Supercontinents: myths, mysteries, and milestones

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Abstract: There is an emerging consensus that Earth's landmasses amalgamate quasi-periodically into supercontinents, interpreted to be rigid super-plates essentially lacking tectonically active inner boundaries and showing little internal lithosphere–mantle interactions. The formation and disruption of supercontinents have been linked to changes in sea-level, biogeochemical cycles, global climate change, continental margin sedimentation, large igneous provinces, deep mantle circulation, outer core dynamics and Earth's magnetic field. If these hypotheses are correct, long-term mantle dynamics and much of the geological record, including the distribution of natural resources, may be largely controlled by these cycles. Despite their potential importance, however, many of these proposed links are, to date, permissive rather than proven. Sufficient data are not yet available to verify or fully understand the implications of the supercontinent cycle. Recent advances in many fields of geoscience provide clear directions for investigating the supercontinent cycle hypothesis and its corollaries but they need to be vigorously pursued if these far-reaching ideas are to be substantiated.

Alfred Wegener (1912, 1915, 1920) proposed the existence of the Late Paleozoic–Early Mesozoic supercontinent, which he called Pangaea ('all lands' in Greek, Fig. 1), and which included almost all the existing landmasses. However, it took the discovery of seafloor spreading (e.g. Vine & Matthews 1963; Vine & Wilson 1965), and the ensuing plate tectonic revolution more than 40 years later, for the existence of Pangaea to become generally accepted. This revolution raised the possibility that other large-scale continents might have existed before it (e.g. Valentine & Moores 1970; Piper 1974, 1975), a proposition that was further developed and corroborated over the next two decades (e.g. Worsley *et al.* 1982, 1984; Nance *et al.* 1986, 1988; Hoffman 1989, 1991; Dalziel 1991, 1992, 1997; Hartnady 1991; Williams *et al.* 1991; Stump 1992; Powell *et al.* 1993; Powell *et al.* 1995).

Worsley *et al.* (1982, 1984) were the first to argue that episodic peaks in continental collisions followed by episodes of rift-related mafic dyke swarms provide a record of supercontinent amalgamation and breakup. Drawing on this concept, they identified geological, biological and climatic trends that accompanied supercontinent assembly, amalgamation, breakup and dispersal (e.g. Nance *et al.* 1986). These papers subsequently led to the widespread recognition that much of Earth history has been punctuated by the episodic amalgamation and breakup of supercontinents (e.g. Murphy & Nance 1991; 2003, 2013; Zhao *et al.* 2002, 2004; Rogers & Santosh 2003, 2004; Condie 2011; Yoshida & Santosh 2011; Huston *et al.* 2012; Mitchell *et al.* 2012; Ernst *et al.* 2013).

This history of episodic supercontinent assembly and breakup, which constitutes the 'supercontinent cycle', may have influenced the rock record more profoundly than any other geological phenomenon (e.g. Condie 2011). As radiometric ages have become more numerous and increasingly precise, the apparent periodicity of supercontinent assembly and breakup was seen to coincide with many of Earth history's major milestones, such as biotic diversification and extinctions (e.g. Ernst & Youbi 2017), magnetic superchrons (e.g. Eide & Torsvik 1996), true polar wander events (e.g. Evans 2003),

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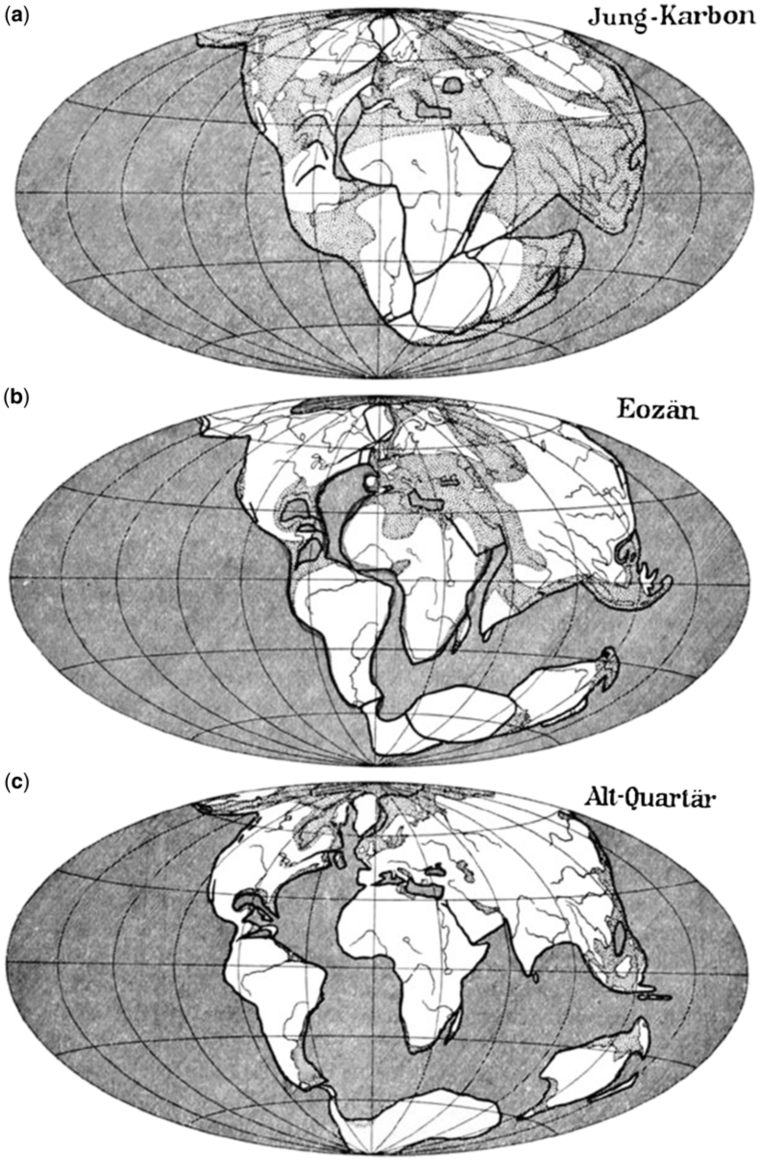

Fig. 1. Wegener's (1915) depiction of the breakup of Pangaea with respect to a fixed Africa. (**a**) His proposed Carboniferous Earth, (**b**) Eocene times, after Pangaea had already broken up, and (**c**) Earth at recent times. White, exposed continents; dotted areas, submerged continental platforms; grey, oceans.

large igneous provinces (e.g. Ernst 2014), continental margin sedimentation (Bradley 2008) and episodes of major climate change (Hoffman 1998). Many have additionally tied the formation and/or disruption of supercontinents to fundamental aspects of Earth's interior dynamics (e.g. Condie 2003; Zhong *et al.* 2007; Santosh *et al.* 2009) and to many of Earth's major geological, climatic and biological developments (e.g. Nance *et al.* 1986; Hoffman 1991; Murphy & Nance 1991; Hoffman & Schrag 2002; Lindsay & Brasier 2002; Condie 2003, 2011; Evans 2003; Dewey 2007; Zhong *et al.* 2007; Condie *et al.* 2009, 2011; Li & Zhong 2009; Santosh *et al.* 2009; Zhang *et al.* 2009; Goldfarb *et al.* 2010; Hawkesworth *et al.* 2010; Santosh 2010*a*, *b*; Bradley 2011; Hannisdal & Peters 2011; Strand 2012; Young 2013*a*, *b*; Ernst & Youbi 2017). The existence of supercontinents, like Pangaea, must be viewed in conjunction with Earth's dynamic system, the resolution of which may greatly improve our understanding of the planet.

