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Estimation of Secular Change in the Size of Continents for Understanding Early Crustal Development

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The size of continents is an essential parameter to understand the growth of the continental crust and the evolution of the solid Earth because it is subject to tectonism and mantle convection and affects the preservation of the crust. This article reviews the secular change in the size of continents on the early Earth, focusing on terrigenous clastic rocks, especially quartzose sandstones occurring on relatively large continents. The earliest continental crust in the Hadean or early Archean was produced with a width of ~200–500 km, similar to modern oceanic island arcs along subduction zones or oceanic islands in hot spot regions by mantle plume heating. Through the collision and amalgamation of such primitive continental crusts, continental blocks over 500 km in width and length evolved and appeared by ca. 3.5 Ga. Through further amalgamation, during ca. 3.3–2.5 Ga, the Archean continents emerged with widths and lengths greater than 1,000 km, which were still smaller than those of modern continents. Continents with widths and lengths of nearly 10,000 km have existed since ca. 2.4 Ga (early Proterozoic). Further analyses of the composition and formation mechanism of clastic rocks will help reveal more quantitative secular changes in the sizes of continents.

Keywords: continental crust, size of continent, sandstone, Archean, early Earth

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INTRODUCTION

The bimodal topography, land masses consisting of several major continents with islands of various sizes and an ocean with an average depth of nearly 4,000 m, is unique to the Earth among other planets of our Solar System. Since the establishment of plate tectonics (Wilson, 1965; Morgan, 1968), the subduction zones of oceanic plates are characterized by the production of granitic continental crust through arc magmatism and its subsequent reworking by sedimentary and metamorphic processes (Dewey and Bird, 1970; Matsuda and Uyeda, 1971). Recycling of the continental crust into the mantle is also significant in the subduction zones through tectonic erosion (subduction erosion), sedimentary subduction, and subduction of the island arc crust (e.g., von Huene and Lallemand, 1990; Scholl and von Huene, 2007, 2009; Clift et al., 2009; Yamamoto et al., 2009; Stern and Scholl, 2010). The growth history of the continental crust is a cumulative result of the production, recycling, and reworking processes; in addition, it has been debated for a long time.

By compiling the worldwide whole-rock geochronological data available before the 1960s, Hurley and Rand (1969) demonstrated the age structure of modern continents. They clarified that Archean crusts occupy only 20% of the total continental volume. Considering the influence of crustal reworking and recycling on the age structure of the modern continents, many studies aiming to reconstruct the growth history of the continental crust appeared (e.g., Fyfe, 1978; Armstrong, 1981;

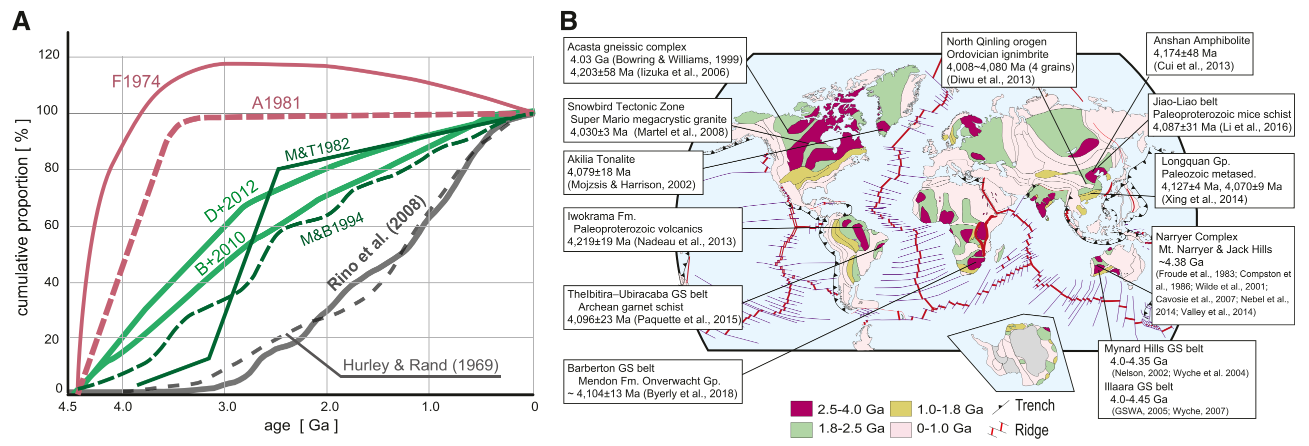

FIGURE 1 | (A) Previous growth models of the continental crust (F1978: Fyfe, s1978; A1981: Armstrong, 1981; MandT 1982: McLennan and Taylor, 1982; MandB 1994: McCulloch and Bennett, 1994; B+2010: Belousova et al., 2010; D+2012: Dhuime et al., 2012). Black solid and dashed lines show the age structure of the modern continents based on geochronological data (Hurley and Rand, 1969; Rino et al., 2008) which commonly indicate the rareness of the ancient crust in modern continents. The growth curves of the continental crust differs largely between studies, considering the great effects of crustal recycling/reworking (red lines) and the limited effects of crustal recycling/reworking (green lines). **(B)** Geotectonic map of the modern continents with localities of Hadean zircon grains modified after Tsutsumi et al. (2018) (GS, Greenstone). The abundant occurrence of Hadean detrital zircon is limited in ca. 3.0–3.3 Ga sandstone and quartzose clastic rocks (Froude et al., 1983; Compston and Pidgeon, 1986; Wilde et al., 2001; Cavosie et al., 2007; Nebel et al., 2014; Valley et al., 2014; Nelson, 2002; Wyche et al., 2004; GSWA; 2005; Wyche, 2007; Paquette et al., 2015; Byerly et al., 2018), even though many odd localities have been reported (Mojzsis and Harrison, 2002; Iizuka et al., 2006; Martel et al., 2008; Diwu et al., 2013; Nadeau et al., 2013; Cui et al., 2013; Xing et al., 2014; Li et al., 2016).

