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GR Focus Review

LIPs, Orogens and Supercontinents: the Ongoing Saga

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Abstract

Of nine large age peaks in zircon and LIP time series < 2300 Ma (2150, 1850, 1450, 1400, 1050, 800, 600, 250 and 100 Ma), only four are geographically widespread (1850, 1400, 800 and 250 Ma). These peaks occur both before and after the onset of the supercontinent cycle, and during both assembly and breakup phases of supercontinents. During supercontinent breakup, LIP activity is followed by ocean-basin opening in some areas, but not in other areas. This suggests that mantle plumes are not necessary for ocean-basin opening, and that LIPs should not be used to predict the timing and location of supercontinent breakups. LIP events may be produced directly by mantle plumes or indirectly from subduction regimes that have inherited mantle-cycle signatures from plume activity. A combination of variable plume event intensity and multiple plume cyclicities best explains differences in LIP age peak amplitudes and irregularities. Peaks in orogen frequency at 1850, 1050, 600 Ma, which approximately coincide with major zircon and LIP age peaks, correspond to onsets of supercontinent assembly, and age peaks at 1450, 250 and 100 Ma correspond to supercontinent stasis or breakup. Although collisional orogens are more frequent during supercontinent assemblies, accretionary orogens have no preference for either breakup or assembly phases of supercontinents. A sparsity of orogens during Rodinia assembly may be related to incomplete breakup of Nuna as well as to the fact that some continental cratons never accreted to Rodinia. There are three groups of passive margins, each group showing a decrease in duration with time: Group 1 with onsets at 2.2-2.0 Ga correspond to the breakup of Neoarchean supercratons; Group 2 with onsets at 1.5-1.2 Ga correspond to the breakup of Nuna; and Group 3 with onsets at 1.5-0.1 Ga not corresponding to any particular supercontinent breakup.

New paleogeographic reconstructions of supercontinents indicate that in the last 2 Gyr average angular plate speeds have not changed or have decreased with time, whereas the number of orogens has increased. A possible explanation for decreasing or steady plate speed is an increasing proportion of continental crust on plates as juvenile continental crust continued to be added in post-Archean accretionary orogens. Cycles of mantle events are now well established at 90 and 400 Myr. Significant age peaks in orogen frequency, average plate speed, LIPs and detrital zircons may be part of a 400-Myr mantle cycle, and major age peaks in the cycle occur near the onset of supercontinent assemblies. The 400-Myr cycle may have begun with a "big bang" at the 2700 Ma, although the LIP age spectrum suggests the cycle may go back to at least 3850 Ma. Large age peaks at 1850, 1050, 600 and 250 Ma may be related to slab avalanches from the mantle transition zone that occur in response to supercontinent breakups.

Key Words

Large igneous province, orogens, supercontinent cycle, zircon age peaks

1. Introduction

We have a great deal of information on the supercontinent cycle, large igneous provinces (LIPs), mantle and crustal cycles, and orogenesis, but we do not fully understand if and how these events are related to each other in space and time. The supercontinent cycle has been with us for at least 2 Gyr and perhaps since the Neoarchean (Worsley et al., 1986; Brown, 2008; Evans, 2013) and has left a profound effect on geologic history. LIPs have left a record in the continental crust from 4 Gyr onwards and are thought by some to track mantle plume activity through time (Herzberg, 1995; Isley and Abbott, 2002). From the earliest remnants of continental crust onwards, we see the imprint of orogenesis and deformation, although the styles of orogeny appear to have evolved with time, and particularly with the onset of the supercontinent cycle. And finally, with the large numbers of precise U/Pb ages from zircons and

LIPs, it is possible to identify cycles in the mantle-crust system, some of which have been with us for at least 4 Gyr (Prokoph et al.,2004; Puetz and Condie, 2019).

In this study we focus on possible relationships between zircon and LIP age peaks, orogeny, plate speed and the supercontinent cycle. With an increasing and more robust paleomagnetic database for supercontinent reconstruction, it is becoming possible to track both LIP activity and orogenic activity during supercontinent assembly and breakup. How good is the alleged correlation with LIP activity and the breakup of supercontinents, and what is responsible for episodic and cyclic igneous activity through long periods of time? What is the timing and geographic relationship of orogens to supercontinent assembly and breakup and is average plate speed changing with time? It is these and related questions that we address in this contribution.

We also include in the Supplement paleogeographic reconstructions showing the distribution of both LIP sites and orogens for the last 2 Gyr (Supplementary Data, Fig. S1) and our current LIP (Supplementary Data, Table S1) and orogen databases (Supplementary Data, Table S2).

2. Methods

2.1 Paleogeographic Reconstructions

Our paleogeographic reconstructions are based on a combination of paleomagnetic data, marine magnetic anomalies, paleopositions of LIPs, passive and active continental margins, fossil correlations, correlation of basement provinces, and sedimentary basin and orogenic history. As new multidisciplinary evidence is continually published, some paleogeographic reconstructions become obsolete, and thus our new reconstructions represent a snapshot in time. These reconstructions from 2 Ga to 400 Ma are compiled at 100-Myr intervals, and at 50-Myr intervals from 400 Ma to the present (Supplementary Data, Fig. S1). The 400-0 Ma reconstructions are from Matthews et al. (2016), the 1000-500 Ma reconstructions from Merdith et al. (2017), and the >1000 Ma reconstructions are novel, but partly based on publications of Pisarevsky et al. (2013; 2014a,b; 2015), Lubnina et al. (2017), Ernst et al. (2013), Cederberg et al. (2016), Hoffman (2014), and on new unpublished paleomagnetic data. Details of paleomagnetic poles and Euler rotations for the time interval 2000-1100 Ma are given in Supplementary Data, Tables S3 and S4. All reconstructions are on Robinson global projections and in a paleomagnetic reference frame. We did not consider true polar wander (Evans, 2003), which is still debated and its implications to global paleogeographic reconstructions, especially in the Precambrian, are speculative. Also, true polar wander involves the rotation of Earth's mantle and crust, so it would not change our conclusions, which are related to the paleogeographic relationships between LIPs and orogens.

The longitudinal uncertainty of paleomagnetic data can be partly overcome by comparing contemporary segments of apparent polar wander paths of two or more continents. However, the small amount of reliable pre-1000 Ma paleomagnetic data results in limitations to this approach (e.g. Pisarevsky et al., 2014a,b). The hypothetical stationarity of two large low shear-wave velocity provinces (LLSVPs) recently was proposed as a reference frame for determination of paleolongitude (Torsvik et al., 2014). However, this hypothesis is still debated and has been applied only to the Phanerozoic. In our 2000-1100 Ma reconstructions, we mostly use minimal continental movement to constrain paleolongitude.

Several published pre-1000 Ma reconstructions consider a long-lived connection between Amazonia, West Africa and Baltica (such as the SAMBA model of Johansson, 2009). However, this model contradicts both paleomagnetic and geological data (Pisarevsky et al., 2014a; Bogdanova et al., 2015; Ibanez-Mejia et al., 2011), and we did not include it in our reconstructions, instead following the results of Pisarevsky et al. (2013, 2014a).

Recent studies have suggested that the supercontinent cycle may not be a simple assembly-breakup cycle. For instance, Li et al. (2019) propose that supercontinent assembly alternates between dominantly extroversion and dominantly introversion and exhibits both the classical short-term cycle of 500-700 Myr and a longer 1000-1500 Myr cycle related to the lifetime of superoceans. Merdith et al. (2019), based on studies of Gondwana assembly, suggest that the supercontinent cycle is either a two-stage cycle or that the last 1 Gyr is dominated by a single supercontinent with brief periods of dispersal and assembly.

One can debate the timing of assembly and breakup of supercontinents based on the oldest craton collisions or fragmentations, respectively, and for that reason we use the first "widespread" collisions or fragmentations to date the onset of these events (Table 1). We calculate the mean angular velocity for each 100-Myr bin by normalizing to the area of large continents on the reconstructions (for details see section 2.2 in Condie et al., 2015a). We also estimate geometrical centers of these continents and calculate distances between each pair of continents for the time slices of 0, 600, 1200 and 1900 Ma. Uncertainties in both plate speed and distances between continental fragments increases with age.