In this paper, we re-examine the definition of a supercontinent and explore some of the major milestones, remaining mysteries and recurring myths that have accompanied the development of the supercontinent cycle in the 35 years since the idea was first proposed (Worsley *et al.* 1982). In doing so, we identify the main arguments in support of (and opposition to) such a cycle, we identify certain incidental associations and we attempt to clarify the relationship between the supercontinent cycle and other episodic tectonic processes, such as the Wilson cycle.

What is a supercontinent?

Although an apparently simple question, precisely defining a supercontinent is a challenge. There are many different opinions about the threshold in size that separates a supercontinent from just a large continent. Some authors advocate for a lenient definition of a supercontinent, which would be assemblies of all, or nearly all, of Earth's continental blocks (e.g. Hoffman 1999). This definition is generous enough to permit researchers some conceptual freedom. Others have suggested a strict threshold value of 75% of the available continental crust at any given age (Meert 2012). This second definition is simple and unequivocal. It includes Pangaea (85–90%) but would exclude other large landmasses, like Gondwana (*c.* 60%) and Eurasia (35%). However, this threshold is quite arbitrary and therefore may discriminate between landmasses that were effectively the same size. Other authors have suggested the term 'semisupercontinent' for the hierarchical level attained by Gondwana, Eurasia and similarly large and long-lived landmasses that were subsets of true supercontinents (Evans *et al.* 2016).

Whilst we appreciate the simplicity of Meert's (2012) 75% threshold and the flexibility of Hoffman's (1999) definition, neither are founded on any particular absolute size or tectonic/geodynamic feature. Hence a hypothetical 'supercontinent' at a time when the spatial extent of continental crust was less than today could be quite small, even at 75%, and exert much less influence on mantle dynamics if the total area of continental crust was, say, just 10% of the world's total surface area (e.g. present-day Eurasia). In addition, we can presently give only very rough estimates of the amount of continental crust at different times throughout Earth history. Consequently, based on these definitions, what we call a supercontinent today could become simply an exceptionally large continent in the future and vice-versa, which is the case for Gondwana.

It is also generally assumed, founded largely on numerical modelling, that a supercontinent represents a single continental superplate (e.g. Gutiérrez-Alonso *et al.* 2008; Li *et al.* 2008; Stampfli *et al.* 2013; Domeier & Torsvik 2014) capable of modifying Earth's mantle dynamics (e.g. Gurnis 1988; Larson 1991; Evans 2003; Coltice *et al.* 2009). However, large-scale continental landmasses and continental superplates are fundamentally different. For example, at present, the continents of Europe, Asia, America and Africa are connected landmasses representing over 80% of the world's continental area. This single landmass, comprising four large plates (Eurasia, Africa, North America and South America) and many microplates, shows intense tectonic activity along the plate boundaries and is being deformed internally. A single continental superplate, however, has no inner boundaries and little to no internal deformation or crust–mantle interactions, and therefore plate interactions would only occur along its outer boundaries, far from the continental core. It has been suggested, however, that superplates may undergo intraplate internal deformation immediately following their amalgamation in order to achieve geodynamic stability (Gutiérrez-Alonso *et al.* 2008).

However, the criterion of a supercontinent as a superplate is not without contradictions. The formation of supercontinents probably produces very-long-wavelength mantle convection cells at spherical harmonic degree 1 (i.e. one hemisphere with downwellings and the other hemisphere with upwellings) or degree 2 (i.e. two antipodal upwellings; Zhong *et al.* 2007). However, the effects of Pangaea on global mantle convection are disputed. According to some, the mantle convection circulation period would be *c.* 50 myr, which would allow Pangaea to modify its cell patterns. (Li & Zhong 2009; Heron *et al.* 2015; Yoshida *et al.* 2015). According to others, however, the amalgamation of Pangaea had only a minor effect on global mantle convection, regardless of the circulation time (e.g. Collins *et al.* 2011). Yet others support longer global mantle circulation periods on the order of 300 myr (Gurnis 1988), which would make long-lived (but debated as a supercontinent) Gondwana, and perhaps Rodinia, the only supercontinents capable of altering convection patterns since most supercontinents are thought to be rather short lived (e.g. Nance *et al.* 2014; Nance & Murphy 2013) or even to break up before they are fully formed (Oriolo *et al.* 2017).

Here, we propose the definition that a 'supercontinent' is a single continental plate of a size capable of influencing mantle convection patterns and core–mantle boundary processes. Accordingly, if a large continent comprising most of Earth's continental crust is unable to influence mantle circulation (e.g. at times when the amount of continental crust was small), then it would not be considered a supercontinent. Whilst recognizing that the required size is likely to have varied through Earth history, the present Eurasian plate may be a good proxy for evaluating the minimum size of such a supercontinent. The Eurasian plate represents more than 10% of Earth's surface and has not been linked to any putative effect on mantle circulation owing to its size. Pangaea, on the other hand, represented a minimum of 25% of Earth's surface area, which is sufficient to influence mantle dynamics in a major way according to numerical modelling experiments (e.g. Coltice *et al.* 2009; Heron *et al.* 2015). Gondwana (*c.* 15% of Earth's surface) may be at the threshold of becoming a supercontinent, especially considering new insights into the statistical cyclicity of supercontinents (Rolf *et al.* 2014) and their putative effects on the lower mantle (e.g. Zhang *et al.* 2009). Although limited, the present-day database for Rodinia and Columbia/Nuna points towards both landmasses being supercontinents under our definition.

One cycle to rule them all

In 1982, Worsley *et al.* were the first to propose the existence of long-term cyclicity in tectonics involving supercontinents, later termed 'the supercontinent cycle' (Nance *et al.* 1988). Along with the name, these workers also gave the cycle a simple kinematic definition: the timespan between (1) rifting and breakup of a supercontinent to (2) reassembly of (nearly) all the dispersed continental plates into a new supercontinent. However the kinematic data for pre-Pangaean supercontinents is, at best, scarce (Mitchell *et al.* 2012; Evans 2013). As a result, the timetable of supercontinental episodicity draws heavily on geochemical and geochronological proxies, such as maxima in orogenic ages (e.g. Worsley *et al.* 1984; Bradley 2011) and global trends of various stable and radiogenic isotopes (e.g. Kemp *et al.* 2006; Collins *et al.* 2011; Dhuime *et al.* 2012; Roberts 2012; Spencer *et al.* 2013, 2014; van der Meer *et al.* 2017).

In addition to its kinematic definition, the original proposal for a supercontinent cycle included a potential driving mechanism, in which the cycle itself modulated mantle heat flow (Worsley *et al.* 1984; Nance *et al.* 1988) as a result of the insulating effect of continent crust (Anderson 1982). Since then, the supercontinent cycle has been both linked to, and decoupled from, long-term mantle circulation (Gurnis 1988; Anderson 1994; Burke & Torsvik 2004; Zhong *et al.* 2007; Torsvik *et al.* 2010; Condie *et al.* 2011; Zhang & Zhong 2011). However, this potential link is still at the core of the geodynamic significance of the supercontinent cycle (Van Kranendonk & Kirkland 2016).

