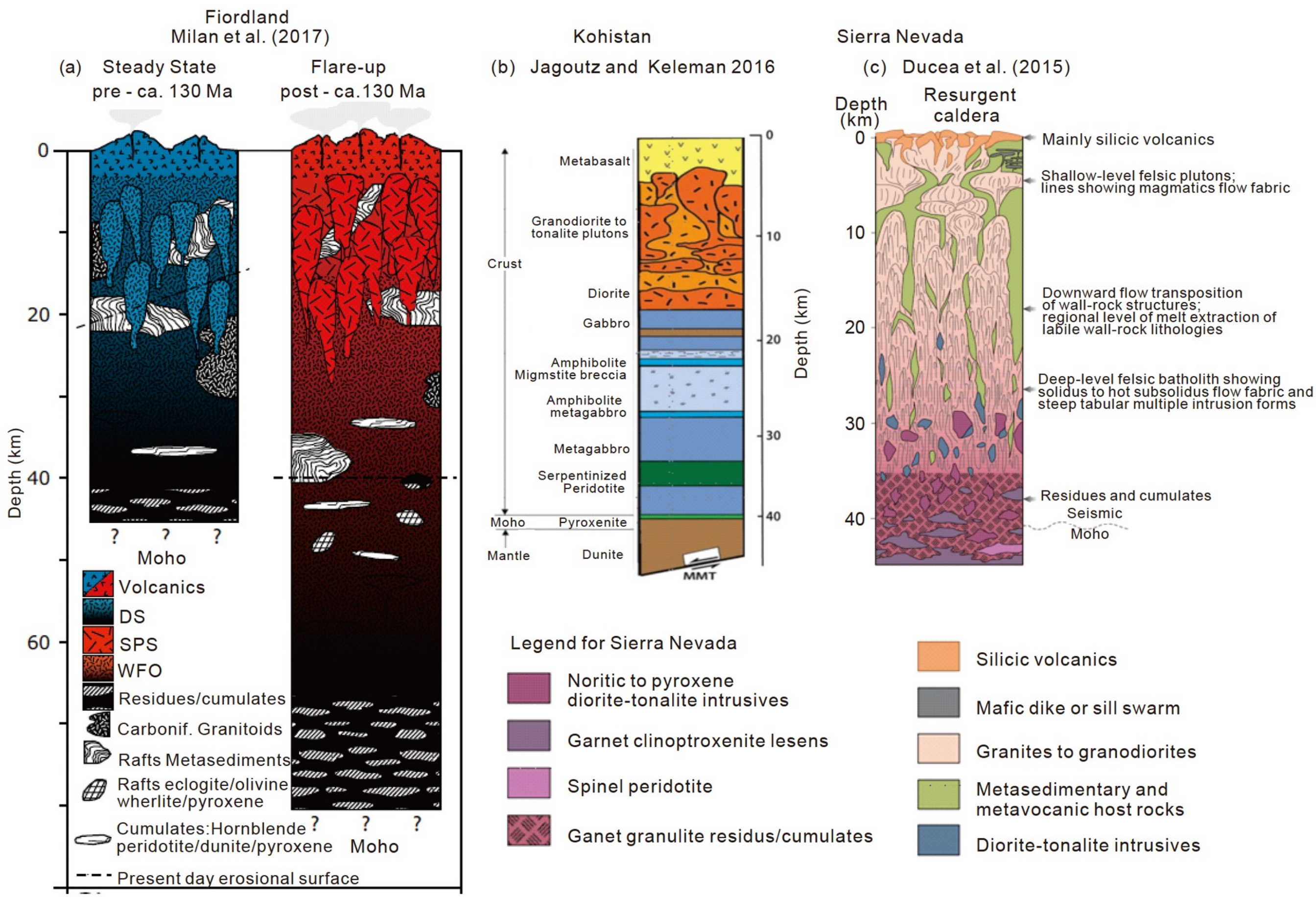

Figure 6 Crustal profile of oceanic and continental-margin arcs. Fiordland from Milan et al. (2017); Kohistan from Jagoutz and Kelemen (2015), and Mooney (2020); Sierra Nevada from Ducea et al. (2015a). Note that the mid-crust of all arcs is dominated by the intrusion of multiple steep-sided plutons of TTG and related rocks, with inter-pluton roof pendants forming down-warped synforms. Abbreviations in Fiordland sections: DS-Darran Suite (Triassic-Early Cretaceous, >129 Ma calc-alkaline rocks), SPS-Separation Point Suite (Early Cretaceous <125 Ma, alkaline to calc-alkaline granitoids), WFO-Western Fiordland Orthogneiss (Early Cretaceous pyroxene ± hornblende diorite and monzodiorite).

blendites, and amphibolite- to granulite facies migmatized remnants of pre-existing metasedimentary sequences (Figures 8, 10), typically with large garnet crystals surrounded by plagioclase-rich leucosomes (Depine et al., 2008). Deepest sections of arc-roots are rarely exposed, with Fiordland (55 km) of South Island New Zealand (Figure 1) perhaps being the deepest continental arc section on the planet (De Paoli et al., 2009). In this section (Figure 6a), between 30–35 km depth the rocks are garnet-rich plagioclase-free residues (Clarke et al., 2013), and below 50 km the rocks are pyroxene-amphibole ± garnet eclogites, garnet clinopyroxenites, and amphibole clinopyroxenites (Figures 5, 6) that are complementary to the melts in the higher crustal levels, thus, colloquially dubbed arclogites (Ducea and Saleeby, 1996; Lee and Anderson, 2015; Ducea et al., 2021a, 2021b). The Carboniferous Black Giant Anorthosite and surrounding granitoid gneisses of Fiordland (Figure 6) share many geological and geochemical features of Archean anorthosite-bearing layered intrusions and associated gneisses (Gibson and Ireland, 1999), suggesting that both Archean and Paleozoic anorthosite-gneiss associations originated in magmatic arcs (Sotiriou and Polat, 2020).

3.4 Models of crustal differentiation in continental-margin accretionary orogens and arcs

Taken on average, continental-margin arcs have major element geochemical compositions that are very similar to the average continental crust (Taylor and McLennan, 1985, 1995). This has been interpreted in some models (Rudnick, 1995; Rudnick and Gao, 2003; DeCelles et al., 2009; Paterson et al., 2011; Ducea et al., 2015a, 2021a, 2021b; Tang M et al., 2019) to suggest that differentiation (or, MASH) in continental arcs is one of the most important steps in the formation of continental crust, where dense mafic cumulates formed during deep-seated magmatic differentiation sink and by some process delaminate, leaving an upper crust more enriched in felsic crustal components. This hypothesized differentiation and delamination follows the first and second mantle melting events that first formed the subducted oceanic lithosphere, then formed the primitive intra-oceanic arcs that amalgamated to form the basement to the continental-margin arcs. The final step in forming continental or cratonic crust in these models is when these arcs collide together in a terminal (*sensu* Dewey, 1977) continent/continent collision, typically producing late-stage granitoids.

This conventional view has been challenged, because most arcs are not thick enough for the mafic cumulates to reach eclogite facies to promote foundering before collision, and there are relatively few examples of preserved dense arc crustal roots (the best are shown in Figure 6), or evidence for the present-day foundering beneath active arcs. Thus, Hildebrand et al. (2018) proposed alternatively that most of the continental crust was not produced by arc magmatism, but was produced by slab failure magmatism (see section 4 below) in the late stages of collision, forming the magmatic suites (typically TTG) that form some of the late- to post-collisional plutons in orogens. It is important to note that slab failure magmatism intruded after the cessation of arc magmatism is typically highly-silicic, of the TTG suite, and also has chemical characteristics that are similar to bulk continental crust- therefore, Hildebrand et al. (2018) suggested that the long-postulated MASH processes, with associated hypothesized delamination of the required but rarely-observed dense melt residues is not as important a process in the generation of continental crust as commonly assumed, and that slab failure has played a greater role in this regard since the early Archean.