2.2 Orogens

In this review we divide orogens into two categories: accretionary and collisional. Accretionary orogens occur along convergent plate boundaries and collisional orogens develop when continental plates collide. There is a continuum between accretionary and collisional orogens, and most accretionary orogens eventually evolve into collisional orogens. A major source of uncertainty in counting orogens is that of what to count as a single orogen (Condie et al., 2015a). Orogens of short strike length could be part of a longer orogen, now displaced by supercontinent breakup, and in this respect, most orogens are really orogen segments (Supplementary Data, Table S2). In some cases an orogen segment may represent a complete orogen, whereas in others, it may represent only part of an orogen that was originally much more extensive. This problem is especially difficult when orogens wrap around cratons with "swirly" patterns as they do in Gondwana. In these cases, no more than one orogen segment is counted along a given craton margin. In very long orogens, such as the Great Proterozoic Accretionary Orogen (Fig. 1b), segments of the orogen are well studied and given names (Supplementary Data, Tables S2 and S7). By definition, collisional orogens end with continent-continent collisions. Accretionary orogens, on the other hand may end by subduction of an ocean ridge, regional plate reorganizations, a change in plate boundary from convergent to transform (such as the San Andreas fault), or collision of a major terrane or continental island arc (Cawood et al., 2011). A major terrane collision may shut down activity in one segment of an orogen and initiate activity along strike in another segment. Very often collisional and accretionary orogens can develop simultaneously with supercontinent assembly.

In supercontinent reconstructions (Supplementary Data, Fig. S1), we show the distribution of orogens that have been described in the literature, and these are summarized in Supplementary Data, Table S2, updated from our 2015 compilation. In Figures 1, 6 and 7 we show possible or probable interconnections of some of these orogens, which we refer to as linked orogens (Supplementary Data, Tables S2 and 2), the number of which increases with time. Because most orogens evolve from accretionary to collisional, the same orogen may be accretionary on one reconstruction and collisional in a later reconstruction.

2.3 Large Igneous Provinces (LIPS)

Large igneous provinces, commonly known as LIPS, are largely mafic magmatic provinces erupted in relatively short periods time of 20-100 Myr. Although many investigators have used areal extent to constrain size (Ernst and Bleeker, 2010), this is difficult to apply to ancient LIPs where most of the LIP is removed by erosion or not exposed. This is especially so when greenstone basalts and komatiites are included as is the case for our database. Because the areal extent of any given LIP is not well constrained for rocks older than about 500 Ma, we adopt a size cut-off of 10^4 km^2. In this study, we include as LIPs, giant dike swarms, continental flood basalts, plume-related basalts and komatiites, and oceanic plateau basalts (Condie et al., 2015b). When more than one period of magmatism is recorded by a single LIP, we use the major period of magmatism as the age, and the range of ages for single LIPs to assign uncertainties to the ages. We use the total geographic distribution of single LIPs of a given age to define a LIP event. Closely spaced LIP age peaks are grouped into single LIP events as described in detail in Condie et al. (2015b). The duration of LIP events is typically 20–50 Myr, but some events may be >100 Myr, and in some cases may represent more than one unresolved LIP event. Although most LIP events also leave a record in ocean basins (as oceanic plateaus and islands), this record is not well preserved before 300 Ma except as minor remnants that were accreted to the continents. Because oceanic LIPs older that about 200 Ma are rarely if ever preserved in the geologic record, our study focusses on continental LIPs only (Supplementary Data, Table S1).

2.4 Cycles and Time Series Analysis

Cycles in Earth history have long been of interest, beginning with the classical studies of Umbgrove (1940) who first suggested that terrestrial cycles may have origins deep within Earth. He recognized a periodicity in both orogenic and magmatic cycles as well as in sea level. Natural cycles occur over a broad range of frequency from days to billions of years and can be broadly divided into four categories (Mitchell et al., 2019). Oribital cycles and oceanic cycles, which are related to changes in Earth's rotational characteristics, occur on a scale of a few days to a few thousand years. Astronomical cycles, such as changes in planetary orbital eccentricities and obliquities occur on time scales of a few thousand to tens of millions years. Geodynamic cycles, which are related to thermal and density changes in Earth's interior, occur on scales of hundreds of millions to billions of years, and it is these cycles we concentrate on in this study.

There has been considerable discussion about the causes of geodynamic cycles, with the most direct information coming from experimental and numerical modeling related to mantle plumes. The rapid evolution of numerical computing in recent years provides a means of verifying complex numerical codes for convection, subduction, plumes, and mantle compositional variation (Davaille and Limare, 2015). Mantle plumes are common features of thermal convection at high Rayleigh number (Olson *et al.*, 1987), and develop on time scales of 30-200 Myr. Experimental studies suggest that there may be two types of mantle plume events: those associated with insulation or isolation of mantle beneath supercontinents and those not associated with supercontinents, but created by the return flow of slab avalanches from the mantle transition zone (Gurnis, 1988; Coltice et al., 2007). As thermochemical plumes rise, the plume material becomes denser and may sink back to the bottom, whereby the whole process can happen again. Several overturn episodes are observed in experiments, but the later ones become progressively more disorganized. Applying these results to the mantle, plume recurrence times of 100–200 Myr are predicted, depending on the viscosity and the amount of internal heating. This is in good agreement with the strong 93 and 187-myr cycles observed in both LIP and zircon time series (Puetz and Condie, 2019).

In this study, we use standard methods of time-series analysis for analyzing plate speed, orogenic activity, and zircon and LIP age distributions. These methods include time-series plots to illustrate variation in a signal over time, lowpass filtering with Gaussian kernels to remove a trend, and cross-correlation analysis to determine the degree to which a detrended time-series leads or lags a periodic model. Details of these methods are given in Puetz and Condie (2019).

3. Results

3.1 The Supercontinent Cycle

3.1.1 Nuna Assembly (1900-1500 Ma)

The first real supercontinent, Nuna, formed in the Paleoproterozoic. Although this supercontinent is also known as Columbia (Meert, 2012), we prefer the name Nuna, which is widely used in the literature today. Assembly of Nuna began about 1900 Ma with collision of small cratons, many derived from the breakup of Neoarchean supercratons (Evans and Pisarevsky, 2008; Evans, 2013), with the final amalgamation at 1600-1500 Ma (Table 1; Fig. 1 and Supplementary Data, Fig. S1). Paleomagnetic data suggest that West Africa, Amazonia and perhaps North China were never part of Nuna, or became part of Nuna only for a short time (Pisarevsky et al., 2014a).

As evidenced by the geographic distribution of collisional orogens between 1900 and 1800 Ma, dispersed micro-cratons collided to produce Laurentia, Siberia, Baltica, Australia-Antarctica and North China, forming the core of Nuna (Figs. 1b and Supplementary Data, Fig. S1). The frequency of both accretionary and collisional orogens increased rapidly at ~1900 Ma and then decreased to ~1800 Ma (Figs. 2 & 3). As expected, collisional orogens are mainly internal and accretionary orogens external in the supercontinent, and the Great Proterozoic Accretionary Orogen (GPAO) began to form around 1800 Ma, propagating along the western coast of Laurentia (present-day coordinates) into western Baltica, and then possibly into India or Amazonia (alternative reconstruction; Johansson, 2009; Evans and Mitchell, 2011; Evans, 2013). The number of cratons decreased from 45 to 22 as the building blocks of Nuna assembled (Condie et al., 2015a). By 1600-1400 Ma, there are only a few collisional orogens still active. The final collisional assembly of Nuna at 1600-1500 Ma occurred at the northern (Australia, Mawson, Laurentia) and southern (India, Kalahari [KPV-ZBW], Congo) ends of the supercontinent (Fig. 1b). These include orogens that reflect the last stages of craton convergence in Antarctica-Australia (such as Kararan, Olarian, Racklan-Forward) and if the age is extended to 1300 Ma, Albany-Fraser and Kibaran orogens can be included [Supplementary Data, Fig. S1]). Five small accretionary orogens (Picuris, Pinwarian, Hallandian-Danopolonian, Gothian, Telemarkian) are all part of the long-lived GPAO, where episodes of activity continued until at least 1450 Ma. Accretionary orogens are also known or likely to have been active along the coasts of Kalahari (KPV+ZBW), North China (Beishan orogen), and Amazonia, although North China and Amazonia may not have been part of Nuna. Although there may have been a long-lived accretionary orogen along the eastern margin of Nuna, there is not enough evidence at present to support the existence of such an orogen (Fig. 1b and Supplementary Data, Fig. S1).

