Formation, Accretion and Reworking of Continents

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Abstract

Felsic continental crust is unique to the Earth in the solar system, but it still remains controversial regarding its formation, accretion and reworking. The plate tectonics theory has been significantly challenged in explaining the origin of continents as Archean continents rarely preserve hallmarks of plate tectonics. In contrast, growing evidence emerges to support mantle plume-derived oceanic plateau models as the models can reasonably explain the origin of bimodal volcanic assemblages and nearly coeval emplacement of tonalite-trondjhemite-granodiorite (TTG) rocks, presence of ~1600ºC komatiites and dominant dome structures, and lack of ultra-high-pressure rocks, paired metamorphic belts and ophiolites in Archean continents. Although plate tectonics seems to fail in explaining the origin of continents, it has been successfully applied to interpret the accretion or outgrowth of continents along subduction zones where new mafic crust is generated at the base of continental crust through partial melting of the mantle wedge with addition of H2O-dominant fluids from the subducted oceanic slabs, and partial melting of the juvenile mafic crust results in the formation of new felsic continental crust, leading to the outside accretion of continents. Subduction processes also cause the softening, thinning and recycling of continental lithosphere due to the vigorous infiltration of volatile-rich fluids and melts especially along weak layers or weak belts, leading to the widespread reworking and even destruction of continental lithosphere. Reworking of continents also occurs in continental interiors due to plume-lithosphere interactions, which, however, leads to much less degrees of lithospheric modification than subduction-induced craton destruction.

Formation, Accretion and Reworking of Continents

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Key Points:

- Archean continental nuclei were likely originated from oceanic plateaus formed by mantle plumes, not from island arcs via oceanic subduction
- Archean continental nuclei underwent accretion/growth at margins by the oceanic subduction, involving juvenile arc formation and accretion
- Continental reworking/destruction occurs mainly at subduction zones by water and volatile infiltration, with minor effects of mantle plumes

Abstract Felsic continental crust is unique to the Earth in the solar system, but it remains controversial regarding its formation, accretion and reworking. The plate tectonics theory has been significantly challenged in explaining the origin of continents as Archean continents rarely preserve hallmarks of plate tectonics. In contrast, growing evidence emerges to support mantle plume-derived oceanic plateau models as the models can reasonably explain the origin of bimodal volcanic assemblages and nearly coeval emplacement of tonalite-trondjhemite-granodiorite (TTG) rocks, presence of ~1600ºC komatiites and dominant dome structures, and lack of ultra-high-pressure rocks, paired metamorphic belts and ophiolites in Archean continents. Although plate tectonics seems to fail in explaining the origin of continents, it has been successfully applied to interpret the accretion or outgrowth of continents along subduction zones where new mafic crust is generated at the base of continental crust through partial melting of the mantle wedge with addition of H2O-dominant fluids from the subducted oceanic slabs, and partial melting of the juvenile mafic crust results in the formation of new felsic continental crust, leading to the outside accretion of continents. Subduction processes also cause the softening, thinning and recycling of continental lithosphere due to the vigorous infiltration of volatile-rich fluids and melts especially along weak layers or weak belts, leading to the widespread reworking and even destruction of continental lithosphere. Reworking of continents also occurs in continental interiors due to plume-lithosphere interactions, which, however, leads to much less degrees of lithospheric modification than subduction-induced craton destruction.

Plain language summary All solid planets in the solar system have a core-mantle-crust structure, but the Earth is a unique planet with felsic continental crust and plate tectonics. Controversy has long surrounded the formation, growth (accretion) and reworking of the continents. The plate tectonics theory has been significantly challenged in explaining the origin of continents as Archean (>2.5 billion years) continents rarely preserve hallmarks of plate tectonics. In contrast,

oceanic plateau models can well explain the origin of bimodal volcanic and plutonic assemblages, presence of ~1600ºC komatiites and dominant dome structures, and lack of ultra-high-pressure rocks, paired metamorphic belts and ophiolites in Archean continents. The accretion of continents can be successfully interpreted by plate tectonics along subduction zones via partial melting of the juvenile mafic crust which itself is generated at the base of continental crust through partial melting of the mantle wedge. Subduction can also cause softening, thinning and recycling of continental lithosphere due to the vigorous infiltration of volatile-rich fluids and melts especially along weak layers or weak belts, leading to widespread reworking and even destruction of continental lithosphere. Minor continental reworking and destruction also occur in continental interiors due to plume-lithosphere interactions.

1. Introduction

Although all solid planets and the Moon in the solar system have a crust of mostly mafic composition in their outermost shell (Campbell & Taylor, 1983), the Earth is unique in that it has a thick (~40 km on average) felsic continental crust as well as a thin (averagely ~7 km) mafic oceanic crust (Table 1). In many areas, the continental crust is much older than the oceanic crust, with the oldest continental rocks up to ~4.0 Ga and the oldest single zircon crystals up to 4.4 Ga (Harrison, 2009; Hawkesworth et al., 2010; Wilde et al., 2001), whereas the oldest oceanic crust is no more than 250 Ma (Roberts et al., 2015). In this sense, the continental crust preserves much more information about the Earth's geological history than the oceanic crust. Moreover, it is the continental crust that provides suitable places for human beings to live and most of the natural resources for human beings to utilize. Therefore, the issues of when, where and how the continental crust was formed, accreted and reworked or destroyed are very important in earth sciences. However, these issues have not been well resolved though the plate tectonic theory has been established for more than half a century.

Table 1

Comparisons of some physical properties and water content among continental and oceanic lithosphere and asthenosphere

<table>
<tr><th rowspan="3">Parameters</th><th colspan="3">Continental lithosphere</th><th rowspan="3">~100-200 Ma Oceanic lithosphere</th><th rowspan="3">Asthenosphere</th></tr>
<tr><th colspan="2">Craton</th><th rowspan="2">Phanerozoic</th></tr>
<tr><th>Archean</th><th>Proterozoic</th></tr>
<tr><td rowspan="2">Crustal thickness (km)[1,2]</td><td colspan="3">~40 (average)</td><td rowspan="2">~7</td><td rowspan="2">N/A</td></tr>
<tr><td colspan="2">~35 - ~50</td><td>~15 - ~80</td></tr>
<tr><td>Lithospheric thickness (km)[3,4,5]</td><td colspan="2">~150 - ~250</td><td>~50 - ~180</td><td>~100</td><td>N/A</td></tr>
<tr><td>Crustal density (kg/m^3)[1,6,7]</td><td colspan="3">2700 - 2800</td><td>2900 - 3000</td><td rowspan="2">~3390</td></tr>
<tr><td>[a]Root density (kg/m^3)[8,9,10]</td><td>~3300-3330</td><td>~3330-3370</td><td>~3340-3390</td><td>~3350-3380</td></tr>
<tr><td>Average heat flow (mW/m^2)[4,11]</td><td>~41</td><td>~48</td><td>~58</td><td>~48-54</td><td>N/A</td></tr>
<tr><td>Mg# [8,12,13,14]</td><td>92-94</td><td>91.5-92.5</td><td>90</td><td>~90-91[b]</td><td>89.3</td></tr>
<tr><td>Water content (wt ppm)[15,16,17]</td><td colspan="2"><24-100 (mostly <30-50)[c]
(mostly <10 in olivine)</td><td>50-100</td><td>50-100</td><td>50-200
(50±20 in olivine)</td></tr>
</table>

[a] at standard temperature and pressure condition. [b] In reference 14, only samples having both Mg# and olivine proportion data are considered. [c] In stable cratonic regions, relatively high-water contents in peridotites (olivine) were associated with localized metasomatism and deformation, and may not represent the overall feature of cratonic mantle root[17]. References include 1: Christensen and Mooney (1995); 2: Dick et al. (2003); 3: Artemieva (2009); 4: Jaupart and Mareschal (2015); 5: Priestley et al. (2019); 6: Carlson and Raskin (1984); 7: Frisch et al. (2011); 8: Griffin et al. (1999); 9: Lee et al. (2005); 10: Poudjom Djomani et al. (2001); 11: Stein and Stein (1992); 12: Boyd (1989); 13: Gaul et al. (2000); 14: Warren (2016); 15: Hirschmann (2006); 16: Hirth and Kohlstedt (1996); 17: Peslier et al. (2017).

Established in the middle 1960's and regarded as a revolution in earth sciences, the plate tectonics theory has been used to interpret various geological phenomena, processes and events that happened during Phanerozoic (0.54 Ga to present) time, and has also been successfully applied to the Proterozoic (2.5-0.54 Ga), but it has been significantly challenged in explaining the origin of Archean (>2.5 Ga) felsic continental crust, which consist mainly of orthogneisses and supracrustals (greenstones) that were metamorphosed from tonalite-trondhjemite-granodiorite (TTG) plutons and mafic-ultramafic volcanic and felsic rocks with minor sedimentary rocks, respectively. For example, plate tectonics can interpret Archean TTG rocks as the products of island arcs in subduction zones, such that the Archean high-pressure-type TTG rocks (adakites) were derived from the partial melting of subducted slabs, whereas the Archean low-pressure-type TTG rocks (equivalent to calc-alkaline granitoids) were derived from the partial melting of juvenile basaltic crust which itself was formed by the partial melting of the mantle wedge with addition of fluids released from the subducted slabs (Arndt, 2013; Condie, 2014; Martin, 1999; Martin et al., 2014; Wyman, 2013). However, such an island arc model is difficult to explain the bimodal volcanic assemblages from the Archean greenstone terranes where basaltic and ultramafic rocks are associated with dacite and rhyolite, without much andesite (Bédard, 2006; Hamilton, 1998, 2007; Van Kranendonk et al., 2007a, 2014; Zhao et al., 1998, 2001). Moreover, island arc models also fail in interpreting domiform structures and anticlockwise P-T paths involving isobaric cooling that characterize the deformation and metamorphism of Archean cratons (craton is an old and long-lived stable continent or continental lithosphere) (Lin & Beakhouse, 2013; Nijman et al., 2017; Van Kranendonk et al., 2007a, 2014; J. Zhang et al., 2014; Zhao et al., 1998, 2001, 2005). For these reasons, more and more researchers are now seeking other tectonic regimes beyond plate tectonics (e.g., mantle plumes, stagnant lid, subduction, etc) to interpret the origin of Archean continents.

Although it is still unclear whether felsic continents were originated from island arcs under the plate tectonics regime or from the non-plate tectonic settings such as oceanic plateaus formed by mantle plumes, it is evident that present-day's large continental blocks developed from the outside accretion of early Archean continental nucleuses or amalgamation of small pieces of early Archean continental nucleuses. However, it remains unknown or controversial regarding the issues of when and how Archean continental nucleuses were grown up to form present-day's large continents through external accretionary or internal collisional orogenic systems.

Growing data have been emerging to demonstrate that continental cratons have not only experienced outgrowth or accretion along their margins, but also been reworked in their interiors with lithospheric thinning, refertilization or destabilization. The fundamental reworking of a craton that results in the loss of its intrinsic tectonic stability has been termed as craton destruction or decratonization (Wu et al., 2008; Zhu et al., 2011). One of the best examples of cratons that have ever been destroyed is the North China Craton, of which its eastern part underwent intensive thinning and extension in association with extensive volcanic eruption, basin filling, granitic emplacement, and gold mineralization during the late Mesozoic time, indicating that the craton has undergone severe reworking or destruction (Wu et al., 2008, 2014; F. Y. Wu et al., 2019; Xu, 2001; Yang et al., 2018; Zhang, 2005; Zheng et al., 2018; Zhu et al., 2011, 2012a, 2012b, 2017, 2020; Zhu & Xu, 2019). However, what kind of geodynamic mechanism caused the reworking and destruction of a continental craton is still a subject of debate.

To resolve the above controversial issues regarding the formation, accretion and reworking of continental cratons, researchers have carried out extensive and integrative geological and geophysical investigations in the past two decades. In particular, the National Science Foundation of China (NSFC) set up a Major Research Program entitled "Destruction of the North China Craton" (2008-2015) that was aimed to determine what kind of geodynamic mechanisms caused the reworking and destruction of the North China Craton and whether the craton destruction is unique to the North China Craton or a common process for other cratons during Earth's evolution (F. Y. Wu et al., 2019). In the past decade, this NSFC Major Research Program supported 66 key projects on the thinning and destruction of the North China Craton which produced large amounts of new data and competing interpretations (Wu et al., 2014; F. Y. Wu et al., 2019; Zheng et al., 2018; Zhu et al., 2011, 2012a, 2012b, 2017, 2020; Zhu & Xu, 2019). These new data and those advancements made in other continents forms the justification for us to write this review in which we review current controversies on the origin and reworking of felsic continents, summarize the various lines of evidence for their mechanisms, and finally propose a new model for the formation, accretion and destruction of continents.

2. Formation of Continents

It is well known that in the solar system, Earth is a unique planet that has both plate tectonics and felsic continental crust (Hawkesworth et al., 2010, 2020; Rudnick, 1995; Sleep, 2000; Taylor & McLennan, 2008). This leads some researchers to speculate a cause-and-result relationship between plate tectonics and felsic continental crust (e.g., Hastie et al., 2016; Tang et al., 2016), but controversy has surrounded the issue of whether plate tectonics or continents first appeared on the Earth. Researchers regarding continents as a result of plate tectonics argue that plate tectonics had appeared before the formation of felsic continental crust on Earth, probably starting during Hadean or Eoarchean time, whereas felsic continental crust developed from island arcs under a plate tectonic regime (Arndt, 2013; de Wit, 1998; Furnes et al., 2009; Greber et al., 2017; Grosch & Slama, 2017; Harrison, 2009; Hastie et al., 2016; Jackson & Fyon, 1991; Kerrich & Polat, 2006; Kusky et al., 2013; Langford & Morin, 1976; Leat & Larter, 2003; Martin, 1999; Martin & Moyen, 2002; Martin et al., 2009, 2014; Nutman et al., 2015; Turner et al., 2014; Wyman, 2013). In contrast, another school of thoughts believes that felsic continental crust had formed long before the start of plate tectonics, as a result of non-plate tectonics, such as mantle plumes, sagduction driven by density difference, delamination of thickened crust, etc. (Bédard, 2006, 2018; Brown, 2006; Brown et al., 2020; Campbell et al., 1989; Cawood, 2020; Condie, 1975, 2005, 2014;

Hamilton, 1998, 2007, 2011, 2019; Hawkesworth et al., 2020; Hill et al., 1992; Larson, 1991; Rozel et al., 2017; Van Kranendonk et al., 2004, 2007a, 2007b, 2014; Zheng & Zhao, 2020). They also argue that it was just the existence of such low-density felsic continental crust that induced the subduction of high-density mafic crust (oceanic crust) beneath felsic continental crust, initializing the operation of plate tectonics (Bédard, 2006, 2018; Brown et al., 2020; Lin & Beakhouse, 2013; Nair & Chacko, 2008). Rey et al. (2014) demonstrated that because the Archean oceanic crust was thick and buoyant, early continents may have produced intra-lithospheric gravitational stresses large enough to drive their gravitational spreading to initiate subduction at their margins and to trigger episodes of subduction. Therefore, the existence of felsic continental crust may be a prerequisite for the initialization of plate tectonics (Bédard, 2006, 2018; Brown et al., 2020; Lin & Beakhouse, 2013; Nair & Chacko, 2008). No matter which model above is valid, researchers have reached a broad consensus that as the major component of Archean felsic continental crust, TTG rocks were derived from the partial melting of thickened mafic crust. The oldest mafic crust on the Earth is also called proto-crust or primitive crust (Grenville, 1922), which may have been formed by crystallization of magma ocean that covered the whole surface of our planet about 4.4 Ga ago (Elkins-Tanton, 2012). This primitive mafic crust most likely existed as a primitive oceanic crust after an initial ocean developed with large voluminous H_2O released out from the magma ocean to form the second atmosphere (Brown, 1949; Zahnle et al., 2010). It is considered that the earliest felsic continental crust was most likely to have developed from a thickened primitive oceanic crust through its partial melting. Modern oceanic settings include oceanic basins, mid-oceanic ridges, island arcs and oceanic plateaus, of which the former two settings have a normal oceanic crust that is too thin to form felsic continental crust via partial melting. Therefore, the first continent with felsic composition was most likely to have developed either from an island arc under a plate tectonic regime or from an oceanic plateau originated from mantle plumes. This depends on which model (plate tectonics or mantle plumes) can more reasonably explain the magmatic, metamorphic and structural features of Archean cratons.

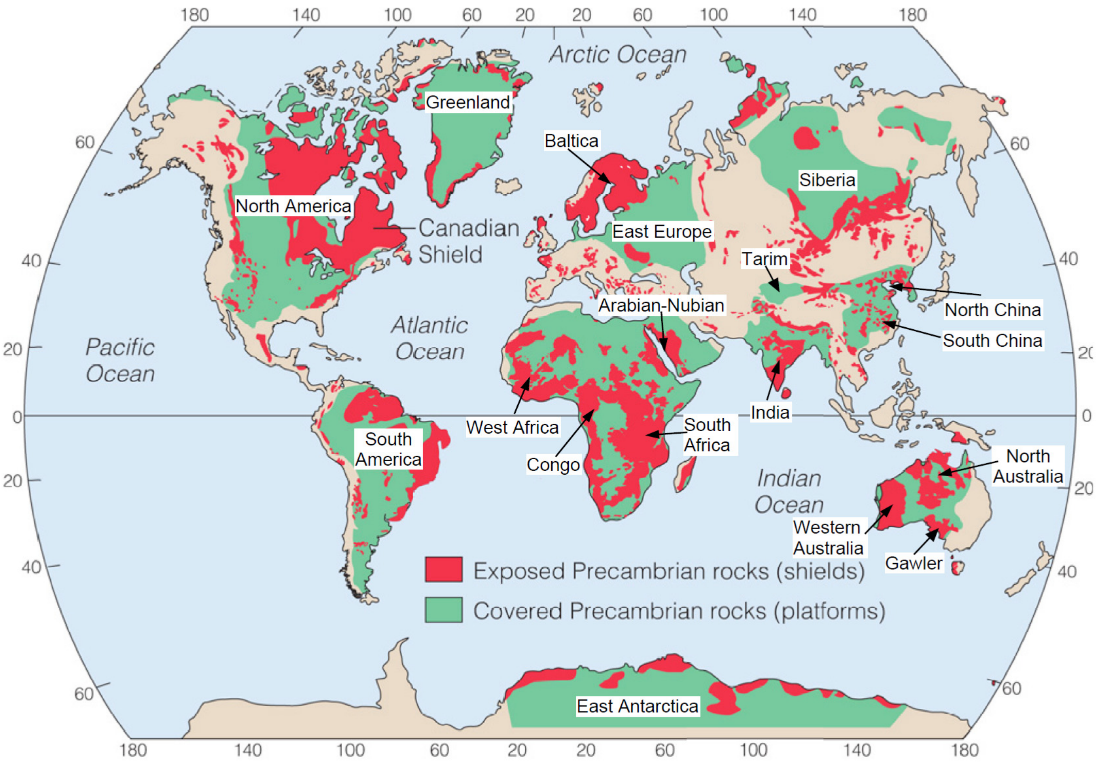

Figure 1. Spatial distribution of major cratons in the world (Wicander & Monroe, 2016).

2.1. Main Features of Archean cratons

Figure 1 shows the spatial distribution of major cratons in the world. Except the ~4.0 Ga Acasta TTG gneisses in North America (Bowring & Williams, 1999), most cratons formed during Archean eon (Arndt, 2013; Hawkesworth et al., 2010). The Eoarchean cratonic crust is represented by the 3.8-3.6 Ga Isua supracrustals and Amitsoq TTG gneisses in Greenland (Nutman et al., 2015), the Paleoarchean cratonic crust is represented by 3.6-3.2 Ga Barberton and Pilbara granite-greenstone terranes in South Africa and Western Australia, respectively, and the Mesoarchean and especially Neoarchean continental crust is widely exposed in all major cratons in the world. The whole-rock Nd and zircon Hf and O isotopic data for Archean felsic rocks from major cratons indicate the existence of Hadean primitive mafic crust (Arndt, 2013; Guitreau et al., 2014; Harrison, 2009; O'Neil & Carlson, 2017; Reimink et al., 2014; Wilde et al., 2001), but nearly whole of such a primitive crust has been destroyed probably either by the Late Heavy Bombardment Event (LHB) or by plate tectonics that appeared later.

Archean TTG rocks are similar to Phanerozoic calc-alkaline granitoids in that they are both high in Si, Ca, Sr, Ba and Na_2O/K_2O and low in heavy REE and depletions in Nb, Ta and Ti, but with temperature decreasing, the former goes to high Na, named the trondhjemitic trend (Tdj), whereas the latter goes to high K, named calc-alkaline trend (CA), indicating that Archean TTG may have been derived from the partial melting of low-K mafic (tholeiitic) rocks under H_2O-saturated conditions with garnet and/or rutile as residual/cumulus phases (Arth & Barker, 1976; Arth et al., 1978; Barker & Arth, 1976; Barker, 1979; Jahn et al., 1981). Garnet is assumed to be residual/cumulus phases in order to explain the strong light to heavy REE fractionation (high La/Yb ratios) of Archean TTG rocks, whereas rutile existing as residual/cumulus phases can explain the striking depletions of Nb, Ta and Ti in Archean TTG rocks(Arndt, 2013; Arth et al., 1978; Barker & Arth, 1976; Jahn et al., 1981; Martin et al., 2014; Moyen & Martin, 2012; Ryerson & Watson, 1987). As shown in Figure 2, most Archean TTG rocks have La/Yb and Sr/Y ratios remarkably higher than those of post-Archean calc-alkaline granitoids, indicating that the former are enriched in light REE (LREE) but depleted in heavy REE (HREE). This led researchers to have proposed that HREE-rich garnet must have existed as the dominant residual phase in the magmas to form TTG, implying that the partial melting of basaltic rocks to form Archean TTG must have occurred under high pressure conditions or in a thickened crust. The metamorphic mafic rocks under such high-pressure conditions are most likely garnet amphibolite and/or eclogite (Arndt, 2013; Arth & Barker, 1976; Arth et al., 1978; Barker & Arth, 1976; Barker, 1979; Condie, 2014; Martin et al., 2014; Moyen & Laurent, 2018). However, recent studies suggest that because of relatively high Mg in Archean crust, garnet may have occurred at a normal crustal thickness (~7 kbar), not necessary to require high-pressure conditions (>15 kbar) at a thickened crustal level (Johnson et al., 2017).

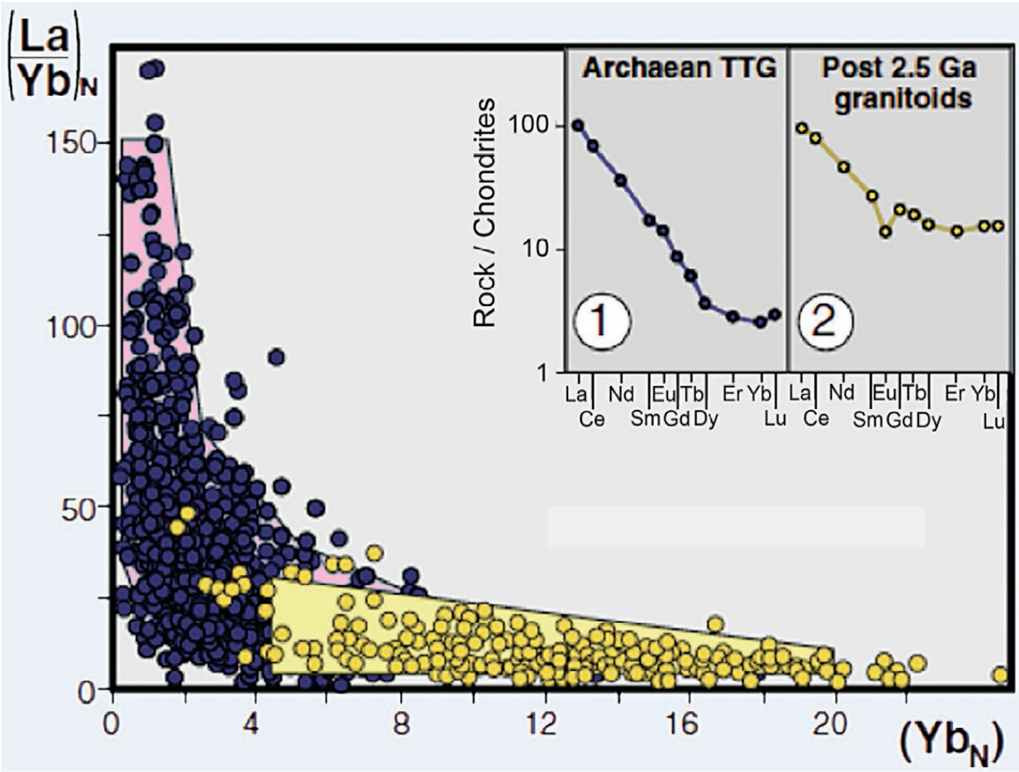

Figure 2. The La/Yb vs Yb diagram to distinguish Archean TTG from post-Archean granitoids (Martin, 1987; Martin & Moyen, 2002; Moyen & Martin, 2012). The subscript N indicates that the ratio or concentration is normalized to primitive mantle. The insets 1 and 2 illustrate the contrasting REE patterns of the two types of granitoid (Moyen & Martin, 2012).

Archean TTG rocks is divisible to low-Al_2O_3 type (Al_2O_2 <15% at SiO_2 = ~70%) and high-Al_2O_3 type (Al_2O_2 >15% at SiO_2 = ~70%). Generally, the low-Al_2O_3 type forms at relatively low-pressure conditions (10-12 kbar), also named low-pressure type, which is characterized by low La/Yb, Sr/Y, Na_2O and Sr and high Y, Yb, Ta and Nd, implying that the dominant residual phases in the magmas to form low-Al_2O_3 type TTG are plagioclase and pyroxene (Arndt, 2013; Condie, 2014; Martin et al., 2014; Moyen & Laurent, 2018). In contrast, the high-Al_2O_3 type TTG is considered to have formed under relatively high pressure conditions (>20kbar), so called high-pressure type, which is characterized by high La/Yb, Sr/Y, Na_2O and Sr and low Y, Yb, Ta and Nd, implying that the dominant residual phases in the magmas to form high-Al_2O_3 type TTG are garnet and rutile (Arndt, 2013; Condie, 2014; Martin et al., 2014; Moyen & Laurent, 2018). There are also Archean TTG rocks with chemical features between the low- and high-pressure types, called medium-pressure type TTG, which formed at 15-20 kbar, with amphibole, garnet and minor rutile (without plagioclase) as dominant residual phases during the partial melting of mafic crust to form TTG magmas (Moyen, 2011). Except minor adakites, Phanerozoic calc-alkaline granitoids have geochemical characteristics (e.g., La/Yb and Sr/Y values) similar to low-Al2O3 (or low-pressure) type Archean TTG rocks (Condie, 2014; Zhang & Zhai, 2012).

In the Archean supracrustals or greenstones, a common protolithic assemblage is bimodal volcanic rock, represented by ultramafic and mafic volcanic rocks (komatiite and tholeiite) at one end, and felsic volcanic rocks such as dacite, rhyolitic dacite and rhyolite at another end, with less or without andesite that is dominant in modern magmatic arcs (Hamilton, 1998, 2019). The komatiite and tholeiite in the Archean supracrustals or greenstones can be derived from the partial melting of mantle. Generally, komatiites in the Archean greenstones contains >18% MgO, which requires the degree of partial melting of mantle as high as 40-60%, indicating that the partial melting must have occurred at high temperatures, which have been estimated at more than 1600°C (Arndt et al., 2008; Campbell et al., 1989; Nisbet et al., 1993). On the other hand, felsic volcanic rocks of Archean greenstones are generally considered to have been derived from the partial melting of mafic lower crust heated by ultramafic to mafic magmas (Arndt, 2013; Campbell et al., 1989; Hamilton, 1998, 2019; Nisbet et al., 1993).

2.1.2. Structural patterns of Archean cratons

There are striking differences in structural styles between Archean cratons and post-Archean orogens, with the former dominated by dome-and-keel structures that mainly resulted from vertical tectonics, whereas the latter are dominated by linear structural belts that resulted from oceanic subduction and continental collision under plate tectonics regime. The dome-and-keel structures in the Archean granite-greenstone terranes consist of circular to elliptical (in plane) domes of granitoid (TTG) gneisses surrounded by trough- shaped keels, containing supracrustal (greenstone) cover rocks (Figure 3), of which the gneiss domes develop from the large-scale diapirism of voluminous TTG magmas and the keels form when the surrounding high-density ultramafic and mafic greenstones sink down, both of which reflect vertical motions as evidenced by vertical lineations in Archean granite-greenstone terranes (Figure 3). Such dome-and-keel structures characterize the major structural patterns of major Archean cratons, including the Barberton and Zimbabwe cratons in South Africa (Figure 3a; Anhaeusser & Wilson, 1981; Van Hinsbergen et al., 2011; Van Kranendonk et al., 2014), the Yilgara and Pilbara cratons in Western Australia (Figure 3b; Collins et al., 1998; Hallberg & Glikson, 1981; Nijman et al., 2017; Van Kranendonk et al., 2004), the Superior craton in North America (e.g., Lin, 2005; Lin & Beakhouse, 2013; J. Zhang et al., 2014), the Eastern Block in the North China craton (Zhao et al., 2001; Zhao, 2014), etc. In these old cratons, the structure in the boundaries between the gneiss domes and the supracrustals (greenstones) is characterized by vertical shearing and stretching, as reflected by vertical stretching lineations or L-tectonites, indicating vertical motions, which contrast with the horizontal motions of plate tectonics as reflected by large-scale thrusting, multiple phases of folding, ductile shear zones, sheath folds, horizontal mineral stretching lineations.

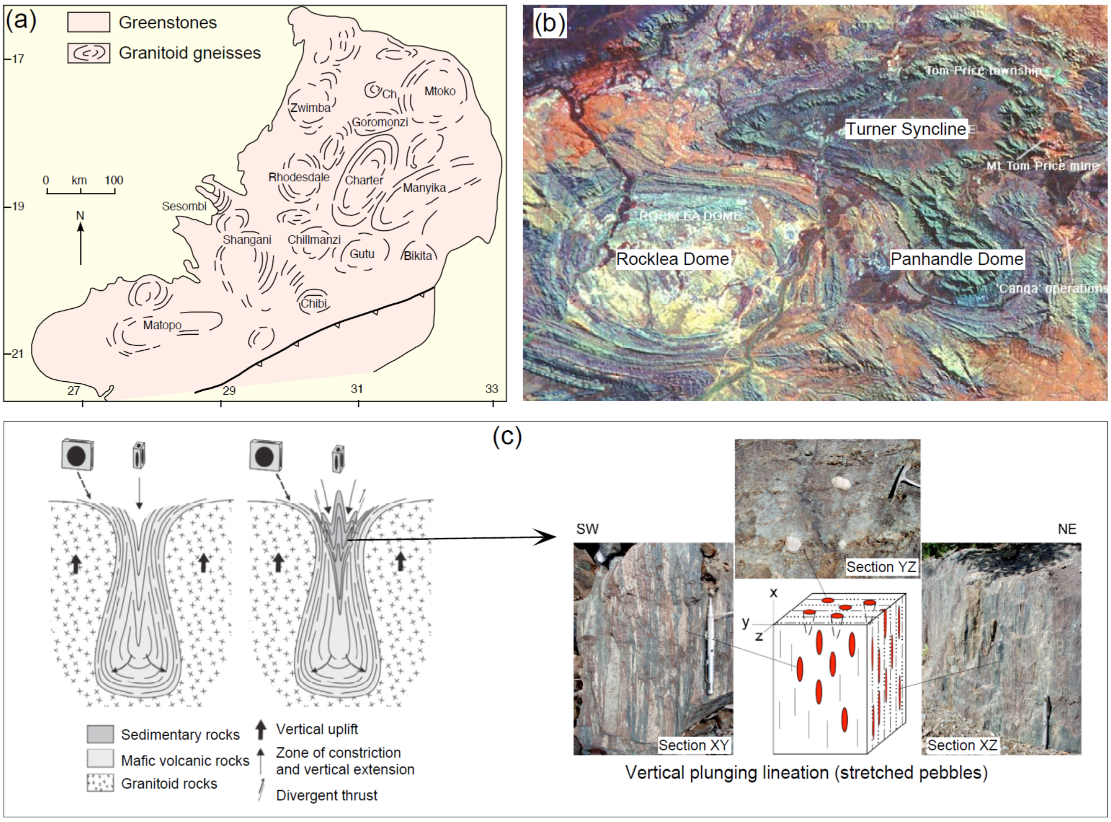

Figure 3. Dome-and-keel structures of Archean cratons. Neoarchean domes in the Zimbabuwei Craton (Anhaeusser & Wilson, 1981); b) Archean domes (Zhao et al., 2001); c) Vertical stretch lineations or L-tectonites developing at boundaries between gneiss domes and supracrustals (J. Zhang et al., 2014).

2.1.3. Metamorphism of Archean Cratons

The metamorphism of Archean cratons is generally characterized by (1) large-scale regional metamorphism, in contrast with Proterozoic and Phanerozoic metamorphism that is limited to linear structural belts (orogens); 2) low- to medium-pressure greenschist facies, amphibolite facies and granulite facies metamorphism that happened at a normal geothermal gradient range (10-30 °C/km); 3) absence of high-pressure blueschist facies and high-pressure or ultrahigh pressure eclogite facies metamorphism; and (4) metamorphic P-T paths dominated by anticlockwise type involving isobaric cooling (Ge et al., 2003; Halpin & Reid, 2016; Jayananda et al., 2000; Kamber et al., 1996; Kramers et al., 2001; Maas & Henry, 2002; Mvondo et al., 2017; Percival, 1994; Raith et al., 1990, 1999; Rollinson, 1989; Sandiford, 1985; Tsunogae et al., 1992, 1999; Zhao et al., 1998, 2001, 2005; Zulbati & Harley, 2007), though clockwise P-T paths were also reported for some Archean terrains. As shown in Figure 4, for example, the metamorphic evolution of Neoarchean terrains in the Eastern and Western blocks in the North China craton, no matter whether they are low-grade granite-greenstone belts or high-grade gneiss terrains, is characterized by anticlockwise P-T paths mostly involving isobaric cooling (IBC). Similar anticlockwise P-T paths have also reported in other Archean cratons, such as the Kaapvaal craton (Kamber et al., 1996; Kramers et al., 2001; Rollinson, 1989; Tsunogae et al., 1992), Gawler craton (Halpin & Reid, 2016), East Antarctica (Sandiford, 1985; Tsunogae et al., 1999; Zulbati & Harley, 2007), Superior craton (Percival, 1994), Wyoming craton (Maas & Henry, 2002), Slave craton (Mvondo et al., 2017), Southern India craton (Raith et al., 1990, 1999).