A hypothesis for the supercontinent cycle

If the term 'supercontinent' is to be defined as a single superplate that affects global mantle dynamics, the supercontinent cycle becomes a geodynamic phenomena, rather than just a kinematic stage in Earth's historical puzzle (Fig. 2). The cycle would

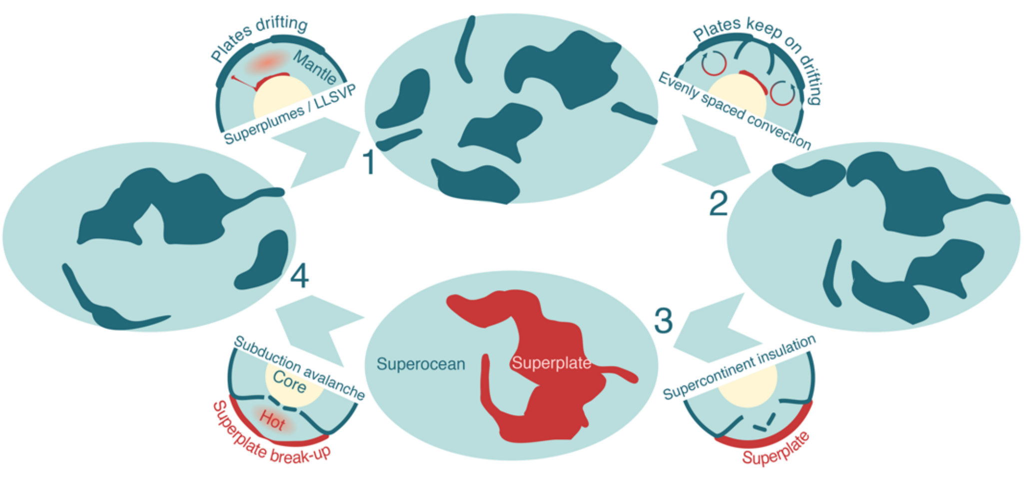

Fig. 2. Proposed influence of the supercontinent cycle on mantle dynamics. (1) *Present-day stage:* plates are dispersed on Earth's surface. (2) *Pre-supercontinent stage*: plates start amalgamating to produce larger continents, eventually merging into a supercontinent. (3) *Supercontinent tenure*: a continental superplate is formed causing mantle upwelling and elevated mantle temperatures. (4) *Supercontinent breakup*: enhanced temperatures below the supercontinent trigger supercontinent breakup. See text for full explanation.

then be part of a long-term cyclicity affecting the geosphere and, consequently, the hydrosphere, atmosphere and biosphere. Otherwise, the cycle would be a kinematic fortuity in which continents occasionally come together and separate out of chance. If such were the case, many of the observed links to other phenomena would have to have a stochastic component.

Figure 2 gives an overall view of the supercontinent cycle hypothesis. Figure 2(1) shows the *present-day stage* – a set of continental plates are dispersed around Earth driven by plate tectonics. Spreading ridges and subduction zones are almost evenly distributed and mantle convection cells are similar in size and dynamism. Figure 2(2) shows the *pre-supercontinent stage* – plates start amalgamating into larger continental plates. Figure 2(3) shows the *supercontinent tenure* – continental plates merge into a continental superplate configuration and a supercontinent is formed. Owing to its size, there is little or no initial crust–mantle interaction recorded in the core of the supercontinent, but mantle upwelling and elevated mantle temperatures are predicted to occur either because the supercontinent acts as an insulator to a large proportion of the mantle (e.g. Anderson 1994; Zhong *et al.* 2007; Heron & Lowman 2011) and/or because the supercontinent becomes largely rimmed by subduction zones, creating a girdle of downwelling which induces upwelling that may cause a rising plume or superplume to develop beneath the supercontinent (Condie 1998; Zhong *et al.* 2007; Li & Zhong 2009; Heron *et al.* 2015). Figure 2(4) shows *supercontinent breakup* – continental insulation and/or the superplume trigger supercontinent breakup, or at least assist other tectonic processes (e.g. roll-back, rifting, ridge subduction; Li & Zhong 2009) that may break the core of the supercontinent, causing the continental fragments to disperse.

Whilst each of these stages is assumed to occur more or less simultaneously in different regions of the globe, some diachroneity is inevitable (e.g. simultaneous continent assembly and breakup in different regions of the supercontinent; Oriolo *et al.* 2017), leading Rogers & Santosh (2009) to advocate the concept of 'maximum packing' to define the point in time when the supercontinent acquires the greatest amount of available continental lithosphere.

Murphy & Nance (2003) suggested two hypothetical end-member scenarios for supercontinent cycle evolution: extroversion and introversion (Fig. 3). Extroverted supercontinents (Murphy *et al.* 2009; Murphy & Nance 2013) are those in which the supercontinent turns 'inside out' (Hartnady 1991; Hoffman 1991). When the parent supercontinent breaks apart, new interior oceans widen whilst the former superocean surrounding the supercontinent closes. All of the pieces then regroup after final closure of the superocean and the interior oceans evolve to become the exterior ocean of the successor supercontinent (Fig. 3a). In contrast, introverted supercontinents are those in which the interior oceans, that is, those oceans that opened during the previous supercontinent breakup, are preferentially consumed to coalesce a new supercontinent (Fig. 3b). Past supercontinents may have formed as a consequence of both processes (Murphy & Nance 2013). Mitchell *et al.* (2012) suggested a process from a mantle perspective called orthoversion, in which the new supercontinent nucleates along a girdle 90° from the centre of its predecessor driven by true polar wander, so if the first supercontinent forms at the equator, the next will form at one of the poles (Fig. 3c). Although strictly there is no kinematic difference between extroversion and orthoversion – movement along a sphere is always a rotation around an Euler pole – the geodynamic implications of orthoversion may vary depending on whether the superplate is centred on the spin axis or lies away from it.

The supercontinent cycle hypothesis is an elegant mechanism for explaining Earth's long-term tectonic cycles and deep-time geodynamic behaviour. However, the limitations of the extant geological record make it difficult to provide an accurate 2- to 4-billion year chronology for Earth's supercontinent cycle and permit many interpretations of major events, including the possibility that the supercontinent cycle does not exist as currently envisaged.

What the supercontinent cycle is not

The concept of periodicity in tectonic processes predates the acceptance of the plate tectonic paradigm (e.g. Umbgrove 1940; Holmes 1951; Sutton 1963). Not until Wilson (1966) asked whether the Atlantic Ocean closed before it opened did such episodicity merged into a plate tectonic framework. Developed and defined as the 'Wilson cycle' by Dewey & Burke (1974), the idea that oceans close and later reopen along former sutures has since become a fundamental concept in plate tectonics (Murphy *et al.* 2006; Buiter & Torsvik 2014).

There is, however, a widespread misconception that the Wilson cycle is synonymous with the supercontinent cycle. The Wilson cycle refers to the history of a single ocean basin and includes the stages of continental rifting (rift valley stage); early oceanization (Red Sea stage); divergence in a mature ocean (Atlantic stage); ocean basin convergence and closure (Pacific stage); and continent–continent collision. The Wilson cycle is not, however, a cycle *sensu stricto* as not all the steps are required; it may abort after the first or second stages and it may span a few million or hundreds of millions of years. Additionally, a Wilson cycle is very