LIPs (large igneous provinces) occur around the perimeter and center of the growing supercontinent as well as in outliers (Fig. 1). During assembly of Nuna, the number of LIP sites drops, especially after 1800 Ma (Fig. 4). This is opposite to what is expected if supercontinent insulation progressively becomes more important as assembly continues. Three peaks in LIP and in both detrital and igneous zircon age spectra at 2120, 2180, and 2215 Ma correspond to breakup of Archean supercratons, all of which may be part of a 90-Myr mantle cycle (Fig. 5; Supplementary Data, Table S5). LIP activity shows four age peaks during the assembly of Nuna at 1880, 1750, 1630 and 1590 Ma, of which only the 1880 and 1630-Ma peaks may be part of a 90-Myr mantle cycle.

3.1.2. Nuna Breakup (1450-1200 Ma)

Although Nuna began to breakup around 1450 Ma (Supplementary Data, Fig. S1), there has been much discussion and debate about the degree to which it fragmented (Pesonen et al., 2012; Roberts, 2013; Meert, 2014; Kirscher et al., 2021). As shown in Figure 6a, based on recent paleomagnetic data (Pisarevsky et al., 2014; Lubnina et al., 2017), it would appear that at 1200 Ma Nuna significantly fragmented, although the core of Nuna may have survived (Kirscher et al., 2021). As pointed out by Meert (2014), some parts of Nuna ("strange attractors") were not fragmented (Congo-Sao Francisco-Tanzania, and Mawson-Australia) or reconnected in similar positions (Laurentia-Baltica). LIPs are widely distributed on continental plates at both maximum packing (1500 Ma, Fig. 1b) and maximum dispersion stages of Nuna (1200 Ma, Fig. 6a). During the breakup of Nuna, the number of LIP sites is variable (~5-20; Fig. 4), but the paleogeographic distribution remains large (Supplementary Data, Fig. S1). There are three large LIP age peaks that occur during supercontinent breakup (1450, 1400, 1260 Ma [Fig. 5]), of which the 1450 and 1260-Ma peaks may be 90-Myr cycle peaks and the 1450-Ma peak may also be a 400-Myr cycle peak (Supplementary Data, Table S5). The 1450 and 1400-Ma peaks are very widespread, whereas the 1260-Ma peak has more limited distribution (Supplementary Data, Fig. S1), and none of the peaks shows preference for the interior or exterior of the dispersing supercontinent. The LIP and the detrital zircon age peaks at 1450 Ma coincide with the onset of Nuna breakup, and maximum plate dispersion at 1200 Ma roughly coincides with a minimum in LIP activity at 1150 Ma.

During the breakup of Nuna, collisional orogens are relatively few in number in the dispersing blocks, whereas accretionary orogens are widespread (Figs. 2, 3, and Supplementary

Data, Fig. S1; Supplementary Data, Table S2). It is possible that at 1450 Ma, the Picuris, Pinwarian and Hallandian-Danopolonian accretionary orogens were still connected as part of the GPAO. Only three collisional orogens are associated with Nuna breakup: 1) the Albany-Fraser-Arunta in Australia-Antarctica, 2) the Southern Grenville-Amazonia Sunsas between Amazonia and Laurentia, and 3) the Kibaran between Congo and Tanzania; collisional activity in these orogens persisted to ≤ 1250 Ma.

3.1.3. Rodinia Assembly (1100-850 Ma)

Most reconstructions of Rodinia agree that much of the core of Nuna either survived or was recombined in a similar configuration in Rodinia (Figs. 6b and Supplementary Data, Fig. S1) (Cawood et al., 2010; Pesonen et al., 2012; Meert, 2014). India, South China, and northern and central Africa appear not to have been part of Rodinia. During the assembly of Rodinia, the number of LIP sites is highly variable, with a striking minimum around 1000 Ma (Figs. 4, 5). Major LIP occurrences are in Congo-Sao Francisco, North China, South China, Siberia, Baltica and Antarctica-Australia-Laurentia. There are three LIP age peaks during assembly (1100, 920 and 810 Ma), all three of which may be part of a 90-Myr mantle cycle, and the 1100-Ma peak may also be part of a 400-Myr cycle (Supplementary Data, Table S5). Corresponding zircon age peaks are around 1050 and 800 Ma, the latter of which approximately corresponds to maximum packing of Rodinia; however, there is a trough in the zircon time series at 900 Ma corresponding to the LIP age peak at this time. The 1100 and 810-Ma LIP age peaks are the only peaks that may be global in extent.

Of the 25 orogens corresponding to Rodinia assembly, 15 are collisional and 10 accretionary and they occur in both the core of Rodinia and in dispersed cratons (Fig. 6 and Supplementary Data, Fig. S1). The oldest collisional orogens record the assembly of the core of Rodinia at 1200-1000 Ma (Grenville, Amazonia-Sunsas, Sveconorwegian, Arunta, Namaqua-Natal orogens), but by 900-800 Ma there are no active collisional orogens in the core. Collisional orogens not part of the assembly of the core of Rodinia include the Eastern Ghats and Central Indian Tectonic zone orogens (1060-900 Ma) recording the collisions of India with Rayner (part of Antarctica) and of North and South India, respectively, and the Qinling and Jiangnan orogens that record amalgamation of the Yangtze and Cathaysia cratons in South China (880-680 Ma) (Fig. 6b). As expected, most of the accretionary orogens occur around the margin of Rodinia

(Arctic, Amazonia Oaxaquia, Yenisei Ridge, Verkhoyansk, Southwest Tarim, Carris Velhos, Xiong'er, and Putumayo), or along the margins of dispersed cratons such as those that comprise central and northern Africa today. Although part of the core of Rodinia, at 950-850 Ma Siberia is surrounded on three sides by accretionary orogens (Fig. 6b).

3.1.4. Rodinia Breakup (750-600 Ma)

Rodinia broke up, probably by extroversion in a relatively short time span of ~150 Myr, but the fragmentation was far from complete (Cawood et al., 2016; Murphy et al., 2020) (Fig. 7a). In particular, Laurentia-Baltica-Amazonia appears to have survived this breakup. There is high variability in the number of LIP sites between 800 and 600 Ma, with a large trough in frequency of LIPs at about 700 Ma, and this time is also a minimum in zircon ages (Figs. 4, 5). The LIP age peaks at 720 Ma and 810 Ma approximately coincide with initial breakup. There are four regions with high LIP concentrations that are associated with Rodinia breakup: 1) Antarctica-Australia-Laurentia (836-788 Ma), 2)Yangtze-Cathaysia (828-756 Ma), 3) Laurentia-Siberia (726-719 Ma) and 4) Laurentia-Baltica (610-556 Ma) (Table 3; Supplementary Data, Fig. S1). All four LIP events are short-lived, and none survives for more than 100 Myr. The Antarctica-Australia-Laurentia, Laurentia-Siberia and Laurentia-Baltica LIP activity are followed at 600-500 Ma by craton breakups, whereas Yangtze and Cathaysia are still joined 600 Ma and beyond. During Rodinia breakup, there are two peaks that may be part of a 90-Myr cycle (720 and 615 Ma), and one peak (775 Ma) that is not part of this cycle (Fig. 5; Table 3). With exception of LIPs concentrated along the Australia-Mawson-Laurentian borders (Fig. 6b), there is no relationship between the locations of LIP cycle peaks and supercontinent breakup. The 775-Ma non-cycle peak is widespread in Yangtze-Cathaysia (775-800 Ma) and does not precede breakup. Between 600 and 500 Ma, Laurentia completely separates from Amazonia and Siberia, as does Siberia from Baltica, and at the same time Gondwana begins to assemble from the cratons now in Africa, South America, Australia and Antarctica (Supplementary Data, Fig. S1).