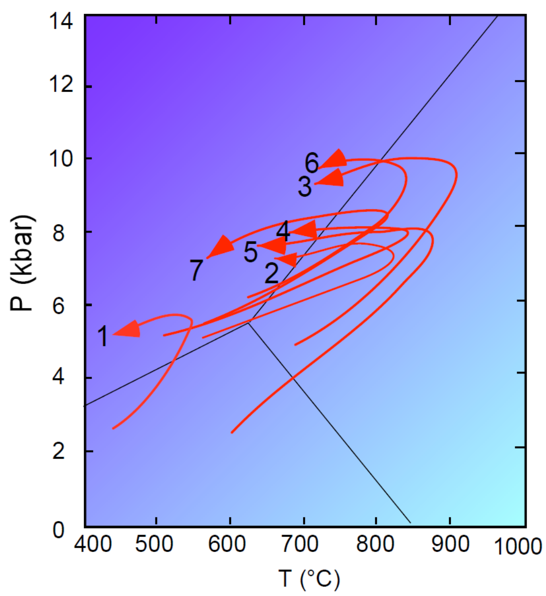

Figure 4. Anticlockwise P-T paths of end-Neoarchean metamorphism in the Eastern Block of the North China Craton (Zhao et al., 2001). 1-Western Shandong; 2-Eastern Hebei; 2, Western Liaoning (Jianping); 4-Northern Liaoning; 5-Eastern Shandong; 6-Miyun; 7-Southern Jilin.

The IBC-type anticlockwise P-T paths is generally considered to reflect metamorphism related to the intrusion and underplating of large voluminous mantle-derived magmas that may occur at the root of an arc or back-arc (Bohlen, 1991; Brown, 2006), intracontinental rifting zones (Sandiford & Powell, 1986), and mantle plumes or hot-spots (Jayananda et al., 2000; Zhao et al., 1998), in which mantle magmas heat continental crust, leading to temperature increasing but meanwhile the mantle magmas may erupt on the surface as continental flood basalts or intrude as sills into the crust, leading to thickening of the metamorphic crust. Therefore, at the early prograde stage, both temperature and pressure increase, but once the mantle magmas cease to rise, the heating stops and the metamorphic crust experiences cooling under a nearly constant pressure condition since there is no much change in the thickness of the crust, leading to an IBC-type anticlockwise P-T path (Zhao et al., 1998, 2001, 2005).

2.2. Island Arc Model for the Formation of Archean Continents

2.2.1. Major Evidence

Whether felsic continental crust originated from island arcs under plate tectonic regime or from some pre-plate tectonics (non-plate tectonics) depends on which model can satisfactorily explain the rocks and their geochemical, structural and metamorphic features as discussed above. Major lines of evidence supporting the island arc model for the formation of Archean felsic continental crust can be summarized as follows:

(1) The metamorphosed igneous plutons in Archean cratons have rock associations similar to those in the roots of Phanerozoic magmatic arcs, with both consisting of Si-rich and Na-rich felsic granitoid (TTG) suites, though Phanerozoic magmatic arcs have more diorite, quartz diorite and monzogranite.

(2) Subduction-related adakites in Phanerozoic magmatic arcs are geochemically similar to Archean high-pressure (or high-Al_2O_3) TTG. They are both characterized by relatively high La/Yb, Sr/Y, Al_2O_3, Na_2O and Sr but relatively low Y, Yb, Ta, Nb and Ti, implying garnet and rutile existing as the residual phases during the partial melting of a mafic crust or as the cumulus phases during the crystallization of TTG magmas. In contrast, the subduction-related calc-alkaline granitoids in Phanerozoic arcs are geochemically similar to Archean low-pressure (or low-Al_2O_3) TTG. They are both characterized by relatively low La/Yb, Sr/Y, Al_2O_3, Na_2O and Sr but relatively high Y, Yb, Ta, Nb and Ti, suggesting plagioclase and pyroxene existing as the residual phases during the partial melting of a mafic crust or as the cumulus phases during the crystallization of TTG magmas. Generally, subduction-related adakites are considered to have been derived from the partial melting of subducted oceanic slabs, whereas calc-alkaline granitoids were derived from the partial melting of juvenile mafic rocks that themselves were derived from the partial melting of hydrated mantle wedge above the subducted slab. Therefore, the island arc model established on plate tectonics can well explain the origins of Archean high- and low-pressure TTG rocks.

(3) Geochemical and petrological experimental data have demonstrated that the source rocks to form Archean TTG rocks via partial melting should be garnet-bearing and/or rutile-beating amphibolites or eclogites, both of which are abundant rocks in subduction zones, though no eclogites have been found in Archean terrains and some previously reported “Archean eclogites” have recently been confirmed to be Paleoproterozoic eclogites (Li et al., 2017).

(4) Although gneiss domes are the dominant structural patterns of Archean cratons, intervening between gneiss domes are some linear structural belts

(keel structures) that underwent intense compressive deformation, similar to the structural styles of Phanerozoic orogens under a plate tectonic regime.

(5) Despite absence of high-pressure and ultrahigh pressure rocks in Archean craton, medium-high pressure rocks, and especially sillimanite-bearing pelitic gneisses/granulites are ubiquitous in Archean high-grade terrains, with metamorphic pressures as high as 8-10 kbar. The sedimentary protoliths of pelitic gneisses/granulites were shales or mudstones that were deposited in the basins, and thus without subduction under plate tectonic regimes, it's hard to imagine how these sedimentary rocks could be brought down to the lower crust depth where they experienced upper amphibolite- to granulite-facies metamorphism to form pelitic gneisses/granulites (Zhao et al., 2012).

(6) Although most cratons underwent metamorphism characterized by anticlockwise P-T paths involving isobaric cooling (IBC), clockwise P-T paths involving isothermal decompression were reported for a few Archean cratons (e.g., Hölttä & Paavola, 2000; Taylor et al., 2010; Valli et al., 2004). Such clockwise P-T paths involving isothermal decompression are generally considered to reflect metamorphism related to subduction or continent-continent collision (Brown, 1993; Harley, 1988, 1989; Zhao et al., 2000).

(7) The last but most important evidence supporting island arc model for the formation of felsic continental crust is that the partial melting of low-potassium tholeiitic rocks to form Archean TTG magmas needs sufficient water (H_2O), which favors subduction zone settings, not mantle plumes/hot-spots (Arndt, 2013; Martin et al., 2014).

2.2.2. Island Arc Model for the Origin of Continents

Magmatic arcs developing in the subduction zones can be divided into three types. The first type is island arc, also called the intra-oceanic arc or Mariana-type arc that develops from the subduction of an oceanic lithosphere below another oceanic lithosphere (ocean-ocean subduction). Island arcs have a simpler crustal structure and composition than arcs built on continental crust, because magmas erupted in intra-oceanic arcs are not contaminated by ancient sialic crust and their compositions more accurately record partial melting processes in the mantle wedge. The second type is continental margin arc, also called the Andean-type arc that develops from the subduction of an oceanic lithosphere beneath a continent (ocean-continent subduction). The third type is a transitional one between the first and second types, also called the Japanese-type arc whose early stage was a continental margin arc (Andean-type arc), but with the further development of a back-arc basin into a baby ocean, the magmatic arc was separated from the mainland and thus the arc is getting more and more similar to the Mariana-type arc, like today's Japan island in which most of old continental material has been assimilated. Obviously, the second and third type arcs are not suitable for discussing the origin of felsic continental crust since they developed on the existence of felsic continents. Thus, only island arc that develops from ocean-ocean subduction is a possible site to generate felsic continents (Zhao & Zhang, 2021).

An island arc generally experiences processes from an immature arc to a mature arc (Stern, 2011). The former represents the initial stage of ocean-ocean subduction, during which the dehydration of subducted oceanic slabs leads to the partial melting of the mantle wedge to form basaltic magmas that rise and penetrate a thin oceanic crust to form basalts on the ocean floor, rather than intermediate to acid volcanic rocks such as andesite, dacite and rhyolite, since the magmas would not change much in composition during their ascent as the magmas and oceanic crust have similar mafic compositions. As a consequence, a large-scale felsic continental crust would not develop from an immature arc, though minor amounts of adakitic rocks may have developed due to the partial melting of the subducted slabs.

With arc magmas on-going, the basaltic arc is getting thicker and thicker so that the root of the arc can experience partial melting to form felsic magmas due to heating from the mantle magmas. The felsic magmas form andesite, dacite and rhyolite when they erupt on the surface and the diorite, quartz diorite, tonalite, trondhjemite, granodiorite suite when they crystallize underground, which are equivalent to low-pressure TTG rocks, whereas the partial melting of the subducted oceanic slabs would lead to the formation of high-pressure TTG rocks. The whole process leads to a transition from an immature arc to a mature arc (Figure 5a). With the further subduction and final closure of the subducted oceanic basins between island arcs, the arcs are growing larger and larger (Figure 5b) and finally collide each other to form a large-scale continent (Figure 5c), which is a generalized island arc model for the formation of felsic continents during Archean eon (Arndt, 2013; Leat & Larter, 2003; Wicander & Monroe, 2016).

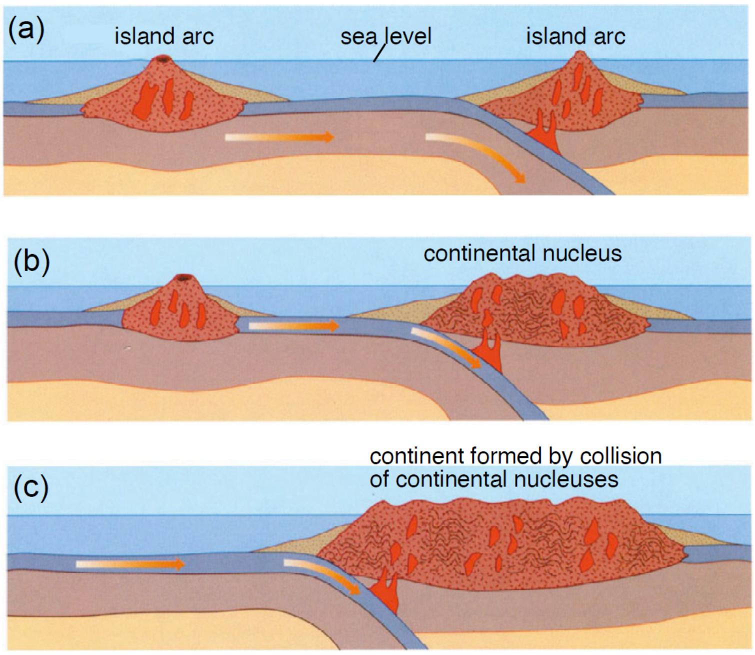

Figure 5. Cartoons showing how continents originated from island arcs under a plate tectonic regime (Wicander & Monroe, 2016). (a) A felsic island arc forms by subduction of oceanic lithosphere and partial melting of basaltic oceanic crust; (b) The island arc with continental composition in (a) collides with a previously formed felsic island arc, thereby

forming a continental core； (c) The process occurs again when the island arc in (b) collides with the evolving continent, thereby forming a craton, the nucleus of a continent.

The above island arc model can reasonably explain the formation of Archean low-pressure (low-Al_2O_3) and high-pressure (high-Al_2O_3) TTG rocks, but typical high-pressure or medium-pressure TTG rocks have not been found in modern island arcs like Mariana island arc in which most igneous plutons are calc-alkaline granitoids and adakitic rocks are mainly volcanic rocks, not plutons (Condie, 2014). In addition, many other geological factors are not supportive for the island arcs model for the origin of felsic continents, which will be discussed in the last section.

2.3. Oceanic Plateau Model for the Formation of Archean Continents

Although the island arc model can well explain the origin of Archean TTG rocks including both the high- and low-pressure types, it cannot satisfactorily explain the bimodal volcanic rocks from the Archean granite-greenstone terranes, which consist predominantly of komatiite-tholeiite at one end and dacite-rhyolite at another end, with less or absence of andesite that is dominant in modern magmatic arcs (Hamilton, 2019). Bimodal volcanic rocks are generally considered to have erupted in an extensional setting, rarely in a compressive tectonic environment.

2.3.1. Mantle Plume Model for the Ultramafic-Mafic Volcanic Rocks from the Archean Greenstones

Extensional tectonic settings allowing large voluminous mantle magmas to rise include intracontinental rifting zones, back-arc basins and mantle plumes/hot-spots. Generally, greenstones in Archean cratons are not limited to some linear structural belts, but irregularly surround granitoids (TTG) gneiss domes (Figure 3a-b), which, together with lack of typical rifting-type sediments in Archean greenstone sequences, precludes the possibility of an intracontinental rifting setting for volcanic protoliths of the Archean greenstones. The possibility of back-arc basins for volcanic protoliths of the Archean greenstones is also very little because few typical subduction zones that should be associated with back-arc basins have been found in Archean cratons (Bédard, 2006, 2018; Hamilton, 1998, 2007, 2019; Stern, 2007, 2008). Classical lithotectonic elements that characterize a subduction zone include high- and ultrahigh-pressure blueschists/eclogites, ophiolites, paired metamorphic belts, etc., of which paired metamorphic belts are regarded as a hallmark for a subduction zone but have not been identified from Archean cratons (Brown, 2006; Brown et al., 2020). Therefore, the most possible tectonic setting for the volcanic protoliths of the Archean greenstones is related to mantle plumes (Campbell et al., 1989; Hill et al., 1992; Kent et al., 1996; Larson, 1991).

A mantle plume is a cylindrical thermal upwelling of large voluminous low-density material that originates deep in the mantle, either from the D" layer located in the core-mantle boundary at a depth of about 2,900 km or from the 670 km discontinuity at the base of the upper mantle (Campbell, 2005; Condie, 2001; Ernst & Buchan, 2003). The concept was originally proposed by W. Jason Morgan in 1971 on the basis of the Hawaii hot spots (Wilson, 1963) to explain the age-progressive chains of volcanic islands that stretch across the ocean basins (Morgan, 1971). A mantle plume is generally composed of a huge head and a narrow tail that is connected to the deep mantle (Campbell et al., 1989; Campbell & Griffiths, 1990, 1992; Hill et al., 1992; Kent et al., 1996; Larson, 1991; Peters & Day, 2017). During the ascent of a mantle plume, the head would trap large amounts of mantle material to enlarge itself and when it reaches the base of lithosphere, it becomes a flattened mushroom shape and experiences decompressional partial melting to form basaltic magmas. The basaltic magmas erupt on the surface within a very short period (< 1 million years), forming continental flood basalts on continents and oceanic plateaus with diameters ranging from 1000-2000 km on ocean floors (Abbott, 1996). When oceanic plateaus rise to the surface from the floors of the ocean basins, they are also called oceanic islands. In current oceans under plate tectonic regimes, oceanic islands originated from a single mantle plume can form an age-progressive chain of volcanic islands, like the Hawaiian Emperor seamount chain, which can be utilized to estimate the speed and direction of plate motions.

The oceanic island basalts (OIB) are distinctively different from mid ocean ridge basalts (MORB) and island arc basalts (IAB) in that the former erupt at high temperature in areas far away from the mid ocean ridges and subduction zones, on a scale of several millions km^3, geochemically characterized by enrichments in LILE, LREE and high Sr^{87}/Sr^{86}, Nd^{143}/Nd^{144}, Pb^{207}/Pb^{204}, and He^3/H^4 ratios (Farley et al., 1992; Hart et al., 1992). Owing to these features, more and more researchers favor an oceanic plateau model for the origin of felsic continental crust because the model can well explain the formation of both bimodal volcanic rocks and TTG-dominant granitoids in the Archean granite-greenstone terranes (Campbell et al., 1989; Campbell & Griffiths, 1990, 1992; Hill et al., 1992; Hill, 1993; Kent et al., 1996; Larson, 1991).

The major volcanic rocks of Archean greenstones are mafic tholeiites and ultramafic komatiites. Some researchers once interpreted Archean komatiites as the early cumulates crystallizing from the tholeiitic magmas, and based on high Mg number of komatiites, they estimated that the temperature of Archean mantle was 200-300°C higher than that of today's mantle (McKenzie, 1984; Sleep & Windley, 1982). However, later studies indicated that the mantle temperature did not reduce so much during the past 3.5 billion years, only dropping 97°C (Abbott & Mooney, 1995; Galer, 1991), and Archean komatiites may not have been the early cumulates for the following considerations:

(1) Nearly all Archean komatiites have MgO contents more than 18%, whereas tholeiites from the Archean greenstones have MgO contents no more than

14%, demonstrating a significant compositional gap between komatiites and tholeiites from the Archean greenstones;

(2) Komatiites and tholeiites from the Archean greenstones exhibit different isotopic compositions (Abbott, 1996; Tomlinson & Condie, 2001);

(3) Experiments have precluded possibility that large amounts of olivine were cumulated from tholeiitic magmas at the early stage (e.g., Drummond, 1988);

(4) As mentioned early, the rock-forming temperature of komatiites is as high as 1600 °C or above, much higher than the rock-forming temperature (1200-1400°C) of tholeiites.

Considering that the rock-forming temperatures and geochemical compositions of komatiites and associated tholeiites from Archean greenstones are similar, respectively, to those of the tail and head of a mantle plume derived from the D" layer of the core-mantle boundary, Campbell et al. (1989) proposed that the Archean komatiites represented the melting products of the tail of a mantle plume, whereas the tholeiites from the Archean greenstones were derived from the decompressive partial melting of the huge head of a mantle plume when it reached the base of lithosphere. In fact, Condie (1975) had proposed a similar model as early as in the 1970s, though at that time he did not establish such a cause-and-result link between the tail and head of a mantle plume and komatiites and tholeiites from Archean greenstones, respectively. So far, this model has been accepted by more and more researchers (Abbott, 1996; Bédard, 2006; Brown et al., 2020; Campbell & Hill, 1988; Condie, 1994, 1997; Desrochers et al., 1993; Hill et al., 1992; Hill, 1993; Rey et al., 2003; Tomlinson & Condie, 2001).

2.3.2. Oceanic Plateau Model for the Origin of Continents

The above successful application of the mantle plume model to the formation of komatiites and tholeiites from Archean greenstones itself is not enough to enable the model to explain the origin of felsic continents because the Archean greenstones may not be part of continental crust, most likely representing Archean oceanic crust. What represents the Archean felsic continental crust are Archean TTG-dominated granitoid rocks that make up the 60-70% of total exposure of Archean cratons (Abbott, 1996; Arndt, 2013; Zhao & Zhang, 2021). Therefore, whether the mantle plume model is successful in explaining the origin of Archean continents depends on whether the model can reasonably explain the formation of Archean TTG-dominant granitoid rocks or not.

As mentioned early, geochemical data and experiments indicate that like other calc-alkaline granitoids, TTG rocks can also be derived from the partial melting of hydrated basaltic rocks. Thus, the origin of Archean TTG rocks depends on two issues: (1) where and how were the basaltic protoliths of Archean TTG rocks extracted from mantle and (2) how was the basaltic crust extracted from the mantle transformed to Archean TTG? For the first issue, mantle plumes may be the best candidate as discussed above. Considering that TTG rocks make up more than 60-70% of Archean cratons and their geochemical differences from Phanerozoic calc-alkaline granitoids (see Section 2.1.1), we think that the transformation of basaltic crust into Archean TTG-dominant felsic continental crust requires the following three conditions:

(1) Existence of large igneous provinces (LIP) with basaltic rocks at least three times as much as the amount of Archean TTG. Petrological experiments demonstrate that, to produce Archean TTG, the degree of partial melting of basaltic rocks must be lower than 30% because, when the partial melting degree is higher than 30%, garnet would be melt and enter into the TTG magmas, which would significantly increase HREE and thus reduce La/Yb ratios, making the magmas geochemically different from Archean TTG rocks that are characterized by high LREE and low HREE, as reflected by high La/Yb ratios. This means that, to make the partial melting degree lower than 30%, one portion of TTG at least needs three portions of basaltic protoliths.

(2) Existence of a very thick basaltic crust to make the partial melting occur in the garnet-stable field (see discussion in Section 2.1.1.).

(3) The basaltic protoliths must have been hydrated so that aqueous partial melting of basaltic crust to produce TTG magmas occurs under H_2O-saturation conditions. Experiments have shown that partial melting of basaltic crust would be going extremely slowly under H_2O-free (dry) conditions.

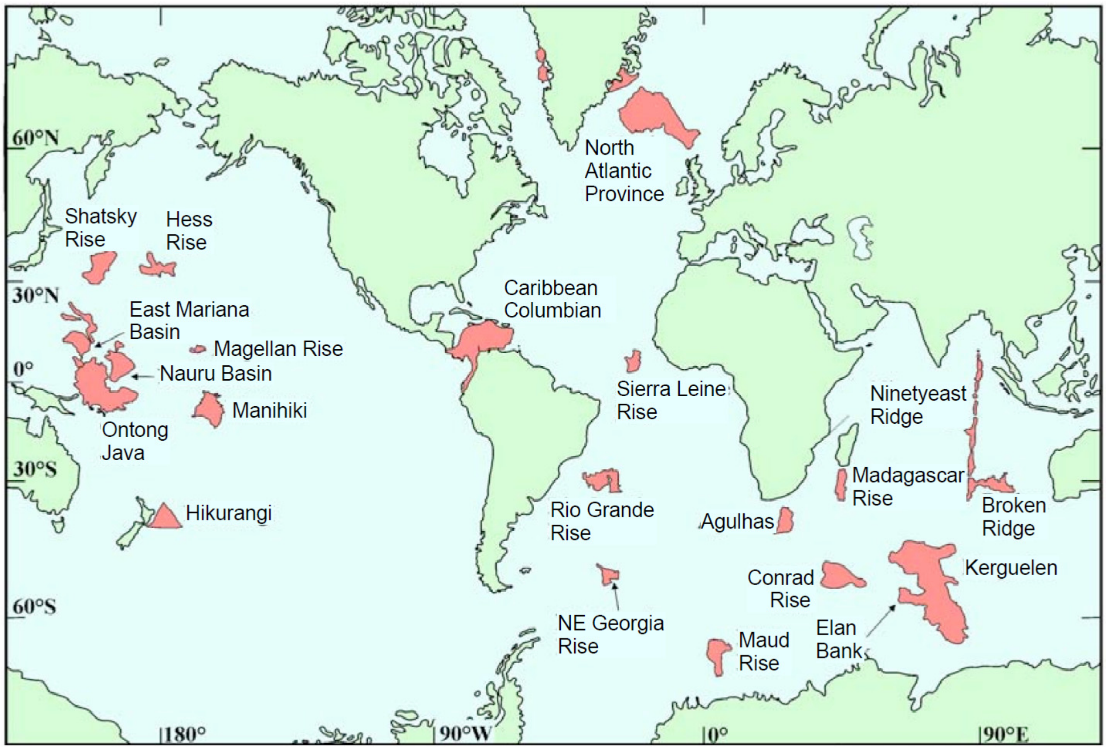

Figure 6. A Map showing all major oceanic plateaus formed within the last 150 Ma (Kerr, 2015).

Oceanic plateaus formed by mantle plumes seem to meet the first and second conditions discussed above. As shown in Figure 6, most of modern oceanic plateaus occur as large igneous provinces (LIP) that make up large areas of over-thickened oceanic crust (Coffin & Eldholm, 1994; Ernst, 2014; Kronenke, 1974). For example, the Ontong-Java oceanic plateau is about 500 km wide and 3000 km long, occupying an area of ~1,900,000 km^2, and the Ice Island and Kerguelen oceanic plateaus also make up more than 1,500,000 km^2 (Figure 6; Abbott, 1996; Coffin & Eldholm, 1994). Such huge volumes of basaltic oceanic crust are enough to meet the above first condition that requires the existence of large igneous provinces (LIP) with basaltic rocks at least three times as much as Archean TTG.

The second condition depends on the thickness of an oceanic plateau. Although the thickness of the Ontong-Java oceanic plateau was estimated at 35-42 km based on seismic wave velocity and rock density (Furumoto et al., 1976), most of other oceanic plateaus have a thickness of 20-30 km (Abbott & Mooney, 1995), which seems to be insufficient to meet the second condition that requires a very thick oceanic crust so that the partial melting to form high-pressure (high-Al_2O_3) TTG magmas can occur in the garnet-stable field. However, there is a broad consensus that the Archean mantle was hotter than today's mantle, which allowed more oceanic crust to have been produced, and thus the Archean oceanic crust must have been thicker than modern oceanic crust (Abbott & Mooney, 1995). Abbot and Mooney (1995) utilized thermal modeling to estimate that the temperature of Archean mantle was at least 93ºC higher than that of the current mantle. Combining this with seismic wave velocity and rock density, Abbot and Mooney (1995) estimated that the thickness of Archean oceanic plateaus was around 40 km. This was supported by Kent et al. (1996) who, using the adiabatic partial melting model of McKenzie and Bickle (1988) at about 1600 ºC, estimated the average thickness of Archean oceanic plateaus at about 43 km, which is close to the minimum crustal thickness that allows the partial melting of basaltic rocks to occur in the garnet field in order to produce medium- and high-pressure type TTG magmas with high La/Yb (Arndt, 2013; Arth et al., 1978; Barker & Arth, 1976; Barker, 1979; Condie, 2014; Martin et al., 2014; Moyen & Laurent, 2018). On the other hand, Kent et al. (1996)'s modeling demonstrated that the Archean crust of basaltic oceanic plateaus are komatiitic with MgO as high as ~19%, and in such high MgO rocks, metamorphic reactions to form garnet may have occurred at 7-8 kbar (Brown et al., 2020; Johnson et al., 2017), which is consistent with the ubiquitous presence of a medium-pressure granulite-facies assemblage (garnet + orthopyroxene + clinopyroxene + plagioclase) in Archean high-grade gneiss terrains (see discussion in Section 1.3), whereas high-pressure granulite-facies assemblage (garnet + clinopyroxene + plagioclase + quartz) or eclogite-facies assemblage (garnet + omphacite) is very rare or absent from Archean terranes. Zellmer et al. (2012) also confirmed that the garnet-stable field can extend to the lower crustal level where mafic crust experiences partial melting to form adakitic or high-pressure type TTG magmas. In addition, based on the results of their partial melting experiment, Qian and Hermann (2013) demonstrated that the most appropriate depth of a mafic lower crust for partial melting to form adakitic and TTG magmas is at 30-40 km, whereas a depth of more than 45-50 km is unfavorable. More recently, Smithies et al. (2019) also argue that there is no direct evidence supporting that the Archean TTG rocks were derived from the partial melting of mafic crust under high-pressure conditions. Taken together, oceanic plateaus can meet the requirement for a thick basaltic crust where the partial melting of basaltic rocks to form Archean TTG rocks occur in the garnet-stable field.

However, oceanic plateaus formed by mantle plumes seem to be difficult to provide enough H_2O for the aqueous partial melting of basaltic rocks to form TTG magmas as the above third condition required. The oxygen isotopic compositions of zircons indicate that most Archean TTG gneisses have $\delta^{18}O$ values ranging between 5.5-6.5‰, with some reaching to 7-9 ‰ in local areas (e.g., Superior craton), much higher than those $\delta^{18}O$ value for the mantle (5.3‰), implying that the basaltic source rocks underwent hydration from the surface before the partial melting (Condie, 2014). This means that the partial melting of basaltic crust to form Archean TTG magmas must have occurred in the upper part that was interacted with or hydrothermally altered by the surface water before the partial melting (Condie, 2014). Such an interpretation favors the partial melting of descent subducting slabs, but disfavors the oceanic plateau model in which the partial melting must have occurred at the dried base of the oceanic plateau, which is regarded as the fatal defect of the oceanic plateau model (Arndt, 2013). Even so, compared with the island arc model for the origin of felsic continents, oceanic plateau models have been applied to interpreted the origins of Archean granite-greenstone terranes in major cratonic blocks in the world (e.g., Abbott & Mooney, 1995; Abbott, 1996; Bédard, 2006, 2018; Brown et al., 2020; Campbell & Hill, 1988; Campbell et al., 1989; Campbell & Griffiths, 1990, 1992; Condie, 1975, 1994, 1997, 2014; Johnson et al., 2017; Moyen, 2011; Moyen & Laurent, 2018; Puchtel et al., 1998; Sanislav et al., 2018; Smithies et al., 2009; Tomlinson & Condie, 2001; Van Kranendonk et al., 2004, 2007a, 2007b, 2010, 2011, 2014; Whalen et al., 2002). Further supports for the oceanic plateau model for the origin of felsic continents have come from discoveries of TTG and relevant felsic rocks in modern oceanic plateaus (Hastie et al., 2010; Ponthus et al., 2020; White et al., 1999; Willbold et al., 2009).

As shown in Figure 7, Van Kranendonk et al. (2007a) applied the oceanic plateau model to interpret the formation of Paleoarchean (3.6-3.2 Ga) granite-greenstone terranes in East Pilbara, Western Australia. In the model, Van Kranendonk et al. (2007a) proposed that the huge head of a mantle plume underwent decompressional melting when it reached the base of the lithosphere, and the melts rose and erupted on the ocean floor, forming a basaltic oceanic plateau. Some komatiites or komatiitic basalts may have formed in the oceanic plateau when the tail of the mantle plume reached the base of the lithosphere. Meanwhile, the heat from the mantle plume caused multiple phases of partial melting of the pre-existing mafic crust and/or the base of the oceanic plateau to form 3.53-3.24 Ga granitoid rocks, leading to the formation of a felsic continental craton (Figure 7a) and a sub-continental residual mantle core (Figure 7b).

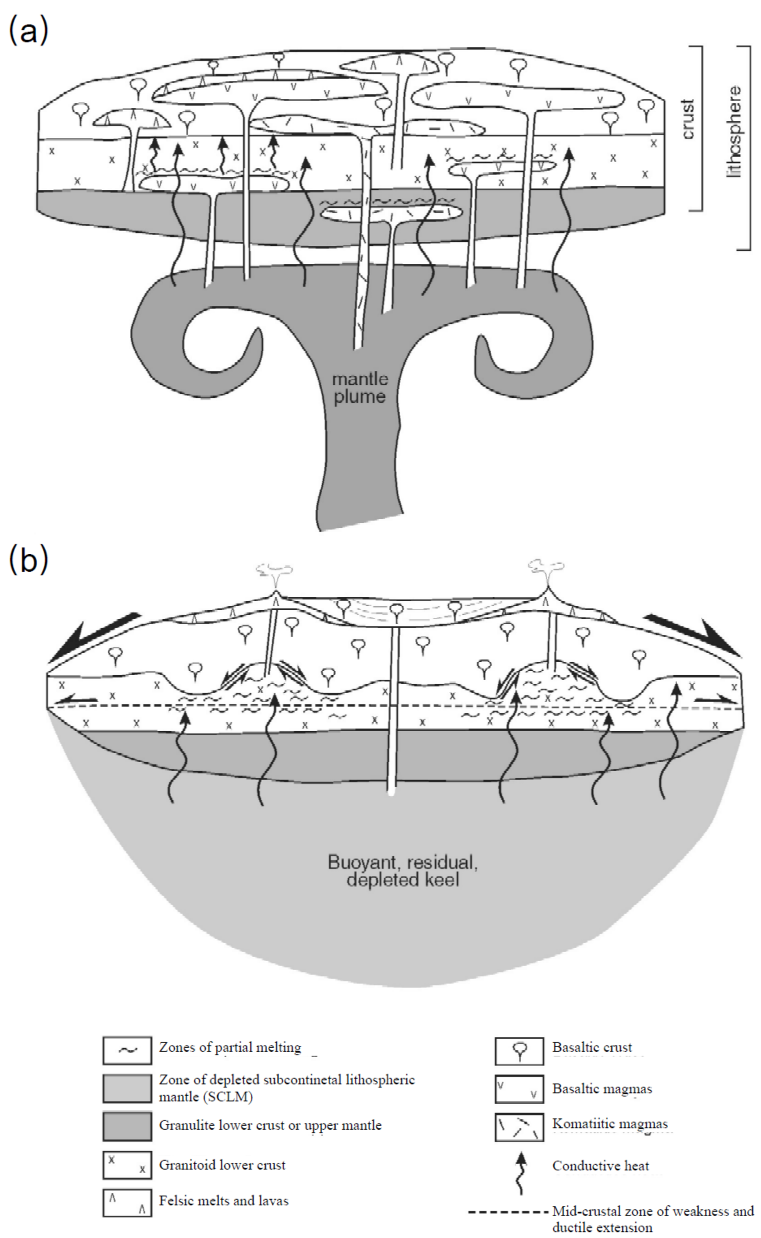

Figure 7. A schematic model showing the development of the East Pilbara granite-greenstone terrane from a thick oceanic plateau that formed by mantle plumes (Van Kranendonk et al., 2007a; Van Kranendonk, 2010).

As one of modified oceanic plateau models for the formation of felsic continents, Bédard (2006) proposed a catalytic delamination-driven model for coupled genesis of Archaean crust and sub-continental lithospheric mantle (Figure 8). In his model, a large mantle plume releases melts that construct a thick volcanic crust (e.g., oceanic plateau). The underplating magmas of a mantle plume cause melting at the base of the basaltic crust to form the first generation of tonalitic magmas with complementary eclogitic to pyroxenitic restites (Figure 8a). The dense eclogitic to pyroxenitic restites delaminate into the deep mantle, triggering the ascent of mantle diapirs, which reach the base of the mafic crust and causes the partial melting to form tonaltic magmas and eclogitic restites again (Figure 8b-c). Similarly, the newly formed eclogitic restites delaminate into the deep mantle and trigger the ascent of mantle diapirs, leading to the partial melting of basaltic crust to form TTG magmas (Figure 8d). A similar model was also proposed by Smithies et al. (2009).