3.5 Migration of magmatic axes of continental-margin arcs

While the locus of active magmatism in continental-margin arcs is usually narrow (25–30 km), the arc axis commonly migrates trench-ward or craton-ward at 1–5 mm/year (Gehrels et al., 2009) making the region affected by the main arc much wider (Figures 4, 5). In addition, continental-margin arcs are just one component of continental-margin accretionary-orogens, which typically consist of several (or more) accreted terranes and arc systems, forming a structurally complex orogen with multiple, in some cases contemporaneous belts of magmatism that are juxtaposed across thin suture zones, but that formed above different subducting slabs (Lee et al., 2007). Considering the fore-arc and backarc regions, continental arcs influence an area up to hundreds of km wide, much larger than IOAS—for example, the tectonically composite section of the North American Cordilleran accretionary orogen containing the Coast Range Mountains batholiths in SE Alaska and British Columbia is 250 km wide, and the post-Jurassic volcanic-plutonic section of the central Andes has a structurally-composite width of 300–800 km (Gehrels et al., 2009; Hildebrand, 2013; Ducea et al., 2015a; Horton, 2018)—larger than the size of many Archean cratons (for example, the Eastern Pilbara craton is only 200 km×200 km), and would easily "disappear" as a normal accretionary/plutonic terrane within the Sierra Nevada batholith (see Kusky et al. (2021) for discussion). Particularly wide areas of crustal thickening form in contractional arcs (Figure 4), where backarc fold-thrust belts can double crustal and mantle thicknesses (DeCelles, 2004), and in arcs that have accreted outboard terranes, or experienced collisions and have underplated arcs or continental fragments, as documented in Japan (Isozaki et al., 2010).

3.6 Life-spans of continental-margin arcs

The integrated lifespan of continental-margin arcs can be exceptionally long. For instance, the Andes grew over a period of ~500 Myr, including the Pampean (Cambrian) and Famatinian (Cambro-Ordovician) peri-Gondwanan arcs in the Central and Southern Andes (Figure 1), later affected by Permian to Mesozoic intrusive events of the Peruvian coastal batholith (Demouy et al., 2012). During their long-lives, the Andes and North American arcs have experienced several periods of arc, micro-continental, plateau, and seamount accretion (Saleeby, 1983; Ramos, 2008; Hildebrand, 2013; Hildebrand and Whalen, 2014a, 2014b), addition of material by tectonic off-scraping, and periods of tectonic erosion. During subduction erosion large amounts of crustal material may be returned to the mantle, but some of it may also be added back to the lower part of the upper plate, in a process called relamination (Figure 5), which if coupled with delamination of the mafic/ultramafic roots (Kay and Kay, 1993; Currie et al., 2015), tends to cause the arc crust to become more silicic with time (Castro et al., 2010; Hacker et al., 2011a, 2011b, 2015). Studies of the volumes of magma intruded into continental arcs (Gehrels et al., 2009; Paterson et al., 2011; Ducea et al., 2015b) show 5–20 Myr long periods of magmatic flare-ups with magma flux rates of 1000 km^2/Myr with longer (30–40 Myr) intervening periods of magmatic lulls with <100 km^2/Myr (Ducea and Barton, 2007).

3.7 Contributions to crustal growth from slab melting

Many studies have examined whether or not melting of subducting oceanic slabs is possible (e.g., Rapp et al., 1999; Rosenthal et al., 2015) and a potential source of crustal growth. Under special conditions involving elevated temperatures, such as during subduction of very young oceanic crust and episodes of ridge subduction (Sisson et al., 2003; Kusky et al., 2003), it has been suggested that slabs can partially melt, and generate adakitic melts, low-K trondhjemites (Li H et al., 2022) associated with diorites and granodiorites, or related magmas (e.g., Defant and Drummond, 1990). This was then taken by many to suggest that the TTG suite, or adakites in the Archean were produced by melting of subducting slabs under hotter mantle conditions in the Archean, a view challenged by Martin et al. (2014).

Recent work in the Sulu orogen, China, has directly documented decompression melting of ultrahigh-pressure (UHP) metamorphic eclogite during its exhumation from mantle depths in the collisional orogen, forming migmatitic eclogite (Wang L et al., 2014, 2021). During the early exhumation of continental crust, the melt component was released from eclogite through dehydroxylation of nominally-anhydrous (NAMs) minerals, and is inferred to have been a supercritical fluid at UHP conditions which was separated into aqueous fluid and granitic melt under high-pressure conditions (Wang S J et al., 2016, 2017, 2020a, 2020b). During later stages of exhumation, however, the partial melting of UHP eclogite is dominated by dehydration of nominally-anhydrous minerals such as omphacite, producing dioritic primary melt components, and minor hydrous minerals (e.g., phengite, zoisite; Feng et al., 2021). This new progress (Wang L et al., 2021) indicates that melting UHP eclogite is feasible under Phanerozoic orogenic conditions, and was compared to possible examples from the Eoarchean Isua Belt (Polat et al., 2015).

Melting of enriched oceanic crust (i.e., formed above a plume, or in Archean conditions) to form oceanic arcs above subduction zones was shown by Gazel et al. (2015) to have formed the juvenile continental crust of the Central American Land Bridge, and was suggested by them to be an important mechanism of generating continental crust in the Archean. Higher mantle temperatures in the Archean would have favored melting of oceanic crust in subduction zones (Defant and Drummond, 1990; Komiya, 2004).

In conclusion, mafic rocks from subducting slabs can melt under special conditions, but much future work remains to be done to determine the relative contribution of slab melts to crustal growth through time.

4. Late- to post-collisional magmatism

While arc magmatism is generally thought to be the largest contributor to crustal growth, there is less certainty about the origins and sources of late-to post-collisional magmas in orogens, and how much they represent reworking of the crust, or new additions to the crust (Davies and von Blanckenburg, 1995; Freeburn et al., 2017; Fu et al., 2019). Post-collisional magmatism differs from arc magmatism, in that it may be distributed for hundreds of kilometers from associated suture zones, and last in duration for tens of millions of years, after the main phases of the collision. The geodynamic causes of melting are varied with many different origins suggested (e.g., Pearce et al., 1990; Brown, 2007; Lee et al., 2012; Freeburn et al., 2017). Some models suggest magma sources from melting of orogenically thickened crust (England and Thompson, 1986), or associated with gravitational collapse of thickened crust (Dewey, 1988; Kusky, 1993). Other late- to post-orogenic triggers of magmatism may include delamination (Bird, 1978), or processes related to slab breakoff (failure) in late stages of collision (Davies and von Blanckenburg, 1995; Freeburn et al., 2017). Yet another setting for post-collisional magmatism is in metamorphic core complexes, which may form on overthickened crust, or over regions such as backarcs experiencing extension from slab rollback (Vanderhaeghe and Teyssier, 1997; Whitney et al., 2004, 2013; Lamont et al., 2020; Soleimani et al., 2021). In most metamorphic core complexes, the magmas form the cores of gneiss domes, typically forming migmatitic to leucogranite sheets derived from remelting of pre-accreted crust with little or no additional contributions from the mantle (Zheng and Chen, 2021; Zheng and Gao, 2021).

4.1 Slab failure magmatism in orogens

The importance of late- to post-collisional plutonic suites intruding into pre-collisional arc-related magmatic/accreted belts has recently been elucidated by detailed syntheses of the Appalachian, North and South American Cordilleran orogens by Hildebrand and Whalen (2014a, 2014b, 2017, 2021) and Hildebrand et al. (2018), and is also well-documented in other orogens such as the Tethyan Alpine-Himalayan system (Fu et al., 2019). In a synthesis of geochemical, temporal and spatial relationships of plutonic suites in these orogens, these authors have shown that many of the magmatic suites traditionally ascribed to arc-magmatism are instead interpreted as slab failure magmas, formed during late stages of collisions, where the subducting slab breaks-off, melts are generated by slab failure (Figure 7), by material moving up through the gap in the slab, from the edges of the detached slab, or from melting of the subducting oceanic lithosphere at depths greater than the ~100 km arc-magma genesis zone (Freeburn et al., 2017). Numerical modeling shows that for rapid slab failure with breakoff occurring below the lithospheric depth of the overriding plate, most melts are generated from the asthenosphere above the detached slab by release of water from the slab edges that are heated by the lower asthenosphere rising through the gap (Freeburn et al., 2017). Other melts are generated by melting of the subducting continental crust, especially where the breakoff occurs near the site of the subducted passive margin (Figure 7). The slab failure melts then experience fractional melting as they rise through the overlying sub-continental mantle lithosphere to generate the late- to post-tectonic suite of magmas (Hildebrand et al., 2018). These relationships are well-exemplified in the classical Taconic orogen of the North American Appalachians (Figure 1), where Middle Ordovician volcanic and sedimentary rocks of the Ammonoosuc arc that collided with North America in the Taconic Orogeny

