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Is the rate of supercontinent assembly changing with time?

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ABSTRACT

To address the question of secular changes in the speed of the supercontinent cycle, we use two major databases for the last 2.5 Gyr: the timing and locations of collisional and accretionary orogens, and average plate velocities as deduced from paleomagnetic and paleogeographic data. Peaks in craton collision occur at 1850 and 600 Ma with smaller peaks at 1100 and 350 Ma. Distinct minima occur at 1700–1200, 900–700, and 300–200 Ma. There is no simple relationship in craton collision frequency or average plate velocity between supercontinent assemblies and breakups. Assembly of Nuna at 1700–1500 Ma correlates with very low collision rates, whereas assemblies of Rodinia and Gondwana at 1000–850 and 650–350 Ma, respectively correspond to moderate to high rates. Very low collision rates occur at times of supercontinent breakup at 2200–2100, 1300–1100, 800–650, and 150–0 Ma. A peak in plate velocity at 450–350 Ma correlates with early stages of growth of Pangea and another at 1100 Ma with initial stages of Rodinia assembly following breakup of Nuna. A major drop in craton numbers after 1850 Ma corresponds with the collision and suturing of numerous Archean blocks.

Orogens and passive margins show the same two cycles of ocean basin closing: an early cycle from Neoarchean to 1900 Ma and a later cycle, which corresponds to the supercontinent cycle, from 1900 Ma to the present. The cause of these cycles is not understood, but may be related to increasing plate speeds during supercontinent assembly and whether or not long-lived accretionary orogens accompany supercontinent assembly. LIP (large igneous province) age peaks at 2200, 2100, 1380 (and 1450?), 800, 300, 200 and 100 Ma correlate with supercontinent breakup and minima at 2600, 1700–1500, 1100–900, and 600–400 Ma with supercontinent assembly. Other major LIP age peaks do not correlate with the supercontinent cycle. A thermochemical instability model for mantle plume generation can explain all major LIP events by one process and implies that LIP events that correspond to the supercontinent cycle are independent of this cycle.

The period of the supercontinent cycle is highly variable, ranging from 500 to 1000 Myr if the late Archean supercratons are included. Nuna has a duration of about 300 Myr (1500–1200 Ma), Rodinia 100 Myr (850–750 Ma), and Gondwana–Pangea 200 Myr (350–150 Ma). Breakup durations are short, generally 100–200 Myr. The history of angular plate velocities, craton collision frequency, passive margin histories, and periodicity of the supercontinent cycle all suggest a gradual speed up of plate tectonics with time.

1. Introduction

The supercontinent cycle as proposed by Worsley et al. (1984, 1985) and now widely accepted is important in understanding the tectonic history of continents, and it also provides a powerful constraint on the climatic and biologic evolution of Earth. The overall characteristics and history of the development of the supercontinent cycle is reviewed in detail by Nance and Murphy (2013) and Nance et al. (2014) and will not be repeated. Lowman and Jarvis (1995) and Gurnis (1988) suggested that continental blocks tend to be drawn to mantle downwellings where they may collide to form supercontinents. Because of their thickness and enrichment in U, Th and K, supercontinents should act as thermal insulators to mantle heat (Anderson, 1982; Gurnis, 1988), but

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consideration based on the thermal budget of the Earth indicates that the actual effect of insulation is likely to be weak, probably no more than an increase of 20 °C in mantle temperature (Korenaga, 2007). Perhaps more important is the global organization of mantle flow pattern caused by the presence of a supercontinent (e.g., Storey, 1995). Recent numerical simulation studies suggest that after a supercontinent is assembled over a downwelling in one hemisphere, circum-supercontinent subduction induces a new major upwelling beneath the supercontinent transforming a degree-1 planform in the mantle into a degree-2 planform with two antipodal downwellings (Zhong et al., 2007; Zhang et al., 2009).

We still have many questions regarding the supercontinent cycle, such as when it began, has continental crust grown in volume with time, and has the period of the cycle been constant or has it changed with time. The timescale of assembly and dispersal of supercontinents is still not well constrained, with estimates of cycle length ranging from 250 Myr to 1000 Myr (Phillips and Bunge, 2007; Zhang et al., 2009; Yoshida and Santosh, 2011). Some investigators have suggested that the supercontinent cycle has speeded up with time (Hoffman, 1997; Condie, 2002), but testing such an idea is not easy because it involves how a supercontinent is defined, and whether or not large blocks of one supercontinent survive during breakup to become incorporated in later supercontinents. The secular change in the supercontinent cycle is, however, an important problem in the evolution of plate tectonics. It is commonly assumed that a hotter mantle in the past resulted in faster plate motions (e.g., Schubert et al., 1980; Davies, 2009), which could be reflected in the formation history of supercontinents.

In this study, we address the question of whether the supercontinent cycle is speeding up, slowing down, or remaining constant with time. Our primary datasets are angular plate speeds as deduced from published paleogeographic reconstructions, paleomagnetic studies, and the frequency of collisional and accretionary orogeny as estimated from extensive geologic and geochronologic data. We also address the question of ocean basin closing and how it may have changed with time, and compare results to the cycles and durations of sedimentation in passive margins. From these data, we discuss the lifetimes of supercontinents and possible relationship to mantle plume activity as deduced from LIP (large igneous province) events through time. We conclude that average plate speed and the collisional frequency of cratons are probably increasing with time, and that the supercontinent cycle, which began about 1750 Ma, is also speeding up with time.