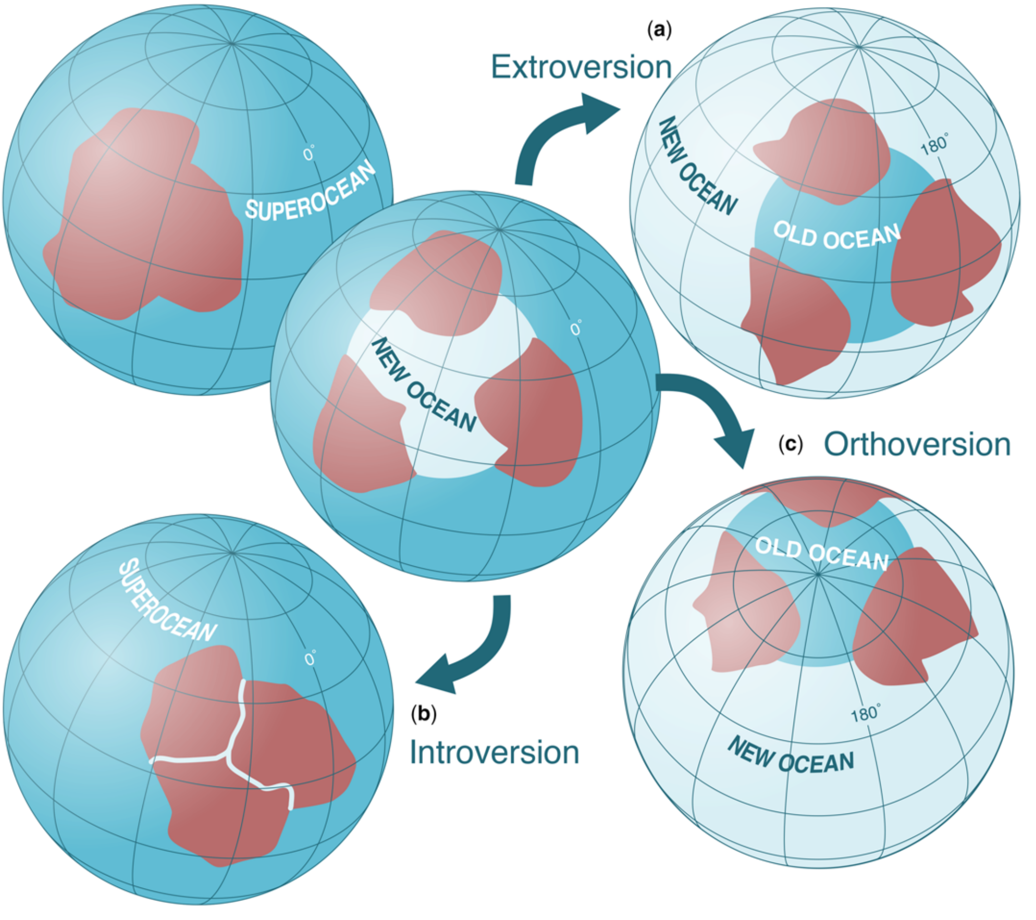

Fig. 3. Three modes of supercontinent amalgamation: (**a**) Extroversion: supercontinent forms through closure of the superocean (supercontinent forms inside-out). (**b**) Introversion: supercontinent forms through closure of the ocean basins responsible for the breakup of the previous supercontinent. (**c**) Orthoversion: supercontinent nucleates along a girdle 90° from the centre of its predecessor. See text for further details.

specific: it starts when rifting and seafloor spreading begin during the separation of two continents, and ends when the last of that seafloor is consumed in a collision (Dewey & Burke 1974).

In contrast, the supercontinent cycle is conceived as a geodynamic mechanism that may explain the long-term behaviour of the entire Earth and involves all of the Earth plates, both continental and oceanic. The process of amalgamation and breakup of supercontinents is thought to happen in a quasi-cyclical manner (*sensu stricto*) with time intervals of *c.* 500 myr (Worsley *et al.* 1984; Nance *et al.* 1988). In addition, whereas there is no hierarchical relationship with the Wilson cycle, a particular supercontinent cycle may involve multiple Wilson cycles, both partial and complete.

Plate kinematic proxies

Studies of plate kinematics use an array of tools that can constrain the time, velocity and location of the plates to provide accurate tectonic, geographic and climatic reconstructions of the past. The breakup configuration of Pangaea is well constrained owing to the preservation of ocean floor from the Jurassic to the present (e.g. Seton *et al.* 2012). However, the precise continental configuration of Pangaea during its amalgamation (cf. Pangaea A, B, C hypotheses; e.g. Muttoni *et al.* 2009; Domeier *et al.* 2012), the number of participating continents (Stampfli & Borel 2002; Stampfli *et al.* 2013; Domeier & Torsvik 2014; Domeier 2016) and their tectonic evolution during the Paleozoic, are still largely unknown and remain controversial (van Staal *et al.* 2009; Isozaki *et al.* 2010; Nance *et al.* 2010; Pastor-Galán *et al.* 2013*a*, b; Waldron *et al.* 2015). Exactly when Pangaea became a bona fide supercontinent, for example, is a matter of debate with age estimates ranging from 330 to 280 Ma (e.g. Veevers 2003; Torsvik 2003; Pastor-Galán *et al.* 2015).

The kinematics of maximum packaging of Gondwana (a.k.a Pannotia/Greater Gondwana) are sufficiently well constrained to indicate that a large landmass merged in the Late Neoproterozoic (e.g. Dalziel 1997; Cordani *et al.* 2013; Tohver & Trindade 2014; Merdith *et al.* 2017; Nance & Murphy 2018). Between 750 and 620 Ma, the East African and Kunga orogens consolidated East Gondwana (Meert 2003; Collins & Pisarevsky 2005), whereas final amalgamation of West Gondwana occurred as the result of numerous continental collisions between 650 and 550 Ma (e.g. Pankhurst 2008; Schmitt *et al.* 2008; Tohver *et al.* 2010). Palaeomagnetic records of the Rodinia to Gondwana transition, however, are limited and their relationships with other geological data are complex (e.g. Meert *et al.* 1993; Evans 2003; Abrajevitch & Van der Voo 2010; Merdith *et al.* 2017). Gondwana remained largely intact until the amalgamation of Pangaea with only ribbon continents separating from it during the Paleozoic (e.g. MacNiocaill *et al.* 1997; van Staal *et al.* 1998, 2009, 2012; Murphy *et al.* 2006; Nance *et al.* 2010; MacDonald *et al.* 2014; Waldron *et al.* 2014). So, in contrast to Pangaea, Gondwana was a very long-lived landmass. Because of this, and its size as a subset of Pangaea, Gondwana's membership in the supercontinent club is hotly debated (e.g. Evans *et al.* 2016). However, by denying the supercontinent status of Pannotia/Greater Gondwana we may diminish the importance of understanding the coupling between the lithosphere and convecting mantle at a critical time in the evolution of Earth's systems within the context of the supercontinent cycle (Murphy 2013; Cawood *et al.* 2016).

There is also growing kinematic evidence for the existence of Rodinia, which formed at *c.* 1.0 Ga as a result of several collisional orogenies sometimes referred to collectively as 'Grenvillian' (e.g. Torsvik 2003; Goodge *et al.* 2008; Halls 2015). However, the configuration of Rodinia is uncertain and, as a consequence, very different plate tectonic reconstructions have been suggested (Weil *et al.* 1998; Kröner & Cordani 2003; Pisarevsky *et al.* 2003; Piper 2007; 2015; Goodge *et al.* 2008; Li *et al.* 2008; Evans 2009; Evans *et al.* 2016; Cawood & Pisarevsky 2017; Merdith *et al.* 2017). Rodinia appears to satisfy our definition of a supercontinent, in that it was large enough to influence mantle convection: (1) the landmass that amalgamated is estimated to have represented a minimum of 18% of Earth's surface area (Li *et al.* 2013), even though many blocks lack palaeomagnetic and other sources of kinematic data (Evans 2013); and (2) the orogens that developed as a result of Rodinia's amalgamation were exceptionally hot, indicating enhanced mantle activity (Beaumont *et al.* 2010).

Shortening from 2.1 to 1.9 Ga, which is common in many of the planet's cratons (Zhao *et al.* 2002), is the main kinematic evidence supporting the existence of a late Paleoproterozoic supercontinent. However, the shape, number of participating cratons, the timing of assembly–disruption, and even the name of the supercontinent – variously Columbia (Rogers & Santosh 2002), Nuna (Hoffman *et al.* 1998) and Paleopangaea (Piper 2013) (see Meert (2012) and Evans *et al.* (2016) for further discussion) – are disputed. In contrast with Columbia/Nuna/Paleopangaea, the hypothetical supercontinent at the Archaean–Paleoproterozoic transition, named Kenor (sometimes written as Kenorland; Barley *et al.* 2005), is supported by some age data but not by reliable kinematic data (e.g. Bradley 2008; Evans *et al.* 2016). The existence and cratonic configuration of such a supercontinent are, therefore, difficult to test. As a result, alternative palaeogeographic models envisage distinct (super)continents like Kenor made up of several large cratons, which were possibly the first original pieces of continental crust (a.k.a. supercratons) named Ur, Superia, Sclavia and Vaalbara (e.g. Bleeker 2003; de Kock *et al.* 2009; Gumsley *et al.* 2017).