O'Nions et al., 1979; Dewey and Windley, 1981; Allègre and Luck, 1980; McLennan and Taylor, 1982; Reymer and Schubert, 1984; **Figure 1**). After the technical development and popularization of *in-situ* zircon U–Pb dating using SHRIMP (Froude et al., 1983; Stern et al., 2016) or LA-ICPMS (Hirata and Nesbitt, 1995) and zircon Lu-Hf isotopic analysis (Thirlwall and Walder, 1995; Vervoort et al., 1996; Knudsen et al., 2001; Griffin et al., 2004), many studies have contributed to the elucidation of continental growth history based on zircon analyses and data compilation (Rino et al., 2004; Rino et al., 2008; Condie et al., 2009; Hawkesworth et al., 2009; Belousova et al., 2010; Voice et al., 2011; Dhuime et al., 2012). By comparing the age structure given by Rino et al. (2004; 2008), older crust (before ca. 2 Ga) is known to be rare in modern continents, even after the exclusion effect of the intra-crustal sedimentary processes, as previously shown by Hurley and Rand (1969). Furthermore, based on compilations of the zircon U–Pb age and Lu–Hf isotopic analyses, several studies considered the crustal reworking in continental growth models (Belousova et al., 2010.; Komiya, 2011; Dhuime et al., 2012; Roberts and Spencer, 2015; Cawood and Hawkesworth, 2019). These studies can possibly evaluate the relative degree of crustal reworking and the fraction of new crust from the depleted mantle at each period. However, they cannot directly determine the volumes of crustal production and destruction because some assumptions for geotectonic processes are required. Subsequently, advanced statistical techniques or box model calculations have been adapted to interpret the zircon data compilation (Cheng, 2017; Puetz et al., 2017; Dhuime et al., 2018; Korenaga, 2018; Puetz and Condie, 2019), but the estimation of crustal production and destruction through time are still unclear.

To improve the understanding of the growth history of continental crusts, this study focuses on the secular change in the size of continental blocks as a major factor in determining the preservation and destruction of continental crust and as a physical parameter to constrain the evolution of the solid Earth. Previous studies on continental growth simplified the crustal differentiation and did not focus on a concrete geological entity of the continental crust. When the total amount of continental crust in the early Earth was not large, continental crusts should have existed as small continental blocks. With the increase in the total continental crust amount, the size of continental blocks should have increased through repeated amalgamations of smaller continental blocks. The size of continents affects the preservation of old crusts and the growth rate of the total amount because the production and destruction of new and preexisting continental crusts have occurred along the plate subduction zones, respectively. In the modern continental crust, the preservation of old crust occurs in the interior of continents, several hundred kilometers away from subduction zones (**Figure 1**). In contrast, if the size of continental blocks on the early Earth was substantially smaller than that of modern ones, the ratio of plate subduction margins to the total mass of the continental crust should have been larger than the present ratio. Additionally, small continental blocks are easily recycled into the mantle through continental subduction. As a result, the old crust was rarely preserved on the surface of the early Earth (**Figures 2A**,B). In the early Earth, plate subduction could have been limited, and the production of felsic rocks was mainly caused by a mantle plume (**Figure 2D**: e.g., Sizova et al., 2010; discussed in the next chapter). Before the start of plate subduction, the production and destruction of continental crust

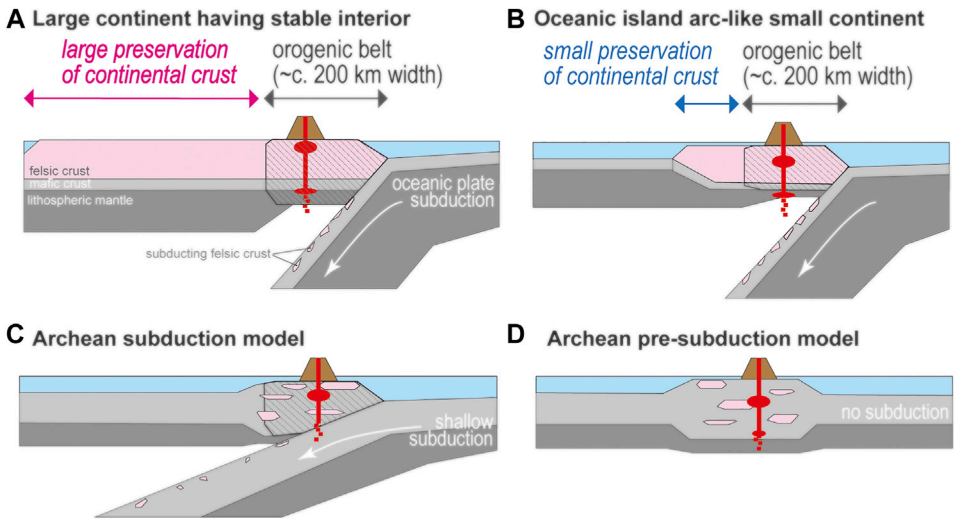

FIGURE 2 | Difference in the preservation of older continental crust from subduction orogen depending on the size of continental blocks **(A)** In a large continent, there are many areas of continental interior far from the subduction orogen and the older crust can be preserved **(B)** In a small continental block like the oceanic island arc, most of it is covered with subduction orogenic belt and there is only limited preservation of relatively older crust **(C)** Archean subduction model similar to modern oceanic island arcs **(D)** Archean pre-subduction model.

were inactive and would have occurred regardless of the size and other topographic features of the continental blocks. Nonetheless, the size of continents would significantly contribute to the preservation of the crust immediately after the plate subduction had begun. Furthermore, the secular change in the size of continents possibly reflects the evolutionary history of the planetary interior because amalgamation and rifting of continental blocks are controlled by plate motion and mantle convection. To determine the growth history of the continental crust, numerical calculations involving mantle convection and tectonic activities are ultimately necessary. Geological observations and petrological-geochemical signatures can work as constraints in the numerical simulation. Among them, the size of continents can be used more directly for such geotectonic simulations, as a special physical parameter. This study aims to examine the existing knowledge regarding the size of continents on the early Earth. Prior to this, the tectonic regime on the early Earth, especially with regard to the time of initialization of plate subduction and the generation of the granitic continental crust, is reviewed in the next section.

START OF PLATE SUBDUCTION AND PRODUCTION OF GRANITIC CRUST

There are many theories on the initiation of plate tectonics or plate subduction. Hereafter, this paper uses the term "plate subduction" to indicate the sheet-shaped continuous dropping of the lithosphere composed of the mafic oceanic crust and ultramafic mantle rocks into the deeper portion of the mantle. This definition offers a clear distinction between the Archean plate subduction and modern-style plate tectonics, which implies a continuous subduction of large, cooled, and rigid oceanic plates (**Figures 2A**,B). Archean oceanic plates and their subduction would have been different from their modern counterparts. Due to the hotter Archean mantle (Herzberg et al., 2010), more extensive partial melting occurred at the mid-oceanic spreading center to form an Archean oceanic crust that was less depleted and over 4–5 times thicker than at present (~7 km) (Sleep and Windley 1982; Abbott et al., 1994; Hastie et al., 2016). The Archean oceanic plates are considered to have been subjected to shallow dip angles underneath the other oceanic plates (**Figure 2C**; de Wit and Hart, 1993; de Wit, 1998; Komiya et al., 2002; Smithies et al., 2018; Martin et al., 2005; Ernst, 2009; Hastie et al., 2016; Ernst, 2017).