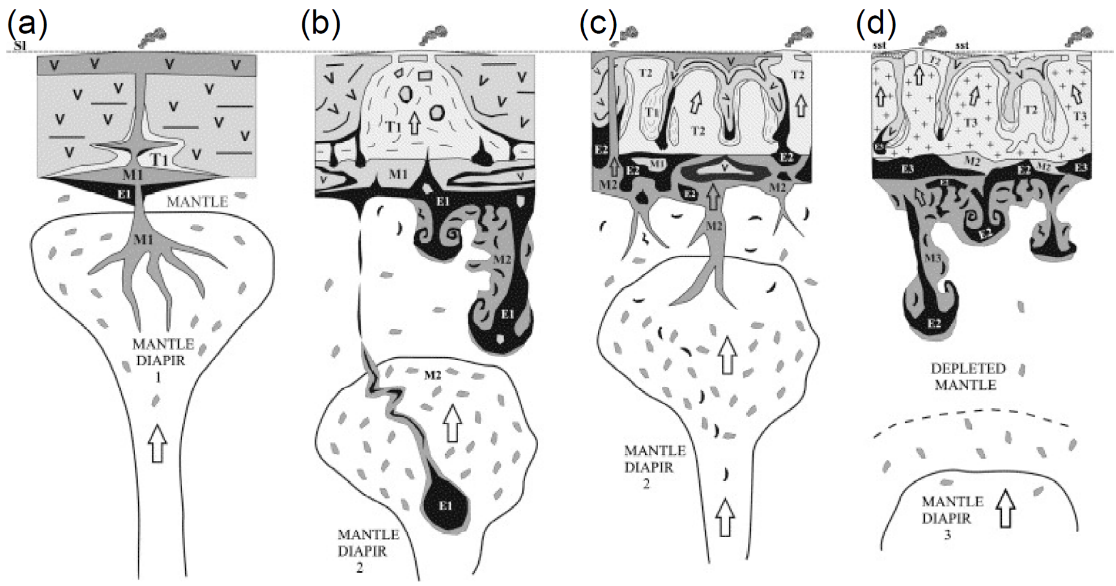

Figure 8. Cartoons illustrating the delamination-driven tectonomagmatic model (Bédard, 2006)

3. Accretion of Continents: A Global Perspective

As discussed above, there is still no broad consensus regarding the origin of felsic continents with one school of thought favoring an island arc model (Arndt, 2013 and references therein), whereas others advocate an oceanic plateau model (Van Kranendonk et al., 2007a and references therein). No matter which model is correct, these two-end-member models only account for the formation of Archean continental nucleuses, whereas present-day's continents must been more comploicated and have developed through the accretion (growth) of these Archean continental nucleuses. Available data have demonstrated that the subduction and associated continental accretion under modern-style plate tectonic regimes play an important role in the development from Archean continental nucleuses to present continents (Hoffman, 1988). In this section, we will discuss where, when and how Archean continental nucleuses underwent accretion to form the present-day's continents.

3.1. Accretion of North American continent

Among all the continents, the North America preserves the best and most complete geological records of how small pieces of cratonic blocks gradually grow to the current continent, mainly by oceanic subduction, arc accretion and continental collision processes. The North American continent is mainly composed of the Precambrian North American Craton in the center, the early Paleozoic Appalachia Orogen in the east and the Meso-Cenozoic Cordillera Orogen in the west (Figure 9). The North American Craton (together with Greenland also known as the Laurentia) is one of the oldest and largest cratons in the world. It was amalgamated by several Archean provinces through Proterozoic collisional and accretionary orogenic belts (Canil, 2008; Hoffman, 1988).

3.1.1. Archean Provinces

The Archean provinces, including the Slave, Rae, Hearne, Wyoming, Superior and Nain, are clustered in the northern two thirds of the North America continent and underlie most of the Canadian shield. Each province has an Archean basement comprising a granite-greenstone terrain, overlain by erosional remnants of early Proterozoic sedimentary cover of platformal facies (Hoffman, 1988).

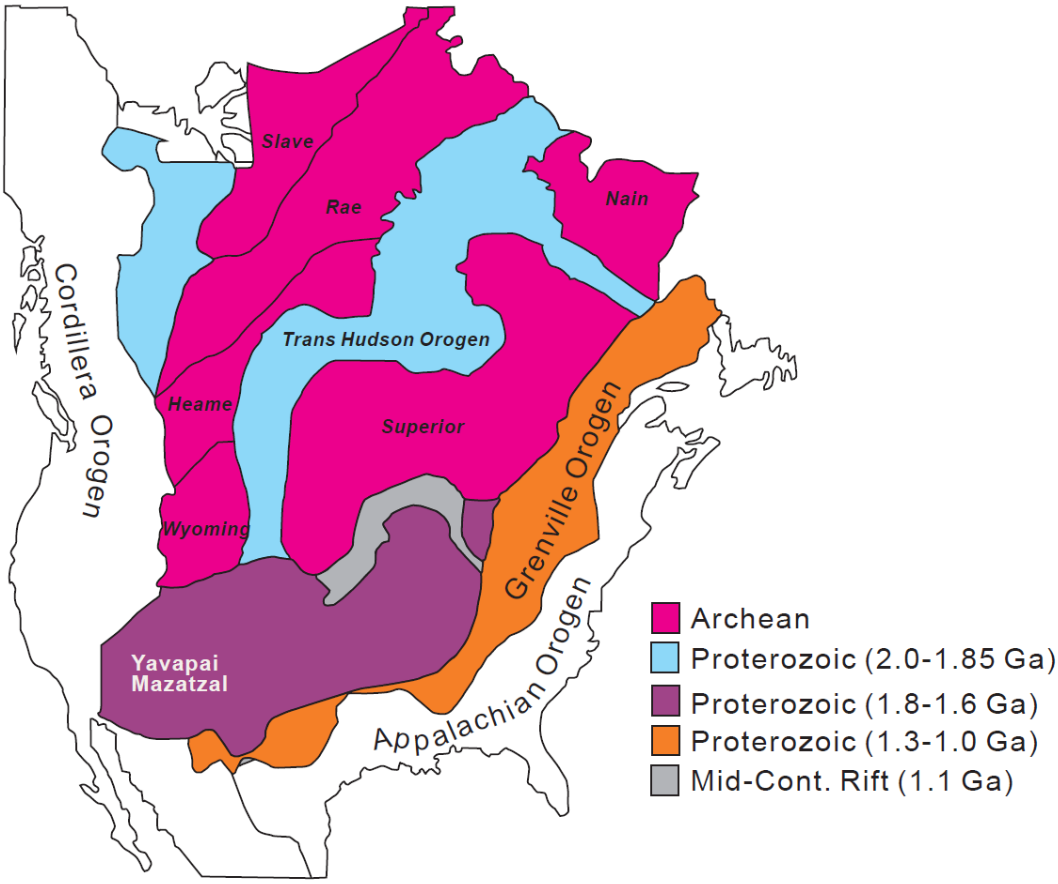

Figure 9. Tectonic sketch map of the North American continent, showing the major Precambrian cratons and Phanerozoic orogens, modified from Hoffmann (1988) and Canil (2008).

The Slave province is an Archean granite-greenstone terrane in the northwestern part of the North America. The main part of the province has an elliptical shape and comprises ca. 172,500 km^2. Its history spans the interval from 3.48 Ga to 2.6-2.5 Ga, with most of the volcanic and sedimentary rocks formed between 2.7 and 2.65 Ga by oceanic subduction and arc accretion (Kusky, 1989; Kusky et al., 2013). Although initially described as a uniform granite-greenstone terrain, the Slave province is more recently viewed as a tectonic collage consisting of a Mesoarchean micro-continental nucleus that was enlarged by the addition of Neoarchean juvenile terranes. The Slave province is subdivided into 5 tectonic domains, at least four of which may be viewed as separate tectonic terranes (Helmstaedt, 2009).

The Rae province can be divided into an ancient (>3.0 Ga) southwestern component and a more juvenile (<3.0 Ga) northeastern component. Evidence for episodic reworking of the evolved southwestern component is recorded in the Sm-Nd and U-Pb signatures of younger Archean and Paleoproterozoic intrusive rocks and sediments (Hartlaub et al., 2005).

To the south of the Rae province, the Hearne province has much in common with the Rae province. It contains gneiss as old as 3.48 Ga and submarine mafic-intermediate-felsic volcanics and associated graywackes of ca. 2.7 Ga.

The Wyoming province was a >100 000 km^2 Mesoarchean craton that was modified by Neoarchean magmatism and tectonism and Proterozoic extension and rifting. Isotopic evidence for widespread old crust and the local occurrence of early and Mesoarchean rocks suggest that the northern and central portions of the Wyoming province were part of a Mesoarchean craton, a southward-younging series of Neoarchean volcanic arcs were subsequently accreted to the continental core of Paleoarchean to Mesoarchean age (Chamberlain et al., 2003). The Beartooth Mountains of the northern Wyoming province contain a suite of ca. 2.8 Ga calc-alkaline and tholeiitic andesite, indicative of a Neoarchean convergent margin process (Mueller & Wooden, 1988). A 2.68-2.67 Ga active margin with accretionary prism, fore-arc basin, magmatic arc and back arc basin was suggested to be developed in the western Wyoming province (Frost et al., 2006). A major episode of Neoarchean crustal growth involved both the lateral accretion of juvenile terranes and the intrusion of arc magmas formed from mantle-derived and (or) juvenile crustal sources occurred within the Wyoming province, and was driven by geologic processes very similar to modern plate tectonics (Frost et al., 2006).

The Superior province is the world's largest Archean craton and provides information on both the nature and scale of ancient process. Five discrete accretionary events assembled fragments of continental and oceanic crust into a coherent Superior craton by 2.6 Ga. They exhibit similar sequences of events at ~10 million-year intervals: cessation of arc magmatism, early deformation, synorogenic sedimentation, sanukitoid magmatism, bulk shortening, regional metamorphism, late transpression, orogenic gold localization, emplacement of crust-derived granites, and postorogenic cooling (Percival et al., 2006). Similarly, many studies suggest that the geological evolution of the southwestern Superior province can be best explained by convergent plate margin processes (Daigneault et al., 2002; Polat et al., 2017). Neoarchean S-type granites were emplaced along some terrane boundary zones in the western Superior province and are interpreted as a marker of Archean collision zones (Yang et al., 2019).

The Nain (or North Atlantic) province lies in the northeastern part of the Laurentia. It is composed of the Saglek block in the north and Hopedalc block in the south, which experienced distinct thermotectonic histories until Neoarchean. The Saglek block comprises amphibolite- to granulite-facies orthogneisses and supracrustal rocks ranging from Paleoarchean (>3.7 Ga) to Neoarchean (ca. 2.7 Ga); the Hopedalc block comprises crust ranging from 3.3 to 2.8 Ga with mainly greenschist- to amphibolite-facies metamorphism. These two blocks juxtaposed to form a single, stable cratonic mass during the Neoarchean (ca. 2.55 Ga) (Connelly & Ryan, 1996).

3.1.2. Paleoproterozoic orogenic Belts

The above Archean provinces record a long history of Archean tectonism, after which many underwent 2.45-2.1 Ga rifting, as shown by thick miogeoclinal successions on the edges of Archean cratons (Corrigan et al., 2005). The collision between the Archean provinces generated a sequence of Paleoproterozoic orogenic belts, which led to assembly of a continental core of a scale compatible with modern continents (Whitmeyer & Karlstrom, 2007).

The Thelon orogen resulted from a dextral-oblique collision between the Slave and Rae provinces, followed by the indentation of the Rae hinterland by the Slave foreland province (Hoffman, 1988). The main magmatism related to this collision event took place at 2.02-1.91 Ga.

Between the Rae and Hearne provinces lies the ~2800-km-long Snowbird Tectonic Zone. It is one of the most controversial tectonic features of the North American Craton. It was originally proposed as a Paleoproterozoic continental suture (Gibb & Walcott, 1971), but was later interpreted as an intracontinental shear zone at 2.6 Ga with limited Paleoproterozoic reworking (Hanmer et al., 1994). Field investigations revealed a major granulite facies mylonite belt that records deformation and metamorphism at several intervals in the Archean and early Proterozoic. Early assemblages are dominated by granulite grade gneisses that were deformed in a major left-lateral strike-slip shear zone at ca. 3.2 Ga. The major episode of deformation occurred at ca. 2.6 Ga, again at granulite facies in a dominantly right-lateral strike-slip shear zone (Kopf, 2002). Eclogites occurs in East Athabasca area in association with a variety of high-pressure granulites that record a complex metamorphic history from 2.6 to 1.9 Ga. The timing of peak eclogite facies metamorphism was constrained at 1.9 Ga (Baldwin et al., 2004). Metamorphic and geochronological data revealed that a 1.9 Ga medium- to high-pressure belt extends along most of this tectonic zone, and it was regarded as a collisional belt marking a pre-1.865 Ga phase of the Hudsonian orogeny involving microcontinent accretion that was fundamental to the growth of Laurentia (Baldwin et al., 2003; Berman et al., 2007).

The 500-km-wide Trans-Hudson orogen forms a convex to the northwest, bounded by the Hearne and Superior provinces (Figure 9). The orogen extends southward to the area between the Wyoming and Superior provinces as far as south Dakota, and to the northeast the main part of the orogen passes beneath the Paleozoic Hudson Bay Basin (Hoffman, 1988). The Trans-Hudson orogen is distinguished by the largest exposure of juvenile, arc-derived Proterozoic crust in the Canadian shield. Isotopic and structural data revealed that the Trans-Hudson orogen resulted from an arc-continent collision followed by terminal continent-continent collision at 1.83-1.80 Ga (Bickford et al., 1990; Corrigan et al., 2005). It has four principle tectonic domains: (1) a narrow eastern foreland; 2) a broad collage of imbricated magmatic arc, marginal basin and collisional basin rock which structurally overlies Archean basement; 3) an Andean-type continental margin batholith; and (4) a broad, reworked northwestern hinterland (Lewry et al., 1994; Lucas et al., 1993). Eclogite rocks with comparative pressure-temperature conditions to that from the Himalaya orogen have recently been identified within the Trans-Hudson orogen, which imply that modern-day plate tectonic processes featuring deep continental subduction occurred at least 1830 million years ago (Weller & St-Onge, 2017). The Trans-Hudson orogen represents the 1.85-1.78 Ga amalgamation of the Hearne, Wyoming, and Superior provinces into a cratonic core of the North American continent (Weller & St-Onge, 2017), and is regarded as a prototype of modern accretionary processes (Corrigan et al., 2009).

In the northeastern part of the North America continent, there are two major Paleoproterozoic orogenic belts, namely the Torngate orogen and New Quebec orogen, respectively. The Torngate orogen is the result of oblique collision between the Nain province in the north and the Rae province in the south (Hoffman, 1988). It is regarded as a narrow doubly divergent collisional orogen (Rivers et al., 1996). Subduction occurred below the Nain province from ca. 1.91-1.87 Ga, and collision between the cratons may have occurred between 1.87 and 1.84 Ga (Scott, 1998). Subduction-related calc-alkaline magmatism of mafic to intermediate composition has been identified along the Torngate orogen (Rawlings-Hinchey et al., 2003). The

New Quebec orogen is bounded to the southwest by the flexurally arched Archean foreland of the Superior province and to northeast by the allochthonous Archean hinterland of the Rae province (Hoffman, 1988). It is a Paleoproterozoic fold and thrust belt active from 2.17 to 1.77 Ga and records the collisional history between the Rae and Superior provinces (Machado et al., 1997).

3.1.3. Paleoproterozoic Accretion

The 2.0-1.8 Ga collision-dominated orogens welded the Archean provinces into the core of the North American continent. At the same time and subsequently, island arcs with mainly juvenile crust developed on the margins of the Archean provinces, which led to outboard growth of the continent.

The Wopmay orogen evolved on the active western margin of the Slave province (Hoffman, 1988). Remnants of three compositionally diverse magmatic arcs (1.94-1.90 Ga, 1.89-1.88 Ga, and 1.87-1.84 Ga) are preserved in the Wopmay orogen, and Nd isotopic data reveal the involvement of 2.0-2.4 Ga crust (Bowring & Podosek, 1989; Cook, 2011). There are no exposures of Archean crust west of the Wopmay fault.

The Penokean orogen records the Paleoproterozoic growth of the southern margin of the Superior province. It comprises a northern domain of sedimentary and minor tholeiitic volcanic rocks deposited on Archean basement of the Superior province and a southern domain composed mainly of Paleoproterozoic volcanic and plutonic rocks belonging to island-arc suites (Hoffman, 1988). The main orogenic events occurred roughly coeval with the Trans-Hudson deformation, which deformed and metamorphosed Archean basement and Paleoproterozoic superacrustal rocks of the Superior province along the southern margin of Laurentia (Whitmeyer & Karlstrom, 2007). The Penokean orogeny began at ca. 1.88 Ga when an oceanic arc (Pembine-Wausau terrane) collided with the southern margin of the Superior province marking the end of a period of south-directed subduction. The docking of the buoyant craton to the arc at ca. 1.87 Ga resulted in a subduction jump to the south and development of back-arc extension both in the initial arc and adjacent craton margin to the north, which was followed by the accretion of the Marshfield terrane to the Superior province at ca. 1.85 Ga resulting in the development of foreland basin (Schulz & Cannon, 2007). Nd isotopic data for igneous rocks revealed the Penokean events involved major growth of new crust from the mantle, with only limited recycled Archean materials (Barovichi et al., 1989).

By ca. 1.80 Ga, the Archean provinces (Slave, Rae, Hearne, Wyoming, Nain and Superior) had amalgamated to form the north-central part of the North American craton by a rapid succession of microcontinent collisions (Hoffman, 1988). The orogens involved are characterized by reworking of Archean crust combined with addition of ca. 2.0-1.8 Ga juvenile crust. Subsequent late Paleoproterozoic (1.80-1.60 Ga) juvenile crust was added mainly to the south of the united Archean provinces. These include the Central Plains, and Yavapai and Mazatzal orogens (Figure 9).

The Central Plains orogen is ca. 1000 km long and ca. 500 km wide which truncates the southern margin of the Superior province and its marginal Paleoproterozoic orogens, the Trans-Hudson (1.95-1.85 Ga) and the Penokean (1.90-1.83 Ga). It is composed of metamorphic and granitoid rocks in the range 1.80-1.63 Ga (Sims & Petermar, 1986).

The Yavapai orogen records the accretion of dominantly juvenile arc crust from 1.80 to 1.70 Ga, including outboard development and collisions of arcs from 1.78 to 1.72 Ga, and an orogenic peak at 1.71-1.68 Ga that resulted in a progressive amalgamation of Yavapai crust to Laurentia (Whitmeyer & Karlstrom, 2007). Accretion and associated deformation took place during several pulses within a long orogenic progression. The oldest rocks include 1.80-1.75 Ga granite-greenstone associations that consist of metabasalt, metaandesite, metarhyorite, and associated volcanic-sedimentary rocks intruded by calc-alkaline granitoids. A near continuum of deformation between 1.78 and 1.68 Ga are recorded in the Yavapai orogen. Volcanic and granitoid batholith from 1.80 to 1.75 Ga are interpreted to record subduction-related accretion and outboard collision to form the 1.80-1.75 Ga arcs. Southeast migration of granitoid magmatism from 1.80 to 1.75 Ga are explained in terms of subduction flip and from south dipping in the Penokean orogeny to north dipping in the Yavapai orogeny (Holm et al., 2005).

The Mazatzal orogen is marked by exposures of Paleoproterozoic rocks younger than 1.70 Ga in New Mexico, southwestern Arizona, and northern Sonora, Mexica. This orogen mainly consists of juvenile arc-related rocks and associated sedimentary rocks. Nd and Pb isotopic data reveal crust that has a slightly younger mantle derivation model age (1.8-1.7 Ga) than the adjacent Yavapai orogen (Aleinikoff et al., 1993). Geochronological data suggest that the Mazatzal orogeny occurred from 1.68 to 1.64 Ga (Karlstrom & Bowring, 1988), followed by post-orogenic magmatism and extension from 1.63 to 1.62 Ga (Amato et al., 2008).

3.1.4. Mesoproterozoic Accretion-Collision - Grenville Orogen

The Grenville orogen is a globally distributed plate convergent event over a protracted period from ca. 1.3 to 0.9 Ga took place in Laurentia and on many other continents, and culminated in continent-continent collisions that facilitated the final assembly of Rodinia. The Grenville orogen in Laurentia is a NE-trending orogenic system truncating the Archean Superior province and Paleoproterozoic Yavapai, Penokean, and New Quebec orogens. It records an 800 m.y. history of accretionary orogenesis along an east- and southeast-facing, predominantly convergent margin. The exposed segment of the orogen was derived largely from reworked Archean to Paleoproterozoic Laurentian crust, products of a long-lived Mesoproterozoic continental-margin arc and associated back arc, and remnants of one or more accreted mid-Mesoproterozoic island-arc terranes (Hynes & Rivers, 2010; Tollo et al., 2004).

The initial stage of the Grenville orogen cycle (Elzeverian orogeny) sutured the Elzevir and Frontenac blocks to the eastern margin of Laurentia ca. 1.3-1.2 Ga, followed by widespread intrusion of the anorthosite-mangerite-charnockite-granite (AMCG) suite from ca. 1.19 to 1.11 Ga likely resulted from orogen collapse of overthickened crust (Whitmeyer & Karlstrom, 2007). The Ottawan orogeny spans the interval ca. 1.09-0.98 Ga and induced renewed northwest thrusting and imbrication of terranes in southeast Canada, involving widespread deformation along the eastern margin of Laurentia, when major collisions appear to have taken place with one or more larger continental masses, probably the Amazonia, at ca. 1.09-1.03 Ga (Hynes & Rivers, 2010).

3.1.5. Paleozoic Accretion and Collision - Appalachian Orogen

The Appalachians are a Paleozoic orogen that formed in a complete Wilson cycle along the eastern Laurentia margin following the breakup of Rodinia and the coalescence of all of the continents to form Pangea (Hatcher, 2010). It began by formation of a Neoproterozoic to early Paleozoic rifted margin and platform succession on the eastern margin of Laurentia. Three orogenesis ultimately produced the mountain chain: the Ordovician Taconic orogeny, which involved arc accretion (Karabinos et al., 1998); the Acadian-Neoacadian orogeny, which involved north-to-south, transpressional, zippered, Late Devonian-early Mississipian collision of the Coralina superterrane in the southern-central Appalachians and the Avalon-Gander superterrane in the New England Appalachians, and Silurian collision in the Maritime Appalachians and Newfoundland (Hibbard, 2000; Murphy & Keppie, 2005); and Alleghanian orogeny, which involved late Mississipian to Permian collision of all previously formed Appalachian component with Gondwana to form supercontinent Pangea (Hatcher, 2010).

3.1.6. Meso-Cenozoic Accretion - Cordillera Orogen

In the western portion of the North American continent the Cordillera orogen is a segment of the Circum-Pacific orogenic belt where subduction of oceanic lithosphere has been underway along a great circle of the globe since breakup of the supercontinent Pangea began in Triassic time (Dickinson, 2004). The length of the Cordilleran orogen from the Gulf of Alaska to the mouth of the Gulf of California is ~5000 km.

The Cordilleran orogen has been built by progressive tectonic addition of crustal fragments along the continental edge in Meso-Cenozoic. Such crustal growth is referred to accretionary tectonics (Saleeby, 1983). Accreted tectonic elements include subduction complexes assembled along the Cordilleran margin, intraoceanic island arcs attached to the continental margin by Jurassic arc-continent collision, and subduction complexes associated with the flanks of the exotic island arcs (Dickinson, 2008). The Cordilleran orogen is thought to represent an incomplete Wilson cycle in that it appears to have developed in the absence of a terminal continental collision. Instead, the Cordillera is interpreted as an accretionary orogen, and its evolution is explained as the result of the incremental, thin-skinned addition of terranes to the continental margin above a landward-dipping subduction zone (Monger, 1997). Johnston (Johnston, 2001) demonstrate that a large portion of the continental foreland of the Cordillera orogen is exotic with respect to the autochthon and forms part of a composite ribbon continent, referred to as SAYBIA. The North American Cordilleran orogen is the result of a two-stage process: (a) Triassic-Jurassic accretion within Panthalassa forming SAYBIA, a composite ribbon continent, and (b) Late Cretaceous collision of SAYBIA with North America (Johnston, 2008).

The geological history of North American continent presents a classical example of how continental crust grow gradually from small-sized cratonic nucleus to the current state of the continent. It is demonstrated by the geological records that accretionary and collisional orogenesis played a key role in enlargement of the continent probably since the first piece of continent was formed from the ocean in Paleoarchean. Other continents grow in a similar way as the North American Continent.

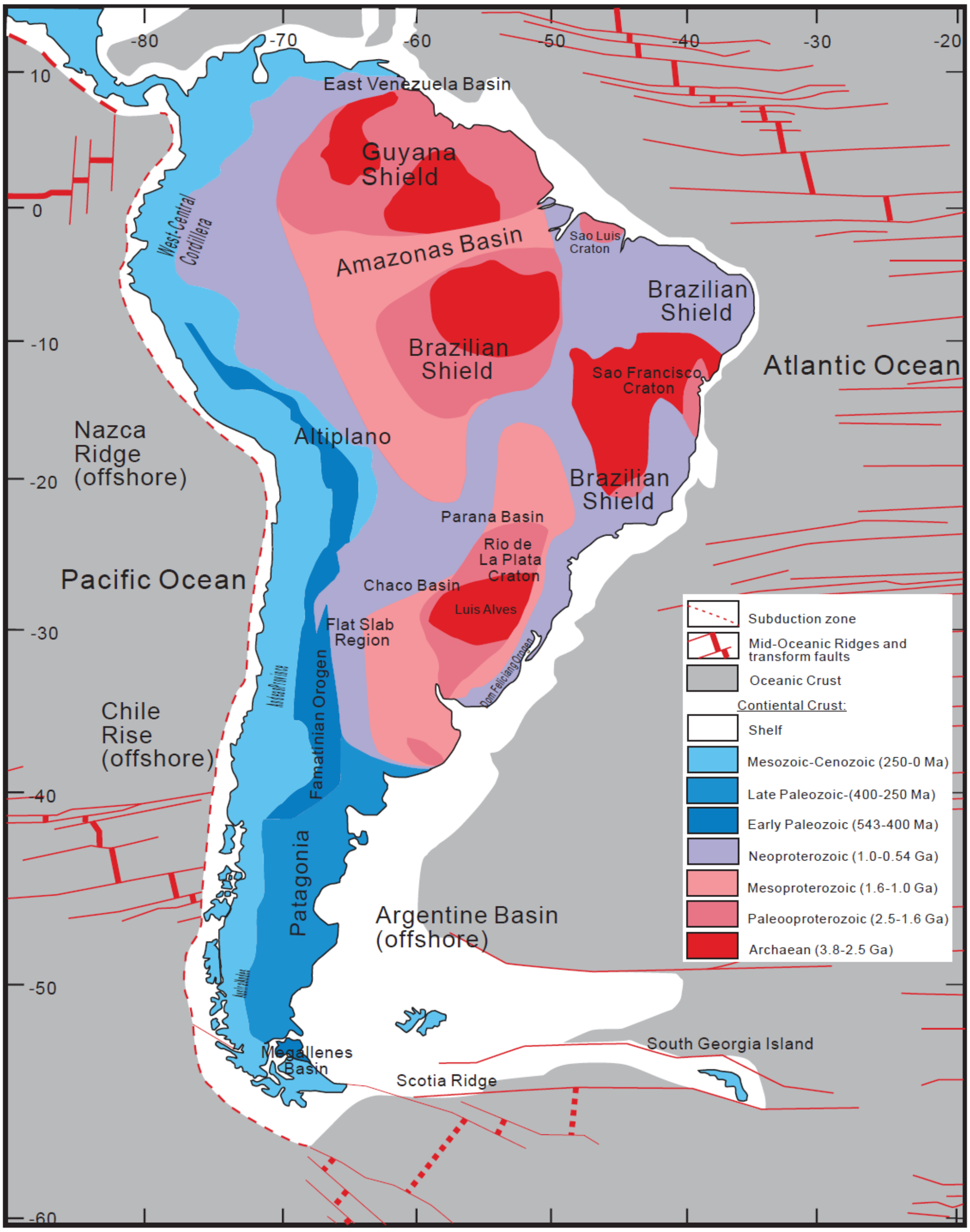

Figure 10. Main geological provinces of the South American continent (after Chulick et al., 2013).

3.2. Accretion of South American Continent

The South American continent consists mainly of the Precambrian Brazilian shield (South American Platform) in the east and late Paleozoic-Cenozoic Andean orogen in the west, with early Paleozoic Patagonia in the middle (Figure 10). The Brazilian shield is composed of the large Amazonian (AM) and São Francisco cratons, the small Rio de La Plata (RP), São Luis (SL) and Luiz Alves (LA) cratonic fragments, and the mobile belts associated to the Brasiliano-Pan African orogenic Cycle (Cordani & Sato, 1999).

The São Francisco craton is represented as a juxtaposition of four Archean continental blocks brought together during convergent and collisional processes at approximately 2.1 Ga during the Trans-Amazonian orogeny: the Gavião block, the granulitic Jequié block, the granulitic Itabuna-Salvador-Curaçá belt, and the Serrinha block (Peucat et al., 2002). The Gavião block is mainly composed of granite-greenstone belts of ca. 2.9-2.8 Ga, as well as an old nucleus with some of the oldest rocks in South America, with ages ranging from 3.4 to 3.2 Ga. The Jequié block is mainly formed of enderbite and charnockite with ages of 2.7-2.6 Ga, as well as migmatite and granulite. The Serrinha block contains 2.9 to 3.5 Ga amphibolite-facies banded gneisses, amphibolites and granitic orthogneisses (Barbosa & Sabaté, 2004).

The Río de la Plata craton is the oldest and southernmost core of South America and is a key piece in the cratonic assemblage of SW Gondwana. It has long been thought to comprise a Paleoproterozoic core surrounded by Neoproterozoic-Cambrian mobile belts (Gaucher et al., 2011). Sm-Nd T_{DM} model ages between 2.6 and 2.2 Ga characterize the Piedra Alta Terrane of this craton. Crystallization ages between 2.2 and 2.1 Ga for the metamorphic protoliths and 2.1-2.0 Ga for the post-orogenic granitoids indicate juvenile crust, followed by a short period of crustal recycling.

Cratonization of this terrane occurred during the late Paleoproterozoic (Oyhantçabal et al., 2011).

Three major orogenic collages are responsible for the formation of the South American platform: Middle Paleoproterozoic Trans-Amazonian orogenic cycle, Late Mesoproterozoic-Early Neoproterozoic, and Late Neoproterozoic-Cambrian Brasiliano-Pan African orogenic cycle (Brito Neves et al., 1999).

The Andes (Andean orogen) extends along the entire western margin of the South American continent, formed as a result of the subduction of Nazca plate beneath the Brazilian Shield, and is the world's second highest orogenic belt. The orogen is divided into tectonic provinces emblematic of Andean-type convergent margins, including a forearc region, magmatic arc, retroarc fold-thrust belt, and foreland basin system that closely follow the western trace of the South American coastline (Jordan et al., 1983). A protracted history of subduction and arc magmatism along the western edge of South America commenced in Late Triassic-Jurassic during breakup of Pangea and westward advance of the South American plate away from Africa (Pepper et al., 2016). The Andes north of the Golfo de Guayaquil, the Northern Andes record an important accretion of oceanic crust during Jurassic, late Cretaceous, and Paleogene times. The Central Andes between the Golfo de Guayaquil (4°S) and the Golfo de Penas (46°30'S) are a typical Andean-type orogen where tectonics was driven by subduction. The Southern Andes (46°30'-52°S) are developed south of the triple junction and are the result of uplift associated with ridge collision, along different ridge segments (Ramos, 1999).

3.3. Accretion of Australian Continent

The North Australian Craton, West Australian Craton and South Australian Craton underlie the western two-thirds of the Australian continent and are sutured by orogenic belts of largely Paleoproterozoic to Mesoproterozoic age. They are bounded on the east by the Phanerozoic Tasmanides, which form the other third of the continent (Figure 11). The cratons have Archean and/or early Paleoproterozoic cores on which are superimposed later Paleoproterozoic orogenic belts and basins; these, along with younger Proterozoic and Phanerozoic successions, conceal relationships between the older blocks (Withnall et al., 2013). The North, West, and South Australian cratons were independently accreted from older crustal fragments by 1.83 Ga (Myers et al., 1996).

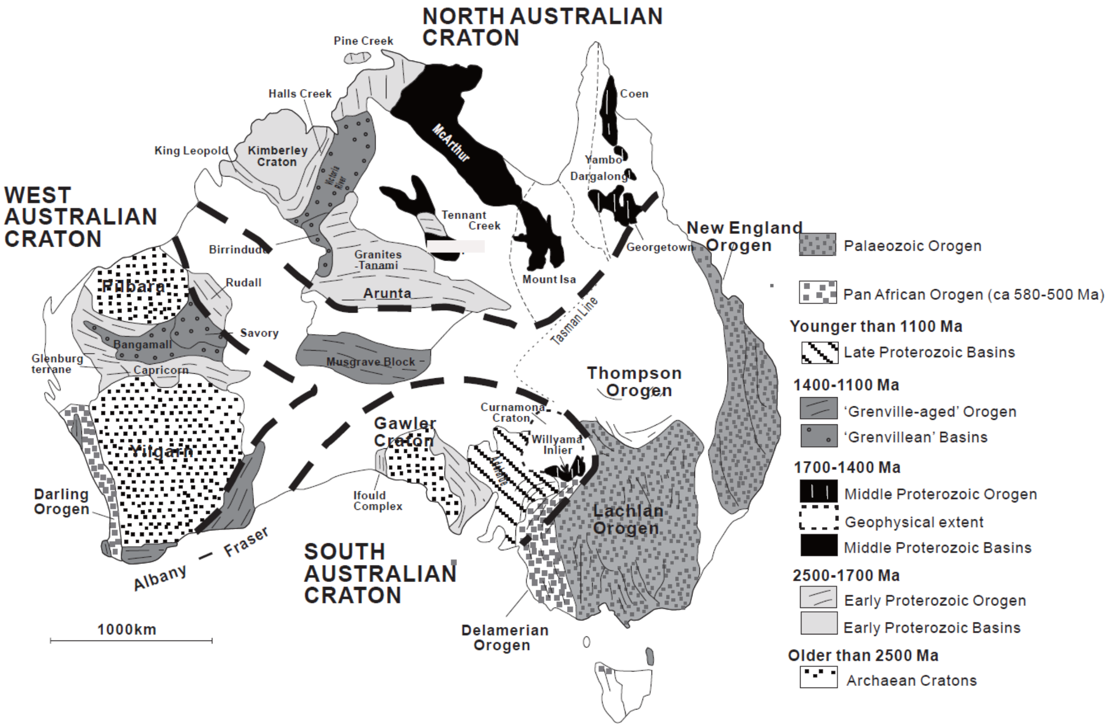

Figure 11. Terrane map of the Australian continent showing major Archaean and Proterozoic terranes, as well as the Paleozoic Tasmanides (adapted after Betts et al., 2002).