Secular trends in the geological record

Owing to the lack of kinematic constraints, episodes of supercontinent amalgamation have been established by way of maxima in the number of collisional orogenies, whereas the timing of breakup has been determined from the ages of large igneous provinces (e.g. Ernst & Bleeker 2010; Ernst *et al.* 2013). Based on this approach, Pangaea, Pannotia/Greater Gondwana (if a supercontinent), Rodinia and Columbia/Nuna can be recognized (Fig. 4). However, a number of other geological features show long-term secular trends that are both quasi-cyclical and broadly coincident with the temporal maxima in orogenesis (e.g. Worsley *et al.* 1984; Nance *et al.* 1986). These features may constitute proxies in the geological record that can be used to track stages within the supercontinent cycle. However, whilst the supercontinent cycle concept is a convenient way to explain these global-scale phenomena, the proxies by themselves cannot demonstrate the existence of the cycle until unequivocally constrained by plate kinematics. These long-term trends have also been explained by geodynamic processes that may reflect specific events within the supercontinent cycle such as

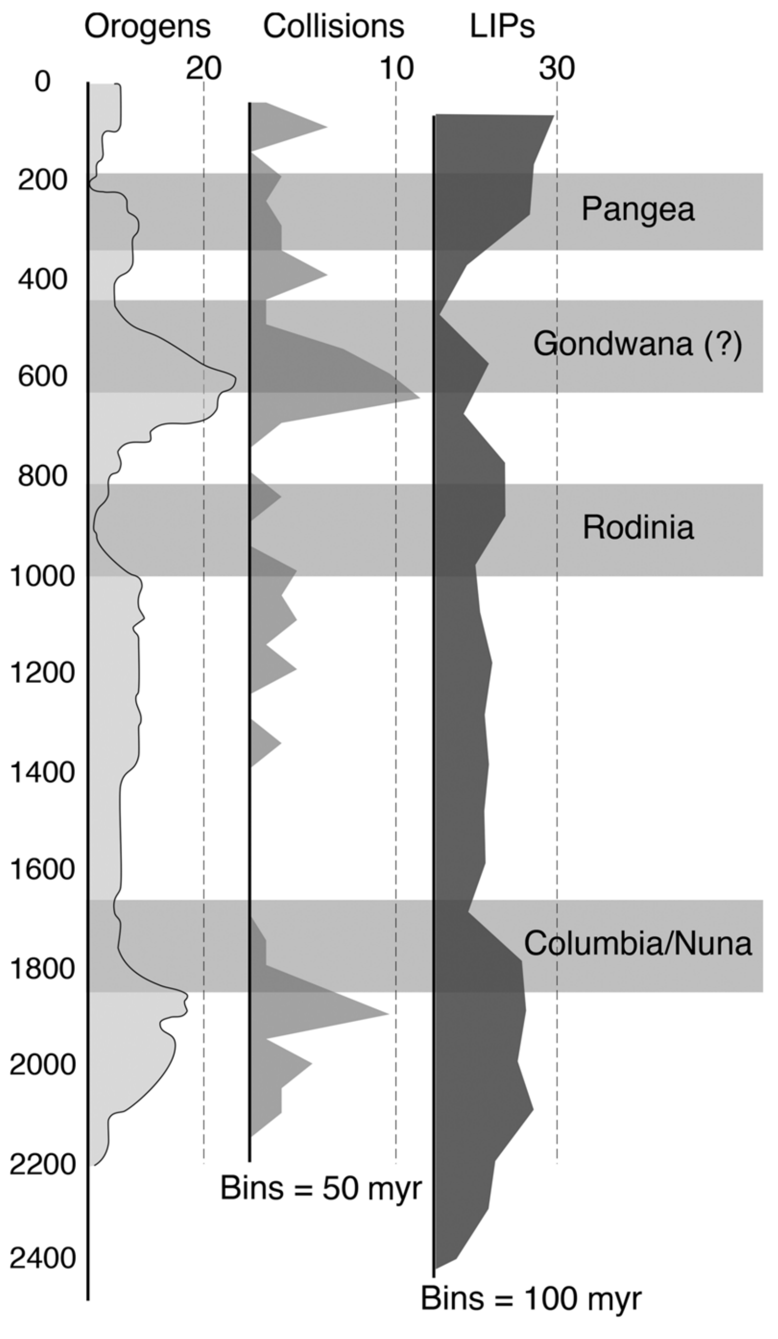

Fig. 4. Temporal distribution of orogens, collisions and LIPs using the databases of Condie & Aster (2013) and Condie *et al.* (2014).

(1) periodic mantle superplumes (Condie 1998), (2) episodic subduction-related slab 'avalanches' (Stein & Hofmann 1992) and (3) intervals during which plate motion and subduction accelerate or slow down (Silver & Behn 2008; O'Neill *et al.* 2009; Dhuime *et al.* 2012). It is important to recognize that, whilst episodicity in collisional orogenesis and the emplacement of large volumes of magmatic

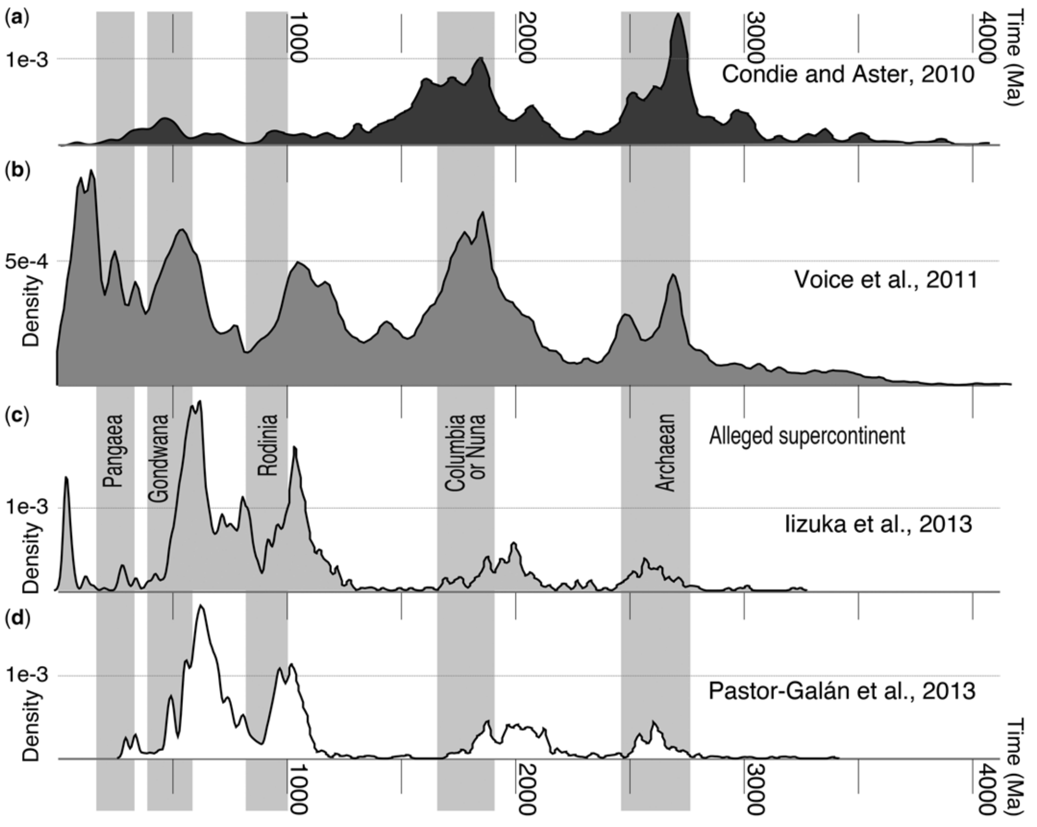

Fig. 5. Zircon-forming events from four independent databases: (**a**) Zircon from igneous rocks (Condie & Aster 2010, (**b**) Global database of detrital zircon (Voice *et al.* 2011), (**c**) detrital zircon from present day rivers (Iizuka *et al.* 2013) and (**d**) detrital zircon from a Paleozoic basin in NW Iberia (Pastor-Galán *et al.* 2013*a*). All four plots are kernel density estimations with a smoothing bandwith of 20. All peaks are broadly coincident despite the differences between databases.

rocks might be an expected consequence of the supercontinent cycle, this is not true in reverse, so the occurrence of peaks in the ages of these phenomena does not demonstrate the existence of such a cycle.