Clear geological records of plate subduction are as old as ca. 3.0 Ga (late Archean). Through geological observations and seismic profiling, the internal structures of cratons have been investigated. Many cratons preserve the crustal structure of terranes arranged in parallel and bounded by low-angle faults, suggesting the downward stacking of crustal blocks by repeated subduction and accretion processes; e.g., the Superior (Ludden and Hynes 2000; White et al., 2003; Angus et al., 2009), Yilgarn (Blewett et al., 2010; Czarnota et al., 2010; Goscombe et al., 2019), and Dharwar cratons (Mandal et al., 2017). A recent paleomagnetism study of the Honeyeater Basalt in the East Pilbara Craton suggests a modern plate motion velocity of ≥2.5 cm/year between ca. 3.4–3.2 Ga (Brenner et al., 2020). Conversely, plate subduction before ca. 3.0 Ga is controversial, as units >3.0 Ga are much rarer and their original geological structures are not well preserved. Several studies have identified ca. 4.0–3.8 Ga accretionary complexes with duplex structures and oceanic plate stratigraphy from the Isua Greenstone belt and Nulliak supracrustal rocks (Komiya et al., 2002, Komiya et al., 2015; Shimojo et al., 2016; Komiya et al., 2017). These could provide an important view of the early Earth subduction tectonics, but these interpretations still do not reach a consensus. Recent studies on the petrology and geochemistry consider that the start of plate subduction was between 3.5 and 3.0 Ga (e.g., Dhuime et al., 2012; Naeraa et al., 2012; Griffin et al., 2014; Dhuime et al., 2015; Tang et al., 2016; Brown and Johnson, 2018; Smithies et al., 2018; Smit et al., 2019), even though other studies argue that subduction was initiated in the early Archean or Hadean eras (e.g., Harrison, 2009; Hoffman et al., 2011; Polat et al., 2011; Furnes et al., 2014; Komiya et al., 2015; Ernst et al., 2016; Koshida et al., 2016; Greber et al., 2017; Maruyama et al., 2018; Sawada et al., 2018; Liu et al., 2020; Nutman et al., 2020). However, most of them have implications on the basis of some geochemical signatures. In these studies, a high occupancy of mafic rocks on the early Archean crust has often been regarded as an indication of no plate subduction. Instead, plume-related crustal production processes, such as oceanic islands or oceanic plateaus, have been envisaged for the ancient period. Even if plate subduction had started, a small amount of felsic crust would have occurred in the initial period because the first plate subduction should have started between two oceanic plates. For example, in oceanic island arcs like the Izu-Bonin-Mariana (IBM) arc occurring on a subduction zone between the Pacific Plate and Philippine Sea Plate, geophysical observations have detected primitive granitic continental crust with a thickness of over 20 km, in contrast to the basaltic or andesitic islands above sea level (Suyehiro et al., 1996). Most of the geochemical studies listed above mainly reflect the composition of materials in limited parts of the crust or mantle where the initiation of plate subduction is difficult to judge. Therefore, the discussion of subduction initiation is still inconclusive.

Limited plate subduction in the early Archean era has been suggested by estimating the buoyant oceanic plates because of the thick basaltic crust, approximately five times thicker than the present crust (Davies, 1992; Davies et al., 1995), or the highly depleted and mechanically strong peridotitic lithosphere (Davaille and Jaupart, 1993; Solomatov, 1995). Under a tectonic regime without plate subduction, a single plate referred to as the "stagnant-lid" covered the surface of the Earth, and crust formation was caused only by mantle plume upwelling (Solomatov, 1995). Recent numerical simulations of mantle convection combined with crustal production processes suggested a slightly more complex model of "lid-plume tectonics," wherein the intermittent sagging and dropping of the mafic crust into the mantle is assumed, known as sagduction (Sizova et al., 2010; Thébaut and Rey, 2013). They considered that Archean felsic continental crusts were generated through the melting of the mafic crust sagging into the mantle. Sagduction was originally proposed by the geological observation of Archean granite-greenstone terranes (Gorman et al., 1978; Goodwin and Smith, 1980). The similarity between the sagduction model for the Archean granite-greenstone terranes and geosyncline model for Pacific-type orogenic belts should be noted, which was predominant as a pre-plate tectonics idea (Dewey and Bird, 1970; Isozaki, 1996).

However, the claims against the Archean plate subduction can be refuted by several factors. Firstly, recent modeling of the MOR melting and thermal structure of the lithosphere based on the assumed Archean geotherm suggests that the Archean oceanic lithosphere was dense enough to subduct into the mantle (Weller et al., 2019). Furthermore, the thick mafic crust on the oceanic lithosphere can promote plate subduction because of the mineral phase-change to eclogite, which is denser than the mantle peridotite (Komiya et al., 2002; Komiya et al., 2004). Secondly, most numerical simulations of mantle convection have a serious problem in explaining plate subduction because they adopt a yielding model to express the behavior of the lithosphere (Tackley, 1998; Tackley, 2000). As observed in many geological records of orogenic belts and rifted basins, a fractured lithosphere does not adhere again, and convergence boundaries continue to exist at almost the same position; however, numerical simulations with a yielding model cannot reproduce such features (Ogawa, 2014). Additionally, the Peierls mechanism in rock rheological strength, which is exponentially dependent on stress, enhances the deformation at a significantly lower stress than the rheology with diffusion creep and power-law creep (Tsenn and Carter, 1987; Katayama and Karato, 2008; Demouchy et al., 2013; Azuma et al., 2017). This is presumed to have contributed to promoting plate subduction in the early Earth. Recently, several researchers have pointed out that the first fracture of the stagnant-lid lithosphere, produced by the solidification of the magma ocean, would have been shredded by meteorite impact events and allowed the initiation of subduction (Maruyama et al., 2018; O'Neill et al., 2020).

The petrology of the Archean tonalite-trondhjemite-granodiorite (TTG) series has been discussed with strong relevance to the initiation of plate subduction. The trace element composition of the Archean TTG is depleted of heavy rare earth elements and lacks Eu and Sr anomalies, which reflect the presence of garnet and the lack of plagioclase in the residual material (Barker and Arth, 1976; Smithies et al., 2003; Rapp et al., 2003; Martin et al., 2008). Two types of environments for the Archean TTG magma genesis have been advocated by many researchers: the hot plate subduction zone (**Figure 2C**; Martin et al., 2008; Hoffman et al., 2011; Nagel et al., 2012; Laurie et al., 2013; Martin et al., 2014; Roman and Arndt, 2020), and mantle plume heating of preexisting thick mafic crust (**Figure 2D**; Van Karnendonk et al., 2007; Kamber, 2015; Palin et al., 2016; Johnsson et al., 1991). The plume heating model is apparently supported by the computer simulation of the early crust and upper mantle evolution. This model showed a similar mechanism to produce felsic magma during the transition from the stagnant-lid regime to the plate subduction regime (Sizova et al., 2010). The computer simulation is based on the stagnant-lid model for the initial stage of the early Earth and possesses the problems discussed above. Conversely, the plate subduction model for the Archean TTG genesis has the strength to supply hydrated metabasite into the deeper regions where garnet is stable. Presently, it is difficult to definitively reject either model. Both plate subduction and plume heating processes for the TTG genesis may have occurred once in the Archean crust.