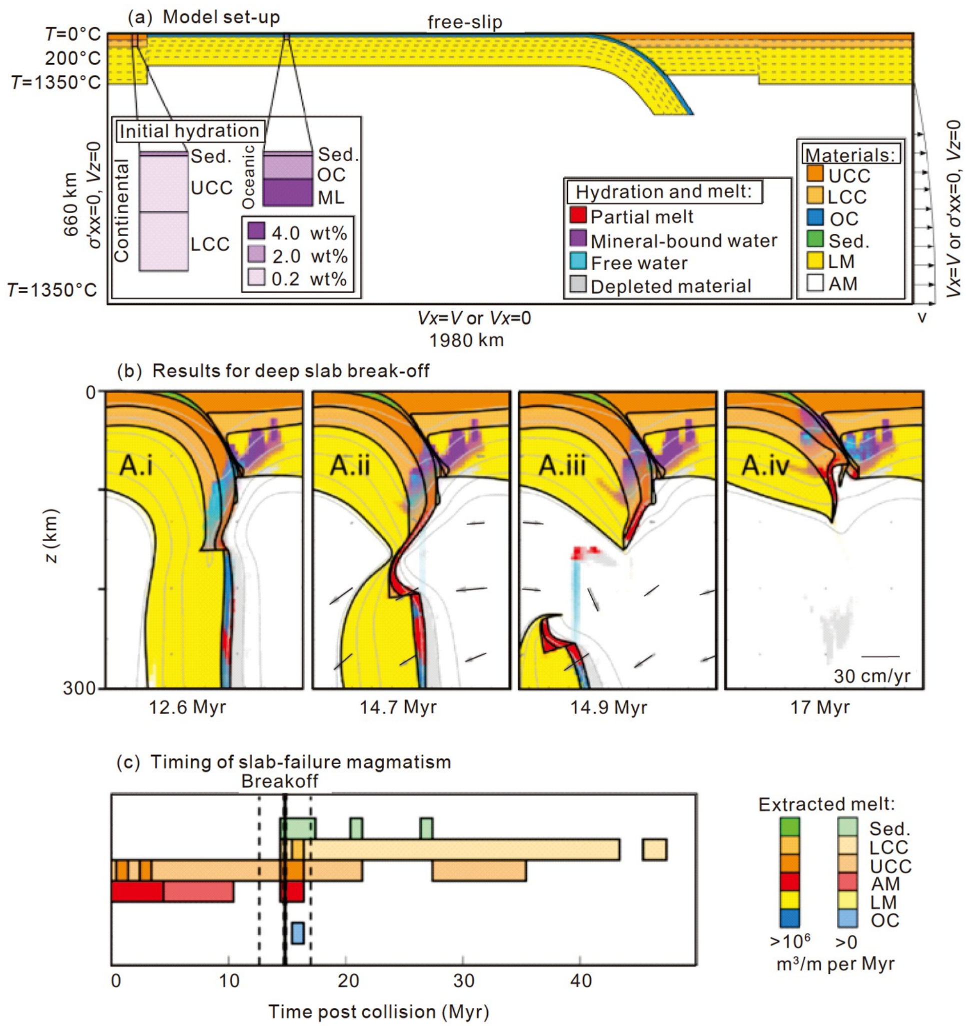

Figure 7 Snippits of numerical models of Freeburn et al. (2017) showing results for slab breakoff induced melting for slab failure at depths below the lithospheric thickness of the overriding plate. (a) Model set-up, with colors as apply to all panels. Abbreviations: sed=sediments, UCC=upper continental crust, LCC=lower continental crust, OC=oceanic crust, LM=mantle lithosphere, AM=asthenospheric mantle. (b) Results for slab breakoff at depths exceeding base of lithosphere of overriding plate. (c) Timing of magmatism from different sources related to slab breakoff. As the continental-margin is partially subducted, the down-going slab breaks off and the upper part containing the partly-subducted continental-margin isostatically rebounds to underplate and cause uplift of the overriding (arc) plate. Magmas are generated from melting of the subducted oceanic slab at >2 GPa in the garnet stability field, forming TTG magmas with distinctive trace element patterns including Nb/Y>0.4, La/Yb>10, Gd/Yb>2.0, Sm/Yb>2.5 and Sr/Y>10 (using the Hildebrand et al., 2018 discriminant plots). The magmas acquire different geochemical traits depending on whether they rise through old or young SCLM of the rebounded partly-subducted slab, and of the character of the overriding plate. Slab failure TTG magmas typically intrude, therefore, as late- to post-deformation plutons and sills, often into the locus of previous arc plutons forming nested plutonic suites, that are the same age as, or slightly younger than, the associated arc magmas and deformation/metamorphic fabrics from subduction. These plutons and sheets are typically deformed into domes soon after their intrusion and during subsequent accretion events. Many variations are possible depending on slab age, dip, and composition of upper and lower plates, and whether or not additional melts are generated from the slab window or mantle wedge.

(Stevens, 1970; Rowley and Kidd, 1981; Stanley and Ratcliffe, 1985; Bradley and Kusky, 1986; Hildebrand and Whalen, 2021) are intruded by an Upper Ordovician plutonic suite known as the Oliverian plutons (Billings, 1937; Tucker and Robinson, 1990), best known from the 400 km-long Bronson Hill anticlinorium that extends from Maine to southern Connecticut (Figure 8). The plutons in New England form a belt including more than two dozen TTG gneiss domes (Hollocher et al., 2002), mantled by the slightly older to contemporaneous volcanics, and deformed into domal structures during the later Devonian Acadian orogeny (Leo G W, 1991) and are generally interpreted to be large-scale east-vergent recumbent fold nappes (Thompson, 1956; Thompson et al., 1968; Robinson et al., 1979, 1991), bringing peri-Gondwanan affinity basement (Tucker and Robinson, 1990; Hibbard et al., 2006) with the Taconic arcs allochthonously onto the North American continent.

The Oliverian gneiss domes (Billings, 1937) of the Bronson Hill anticlinorium (Figure 8) bear remarkable similarities to the dome-and-basin structures of the Eastern Pilbara, which are often used to claim that a different tectonic regime operated on the early Archean Earth than at present (this is