The North Australian Craton (NAC) includes Paleoproterozoic orogens and basins including the Halls Creek, Pine Creek, McArthur, Mount Isa, Tennant Creek, Tanami, and Aileron geological regions (Myers et al., 1996). Archean basement to the NAC crops out in the Pine Creek and Tanami regions, with ages in the range 2.67 Ga-2.50 Ga. An early phase of basin development at 2.05-2.00 Ga is reflected in the basal units of the Pine Creek Orogen (Scrimgeour, 2006). The Barramundi Orogeny represents a continent-wide accretionary event in which several Archaean cratons amalgamated to form the NAC in a manner that is analogous to the Paleoproterozoic amalgamation of Laurentia along the Trans-Hudson Orogen (Betts et al., 2002). A continental strip accreted to the southern margin of the NAC between 1.68 and 1.65 Ga. The West Australian Craton formed during the Paleoproterozoic (2.0-1.8 Ga) by the collision and combination of the Pilbara and Yilgarn cratons which are fragments of formerly more extensive Archean continents. The collision zone is marked by the Capricorn Orogen. The West Australian Craton was itself fragmented by rifting between 1.6 and 1.3 Ga, and then involved in collision and amalgamation with other continental fragments between 1.3 and 1.1 Ga along the Albany-Fraser and Rudall-Musgrave Orogens (Myers, 1993). The Yilgran Craton mainly consists of granite-greenstone terranes that formed between 3.0 and 2.6 Ga. The Pilbara Craton, the largest Archean nucleus in Australia, comprises a basement of 3.6 to 2.8 Ga granite-greenstone terrane unconformably overlain by a cover sequence between 2.76 and 2.4 Ga. Crustal growth occurred throughout the eastern part of the craton between 3.65 and 3.15 Ga. In the western parts of the craton, crustal growth is interpreted to have occurred in tectonic environments comparable to modern magmatic arc and backarc tectonic settings (Barley et al., 1998) and accretion of outboard island arcs and collisions of continental fragments (Smith et al., 1998). The South Australian Craton formed during the Kimban orogeny between ca. 1.84 and 1.70 Ga associated with collision and amalgamation of the Gawler and Curnamona cratons (Myers et al., 1996).

The North, West and South Australian cratons were combined at ca. 1.3 Ga as an early component of the Rodinian supercontinent. The NAC was first joined to the northeastern margin of the West Australian Craton. The combined West and North Australian cratons were joined to the South Australian Craton along the Albany-Fraser orogen (Myers et al., 1996). The Musgravian Orogen may represent a former continuation of the Albany-Fraser Orogen along the northern margin of the

South Australian Craton that was later displaced westward during younger intracratonic deformation (Myers et al., 1996).

Eastern Australia is made of Cambrian to Triassic (~550-220 Ma) subduction-related orogens that are collectively referred to as the Tasmanides (Fergusson & Henderson, 2015; Rosenbaum, 2018). This orogenic system is commonly subdivided into five orogens (Delamerian, Thomson, Lachlan, Mossman, and New England Orogens) that successively accreted to the Australian continent (Glen, 2005). The Tasmanides provide a geological record of the tectonic evolution of the eastern Gondwanan margin, from the Neoproterozoic breakup of Rodinia to the fragmentation of Pangea in the Permian-Triassic (Rosenbaum, 2018). The five orogens that define the Tasmanides generally become younger from west to east, with the oldest (Cambrian) components of the Tasmanides dominant in the Delamerian and Thomson Orogens, and the youngest (Triassic) components found in the New England Orogen.

The Delamerian Orogen covers the area of the southwestern Tasmanides and western Tasmania, and separates the Australian Precambrian cratons from the younger Paleozoic to Mesozoic orogenic belts of eastern Australia. It records the late Ediacaran (latest Neoproterozoic) to Cambrian evolution of the paleo-Pacific margin of Gondwana from a passive to a convergent margin (Turner et al., 2009). The protoliths of many rocks in the Delamerian Orogen are associated with rift-related Neoproterozoic sedimentary successions (Foden et al., 2006), which were subjected to deformation and metamorphism in the early to middle Cambrian.

The Thomson Orogen is the largest tectonic domain of the eastern Australian Tasmanides and represents approximately one-eighth of the Australian continent (Spampinato et al., 2015). It was a tectonically active area with several episodes of deposition, deformation and plutonism from Cambrian to Carboniferous time (Murray & Kirkegaard, 1978). The Thomson Orogen remains the least-known component of the Tasmanides because a vast majority of the orogen is covered under nonmetamorphosed Devonian to Cenozoic sedimentary rocks.

The Lachlan Orogen occupies the central part of the north-south trending Tasmanides along the eastern margin of Australia. It has long been described as accretionary orogen that provides an unmodified example of Paleozoic Circum-Pacific tectonics (Coney, 1992), or as a retreating subduction zone orogen; it did not suffer a terminal continent-continent collision (Foster & Gray, 2000). The Lachlan Orogen consists of three separate and distinct subprovinces, each with differences in rock type, metamorphic grade, structural history, and geological evolution. The western and central subprovinces are dominated by a turbidite succession consisting of quartz-rich sandstones and black shales. The eastern subprovince consists of mafic volcanic, volcaniclastic, and carbonate rocks, as well as quartz-rich turbidites and extensive black shale in the easternmost part (Foster & Gray, 2000; Q. Zhang et al., 2019). Shortening and accretion of the Lachlan occurred through stepwise deformation and metamorphism from Late Ordovician (~450 Ma) through early Carboniferous times, with dominant events at about 440-430 Ma and 400-380 Ma (Foster & Gray, 2000).

The Mossman Orogen, in the northeastern Tasmanides, mainly comprises deformed Silurian-Devonian sedimentary and igneous rocks of the Hodgkinson and Broken River Provinces. Deformation in the Mossman Orogen predominantly took place from the Late Devonian to earliest Carboniferous; the eastern part of the Mossman Orogen was subjected to a younger phase of deformation in the Permian and Triassic (Rosenbaum, 2018).

The New England Orogen is located in the easternmost part of the Tasmanides. Most of the rocks in the New England Orogen are Devonian to Triassic with a minor component of Neoproterozoic to Silurian rock units (Harrington & Korsch, 1985). The Devonian and Carboniferous units likely represent west to east continental arc and forearc assemblages (forearc basin and accretionary complex). Early Permian rocks are dominated by voluminous granitic intrusions, bimodal volcanism, and clastic sedimentary rocks, which were deposited in rift-related basins (Rosenbaum, 2018). A series of sharp bends (oroclines) are recognized in the Paleozoic to early Mesozoic New England Orogen (Cawood et al., 2011; Rosenbaum et al., 2012). These oroclines are formed ascribed to an early stage of subduction curvature during slab rollback at 300-285 Ma, followed by bending associated with dextral transpression, and final tightening by E-W shortening during the Permian to Triassic (265-230 Ma) Hunter-Bowen orogeny (Rosenbaum et al., 2012).

3.4. Accretion of African Continent

The African continent mainly consists of the Kalachari Craton in the south, the Congo Craton in the center, the West African Craton in the west, and the East African Orogen (Figure 12). The Kalahari Craton was spawned from a small composite Archaean core (Zimbabwe and Kaapvaal) which grew by prolonged crustal accretion in the Paleoproterozoic along its NW side (Magondi-Okwa-Kheis Belt, Rehoboth Subprovince) to form the Proto-Kalahari Craton by 1.75 Ga. From ca. 1.4 to 1.0 Ga, all margins of this crustal entity recorded intense tectonic activity: the NW margin was a major active continental margin between ca. 1.4 and 1.2 Ga and along the southern and eastern margins, the Namaqua-Natal-Maud-Mozambique Belt records a major arc-accretion and continent-collision event between ca. 1.1 and 1.05 Ga. By ca. 1.05 Ga, the Proto-Kalahari nucleus was almost completely rimmed by voluminous Mesoproterozoic crust and became a larger entity, the Kalahari Craton (Jacobs et al., 2008). Zircon U-Pb-Hf study clearly indicate that the terranes of the Kalahari Craton lie on distinct crustal evolution trends, which requires that the continental crust that constitutes these terranes formed at different times, and subsequently evolved differently. Furthermore, it requires that the terranes of the Kalahari Craton were successively accreted (Zeh et al., 2009).

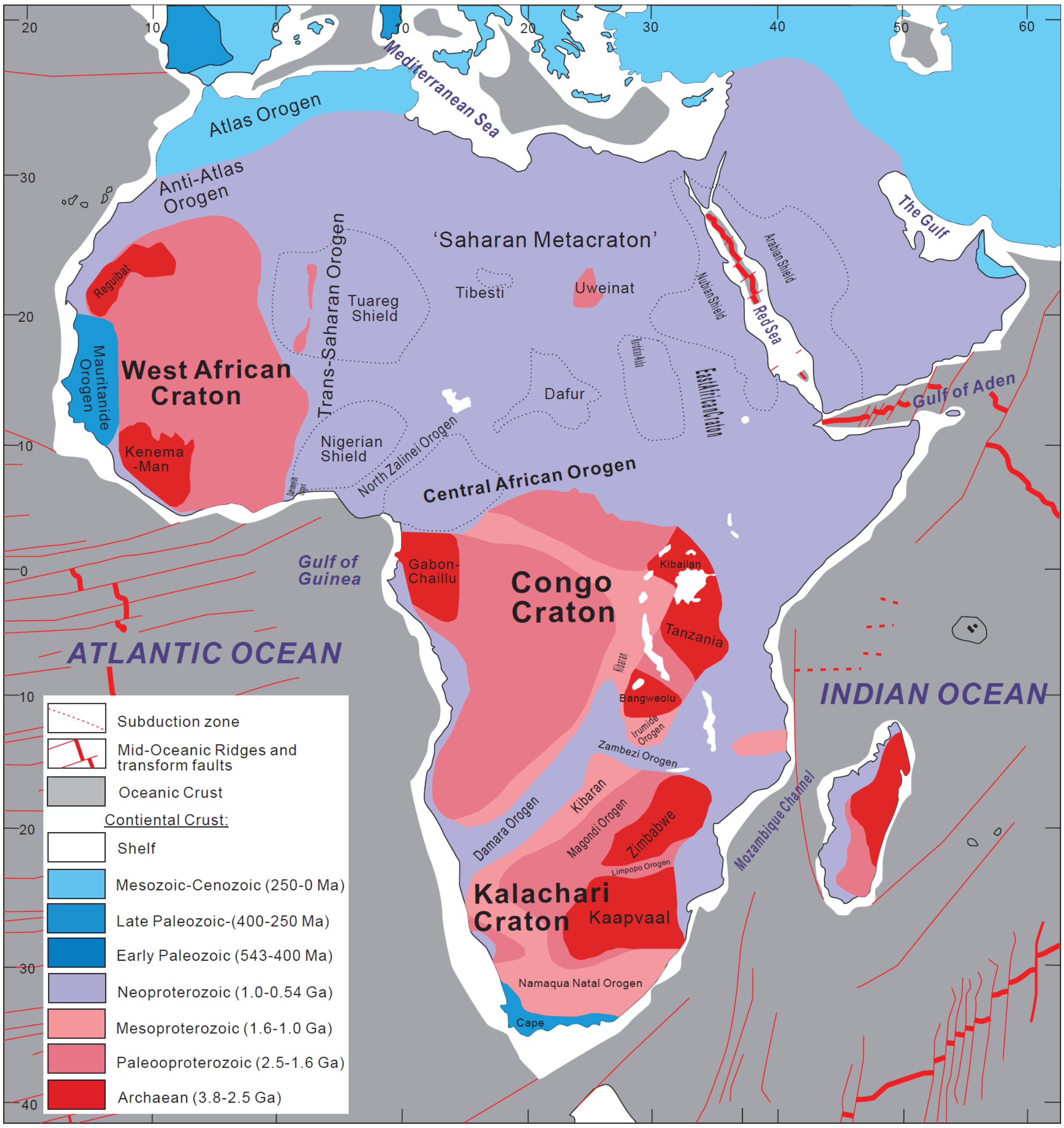

Figure 12. A sketch map showing the major constitutes of the South African continent (After Van Hinsbergen et al., 2011).

The Congo Craton is defined as the amalgamated central African landmass at the time of Gondwana assembly. The nucleation of the Congo Craton from its composing cratonic blocks, which include the Angola-Kasai Block, the NE-Congo-Uganda Block and the Cameroon-Gabon-Congo-São Francisco Block to the west and northwest of the Mesoproterozoic Kibaran Belt, and the Bangweulu Block and Tanzania Craton, to the east and southeast, was at the latest completed after peak compressional tectonism in the Kibaran Belt at 1.38 Ga. Late Mesoproterozoic tectonism along the southern margin of this proto-Congo Craton, in a region called the Irumide Belt, marks compressional tectonism at ca. 1.05-1.02 Ga, which produced extensive reworking along this margin, possibly linked to the participation of the Congo Craton in the Rodinia supercontinent (De Waele et al., 2008).

The West African Craton is a large slice of Archean to Paleoproterozoic continental crust divided into two main parts; the Keneman Rise to the south and the Reguibat Rise to the north, separated by the Neoproterozoic to Devonian Taoudeni Basin. The Keneman and the Reguibat Rises both consist of a western Archean domain that was affected by the subsequent Leonian and Liberian tectonothermal events and an eastern Paleoproterozoic domain accreted during the Birimian cycle around 2.1 Ga (Thiéblemont et al., 2001). The discovery of blueschist metamorphism in the 2.2-2.0 Gyr old West African metamorphic province suggests that modern-style plate tectonics existed during the Paleoproterozoic era (Ganne et al., 2012).

The Neoproterozoic orogeny, collectively termed the Pan-African Orogeny, led to the amalgamation of the Kalachali, Congo, and West African cratons and the final consolidation of the African continent. The Damara Orogen joins the Kalahari Craton in the south and the Congo Craton in the north. It contains a Neoproterozoic succession that was intensely deformed and metamorphosed in the protracted Neoproterozoic to Cambrian Pan-African Orogeny (Goscombe et al., 2004).

The East African Orogen (EAO), extending from southern Israel, Sinai and Jordan in the north to Mozambique and Madagascar in the south, is the world's largest Neoproterozoic to Cambrian orogenic complex. It comprises a collage of individual oceanic domains and continental fragments between the Archean Sahara-Congo-Kalahari Cratons in the west and Neoproterozoic India in the east. Orogen consolidation was achieved during distinct phases of orogeny between ca.

850 and 550 Ma (Fritz et al., 2013; Stern, 1994). The EAO is subdivided into the Arabian-Nubian Shield (ANS) in the north, composed largely of juvenile Neoproterozoic crust, and the Mozambique Belt (MB) in the south comprising mostly pre-Neoproterozoic crust with a Neoproterozoic–early Cambrian tectonothermal overprint (Fritz et al., 2013).

3.5. Accretion of Eurasian Continent

The Eurasian continent is currently the largest continent in the world. It mainly consists of six cratonic blocks: the Russian and Siberian cratons in the north, the Tarim and North China cratons in the middle, and the Afro-Arabian and Indian cratons in the south, and two huge Phanerozoic orogenic systems: the Altaids in the north and the Tethysides in the south (Figure 13).

Figure 13. Tectonic units of Eurasia, showing the locations of major cratonic blocks and orogenic belts, modified after Şengör et al. (2018).

3.5.1. Precambrian Cratonic Blocks

The Russian Craton (also called the East European or Baltica Craton) is the coherent mass of Precambrian continental crust that occupies almost the entire northeastern half of the European continent. It was assembled by the successive collision of three once autonomous crustal segments (Fennoscandia, Sarmatia and Volgo-Uralia) at ca. 1.8-1.7 Ga, roughly concomitantly with the formation of the Paleo- to Mesoproterozoic Columbia/Nuna supercontinent, which appears to have persisted until ca. 1.4 Ga. After its formation, the Russian Craton has never been dismembered completely, but signs of accretion of new crust and, in particular, rifting along its margins as well as truncated tectonic trends indicate that its size and shape have changed repeatedly (Bogdanova et al., 2008). Isotopic data indicate repeated episodes of generation of juvenile continental crust along and outboard of the margin of the dominantly Archaean Craton in Sarmatia, which are best explained by accretionary plate tectonics. An Andean-type geodynamic setting between 2.1 and 2.0 Ga ago was succeeded by the formation of several island arcs and accretion of these onto the older continent between 2.0 and 1.8 Ga. Subsequent compression of the newly formed crust lasted until ca. 1.7 Ga (Claesson et al., 2001).

The Siberian Craton was assembled at ca. 2.1-1.8 Ga by Archean blocks through orogenic belts and suture zones. Similarities in ages between the Paleoproterozoic orogenic belts of Siberia and Laurentia cratons suggest that they could have originated from the same proto-craton (Gladkochub et al., 2006). The present-day structure of craton is generally considered as a result of collision, amalgamation and accretion of microcontinents different in age, which turned into heterogeneous tectonic blocks in the course of collision that is consistent with the current concept of accretionary tectonics. The craton comprises the Tungus, Anabar, Olenek, Aldan, and Stanovoi provinces. The latter consist of heterogeneous tectonic blocks with superimposed Paleoproterozoic fold and orogenic belts sometimes. Areas between terranes correspond to Paleoproterozoic island-arc fragments, the largest among which is the Akitkan orogenic belt. The belt represents a Paleoproterozoic island-arc system sandwiched between the Anabar and Aldan superterranes (Rosen et al., 2006). The system development terminated by the collision prism collapse 1.8 Ga ago.

The Tarim Craton, located in northwestern China, is one of the largest Precambrian cratonic blocks in East Asia. It is largely covered by Cenozoic desert, and the Precambrian basement mainly expose along its southwestern and northeastern margins, i.e. Altyn and Quruqtagh regions. The Tarimian Orogeny resulted in the final cratonization of the Tarim Craton, and produced an unconformity between the metamorphosed basement (Archaean to early Neoproterozoic) and the sedimentary cover (middle Neoproterozoic to Phanerozoic) (Lu et al., 2008).

The North China Craton (NCC, or Sino-Korea Craton) is the largest Precambrian cratonic block in east Asian continent. It contains some fragments of the oldest continental crust in the world dated back to ca. 3.8 Ga (Liu et al., 1992). Although different tectonic models exist, it is generally agreed that the NCC was formed through amalgamation of different small Archean blocks during the Paleoproterozoic accretionary and collisional orogenesis (Kusky, 2011; Zhai & Liu, 2003; Zhao et al., 2001). The Archean blocks are dominated by tonalite-trondhjemite-granodiorite (TTG) gneisses and granite-greenstones. Based on age, lithological assemblage, tectonic evolution and P-T-t paths, Zhao et al. (2000, 2001) divided the NCC into the Western Block, the Eastern Block and the intervening Trans-North China Orogen. The Western Block forms a stable platform composed of late Archean to Paleoproterozoic metasedimentary belts that unconformably overly Archean basement, whereas the Eastern Block consists of granulite facies TTG gneiss and charnockite with minor mafic granulite and amphibolite. The Trans-North China Orogen is composed of late Archean amphibolite and granulite and 2.5 Ga granite-greenstone terrains (Zhao et al., 2000, 2001). The final amalgamation of the Western and Eastern blocks of the craton is suggested to have occurred at ~1.85 Ga (Zhao et al., 2012), or 2.5 Ga (Kusky & Li, 2003; Kusky, 2011) during an arc/continent collision. The style of tectonic accretion in the NCC changed at circa 2.5 Ga, from an earlier phase of accretion of arcs that are presently preserved in horizontal lengths of several hundred kilometers, to the accretion and preservation of linear arcs several thousand kilometers long with associated oceanic plateaus, microcontinents, and accretionary prisms (Kusky et al., 2016). The style of progressively younger and westward outward accretion of different tectonic components is reminiscent of the style of accretion in the Superior Craton.

3.5.2. Phanerozoic Orogenic Systems

The Altaids (or Altaid Tectonic Collage), also known as the Central Asian Orogenic Belt (CAOB), sandwiched between the Russian Craton, the Siberian Craton and the Tarim-North China Craton, is considered as one of the largest Phanerozoic accretionary orogens in the world (Şengör et al., 1993; Windley et al., 2007). It is proposed that the Asia continent grew by 5.3 million km^2 during the Paleozoic through the growth of subduction-accretion complexes along a single magmatic arc now found contorted between Siberia and Baltica (Şengör et al., 1993; Şengör & Natal'in, 1996). Isotopic data show that most of the granitoids in the Altaids have depleted Nd-Hf compositions, consistent with the large-scale distribution of juvenile continental crust (Jahn et al., 2000).

The Altaids is formed through complicated amalgamation between subduction-accretion complexes, magmatic arcs, seamounts, forearc and backarc basins, and continental fragments during the progressive subduction and final closure of the Paleo-Asian Ocean (Windley et al., 2007). There are different types of magmatic arcs, including Andean-type arcs, intro-oceanic island arcs and Japan-type island arcs within the Altaids (Song et al., 2015; Xiao et al., 2020). Pre-Altaid basement, although exist, are generally paucity in the Central Asia. In contrast, most terranes are composed of juvenile crust newly exacted from the mantle, either through 'syn-subduction lateral accretion of arc complexes' and 'post-collisional vertical accretion of underplated mantle material' (Long et al., 2011). In addition, ridge subduction has also played an important role in the continental growth of Central Asia (Windley & Xiao, 2018).

The timing of accretion of the Altaids generally become younger southwards (present coordinates), and the final closure of the Paleo-Asian Ocean took place along the South Tianshan-Beishan-Solonker suture in the southernmost Altaids. The termination of the Paleo-Asian Ocean was scissor-style eastward from Permian to Middle Triassic (Xiao et al., 2009).

The Tethysides is situated to the south of the North China and Tarim cratons. It contains a record of the progressive closure of the Proto-Tethys, Paleo-Tethys, Meso-Tethys, and Neo-Tethys Oceans resulting from the convergence between the Gondwana and Laurasia continents (Şengör, 1987). In the east the Qinling and Dabie-Sulu orogens formed through closure of the ocean between the North China and South China cratons in the Mesozoic (Ernst & Liou, 1995). In the west there are the Kunlun-Qilian orogen in the north, the Himalayan-Tibetan orogen in the south and southeast, and the Sanjiang orogen in SW China, which formed by progressive closure of oceans caused by Paleozoic to Cenozoic convergence between blocks rifted from the northern margin of the Gondwana and Eurasian continents. The Tethys Ocean was finally terminated by collision between the Indian continent and the southern margin of the Eurasian continent giving rise to the bulk of the Tibetan Plateau (Ding et al., 2005). The India-Asia collision leads to tremendous reworking and transforming of the Asian Continent expressed by high-topography in south-central Asia and large-scale strike-slip deformation in SE Asia.

4. Reworking of Continents

Continents have evolved for several billion years ever since their formation in the Archean time (see section 2). Such long-term evolution of continents involves multiple phases of orogeny and frequent tectono-thermal processes, resulting in not only episodic additions of new continental materials (continental accretion or growth; see section 3), but also repeated modifications of the existing continental lithosphere (continental reworking) (e.g., Holdsworth et al., 2001). Continents are therefore characterized by both longevity in their evolution and complex modification in the lithosphere.

4.1. Longevity of Continents

The majority of continents consists of long-lived crust and lithosphere, namely cratons that have been isolated from the convecting mantle since the Precambrian, mostly for two or three billion years (e.g., Lee et al., 2011; Lenardic & Moresi, 1999). There have been numerous lines of evidence suggesting the preservation of Precambrian crustal basement and similarly old deeper mantle roots beneath cratons (e.g., Pearson, 1999; Richardson et al., 1984; Spetsius et al., 2002; Walker et al., 1989). Compared to the short-lived oceanic lithosphere that is recycled back into the mantle on a timescale of ~100 Ma, continental lithosphere is capable of resisting deformation and avoiding recycling as a whole over a timescale of about one order of magnitude longer (Lenardic et al., 2003; O'Neill et al., 2008). The longevity and stability of continents, in particular cratons, have been primarily attributed to the structure and intrinsic (physical and chemical) properties of the continental lithosphere, relative to the asthenosphere and oceanic lithosphere (Figure 14 and Table 1).

4.1.1 Intrinsic buoyancy of continental lithosphere

One relevant property to the longevity and stability of continents is thought to be the intrinsic buoyancy of continental lithosphere (Carlson et al., 2005; Jordan, 1978). This property is closely related to the lithospheric structure and composition of continents that much differ from their oceanic counterparts. Continental lithosphere is composed of a sialic crust of ~40 km thick on average and a mantle keel of melt depleted residuum composition with highly variable thicknesses (Table 1). The majority of continental lithosphere is thicker than ~120 km, even up to >200 km beneath Archean cratons, with thin lithosphere of a few tens of kilometers in continental margins or active rift zones (Figure 14). In contrast, oceanic lithosphere consists of a ~7-km basaltic crust and a thin mantle keel with a total thickness of mostly no more ~100-110 km (Figure 14, Table 1). These large differences in the crustal and lithospheric thickness and composition are essential for the distinct fates of continents and oceans. With progressive cooling, oceanic lithosphere becomes denser and thicker as ages and eventually downwells to form a subducting slab. In contrast, the continent lithosphere is intrinsically chemically buoyant, which likely offsets its negative buoyancy due to lower temperatures than the underlying asthenosphere (Figure 15a) and makes it in isostatic equilibrium (e.g., Jordan, 1978, 1988). Continental lithosphere is thus not prone to subduct as oceanic lithosphere does.

The chemical buoyancy of continental lithosphere is contributed from both the melt depleted residuum composition of the mantle root and the sialic composition of the crust. The melt depletion of the mantle root is characterized by less Fe relative to Mg (or high Mg#, Table 1) and a high proportion of olivine to clinopyroxene+garnet, as a result of extensive melt extraction (Lee et al., 2005; Poudjom Djomani et al., 2001). This chemical depletion generally increases with age of continent and leads to the most refractory nature of Archean mantle root (Lee et al., 2011; O'Reilly et al., 2001). Accordingly, the densities of subcontinental lithospheric mantle (SCLM) decrease in order and are averaged at standard P-T conditions as ~3310 kg/m^3, ~3340 kg/m^3 and ~3360 kg/m^3 for the Archean, Proterozoic and Phanerozoic periods, respectively (Table 1), corresponding to ~2.5%-1% density reduction compared to asthenospheric mantle at similar depths (Artemieva, 2009; Deen et al., 2006; Poudjom Djomani et al., 2001). Compared to the mantle root, the averagely ~40-km sialic continental crust, which is much more buoyant than the ~7 km basaltic oceanic crust and asthenospheric mantle (Table 1), contributes a larger, even dominant part (>70%) to the overall buoyancy of the continental lithosphere. A significant reduction of the crustal contribution by scraping off the most buoyant upper crust would dramatically increase the density of the continental lithosphere and make it readily subductable (e.g., Capitanio et al., 2010). However, recent numeric modeling studies indicate that, although intrinsic buoyancy improves the stability of continental lithosphere, it is insufficient by itself to account for the longevity of cratons (Doin et al., 1997; François et al., 2013; Lenardic & Moresi, 1999; O'Neill et al., 2008).

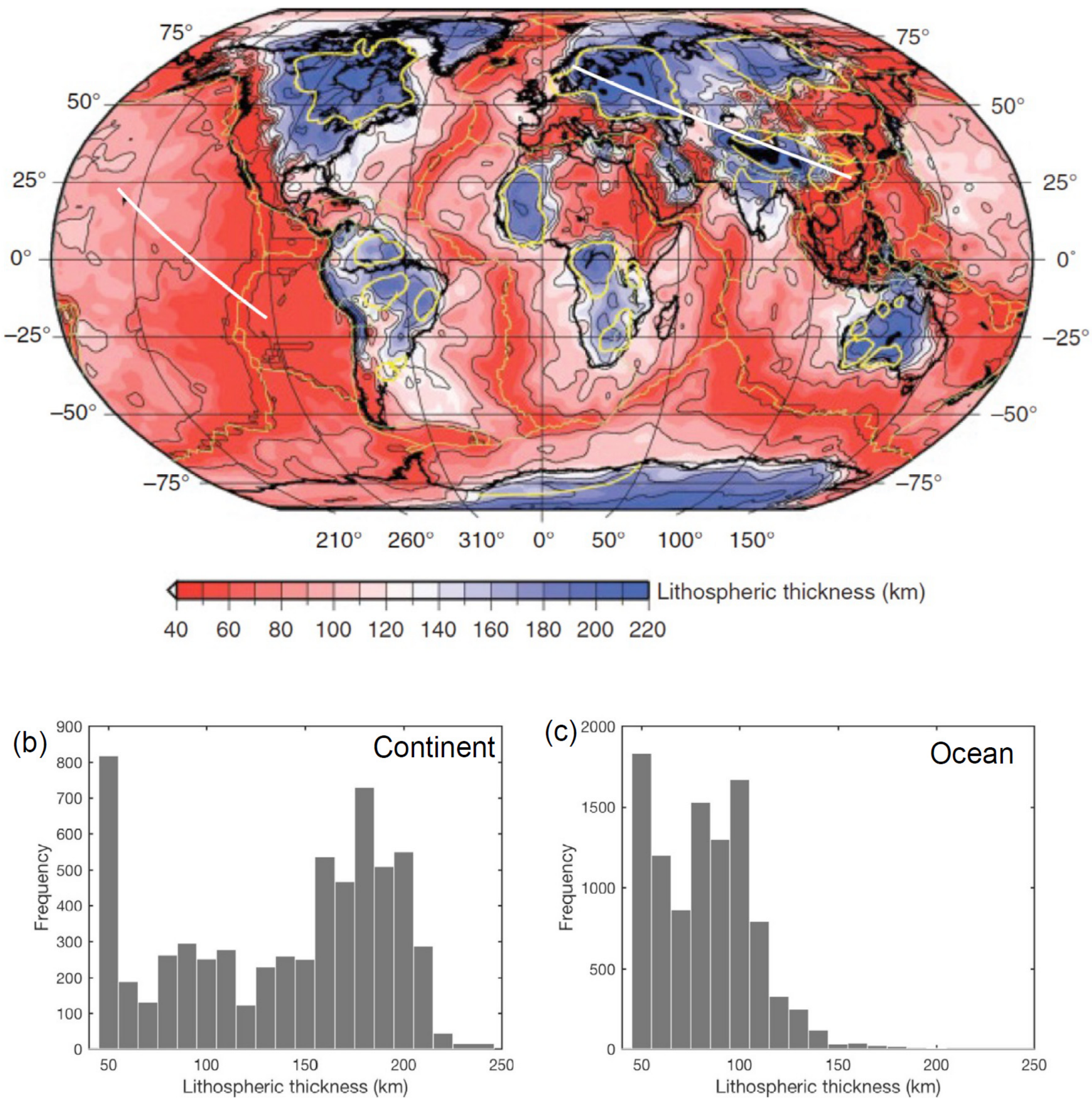

Figure 14. Global lithospheric thickness map (a), and histograms of lithospheric thicknesses beneath continents (b) and oceans (c), respectively. The lithospheric thicknesses were obtained from the upper mantle Vsv model CAM2016Vsv by surface wave tomography and the Vs(T) relationship given in Priestley et al. (2019), and the thickness data were downloaded from http://ds.iris.edu/ds/products/emc-cam2016/ provided by Keith Priestley and Tak Ho. (a), modified from Priestley et al. (2019). The yellow contour denotes the geologically mapped boundary of the shields at the surface. The two while lines mark the locations of two cross-sections in Figure 16. (b) and (c) are plotted by separating the lithospheric thickness data for continents and oceans. Note that the lithospheric thicknesses derived by surface wave tomography have uncertainties of a few tens of kilometers (Fichtner et al., 2010).