Zircon forming events

The recent profusion of precise U–Pb zircon geochronological analyses has permitted the creation of global databases of both detrital and magmatic zircon ages (e.g. Condie & Aster 2010; Bradley 2011; Dhuime *et al.* 2011; Voice *et al.* 2011). Many studies interpret zircon age maxima as major zircon-forming events (e.g. Condie *et al.* 2009; Bradley 2011) and most of the studies correlate these events with the assembly of supercontinents (Fig. 5; e.g. Condie & Aster 2010; Hawkesworth *et al.* 2010; Roberts 2012; Cawood *et al.* 2013; Van Kranendonk & Kirkland 2016). Some authors have claimed that supercontinent cycles can be seen in detrital zircon ages collected from present-day rivers (e.g. Iizuka *et al.* 2013) and even from single localities if sampled sufficiently (Pastor-Galán *et al.* 2013*a*).

The possibility of a major bias in the zircon record exists, since global data coverage is uneven and data are scarce in remote areas. Despite minor differences, however, the age of zircon-forming events is consistent in independent databases (Fig. 5; e.g. Condie *et al.* 2009; Voice *et al.* 2011; Iizuka *et al.* 2013; Pastor-Galán *et al.* 2013*a*). It is, therefore, difficult to argue that the zircon maxima are artefacts of uneven sampling.

There are, however, other possible artefacts in the zircon record. Zircon generated in subduction-related environments developed prior to collision may be more likely to be preserved than those generated in other tectonic settings (Hawkesworth *et al.* 2009, 2010), making the zircon-forming events a prelude to supercontinent formation. In contrast, Spencer *et al.* (2015) argue that, in the case of the Grenville orogeny, the zircon age peak can be strictly ascribed to collisional and post-collisional magmatism. The zircon-forming events could also be related to major accretionary orogenies (e.g. Cawood *et al.* 2009; Fernández-Suárez *et al.* 2014). Although supercontinent amalgamation is thought to be preceded by accretionary orogenesis as continents converge, accretionary orogenesis does not necessarily imply the formation of supercontinents. Hence, the zircon maxima, whilst correlating with supercontinents, could be a consequence of preservation bias rather than being a strict record of zircon-forming events.

Hf and O isotopes in zircon

The combination of U–Pb ages and the isotopic systems of Lu–Hf and O in zircon has been used to better understand the growth of continental crust and the amalgamation of supercontinents (e.g. Condie & Aster 2010; Condie *et al.* 2011; Voice *et al.* 2011; Dhuime *et al.* 2012; Spencer *et al.* 2014). Hf and O isotopes are powerful tools for revealing whether zircon crystallized from juvenile magma during crustal generation or from magma generated by crustal reworking; radiogenically enriched Hf signatures occur in continental crust whereas radiogenically depleted Hf signatures characterize juvenile mantle. As a result, the $^{176}Hf/^{177}Hf$ ratio (commonly expressed as εHf with respect to bulk silicate Earth) can be used as a measure of crustal residence age (i.e. the time from mantle extraction to crustal formation; Hawkesworth & Kemp 2006). Used together, U–Pb ages and Hf depleted mantle model ages can be used to evaluate the length of time between the crystallization of the zircon and its initial separation from the mantle. In this way, combined isotope signatures are useful tools for studying the processes of crust formation and crustal residence time. In plots of εHf v. time, positive excursions (i.e. towards high εHf values) indicate a mantle-like signature, whereas negative excursions indicate crustal reworking (Fig. 6; Collins *et al.* 2011; Roberts 2012; Henderson *et al.* 2016). Compiled zircon Hf data (Belousova *et al.* 2010; Lancaster *et al.* 2011; Dhuime *et al.* 2012; Roberts & Spencer 2015) show several periods of presumed growth and reworking of continental crust that are quasi-cyclical and, in some cases, coincident with maxima and minima in zircon-forming events and thus periods of alleged supercontinent assembly. However, the correlation is not straightforward since zircon-forming events may correlate with either maxima (e.g. 350 Ma; Fig. 6) or minima (e.g. 550 or 1000 Ma) in zircon εHf. Collins *et al.* (2011) suggested that Hf may be controlled by orogenic style; accretionary-Pacific-style orogens forming new juvenile crust (positive Hf excursion) whereas continental collisions rework old crust resulting in fanning arrays in εHf diagrams. Alternatively, Spencer *et al.* (2013) and Gardiner *et al.* (2016) respectively posit that subduction polarity or subduction angle immediately prior to continental collision is responsible for the isotopic signature.

Some researchers have tried to understand supercontinent amalgamation by combining Hf isotope data with O isotopes, the fractionation of which is time independent. The ratio between oxygen isotopes (^{18}O and ^{16}O expressed as $\delta^{18}O$) in mantle-derived melts is 5.3 ± 0.6‰ (2σ; e.g. Hawkesworth & Kemp 2006). The isotopic system, however, is sensitive to low-temperature subaerial processes driven by meteoric water. Therefore supracrustal rocks are enriched in ^{18}O, giving higher $\delta^{18}O$ values. Zircon formed in magma with appreciable supracrustal assimilation have $\delta^{18}O > c.\ 6.0$‰, whereas mantle-derived zircon without significant crustal assimilation will have $\delta^{18}O$ values of *c.* 5.5‰ (Hawkesworth & Kemp 2006). Therefore, intervals of continental (supracrustal) reworking and enhanced mantle output should respectively yield opposite $\delta^{18}O$ excursions. Maxima and minima in zircon $\delta^{18}O$ values through geological time also show a somewhat cyclic pattern that broadly coincides with zircon-forming events and excursions of in εHf (Spencer *et al.* 2014; Payne *et al.* 2016). Importantly, the peaks of $\delta^{18}O$ seem to occur during supercontinent assembly and not during the tenure of the supercontinent.

Seawater Sr and Nd

The $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ ratios in Earth's ocean water are the result of juvenile input from the mantle along ocean ridge, oceanic arcs and large igneous provinces, and more geochemically diverse riverine input from continents. Consequently, when mantle-derived magmatism dominates the Earth system (i.e. when continental arcs and collisions are few), seawater shows a depleted radiogenic signal. In contrast, if global tectonics is marked by multiple continent–continent collisions, the radiogenic isotopic signatures are enriched. In the context of the supercontinent cycle, the former should correspond to breakup whereas the latter would be expected during amalgamation. Sr and Nd isotopes in marine sediments provide a record of the oscillations between mantle and erosional sources through Earth history (Nance *et al.* 1986; Keto & Jacobsen 1988; Veizer 1989; Peucker-Ehrenbrink *et al.* 2010; Peters & Gaines 2012; Condie & Aster 2013; van der Meer *et al.* 2017 and references therein).