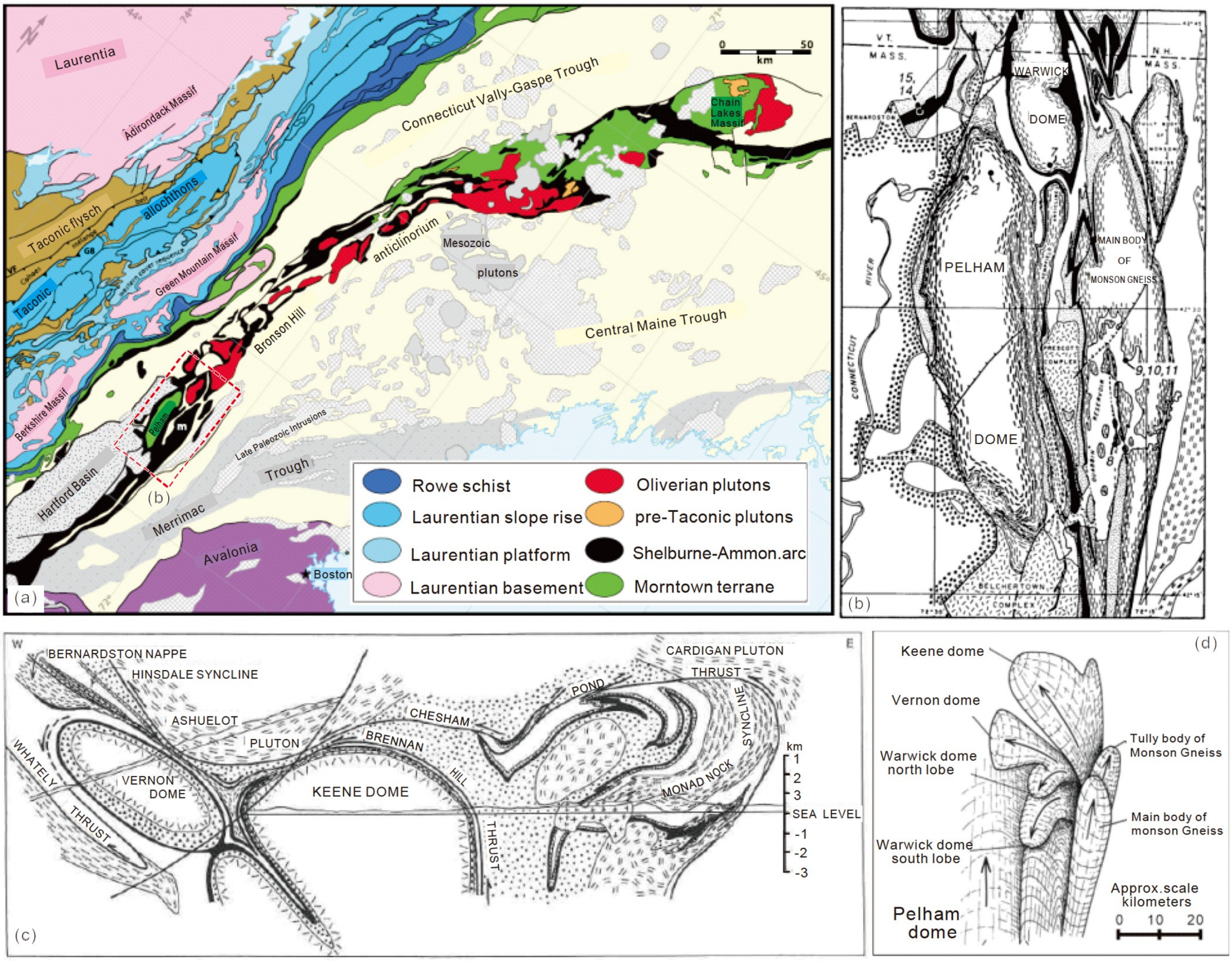

Figure 8 (a) Map of the New England Appalachians showing the location of the Oliverian domes in the Bronson Hill Anticlinorium. Main map from Hildebrand and Whalen (2021), modified from Hibbard et al. (2006) with modifications from Karabinos et al. (2017). The Laurentian (North American) basement of general Grenvillian affinity is best exposed in the Adirondack dome and Green Mountains and Berkshire massifs, whereas the Laurentian passive margin and foreland basin (shown as the Ordovician Taconic flysch on map) overlie this basement gneiss complex. The Taconic allochthons are comprised of the Laurentian slope and rise facies, thrust over the Laurentian margin. The Rowe schist represents the suture zone between Laurentia and the accreted Taconic arc, which is formed by the Moretown-Shelbourne and Ammonoosuc volcanics in thrust sheets with associated plutons. These are intruded by the late- to post-collisional TTG plutons of the Oliverian suite. (b) Detailed map of the Pelham, Monson, and Warwick domes located in the southern end of the main map (modified from Robinson et al., 1992), showing details of typical dome-and-keel structures, with gneissic domes with older basement, intruded by plutons that are the same age and slightly younger than the mantling, multiply deformed and metamorphosed volcano-sedimentary sequences. The domes were flattened from more rounded to elliptical shapes in the younger (Devonian) Acadian orogeny. (c) Robinson's (1991) restored cross section across the Vernon and Keene Domes, and (d) Robinson's (1991) spectacular restored structural relief model with the inferred flow lines for the major gneiss domes.

documented in detail in Kusky et al., 2021). The Oliverian domes are comprised of gneissic plutonic rocks that are strongly deformed and metamorphosed, and have been the source of controversy about their origin for a century (Robinson et al., 1991). Some of these, such as the Pelham dome in Massachusetts (Figure 8) have cores of Gondwanan-affinity circa 613 Ma gneisses (Tucker and Robinson, 1990), intruded by Late Ordovician (454–442 Ma) plutonic suites, and overlain by metasedimentary rocks as young as Devonian before they were deformed into the domal fold nappes (Robinson et al., 1992). Some of the domes are intruded by several generations of intermediate composition magmas of the TTG suite, with ages as young as 445 Ma (Hollocher et al., 2002) which is up to 10 Myr younger than the 455–451 Ma Taconic arc/continent collisional orogeny (Jacobi and Mitchell, 2018; Hildebrand and Whalen, 2021), and with geochemical signatures (no Eu anomalies, low Y, depleted in heavy REE) that are characteristic of magmas worldwide related to slab failure (Hildebrand and Whalen, 2014a, 2014b, 2021). The mantling sequence (Ammonoosuc volcanics) include generally tholeiitic basalt, some andesite and dacite, metamorphosed sea-floor hydrothermal deposits now consisting of cummingtonite-gedrite-cordierite (Robinson et al., 1998), with minor carbonate and conglomerate near the top of the lower sequence. These are overlain by garnet-amphibole-bearing quartzites (interpreted to be volcanogenic exhalative deposits (Hollocher, 1993), dacites, rhyolites, and amphibolites. These, in turn, are overlain by sulfidic black shales, arkosic sandstone and bimodal volcanic rocks of the Partridge Formation (Figure 8) (Hollacher, 1993).

In their geochemical synthesis of the early Paleozoic magmatic rocks of western New England, Hildebrand and Whalen (2021) found that they belong to two groups- rocks with arc compositions of the Ammonoosuc arc have ages>450 Ma, whereas the plutons with ages of 450 Ma and younger (to 445 Ma) have geochemical signatures of arc-failure magmatism. Melts generated in this way explain the age discrepancy with the plutons being younger than the arc volcanics, and also explains why the Oliverian plutons have long been known to have a Laurentian isotopic signature (see Dorais et al., 2008, 2012). As the melts formed and escaped from the deeply sinking detached slab, they rose through old Laurentian lithosphere that was underplated during isostatic rebound after slab breakoff (see model in Figure 7), and as the magmas rose through this old sub-continental lithospheric mantle, they assimilated the old enriched LILEs, explaining the isotopic and trace element signatures (Hildebrand and Whalen, 2021).

The rock assemblages in the Bronson Hill anticlinorium, and its intrusive TTG suite and overall structure of mantled gneiss domes forming a "dome-and-basin-type" regional pattern is remarkably similar-to that of the Mesoarchean Eastern Pilbara craton (Kusky et al., 2021), which have been widely used to suggest that Earth's early-mid Archean tectonic regime was different from younger times (e.g., Johnson et al., 2014; Hawkesworth et al., 2020). The similarities of the Oliverian domes to the Archean domes-and-basins include rock types, polyphase structures, and relative ages of the gneissic to granitoid cores and mantling volcano-sedimentary sequences. Likewise, Kusky et al. (2021) showed additional similarities between the Pilbara dome-and-basin structures with plutonic domes with mantling volcano-sedimentary sequences of the Sierra Nevada and Coast Ranges batholiths, documenting no changes in structural styles, rock types, or relative age relationships between Mesoarchean, Paleozoic, and Mesozoic plutonic belts.

The magmas formed in this way are typically TTG suites, which typify orogens of all ages, back to the Eoarchean (Windley et al., 2021; Sotiriou et al., 2022). Through a compilation of arc failure magmas of the North and South American Cordilleras, Hildebrand and Whalen (2014a, 2014b, 2017) and Hildebrand et al. (2018) have developed a general petrogenetic model for the formation of most late- to post-orogenic arc-failure magmas from partial melting of the garnetiferous plagioclase-free metabasalt and gabbro (i.e.,> 2 GPa) from the detached slab, that create siliceous, generally TTG magmas with distinctive trace element patterns including: Nb/Y>0.4, La/Yb>10, Gd/Yb>2.0, Sm/Yb>2.5 and Sr/Y>10 (Hildebrand and Whalen, 2021). Since the magmas are generated in the slab, they must rise through the overlying SCLM of the partially-subducted leading edge of the subducting plate attached to the thinned continental-margin (Figure 7), which may have different characteristics-if the overriding SCLM is old and enriched the rising melts may assimilate LILE and acquire crustal-type Sr and Nd signatures (Hildebrand et al., 2018).

4.2 Orogenic collapse related granitoids

In young orogens, upper crustal extension often follows, or accompanies, crustal thickening and can be related with the formation of graben at high angles to the orogenic strike as in Tibet, or can be linked with the formation of metamorphic core complexes and the emplacement of late-collisional minimum melting anatectic granitoids (Dewey, 1988; Burchfiel et al., 1992; Zheng and Chen, 2017, 2021; Zheng and Gao, 2021). The granites are typically K-rich, cross-cutting structures, similar to those that mark one of the final stages in the evolution of many cratons (Kusky, 1993). In addition, orogenic thickening of crust can in some cases lead to lithospheric delamination, or the formation of "drips" of the dense lower part of the lithosphere that decouples from the overlying crust and sinks into the asthenosphere and deeper levels of the mantle during collisions, thereby placing hot asthenosphere directly against the base of the collisionally-thickened crust (Bird, 1978; England and Houseman, 1989; Göğüş et al., 2017). Petrological arguments (Ducea and Saleeby, 1996, Saleeby et al., 2003) and deep seismic profiles beneath the Alps and Pyrenees are compatible with lithospheric delamination beneath active orogens (Nicolas et al., 1990; Nelson, 1991; Prodehl and Mooney, 2012) suggesting that crust thickened to more than 65 km becomes gravitationally unstable (because the strength of quartz and olivine can no longer support the thick crust, and the deformation mechanism changes to power-law creep (England and Bickle, 1984) and will begin to collapse towards the topographic free face if the plate-boundary forces are relaxed and there are no increased vertical forces to buoy up the crust (Dewey, 1988). In early studies McKenzie and Bickle (1988) and Nelson (1991) suggested that upper crustal extension coupled with delamination of the lower crust or sub-continental lithospheric mantle will produce decompression melting in the upper mantle. This in turn will form mantle-derived basaltic melts that pond along the Moho, and in the lower crust, causing anatectic melting that generates intermediate to silicic granitoids. Kusky (1993) related these to the common late- to post-kinematic granitoids that intrude many accretionary greenstone terranes worldwide, and suggested that these are a post collisional response that may be recognized as a sign of orogenic crustal thickening from collisional orogenesis.