There are over 20 orogens accompanying the breakup of Rodinia (750-600 Ma) (Supplementary Data, Table S2; Supplementary Data, Fig. S1), and as expected, most are accretionary; only after 650 Ma do collisional orogens become widespread as Gondwana begins to assemble from cratons largely in Africa, South America, Antarctica and Australia. Between 900 and 750 Ma, accretionary orogens developed along the margins of Congo, Laurentia, Siberia, Baltica, North China and West Africa (Fig. 6b and Supplementary Data, Fig. S1), and between 750 and 550 Ma, most orogens and major LIP activity are geographically widely separated.

3.1.5. Gondwana (600-450 Ma) and Pangea (400-300 Ma) Assembly

Numerous studies have addressed the assembly of Gondwana, an in part, we rely on results of these studies in our reconstructions (Stern, 1994; Meert, 2003; Gray et al., 2008; Brito-Neves et al., 2014). As with Nuna, the number of continental LIP sites drops rapidly as Gondwana assembles in the Southern Hemisphere (650-450 Ma), but then increases again beginning about 300 Ma as Pangea assembles (Fig. 4). There is a major age peak at about 600 Ma in both LIPs and zircons and this coincides with the beginning of assembly of Gondwana (Fig. 5). This peak is well represented by the LIP activity (Central Iapetus Magmatic Province) preceding the breakup of Laurentia and Baltica and opening of the Iapetus Ocean at 550 Ma (Fig. 7 and Supplementary Data, Fig. S1). At 511 Ma, new LIP activity appears in Australia-East Antarctica. All LIP age peaks accompanying the assembly of Gondwana-Pangea (560, 440, 380, and 260 Ma) may be part of a 90-Myr mantle cycle and the 440 and 260-Ma peaks may also be part of a 400-Myr cycle (Table 3). Although assembly of Pangea begins at ~400 Ma, the major step in this assembly is the collision of Gondwana with Laurentia-Baltica at about 320 Ma, and a large age peak in both zircons and LIPs occurs at 260 Ma, coincident with completion of the supercontinent (Fig. 5).

Between 600 and 500 Ma there are 42 orogens of which 24 are collisional and 18 are accretionary (Supplementary Data, Table S2; Figs. 2 and 3). The similar frequency of each type of orogen reflects the ongoing breakup of Rodinia while Gondwana is beginning to assemble in the Southern Hemisphere. Peaks in frequency of both accretionary and collisional orogens occur at 650-500 Ma. Most of the collisional orogens are part of the Pan-African System widespread in Africa, Antarctica and South America. During assembly of Gondwana, accretionary orogens occur along the margins of the opening Iapetus Ocean, in dispersed cratons, and along the southern margin of Gondwana (Fig. 7b and Supplementary Data, Fig. S1). Between 400 and 250 Ma, there are 31 orogens, of which 4 are collisional and 28 are accretionary; however, many of the accretionary orogens rapidly evolved into collisional orogens as Northern Hemisphere landmasses collided. As Pangea assembled, orogenic action is mostly around Laurentia and Siberia as they collided with each other and with Gondwana. By 250-200 Ma, the Terra Australis superorogen had propagated from the southern margin of Gondwana all the way into Laurentia (Cawood, 2011). Also, the geographic relationship of LIPs to LLSVPs begins only at 300-200 Ma (Fig. 7b) (Li and Zhong, 2009; Doucet et al., 2020), suggesting that the Atlantic LLSVPs did not form before this time.

3.1.6. Pangea Breakup (180-0 Ma)

At 200 Ma, LIP action shifts from around the Tethys to the Atlantic LLSVP where the Atlantic Basin will open. LIP action begins in the South Atlantic at 130-110 Ma (Ulvrova et al., 2019), and LIPs in South America and South Africa at 135-130 Ma precede opening of the Atlantic Ocean. Opening of the Central Atlantic at 190-170 Ma is preceded by LIP activity at 200 Ma in Laurentia and Northwest Africa. The breakup of North America and Greenland at 70-50 Ma is preceded by major LIP activity at 100-80 Ma, and the breakup of the Northeast Atlantic (50-30 Ma; Ulvrova et al., 2019) is preceded by LIP activity at 60-50 Ma. In contrast to opening of the Atlantic, opening of the Indian Ocean begins about 150 Ma, but most LIP activity occurs after this time at about 100-90 Ma (Figs. 4, 7 and Supplementary Data, Fig. S1). The number of LIP sites on continents and ocean basins increases rapidly from 200 to 100 Ma (Fig. 4), but from 50 Ma onwards, most LIP action on the continents decreases in frequency.

The LIP age peak at 100-90 Ma may be part of a 90-Myr mantle cycle (Table 3), and this peak is best represented by LIP activity in opening ocean basins related to the breakup of Pangea (Supplementary Data, Fig. S1). It is represented on eight continental plates and is also well developed in the zircon age spectra (Fig. 5). This peak correlates with abrupt widespread changes in plate motions and boundary configurations (Matthews et al., 2012), and it may reflect, in part, increasing frequency of collisional orogens (Figs. 3 and 5) and associated high relief of mountains (Condie et al., 2015a), particularly in South-Central and Southeast Asia. Most LIPs at 100-90 Ma appear to be related to plumes coming from the north and south ends of the Atlantic LLSVP and also from the South Pacific LLSVP. The 260 and 440-Ma LIP age peaks may be part of a 400-Myr cycle (Table 3).

There are 34 recognized orogens between 385 Ma and 34 Ma, of which 15 are collisional and 19 accretionary. However, this division is rather arbitrary since all of the collisional orogens began life as accretionary orogens, and the classification is really a snapshot of a stage in orogen evolution. Most of the collisional orogens are associated with the Alpine-Himalayan System, and most of the accretionary orogens are part of the peripheral orogenic system surrounding Pangea, just before it began to fragment (Fig. 7b and Supplementary Data, Fig. S1). These occur along the west coasts of North and South America and along the coast of Eastern Asia. Orogens in which the accretionary phase rapidly evolved into a collisional phase are produced as terranes were rifted off Gondwana and traveled north across the Tethys Ocean basin to collide with Asia.

4. Large Igneous Provinces (LIPS)

We now have 916 entries in our ever-growing LIP database (Supplementary Data, Table S1). The time series of this database (Fig. 5) is very similar to our 2019 database (Puetz and Condie, 2019) with only 529 entries, attesting to the representativeness of the earlier database. Overall characteristics of both the LIP and zircon age time series are described in earlier publications (Puetz and Condie, 2019; Condie and Puetz, 2019). Our data show that there is no consistent pattern between the number of LIP sites, the number of cratons on which LIPs occur, and LIP age peaks (Fig. 8). However, there is an overall decrease in the number of total LIP sites with time from 2500 Ma to about 500 Ma (Fig. 4), a feature that may record cooling of the mantle. In supercontinent reconstructions, LIP sites are strongly concentrated in some geographic regions (Supplementary Data, Fig. S1), a feature may be due to either or both, 1) a localized concentration of mantle plumes of approximately the same age, or 2) a high frequency of studies in these regions.

Because oceanic lithosphere covers most of Earth's surface, and most of the evidence for plumes is oceanic LIPs, most of this record has been lost by subduction. Thus, the only evidence for most LIP events > 200 Ma is the continental LIP record (Ernst and Bleeker, 2010). With our new database (using a 20-Myr bin size and excluding oceanic sites), we now recognize eight large detrital zircon age peaks < 2300 Ma in a detrended time series (2150, 1850, 1450, 1050, 800, 600, 250 and 100 Ma), of which all eight have corresponding LIP age peaks (1050 Ma zircon peak is at 1100 Ma; Figs. 5 and 8). In addition, there is a strong 1400-Ma peak in the LIP spectrum. Only three peaks are geographically widespread (> 10 sites per peak; 2180, 2050, and 1850 Ma) (Supplementary Data, Table S1). There are 31 LIP age peaks < 2300 Ma, and each peak is represented by 5-25 sites, except for peaks at 1850 and 1050 Ma, which have ≥ 35 sites per peak. It is noteworthy that these two peaks occur at the onsets of assembly of Nuna and

Rodinia, respectively. Overall, significant zircon and LIP age peaks occur near the onsets of supercontinent assembly at 1850, 1100-1050, 600 and perhaps at 100 Ma, if a new supercontinent began to assemble at this time (Figs. 5 and 8).