4.1.2. High Strength of Continental Lithosphere

It has been suggested that high strength (viscosity) of continental mantle roots is essential for the long-term stability of cratons (Carlson et al., 2005; Karato, 2010; Pollack, 1986). High root strength compared to both oceanic lithospheric mantle and asthenosphere is primarily due to the low temperatures, but also attributed to the exceptional melt depletion and thus dehydration of cratonic roots (Table 1) (Figure 15b-c). The effect of dehydration is particularly pronounced by a sharply increase in rheological contrast (viscosity ratio) to more than one order of magnitude between the lowermost mantle root and the top of the asthenosphere (inset in Figure 15c) with typically reported water contents in both cratonic lithospheric mantle and asthenosphere (Table 1). This feature differs markedly from the continuously variation of viscosity ratio with depth if only temperature effect is considered (Figure 15c). Geodynamical models argue that a plausible range of rheological contrast between a cratonic root and asthenosphere (2-3 orders of magnitude for a constant stress) could prevent the cratonic root from convecting mantle erosion for

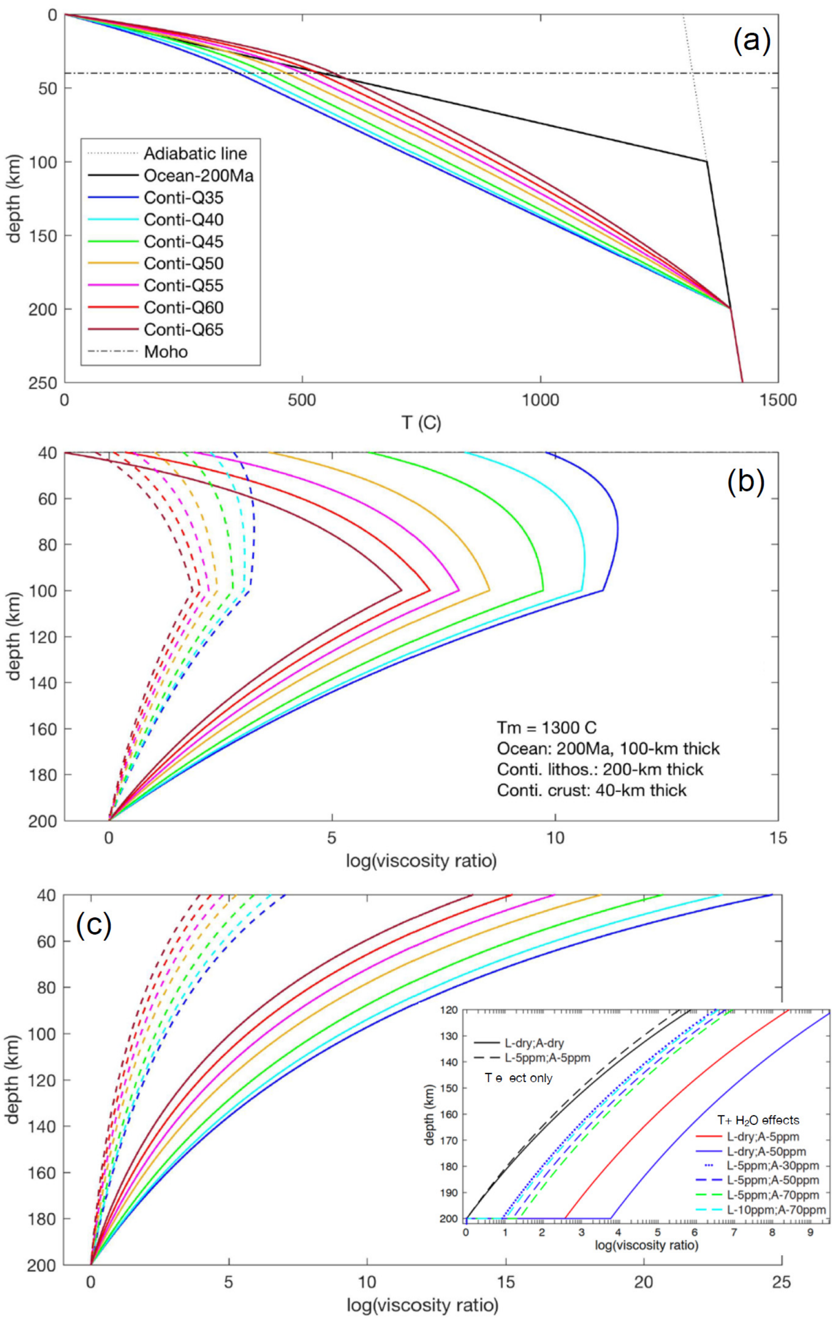

Figure 15. (a) Simplified temperature profiles for a 200Ma old and 100-km thick oceanic lithosphere and a serious of 200-km thick continental lithosphere with various surface heat flows (from 35 mW/m2 to 65 mW/m2 with 5 mW/m2 increment); b) Rheological contrasts between the continental lithospheric mantle and oceanic lithospheric mantle due to the temperature differences shown in (a); c) Rheological contrasts between the continental lithospheric mantle and the top of the asthenosphere with only temperature effect (main panel) and considering the effects of both temperature and various water contents in olivine

in the cratonic lithosphere (dry to 10 ppm) and asthenosphere (dry to up to 70 ppm) (insert). Solid lines and dash lines in (b) and the main panel of (c) are for constant stress and constant strain rate cases, respectively. The insert in (c) is for surface heat flow of 45 mW/m2 and constant stress case. The values of water contents are based on estimates from Peslier et al. (2017) and Hirth and Kohlstedt (1996). See also Table 1. A dislocation creep rheology with the parameters given in Dixon et al. (2004) are used here. The temperature effect is calculated under dry condition.

over billions of years, even without considering the effects of the chemical buoyancy of the root (e.g., Lenardic & Moresi, 1999; O'Neill et al., 2008; Wang et al., 2014). If the intrinsic buoyancy is further involved in modeling, the value of required viscosity ratios could be further reduced to below two orders of magnitude (~50 for a constant stress, Wang et al., 2014). It roughly falls into the range both calculated here at around the bottom of the cratonic root (inset in Figure 15c) based on the data of mantle water distribution (e.g., Peslier et al., 2017; Xia & Hao, 2013) and provided by laboratory experiments (10s up to 10000) for dehydration strengthening for upper mantle minerals, mostly olivine (Fei et al., 2013; Hirth & Kohlstedt, 1996; Karato, 2010). However, experimental results reported a rather large span of strength drop (> 1-2 order of magnitude) due to similar water abundance in the dislocation creep regime (e.g., Demouchy & Bolfan-Casanova, 2016; Fei et al., 2013), which indicates that, although the dominant role of high root strength for craton longevity has been a consensus, the effects of volatiles, especially water on mantle rheology is yet not well constrained (Brodholt, 2013).

In addition to the high strength of the mantle root, the high strength of the crust and strong coupling between crust and lithospheric mantle also contribute to the rigidity and long-term stability of continental lithosphere. This was previously indicated by both numerical simulations on mantle dynamics (e.g., Lenardic et al., 2003) and theoretical calculation of rock strength (Burov & Diament, 1995). Isotopic dating (Aulbach et al., 2004; Pearson et al., 1995) and seismic imaging results of continental lithosphere of various ages (Darbyshire et al., 2017; Durrheim & Mooney, 1994; Nguuri et al., 2001) provide further evidence for the crust-mantle coupling during the evolution of continents. In cratonic regions, it is suggested that the crust and lithospheric mantle behave in a coupled manner to keep the lithosphere strong and stable. On the one hand, a strong crust helps to maintain the high strength of the continental lithosphere as a whole and prevents the crust detaching from the subcrustal mantle (Burov & Diament, 1995; Lenardic et al., 2003). On the other hand, the thick and high-viscosity lithospheric mantle separates the crust from the convecting mantle and serves as a stiff root to protect the crust from being heated and thus weakened and being destroyed during mantle processes (e.g., Carlson et al., 2005). Overall, both the strong crust and lithospheric mantle and therefore strong crust-mantle coupling are all of importance for the longevity of continents.

4.1.3. Structural Heterogeneities Within Continental Lithosphere

Beside the intrinsic chemical buoyancy and high strength of cratonic mantle roots, the most distinct feature of continental lithosphere compared to the oceanic counterpart is probably the ubiquitous structural heterogeneities, as revealed increasingly by geophysical observations (e.g., Fichtner et al., 2010; Priestley et al., 2019; Schaeffer & Lebedev, 2014). The heterogeneous structural feature is closely related to the early formation and long-term evolution of continents, which involves complex lithospheric processes of amalgamation and accretion of smaller blocks of various ages and subsequent thermo-tectonic modifications (see sections 2-3). Old cratons are typically surrounded by younger mobile belts of distinctly different evolution histories, resulting in complicated lateral structural variations especially at around the margins of cratons (Figure 16a-d). Ocean basins, on the other hand, have a relatively simple history of evolution, forming at middle ocean ridges, expanding towards both sides and diminishing at subduction zones. Accordingly, the lithospheric structure of ocean basins mainly monotonously varies with age (Figure 16e), exhibiting relatively more simplified features than the continental lithosphere at similar scales (Figure 16d).

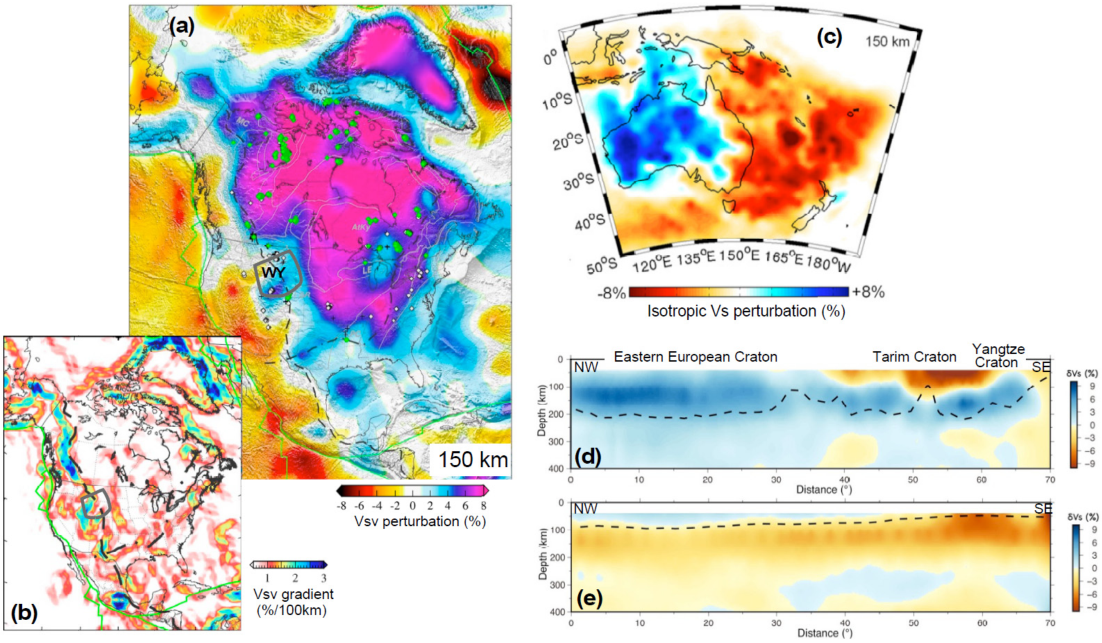

Figure 16. Structural heterogeneities of continental lithosphere and comparison with oceanic lithosphere. (a) Shear wave velocity (Vsv) structure at 150-km depth beneath North America, showing high-velocity cratonic core and lower-velocity younger orogenic belts surrounding the core. Green filled and white diamonds mark the locations of diamondiferous and non-diamondiferous kimberlites and lamproites, respectively. (b) Lateral Vsv gradients shown in (a), plotted in percentage per 100 km. Larger velocity gradients concentrate at around cratonic margins or plate boundary areas. Dashed black line in (a,b) separates the little deformed stable cratonic core from the strongly deformed tectonic region to the west. The Wyoming craton (WY) is outlined in thick gray. (a,b) are modified from Schaeffer and Lebedev (2014). (c) Isotropic shear wave velocity perturbations at 150-km depth beneath Australia (Fichtner et al., 2010), showing structural variations between cratons and surrounding younger belts similar to (a). (d,e) Shear wave velocity cross-sections within the Eurasian continent (d) and Pacific ocean (e). See Figure 1a for the locations of the cross-sections. Dash line in (d) and (e) marks the base of the

lithosphere from Figure 1a. The velocity model is the global upper mantle tomographic Vsv model CAM2016Vsv downloaded from http://ds.iris.edu/ds/products/emc-cam2016/ provided by Keith Priestley and Tak Ho. A-A’: (63.562N, 15.204E) -- (27.241N, 115.088E); B-B’ : (23.168N, 158.876W) -- (18.460S, 102.600W).

Structural heterogeneities reflect different nature and properties, in particular mechanical strength (rheology) of the lithosphere within continents (e.g., Hieronymus et al., 2007) and have been invoked to explain the longevity and stability of cratons (e.g., Lenardic et al., 2003; Yoshida, 2012). Mobile belts surrounding a craton are mechanically weak zones, relative to the craton. Because of their weakness, peripheral mobile belts may themselves become the preferential/favorable loci of intensive heating and strain localization (Tommasi et al., 2001; Vauchez et al., 1997), and thus could shield the craton from tectonic stressing and thermo-mechanical erosion. Such processes may lead to further sharpening of the already existed variations in lithospheric structure and properties within continents, consistent with plenty of seismic observations that reveal distinct structural changes across cratonic boundary zones (Figure 16a-d), such as Europe (Shomali et al., 2006; Wilde-Piórko et al., 2010), East Asia (Chen, 2010; Tao et al., 2018), Australia (Fichtner et al., 2010; Fishwick et al., 2008), North America (Schaeffer & Lebedev, 2014; Schmandt & Lin, 2014), and South Africa (James et al., 2001; Ortiz et al., 2019). The overall effects of structure and property heterogeneities for the longevity and stability of cratons mainly depend on the differences and interactions between a craton and its adjacent mobile zones rather than on the craton itself (thus regarded as external factors, Wang et al., 2014), which is closely related and probably complementary to the roles played by the high strength and intrinsic chemical buoyancy of cratonic roots (internal factors).

4.2 Evidence for Continental Reworking

Continents, even the most stable cratons are not invariable forever. Despite the long-term stability of continents (cratons) compared to their oceanic counterparts, continental lithosphere is also characterized by episodic tectono-thermal disturbations or reworking throughout its history. Continental reworking typically involves interactions of between lithosphere and asthenosphere, crust and mantle or among tectonic blocks, accompanying mountain building, crustal growth or accretion (e.g., Griffin et al., 2009; Holdsworth et al., 2001; Peslier et al., 2017; Stern & Scholl, 2010). The styles and processes of reworking are expressed by the way in which the existing lithosphere responses to the tectono-thermal events (Holdsworth et al., 2001). Significant continental reworking may not only affect the structure and properties of the involved continental lithosphere, but also link to or exert impacts on both the surface processes and mantle dynamics of a broader region, even globally (e.g., Allen & Armstrong, 2008; Hu et al., 2018; Liu, 2014; Tassara et al., 2017).

Evidence of continental reworking is ubiquitous both in space and time. In the following we provide a number of examples from shallow to deep lithospheric depth, from an integrated geological and geophysical point of view.

4.2.1 Surface Geology and Tectonic Deformation

Reworking of continents is reported to be commonly associated with tectonic processes, i.e., mostly oceanic subduction and continental collision occurring at plate boundaries. Well-known examples come from the circum-Pacific subduction zones and the Tethyan collisional belt (Figure 17, Müller et al., 2019). In western North America, the Late Mesozoic to Cenozoic subduction of the Farallon plate and associated processes have severely reworked the overlying continental plate. This includes the creation of thick-skinned thrust belts with strong lithospheric deformation over 1000-2000 km inboard by flat Farallon subduction during the 80-40 Ma Laramide Orogeny (Copeland et al., 2017; Dickinson, 2004; Yonkee &

Weil, 2015), occurrence of mid-Tertiary ignimbrite flare-up following slab rollback and subduction of the Farallon-Pacific mid-ocean ridge (MOR) (e.g., Dickinson & Snyder, 1979), development of the Cordilleran metamorphic core complexes as a result of crustal spreading of an early thickened lithosphere (Coney, 1980; Lister & Davis, 1989; Liu, 2001) and formation of the extremely extensional Basin and Range Province over the slab window due to MOR subduction (e.g., Dickinson & Snyder, 1979). Oceanic subduction with distinct slab morphologies and slab-mantle interactions also took place in other circum-Pacific regions since late Mesozoic. It has resulted in the uplift of the Andes including the Altiplano Plateau (second largest plateau in the world) with overwhelming compressional deformation but episodic extensions in western South America (Chen et al., 2019).

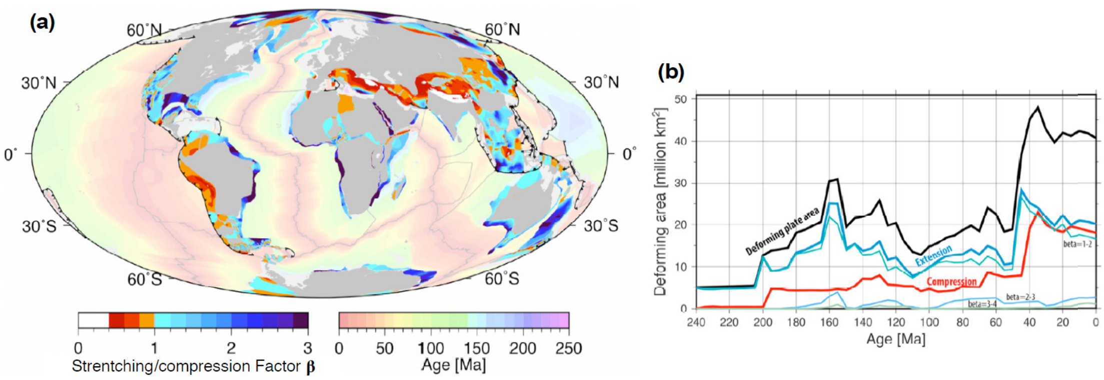

Figure 17. Distributed continental deformation over the last 240 million years (Müller et al., 2019). (a) Spatial distribution accumulated over time, with areas in extension represented by the stretching/compression factor **β** > 1 and those in compression by **β** < 1 ; b) Total deforming area (black), that in extension (blue) and that of shortening (red) through time. Extensional areas are further classified into three groups according to the stretching factor **β**: **β** = 1–2 (turquoise), **β** = 2–3 (light blue), and **β** = 3–4 (light green). Note that continental extension occurred mainly in the Late Jurassic and after 50 Ma, and compressional deformation of continents also increased sharply after ~50 Ma. Continental stretching and shortening since ~50 Ma dominantly reflect lithospheric deformation associated with circum-Pacific subduction and Tethyan collision.

In the western Pacific-eastern Asian region, on the other hand, developed large-scale extensional basins on land and back-arc basins offshore, eventually giving rise to the formation of the extension-dominated trench-arc-basin systems with vigorous volcanism both along arcs and within the continent (Liu et al., 2017). These circum-Pacific subduction processes are thought to be responsible for the basement-involved deformation and marked changes in the structure, thermal state and rheology of not only young tectonic belts but also the presumably strong cratonic lithosphere, as reported in the western part of the North American craton including Wyoming craton (e.g., Carlson et al., 2004; Dave & Li, 2016); Figure 16a-b), Brazilian craton in South America (Beck & Zandt, 2002) and North China Craton (NCC) in eastern China (e.g., Xu, 2001; Zhu et al., 2011). This will be further described and discussed in 4.2 and 4.3. Along the Tethyan collisional belt, the subduction and Cenozoic closure of the Neo-Tethys ocean and subsequent collisions of the Indian, Arabian plates and the micro continental ribbon of Africa with Eurasia have led to not only the formation of the strongly deformed thrust belts of the Alps-Zagros-Himalayan orogens at the collisional front, but also severe reactivation of pre-existing weaknesses and diffuse deformation within the continental interior in south to central Eurasia (Figure 17), accompanying widespread syn- and post-collisional igneous activities (Faccenna et al., 2014; Müller et al., 2019; Yin, 2010). Such processes have given rise to the development of the giant Tibetan plateau in south Asia featured by the highest topography (~5 km) and thickest continental crust (~60-80 km) over the world (Figure 18), which has exerted profound impacts on the tectonic framework, climate and life on Earth (e.g., Guo et al., 2002; Jolivet et al., 2018; Molnar et al., 1993).

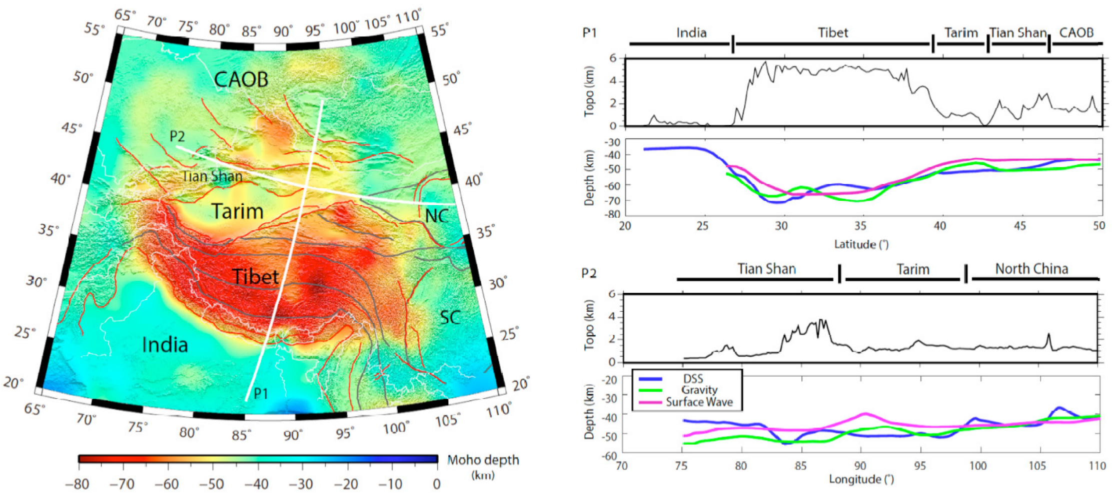

Figure 18. Variations of Moho depth in Tibet and adjacent areas mainly from deep seismic sounding (DSS) data (left) and along two profiles (right). Results from gravity inversion and surface wave tomography are compared with that of DSS along the profiles. The figure is modified from Teng et al. (2013). Note that the Moho is the deepest in Tibet (~60-80 km) and also depressed beneath Tian Shan (>50 km), roughly mirroring the topography. The deepening of the Moho reflects the crustal deformation in both areas induced by the Cenozoic India-Eurasia collision.

As mentioned above, continental reworking happens in broad regions including both active margins and continental interiors, even stable cratons. While reworking of continental margins is always associated with plate boundary processes of subduction or collision, reworking of intracontinental regions may or may not. The latter includes both compressional and extensional types, corresponding to mountain building and rifting, respectively, in regions that are far from plate boundaries. Compression-type intracontinental reworking has been reported worldwide, such in Central Asia, South China, Central Australia, etc., and is characterized by basin inversion, ductile shearing, middle-high temperature metamorphism and widespread magmatism (Raimondo et al., 2010; Ziegler et al., 1998). This type of reworking is typically related to plate boundary processes. The Cenozoic Tian Shan range, which is located within the Eurasian plate and originally developed in the Paleozoic as part of the Central Asian Orogenic Belt, experienced intense deformation and exhumation as a result of the India-Eurasia collision (Figure 18, Avouac et al., 1993; Tapponnier & Molnar, 1977, 1979). In South China, the Early Mesozoic Paleo-pacific subduction reactivated and deformed a large area around the Neoproterozoic suture zone in between the Yangtze craton and the Cathaysian block to form the Xuefengshan orogenic belt (Chu et al., 2019). Central Australia also contains valuable records of continental reactivation and reworking. The late Neoproterozoic to early Paleozoic Petermann Orogen and the mid-Paleozoic Alice Springs Orogen there were all characterized by strain localization, crustal thickening and subsequent exhumation of mid-lower crust, accompanying spatial and temporal changes in the thermal regime of the lithosphere (Aitken et al., 2019; Hand & Sandiford, 1999; Sandiford & Hand, 1998), which overprinted the complex Paleo- and Mesoproterozoic orogenic belts of the region (Holdsworth et al., 2001).

Continental reworking associated with extensional deformation has also been observed widely within continental interiors, some of which are temporarily correlated with plate boundary processes that are dominated, however, by compressional deformation. For example, during the late Eocene to Oligocene coeval with the Alpine and Pyrenean orogens formed in western Europe was the development of a large rift system in the foreland of the Alps with Variscan basement (European Cenozoic Rift System, ECRIS), extending over 1100 km from the North Sea coast to the Mediterranean (Dèzes et al., 2004). Similarly, intracontinental rift systems have developed in a large area within Asia as the India-Eurasia collision proceeded in the late Cenozoic, including the N-S trending rifts in southern Tibet (Armijo et al., 1986), circum-Ordos rift systems in central China (Zhang et al., 1998), and Baikal rift system at the boundary between the stable Siberian craton and the deforming Baikal-Sayan fold belt (Logatchev & Florensov, 1978). The relationships between these intracontinental rift systems and the continental collisional processes have not been well understood, and whether or not a mantle plume regime is required for the formation of the rifts remains debated (e.g., Lebedev et al., 2006; Merle & Michon, 2001; Molnar & Tapponnier, 1975; Ziegler & Dèzes, 2007).

The continental lithosphere can also be significantly extended, faulted, metasomatized and weakened during either passive or active rifting processes. The present-day Atlantic passive margins are commonly characterized by increasingly high degrees of structural extension seaward, in many cases presenting a zone of marked crustal thinning (with thickness < 15-20 km) of more than 100 km wide in between the oceanic and normal continental crust (Peron-Pinvidic et al., 2013). Particularly near the ocean-continent transition, the hyperextended continental basement is severely thinned (up to <10 km) and composed of rocks of not only continental crust but also exhumed and variously serpentinized mantle as well as of magmatic intrusions and infiltrations, with faults cutting from the surface to the mantle. These features suggest that the presumably thick and strong pre-rifting intracontinental lithosphere has been modified and thinned or reworked by the rifting process. One typical example is from the Iberian margin of the southern North Atlantic that underwent extreme crustal thinning and mantle exhumation during seafloor spreading in the early Cretaceous, such that the rifting process overprinted the thermal, compositional and structural inheritance of the lithosphere from the earlier Variscan Orogeny (Sutra & Manatschal, 2012). Similar reworking was also proposed for other passive margins, such as those of the Indian plate and the South China Sea in association with rifting and seafloor spreading processes (Calvès et al., 2011; Gao et al., 2015). Generally, deep-rooted mantle plumes are considered to have played an important role in many cases of such continental rifting and subsequent reworking and final breakup under extensional regime, based on the spatial and temporary coincidence between LIPs and continental breakup (e.g., Buiter & Torsvik, 2014; Storey, 1995). There are, however, also intracontinental rifts that may have not succeed in splitting a continent apart (called aulacogen). Examples of failed rifts include the ~1100 Ma-old Midcontinent Rift System in the center of the North American continent (Green, 1983), Early Cambrian Southern Oklahoma aulacogen in the southern United States (Gilbert, 1983) and Late Devonian Dniepr-Donets aulacogen in the southern part of the East European craton (Stovba et al., 1996). These failed rifts all experienced crustal thinning and extensive volcanism possibly associated with plume impingement in the early stage and crustal thickening and basin inversion with compressional deformation and faulting during later evolution (Keller & Stephenson, 2007; Stein et al., 2015), presenting complex and long-lasting reworking processes. A common tectonic setting of these failed rifts is pericratonic, for they lie within cratons but near the margins of cratons (Keller & Stephenson, 2007). The thick and rigid cratonic lithosphere may have hindered rapid upwelling of plumes and probably contributed to the failure of continental rifting and breakup in these regions.

4.2.2. Crustal Reworking

The rock units and structures of continental crust record multiple tectono-thermal events from the early formation through the long-term evolution of continents that led to various degrees of creation, reworking and destruction of continental crust (Hawkesworth et al., 2016; Stern & Scholl, 2010). Felsic granites are widespread on Earth and are the key constituent of continental crust, making the Earth unique among other planets (Campbell & Taylor, 1983; Taylor, 1989; Wu, Li, et al., 2007). It is generally believed that Archean continental crust is stable due at least partially to its lower-density felsic component, including older Na-rich tonalite-trondhjemite-granodiorite (TTG) and younger K-rich granites (Campbell & Taylor, 1983). The K-rich granites, which occupy less than 10% of the surface areas in the Archean but have dominated the exposed upper crust since 2.5 Ga, are essentially the result of intra-crustal melting (Taylor & McLennan, 1981; Wu, Yang, et al., 2007) and thus a representative product of continental crust reworking, especially since the Archean. New geochemical observations also consistently reveal a marked decrease in the MgO content of the upper continental crust from >11 wt % before the Neoarchean to ~4 wt % at the end of Archean, reflecting a transition from a mafic upper continental crust to a felsic one like the present-day during the Neoarchean (Tang et al., 2016).

A recent combination of oxygen and hafnium isotope ratios in zircons of different ages (Dhuime et al., 2012) further reveal the ubiquitous global crustal reworking through time (Figure 19a), showing higher reworking rates (> 50% proportions of reworked crust versus newly created crust) from around the end of Archean to the present day (Figure 19b). The sharp increase in the degree of crustal reworking in the Neoarchean well coincides with the rapid compositional change of the upper continental crust, as well as the springing-up of UHT metamorphism records (Figure 19c). The significant reworking of continental crust in the Proterozoic and Phanerozoic accompanied the slowdown of continental growth (Figure 19d), probably associated with subduction-driven plate tectonics gaining maturity from the Late Archean to Paleoproterozoic and being dominant since then on the Earth's evolution (e.g., Dhuime et al., 2012; Hawkesworth et al., 2016; Sizova et al., 2010).

Crustal reworking not only affects the rock units and geochemical features, but also exerts significant impacts on the structure and properties of continental crust. For example, reworking typically led to the development of structural and compositional layering in the continental crust. It has been widely accepted that mature continental crust with an intermediate composition evolved from juvenile basaltic crust (as oceanic crust) by refining and differentiation processes, in which the occurrence of fractionation and anatectic melting give rise to the felsic granodioritic upper crust and residual and cumulate mafic gabbroic lower crust accompanied by loss of dense, pyroxene- and garnet-rich mafic to ultramafic material (Hacker et al., 2015; Stern & Scholl, 2010). Although continental crust formation and evolution may follow this general regime, the real processes are more sophisticated and diversified from region to region, resulting in complex layering and lateral variations in structure, composition as well as deformation of continental crust.

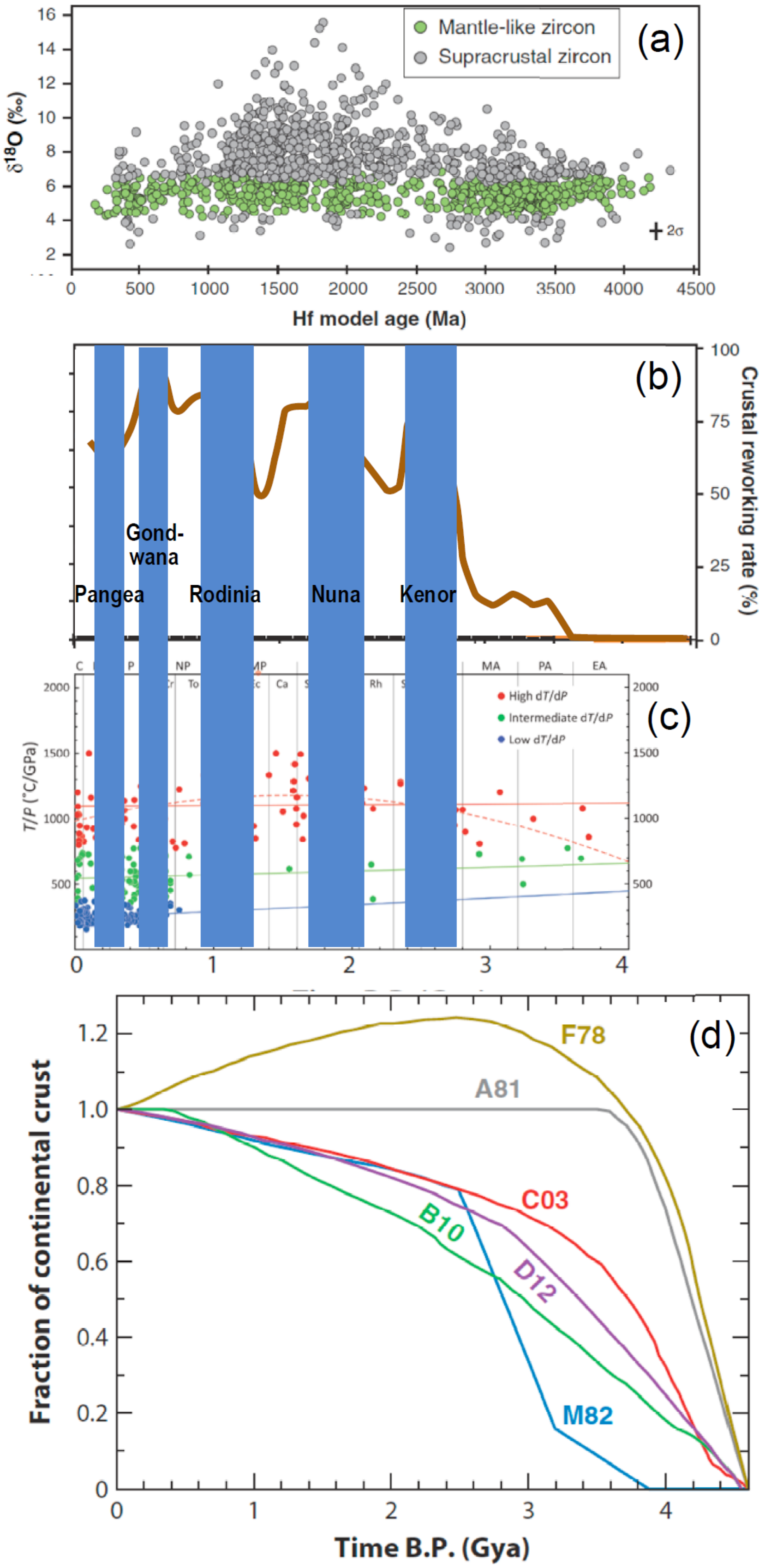

Figure 19. (a) $\delta^{18}O$ versus Hf model ages in 1376 detrital and inherited zircons from Australia, Eurasia, North America, and South America (Dhuime et al., 2012). The Hf model ages of mantle-like zircons record periods of new crust formation, whereas those of supracrustal zircons are referred to as hybrid model ages that record periods dominated by crustal reworking. (b) Variation in the reworking rates of continental crust through time calculated from the preserved proportions of reworked crust and new crust based on U-Pb and Hf analyses of zircons from Phanerozoic sediments (Dhuime et al., 2012); c) Metamorphic thermal gradient (T/P) versus age (Brown & Johnson, 2018). Three types of metamorphism are considered: high dT/dP or UHT (red), intermediate dT/dP (green), and low dT/dP or UHP (blue). Shaded areas mark the ages of supercontinents (from Hawkesworth et al., 2016). (d) Models of continental crust growth from selected literatures (Korenaga, 2013). F78 (Fyfe, 1978), A81 (Armstrong et al., 1981), M82 (McLennan & Taylor, 1982), C03 (Campbell, 2003), B10 (Belousova et al., 2010), and D12 (Dhuime et al., 2012). B.P., before present. Note that slow or even negative crust growth rates (slops of the growth curves) in Proterozoic and Phanerozoic correspond to higher reworking rates, as exemplified in (b).