4.3 Post-collisional lithospheric foundering magmatism

Recent studies have suggested a temporal relationship between lithospheric foundering and felsic magmatism at convergent plate margins (Zheng and Gao, 2021). This leads to the suggestion that post-collisional thinning of orogenic lithospheres would be associated with thinning of mantle lithosphere through removal of the orogenic root, leading to active rifting and rift failure magmatism at shallow crustal levels (Zheng et al., 2022). This mechanism may also have operated in the Archean in response to lithospheric foundering. In addition, this deep-foundering and related shallow rift magmatism may also be caused by slab rollback in the late stage of oceanic subduction and slab breakoff in the terminal stage of oceanic subduction (Hildebrand et al., 2018; Zheng et al., 2022). Substantially, rift failure magmatism is a shallow manifestation of such deep processes as slab rollback, slab breakoff and orogenic lithospheric root foundering at mantle depths (Zheng et al., 2022). Therefore, Archean active rifting along convergent plate margins would be caused not only by slab rollback and slab breakoff in the syn-collisional stage but also by lithospheric foundering in the post-collisional stage.

4.4 Alkaline and carbonatitic magmatism

An additional setting of magmatism in orogens is related to suture zones, where original rift-related magmas (ARCs, or alkaline igneous rocks, and carbonatites, in the definition of Burke and Khan, 2006) are associated with suture zones, and hence are named DARCs (deformed alkaline igneous rocks and carbonatites). In classic papers, Burke et al. (1977, 2003), Burke and Khan (2006), and later Zheng and Gao (2021) suggested that older ARCs can be deformed in suture zones, marking sites where rifts have been reactivated, and the alkaline rocks are deformed into DARCs. When these alkaline rocks are subducted to > 100 km during collision, they provide the material for melting for later magmatism (Burke et al., 2003).

One under-appreciated example of a DARC in China may be the circa 1.4 Ga Bayan Obo carbonatite on the north margin of the North China block, formed during the breakup of the Columbia Supercontinent, then remobilized during subduction and collisions of the Paleo-Asian Ocean in the Silurian and Permian (Ling et al., 2013). Initial deposition was in the Bayan Obo rift at about 1.75 Ga (Kusky and Li, 2003; Li et al., 2007; Kusky and Santosh, 2009), and the carbonatite magmatism was widespread along the north margin of the craton from 1.4 to 1.2 Ga during the final break-up of Columbia (Fan et al., 2016). The deformation of the Bayan Obo carbonatite bodies was a result of a possibly highly oblique Permian collision on the northern margin of the North China block. They resemble other DARCs (Deformed Alkaline rocks and Carbonatites) in being greatly elongate in length to breadth proportions compared with their postulated ARC (alkaline rock and carbonatite) parent bodies. Those are usually close to round in outcrop which is considered normal in their sub-volcanic environments.

Metasedimentary carbonate rocks in highly deformed environments are not generally as highly deformed as carbonatites. The great elongation of DARC's is here suggested to be a consequence of lack of strength as a consequence of the fluidity of associated material part of which, after eruption, is preserved as metasomatic fenite, as at Bayan Obo (Fan et al., 2016). Structure similar to that of Bayan Obo is familiar elsewhere: For example, in (1) The nepheline gneiss and carbonatite Kpong DARC of SE Ghana which is mapped as ~50 m thick over a length of ~60 km; (2) The Khariar body of the Eastern Ghats of India which is mapped as 30 km long (Leelanandam et al., 2006), (3) The carbonatites of the Grenville Province in Ontario (Burke and Khan, 2006). It seems likely that, as at Bayan Obo, these three highly deformed DARCS will prove discontinuous along strike when mapped in detail.

5. Crustal growth by arc magmatism recorded in dome-and-basin structures of some Archean terranes

One of the most contentious issues in the Earth Sciences is what was the style of tectonics, or non-tectonic regimes on the early Earth (see review by Windley et al., 2021). Proponents of the stagnant lid, heat pipe and other non-plate tectonic models discussed above typically use the "dome-and-basin" structures of the Mesoarchean Pilbara craton, to claim that they are different from anything on the present Earth, and therefore that plate tectonics did not produce them (e.g., Nebel et al., 2018; Hawkesworth et al., 2020; Johnson et al., 2014, 2019). In contrast, Kusky et al. (2021) examined the structural and magmatic evolution of the domes and surrounding greenstones, and claimed that they are complexly and multiply deformed deep sections of arcs, produced in very much the same way as the intra-oceanic and continental-margin arc systems in accretionary orogens described above. Here, we briefly review the salient points.

Map patterns of Archean terranes (Figure 9a) can be generally divided into three categories (Kusky and Vearncombe, 1997), including dome-and-basin regions (e.g., parts of Zimbabwe and Pilbara), broad tracks with internally bifurcating thrust-bound units (e.g., parts of Slave, North China, and Yilgarn Provinces), and long-linear belts with

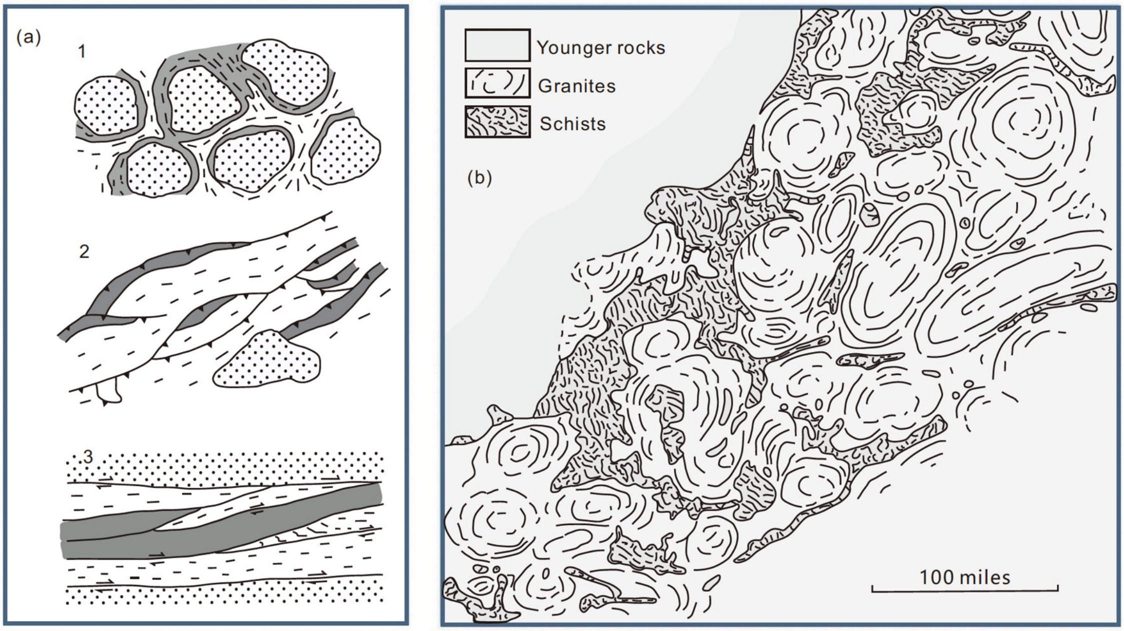

Figure 9 (a) The three main types of structural patterns found in Archean greenstone-granite terranes, globally (from Kusky and Vearncombe, 1997). 1, broad domal granitoids with inter-domal greenstones defining the "dome-and-basin" pattern; 2, broad greenstone tracts with internally bifurcating lithological domains, and irregular granitoid contacts; 3, long narrow and linear patterns of alternating metasedimentary, metavolcanic plutons, high-grade gneissic, and plutonic terranes. (b) Idealized sketch of the so-called gregarious batholiths of Zimbabwe, by MacGregor (1951). This was drawn before detailed regional mapping of the craton.

bounding strike-slip fault systems (e.g., Superior, parts of Yilgarn). Other terranes, such as the North Atlantic, preserve various elements of all of these, but are complexly refolded into interference patterns. The dome-and-basin patterns are relatively rare compared to the others (Kusky et al., 2021).