2. Methods

2.1. Characteristics of orogens

In previous papers we have discussed the compilation of orogen characteristics and uncertainties, and this will not be repeated (Condie, 2013, 2014). One of the main sources of uncertainty in counting orogens is that of what to count as a single orogen. Collisional orogens of short strike length could be part of a longer orogen, now displaced by later supercontinent breakup. Hence, most of the orogens listed in Appendix 1 are really "orogen segments". 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 (Condie, 2013) (Fig. 1, number 35), some portions of the orogen that have been well studied are designated as segments.

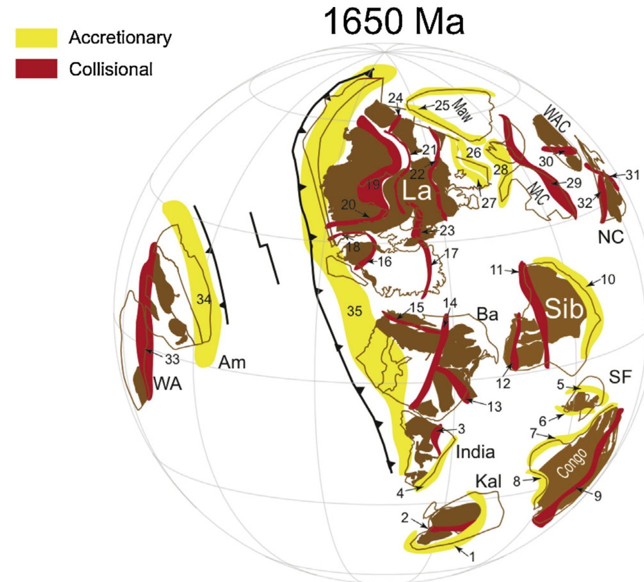

Fig. 1. Paleogeographic reconstruction of cratons during the late stages of Nuna assembly at 1650 Ma (from Pisarevsky et al., 2014a). Orogens from Appendix 1 (Pts 1 and 2). *Key*: cratons: Kal, Kalahari; SF, São Francisco; Sib, Siberia; La, Laurentia; Maw, Mawson; NAC, North Australian craton; WAC, West Australian craton; NC, North China; Am, Amazonia; WA, West Africa. orogens: 1, Magondi-Kheis (2.04–1.96 Ga); 2, Limpopo (2.06–1.97 Ga); 3, Aravalli (1.87–1.85 Ga); 4, Lesser Himalaya (1.88–1.78 Ga); 5, Borborema (2.35–2.30 Ga); 6, Mineiro (2.45–2.36 Ga); 7, West Congo (2.1–2.0 Ga); 8, Luizian (2.1–2.0 Ga); 9, Ubendian (1.88–1.85 Ga); 10, Angara (1.9–1.85 Ga); 11, Akitkan (1.9–1.87 Ga); 12, Sutam (1.9–1.85 Ga); 13, Volga-Don (2.05–2.0 Ga); 14, Volhyn-Central Russian (1.80–1.78 Ga); 15, Lapland Granulite Belt (1.92–1.87 Ga); 16, Nagssugtoqidian (1.87–1.84 Ga); 17, Inglefield (1.95–1.92 Ga); 18, Torngat (1.87–1.84 Ga); 19, Trans-Hudson (1.85–1.80 Ga); New Quebec (1.87–1.82 Ga); 21, Arrowsmith (2.4–2.3 Ga); 22, Thelon (1.96–1.91 Ga); Foxe (1.87–1.85 Ga); 24, Big Sky (1.8–1.7 Ga); 25, Nimrod-Ross (1.84–1.73 Ga); Racklan-Forward (1.64–1.60 Ga); 27, Wopmay (1.9–1.84 Ga); 28, Olarian (1.58–1.54 Ga); 29, Kimban-Yapungku (1.83–1.70 Ga); 30, Glenburgh (1.97–1.94 Ga); 31, Trans-North China (1.89–1.85 Ga); 32, Khondalite (1.95 Ga); 33, Birimian-Transamazonian (2.1–2.05 Ga); 34, Amazonia (2030–1000 Ma); 35, Great Proterozoic Accretionary Orogen (1900–1100 Ma).

In this study, a major distinction between collisional and accretionary orogens is made based on how they end: collisional orogens end with continent-continent collisions (Appendix 1, Pt 1). Accretionary orogens, on the other hand, do not always end with a continent-continent collision as did India and Tibet. Rather they 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 (Condie, 2007; Cawood et al., 2009; Moores et al., 2013). 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 the last 300 Myr, for instance, peripheral accretionary orogens have developed simultaneously with collisional orogens responsible for aggregation of Pangea (Cawood and Buchan, 2007; Cawood et al., 2009).

2.2. Plate speeds

Paleomagnetic data provide a quantitative tool for paleogeographic reconstructions. However, the number of high-quality paleomagnetic results is limited, especially for the Early-Middle Paleozoic and Precambrian (e.g. Van der Voo, 1993; McElhinny and McFadden, 2000; McElhinny et al., 2003; Pisarevsky et al., 2003, 2014a,b; Li et al., 2008). Consequently the most complete and reliable published global paleogeographic reconstructions for