For instance, anomalous low-velocity zones (LVZs) have been widely documented in the mid-lower crust of various tectonic settings (Figure 20a-c). In active orogens such as the Himalayan-Tibetan orogen and the central Andes and young tectonic belts or volcanic areas such as the Yellowstone and northeastern China, crustal LVZs are commonly attributed to the presence of partial melts and/or aqueous fluids associated with crustal shortening and thickening (e.g., Beck & Zandt, 2002; Liu et al., 2014; Nelson et al., 1996; Yuan et al., 2000) or induced by intensive heating from mantle (e.g., Fan & Chen, 2019; Huang et al., 2015). Beneath stable cratons or old orogens, on the other hand, partial melts or aqueous fluids are not favored to explain the observed LVZs in the mid-lower crust given the in situ low temperatures and mobility of melts and fluids. Instead, felsic composition resulting from crystallization of granitic intrusions by crustal remelting associated with ancient tectono-thermal events was thought to be more plausibly responsible for the intra-crustal LVZs in these regions (Chen et al., 2015; Yuan & Bodin, 2018). High-velocity zones (HVZs) have also been observed in continental crust of various ages, especially in the lower crust, often accompanying a transitional Moho character (Figure 20a,c-e). This structural feature is generally attributed to crustal reworking and vertical growth by mafic underplating and intrusion (Thybo & Artemieva, 2013), either during subduction-collision processes for continental assembly (e.g., Darbyshire et al., 2017; Petitjean et al., 2006; Yuan & Bodin, 2018), or associated with rifting to break-up of continents (Bronner et al., 2011; Delph & Porter, 2015; Mjelde et al., 2009; Thybo & Nielsen, 2012) or reactivation of continental interiors (e.g., Cheng et al., 2013; Zheng et al., 2006).

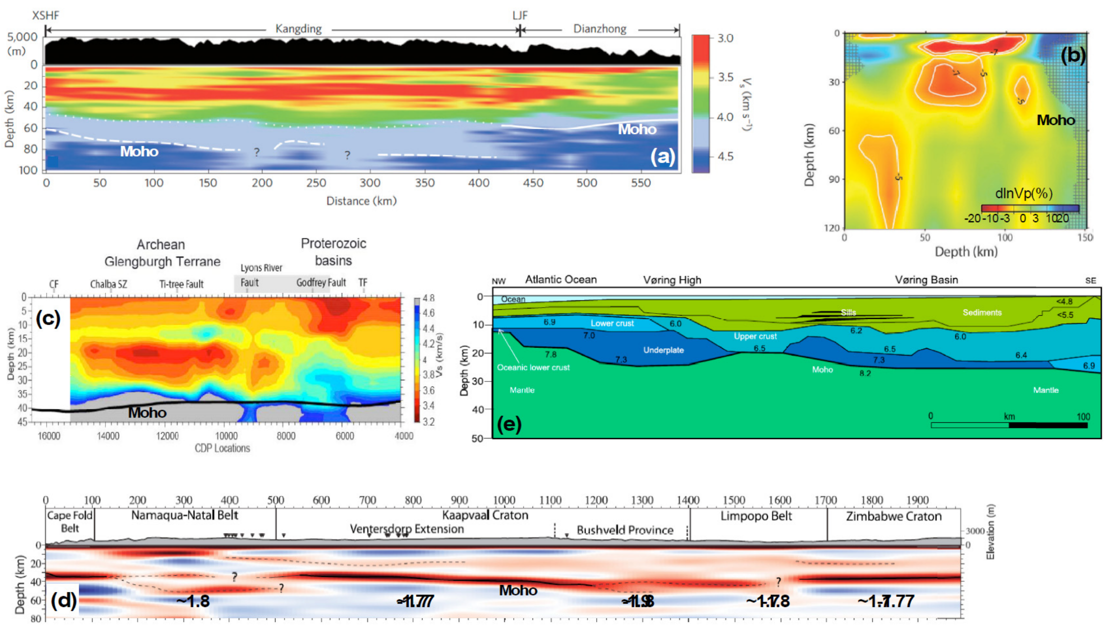

Figure 20. Crustal structures resulted from continental reworking. (a) NNW-SSE S-wave velocity cross section in Eastern Tibet (Liu et al., 2014), showing both a low-velocity zone (LVZ) in the mid crust and high-velocity zone (LVZ) in the lowermost crust (outlined by the white dotted and dashed lines). (b) NE-SW P-wave velocity cross-section traversing Yellowstone (modified from Huang et al., 2015), showing two LVZs in the upper and

mid-lower crust, respectively. These crustal LVZs are interpreted as separate magma reservoirs sourced from mantle. White lines denote the 5% and 7% P-wave velocity reduction contours. (c) S-wave velocity model along a NE-SW transect with annotated surface geological features in the western Australian craton (Yuan & Bodin, 2018). A slow midcrust is imaged beneath the Archean Glengburgh Terrance, whereas a fast lower crust is observed beneath the Paleoproterozoic basin area. SZ: shear zone. (d) NE-SW cross-section of receiver function stack and average crustal Vp/Vs ratios (numbers) in southern Africa (modified from Delph and Porter, 2015). The coincidence of thicker crust, higher Vp/Vs ratios and a gradual Moho discontinuity suggests the presences of HVZs in the lower crust beneath the Bushveld Province and Namaqua-Natal Belt. (e) P-wave velocity structure of the crust along a profile across the Vøring margin at the Norwegian shelf in the North Atlantic (Mjelde et al., 2009; Thybo & Artemieva, 2013). HVZs are observed at around the base of the crust.

Reworking of continental crust often involves significant deformation and metamorphism at depth, accompanying magmatic activities. Geophysical and geological studies suggest marked ductile deformation of the slow and weak mid-lower crust, either induced by channel flow (Bai et al., 2010; Liu et al., 2014; Royden et al., 1997; Searle et al., 2011) or from shearing along a midcrustal detachment (e.g., Gao et al., 2019; Klemperer, 2006), or partial eclogitization of the lowermost crust (e.g., Hetényi et al., 2007; Z. Zhang et al., 2014) at different areas in the Tibetan plateau. Such kind of crustal reworking is attributed to the Cenozoic to present convergence and collision between the Indian and Eurasian plates, which has caused dramatic crustal thickening (Figure 18), strong mantle deformation and crust-mantle decoupling (e.g., Gao et al., 2019; C. Wu et al., 2019). Recent seismic observations also reveal thicker crust (>40 km up to 60-70 km), marked crustal layering in both structure and deformation and a variable Moho character beneath both the Mesoproterozoic Grenville and Paleoproterozoic Trans-Hudson orogens. It was then proposed that mid-lower crustal flow may also have taken place in these older orogens, similar to that beneath the present-day Tibetan plateau, probably associated with the extensional collapse of the orogenic plateaus during the Proterozoic continental assembly processes (Darbyshire et al., 2017; Pawlak et al., 2012; Petrescu et al., 2016). Seismic and gravity data further suggest that at least parts of the thickened crustal root beneath these and other Proterozoic orogens or tectonic zones and some Paleozoic orogens (such as the Ural Mountains) may have undergone high-grade metamorphism even partial eclogitization (Darbyshire et al., 2017; Delph & Porter, 2015; Fischer, 2002; French et al., 2009; Petrescu et al., 2016), which may have been preserved (e.g., Figure 20d) under fluid-absent conditions and/or possibly with the aid of strong subcontinental lithospheric mantle (Leech, 2001). Such a deformation style of crustal thickening and flow in older orogens is comparable to that inferred beneath the Cenozoic to present Tibetan plateau, but appears distinct from the processes in the Archean time. Lower crustal flow was speculated to also exist in the Archean, but thought to have presumably prevented significant thickening of the hotter and weaker crust and led to a relatively shallow and flat Moho at ~35-40 km depths (Calvert & Doublier, 2018; Rey & Houseman, 2006), as observed today beneath Archean cratons without significant later crustal deformation (Kaapvaal and Zimbabwe cratons in Figure 20d). That is, continental crust may have never been noticeably thickened during its Archean evolution.

Metamorphism and deformation during crustal reworking is probably best represented by UHP or UHT metamorphism and relevant deformation that crustal rocks experienced exclusively at convergent margins associated with subduction to collision orogenesis. UHP metamorphism, which registers low thermal gradients (≤ 10-12ºC/km or 350-375ºC/GPa) and deep subduction of continental crust, was documented mostly in Neoproterozoic to Phanerozoic orogens such as Himalaya-Tibet (Guillot et al., 2008)and references therein), Alps (Chopin, 1984), Dabie-Sulu (Wang et al., 1989; Xu et al., 1992), Urals (Leech, 2001), etc. On the other hand, UHT metamorphism, the most thermally extreme type of crustal metamorphism (temperatures ≥ 900ºC, thermal gradients ≥ 20-25 ºC/km or 750-775ºC/GPa), occurred throughout much of the Earth history, with ages mainly ranging from Neoarchean to Cenozoic (Figure 19c, Brown, 2006; Brown & Johnson, 2018; Clark et al., 2011; Kelsey, 2008). Recently, growing geochemical evidence suggests that the deep crust of Precambrian UHT metamorphic terranes could have persisted with the presence of melts under suprasolidus temperatures for >60-200 Ma, accompanying slow cooling and pervasive deformation (Clark et al., 2018; Horton et al., 2016; Jiao et al., 2020; Taylor et al., 2020; Walsh et al., 2015). This is distinctly different from the Phanerozoic UHP metamorphism that sustained only about 10-20 Ma with rapid exhumation (e.g., Ernst et al., 1997). The fact that UHT and UHP metamorphism teemed in the Neoarchean and Neoproterozoic, respectively (Figure 19c) and the contrasting features between the two may mark significant transitions of tectonic regime at around the end of the Archean and Proterozoic, respectively (e.g., Brown, 2006; Dhuime et al., 2012; Zheng & Zhao, 2020), consistent with the secular variations in the rates of continental crust reworking and growth (Figure 19b and 19d).

4.2.3 Reworking of subcontinental lithospheric mantle

Continental reworking not only affects the structure and properties of the crust, but also changes the characteristics and even nature of subcontinental lithospheric mantle. Reworking of the continental lithospheric mantle is dominantly reflected with ubiquitous modifications in its composition through melt or fluid-induced metasomatism. This is exemplified by the widely varying major element compositions of mantle peridotites in both old cratons (kimberlite-borne xenoliths) and orogenic belts (basalt-borne xenoliths) (Figure 21, Tang, Zhang, Ying, & Su, 2013 and references therein).

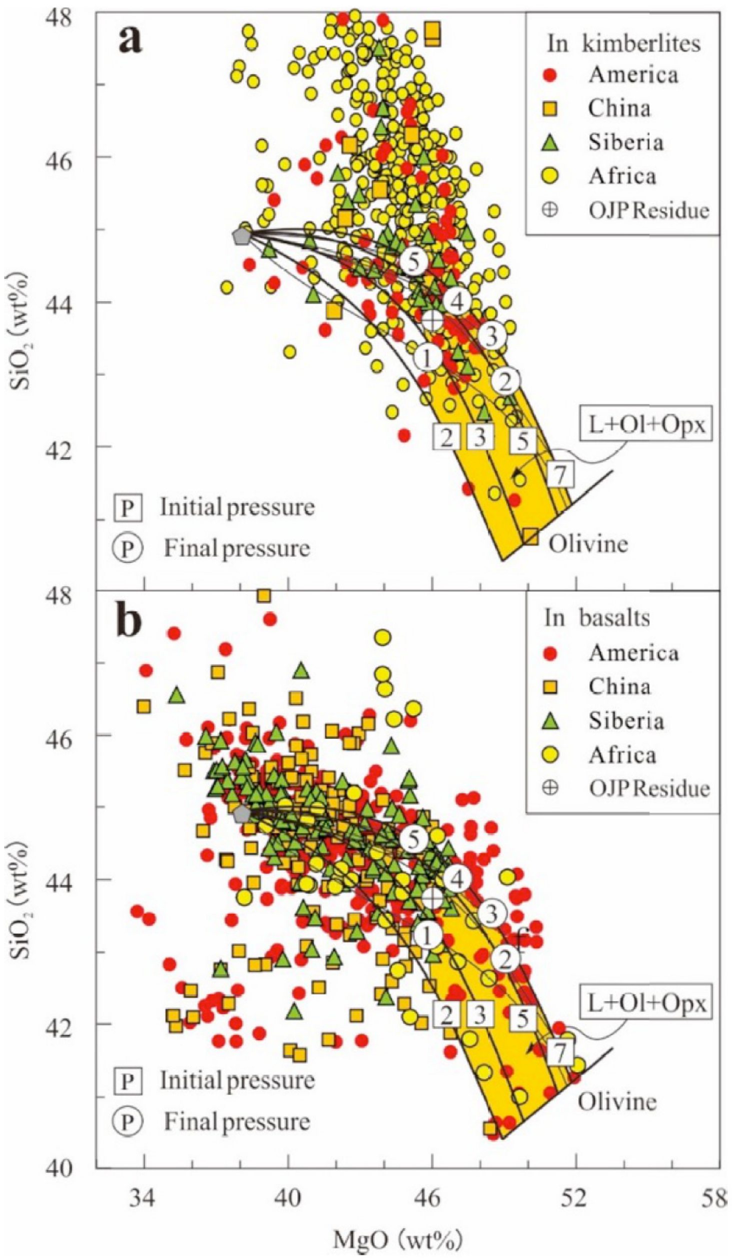

Figure 21. SiO_2 and MgO contents of peridotites in global continental lithospheric mantle. Data sources: Peridotite xenoliths in kimberlites and basalts (Doucet et al., 2012; Gornova et al., 2013; Tang, Zhang, Ying, & Su, 2013 and references therein; Howarth et al., 2014; Lin et al., 2019; Xiao et al., 2013; Yang et al., 2018; Zou et al., 2016). Bold lines labeled with squares, initial melting pressures; light lines labeled with circles, final melting pressures; brown shaded area, compositions of residual harzburgite designated as [L+Ol+Opx] after melting of lherzolite (Herzberg, 2004). The curves are individual melting curves of normal mantle peridotite, and the area covered by the melting curves gives the composition range of

residual phases after partial melting of normal mantle peridotite under different temperature and pressure conditions. The data points falling within the range of partial melting curves denote the residual of partial melting of normal mantle peridotite, whereas those falling outside the melting range do not represent the residual of normal mantle melting, but reflect the influence of melt metasomatism.

In particular in the typical ancient cratons of South Africa, Siberia and North America, mantle peridotites have a large range of SiO_2 contents with a significant portion being much higher than the composition range of normal mantle peridotites (Figure 21a). Such a feature indicates that these peridotites are not the pure remnants of partial melting of normal mantle peridotites, but have undergone various degrees of metasomatism with the introduction of silica-rich fluids or melts, probably related to subduction (e.g., Bell et al., 2005; Simon et al., 2007). On the other hand, the common occurrence of kimberlites especially in Archean cratonic regions (Sparks, 2013) and the addition of clinopyroxene, lherzolitic garnet and phlogopite etc. within the lithospheric mantle suggest another kind of metasomatism by volatile-rich and silica-poor melts, mostly associated with rising plumes (e.g., Griffin et al., 2009; Simon et al., 2007).

The Re-Os isotopic data of multiple sulfides in the global mantle peridotites also support the view that the subcontinental lithospheric mantle is widespread modified by peridotite-melt reaction. A large number of data show that most of the mantle peridotite xenoliths from the cratonic lithospheric mantle of Archean age have younger Re-depletion ages of Proterozoic to Cenozoic, which is well correlated with the propensity of Fo (Mg number) reduction in olivine (Figure 22). Both observations consistently reflect refertilization of the subcontinental lithospheric mantle by peridotite-melt reaction, with addition of Fe, Ca Al and Re to originally depleted protoliths (e.g., An et al., 2017; Aulbach et al., 2004; Pearson, 1999; Smith

& Boyd, 1987; Xu et al., 2008; Zhang et al., 2012). Therefore, the re-depletion model ages of these peridotites do not represent the real formation ages of mantle peridotites, but are the result of the multiple-phases of interactions between Archean peridotites and melts (Shu et al., 2019; Zhang, 2009).

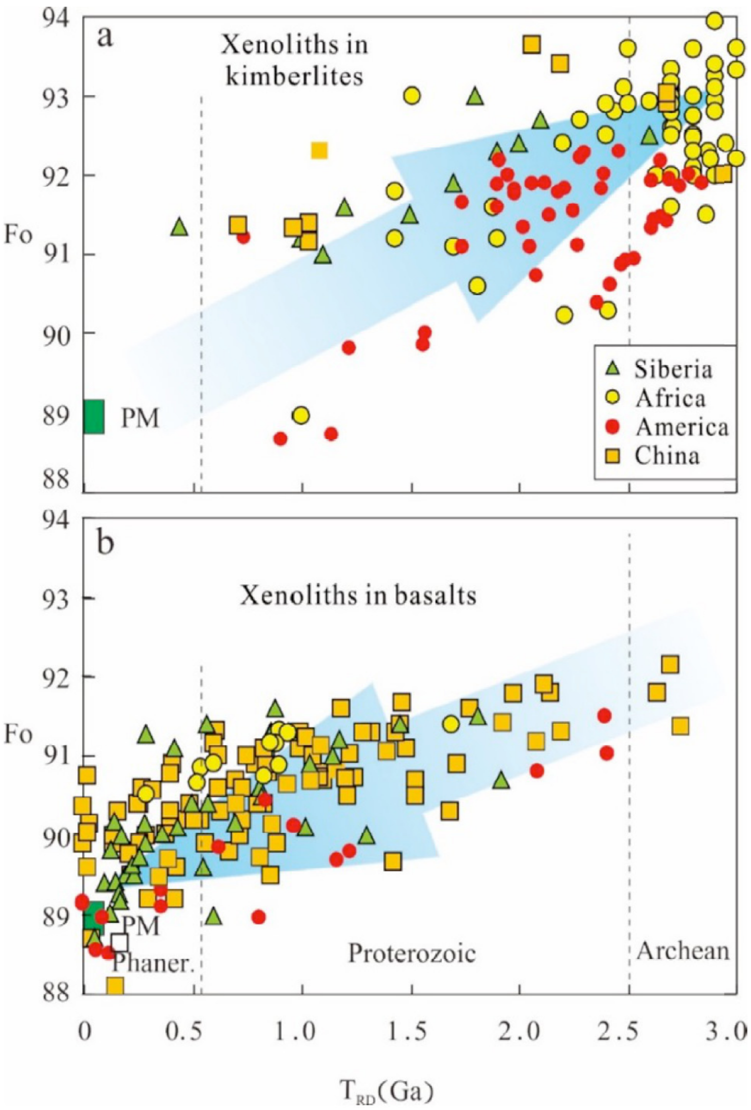

Figure 22. Relationship between Re-depletion model age and Fo in olivine of peridotites in global continental lithospheric mantle. Data sources: (Liu et al., 2015; Tang, Zhang, Ying, Su, et al., 2013 and references therein). PM represents the primitive upper mantle.

The refertilization of the ancient depleted lithospheric mantle is mainly a bottom-up process as the upwelling of fertile asthenospheric material, and likely follows previously existing lithospheric weaknesses such as suture zones between ancient blocks or breaks in the Archean root (Foley, 2008; O'Reilly et al., 2001; Xiao & Zhang, 2011). Such kind of refertilization is indeed an asthenospherization of lower parts of the lithospheric mantle, which may be controlled by the topography of the lithosphere base (Foley, 2008). This process could result in a change in geochemical composition and properties (e.g., buoyancy, rheological strength, etc.) of the reworked part of the subcontinental lithospheric mantle, and thus potentially affect the stability and evolution of continents (e.g., Carlson et al., 2004; Chesley et al., 1999; Griffin et al., 2009).

The widespread occurrence of reworking on the subcontinental lithospheric mantle via melt or fluid-induced metasomatism inevitably leads to enhanced water contents in the lithospheric mantle beneath continents (Peslier et al., 2017; Xia & Hao, 2013). This is evidenced by the mantle xenolith data, showing a large range of water concentrations in olivine for peridotites from several cratonic regions (Figure 10a). Although the water contents of olivine at near the base of the lithosphere (~200-km depth) in these regions are generally lower than that estimated for the asthenosphere (<10 ppm vs. ~30-70 ppm, Table 1), there do have cases of high water concentrations, such as locally beneath the Siberian craton (Figure 23). At shallower depths, i.e., ~120-150 km, anomalously high water contents (>50 ppm) in olivine appear within the lithospheric mantle of both the Kaapvaal and Siberian cratons, well exceeding the amount for the asthenosphere (Figure 23a) and inconsistent with the presumed dehydration nature of cratonic roots. The elevated water contents of olivine display positive trends with indices of metasomatism (such as modal proportion of clinopyroxene, Ni and Ti contents in peridotites) (Doucet et al., 2014; Jean et al., 2016), suggesting that the corresponding mantle xenoliths represent water-rich metasomatized peridotites rather than that of the typically dehydrated cratonic mantle. On the other hand, the olivine from diamond inclusions in the mantle xenoliths from the Siberian craton has a much lower water content (Figure 23a), possibly due to the protection of diamonds for the inclusions from interactions with water-bearing metasomatic agents (Jean et al., 2016; Novella et al., 2015; Taylor et al., 2016). Therefore, water concentrations of olivine inclusions in diamonds are more representative for that of cratonic lithospheric mantle. Similar features are also observed for other peridotitic minerals, and the water contents in bulk-rock peridotites calculated from data measured for individual minerals show large differences in between metasomatized peridotites and those unaffected by metasomatism (Figure 23b).

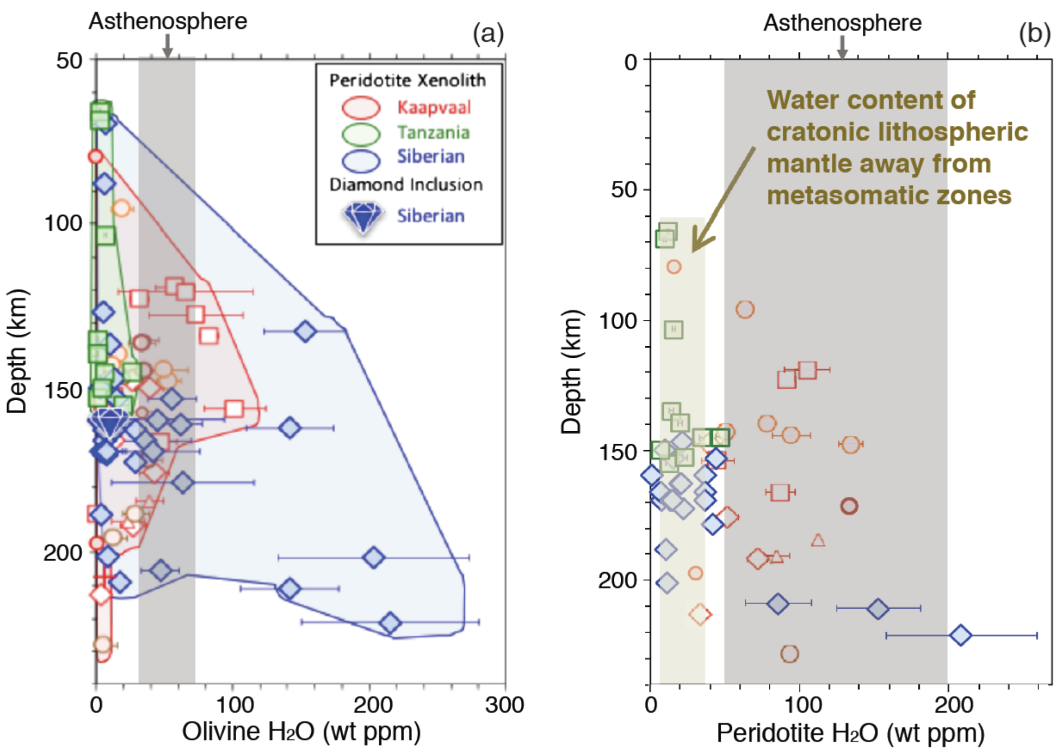

Figure 23. Water concentrations in olivine (a) and for bulk-rock peridotites (b) as a function of depth in the cratonic lithospheric for mantle xenoliths from the Kaapvaal, Siberian and Tanzanian cratons (modified from Peslier et al., 2017). Grey shaded areas represent the range of the corresponding water contents in the asthenosphere (Table 1).

The water contents from xenoliths within subcontinental lithospheric mantle appear highly variable from region to region and also heterogeneous vertically (Figure 23), indicating complex and varying degrees of mantle metasomatism and thus reworking. The metasomatism-induced addition of water may have effects on the strength and rheological behavior of subcontinental lithospheric mantle. Such effects lie largely on the water concentration in olivine, the most abundant mineral of the upper mantle (>50% in volume). This is because even a trace amount of water weakens olivine (Dixon et al., 2004; Faul et al., 2016; Hirth & Kohlstedt, 1996; Karato, 2010), but water does not weaken clinopyroxene, and its weakening effect on orthopyroxene is complex (Gavrilenko et al., 2010; Ohuchi et al., 2011) and not well understood (Demouchy & Bolfan-Casanova, 2016). The water content of garnet is very limited compared with that in other phases of mantle peridotite (Demouchy & Bolfan-Casanova, 2016), and thus often not considered when evaluating the strength of the lithospheric mantle. It is worth to note that, although the large range of water concentrations in olivine of mantle xenoliths (Figure 23a) suggests high heterogeneity and complexity of lithospheric mantle reworking, primarily associated with the melting and metasomatic history of the mantle sampled by the xenoliths, it may not be indicative of large differences in the overall strength and rigidity of the cratonic lithospheric mantle, as the cratons do not show an enhanced instability with water contents of olivine.

The typical cratonic lithospheric mantle is believed to remain relatively dry and strong (Peslier et al., 2017 and references therein). Taking the typical values of water contents in olivine from diamond inclusions (<10 ppm) for unaltered cratonic lithospheric mantle results in about one order of magnitude higher in viscosity at the base of the lithosphere than that of the asthenosphere (Figure 15c), consistent with the stability of most cratons. Moreover, seismic anisotropy observations show a dominant A-type fabric of olivine in the cratonic lithosphere (Baptiste & Tommasi, 2014; Gung et al., 2003; Mainprice & Silver, 1993). Laboratory experimental results suggest that the A-type fabric of olivine is generally associated with a dry environment (Karato et al., 2008). All these together provide evidence for the general dry nature of the cratonic mantle lithosphere, and also indicate that, despite its common appearance among cratons, reworking by melt or fluid-induced metasomatism may be localized, possibly at lithospheric weaknesses (Peslier et al., 2017), and thus not alter the nature of the majority of the cratonic lithosphere. It is only in rare cases that severe reworking results in extremely high water contents and significant weakening of a large part of cratonic lithospheric mantle, leading to fundamental destruction of a craton. For instance, it was reported that the lithospheric mantle of the eastern NCC was highly hydrated (>1000 ppm) and metasomatized in the Early Cretaceous (~120 Ma) (e.g., Xia et al., 2013), a time period when this part of the craton was destroyed with a loss of more than 100 km of its mantle root (see 4.2.3 and 4.3).

Complex metasomatism and reworking of continents is also manifested by the structural features and detailed fabrics of deformation in the lithospheric mantle. As mentioned before, continental lithosphere is characterized by large structural and rheological contrasts particularly between cratonic cores and surrounding mobile belts (4.1.3, Figure 16a-c). Such a characteristic represents the tectonic imprints of multiple phases of reworking mainly concentrated at weak zones in the lithosphere during the long-term evolution of continents, possibly superposed upon intrinsically different structural features associated with diverse continental formation processes (e.g., Begg et al., 2009; Chen, 2010; Schaeffer & Lebedev, 2014). Strong reworking especially in relatively weak continental lithosphere is evidenced by observations of localized softening and deformation to form mantle shear zones with various grain sizes in orogenic belts (e.g., Dickinson, 2004; Kaczmarek & Tommasi, 2011; Reuber et al., 1982), which often involves melt-rock reactions and fluid infiltration and thus further reduces the strength of the lithosphere (Dickinson, 2004). Moreover, the occurrence of earthquakes within the presumably strong and aseismic lithospheric mantle beneath continents, e.g., the 2013 Mw 4.8 Wyoming earthquake at ~75-km depth (Prieto et al., 2017), deep seismicity below the Moho both along the Newport-Inglewood fault in southern California (Inbal et al., 2016) and in the Himalayan collision zone (Schulte-Pelkum et al., 2019), probably all reflects on-going weakening and reworking of the subcontinental mantle lithosphere in these areas.

Mobile belts surrounding cratons or tectonic boundary zones, which are mechanically weaker than cratonic interiors, generally experience more intensive reworking and exhibit lower-velocities or lower-resistivities, with sharp structural changes or distinct deformation patterns in the deep lithosphere compared with adjacent cratonic nuclei (e.g., Figure 16a-c). Experimental and theoretical studies indicate that, lithospheric weak zones could be long-lived structures, due to the extremely slow grain growth and healing of mantle rocks after being weakened (several 100 Myrs or more) compared to the fast weakening processes (in 1 Myrs or less) (e.g., Bercovici & Ricard, 2012; Chu & Korenaga, 2012). Therefore, pre-existing structures in the lithosphere have been the primary control on the tectono-thermal evolution and reworking of continents (Audet & Bürgmann, 2011; Buiter & Torsvik, 2014; Frizon de Lamotte et al., 2015; Heron et al., 2016; Thomas, 2006). The rheological weakness and mechanically anisotropic behavior of the pre-existing structures make them function as the sites of strain localization and intensive heating in the lithosphere during tectono-thermal events, leading to further weakening of these zones (Thomas, 2006; Tommasi et al., 2001; Vauchez et al., 1998). Such effects may repeatedly intensify the contrast in the lithospheric structure and properties across cratonic margins or tectonic boundary zones, and are considered to be responsible for the distinct present-day structural heterogeneities of continental lithosphere (Figure 16; Keranen & Klemperer, 2008; Schaeffer & Lebedev, 2014; Snyder et al., 2017; Zhu et al., 2011). This also well explains the observations that, during episodic assembly and break-up of continents in association with the closure and opening of ocean basins, continental margins are repeatedly reworked with focused deformation and magmatism during subduction, collision and rifting, whereas continental cores remain largely intact (Buiter & Torsvik, 2014; Frizon de Lamotte et al., 2015; Houseman & Molnar, 2001; Li et al., 2008).

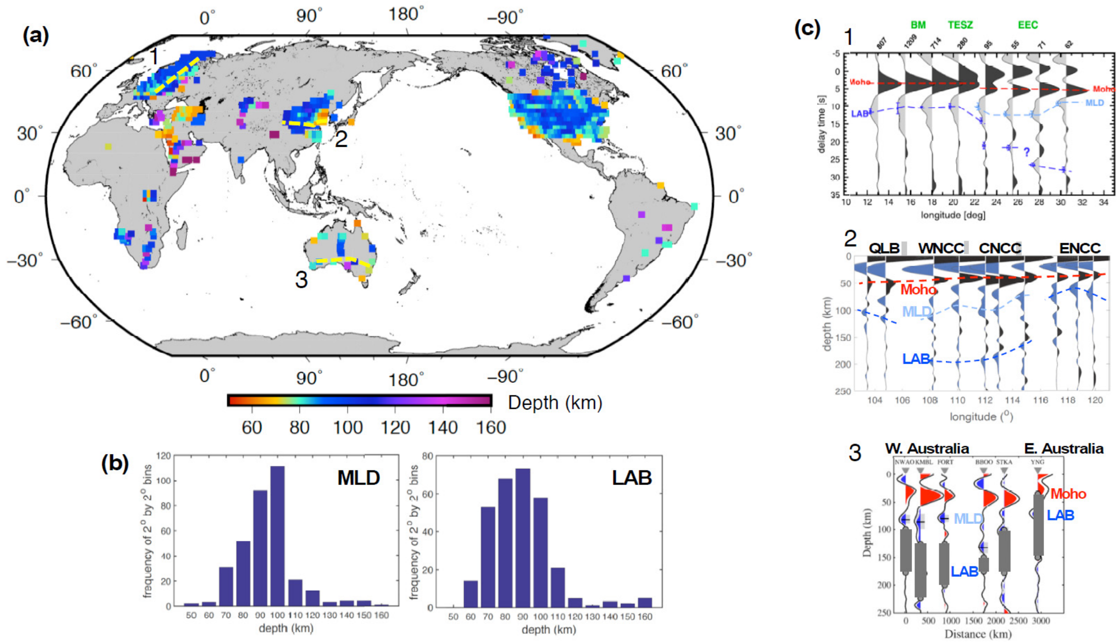

Figure 24. (a) Depth map of the discontinuity with a downward velocity decrease in the shallow upper mantle (≤160 km) beneath continents (modified from Chen, 2017). Only results from teleseismic converted wave (Sp or Ps) studies are considered to ensure similar resolutions, and recent studies, i.e., Yuan et al. (2017) for northwestern Namibia, Kennett and Sippl (2018) for central Australia, Chen et al. (2018) and Hopper and Fischer (2018) for U.S. are complemented to and Y. Zhang et al. (2019) for the North China Craton is used to update the dataset of Chen (2017). The locations of the three cross-sections in (c) are shown as yellow dashed lines; b) Spatial frequency of the depths in (a) but only for the interpreted MLD beneath cratonic regions (left) and the LAB beneath tectonically active regions (right); c) Three cross-sections of Sp waveforms: NE-SW oriented across the Trans-European Suture Zone (TESZ) (Knapmeyer-Endrun et al., 2017), E-W oriented across the North China Craton (modified from Chen et al., 2014), and E-W oriented in southern Australia

(modified from Ford et al., 2010). Moho, MLD and LAB are marked. These cross-sections show similar depths of the cratonic MLD and the LAB in adjacent tectonic regions.