5. Episodic Mantle Events

Some investigators (Isley and Abbott, 2002; Prokoph et al., 2004; Puetz and Condie, 2019) have proposed that the intensity of global LIP activity and global magmatism (U-Pb zircon age frequency) fluctuate with multiple periodicities ranging from 15 to 820 Myr. These cycles are sometimes interpreted as episodes of heating and cooling within the mantle (Isley and Abbott, 2002; Condie and Puetz, 2019). In this study we refer to them as mantle cycles. From 2300 Ma onwards, spectral analysis of LIPs and zircon age distributions generally show 24 repetitions of a of 90-to-93-myr cycle. There is no obvious relationship between this cycle and the breakup and assembly supercontinents, and there is no obvious secular relationship between the number of LIP sites and the number of cratons/orogens on which they are found (Fig. 8). If LIP age peaks result chiefly from thermal insulation effects of the continental lithosphere (Lenardic et al., 2011; Brandl et al., 2013), they should be more frequent in the cores of supercontinents, which however, is not observed (Supplementary Data, Fig. S1). Instead, the 90-Myr cycle appears to be driven by bottom-up mantle events (Condie et al., 2015b; Condie and Puetz, 2019).

Periodicities in zircon and LIP ages are recognized at about 90, 105, 140, 185, 270, and 400 Myr (Isley and Abbott, 2002; Prokoph et al., 2004; Puetz and Condie, 2019), and a 400 Myr cycle is now detected in the age distributions of U-Pb detrital zircons and two-stage Hf model ages for εHf values near the depleted mantle growth line (Puetz and Condie, 2019). This further suggests the 400-Myr cyclicity is linked to episodic mantle events. In addition, a cycle at 800 Myr may be associated with the supercontinent cycle (Isley and Abbott, 2002; Prokoph et al., 2004; Chen and Cheng, 2018; Puetz et al., 2018; Condie and Puetz, 2019). Two age peaks in the 800-Myr cycle at 250 Ma and 2700 Ma (Condie and Puetz, 2019) correspond to, respectively, the end of assembly of Pangea, and the assembly of one or more supercratons in the Neoarchean (Evans and Pisarevsky, 2008).

The 400-Myr cycle is linked to maxima in orogen frequency and plate speeds (Figs. 5 and 8), approximately corresponding to the onsets of supercontinent assembly for Nuna, Rodinia and

Gondwana-Pangea. Peaks in LIP activity associated with the 400-Myr cycle occur at the onset stage of supercontinent assembly for Nuna and Rodinia, but not for Gondwana-Pangea. Also, there is a slight tendency for both the number of sites and the number of cratons/orogens on which LIPs are found to be higher during the breakup of Neoarchean supercratons (2.2-2.0 Ga) than for breakup of later supercontinents. As with the 90-Myr mantle cycle, the 400-Myr cycle appears to be driven by bottom-up processes (Condie et al., 2015b). Initiation of the supercontinent cycle at about 2 Ga, may have been influenced by an already existing 400-Myr mantle cycle.

The relative importance and accuracy of each cycle mentioned here will require further research to define the periodicities and their significance more rigorously.

6. Angular Plate Speeds

Hoffman (1997) suggested many years ago that the recurrence interval of supercontinents has become shorter with time and Condie et al. (2015a) suggested from paleomagnetic data that average angular plate speed has been increasing with time for the last 2 Gyr. One of the main problems with an increasing rate of the supercontinent cycle is that it would seem to require an increase in the rate of plate tectonics, which is counterintuitive for an Earth that has been cooling for 4.5 Gyr. As paleomagnetic data are the quantitative basis for continental reconstructions, we calculate angular plate velocity to analyze the motion of continental cratons (Supplementary Data, Table S6). Our plate speed estimates are based on a divergent set of approaches and reconstructions (Ernst et al, 2013; Pisarevsky et al, 2013, 2014a,b, 2015; Hoffman, 2014; Cederberg et al. 2016; Matthews et al, 2016; Merdith et al, 2017; Lubnina et al, 2017). Despite the diversity, all of these generally assume minimal movement of continents. For this reason, our estimates serve as minimum constraints for plate speeds. This provides one of many preliminary steps toward the ultimate goal of attaining reliable full plate reconstructions.

Using our new timing for the supercontinent cycle (Table 1) based on new and more precise paleogeographic reconstructions (Li et al., 2008; Pisarevsky et al., 2014; Meert, 2014; Keppie, 2016; Matthews et al., 2016; Merdith et al. 2017; Lubnina et al., 2017), average plate speed appears to have decreased rather than increased with time (Fig. 9). Although the background plate speed may have remained nearly constant at about 35 deg/100 Myr, the average plate speed decreased from about 50 to 40 deg/100 Myr between 1900 and 100 Ma as shown by the linear regression. Although the results show plate speed decreasing with time, the small r value (0.31 with peaks, 0.15 without peaks) and uncertainties in supercontinent reconstruction also allow approximately constant plate speeds with time. In addition, we see a remarkable increase in angular plate speed near the onset of supercontinent assemblies at 1850, 1050 and 650 Ma. As the three supercontinents (Nuna, Rodinia, Pangea) continued to assemble, plate speed rapidly dropped as moving cratons collided. There are also small peaks in plate speed near the onset of breakup of Nuna (1450 Ma) and near the onset of assembly of Pangea (450 Ma).

7. Orogens

Our results suggest that the average frequency of both accretionary and collisional orogens increases from the end of the Archean, but at a rate less than proposed by Condie et al. (2015a) (Figs. 2 and 3). In addition, there are large peaks in frequency at 2000-1900 Ma and about 600 Ma, roughly corresponding to zircon and LIP age peaks and to the onsets of supercontinent assembly (Fig. 5). Peaks at 500 and 100 Ma in collisional orogens and at 400 Ma in accretionary orogens correspond to the onset of growth of Pangea and the possible onset of assembly of a new supercontinent (~100 Ma) (Australia colliding with Asia). The heights of the age frequency peaks are not as important as their ages, because height is, in part, related to number of orogen segments counted as discussed by Condie et al. (2015a). Because orogen size may also be important, we consider the possible effects of preserved orogen lengths on the frequency of orogens with time (Supplementary Data, Fig. S2 and Supplementary Data, Table S7). The length-normalized results show a similar secular curve to the number of orogen curves (Figs. 2 and 3) and exhibit the same two spikes in orogen frequency at about 1900 and 600 Ma. The major difference in frequency between accretionary and collisional orogens is shown during the breakup of Rodinia (750-600 Ma) and the Neoarchean supercratons (2200-2000 Ma): accretionary orogens show peaks in frequency at 700 Ma and 2200 Ma, whereas collisional orogens show troughs at these times (Figs. 2 and 3). Orogens do not show a preference for assembly or breakup stages of supercontinents, and the minima in collision frequency do not always correspond to supercontinent breakup. For instance, the minima at 800-700 Ma and 1550-1450 Ma (Fig. 3) correspond to periods of supercontinent stability or the beginning of breakup of Rodinia and Nuna, respectively. The bottom line is that although collisional orogens are more frequent during supercontinent assemblies, accretionary orogens have no preference for either breakup or assembly phases of supercontinents.