To date, there is no consensus on the timing of the start of plate subduction in the Archean or Hadean eras. Ancient plate subduction was probably intertwined with the mantle plume activity (Wyman et al., 2002), which would have been similar to the back-arc spreading activity in the Phanerozoic Earth. Based on the two models of the early Earth tectonic regime, the next chapter discusses how the primitive continental crust, such as the oceanic island arc or hot spot oceanic islands, developed into massive continental blocks.

SIZE OF CONTINENTS OVER TIME

Size of Continents and Terrigenous Clastic Rocks

It has long been postulated that continents in moderate size formed through collision and amalgamation of small continental blocks (de Wit and Hart, 1993; Hoffman, 1988; Hoffman, 1989), but the actual size of these small blocks has not been well understood yet. Based on modern continental blocks and plate margins (Inman and Nordstrom, 1971), this study classifies continental blocks as follows. Class 1 includes ca. 200 km wide continental crusts without large exposure above sea level like the IBM island arc along the Pacific subduction zone or the Hawaiian islands on a hot spot. Class 2 includes ca. 500 km wide continental blocks like the Japanese islands and Greater Sunda islands along the subduction orogenic belts or Iceland on a hot spot. Class 3 includes continental blocks greater than ca. 1,000–3,000 km in width and length like Greenland and Australia. Class 4 continental blocks are above ca. 5,000–10,000 km in width and length like North America. Considering this classification of modern geographic features as a guideline, we discuss the secular change in the size of continents over time.

The modern extant pre-Cambrian terranes can constrain the minimum size of continents at that time, but we need to investigate the actual size of the continents from geological records taking into account later continental break-up. To estimate the land surface area of continental blocks indirectly, terrigenous clastic rocks are useful because their composition and stratigraphy reflect the size of the provenances. Small continental blocks with lengths and widths less than several tens of kilometers yield only immature clastic rocks with compositions close to rocks in the provenance and various grain sizes. Larger continental blocks with lengths and widths of several hundred kilometers can produce clastic rocks comprised of detrital grains sorted by mineral species and grain size through prolonged weathering. On even larger stable continental areas, highly quartzose sandstones are characteristic of the cratonic cover (Dickinson et al., 1983). Subsequent studies, especially those of the Orinoco River in South America, have indicated that tropical weather and low relief topography play major roles in the formation of quartzose sandstone and the size of continents do not relate to its formation directly (Johnsson et al., 1991). Nonetheless, it is important to note that many quartzose sandstones have actually been observed as cratonic cover sequences on large continents. This geological observation is likely related to preservation potential of sedimentary basins. Additionally, the thickness of such quartzose sandstone strata is mainly controlled by tectonic setting. In Proterozoic and Phanerozoic continents of class 3 or 4, several kilometers thick quartzose sandstone strata deposited on continental rift basins and passive continental margins over nearly 100 million years as the continents were large enough to exist stably for a long period of time. Based on these considerations, here we focus on the quartzite sandstone for the estimation of the size of continents in conjunction with lithology of the sedimentary sequences.

Detrital zircon age patterns from sandstones also have the potential to estimate the size of the provenance area. It is well known that the complexity of a detrital zircon age pattern is largely controlled by the topographic features and tectonic setting (Cawood et al., 2012; Aoki et al., 2014). In sedimentary basins on small continental blocks, such as island arcs and oceanic islands, most detrital zircons are derived from young igneous rocks related to arc magmatism, and zircons far older than the depositional age are not found. In contrast, sedimentary basins in rift settings or passive margins on large continents have a supply of detrital zircon from multiple aged crusts in their provenance.

~1.8 Ga Continents in Modern Size

The Rodinia Supercontinent, formed ca. 1.2–0.8 Ga and rifted ca. 0.75–0.6 Ga, is the oldest supercontinent whose paleogeography is well constrained by paleomagnetic analysis and the succession of geological units (Moores, 1991; Karlstrom et al., 2001; Li et al., 2008; Evans, 2009). Although several Proterozoic supercontinents before 1.3 Ga have been proposed such as Columbia (Rogers and Santosh, 2003), Sclavia, Superia and Vaalbara (Bleeker, 2003), the paleogeography and size of the continents are still under debate (e.g., Bradley, 2011; Nance and Murphy, 2013; Piper, 2018). Nonetheless, the 2.0–1.8 Ga shield of the "Nuna continent" is preserved in modern North America and is one of the most substantial records for the existence of a continent above 5,000 km in width and length at least. Moreover, it also provides evidence for the successive growth of the continent size during the Proterozoic era (Hoffman, 1988; Whitemeyer and Karlstrom, 2007). The ~1.8 Ga middle Proterozoic quartzose sandstone stratum is over 1,000 m thick that occurs worldwide; for example, the ca. 1.7–1.6 Ga Sioux and Barron quartzites in the southern Lake Superior region of North America (Southwick and Mossler, 1984; Holm et al., 1998), ca. 1.8–1.6 Ga Waterberg Supergroup in southern Africa (Meinster and Tickell, 1975; Tankard et al., 1982), the ca. 1.7–1.6 Ga McArthur Group in Australia (Rawlings, 1999; Rawlings et al., 2004; Jackson et al., 2007), and ca. 1.8–1.3 Ga Cuddapah, Kaladgi, and Pranhita-Godavari basins in the Indian Peninsula (Saha et al., 2016). Deposition of these thick sedimentary units is considered to be due to stable large continents during the Proterozoic era.