The dome-and-basin concept had its early origins in Macgregor's (1951) gregarious batholith concept (Figure 9b), where he drew stylized (later, proven by his own mapping to be incorrect) rings around gneiss complexes in Zimbabwe that he interpreted to have risen through surrounding cover sequences. This same principle, of drawing idealized rings around normal plutons (not real foliation or lineation measurements), and calling them gneiss domes, continues in many cratons to this day, without proper supporting structural measurements of foliations and lineations, nor determination of the timing of magmatic and structural events that are essential to prove that the structures are gneiss domes that rose through an overlying carapace (e.g., see Whitney et al., 2004, and Kusky et al., 2021 for discussions). Horizontal plate tectonic models that explain the map patterns of the Zimbabwe craton were later presented by Stowe (1971, 1974), Kusky and Winsky (1995), Kusky (1998), Jelsma and Dirks (2002), and Sawada et al. (2021).

Later, the eastern Pilbara craton became recognized as the most "typical" (because others are so rare) dome-and-basin area (Hickman, 2016; Van Kranendonk et al., 2015; Kusky et al., 2021). The part of the eastern Pilbara with the dome-and-basin structures is a mere 200 km×200 km in area (smaller than the seamount of Hawai'i (Figures 10, 11), and six times smaller than just the Sierra Nevada batholith), yet has become a focus of recent models of the early Earth, claiming that the outcrop patterns of domes-and-basins are unique to the early Archean, and not found elsewhere on Earth (Bédard, 2006, 2018; Johnson et al., 2014, 2019; Hawkesworth and Kemp, 2006; Hawkesworth et al., 2020). This is incorrect.

In a recent synthesis, Kusky et al. (2021) assessed the structural evolution and magmatic succession of the eastern Pilbara, and concluded that the domes and magmatic rocks formed in a typical oceanic to continental-margin accretionary orogen/arc system (Figure 10). The domes formed during multiple horizontal shortening events, punctuated by, and in some cases, contemporaneous with magmatic intrusions. The result is a regional Type-I fold-interference pattern. The oldest rocks in the eastern Pilbara belong to an ophiolitic affinity ultramafic-gabbro-anorthosite suite known as the Mount Webber Gabbro (3.56–3.48 Ga), which is the same age as the lower mafic/ultramafic volcanic rocks of the Warrawoona Group. This is interpreted by Kusky et al. (2021) as an imbricated ophiolitic/oceanic/OPS sequence, that when thickened during accretion, yielded early felsic melts along thrust planes. There is no evidence that the mafic/ultramafic sequences were erupted through an older sialic basement, forming a 40 km thick dense and heavy layer that sank though soft warm and mushy older sial as required by the drip-tectonic models of Van Kranendonk et al. (2015), Nebel et al. (2018), Hawkesworth et al. (2020),

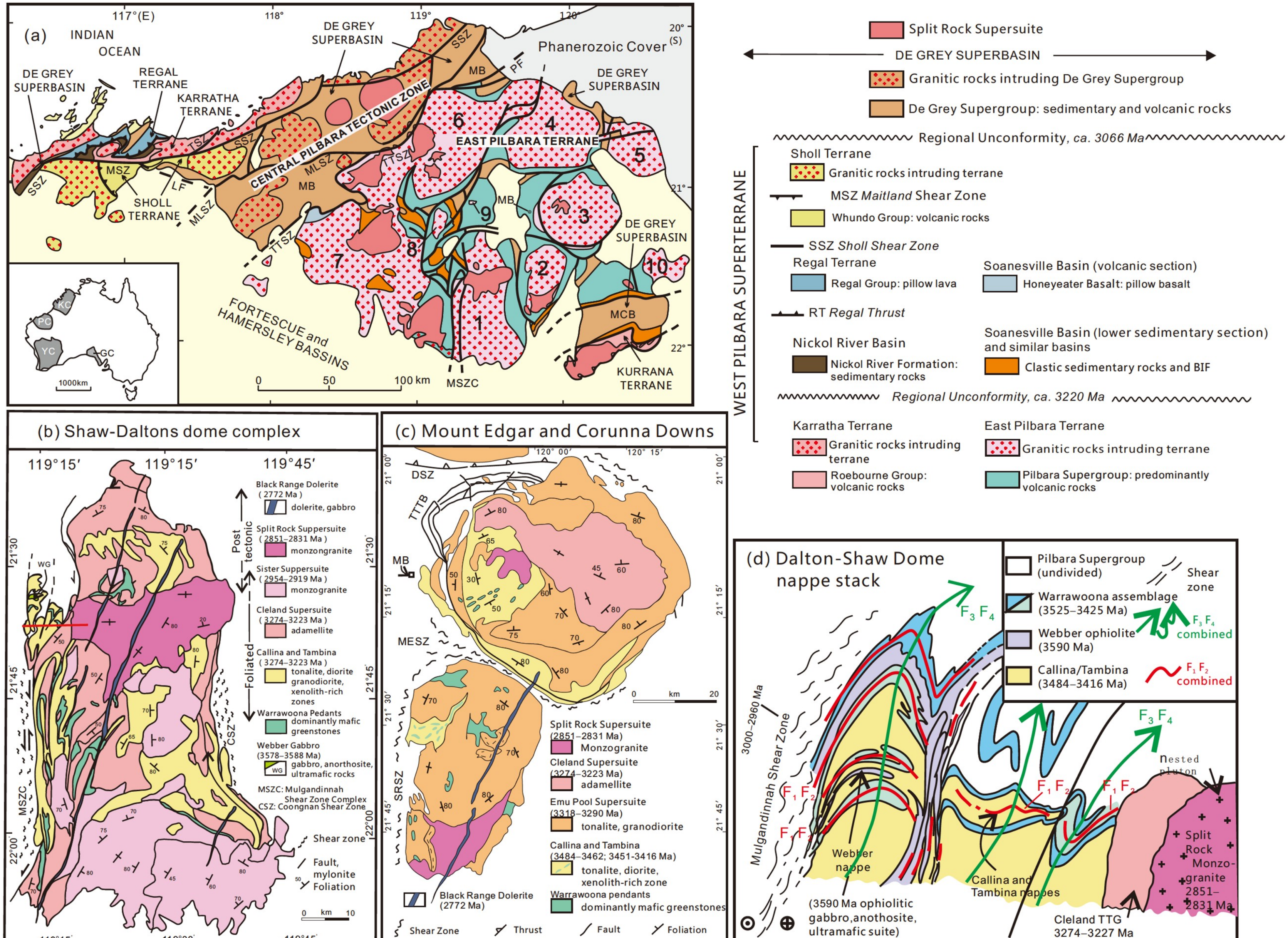

Figure 10 (a) Simplified geological map of the Pilbara craton (from Kusky, 2021a). (b) Litho-structural map of the Shaw-Daltons dome complex (number 1 on (a)). (c) Litho-structural map of the Mount Edgar and Corunna Downs dome complexes (2 and 3 on (a)). (d) Structural profile across the NW part of the Shaw Daltons Dome, showing that it consists of a pile of stacked nappes later folded by upright folds by transpression along the Mulgandinnah shear zone, forming a dome structure. Line of profile for cross-section in (d) is shown as red line in the NW part of (b).

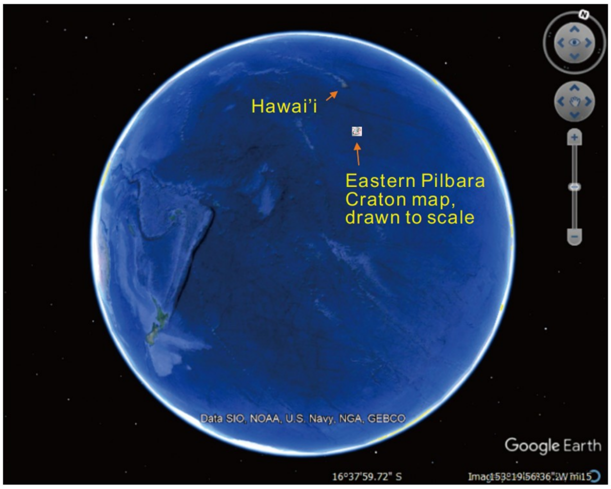

Figure 11 The Pacific plate, view from Google Earth. Plate tectonics is a local phenomenon on the present Earth, with large parts of the planet showing heat loss through conduction or plumes, like Hawaii, located near the center top of the image. The small square pasted south of Hawai'i is the map of the eastern Pilbara dome-and-basin province, drawn to scale, for comparison.

and Johnson et al. (2014, 2019). As subduction and accretion continued, arc-related magmas of the Callina and Tambina Supersuites intruded the structural stacks of imbricated oceanic assemblages between 3484 and 3416 Ma, then were deformed in a likely arc/arc collision, folded, and intruded by slab failure magmatism of the Emu Pool Supersuite during late deformation at 3324–3277 Ma and the Cleland Super-suite at 3274–3223 Ma, with magmas intruding along thrust planes, and refolding the structures into domes. A large part of the 200 km×200 km area of the eastern Pilbara was affected by transcurrent faulting of the Mulgandinnah Shear System (Figure 10) at 3000–2960 Ma, with domes being further uplifted in an en-echelon transpressional fold belt associated with the major shear zone (Kusky et al., 2021). Post-orogenic magmatism (collapse?) is recorded by the Split Rock and related suites from 2851 to 2831 Ma, and craton-wide (200 km×200 km) rifting is recorded by the intrusion of the Black Range dolerite dike swarm at 2772 Ma.