Accretionary orogen durations are mostly 50-200 Myr (Md = 100 Myr), while the collisional phase is mostly 20-100 Myr (Md = 55 Myr) (Supplementary Data, Figs. S3 and S4; Supplementary Data, Table S2). In contrast, long-lived linked orogens have durations of ≥ 275 Myr, the longest of which is the Great Proterozoic Accretionary orogen (GPAO) with a duration of almost 1 Gyr (Table 2); the GPAO accompanies the assembly of Nuna and some components persist into the assembly of Rodinia. In general, long-lived orogens develop on cratons that did not significantly fragment during supercontinent breakup, such as Amazonia, Baltica, and Laurentia. Accretionary orogens, or segments thereof, may end in one of four ways: 1) collision between cratons, 2) large terrane collisions in which the subduction zone may step oceanward, 3) rifting in the backarc, which leads to a new passive margin and a dispersing arc, and 4) subduction of an ocean ridge which leads to either a transform fault system or rarely (if collision has no transcurrent component) to loss of a convergent plate margin. Of the 194 orogens in our database (Table S2), 87% have collisional terminations, 7% have unknown terminations, and 6% are ongoing today. Of the 87% collisional orogens, 45% are linked as “orogen segments” in long-lived orogens and 42% are non-linked orogens with terminal collisions (Supplementary Data, Tables 2 and S2).

8. Passive Margins

Bradley (2008) suggested that the lifetimes of passive continental margins decrease with time, consistent with an increasing speed of the supercontinent cycle. However, he did not recognize some passive margins in the time interval of 1900-1000 Ma, and more recent supercontinent reconstructions require 12 or more passive margins that came into existence during this time (Condie, 2020). On a graph of passive margin onset age versus duration (Fig. 10), there seems to be three groups of passive margins, each group showing a decrease in passive margin duration with time. Group 1 is possibly associated with the breakup of Neoarchean supercratons (2.2-2.0 Ga) and Group 2 may be associated with the breakup of Nuna (1500-1200 Ma). However, Group 3 does not appear to be associated with any particular supercontinent, although it is dominantly associated with the breakup of Rodinia and assembly and breakup of Gondwana-Pangea. The longest lived passive margins (durations of 400-600 Myr, Fig. 10) are part of Group 3 and occur along the northern and eastern margins of Baltica and along the margins of Siberia. It is important to note that passive margin onsets in each group occur during both supercontinent breakup and assembly phases, attesting to the overlap of these phases. The median duration of passive margins in a 100-Myr moving window remains relatively constant at 150-200 Myr (Supplementary Data, Table S8).

9. Discussion

9.1 LIPS

LIP activity followed by continental separation and ocean basin opening occurs chiefly during the breakup phases of supercontinents. Most high-density LIP activity (90%) is associated with a LIP age peak, but fragmentation of plates does not always follow major LIP activity. For instance, LIP activity associated with the large 1400-Ma and 1450-Ma age peaks, during Nuna breakup, is not always followed by continental separations (Table 3). Of the 32 LIP age peaks represented along craton boundaries, only 12 (38%) are followed by craton separation, and most of those not followed by continental separation occurred during assembly or transition phases of supercontinents. Except for the 130-Myr separation time of Australia from Siberia at 1750 Ma, the time between LIP age peaks and continental separation is relatively constant in the range of 50-75 Myr. These results suggest that major LIP activity occurs during both assembly and breakup phases of supercontinents, but rarely does it result in continental separation during assembly or transition phases. Also, there is no evidence that LIPs can be used to predict the timing and location of supercontinent breakup as originally suggested by Ernst and Bleeker (2010).

There are many regions that show multiple pulses of LIP activity, each pulse lasting for 50-100 Myr and separated by ≥ 100 Myr. For instance, East Laurentia-Greenland had eight periods of episodic activity from 1600 to 50 Ma. Having six pulses of LIP activity are centers in Scandinavia (1785 to 940 Ma), Siberia (1800 to 200 Ma), North Australia (1800 to 400 Ma), South Africa (1400 to 200 Ma), and India (1800 to 50 Ma) (Table 3). There are many cratons with episodic LIP activity over long periods of time. Examples are the Slave craton in Canada (10 pulses ranging in age from 2037 to 723 Ma), the Amazon craton (13 pulses between 1890 and 80 Ma), the Siberian craton (14 pulses between 1780 and 252 Ma), and the North China craton (11 pulses between 1800 and 27 Ma). In Scandinavia there are two regions with a high density of Proterozoic LIPs, one in southern Finland and another in southern Sweden, with activity mostly at 1650-1450 Ma. Could they both come from the same group of mantle plumes? If so, there are two problems: there is no plume track between the two centers, and there is some activity in both centers at the same time. An alternative explanation is that each center results from a different plume or group of plumes. Since it is unlikely that even large plumes can survive for more than 200 Myr (Arnould et al., 2020), it is unlikely that long-lived LIP activity reflects single or multiple mantle plumes of the same age. The common association of some LIPs with subduction-related greenstones is consistent with the possibility that some LIPs are subduction-related rather than plume-related. Scandinavia, for instance, was part of an accretionary orogen from 1800 to about 1000 Ma. Perhaps each pulse of LIP activity was related to an episode of subduction in this region (Wang et al., 2014). And of course it is possible that some of the LIPs are plume-related while others are subduction-related.

9.2 Mantle Cycles

We still do not understand what produces mantle cycles such as the 90-Myr and 400-Myr cycles (Fig. 5). It is possible that the 400-Myr cycle is a harmonic of the 90-Myr cycle. If the LIP geographic distribution of these two cycles is representative, neither cycle appears to be global in extent. However, lack of preservation of oceanic LIPs and sample biases may be responsible for the limited geographic distribution of LIPs that track these cycles. In contrast, detrital zircons track age peaks in large geographic areas, and suggest that both cycles may be widespread. Another question that we really do not have a satisfactory answer to yet is what controls the amplitude of LIP age peaks, At least three possibilities need to be considered: 1) intensity of a mantle plume event, 2) multiple cyclicity reinforcements, and 3) an increased number of studies in particular geographic areas (for whatever reason). With the possible exception of the 1880 and 1100-Ma age peaks, peak amplitude (Figs. 5 and 8) does not appear to be related to either the number LIP sites nor the number of continents/orogens on which an age peak is represented (Supplementary Data, Table S5), and thus possibility 3) seems least likely of the three causes. Probably some combination of variable plume intensity and multiple plume cyclicities offers the best explanation for differences and irregularities in amplitudes of LIP age peaks (Puetz and Condie, 2020).

Both numerical and experimental models related to mantle plume generation (Davaille, 2005; Li et al., 2018) are consistent with mantle events with a cyclicity of 100-200 Myr (Condie et al., 2015b). Furthermore, both cycle and non-cycle plumes are generated in numerical models (Li et al., 2018). Possible events responsible for zircon and LIP age peaks include mantle overturn, thermochemical destabilization in the deep mantle producing plumes, and mantle avalanches when slabs suddenly sink through the 660-km discontinuity (Davies, 1995; Condie, 1998; Machetel and Humler, 2003; Davaille et al., 2005). For all three possibilities, we must address the question of how mantle cycle peaks are transferred to subduction-related magmas, which are the chief sources of zircon. It is probable that increases in subduction rate or/and the number of subduction zones is required by any model. Perhaps some plumes move laterally upon hitting the base of the lithosphere (Bagley and Nyblade, 2013) increasing rates of subduction, thus transferring the signals of mantle cycles into arc magmatism (Arndt and Davaille, 2013). Another possibility, as suggested by geodynamic models, is that plumes may contribute to the initiation of subduction by focused magmatic activity weakening and thinning the lithosphere, thus increasing the total number of subduction zones (Gerya et al., 2015).