2.4–1.8 Ga Early Proterozoic Sandstone and the Size of Continents

During the early Proterozoic era, between ca. 2.4 and 1.8 Ga, large-scale sedimentary sequences had formed on many continental rift basins and passive continental margins (Martin et al., 2013). The increased sedimentary sequences caused the great oxygenation of the surface environment of the Earth (Rye and Holland, 1998; Bleeker et al., 1999), and possibly the emergence of eukaryotes (Han and Runnegar, 1992; El Albani et al., 2010; Edou-Minko et al., 2017). The thickness of the quartzose sandstone strata in the continental rift basins and passive continental margins are over 1,000 m and are comparable to those after 1.8 Ga. For example, the 2.4–2.2 Ga Huron Supergroup on the Superior Craton and the Snowy Pass Supergroup on the Wyoming Craton deposited in a broad basin on the passive continental margin contain quartzose sandstone strata that are over several thousand meters thick (Roscoe and Card, 1993). The size of continents during the early Proterozoic period should have been similar to that after 1.8 Ga, i.e., continents of class 4. The continents may correspond to the previously named Sclavia and Superia (Bleeker, 2003) based on age dating and the paleomagnetic analysis of radial dyke swarms. Although the paleogeography of these continents is not well constrained, it is certain that continents as large as those after 1.8 Ga have existed since ca. 2.4 Ga.

Quartzose Sandstone and the Size of Continents in the Archean Crust

Before the ca. 2.5 Ga Archean-Proterozoic boundary, direct geological records for the size of continents are rarely preserved. Some continents larger than the modern-extant Superior Craton, over 1,000 km in width and length, must have existed at ca. 2.7 Ga (late Archean). Other, mostly Archean, cratons have a width and length of 500–1,000 km or less (**Figure 1**). The presence of Archean quartzose sandstone has been interpreted as an important indicator of cratonic interior on relatively stable large continents at the depositional age (Rogers, 1996; Eriksson et al., 2013; Cawood et al., 2018), even though the origin of Archean quartzose sandstone is subject to debate because the Archean surface environment and climate is different from modern ones (Condie, 1981). The absence of vegetation on the Precambrian land possibly hindered chemical weathering by accelerating the transport of clastic particles (Dott, 2003). On the other hand, a Precambrian CO_2-rich atmosphere could have promoted chemical weathering (Sleep and Hessler, 2006). Due to these differences in the continental surface, the conditions for the formation of quartzose sandstone would have been different between the modern and Precambrian environments. In the Archean supracrustal units, we can find both immature non-quartzose sandstones such as graywacke and mature quartzose sandstones as a cratonic cover for the Archean terranes (Condie, 1993). This indicates that the formation of quartzose sandstone requires flat topographic features even under the Archean atmospheric conditions. We proceed on the assumption that the conditions for quartzose sandstone formation were not extremely different from those at present.

Archean quartzose sandstone on a granitic basement with a clear unconformity can be traced back to 3.0 Ga; for example, the ca. 2.6 Ga Moeda Formation of the Minas Supergroup in the Sao Francisco Craton (Minter et al., 1990; Alkmim and Marshak, 1998; Alkmim and Martins-Neto, 2012; Koglin et al., 2014), ca. 2.7 Ga Hardey Formation of the lowermost Fortescue Group in the Pilbara Craton (Thorne and Trendall, 2001; Hall, 2005), ca. 2.8–2.7 Ga Central Slave Cover Group in the Slave Craton (Bleeker et al., 1999), ca. 2.9 Ga Steep Rock, Lumby Lake, Savant Lake, and other greenstone belts of the Superior Craton (Wilks and Nisbet, 1988; Davis and Jackson, 1988; Davis et al., 1988; Donaldson and de Kemp, 1988), ca. 3.0–2.8 Ga Lower and Upper Burawayan greenstone units on the Zimbabwe Craton (Wilson et al., 1995; Fedo et al., 1996; Hunter et al., 1998), ca. 3.0–2.9 Ga Witwatersrand and Pongola groups in the Kaapvaal Craton (Beukes and Cairncross, 1991; Robb and Meyer, 1995; Kositcin and Krapez, 2004), and so on. These quartzose sandstone strata mostly occupy the lowermost part of sedimentary sequences and are shallow marine deposits overlain by pelitic rocks, basaltic-komatiitic lava flow including pillow structure, and banded iron formation (Bleeker, 2003). This characteristic lithology indicates that the sedimentary sequences deposited on the continental rifting basin or passive continental margin. Quartzose sandstones within a similar lithological combination are also found in highly metamorphosed Archean terranes; for example, ca. 3.1 Ga Beit Bridge Group of the Limpopo belt (Eriksson et al., 1988), ca. 3.2 Ga Mt. Narryer complex, Jack Hills belt and Reynard Hills belts on the Yilgarn Craton (Myers, 1988; Spaggiari et al., 2007), ca. 3.2 Ga Beartooth Mountain on the Wyoming Craton (Maier et al., 2012), ca. 3.2–3.0 Ga Bababudan Group in the Dharwar Craton (Srinivasan and Ojakangas, 1986), ca. 3.0 Ga Keonjhar Quartzite in the Singhbhum Craton (Ghosh et al., 2016), ca. 3.0 Ga Naharmagra quartzite of the North Delhi belt in the Aravalli Craton (Raza et al., 2010), and ca. 3.3 Ga Sebakwian quartzite on the Zimbabwe Craton (Bolhar et al., 2017). Older Archean siliceous sandstones also exist, for example, the ca. 3.4 Ga Noisy Formation in the Kaapvaal Craton (de Wit et al., 2011), and ca. 3.5 Ga Mt. Goldworthy quartzite in the Pilbara Craton (Sugitani et al., 2003); however, these quartzose sandstone strata do not have the lithology of ~ca. 3.3 Ga continental rift basins or passive continental margins. Local hydrothermal silicification is considered to be a major factor for the origin of the ~ca. 3.3 Ga quartzose sandstone shown above.

It is notable that most of the Archean quartzose sandstone layers are less than a hundred meters thick, in contrast to Proterozoic ones (**Figure 3**). For example, the basal quartzite of the 2.8 Ga Central Slave Cover Group on the Slave Craton is a typical example, which is less than ~100 m thick (Bleeker et al., 1999; **Figure 3**). The Moeda Formation in the lowermost Minas Supergroup is less than 350 m thick (Minter et al., 1990). Most of the other Archean quartzose sandstone strata are similar or thinner than the Moeda Formation (Bleeker, 2001). The change in thickness of quartzose sandstone on continental rift basins or passive continental margins suggests that such basins with continental or shallow marine environments for deposition in the Archean era did not remain for long periods. The short life span of such basins can be explained by the smaller size of continents or thinner continental crust than those after the Proterozoic period. The model of thin continental crust is denied by seismic data of Archean terranes, which indicates that the Archean cratons were nearly 30 km thick since ca.