Kusky et al. (2021) presented this tectonic summary of the development of the eastern Pilbara, but also challenged that the map patterns, rock types, and spatial – temporal relationships are any different from typical continental-margin accretionary orogens and arc systems. By showing maps at similar scales, it was shown clearly that the dome-and-basin map pattern is found in continental-margin batholiths, if eroded to sufficient level (Figure 5), and the vertical magma transport and formation of structural domes is a normal consequence of horizontal plate tectonics. Further, if one examines the scale of the whole craton, it is evident that the batholith belt, with dome-and-basin structures, is linked outboard with an accretionary orogen, with the same tectonic zonation and scale, as the American Cordillera (Kusky et al., 2021). Thus, it is concluded, with five key geological indicators (rock types, processes, structures, and spatial and temporal scales of magmatism and deformation) having high degrees of similarity, that the processes that produced arcs and accretionary orogens on the modern Earth, are the same as those that produced the similar structures in the Mesoarchean of the Pilbara.

6. Discussion

6.1 Summary of crustal evolution by arc magmatism and accretion above subduction zones

Above, we describe processes at predominantly modern convergent margins, where subduction of an oceanic plate beneath another oceanic plate forms intra-oceanic arc systems, and subduction of an oceanic plate beneath a more evolved arc or continental-margin produces a continental-margin arc. There is a continuum of crustal thicknesses between oceanic arcs and continental-margin arcs, ranging from a few tens of km in primitive oceanic systems (15–35 km), to more typical 30–50 km in continental-margin arcs. The Andes of South America currently preserve some of the thickest crust of any arc systems on Earth (up to 70 km), and we refer to these systems as over-thickened continental arcs in accretionary orogens. Similar processes occur in all arc systems, primarily off-scraping and accretion or subduction erosion at the trench, release of volatiles at depth to produce melting in the mantle, which generates mafic melts that rise to the crust. The melts that emerge in oceanic arcs are primarily mafic, including IAT basalts, and a mid-crust layer of TTG-type intrusive suites forming a complex mid-crustal dome-and-basin type structural pattern as found in a few Archean terranes such as the Pilbara and Zimbabwe cratons (e.g., Kusky et al., 2021). As the crust becomes thicker by either MASH processes, or by structural thickening during accretion of other arc terranes, the melts must rise through and assimilate more crustal material, and become more felsic with time, and the early plutons and sheet-like intrusions are deformed into domal structures.

The crustal composition paradox is currently unsolved, as most melts in arcs are mafic, but the continental crust has a composition of andesite or tonalite/granodiorite (Rudnick and Gao, 2003; Gazel et al., 2015). One popular set of models suggests that this can be resolved by delaminating a thick heavy arc crust root (arclogites and related rocks), removing the mafic/ultramafic component, and leaving an upper crust that resembles the bulk continental crustal composition (Rudnick and Gao, 2003; Ducea et al., 2015a, 2021a, 2021b). Another model suggests that most of the continental crust formed by late-stage slab failure magmas of TTG composition, and most arcs never were thick enough to develop a thick dense mafic/ultramafic root, or to reach eclogite facies, a *prima facie* requirement for delamination (Hildebrand et al., 2018).

Throughout this review, we make comparisons between the modern systems, and Archean systems, and note remarkable similarities in the rock types, processes, structures, and spatial and temporal scales of magmatism and deformation in arc systems throughout geological time. This comparison applies to all parts of the arc as a linked system, from the accretionary wedge, to the magmatic axis (from surface to arc root), to back arc regions, and for the duration of arc magmatism. Detailed reviews of these processes in preserved accretionary wedges and accreted arcs in the Archean can be found in Windley et al. (2021), Kusky et al. (2018, 2020, 2021a), and Sotiriou et al. (2022). Thus, holistically, we can say with a high degree of confidence that the continental crust has grown predominantly through additions to the crust by accretion in forearcs (e.g., reviews in Kusky et al., 2013, 2020), and magmatic additions in juvenile and mature arcs. Other contributions, such as from massive outpourings of mafic lavas in large igneous provinces (e.g., Ernst et al., 2021) contribute mostly mafic magmas to the continental crust, so are not the main source of continental growth. Thus, we suggest that processes of arc tectonics and subduction have been in operation throughout the preserved geological record, and application of the null hypothesis states that we can apply the plate tectonic paradigm, and search for any difference with time. This is because of higher heat production in early times, and possible slightly higher mantle potential temperatures in the Archean than in the Phanerozoic, with differences varying from < 100°C (Aulbach and Arndt (2019) to as large as 250°C (Korenaga, 2011, 2016, 2017). Plate tectonics apparently has been able to buffer the effects of decreasing energy from orbital dynamics (Hofmeister et al., 2022), a cooling mantle, and maintain surface conditions to such a degree (e.g., Wang Z S et al., 2020) that enabled the development of life on a habitable planet.

6.2 Evidence for subduction and crustal growth by arc magmatism in Archean accretionary orogens through time; deeper global considerations

Many recent arguments against the operation of plate tectonics in the Archean, in favor of some kind of alternative Earth with stagnant lid behavior, state that those who wish to use the plate tectonic paradigm to explain the Archean geological record need to prove the existence of a globally-linked network of narrow plate margins, including ridges, trenches, and transforms (Bercovici et al., 2015; Lenardic, 2018a, 2018b; Brown et al., 2020). However, this is an untestable requirement to demonstrate the operation of plate tectonics on the early Earth, since so little of the plate mosaic from any given interval is preserved, and the remnants of the Eoarchean world are typically a few to tens of km across (e.g., Nuliak, Nuvuagittuq, or tens by a couple hundred km (Isua and SW Greenland). Thus, this is an untenable argument.

In separate reviews, we have clearly documented the presence of all types of plate boundaries in the Archean (Kusky et al., 2018; Windley et al., 2021), with clear evidence for extensional and convergent boundaries operating back to the Eoarchean, forming the accretionary orogens and their components that form Earth's oldest rocks. All Archean crustal fragments preserve evidence for formation and deformation at plate boundaries (Kusky et al., 2018; Windley et al., 2021). Despite the evidence summarized in those papers for the operation of plate boundary processes throughout the geological record, some of the proponents of the stagnant lid hypothesis argue that these clearly documented examples are just "local examples" of subduction, or "local spreading centers" on a planet dominated by stagnant lid behavior (e.g., Bédard, 2018; Johnson et al., 2014, 2019; Nebel et al., 2018; Brown and Johnson, 2018, 2019; Cawood et al., 2018; Hawkesworth et al., 2020). While the first so-called "requirement" to prove the operation of plate tectonics to the proponents of a stagnant lid Earth is non-testable, the second argument amounts to a dismissal of the overwhelming evidence for the operation of plate tectonics founded in the geologic record, and the same argument could be used to suggest that the present Earth is a stagnant lid, with only "local operation of subduction or spreading" (Figure 11). So, we ask, on the present surface of Earth, how abundant are plate boundaries, and how likely are they to be encountered by a geologist randomly dropped from space?

6.3 Plate boundaries are local phenomena with global implications

A simple look at the areas of present-day plates and plate boundaries (not including the wide diffuse plate boundary deformation zones, *sensu* Kremmer et al., 2014) high-light the misconception that well-documented examples of Archean paleo-plate boundaries may simply be local phenomena that do not relate to a global tectonic network. Present-day convergent margins have a length of 61,900 to 61,940 km (Bird, 2003; Matthews et al., 2016). The width of subduction zones (arc-trench gap) ranges from about 100 km to a maximum of 590 km, with an average of about 200 km. This means that convergent margins occupy 12,388,000 km^2, or about 2.4% of Earth's present surface (Earth's surface area is 510,100,000 km^2). So, on the present Earth, subduction is clearly a local phenomenon! Does this mean the present planet is a stagnant lid? The size of the largest plate (Pacific) is 103,000,000 km^2, or 20.19% of Earth's surface, and 10 times larger than the total subduction zone area! Clearly, a large part of the present Earth could be interpreted from surface geology as an internally stagnant lid (Figure 11), losing heat through conduction and advection in plumes including Hawai'i. Now, if we take the seven largest plates (Pacific, North American, Eurasian, African, Antarctic, Indo-Australian, and South American), they occupy 471,700,000 km^2, or 92.5% of Earth's stagnant surface! Going further, and adding in the smaller Nazca, Philippine Sea, Arabian, Caribbean, Cocos, Caroline, and Scotia plates, we have an area of> 500,000,000 km^2, or >95% of Earth's surface that is not occupied by an active plate boundary.