In addition to reworking of cratonic margins and tectonic boundary zones that gives rise to significant lateral structural heterogeneities, continental evolution and reworking also result in marked vertical variations or layering of the structure and properties in not only the crust (e.g., Figure 20) but also the subcontinental lithospheric mantle. Recent seismic studies consistently reveal a sharp velocity reduction in a depth range of ~60-150 km but mostly clustered at ~70-100 km depth within the thick mantle root beneath cratonic regions (Figure 24; Chen, 2017; Fischer et al., 2010; Selway et al., 2015 and references therein). The corresponding velocity discontinuity is thus named the mid-lithospheric discontinuity (MLD), marking the top of a relatively low-velocity layer (LVL) within the overall high-velocity cratonic lithospheric mantle. The common observations of the MLD and the underlying LVL beneath cratons indicates pronounced vertical structural variations or layering of continental lithospheric mantle (Chen, 2017; Yuan & Romanowicz, 2019 and references therein). The MLD is featured as a rather sharp discontinuity, with a shear-wave velocity reduction of several to over 10% generally within a 30-40 km depth range (Selway et al., 2015).

In cases, it was reported that the velocity drop at the MLD broadly coincides with vertical variations in other physical (e.g., seismic anisotropy, electrical resistance) or chemical properties (e.g., petrological characteristics) (e.g., Chen et al., 2009; Fichtner et al., 2010; Wirth & Long, 2014), possibly reflecting a same origin. In particular in the Kaapvaal craton, the structural layering of the lithospheric mantle observed seismically (Sodoudi et al., 2013) appears correlated with the vertical variations of both Mg# in olivine and shear-wave velocity calculated from whole-rock composition (Griffin et al., 2009), and also of the water content in olivine that appears high (> 50 ppm) at ~4-5 GPa (~120-150 km depth) and obviously decreases below ~5.5 GPa to almost dry (<10 ppm) at the base of the lithosphere (Figure 10a, Peslier et al., 2010, 2017). Collectively, the structural feature of the MLD was typically attributed to the accumulations of hydrous or other volatile-rich minerals that are particularly stable at the MLD depths through melting or fluid-induced metasomatism (e.g., Aulbach et al., 2017; Saha et al., 2018; Selway et al., 2015), although other mechanisms were also proposed (e.g., Karato et al., 2015; Rader et al., 2015; Thybo, 2006). In combination with the geochemical observations showing widespread interactions between Archean peridotites and melts in cratonic regions (Figures 21-22), it may be reasonable to speculate that the velocity reduction and probably also other property variations at the MLD mainly results from the long-term multiple stages of refertilization of the lithospheric mantle, as previously suggested (Aulbach et al., 2017; Griffin et al., 2009). It is possible that inherited fabrics related to the Archean formation of continental lithosphere may also contribute (e.g., Chen et al., 2009; Lee et al., 2011; Rader et al., 2015).

It is also interesting to note that both the depths and structural feature of the MLD beneath cratons are comparable to those of the LAB in tectonically active regions (Calò et al., 2016; Chen et al., 2014; Foster et al., 2014; Rychert et al., 2010; Thybo, 2006), even similar to the LAB beneath oceanic lithosphere (e.g., Kawakatsu et al., 2009; Olugboji et al., 2016). Such observations probably reflect a genetic link between these discontinuities (Chen, 2017). A number of models have been proposed to relate of the cratonic MLD with the LAB beneath tectonic regions, such that a new and shallow LAB forms following the original MLD after the removal of the lower part of a weakened cratonic mantle root (Aulbach et al., 2017; Chen et al., 2014), or that the LAB of a young and thin lithosphere gradually transforms to a MLD as the lithosphere progressively cools and thickens (Rader et al., 2015). In either model, the development of the MLD or the LAB is closely linked to the evolution and reworking of continents, which is yet to be fully understood. In addition to the MLD commonly observed at ~70-100 km depth, recent studies also detect other deeper discontinuities with either velocity increase or decrease with depth within or immediately below the continental lithospheric mantle (e.g., Calò et al., 2016; Sodoudi et al., 2013; Wu et al., 2020), implying more complicated vertical structural variations beneath continents.

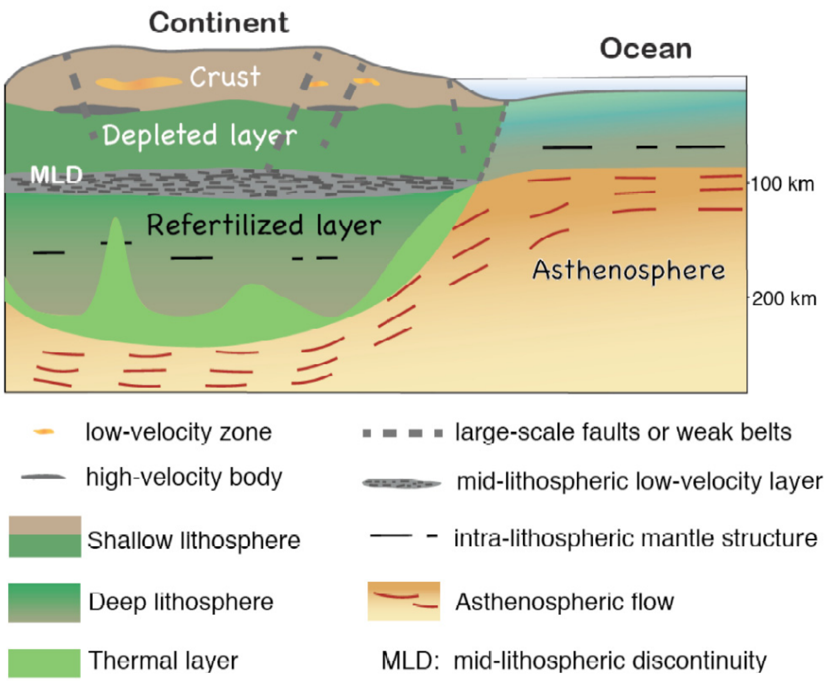

Figure 25. Conceptual model of continental lithosphere, showing lateral and vertical structural heterogeneities resulted from long-term evolution and reworking in comparison with oceanic lithosphere (modified from Yuan and Romanowicz, 2019).

Overall, it is the long history of evolution and particularly complex reworking of continents that result in the present-day striking structural heterogeneities, not only laterally but also vertically, of the continental lithosphere compared to its oceanic counterpart (Figure 25). In cratons, such heterogeneities likely mark mechanically weaknesses of various scales inside the overall strong cratonic lithosphere. The vertical heterogeneities, such as the weak layer topped by the MLD at ~70-100 km depth, may behavior similarly and play a similar role during deformation and magmatic processes as the laterally weak boundary zones discussed before (e.g., Chen, 2017; Hu et al., 2018). Both these vertical and lateral weaknesses, once created, could have combined to exert significant impacts on the ensuing evolution and facilitate reworking of cratons.

4.2.4 Temporal and Spatial Extent of Continental Reworking

Reworking of continents occurs widely in both time and space. The reworking processes are mostly associated with the operation of plate tectonics, as reflected by the sharp increase in the reworking rate of continental crust (Figure 19b) and the teeming of UHT metamorphism in the Neoarchean (Figure 19c), a time period during which plate tectonics was thought to have become vigorous (e.g., Dhuime et al., 2012; Hawkesworth et al., 2016; Sizova et al., 2010; Tang et al., 2016). The degree of continental crustal reworking appears to be well correlated with the evolution of supercontinents, showing increased reworking coeval with the supercontinent assembly (Figure 19b). The periods of intensified continental crust reworking during the supercontinent assembly witnessed the more frequent occurrence of deep crustal metamorphism (Figure 19c) and enhanced reworking of sedimentary materials as represented by elevated $\delta^{18}O$ of zircons (Spencer et al., 2014). Moreover, the U-Pb crystallization ages of detrital zircons over the Earth's history also peak at the same periods (Voice et al., 2011), which is considered to reflect better preservation of magmatic rocks generated in the late stage of subduction and collision associated with crustal thickening. These observations were interpreted to reflect more pronounced reworking of pre-existing continental crust during the periods of continental collision and development of supercontinents (Hawkesworth et al., 2016). On the other hand, more intensive magmatism with higher rates of juvenile crust generation happens during oceanic subduction before collision (Clift et al., 2009; Stern & Scholl, 2010). However, it has been widely documented that continental crust is also destroyed by erosion, subduction and in cases delamination along subduction zones, at rates comparable to, or even greater than the rates of new crust generation (Clift et al., 2009; Scholl & von Huene, 2009; Stern, 2011). Collectively, both destruction and reworking of continental crust, which reshape the continents, mostly occur in the processes of oceanic subduction and continental collision during the assembly of supercontinents. Considering that the protection role of subcontinental lithospheric mantle to the overlying crust and the importance of crust-mantle coupling for the longevity of continents (see 4.1.2), it is expected that reworking and destruction of the subcontinental lithospheric mantle would be almost simultaneous to that of the continental crust.

Spatially, reworking of continents preferentially takes place at continental margins, mostly along subduction and collision zones as indicated above, and to a lesser extent also happens at intracontinental zones and rifting margins. These reworked zones are usually pre-existing weaknesses (e.g., faults, ancient sutures) of continental lithosphere (Audet & Bürgmann, 2011; Bercovici & Ricard, 2012; Buiter & Torsvik, 2014; Frizon de Lamotte et al., 2015; Heron et al., 2016; Holdsworth et al., 2001), as demonstrated before. During plate tectonic processes, the lithospheric fabrics and properties of the continental margins could be significantly altered, whereas the cratonic interior surrounded by these marginal zones are usually not noticeably affected. For example, the major part of the cratons in North America and South America remain rigid, thick and stable despite long periods of oceanic subduction to the west (Figure 16a, Carlson et al., 2005). Continental rifting also mostly occurs along tectonic belts that wrap around cratonic cores, such as the two branches of the East Africa rift system encircling the Tanzanian Craton (Fletcher et al., 2018), the Yinchuan-Hetao and Fenwei rift systems surrounding the stable Ordos block of the western NCC (Zhang et al., 1998), and the Baikal rift developed at around the southern boundary of the Siberian Craton (Logatchev & Florensov, 1978). And in many, if not all, cases of oceanic opening, for examples for the Atlantic and Indian Oceans, passive margins break up often along former sutures with or without the help from rising plumes (Buiter & Torsvik, 2014; Frizon de Lamotte et al., 2015). Moreover, as mentioned before, although reworking of the lithospheric mantle by melt- or fluid-induced metasomatism is commonly observed and results in elevated water contents in stable cratons (Figure 23), it does not obviously affect the overall nature of the cratonic lithosphere. This could be because such kind of metasomatism more focuses at pre-existing weaknesses (Foley, 2008; Peslier et al., 2017) or intralithospheric depths, such as the MLD layer that does not directly contact with and thus is not easily affected and weakened by the convecting asthenosphere (Figure 25).

Although continental reworking is usually concentrated at lithospheric weaknesses, in specific cases it does happen in cratonic regions, resulting in reactivation and modification, even destruction of the presumably strong cratonic lithosphere. Following previous studies (e.g., Carlson et al., 2005; Wu et al., 2008; Zhu et al., 2011), here the phrase of craton destruction or decratonization refers to a kind of geological phenomena that both the lithospheric mantle and the crust of a craton are severely reworked, such that the craton loses its intrinsic stability. As demonstrated in 4.2.1, the western North American craton especially Wyoming craton, the Brazilian craton in South America and the NCC in East Asia are such examples that have experienced at least partial destruction since the Mesozoic.

The NCC is considered to be the most significantly destroyed craton in the world (Carlson et al., 2005; Zhu et al., 2011). After its formation at ~1.85 Ga by the amalgamation of the eastern and western blocks (eastern and western NCC) of Archean ages along the Trans-North China Orogen (central NCC) (Zhao et al., 2001, 2005), the NCC was not noticeably disturbed and remained stable like other typical cratons until the Mesozoic. However, severe reworking of the eastern NCC, as manifested by significant lithospheric extension with the formation of widespread metamorphic core complexes, pull-part basins and grabens, intensive magmatism and large-scale gold mineralization, took place in the early Cretaceous (~135-115

Ma), leading to vital destruction of this part of the craton (e.g., F. Y. Wu et al., 2019; Zhu et al., 2012a). It was suggested that the eastern NCC lost more than 100 km of its mantle root and its crust was also thinned, chemically modified and weakened during the destruction, whereas the lithosphere of the central and western NCC was only locally thinned and modified and has largely retained its cratonic nature over the long-term evolution (e.g., Chen, 2010; Xu, 2001; Zhu et al., 2011).

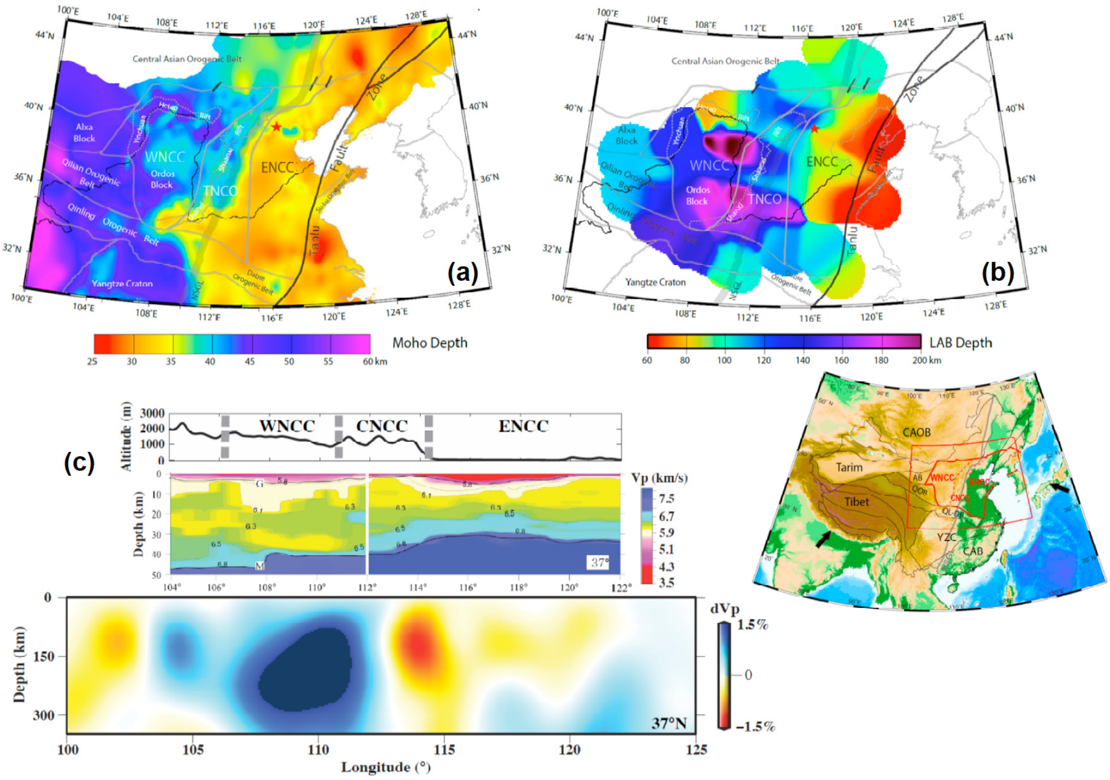

Figure 26. Depth distributions of the Moho (a) and the LAB (b) in the NCC and crustal and upper mantle P-wave velocity structure together with altitude along 37°N latitude across the craton (c), showing large differences in crustal and lithospheric thickness and structure and topography as well between the eastern NCC (ENCC) and the central (CNCC or TNCO) - western NCC (WNCC). The Moho depth data used in (a) are from Wei et al. (2015, 2016); b) is modified from Y. Zhang et al. (2019). The crustal and upper mantle velocity models in (c) are from Zheng et al. (2017) and Xu et al. (2018), respectively. G – base of sediments; M – Moho. Inset shows the distinct E-W difference in topography (low land in the east and plateaus or mountains in the west separated roughly by the North-South Gravity Lineament,

or in brief, NSGL) in eastern China, which broadly coincides with the deep structural variations as observed in the NCC (a-c). Red lines outline the NCC and the spatial range of (a,b). AB – Alxa block; CAB – Cathaysia block; CAOB – Central Asian Orogenic Belt; QL-DB – Qinling-Dabie Orogenic Belt; QOB – Qilian Orogenic Belt; YZC – Yangtze Craton.

Today the eastern NCC shows a thin crust (<35 km) and lithosphere (mostly <100 km) with a slow, fertile and young lithospheric mantle, in contrast to the thicker crust (~35-50 km) and generally fast, thick and stable mantle root with localized thinning and modification in the central and western NCC (Figure 26a-c) and also differing from the crustal and lithospheric structure of typical cratons in the world (e.g., Table 1). It is interesting to note that the characteristic variation of surface topography is broadly concordant with the E-W difference in the crustal and lithospheric structure (Figure 26) and associated spatially uneven reworking and destruction of the NCC (Chen, 2010; Xu, 2001; Zhu et al., 2011), suggesting the involvement of the entire lithosphere from the surface to the base during the craton destruction.

The Wyoming craton in western U.S. was also reported to have considerably reworked and partially destroyed during the Late Cretaceous to Paleocene Laramide orogeny and later processes, with a significant portion of its lower lithosphere eroded and partially replaced by juvenile mantle (Figure 16a-b; Carlson et al., 2004; Dave & Li, 2016; Snyder et al., 2017). Similarly, the lithosphere of the western Brazilian craton was documented to has been thinned and destroyed associated with the Cenozoic Nazca plate subduction (Beck & Zandt, 2002), but less information is available.

The cratons that have experienced various degrees of destruction since the Mesozoic are all along the circum-Pacific subduction zones, and the reworking and destruction of their cratonic lithosphere are suggested to be triggered largely by oceanic subduction (e.g., Beck & Zandt, 2002; Snyder et al., 2017; F. Y. Wu et al., 2019; Zhu et al., 2011). This will be discussed later in 4.3. Recent studies on the Wyoming craton and NCC suggest that the destruction of these cratons may have been closely related to a rapid change of subduction regime from an early flat or shallow-angle subduction to a later steep subduction following fast trench retreat, accompanying intensive lithospheric deformation from compressional to dominantly extensional, infiltration of large volumes of fluids and voluminous magmatism on the overlying continental lithosphere (e.g., F. Y. Wu et al., 2019; Yonkee & Weil, 2015). Although the details need further investigation, it indicates that destruction of cratons may not take place casually and evenly in time and space, but probably happen with specific conditions or features of plate tectonics, particularly of oceanic subduction. That means that craton destruction would be temporarily and spatially confined, consistent with less such observations than those of the long-term stability and preservation of cratons (Carlson et al., 2005).

4.3. Mechanisms of Continental Reworking

Two main mechanisms or dynamic triggers have been proposed for the reworking of continents, in particular cratons. One is oceanic subduction, and the other is mantle plume activity, representing processes at plate boundaries and underneath plates, respectively. The key factors dominating these processes and their effects on the cratonic stability are, however, still controversial.

4.3.1. Mantle Plume Perspective

It has been proposed that mantle plume impingement can cause thinning, refertilization and destabilization of the continental lithosphere through thermo-mechanical-magmatic erosion (e.g., Foley, 2008; Lee et al., 2011). Numerical simulation studies suggest that a typical thick Archean lithosphere cannot be thinned significantly via thermomechanical erosion by a hot mantle plume over a short period of time (<100-200 Ma, Petitjean et al., 2006; Wang et al., 2015). For a more effective lithospheric erosion and refertilization, sufficient amounts of melts are required. However, such a process is largely suppressed beneath cratons with thick lithospheric roots. This is because 1) mantle plumes are preferentially deflected away to areas of thin lithosphere, e.g., craton margins compared to the craton interior, making the interactions between mantle plumes and thick cratonic root short-lived and less intensive and hence preventing erosion of the thick root from below (Foley, 2008); and 2) decompression melting is greatly reduced beneath thick cratonic mantle root, due to relatively high pressure and the lack of adequate space for decompression above the solidus (Lee et al., 2011). Therefore, the influence of mantle plumes on thick lithosphere of Archean cratons by means of thermomagmatic erosion is expected to be limited, and may not noticeably contribute to craton destruction.

Structurally heterogeneities commonly presented in cratonic lithosphere may help to enhance the effects of mantle plumes. It has been reported that laterally weak belts in continental lithosphere are often the areas of intense heating and strain concentration during tectono-thermal events (e.g., Tommasi et al., 2001). The weak layer topped with the MLD within cratonic lithospheric mantle was also suggested to be the focus of ductile deformation at mantle depths (Chen, 2017; Prieto et al., 2017). A recent geodynamic study proposed that energetically rising mantle plumes could preferentially concentrate to pre-existing weak belts in stable cratonic regions and trigger the removal of the lower part of the thick lithosphere by progressively peeling off the cratonic root along the weak MLD layer, causing widespread surface uplift and volcanism (Hu et al., 2018). Such a process and later vertical accretion and thickening of the lithosphere were invoked to link the evolution of cratons with mantle plume activities in terms of plume-lithosphere interaction, which indeed well explains the kimberlite volcanism, sedimentation and unroofing history of the western Gondwana cratons since the Cretaceous and the present-day crustal and mantle structure of the region (Hu et al., 2018). Seismic observations in South Africa, a region that was once part of western Gondwana, further show that the mantle plume-related kimberlite volcanism and corresponding lithospheric modifications apparently cluster at relatively weak and low-velocity belts within cratons or craton margins, without changing the lithospheric structural coherency and long-term stability of the craton cores (Begg et al., 2009; Griffin et al., 2009). Such a feature is also present in North America where diamond-bearing and non-diamondiferous kimberlites tend to occur within or at the boundaries or surrounding areas of the high-velocity cores of cratons with various degrees of velocity reduction (Figure 16a; Schaeffer & Lebedev, 2014), suggesting lithospheric reworking to different extents on relatively weaker parts of the lithosphere but without affecting the overall structure and nature of the craton interior.

Indeed, there is little evidence for any Archean craton having been destroyed by underlying mantle plumes, even superplumes (Lee et al., 2011; Wu et al., 2014). In addition to the western Gondwana cratons and those in North America mentioned above, a plenty of other cratons in the world, such as the Siberian craton (Howarth et al., 2014; Reichow et al., 2009), Indian craton (Chalapathi Rao et al., 2013; Lehmann et al., 2010), Eastern European craton (Downes et al., 2005; Wu et al., 2013), Tarim craton (Xu et al., 2014), western Yangtze craton(Xu et al., 2004), etc., which all have been noticeably affected by plume activities with or without significant lithospheric thinning, remain rigid and stable today. This indicates that mantle plumes may not be efficient in causing severe refertilization and destabilization of cratons, even being capable of inducing marked thinning of cratonic lithosphere by making use of both lateral and vertical within-lithosphere weaknesses (Figure 27a). Probably the accretion of remnant mantle plume materials from below could even result in a rapid thickening of the thinned lithosphere in the wake of the mantle plume activity, as suggested in some cratonic regions of eastern

Europe and western Gondwana (e.g., Hu et al., 2018; Wu et al., 2014; Yuan et al., 2017).

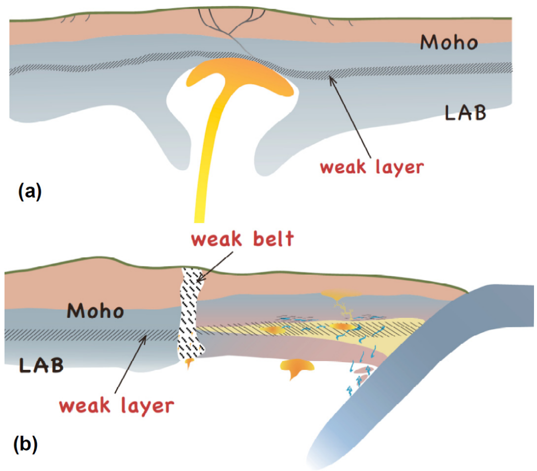

Figure 27. Schematic models of lithospheric delamination and thinning caused by plume-lithosphere interaction (a) and lithospheric destruction of a craton caused by oceanic subduction (b). In (a), the cratonic lithosphere is not severely refertilized and destabilized by the plume, although delamination of the lower part of the cratonic lithosphere may happen along the MLD-topped weak layer, caused by the upwelling of the plume along a horizontal weak belt within the cratonic lithosphere. After delamination, melt-depleted plume materials could accrete from below to the bottom of the thinned lithosphere and result in a rapid lithospheric thickening in the wake of the plume activity. Modified from Hu et al. (2018). In (b), the composition and properties of the cratonic lithosphere is greatly changed during the long-term oceanic subduction. The lower part of the lithospheric mantle is severely metasomatized via infiltration of fluids released by subducting slabs and melts in the mantle wedge. The portion that is surrounded by the MLD-topped weak layer and a major weak belt in the lithosphere is eventually destroyed by rapid erosion and/or delamination. The upper part of the lithospheric mantle and the crust are also modified and weakened by subduction-related magmatic and deformation processes, especially after delamination.

4.3.2. Plate Tectonics Perspective

Compared with mantle plume activities, plate tectonic processes are more efficient for reworking as well as accretion of continents (see section 3 and also (Stern & Scholl, 2010; Tang et al., 2016), due largely to the vigorous recycling of water and other volatiles and involvement of more significant deformation and metamorphism.

Water abundance or accessibility of fluids is probably the most important factor in controlling the style, degree and process of continental reworking. For example, fluid flows and associated fluid-rock interactions are of central importance in controlling the reactivation of shear zones and fault systems as well as deep crustal deformation and metamorphism during orogenic processes (Austrheim, 1998; Cartwright et al., 2000; Stewart et al., 2000). In particular, fluid is required for efficient eclogitization, and thus accessibility of fluids is key for whether or not crustal roots can survive from delamination to arrest orogenic collapse (Leech, 2001). Moreover, the presence of water promotes partial melting and facilitates melt- or fluid-induced metasomatism to modify the composition and affect the density and rheology of subcontinental lithospheric mantle (Karato, 2010; Peslier et al., 2017; Tang, Zhang, Santosh, et al., 2013), which are key factors to influence the stability and longevity of cratons. Large amounts of water and easy accessibility of fluids are mostly favored at convergent boundaries, especially in subduction zones (Cai et al., 2018; Dixon et al., 2002; Hirschmann, 2006; Karato, 2010). Moreover, the prevailing oxidation conditions above subducting slabs were also suggested experimentally to be an important factor in softening mantle peridotites (Cline II et al., 2018). These features thus make subduction zones the most significant sites of both creation of juvenile continental crust and lose and reworking of pre-existing one over other tectonic settings (Clift et al., 2009; Hawkesworth et al., 2016; Stern & Scholl, 2010).

Another main difference between subduction processes and mantle plume activities in affecting continental evolution is probably their different durations in function. Mantle plume activities do not focus on the same continental region for a sufficient long period. Considering a typical plate speed of 5 cm/yr, the distance from the mantle plume to the original site of plume impingement after 50 Ma would be 2500 km, which is too far for the mantle plume's effects to remain on the lithosphere of the original site. On the other hand, subduction zones can be much longer-lived along a continental margin and the subduction processes can continuously affect the physical and chemical properties and thus control deformation of the overlying continental lithosphere. For instance, the eastern margin of East Asia has been an active continental margin associated with Paleo-Pacific subduction since ~200 Ma (Wu, Li, et al., 2007; Xu et al., 2013), and the Cordilleran active continental margin in eastern Pacific started in Late Triassic or even earlier (Barth et al., 2011; Boschman et al., 2018; Riggs et al., 2016). These suggest that both continental regions have been influenced by oceanic subduction in the Pacific tectonic domain for at least ~200 Ma. If considering earlier subduction events that may have also imposed effects on the two regions, e.g., the Paleo-Asian ocean subduction from north and Paleo-Tethyan ocean subduction from south beneath East Asia during Paleozoic to early Mesozoic (Meng & Zhang, 1999; Xiao et al., 2015; Zhao et al., 2018), and the late Permian-early Triassic Sonoma orogeny in Cordillera (Dickinson, 2004), the duration of subduction influence there could be much longer. The region of South Eurasia along the Tethys belt has also experienced long-term ocean subduction and repeated collision and accretion of continental ribbons from Gondwana since the Paleozoic (Wan et al., 2019). A longer period of subduction and associated processes would lead to larger volumes of water or fluids released from subducting slabs, and are expected to exert stronger impacts on the overlying continental lithosphere, not only margins but also interiors. This may explain why significant deformation and reworking of continents including some cratons have occurred and even are on-going processes today in the circum-Pacific and Tethys domains (Figure 17).

Similar to the mantle plume case, structural heterogeneities in the lithosphere may also significantly influence the extent and processes of the subduction-associated reworking of continents. On the one hand, both lateral (e.g., faults, boundary zones) and vertical weaknesses (e.g., decollements, MLD layer) could intensify and speed up the thermo-chemical-mechanical modification of the continental lithosphere by providing paths for the infiltration of fluids and melts and concentrating deformation due to their weaker rheology (e.g., Liu et al., 2018a; Peslier et al., 2017; Wang et al., 2018). Moreover, the presence of lithospheric weaknesses would promote mechanical decoupling between juxtaposed tectonic units or between the upper and lower parts of the continental lithosphere during the reworking process (e.g., Heron et al., 2016; Lenardic et al., 2003; Liao & Gerya, 2014). For a generally short period (e.g., < 30-50 Ma comparable to the duration of a plume activity) or a limited influence of subduction on the interior of continents or core parts of cratons (e.g., a craton being separated from a subduction zone by a wide marginal belt or the cratonic crust and upper lithospheric mantle separated from convecting asthenosphere by the lower mantle root), these effects of structural heterogeneities on continental reworking are essentially similar to those in the mantle plume-lithosphere interactions. In such cases, the lithospheric reworking is primarily intensified at the weaknesses themselves, and thus these weaknesses function as buffer zones to protect cratons from tectonic deformation and modification (e.g., Lenardic et al., 2003; Zhu et al., 2011).

However, with a longer period and more continuous and stronger influence of oceanic subduction, lithospheric weaknesses appear to fail in preventing craton from significant reworking to destruction (e.g., the NCC, Wyoming craton, etc.). Rather, the major weaknesses in the lithosphere may control the process, degree and spatial extent of craton destruction (Figure 27b). Under the influence of long-lived subduction, the overlying continental lithosphere could be significantly reworked, such that the lithosphere of the trench-side mobile belt is largely destroyed and the cratonic lithosphere is directly exposed to the subduction. The lower part of the cratonic mantle root may thus be strongly metasomatized from below by melts and fluids sourced from the subducting slab, and the weak MLD layer may be further weakened and enlarged by hydrous inflow from nearby asthenosphere in the mantle wedge, as suggested by numerical modeling results (e.g., Liu et al., 2018a; Wang et al., 2018). This simultaneous reworking both in the middle and lower parts of the cratonic lithosphere is expected to hasten gravitational instability and induce a rapid delamination or whole-sale erosion of the lower mantle root along the MLD layer. It is worth noting that after the loss of the lower mantle root, a new LAB could develop at the previous MLD depths, and the overlying original upper lithospheric mantle and the crust may be heated and modified by the hot rising asthenospheric mantle as the subduction process continues (Figure 27b, Chen et al., 2014; Liu et al., 2018b). On the other hand, such lithospheric reworking and destruction are possibly spatially confined by a major vertically extended weak belt within the upper continental plate, and the hinterland to the other side may not be significantly modified, due probably to both the protection effect of the weak belt and the fading away of the subduction influence (Figure 27b).

Overall, under the long-lived and strong influence of subduction and important effects of pre-existing lithospheric weaknesses, a craton could be severely modified even destroyed but likely with a spatially uneven reworking and destruction (Figure 27b). This explains the sharp differences in both tectonic processes (e.g., F. Y. Wu et al., 2019; Xu, 2001; Zhu et al., 2011) and structure from the surface to the base of the lithosphere between the eastern and central-western NCC (Figure 26; Chen, 2010; Wei et al., 2015; Xu et al., 2018; Y. Zhang et al., 2019; Zheng et al., 2017), and may also be responsible for the similar observations across the boundaries of distinctly different structures and/or deformation in western North America, including the Wyoming craton (Figure 16a a-b). The scenario proposed in Figure 27b is consistent with the observed similar depths of the MLD in stable cratonic regions and the LAB in adjacent destroyed cratonic parts or tectonic regions (Figure 24, Chen, 2017; Thybo, 2006 and references therein).

In both cases of the NCC and the western North American craton, the MLD layer is thought to be the key horizontally extended weakness within the lithosphere affecting the destruction of the cratons (e.g., Chen et al., 2014; Liu et al., 2018a; Wang et al., 2018). The effects of the MLD layer to facilitate decoupling and delamination of the lower mantle root from the upper part could be particularly pronounced in the compressional setting (e.g., Gorczyk et al., 2012; Wang et al., 2018) during the earlier flat oceanic subduction beneath both regions (e.g., F. Y. Wu et al., 2019; Yonkee & Weil, 2015). It is also possible that the weak lower crust with compressional thickening and eclogitization may further aid the deformation and destruction of the cratonic lithosphere (e.g., Bird, 1979; Le Pourhiet et al., 2006; Wang et al., 2018). As for the major vertically extended weaknesses, the Paleoproterozoic Trans-North China Orogen (central NCC) that sutured the Archean western and eastern NCC may be reactivated and play an important role in controlling the spatial extent of the craton destruction, as all the sharp E-W changes in topography, gravity (marked by the North South Gravity Lineament, or NSGL) and crustal and lithospheric structures spatially coincide with this belt (Figure 26). In the western North America especially around the Wyoming craton, large velocity gradients in the shallow upper mantle (100-150 km depth) and the boundary of stable cratonic core appear roughly following the Paleoproterozoic Trans-Hudson Orogen (Figure 16a-b), indicating a possibly similar role of the orogenic belt to the Trans-North China Orogen during the destruction of the western North American craton.

Overall, the combination of current observations of continental reworking and destruction of cratons and results of numerical modeling suggests that both processes at plate boundaries (i.e., plate tectonics) and underneath plates (e.g., mantle plume activities) can cause modifications of the continental lithosphere to various degrees, but mostly at around continental margins or lithospheric weak zones. However, specific plate tectonic processes, in particular long-lasting oceanic subduction, may induce significant softening and deformation of both the cratonic crust and lithospheric mantle by involving great amounts of water and other volatiles and large tectonic stresses. Further due to the important effects of reactivated pre-existing lithospheric weaknesses, these processes can eventually lead to severe but spatially heterogeneous reworking and destruction of cratons.