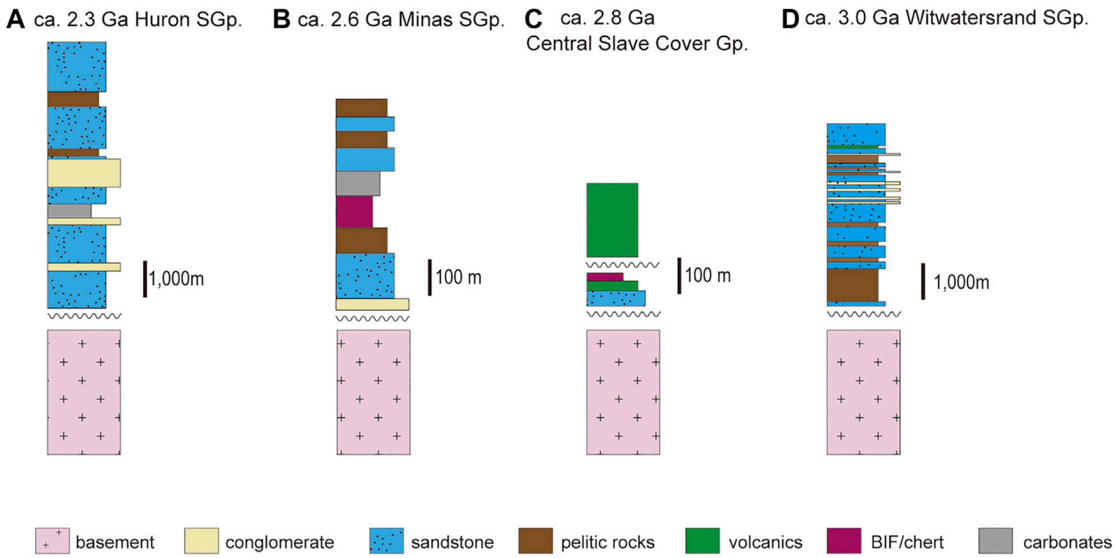

FIGURE 3 | Stratigraphic columns of representative quartzose sandstone layer from ca. 3.0 Ga to ca. 2.3 Ga. **(A)** ca. 2.3 Ga Huron Supergroup on the Superior Craton (Roscoe and Card, 1993). **(B)** ca. 2.6 Ga Minas Supergroup in the Sao Francisco Craton (Minter et al., 1990; Koglin et al., 2014); **(C)** ca. 2.8 Ga Central Slave Cover Group in the Slave Craton (Bleeker et al., 1999); **(D)** ca. 3.0 Ga West Rand Group of the Witwatersrand Supergroup in the Kaapvaal Craton (Frimmel, 2010).

3 Ga (Drummond, 1988). The smaller size of continents is more plausible for the reason of short life span of the depositional basins because the model can explain the short-lived continental break-up and subsidence. Therefore, the relatively thin ca. 3.3–2.5 Ga quartzose sandstones in the continental rift basin and passive continental margin imply the emergence of relatively stable continental interior environments, but the size of them was still not as large as that of the continents after the Proterozoic era, probably ~1,000 km in width and length, similar to class 3 continents. The exceptionally thick the ca. 3.0 Ga West Rand Group of the Witwatersrand Supergroup in the Kaapvaal Craton may indicate a slightly larger continent at that time than in the previous and subsequent ages, or that a local continental collision affected to increased sedimentary supply (Burke et al., 1986; Catuneanu, 2001).

Some of the older Archean supracrustal units contain well-classified clastic rocks, for example, the conglomerate in the ca. 3.5 Ga Warawoona Group on the Pilbara Craton (Buick et al., 1995). The well-sorted clastic rock indicates that the size of the continental block at that time was similar to that of class 2 continents. Before ca. 3.4 Ga, the size of the continents might have been less than several hundred-kilometers in width and length, similar to that of classes 1 and 2.

Additionally, detrital zircon age patterns in Archean quartzose sandstone and other clastic rocks are also important geological records for the estimation of the size of continents. The ca. 3.3–2.5 Ga quartzose sandstone and quartzite in the continental rift basin or passive margin show complex detrital zircon age patterns and often contain Eoarhcean and Hadean ones (e.g., Sircombe et al., 2001; Maier et al., 2012; Zeh et al., 2014; Bolhar et al., 2017), indicating that their detrital grains are widely provided from relatively large areas. Furthermore, it is noteworthy that Hadean detrital zircon grains occur in ca. 3.3–3.1 Ga sandstones. Ancient detrital zircon U–Pb ages are extremely rare in clastic rocks with depositional ages younger or even older than ca. 3.3–3.1 Ga (**Figure 1**). A lesser occurrence of Hadean zircon in clastic rocks younger than 3.0 Ga (late Archean) simply indicates that the Hadean geological units had been destroyed by crustal reworking and recycling by 3.0 Ga. Conversely, no occurrence of Hadean zircon in >3.3 Ga clastic rocks suggests that only small provenances existed because continental blocks were fragmented through rifting and sedimentary basins did not receive detrital grains from Hadean units. The detrital zircon age pattern in Archean rocks would also reflect the secular change in the size of continents.

Estimated Secular Change in the Size of Continents

The secular change in the size of continents estimated above is summarized as follows: 1) mature clastic rocks older than ca. 3.4 Ga have not been found, but ca. 3.5 Ga well-sorted clastic rocks occur in several cratons. This indicates that ca. 3.5–3.4 Ga continental blocks were in scales of several hundred kilometers or less. 2) During ca. 3.3–2.5 Ga (late half of Archean), quartzose sandstone of the continental rift basin or passive margin thinner than a hundred meters occurred on many cratons, indicating that the continents were nearly 1,000 km in width and length or more. 3) After ca. 2.4 Ga, quartzose sandstone strata thicker than 1,000 m are widely deposited worldwide. Since this period, the width or length of continents has been nearly 10,000 km until the present. **Figures 4A and B** are schematics of the above-described secular change in the occurrence of quartzose sandstone and the size of continents.

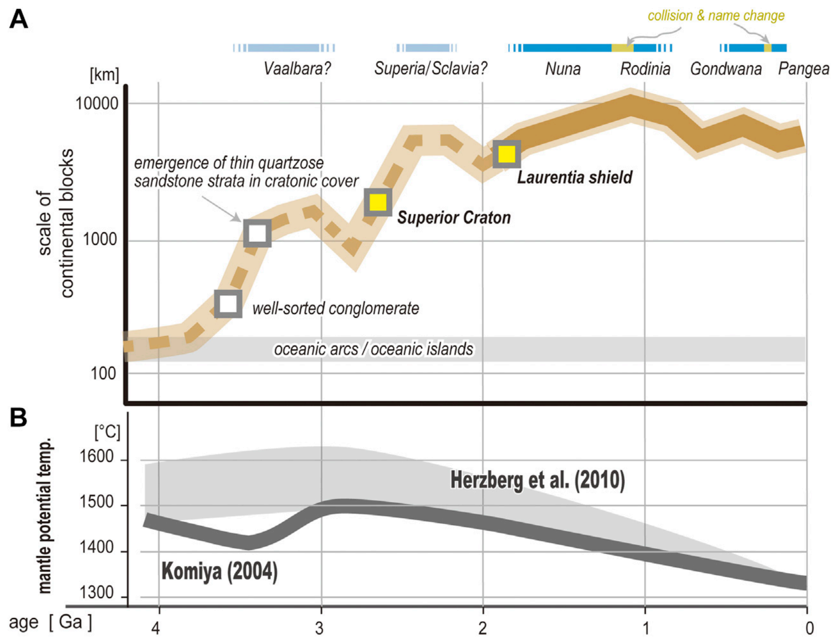

FIGURE 4 | (A) Schematic diagram of the maximum scale of continental blocks through the history of Earth in Supercontinental periods (Bradley, 2011). **(B)** Secular trend in the mantle potential temperature estimated through petrological methods (Komiya, 2004; Herzberg et al., 2010).