Thus, the argument that identification of a paleo-subduction zone (a very lucky find indeed), or paleo- sea-floor spreading axis are just local phenomena, on a planet that deceptively appears to the untrained eye to presently have a stagnant lid, probably applies more to the present-day Earth, than the Archean Earth, when the planet likely lost the extra heat produced through a greater plate boundary length, and faster plate motions (e.g., Burke and Kidd, 1978). Thus, the wisdom of the early models of Sleep and Windley (1982), and Abbott and Hoffman (1984) and others, who suggested that on the early Earth, which had to remove more heat than at present because of the higher heat production and secular cooling, the planet probably responded by more vigorous convection and plate motions, a greater ridge length, thicker oceanic crust, and smaller plates (Burke and Kidd, 1978; Sleep, 2000; Harrison, 2020), seems to hold. In contrast, the models of stagnant lids have no geologic evidence upon which to stand.

The descriptions of intra-oceanic arc systems, accretionary orogens, continental-margin arc systems, and micro-continental collisions presented above suggest that geological processes at paleo-convergent plate boundaries were fundamentally quite similar in the Archean to those of today, and oceanic slabs may subduct to subarc depths where the release of volatiles generates arc magmatism. Although it is possible to produce "arc-like" geochemical signatures using other pressure-temperature-fluid melting conditions (e.g., van Hunen and Moyen, 2012; Bédard, 2006), the combination of geology, lithology, structural geometry, metamorphism, and geochemistry strongly argues for a subduction-related origin for arc-like magmas in Archean terranes (Windley et al., 2021; Sotiriou et al., 2022). For instance, the famous dome-and-basin province of Eastern Pilbara, is only 200 km×200 km in area, and would fit easily inside a Cordilleran batholith like the Sierra Nevada, with little or no discernable difference in rock types, structural patterns, or temporal evolution of different magmatic and structural stages (Kusky, 2021a). While the overall first-order similarities are striking, there are measurable apparent difference in thermal characteristics of metamorphism (Brown and Johnson, 2018, 2019) and warmer slab-top geotherms in many Archean orogens (Wang et al., 2018; Huang et al., 2020; Deng et al., 2022), while others are in the same range as cold, or Alpine-style geotherms of the modern Earth (Ning et al., 2022).

Deeper than the arc-magma generation depth is a "depth of speculation" about how deep Archean oceanic lithospheric slabs may have been subducted after they returned to the mantle, leaving no trace, except for perhaps geochemical signatures in the depleted mantle, isolated mineral xenocrysts (Kusky et al., 2021b, 2022), or stagnant slabs in the transition zone or along D″ at the core-mantle boundary (Torsvik et al., 2014; Wang Z S et al., 2018; Mitchell et al., 2021). In a few examples from the Kaapvaal, Slave and other cratons, samples of eclogites that are interpreted to be deeply subducted Archean oceanic lithosphere (Dongre et al., 2015; Aulbach et al., 2017; Richardson et al., 2001; Shirey and Richardson, 2011) have been returned to the surface, where oceanic slabs have underplated the overlying continents forming the sub-continental lithospheric mantle (SCLM) (e.g., Kusky, 1993; Wang Z S et al., 2022), and kimberlites have entrained samples of the underlying eclogite and peridotite as mantle xenoliths (Helmstaedt and Schulze, 1989; Richardson et al., 2001; Shirey and Richardson, 2011). Only recently have orogenic eclogite facies metamorphosed oceanic MORB been identified *in-situ* in Archean orogens (Ning et al., 2022).

Additional demonstration of the subduction of oceanic lithosphere from Archean cratons comes from geophysical data. Perhaps the best-documented example comes from combined deep seismic reflection-refraction and geological data on a transect across the Superior Province, that clearly documents three Archean paleo-subduction zones. The first two suture zones offset the Moho at depth, and are traced to the tectonic boundaries between the 2.7 Ga Quetico and Wawa terranes and the 3.2 Ga Winnipeg River terrane on the surface (Percival et al., 2012). The surface geology along these boundaries is marked by slabs of Archean oceanic crust, including greenstone belts with oceanic (MORB) pillow lavas, bordered by the metasedimentary, accretionary prism-like Quetico Domain, a clear indication of shallow level convergent margin tectonics to deep subduction in the Archean (Percival et al., 2004a, 2004b). Magnetotelluric (MT) data shows that these remnant slabs extend to circa 300 km depth (Kendall et al., 2002; Sol et al., 2002; White et al., 2003), and they show a strong S-wave anisotropy (Musacchio et al., 2004) that is only known to be exhibited by oceanic lithosphere. The third suture zone imaged seismically corresponds to the boundary between the 3.0 Ga North Caribou and 3.5 Ga Hudson Bay terranes. Similar integration of geophysical and geological data has revealed fossil, or paleo-subduction zones North China (Kusky et al., 2014a, Kusky, 2011), and Slave cratons (Kusky, 1989; Cook et al., 1999), in the Archean Yilgarn (Drummond et al., 2000; Kusky et al., 2018). We emphasize that the coincidence of these Archean mantle geophysical anomalies with surface suture zones is no different from comparable and well-accepted examples in the Proterozoic of Sweden, the Caledonides of Scotland, the Himalayas of India, and the Andes of South America (Prodehl and Mooney, 2012).

7. Conclusions

(1) Magmatic arcs are a central part of larger intra-oceanic and continental arc systems, which may also include a forearc accretionary prism, forearc region of rapidly generated forearc subduction initiation ophiolites, and forearc basins. This may be bordered on the backarc side by a backarc basin in extensional arcs where the slab is rolling back or retreating with respect to the arc, or a retro-arc basin where the arc is compressional or advancing with respect to the subducting slab. Arc systems are either accreted, subducted, or form on the margins of previously accreted arc systems, in accretionary orogens of all ages.

(2) Arc magmatism and deformation in arc-bearing accretionary orogens is recognized in all cratons of all ages, showing that the geological evidence for the operation of plate tectonics throughout the geological record is overwhelming. Continental crust would have mainly grown through arc magmatism through time.

(3) Not all arcs are accretionary, some experience large amounts of subduction erosion, or whole-scale subduction to be recycled into the mantle.

(4) TTG type steep-walled plutons form a dome-and-basin type pattern at 15–50 km depth in most arcs, which is remarkably similar to that found in the dome-and-basin patterns in some Archean terranes such as the Eastern Pilbara, Zimbabwe, and parts of the North China cratons.

(5) The rock types, petrogenesis, structural relationships between magmatism and tectonic accretion events in modern arcs are the same as those recorded in accreted arcs that form the bulk of the Precambrian continental lithosphere, preserved at different depths of erosion, and deformation.

(6) Calls for Archean geologists to demonstrate “globally linked networks of weak plate boundaries” to demonstrate the operation of plate tectonics in ancient terranes are not practical. This is because the preserved sizes of the oldest Archean crustal remnants are typically only tens to hundreds, and rarely approaching two thousand kilometers in dimensions, and are only the remnant preserved fragments of former planetary-wide systems. In addition, many of the tools, such as paleontologically-based paleogeography, GPS and paleomagnetism, used to demonstrate plate motions in the post-Mesozoic and extant plate mosaics are not applicable to the Archean, where geologists must rely on the rock record of structures, metamorphism, and rock types associated with plate boundary interactions. These records are strikingly similar from post-Mesozoic through early Archean times. Plate tectonics in the extant plate mosaic can only be proven locally when the geologist must rely on purely geological evidence. The geological record speaks, and states clearly, that plate tectonics is responsible for the structures and rocks produced on Earth through time.

(7) Based on the null hypothesis and plate tectonics approach, geologists can begin to understand better the differences in the types of magmatism, deformation, metamorphism, duration and rates of processes at convergent plate margins through time, because of the fundamental changes through time of heat production and flow from Earth and the rotational dynamics of the Earth-Sun-Moon system, and better understand links with changes in atmospheric and ocean chemistry, and the development of life on Earth.

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