Global compilations of Sr (McArthur *et al.* 2012) and Nd (Keto & Jacobsen 1988; Fig. 7) isotopes

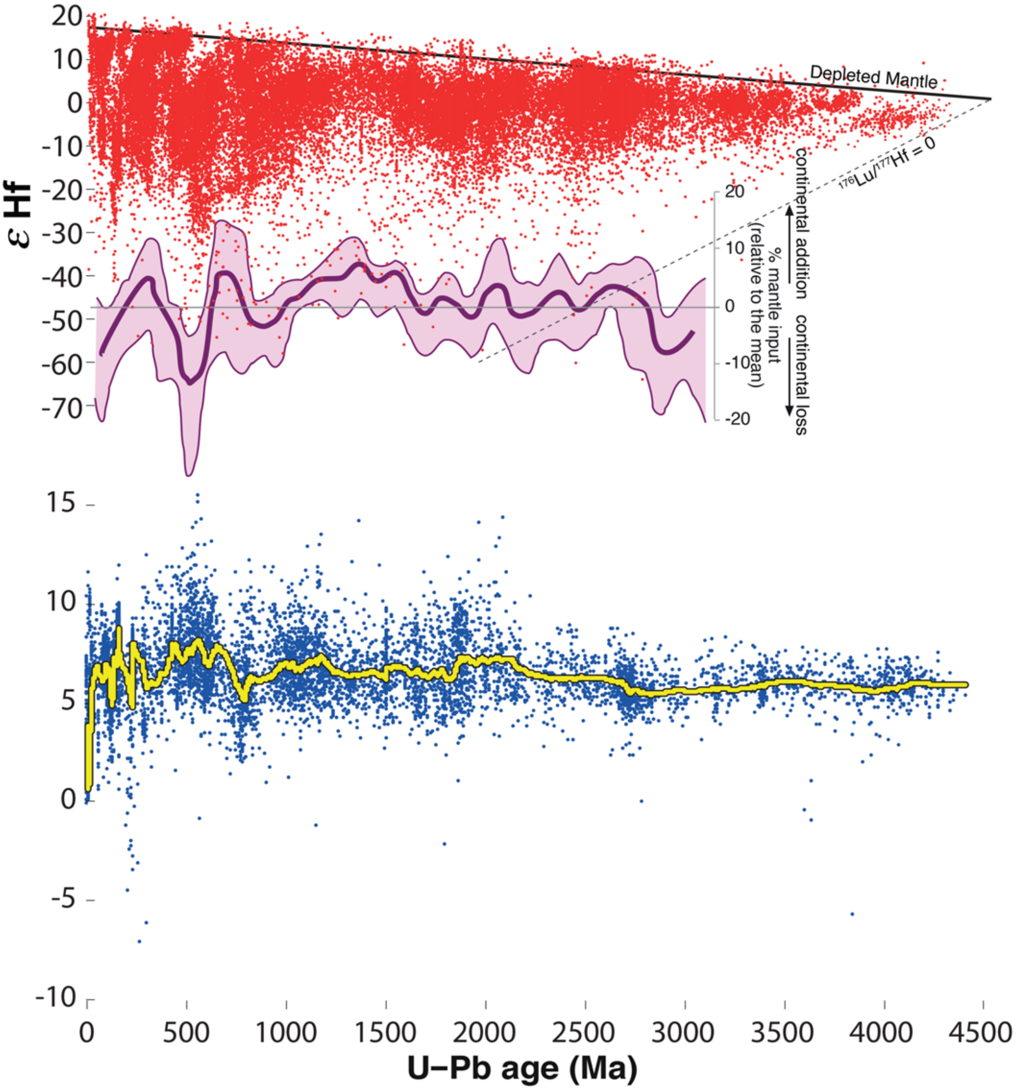

Fig. 6. Compilation of zircon εHf after Roberts (2012) and Roberts & Spencer (2015), and $\delta^{18}O$ (new compilation augmented from Spencer *et al.* 2014 and Payne *et al.* 2016; see DR1 for references). Red dots = compilation of global zircon Hf data from Roberts & Spencer (2015), plotted as U–Pb zircon age v. $\varepsilon Hf_{(initial)}$. For details about percentages of continental growth and loss see Roberts (2012). Blue dots = $\delta^{18}O$ values from this study together with the moving average.

in authigenic marine minerals show good resolution from the Neoproterozoic to the present (Fig. 7). The global trend in both isotopes broadly correlates (Fig. 7) despite the significantly shorter residence time for Nd (300–600 years; Arsouze *et al.* 2009; Charbonnier *et al.* 2012) over Sr (1 to 20 myr since the Late Cambrian; Vollstaedt *et al.* 2014), which makes Nd a more immediate recorder but also more easily affected by local events. Oscillations in the Sr and Nd isotopic signatures, as expected, show opposite trends and the peaks and troughs are coincident with those shown by zircon-forming events and Hf and O isotopes (Fig. 7). Surprisingly, however, the lowest strontium ratios and highest εNd values coincide with the amalgamations of Pangaea and Rodinia, which is the opposite correlation to the one that would be expected in a world dominated by collisions and orogenesis.

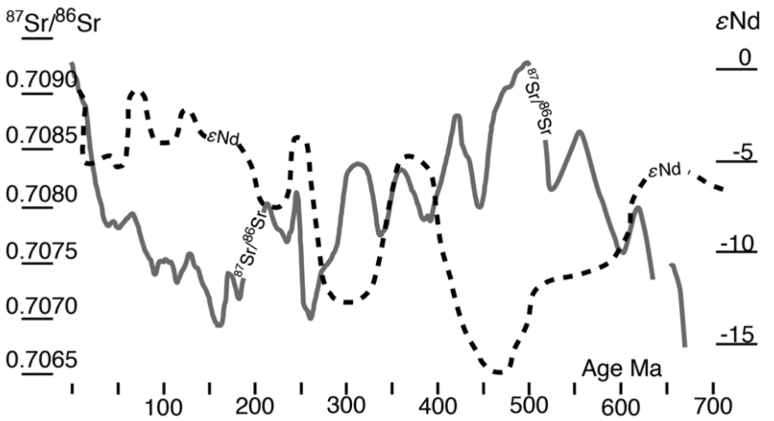

Fig. 7. Global compilation of Sr (McArthur *et al.* 2012) and Nd (Keto & Jacobsen 1988) in seawater showing a strong correlation with the Pangaea cycle in spite of their very different residence times in ocean water.

Metallogenic provinces

The temporal and geographic distribution of mineralization (particularly beryllium, boron, copper, mercury and molybdenum) also reveals an episodic pattern of development and diversification with time (Barley & Groves 1992; Goldfarb *et al.* 2001; Leach *et al.* 2005; Hitzman *et al.* 2010; Huston *et al.* 2010; Maynard 2010; Slack 2013; Hazen *et al.* 2014; Cawood & Hawkesworth 2015). A statistically significant increase in the number of reported mineralization localities and/or the appearance of new mineral species occurs at times (*c.* 2800–2500, *c.* 1900–1700, *c.* 1200–1000, *c.* 600–500 and *c.* 430–250 Ma) that broadly coincide with peaks in zircon-forming events. These time intervals are also broadly coincident with proposed episodes of supercontinent assembly and tenure, including Gondwana. In contrast, fewer deposits and/or fewer new mineral species containing these elements have been reported from the intervals *c.* 2500–1900, *c.* 1700–1200, 1000–600 and 500–430 Ma. The database further suggests that not all these metallogenic peaks are equal (Hazen *et al.* 2014), perhaps indicating that the not all continents formed equally and that their particularities may also influence mineralization (e.g. Spencer *et al.* 2013).

Large igneous provinces, large low-shear-velocity provinces and plate tectonics

The supercontinent cycle has been traditionally interpreted on the basis of tectonic and geodynamic processes comparable with those operating today, assuming that some form of plate tectonics has operated at least since the end of the Archaean Eon. However, whilst modern plate interactions are relatively well understood, the underlying long-term global tectonics and geodynamics are not. As a result, understanding how the supercontinent cycle works is prerequisite to understanding how the inner Earth works.