The secular change in the size of continents is shown in **Figure 5** as a global map. **Figures 5A and B** show the model of the early initiation of plate subduction. After the start of plate subduction in the Hadean and early Archean era, a narrow and thin continental crust formed like that in modern oceanic island arcs (**Figure 5A**). These primitive continental crusts had amalgamated to form slightly larger continental blocks over 500 km in width by ca. 3.5 Ga at the latest (**Figure 5B**). In contrast, **Figures 5A′ and B′** show the model of limited plate subduction on the early Earth. Early felsic crust was formed in hot spot regions by a mantle plume, dominated by a mafic crust and ~500 km in diameter (**Figure 5A′**). After ca. 3.5 Ga initiation of the plate subduction, the plume crusts had amalgamated to form felsic continental blocks (**Figure 5B′**). In any case, the primitive continental crust had further amalgamated and the first continents with a width greater than 1,000 km, larger than modern Greenland or Australia, appeared around ca. 3.3 Ga. The Archean continents repeatedly rifted and collided and had been smaller in size than those after the Proterozoic. A small peak in the size of continents might have occurred at ca. 3.2–3.0 Ga, as suggested by thick quartzose sandstone strata with eolian structures. The continents were likely rifted into smaller blocks after 3.0 Ga, which was related to the late Archean peak in the mantle potential temperature (Komiya, 2004; Herzberg et al., 2010; **Figure 4C**). By ca. 2.4 Ga, the continental blocks further amalgamated to form larger continents greater than 5,000 km, like modern North America. The stabilization of large continents during the early Proterozoic era was also possibly related to the decreased mantle temperature (Höning and Spohn, 2016).

WHAT WE NEED TO DO FOR FURTHER UNDERSTANDING

This study attempts to estimate the secular change in the size of continents by using limited information, mainly from the occurrence of quartzose sandstones. However, this estimation is still crude, and the size of continents before ca. 1.8 Ga is not well constrained. Unfortunately, most parameters in the petrology and geochemistry of igneous and metamorphic rocks and their minerals do not contain any information on the length, area, or volume, but only on the temperature, pressure, time, and the relative amount of fractionation. To obtain physical parameters related to the size of continents, more investigation on the Archean clastic sedimentary rocks is required.

Further information about the size of continents can be obtained from the U-Pb age pattern of detrital mineral grains because it can reflect the age structure of the provenance area on a continent. A compilation of detrital zircon ages and isotopic data were undertaken by combining large datasets and investigating the peaks in the compiled datasets (e.g., Condie et al., 2009; Belousova et al., 2010; Roberts and Spencer, 2015). However, such data compilations, without checking the depositional ages of the host clastic rocks, would obtain superimposed signals of the production and destruction of the continental crust over time, resulting in a misunderstanding of the true evolution of continents. To disclose the secular change in the age distribution pattern of detrital zircon grains, the age structure of continents, evolution of continents, and time-lapse analysis of the detrital zircon age pattern are important (Parman, 2015; Spencer et al., 2017; Sawada et al., 2018; Puetz and Condie, 2019).