9.3 Plate Speeds and Orogens

Based on numerical modelling, Korenaga (2006) suggested that plate speed is increasing with time because of thicker plates in the past that result in less efficient heat transport, and thus lower average plate velocities. Yet our new results suggest that average plate speed remains unchanged or has been decreasing with time. The results of Agrusta et al. (2018) are also consistent with decreasing plate speeds with time if trench migration has increased with time as plates have strengthened. We consider two possibilities to account for the difference in our results and the model of Korenaga (2006) in the last 2 Gyr: 1) progressively decreasing distances between fragmenting plates, and 2) an increasing proportion of continental crust on plates with time. As a test of the first possibility, the median distance between continents or cratons at maximum dispersion of each supercontinent is given in Figure 11 (data in Supplementary Data, Table S9). The results show that craton distance has not decreased with time, but remained relatively constant with a median distance of ~ 9000 km. During supercontinent breakup some continents do not fragment or at least do not fully separate from each other, and some continents remain close to each other during breakup and assembly of later supercontinents (Meert, 2014). For instance, Laurentia and Baltica, although slightly rotated, have similar configurations in Nuna, Rodinia and Pangea. Yet the dispersion of cratons (as measured by 1s of the mean) was much greater at the onset of Nuna assembly than at the onsets of assembly of the other two supercontinents (Fig. 11). During the assembly of the building blocks of Nuna (1900-1800 Ma), the number of cratons rapidly decreased as they collided with each other, and after 1200 Ma, the number of cratons remained relatively constant at 13-15 (Supplementary Data, Fig. S5). Thus, decreasing distances between cratons could contribute to the "apparent" decrease in plate speed during the growth of Nuna (Fig. 9), but not to later supercontinents where the median and dispersion of cratons distances are similar.

Another factor that may affect plate speed in the past is the proportion of plate area comprising continental crust. Zahirovic et al. (2015) show an inverse relationship between proportion of continental crust on young plates and average plate speed. Although most of the continental crust formed by the end of the Archean, as much as 30% of the present volume of continental crust may have been added after 2 Ga (Dhuime et al., 2017), and much of this in the Great Proterozoic Accretionary Orogen (Condie et al., 2015a). Thus, an increasing volume of continental crust after 2 Ga may have contributed to relatively steady or decreasing plate speeds after this time, thus overpowering the effect of decreasing plate thickness (Korenaga, 2006) on increasing plate speed.

As expected, the frequency of collisional orogens increases during supercontinent assembly, with an especially large peak at 1850 Ma approximately coinciding with large peaks in LIPs and detrital zircons at the onset of assembly of Nuna (Figs. 2, 3, and 5). Why there are so few collisional orogens (only 15) related to the assembly of Rodinia remains a problem (1100-800 Ma). Although the number could be increased to 20 if predicted orogens (Fig. 1a) are also included (Congo-W Africa, India-S China, North China-Siberia, and Siberia-Laurentia), Rodinia assembly still has many fewer orogens than the assemblies of Nuna (45) or Gondwana-Pangea (70). Possibly the small number of orogens during Rodinia assembly is related to an extroversion origin for this supercontinent (Liu et al., 2017). The core of Nuna largely remained intact during Rodinia assembly (Evans and Mitchell, 2011; Roberts, 2013), thus far fewer orogens were necessary to assemble Rodinia. Also, some continental fragments may not have accreted to this core (such as the African blocks, South China, and India; Fig. 6).

9.4 Is there and Underlying Cause?

The correlations in time-series for plate speeds, LIPs, detrital zircons and orogen frequency (Fig. 5) seems to require a common global process to manifest itself via these four possibly related geological processes. All four time-series exhibit approximately coincident peaks with the onset of supercontinent assembly (1850, 1050, 650-600 and 100 Ma) and may result from episodic deep mantle events such as mantle overturn, plume, or avalanche events (Davies, 1995; Condie, 2000; Machetel and Humler, 2003; Davaille et al., 2005). Slab avalanches at the 660-km discontinuity may result in intense plate convergence during supercontinent assembly (Faccenna et al., 2013). Li (2020) shows that slabs being subducted deep into the mantle may produce localized thermal anomalies just above the core-mantle boundary. In some cases, these may produce mantle plumes, and thus possibly produce widespread mantle plume events. East et al. (2020) show that new ocean ridges produced during supercontinent breakup contribute to increased subduction rates. And finally, zircon age peaks may be linked to rates of convergent margin magmatism as shown by correlations with subduction flux at 250 and 100 Ma (Hounslow et al., 2018).

A very intriguing correlation has emerged from our study: since 2300 Ma, five major age peaks are observed with ~400-Myr periodicity (Fig. 5). Just what drives this cycle is not yet clear. It may have begun with a "big bang" at the 2700-Ma global event, although the LIP age spectrum suggests the cycle was already operational by at least 3850 Ma (Supplementary Data, Fig. S6). In either case, the supercontinent cycle may have adapted to an existing 400-Myr cycle beginning with the breakup of Archean supercratons at 2200-2000 Ma, which coincides with the widespread propagation of plate tectonics (Condie, 2020). However, the relationship between age peaks in LIPs, detrital zircons, and orogens and the supercontinent cycle remains a subject of debate and uncertainty. First of all, the supercontinent cycle should be considered as quasi-periodic rather than perfectly periodic because the interval between assemblies decreases with time (from 800 to 500 Myr, Fig. 5). If the supercontinent "cycle" adapted at least partly to an already existing 400-Myr cycle, it soon became detached from this cycle, especially after 1 Ga. This agrees with the numerical models of Rolf et al. (2014), which suggest that any regularity in the timing of the supercontinent cycle is prevented by the chaotic nature of mantle convection.

Once established, supercontinent breakup may lead to increased subduction rates and accumulation of slabs in the mantle transition zone, later followed by slab avalanches. As an example, the breakup of Archean supercratons at 2200-2000 Ma may have triggered slab avalanches. Then, 100-200 Myr later, this initiated a widespread mantle plume (LIP) event, which in turn, may be responsible for the large 1850-Ma age peak in zircons, frequency of orogens, and average plate speed. We propose a connection between slab avalanches and zircon ages as follows: avalanche (plus 100-200 Myr) → plume (LIP) event → increasing subduction rate and plate speed → more orogens → more granites → more zircons. The same sequence of events may have been repeated three more times with avalanches at 1300, 800 and 300 Ma, giving rise, respectively, to mantle events at 1100, 650-600 and 100 Ma. A weak to moderate age peak at ~1450 Ma in all four time-series coincides with the onset of breakup of Nuna, but the major LIP age peak at 800 Ma is not found in the other three time-series. This apparent inconsistency is not easily explained unless it indicates a developing slab avalanche at 800 Ma.

10. Conclusions

1) Of nine large age peaks in either or both zircon and LIP time series <2300 Ma at 2150, 1850, 1450, 1400, 1050, 800, 600, 250 and 100 Ma, only four are geographically widespread (1850, 1400, 800 and 250 Ma). These age peaks occur both before and after the onset of the supercontinent cycle, and during both assembly and breakup phases of supercontinents.

2) Significant age peaks in orogen frequency, average plate speed, LIPs and detrital zircons may be part of a 400-Myr mantle cycle, and major age peaks in the cycle occur near the onset of supercontinent assemblies. The 400-Myr cycle may have begun with a "big bang" at the 2700 Ma event, although the LIP age spectrum suggests the cycle may go back to at least 3850 Ma.

3) A prominent 90-Myr mantle cycle is recorded by continental LIPs, which are geographically more widespread than LIPs that are not part of this cycle. The 90-Myr cycle shows no preference for breakup or assembly stages of supercontinents.

4) LIP age peaks occur during both assembly and breakup phases of supercontinents, but rarely if ever does LIP activity result in plate separation during assembly or transitional phases. There is no consistent pattern between the number of LIP sites, the number of continents/orogens on

which LIPs occur, and LIP age peaks. During supercontinent breakup, LIP activity is followed by ocean-basin opening in some areas, but not in other areas. This suggests that mantle plumes are not necessary for ocean-basin opening, and that LIPs should not be used to predict the timing and location of supercontinent breakups.

5) LIPs recording mantle cycles may be produced directly by mantle plumes or indirectly from subduction regimes that have inherited mantle-cycle signatures from plume activity.   Some combination of variable plume event intensity and peak enhancement by multiple cyclicities offers the best explanation for differences in LIP age peak irregularities and amplitudes.

6) New paleogeographic reconstructions of supercontinents indicate that in the last 2 Gyr average angular plate speeds have not changed or have decreased with time, whereas the number or orogens has increased.

7) Decreasing distances between cratons with time could contribute to an “apparent” decrease in plate speed during the growth of Nuna, but would not affect later supercontinents where the median and dispersion of craton distances are similar. A possible explanation for decreasing or steady plate speed is an increasing proportion of continental crust on plates as juvenile continental crust continued to be added in post-Archean accretionary orogens.