5. Discussion and Conclusions

5.1. Origin of Continents

As discussed in Section 2, both the island arc and oceanic plateau models can be applied to interpret the formation of Archean granite-greenstone terranes, but neither of them can satisfactorily explain all features of Archean continental cratons. For example, the island arc model under a plate tectonic regime cannot reasonably explain the following features of Archean cratons:

(1) Bimodal volcanic assemblages in the Archean greenstone terranes. In the Archean greenstones, common volcanic rock associations are bimodal, where basaltic and ultramafic rocks are associated with dacite and rhyolite, without much andesite. They are different from Phanerozoic volcanic arc assemblages which are commonly unimodal with andesite as the most typical volcanic-rock type in many mature magmatic arcs.

(2) Widespread presence of komatiites and komatiitic rocks in the Archean greenstone terranes. Plate tectonics is incompetent to explain the origin of these Archean komatiites because their formation needed a melting process that occurred at 1600-1900°C. Most Archean greenstones contain komatiites. Komatiites is characterized by high MgO (>18%), which requires at least 40-60% partial melting of the mantle. Such high degree partial melting of the mantle requires the partial melting temperatures at 1600-1900°C. Such high temperature conditions are not

common in island arc environments, though minor komatiites with high silicon contents and hydration may have been produced in subduction zones.

(3) TTG gneisses making up 60~70% of the basement exposure in Archean terranes. In major Archean cratons, similar-aged TTG gneisses are exposed over a whole craton and do not show any systematic progression in age across the 500-1000 km wide, all emplaced within a very short period. This is inconsistent with a successively-accreted arc model in the plate tectonics context.

(4) Mass-balance problem with the Archean TTGs that make up 60-70% exposure of Archean cratons. Due to their high LREE/HREE ratios, high large ion lithophile element contents and negative Nb–Ta–Ti anomalies, Archean TTG rocks could be either derived from the partial melting of eclogites or rutile-bearing garnet amphibolites of subducted oceanic crust, or generated by melting the base of thick basaltic plateaus above mantle plumes, leaving behind restites containing pyroxene, garnet, and rutile (Bédard, 2006). However, modern petrological experiments have demonstrated that if TTG rocks were formed by partial melting of eclogites or garnet-/rutile-bearing amphibolites, the partial melting degree should be lower than 30%; otherwise, the resultant TTG rocks would not possess above geochemical features. If the partial melting degree is no higher than 30%, it requires that the actual volume of eclogites or garnet-/rutile-bearing amphibolites is three times of the volume of TTG rocks, which would create a mass-balance problem with the Archean TTGs that make up 60-70% exposure of Archean cratons, especially considering the difficulty of extracting such a large volume of TTG melts from subducting slabs within such a short period.

(5) Anticlockwise P-T paths involving isobaric cooling reconstructed for many Archean cratons. Available data show that the metamorphic evolution of Archean rocks from many Archean cratons is characterized by anticlockwise P-T paths involving isobaric cooling (Ge et al., 2003; Halpin & Reid, 2016; Jayananda et al.,

2000; Kamber et al., 1996; Kramers et al., 2001; Maas & Henry, 2002; Mvondo et al., 2017; Percival, 1994; Raith et al., 1990, 1999; Rollinson, 1989; Sandiford, 1985; Tsunogae et al., 1992, 1999; Zhao et al., 1998, 2001, 2005; Zulbati & Harley, 2007), which reflect the metamorphism related to the intrusion and underplating of large amounts of mantle-derived magmas (Bohlen, 1991). Although anticlockwise P-T paths involving isobaric cooling may also characterize the metamorphism occurring at the root of magmatic arcs or under back-arc basin setting, it is argued that such metamorphism should be paired with the relatively high pressure metamorphism of subducted zones that is characterized by clockwise P-T paths involving isothermal decompression, and they together form the paired metamorphic belts like what we have observed in modern magmatic arcs (Brown, 2006, 2008; Brown et al., 2020). Brown (Brown, 2008) argued that paired metamorphic belts could be regarded as a hallmark for tracing plate tectonics in the early history of the Earth. This is not the case in many Archean cratons where the Archean metamorphism is characterized by anticlockwise P-T paths involving isobaric cooling, without any clockwise P-T paths involving isothermal decompression to form paired metamorphic belts.

(6) The dominant structural patterns of Archean cratons are gneiss domes (dome-and-keel structures), which do not resemble linear orogenic belts that typify Proterozoic and Phanerozoic accretionary (subduction) orogens that resulted from metamorphosed and deformed continent margin arcs.

(7) Absence of ophiolites and ultra-high pressure (UHP rocks) in Archean cratons. Ophiolites are an unequivocal index of plate tectonic activity as they are fragments of oceanic lithosphere tectonically emplaced on continental crust through horizontal movements, but evidence for Archean ophiolites is sparse and often controversial. It is the same case with UHP rocks that are common in Phanerozoic and even Proterozoic arc terranes, but have not been reported with reliable Archean metamorphic ages from Archean terrains.

In summary, Archean cratons do not exhibit typical lithotectonic elements that are observed in Phanerozoic island arcs. In contrast, although oceanic plateaus formed by mantle plumes fail to provide enough H_2O for the partial melting of basaltic rocks to form TTG magmas, the mantle plume-driven oceanic plateau model can well explain: (1) the exceptionally large exposure of TTG and granitoid intrusions that were emplaced over a short time period without systematic age progression across a craton-scale; 2) the presence of MgO-rich komatiitic melts with eruption temperatures as high as 1650°C; 3) bimodal volcanic assemblages; 4) dominant diapirism-related domal structures; 5) metamorphism (with anticlockwise P-T paths involving isobaric cooling) related to the intrusion and underplating of large amounts of mantle-derived magmas; and (6) affinities of mafic rocks (metamorphosed to be amphibolites and mafic granulites) to continental tholeiitic basalts. Taken together, we favor a mantle plume-driven oceanic plateau model for the formation of Archean felsic continents. In our view, felsic continents appeared before plate tectonics started on our planet, and it was just felsic continents that created favorable conditions for initialing plate tectonics through subduction of a dense oceanic crust below a buoyant continental crust. In this sense, modern-style plate tectonics contributed little to the formation of ancient continents, but it played an important role in the accretion and reworking of continents that will be discussed in the next two sections.

5.2. Growth of Continents

Through the previous review of major geological events leading to the formation and accretion of the five major continents in the world, we can summarize some common basic laws for craton formation and continental growth. It is obvious that each continent was composed of several Archean cratonic nucleuses welded by post-Archean orogenic belts. The Archean cratons were formed by older cores, i.e. cratonic nucleuses, which were amalgamated together along Proterozoic orogenic belts. Although there is controversy on the origin of continents (see Section 2), it is evident that small pieces of early Archean nucleuses joined together to form larger Archean provinces through tectonic processes similar to terrane accretion in Phanerozoic orogenic systems. The accretionary processes of Archean cratons were best illustrated by the North American continent where Slave, Rae, Hearne, and Wyoming cratons each grew on its margin by oceanic subduction and arc accretion, and subsequently collided together to form a larger Archean province.

The Archean nucleuses grew laterally through arc magmatism in the Paleoproterozoic, which happened in all the continents worldwide. Through the subduction and accretionary processes, the continents grew progressively. This kind of enlargement of continents is well recorded in the South America, Australia and South Africa continents (Figs. 10-12). Paleoproterozoic (2.1-1.8 Ga) continent-continent collisional orogeny happened globally, which led to the formation of the Earth's first supercontinent Columbia/Nuna (Zhao et al., 2002). In the North America continent, Paleoproterozoic (~1.8 Ga) high-pressure, low-temperature collisional metamorphism and crustal layering in both structure and deformation comparable to the Himalaya orogeny was discovered in the Trans-Hudson orogen, which implies that modern-style plate tectonic processes featuring deep continental subduction and subsequent extensional collapse of an orogen occurred at least about 1800 million years ago (Darbyshire et al., 2017; Weller & St-Onge, 2017). Similar collisional process has also been well-documented in the Paleoproterozoic Trans-North China Orogen in the North China Craton (Zhao et al., 2012).

Two contrasting orogenic systems (external and internal) were proposed to describe the global Phanerozoic orogens (Collins et al., 2011), of which continental accretion mainly occurs in the external orogenic systems, also known as accretionary orogens that are major sites for Phanerozoic continental growth, particularly important for the enlargement of the Eurasia, North America, South

America and Australia continents during Phanerozoic time. Growth of giant subduction-accretion complexes, slab roll-back, oceanic ridge subduction and formation of intra-oceanic subduction systems are major processes that lead to continental growth during accretionary orogenesis (Jahn et al., 2000; Şengör et al., 1993; Xiao et al., 2015). These have been well illustrated in the Altaids in Central Asia, the Cordillera in North America, the Andes in South America and the Tasmanides in Australia.

We propose that accretionary processes in ocean-continent convergent margins have been playing important roles in the formation and growth of the global continents since Archean when the size of continents was small. In contrast, modern-style continent-continent collisional orogeny, marked by deep subduction of continental lithosphere, might have occurred much later since the Paleoproterozoic (1.9-1.8 Ga) when the continents grew to considerable size through earlier accretionary processes. The collisional orogeny is likely a hallmark for the change from Archean-style plate tectonics to modern-style plate tectonics. In the early Earth, accretionary processes were the major mechanism for continental growth, while both accretionary and collisional orogenesis have been important in shaping the Earth. The secular change of plate tectonics style has changed the behavior of Earth's lithosphere, which in turn affects the mechanism and rates of continental growth.

5.3. Reworking of Continents

5.3.1. Continental Reworking vs. Continental Growth

The generation, growth (accretion), reworking and destruction of continents are mutually correlated over time. At present, only a small volume (<25%) of the continental crust is older than ~3.0 Ga (e.g., Belousova et al., 2010; Korenaga, 2018; Voice et al., 2011), whereas about 60-80% of the present volume of the continental crust may have formed after 3.0 Ga (Campbell, 2003; Dhuime et al., 2012; Pujol et al., 2013). This means that significant volumes of old crust have been destroyed, accompanying obvious variations in the rates of growth and reworking of the continental crust since ~3 Ga (e.g., Dhuime et al., 2018; Hawkesworth et al., 2020 and references therein). In particular, dramatic increases in the rates of reworking and destruction and simultaneous reduction in the rates of net growth of the continental crust took place during Neoarchean time (3.0-2.5 Ga), coeval with the marked changes in the global tectonics. For instance, with the continuous cooling of the Earth since ~3.0 Ga (Korenaga, 2013), the crust and lithosphere of continents could have attained a certain rigidity in the Neoarchean, as reflected by the formation of the oldest major sedimentary basins (~2.8 Ga) and oldest regional dyke swarms (~2.6 Ga) (Cawood et al., 2018). The appearance of the oldest granulite (Holder et al., 2019) coincided with the gradual increase in the thickness and buoyancy of the continental crust as a result of a compositional change from mafic to more felsic in the Neoarchean (Dhuime et al., 2018; Tang et al., 2016). Such a coincidence suggests that it was probably till the Neoarchean that the continental crust was sufficiently strong to sustain crustal thickening and deep-crustal metamorphism. This may explain the rapid emergence of UHT metamorphism since ~2.8 Ga (Figure 19c). The strong and rigid continental crust and stabilized lithosphere in the Neoarchean would have facilitated the onset of plate tectonics as the dominant global regime, leading to the reduction in the net crustal growth and the development of the supercontinents/supercratons, as reflected by the dramatical increases in the rates at which differentiated continental crust was reworked and destroyed (Hawkesworth et al., 2020).

The development and continuous operation of plate tectonics over time is thought to generally increase the destruction and reworking rates and in turn reduce the growth rates of continental crust, with much less effects on the overall rate of crustal generation (Dhuime et al., 2018; Hawkesworth et al., 2020). This is largely attributed to the widespread subduction and associated tectonic processes. In addition to this long-term trend, it has demonstrated that the degree of continental crustal reworking exhibits shorter-period variations, closely associated with the evolution of supercontinents (Figure 19b, also see 4.2.4). However, it is still unclear whether or not the sharp changes of the continental crust at ~3.0-2.5 Ga accompanied similar changes in the continental lithospheric mantle, since the origin, timing and formation processes of the mantle lithosphere and its link with the crustal formation during the early Earth remain unknown or controversial (e.g., Gerya, 2014; Perchuk et al., 2020 and references therein). On the other hand, in the later evolution of continents under the plate tectonics regime, during which strong crust-mantle coupling has proven to be a pre-request for the rigidity and long-term stability of continental lithosphere (see 4.1.2), the growth, reworking and destruction of the continental lithospheric mantle would closely link to that of the continental crust in both time and space.

5.3.2. Cratons vs. Orogens

Under the plate tectonics regime, the processes of generation, accretion, reworking and destruction of continents took place mostly at convergent plate margins during the development of accretional and collisional orogens (Clift et al., 2009; Hawkesworth et al., 2016, 2020; Stern & Scholl, 2010). Orogenic processes in the Precambrian played essential roles in shaping the lithosphere of cratons, which differs in structure and properties from that of Phanerozoic orogens/tectonic belts (see 4.1, Table 1). Temperature is considered the most significant factor responsible for the differences between Precambrian and Phanerozoic orogens and thus between old cratons and young tectonic belts. In the late Archean and early Proterozoic (>1.6 Ga), during which most cratons developed without modern analogues (Arndt et al., 2009; Eaton & Claire Perry, 2013; Lee et al., 2011), high mantle temperatures (~150–250 degrees Celsius) than the present-day values (Herzberg et al., 2010; Korenaga, 2013) resulted in hot orogens characterized by higher-degree mantle melting, weaker crust and upper mantle, distributed shortening, high- to intermediate-temperature metamorphism, and efficient elimination of high-density, low-viscosity lithologies (Arndt et al., 2009; Chardon et al., 2009; Condie, 2007; Gerya, 2014). The continental lithosphere established at such old orogens appears to be sufficiently buoyant and strong to survive from mantle erosion and tectonic disturbance as the Earth cools (Eaton & Claire Perry, 2013; Perchuk et al., 2020). In contrast, Neoproterozoic to Phanerozoic orogens are typically cold orogens with temperatures too low (<150 degree than the present-day values) to produce craton-type of lithosphere, but favorable for the stabilization of continental subduction and generation of UHP metamorphic complexes that are nearly absent in the Precambrian (Ernst et al., 1997; Gerya, 2014; Guillot et al., 2008).

Despite the large differences, there are some observations on the commonalities between structures produced by Phanerozoic orogens and those preserved within old, cratonic lithosphere, based on which comparable tectonic processes have been suggested among orogens of Precambrian and Phanerozoic ages. As mentioned in 4.2.2, the common features of layering in the crustal structure and deformation have led to the suggestion that similar mid-lower crustal flow may have taken place in both the Mesoproterozoic Grenville and Paleoproterozoic Trans-Hudson orogens and the present-day Himalayan-Tibetan orogen, probably under sufficiently high temperature conditions that made the crust much weaker than usual, as indicated by geophysical observations(Bai et al., 2010; Darbyshire et al., 2017; Liu et al., 2014; Pawlak et al., 2012; Petrescu et al., 2016). Considering the gradual cooling of the Earth, it is expected that such kind of high-temperature conditions might be more frequently met in the Precambrian than the Phanerozoic to present day. It is also proposed based on geodynamic modeling (e.g., Betts et al., 2015; Cooper & Miller, 2014) that modern terrane accretion, such as the Paleozoic accretion of the Selwyn block to the Gondwana margin in SE Australia (Cayley, 2011; Foster & Gray, 2000) and the ongoing accretion of the Yakutat terrane in the eastern end of the Aleutian subduction zone (Colpron et al., 2007; Eberhart-Phillips et al., 2006), introduces features reminiscent of the internal dipping structures (such as dipping MLDs) widely observed in the shallow cratonic lithospheric mantle (Copeland et al., 2017). Therefore, the dipping structures observed beneath cratonic regions are considered a kind of tectonic imprints of Precambrian accretionary orogens (e.g., Bostock, 1998; Chen et al., 2009; Snyder et al., 2017 and references therein), which contribute strongly to the assembly and stabilization of cratons and growth of continents owing to their high production rates of juvenile crust compared to Phanerozoic accretional orogens (Chardon et al., 2009; Condie, 2007). However, to what extent Precambrian accretionary orogens are comparable in terms of tectonic regime and processes to modern counterparts is still an open question.

5.3.3. Lithospheric Processes vs. Mantle Dynamics

Lithospheric processes especially at orogens of various styles and ages, which are responsible for the formation and long-term evolution of continents, have been increasingly evidenced to be largely controlled by and strongly interact with sublithospheric mantle dynamics (e.g., Chen et al., 2019; Copeland et al., 2017; Faccenna et al., 2013; Handy et al., 2010). Under the plate tectonic regime, variations in crustal and lithospheric deformation, magmatic activity and surface topography, or collectively in the pattern of orogeny in different orogens, and also episodic changes along a single orogenic belt at convergent plate margins, are often attributed to different morphologies of the subducting slabs and slab-mantle interactions at depth. Typically the subduction dynamics in the upper mantle, such as trench retreat or advance, slab flattening or steepening, continuous subduction or break-off, slab stagnating in the mantle transition zone or penetrating into the lower mantle, etc., is considered of key importance in determining the orogenic style in the overlying plate, based on the studies of both circum-Pacific (Copeland et al., 2017; Horton et al., 2016; Li & Li, 2007; Yang et al., 2018) and Tethyan orogens (Handy et al., 2010; Li et al., 2011; Mouthereau et al., 2012). Recent geological and geophysical observations and geodynamic modeling have further explored the significant roles of slab subduction in the lower mantle in driving plate motion and shaping orogens on the surface, with particular emphasizes on the different tectonic responses of the overlying continental lithosphere to slab-upper mantle and slab-lower mantle interactions (e.g., Chen et al., 2019; Faccenna et al., 2013; Schellart, 2017). It has been suggested that the driving forces provided by subducting slabs function differently in the upper and whole (upper + lower) mantle subduction systems, which may result in distinct patterns of orogeny (Conrad & Lithgow-Bertelloni, 2004; Faccenna et al., 2013). In case of an upper mantle subduction, the pull force that the subducting slab exerts directly on the surface plate dominates in the system, such that the overriding plate moves in the same horizontal direction as the subducting slab and slab rollback/trench retreat leads to extensional deformation in the overriding plate without obvious crustal thickening (Faccenna et al., 2013). This indeed explains the main features of the Cenozoic Mediterranean mobile belt (Faccenna et al., 2014; Jolivet et al., 2003) where seismic tomography shows that high-velocity slabs are mainly confined in the upper mantle (e.g, Zhu et al., 2012a). In case of a slab penetrating to the more viscous lower mantle, on the other hand, the down-going slab exerts indirect suction force on the surface plate by exciting mantle circulations through viscous coupling. The slab suction force is thought to play a more important role than slab pull in this subduction system, inducing opposite horizontal movements of the overlying plate and subducting slab, with compressional deformation and obvious thickening of the crust in the overriding plate (Faccenna et al., 2013). Such a scenario is consistent with the observations at a number of orogens controlled by whole-mantle subduction, typically the Himalaya-Tibet and Central Andes (e.g., Allmendinger et al., 1997; Hatzfeld & Molnar, 2010). The differences in the upper- and whole-mantle subduction dynamics and corresponding lithospheric responses have been invoked not only to decipher the distinct characteristics of orogens (e.g, Faccenna et al., 2013), but also to understand the temporal variations of lithospheric tectonics in a single orogenic system, such as the Andes during the Mesozoic-Cenozoic evolution (e.g, Chen et al., 2019; Schellart, 2017).

In addition to slab subduction at depth, other mantle processes, particularly plume upwelling, have also been widely investigated and suggested to have played important roles in the growth of Archean cratonic mantle keels (e.g., Arndt et al., 2009; Griffin et al., 2003; Stein & Hofmann, 1994) and exerted considerable impacts on the later growth and reworking of continents (e.g., Chen et al., 2020; Hu et al., 2018; Wang et al., 2015; Wu et al., 2014; Yuan et al., 2017). As demonstrated in 4.3.1, rising plumes help to speed up the thinning and in some cases later thickening of cratonic lithosphere, especially at pre-existing lithospheric weak zones (Hu et al., 2018; Wang et al., 2015), although plume impingement alone appears incapable of contributing to an efficient weakening and destruction of stable cratons. It is also suggested that strong interactions between plumes and overlying plates could result in an acceleration of plate motion by reducing the lithosphere-asthenosphere coupling (Kumar et al., 2007), or trigger rifting and even break-up of continents with the help of lithospheric weaknesses (Buiter & Torsvik, 2014; Frizon de Lamotte et al., 2015), or both. This has been invoked to explain the globally fastest plate motion (Müller et al., 2016) and longest continental rift system (Brune et al., 2017) coeval in the early Cretaceous (~140-120 Ma) over the past 200 Myrs, for this time period was characterized by significantly enhanced plume activities and mantle temperature (Humlera et al., 1999). Moreover, the coupling between superplume and supercontinent cycles throughout the geological history (e.g., Li & Zhong, 2009; Li et al., 2019) further indicates that the lithosphere-plume interactions and strong influences of plume activities on the structure and tectonics of continents not only are evident in the recent short era, but also persist during the Earth's long-term evolution. Indeed, an integration of various observations on plate and plume tectonics and their interactions at multiple time-space scales would help to gain deeper insight into the correlation between lithospheric processes and mantle dynamics and the co-evolution of the shallow and deep Earth (Chen et al., 2020).

5.4. A Scenario for the Formation, Accretion and Reworking of Continents

On the basis of the above discussion, we propose the following scenario for the formation, accretion and reworking/destruction of continents:

(1) ***Formation of continents***: During the Eoarchean-Mesoarchean time when plate tectonics had not started on our planet, felsic continental crust was most likely originated from oceanic plateaus formed by mantle plumes that were originated from the core-mantle boundary due to tremendous accumulation of radioactive heat in the ~200 km thick D" layer following the formation of core-mantle-crust and associated magma ocean event at about 4.45 Ga. In many Archean cratons, TTG rocks were several hundred million years younger than greenstones that are considered to have formed from oceanic island basalts (OIB), indicating that the dominant ultramafic-mafic and minor felsic volcanic rocks (greenstones) and TTG rocks in the Archean cratons were not the products of a single mantle plume event. For this reason, we favor a two-stage model for the formation of Archean felsic continents, which is shown in Figure 28a-c. As shown in Figure 28a-b, the first stage represented the formation of an oceanic plateau by an early mantle plume. When the total heat from the accumulation of radioactive decay in the D" layer at the core-mantle boundary and released from the outer core reached a critical value, causing partial melting of the mantle material, a plume was induced with the upwelling of the molten mantle material. During the ascent, the plume would trap large amounts of mantle material to enlarge itself as a mushroom-shaped body that had a huge head with a long and narrow tail connecting to a magma chamber at the core-mantle boundary (Figure 28a). When the huge head reached the base of lithosphere, it became a flattened

mushroom shape and then experienced decompressional partial melting to form basaltic magmas. The basaltic magmas erupted on the surface within a very short period (< 1 Ma), forming oceanic plateaus with diameters ranging from 1000-2000 km on ocean floors (Figure 28b).

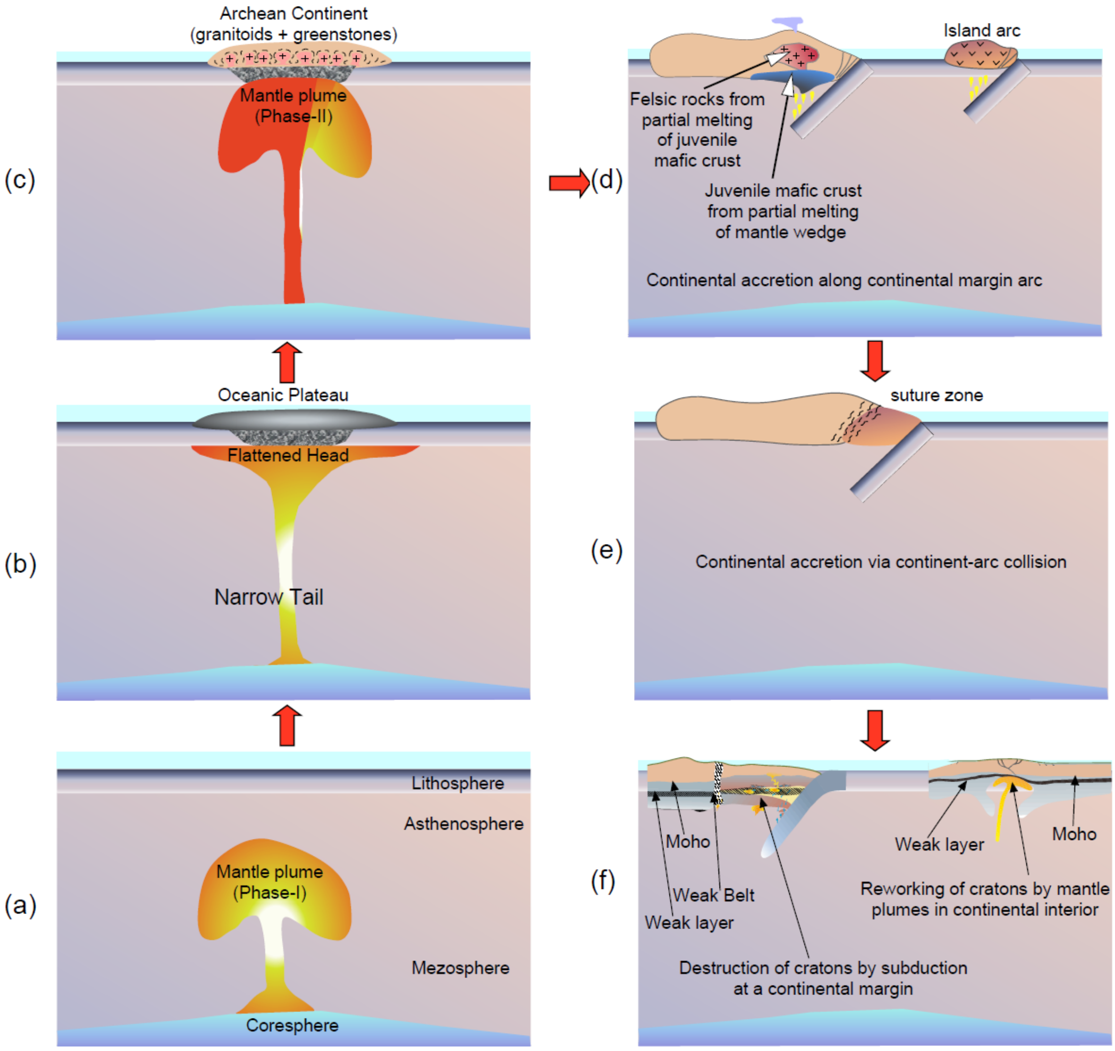

Figure 28. Schematic cartoons showing proposed geological models for the origin, accretion and reworking of continents.

At the later stage when the narrow tail of the plume reached the base of lithosphere, the molten material from the tail erupted on the surface, forming high-temperature (~1600°C) komatiites that are important components of

Archean greenstones. Meanwhile, the high-temperature komatiitic magmas may have also caused the partial melting of mafic oceanic crust to form minor TTG rocks, like minor 2.75-2.65 Ga TTG rocks from the North China Craton. The first stage would end when all mantle magmas were consumed from the chamber at the core-mantle boundary. The second stage represented the partial melting of the base of the oceanic plateau to form TTG magmas, possibly caused by a later mantle plume (Figure 28c). At the second stage, when another mantle plume reached the base of the oceanic plateau, it would cause extensive metamorphism, forming greenstones, amphibolites and garnet-bearing granulites from the metamorphosed upper, middle and lower crust of the oceanic plateau, respectively. The mantle plume would also cause partial melting of the lower crust of the oceanic plateau to form TTG magmas. Driven by density difference, the light TTG magmas would rise and the heavy greenstones would sink down (sagduction), forming unique dome-and-Keel structures in Archean cratons in which the diapiric TTG domes were surrounded by sinking down greenstone belts (Figure 28c). When the mantle plume ceased to provide magmas, the heating stopped and the metamorphic crust experienced cooling without variations in the crustal thickness, leading to the metamorphism characterized by an anticlockwise P-T path involving isobaric cooling. Therefore, the oceanic plateau model can not only reasonably interpret the formation of Archean TTG and greenstones, but also satisfactorily explain the metamorphic and structural features of Archean cratons.

(2) ***Accretion of continents***: Once felsic continents developed from oceanic plateaus, they become continental nucleuses of continental cratons in the early stage. In the following stage, the density difference between felsic continental crust and oceanic crust may have induced subduction of the oceanic lithosphere beneath the continental lithosphere, which may indicate

the start of modern-style plate tectonics in our planet. Subduction zones along the margins of continents are the main sites for docking various terranes including island arcs, seamounts and oceanic plateaus, and micro-continental blocks, which are emplaced into, and/or connected by, accretionary complexes formed by the subduction zones. These long-lived geodynamic processes are mainly characterized by multiple phases of orogeny, sometimes with orogen-scale oroclines bending long linear arcs and/or continental ribbons, generating outside large-scale accretion or massive growth of continental cratons as discussed in Section 3. When an oceanic slab is subducted to a deep level, large amounts of H_2O-dominant fluids would be released out and enter into the mantle wedge, leading to the partial melting of the mantle wedge, producing basaltic magmas. The underplating of basaltic magmas lead to addition of a juvenile mafic crust to the base of upper plates, including continental lithosphere and intra-oceanic arcs (Figure 28d). Some of the basaltic magmas derived from the partial melting of the mantle wedge may also ascend to the surface along deep fractures or faults, forming basalts and andesites due to felsic continental crustal contamination during the ascent, forming continental marginal volcanic arcs (Figure 28d). In many long-live subduction zones, the juvenile mafic crust above the mantle wedge will be heated and undergo partial melting, producing calc-alkaline magmas that are emplaced into the crust to form felsic plutons, leading to continental marginal accretion (Figure 28d). During Proterozoic and especially Phanerozoic times, the accretion or growth of continents may also result from early stage of continent-continent collision or continent-arc collision (Figure 28e).

(3) ***Reworking or destruction of continents***: While juvenile continental crustal material was extracted from the mantle wedge in subduction zones, leading to the outside accretion/outgrowth of continents, the base of the continental

lithosphere may experience softening, thinning and delamination due to the vigorous infiltration and recycling of H_2O-dominant fluids (released out from subducted slabs) along weak layers or weak belts, leading to the widespread reworking and even destruction of continental lithosphere (Figure 28f). Minor reworking or modification of continental lithosphere may also occur in the interiors of continents due to thermo-mechanical-magmatic erosion by mantle plumes (Figure 28f).

5.5. Conclusions

(1) Archean continents with felsic crust must have been originated either from island arcs under a plate tectonic regime or from oceanic plateaus derived from mantle plume.

(2) The island arc model can well explain the formation of Archean TTG rocks, of which the high-pressure-type TTG rocks are considered to have been derived from the partial melting of subducted slabs, whereas the low-pressure-type TTG rocks (equivalent to calc-alkaline granitoids) were derived from the partial melting of juvenile basaltic crust which itself formed by the partial melting of the mantle wedge with addition of fluids released from the subducted slabs. However, the island arc model is in failure to explain the absence of andesites from the Archean greenstone terranes, the presence of komatiites with temperatures of 1600ºC, nearly coeval emplacement of TTG plutons on a cratonic scale, dome-and-keel structures and anticlockwise P-T paths involving isobaric cooling that characterize the deformation and metamorphism of Archean continental cratons.

(3) The mantle-plume oceanic plateau model can reasonably interpret the origins of both Archean greenstones and TTG rocks, of which the tholeiites and komatiites from the greenstones were derived from partial melting of the head and tail of a mantle plume, respectively, the felsic dacite, rhyolitic dacite and

rhyolite from the greenstone were derived from the partial melting of the crust, and TTG rocks were derived from the partial melting of basaltic rocks from the lower part of oceanic plateau, though oceanic plateaus seem to be difficult to provide enough H_2O for the aqueous partial melting of basaltic rocks to form TTG magmas. In addition, the oceanic plateau model can also well interpret the Archean dome-and-keel structure, anticlockwise P-T paths, and absence of blueschist and paired metamorphic belts.

(4) Since plate tectonics appeared in Earth, Archean continental nucleuses underwent accretion or growth along their margins through the subduction of oceanic lithosphere, and accretionary processes involving juvenile arc formation and arc accretion were the major mechanism for continental growth in the early earth's history.

(5) Archean nucleuses grew laterally through arc magmatism and amalgamation during Paleoproterozoic to form sizeable landmass. Globally continent-continental collisional orogeny happened since Paleoproterozoic (2.1-1.8 Ga), which led to the formation of the Earth's first supercontinent Nuna. Phanerozoic continental growth is best illustrated in the Central Asia, the North America Cordillera and the Tasmanides of East Australia.

(6) Growth of giant subduction-accretion complexes, slab roll-back, oceanic ridge subduction and formation of intra-oceanic subduction systems are major processes that lead to continental growth during accretionary orogenesis. The collisional orogeny is likely a hallmark for the change from Archean-style plate tectonics (accretion) to modern-style plate tectonics (accretion and collision).

(7) In addition to lateral accretion or growth along their margins, continental lithosphere also experiences episodic reworking throughout its evolution history, resulting in ubiquitous lateral and vertical structural heterogeneities. This and the

intrinsic chemical buoyancy and high strength of both the crust and mantle lithosphere are essential factors for the longevity and stability of continents.

(8) Reworking of continents has become increasingly significant but with simultaneous reduction in the rates of continental growth since the operation of plate tectonics at ~3.0-2.5 Ga, mostly associated with oceanic subduction and continental collision and related mantle processes at depth during the assembly of supercontinents.

(9) Usually reworking is concentrated at lithospheric weaknesses in either continental margins or interiors, which does not affect the overall stability of continents. However, significant reworking does happen in presumably stable cratonic regions, resulting in severe reactivation and modification, even destruction of the strong cratonic lithosphere.

(10) Destruction of cratons is largely attributed to long-lasting oceanic subductions that could induce marked softening and deformation of the cratonic lithosphere by involving vigorous recycling of water and other volatiles and large tectonic stresses. Plume activities underneath plates play a relatively minor role in continental reworking and destruction, though in cases do affect plate motion and the evolution of cratonic lithosphere.

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Conflict of Interest

The authors declare no competing interests.

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No new data were used in the review article, which is based on existing data from previously published sources. The sources of the public or published data used in the figures are detailed in the legend.

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