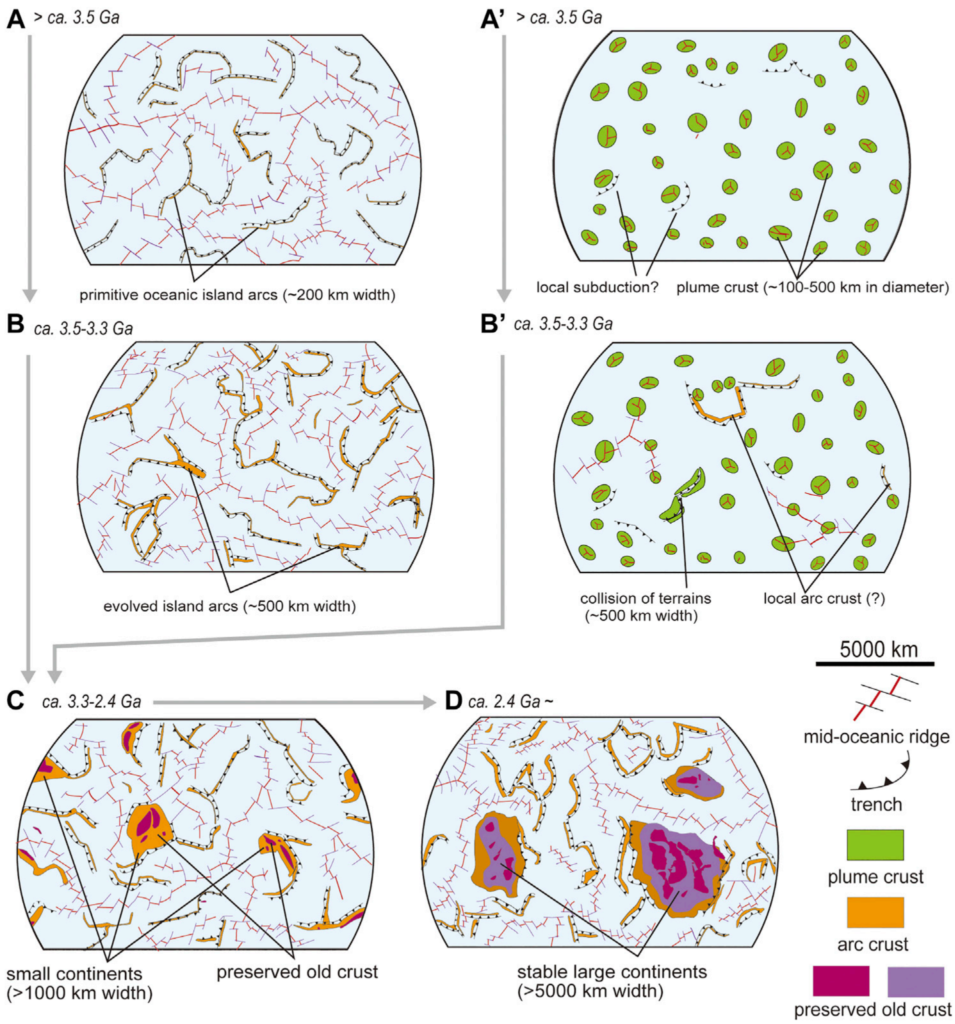

FIGURE 5 | Map view of secular change in the size of continents modified after Sawada et al. (2018). **(A)** After the start of the plate subduction, many narrow and thin oceanic island arc-like continental crust were produced. **(B)** Through the amalgamation of the narrow, thin continental crust, evolved continental blocks with a width of over 500 km. Old Hadean and early Archean crusts had been preserved in the interior of the continental blocks. **(A′)** Plate subduction was still limited and the thin felsic crust was conceived in plume-related mafic crust. **(B′)** After the start of plate subduction, the plume crust gradually amalgamated and larger continental blocks emerged. Production of arc crust likely started. **(C)** Larger continents with over a 1,000 km width appeared around 3.3 Ga. Most of the continents during this period were not as large as the modern ones, although some exceptions with nearly a 5,000 km width likely existed during 3.2–3.0 Ga. **(D)** The Archean continents finally amalgamated to form large continents similar to modern ones.

The differences in detrital zircon age patterns have the potential to determine the size of continents, especially during the early period in which geological units and structures are poorly preserved. Although a perfect correspondence between the size of continents and detrital zircon age patterns is not always achieved, it is possible to detect the general trend in the size of continents by using a large dataset. Detrital zircon Lu-Hf isotopic studies to estimate crustal reworking through time should be reexamined because Hf isotopic values are likely affected by the mineral composition of residual rocks during the partial melting and crustal assimilation processes according to recent detailed zircon analysis combined with geological and petrological observations (Chen et al., 2015; Petersson et al., 2019). Trace element analyses of zircon combined with Lu-Hf isotopic analysis could help in solving this problem as they reflect the parental magmatic composition and residual mineral assemblage well (Grimes et al., 2015). The detrital monazite U-Pb age also has the potential to reveal the crustal evolution, especially by metamorphism during the orogeny (Itano et al., 2016; 2018). Other minerals, such as apatite, titanite, allanites, and garnet also have the potential to contribute to our understanding of the early Earth. However, they are challenging to use for Archean detrital studies because of the high content of initial Pb and weaknesses for alteration (e.g., Chen et al., 2015; Seman et al., 2017). Presently, it is not easy to prelaunch the success of the massive chronological data by minerals other than zircon.

This study used the quartzose sandstone strata for the estimation of the size of continents, but the correlation between the occurrences and size of continents was not always perfect because of the intense chemical weathering caused by the Precambrian atmosphere. The sedimentary structures of the Archean sandstone strata depend on the preservation of the sedimentary basin in the Archean terranes. In contrast, the geochemistry of clastic sedimentary rocks is worth exploring. For example, by using whole-rock Zr/Sc vs. Th/Sc plots for modern turbidites, McLennan et al. (1990) pointed out that relatively hard zircon is more concentrated in sediments along passive continental margins than those along active margins. This is because the narrow and limited provenance along active margins allow the composition of clastic rocks to be close to igneous rocks, whereas clastic rocks on wide and large provenances along passive margins are strongly affected by sedimentary sorting. Likewise, it is possible to detect the size of the provenance from a whole-rock composition of clastic rocks, especially from relatively coarse-grained ones such as sandstones. Based on the sedimentary structure and mineralogical-chemical composition of relatively young, well-preserved Phanerozoic and Proterozoic quartzose sandstone and quartzite, chemical proxies need to be developed and applied to Archean ones. These expected chemical proxies should be applicable for metamorphosed and fragmented Archean quartzite, and more information for older continents can be obtained.

In the above estimation of the size of continents, some significant difficulties still exist in obtaining the actual evolutionary history of the solid Earth. Constraints on the amount of continental crust produced through magmatism and recycled into the mantle has been a major issue for solid Earth studies. Due to the small and thin continental crust in the initial stage of the plate subduction, like that of modern oceanic island arcs or hot spot islands, most of the primitive continental crust should have been subducted to the mantle without any detectable geological records (Santosh et al., 2009; Yamamoto et al., 2009; Spencer et al., 2017; Sawada et al., 2018). Although some geochemical signatures of subducted continental crust have been found from modern hot spot lavas (Jackson et al., 2007; Workman et al., 2008; Willbold and Stracke, 2010), it seems almost impossible to estimate the total amount and influx of the subducted continental crust during the Hadean and Archean periods through conventional, petrological, or geochemical methods for geological samples. To understand the surficial environments and the interior of the early Earth, more realistic computer simulations of the crust–mantle evolution are required. Modeling of the mantle convection in a 3D spherical shell is necessary, which is free from significant approximation and simplification, and is able to reproduce plate subduction based on the significant contrast of viscosity. Through these simulations, many parameters can be calculated, such as the speed of plate subduction, size and life span of plates, production and destruction rates of new continental crust, and movement and amalgamation of the generated continental crust. Geochemical and petrological studies can determine some of these parameters and evaluate the results of computer simulations. Combined with numerous petrological and geochemical data reported previously, estimating the size of continents must be relevant in the near future, and research on coarse-grained clastic sedimentary rocks can offer a solution for this purpose.

SUMMARY

This paper reviewed the crustal development in the early Earth and the current understanding of the size of continents over time. The size of continents is emphasized as an essential parameter to determine the geotectonic evolution of the continental crust. Clastic sedimentary rocks are possibly the key for estimating the size of continents because their composition reflects the physical processes occurring on the continental surface, such as transportation to the depositional basin. The present paper summarizes the tentatively estimated secular changes in the size of continents in the Archean and Proterozoic eras based on the occurrence of quartzose sandstones, which indicate a high degree of mineral sorting. We can trace back the existence of massive continents with widths and lengths of nearly 10,000 km like modern North America to ca. 2.4–2.3 Ga. The first continents, which were over 1,000 km in width and length, appeared at ca. 3.3 Ga. Most of the Archean continents were still not as large as those after ca. 2.4 Ga. Further analyses of clastic rocks can enable a more quantitative estimation of the size of continents, which can contribute to understanding the geotectonic history of the early Earth.

AUTHOR CONTRIBUTIONS

The first author, HS, constructed the discussion and interpretation of the data and participated in the preparation of the manuscript.

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