8) Peaks in orogen frequency at 1850, 1050, 600 Ma, which approximately coincide with major zircon and LIP age peaks, correspond to onsets of supercontinent assembly, and age peaks at 1450, 250 and 100 Ma correspond to supercontinent stasis or breakup. Although collisional orogens are more frequent during supercontinent assemblies, accretionary orogens have no preference for either breakup or assembly phases of supercontinents. Most accretionary orogens terminate with continental collisions.

9) During assemblies of Nuna and Gondwana-Pangea, there is a high concentration of both accretionary and collisional orogens. The sparsity of orogens during Rodinia assembly may be related to incomplete breakup of Nuna as well as to the fact that some cratons never accreted to Rodinia.

10) There are three groups of passive margins, each group showing a decrease in passive margin duration with time: Group 1 with onsets at 2.2-2.0 Ga corresponding with the breakup of Neoarchean supercratons; Group 2 with onsets at 1.5-1.2 Ga corresponding to the breakup of Nuna; and Group 3 with onsets at 1.5-0.1 Ga not corresponding to any particular supercontinent breakup.

11) Large age peaks at 1850, 1050, 600 and 250 Ma may be related to slab avalanches from the mantle transition zone that occur in response to supercontinent breakups.

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Figure Captions

1a. Paleogeographic continental reconstruction at 1900 Ma. Modified after Pisarevsky et al. (2014a) and Lubnina et al. (2017). Shown are collisional and accretionary orogens and large igneous provinces (LIPs) in the age range of 2000-1800 Ma.
Cratons: WYO, Wyoming; SUP, Superior; NAT, North Atlantic; HRN, Hearne; SLV, Slave; YLG, Yilgarn; PBR, Pilbara; TUN,Tungus; ALD, Aldan; DAL, Daldyn; MAG, Magan; DRW, Dharwar; KKR, Karelia; KPV, Kaapvaal; ZBW, Zimbabwe; KPV+ZBW = Kalahari; AMZ, Amazonia; REG, Reguibat; HOG, Hoggar-Tuareg; VUR, Volga-Uralia; SAR, Samartia; SNB, Singhbhum; ARV-BDH, Aravalli-Bundelkhand; BST, Bastar; CGO, Congo; TAN, Tanzania; MAD, Madagascar; RPL, Rio de la Plata; RNR, Raynar; YTZ, Yangtze; CAT, Cathaysia; TAR, Tarim; SFR, São Francisco; MAW, Mawson; NAC, North Australia; SAC, South Australia; NCH, North China.

b. Paleogeographic continental reconstruction at 1500 Ma, corresponding the maximum packing of Nuna. Modified after Pisarevsky et al. (2014a) and Lubnina et al. (2017). Shown are collisional and accretionary orogens and large igneous provinces (LIPs) in the age range of 1600-1400 Ma. Other information in Figure 1a and Table 1.

2. Frequency of accretionary orogens expressed as number of orogen segments per 50-Myr bin moving in 50 Myr increments. Red Line, linear regression analysis: $n = 4.27 - 0.000321a$, $r = 0.65$ (n, number of orogens; a, age in Ma). Supercontinent assembly, yellow; breakup, blue. Data from Table S2.

3. Frequency of collisional orogens expressed as number of orogen segments per 50-Myr bin moving in 50 Myr increments. Red Line, linear regression analysis: $n = 2.75 - 0.00048a$, $r = 0.13$ (n, number of orogens; a, age in Ma). Supercontinent assembly, yellow; breakup, blue. Data from Table S2.

4. Histogram showing the frequency of LIPs (large igneous provinces) with time in 50-Myr bins. Supercontinent assembly, yellow; breakup, blue. Data from Table S1.

5. Detrended time-series using 20-Myr bin sizes, with one line also smoothed with a 7-weight Gaussian kernel. a) U/Pb detrital zircon ages ($n = 443{,}259$), records accepted with absolute discordance <50 Myr and 2σ uncertainty <70 Myr, from Puetz and Condie (2019); b) continental large igneous province (LIP) ages ($n = 915$) from Table S1; c) number of orogens (from Table S2); c) Mean angular plate speeds (Table S5) at 100-Myr intervals. Vertical dashed lines show a 400-Myr mantle cycle. Supercontinent assembly, yellow; breakup, blue.

6a. Paleogeographic continental reconstruction at 1200 Ma, corresponding to the maximum dispersion of Nuna. Modified after Pisarevsky et al. (2014a) and Lubnina et al. (2017). Other information in Figure 1a.

b. Paleogeographic continental reconstruction at 800 Ma, close to maximum packing of Rodinia (Table 1). Modified after Pisarevsky et al. (2014a), Lubnina et al. (2017) and Merdith et al. (2017). Shown are collisional and accretionary orogens and large igneous provinces (LIPs) in the age range of 900-700 Ma. Other information in Figure 1a.

7a. Paleogeographic continental reconstruction at 600 Ma, approximately corresponding to maximum dispersion of Rodinia and onset of assembly of Gondwana. Modified after Pisarevsky et al. (2014a), Lubnina et al. (2017) and Merdith et al. (2017. Shown are collisional and accretionary orogens and large igneous provinces (LIPs) in the age range of 700-500 Ma. Other information in Figure 1a.

b. Paleogeographic continental reconstruction at 600 Ma, approximately corresponding to maximum packing of Pangea. Modified after Merdith et al. (2017) and Matthews et al. (2016). Shown are collisional and accretionary orogens and large igneous provinces (LIPs) in the age range of 350-150 Ma. Some of these do not show on the map because of distortions on the Robinson projection in the polar regions. Other information in Figure 1a.

8. Large igneous province (LIP) age peak frequency versus peak age expressed as the number LIP sites per peak and the number of cratons/orogens on which each peak is found. Supercontinent assembly, yellow; breakup, blue. Data from Table S1. Vertical lines represent a 400-Myr cycle. Peaks are defined with a 20-Myr bin size and assumes a 20-Myr uncertainty on peak location.

9. Average area-weighted plate speed (deg/100 Myr) as a function of age. Red line, linear regression analysis: s = 38.1 + 0.0082a, r = 0.31 with peaks and s = 34.8 + 0.0018a, r = 0.15 without peaks (s, plate speed; a, age in Ma). Each point is the average plate speed in a 100-Myr moving window (data in Table S5). Supercontinent assembly, yellow; breakup, blue.

10. Passive continental margin duration as a function of onset age. Data from Bradley (2008) and Table S7. Supercontinent assembly, yellow; breakup, blue.

11. Median distance (in km) between cratons as a function of supercontinent maximum dispersion age (Table 1). Vertical lines, one standard deviation of the mean. Reconstruction references: Li et al., 2008; Pisarevsky et al., 2014a; Meert, 2014; Keppie, 2016; Matthews et al., 2016; Merdith et al. 2017; Lubnina et al., 2017. Supercontinent assembly, yellow; breakup, blue.

Supplementary Data

Supplementary Figures

S1. Paleogeographic Maps

S2. Craton Distances

S3. Duration accretionary orogens

S4. Duration collisional orogens

S5. Number of cratons

S6. Detrended time series

Supplementary Tables

S1. LIP summary

S2. Orogen summary

S3. Euler rotations

S4. Paleomagnetic poles

S5. LIP geographic distributions

S6. Plate speeds

S7. Summary of orogen lengths

S8. Passive margins

Kent Condie is emeritus professor of geochemistry at New Mexico Institute of Mining and Technology, Socorro, NM where he has taught since 1970. His textbook, Plate Tectonics and Crustal Evolution, was first published in 1976 and has gone through four editions. In addition Condie has written seven other professional books the most recent of which, Earth as an Evolving Planetary System is now in the fourth edition. He is author or co-author of over 750 articles published scientific journals. He was awarded NMT's Distinguished Research Award in 1987. In addition, he was elected the Vice President of the International Association for Gondwana Research in 2002 and in 2007 was bestowed an Honorary Doctorate Degree from the University of Pretoria in South Africa. He was awarded the Penrose Medal of the Geological Society of America in 2018. Condie is a member of the American Geophysical Union, the Geological Society of America and the Geochemistry Society.

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