Top-down and bottom-up

The occurrence of intraplate hotspot volcanism cannot be explained by classical plate tectonic processes, despite the fact that hotspots have been key to providing reference frames for absolute plate reconstructions (Wilson 1963; Morgan 1972; Doubrovine *et al.* 2012). Present day intraplate hotspot volcanism is relatively common but, in general, represents a minor input of new volcanic rocks compared with arc volcanism. However, there have been numerous episodes in space and time of intense intraplate volcanism, usually referred to as large igneous provinces (LIPs) and commonly attributed to superplumes (e.g. Bryan & Ernst 2008; Ernst *et al.* 2008). Recent plate-kinematic reconstructions in a mantle reference frame for the Mesozoic and Cenozoic have linked eruption sites of LIPs and kimberlites with present-day plume locations (Fig. 8; Torsvik *et al.* 2010). These reconstructions suggest that most plumes form at the edges of the large low-shear-velocity provinces (LLSVPs), which are two antipodal areas located along the core–mantle boundary (below the Pacific and Africa) that have been identified through mantle tomography (e.g. Ritsema *et al.* 1999; Burke & Torsvik 2004). Furthermore, geological evidence links LIP occurrences and supercontinents (e.g. Bleeker 2003; Ernst *et al.* 2008; Li & Zhong 2009). The long-term stability of LLSVPs and their connection to LIPs and the

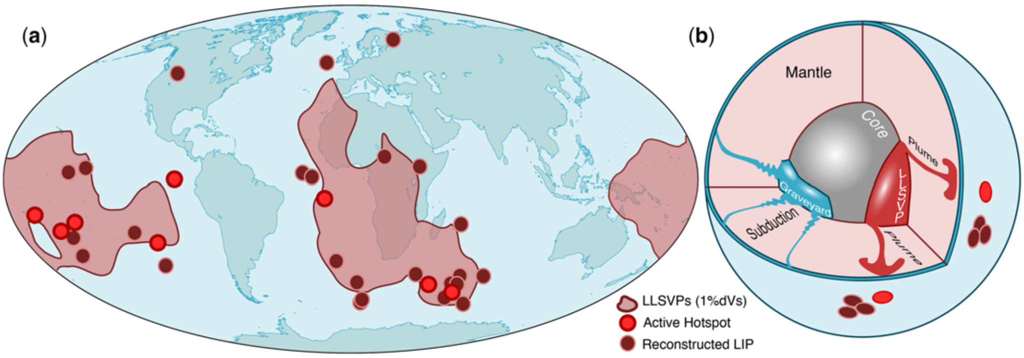

Fig. 8. (**a**) Location of LLSVPs within the Earth and their spatial relation with major hot spots and reconstructed LIPs (after Torsvik *et al.* 2010). Most of hotspot volcanism appears to occur at the edges of the LLSVPs. (**b**) Cartoon depicting the location of LLVPs.

supercontinent cycle are now the subject of intense debate, in which there are three main competing hypotheses:

(1) Plumes are shallow features in the upper mantle and are the result of plate tectonics, and LIPs (superplumes) are the result of thermal insulation of the supercontinent's continental lithosphere (e.g. Anderson 1994). This implies a 'top-down' tectonic system in which the formation of supercontinents is responsible for the formation of LIPs, but without involving the lower mantle. This hypothesis implies that LLSVPs have little or no influence on plumes and LIPs (Foulger 2010, 2012; Julian *et al.* 2015). The hypothesis has been tested through numerical modelling with contrasting results; several authors have found that the formation of supercontinents would cause sufficient thermal insulation of the mantle to raise temperatures enough to produce melting (e.g. Coltice *et al.* 2009; Rolf *et al.* 2012), whereas others have found that this is not the case and that the formation of LIPs after supercontinent amalgamation is likely to be a response to subducting slabs surrounding the supercontinent (e.g. Heron & Lowman 2010; Heron *et al.* 2015).

(2) The supercontinent cycle and the formation of antipodal LLSVPs are linked. In this model, LLSVPs are the result of slab graveyards in the circum-supercontinent subduction system (Evans 2003; Li *et al.* 2008; Li & Zhong 2009). This hypothesis implies a dynamic feedback between 'top-down' and 'bottom-up' tectonics. The formation of supercontinents ringed by subduction zones would be responsible for the formation/fuelling and location of LLSVPs (Maruyama 1994; Maruyama *et al.* 2007), which in turn would form superplumes. This hypothesis has also been tested numerically with contrasting results. Some modelling studies support LLSVPs being the result of slab graveyards (e.g. Zhong *et al.* 2007; Tackley 2011), whereas other models find accumulating slabs at the core–mantle boundary to be very difficult (Li & McNamara 2013). Heron *et al.* (2015) have suggested that the ring of subduction around supercontinents may form plumes regardless of the nature of LLSVPs. This model predicts major True Polar Wander events during supercontinent cycles (e.g. Evans 2003; see the section 'Supercontinent and Earth's magnetism: True polar wander and superchrons').

As 'top-down' geodynamic models, hypotheses 1 and 2 both have supercontinents breaking up over geoid highs and reassembling over geoid lows (represented at the surface by subduction zones) (Anderson 1994; Gurnis 1988). Murphy *et al.* (2009) argue that 'top-down' tectonics would explain supercontinent formation by extroversion (closure of the circum-supercontinent ocean), but fail to explain supercontinent formation by introversion (closure of the oceans formed by supercontinent breakup), as was the case with Pangaea.

(3) Hotspots are generated at the edges of LLSVPs and are linked to kimberlites. In this model LLSVPs are stable, long-lived features, lasting at least 500 myr (Torsvik *et al.* 2010) and possibly since early in Earth's history (Dziewonski *et al.* 2010; Burke 2011), in which case they would predate the supercontinent cycle and even plate tectonics. In this hypothesis, LLSVPs play a major role in long-term mantle circulation (Dziewonski *et al.* 2010). The

hypothesis consequently employs 'bottom-up' tectonics. The idea that LLSVPs and kimberlites are fixed has enabled their use as a mantle reference frame to constrain palaeolongitude (Torsvik *et al.* 2012; Domeier & Torsvik 2014). Bull *et al.* (2014) used numerical modelling and found coherence between these reconstructions and the hypothesis of fixed LLSVPs. Amongst the critics, Julian *et al.* (2015) suggested the hypothesis involved circular reasoning and *ad-hoc* interpretations, and Evans (2010) pointed out that the formation of Pangaea in the context of bottom-up tectonics over fixed convection supercells (Collins 2003) is counter-intuitive, since, according to this hypothesis, it would have to have formed over an upwelling-spreading location (present day African LLSVP).

Supercontinents and the start of plate tectonics

The question of when modern-style plate tectonics started is unresolved, with suggestions ranging from very early in Earth's history (e.g. Harrison 2009; Bercovici & Ricard 2014) to the late Neoproterozoic (Stern 2005, 2007; Stern *et al.* 2016; Hamilton 2011). The start of plate tectonics is arguably one of the most important questions in Earth science and is crucial to understanding the supercontinent cycle. For instance, a late start for modern-style plate tectonics (at *c.* 750 Ma) would invalidate our definition of the supercontinent cycle, but at the same time would require that most of Earth's continental crust did not form by subduction. The latter is a common thought for Hadean and early Archaean times (e.g. Bédard 2018; Johnson *et al.* 2017) and requires a non-uniformitarian mechanism to explain the formation of major orogenic belts and secular trends. However non-uniformitarian mechanisms are commonly invoked for Hadean–Early Archaean times but not usually considered in late Archaean and Proterozoic times (Condie & Kröner 2013; Spencer *et al.* 2014). The present database actually permits an Early Archaean start of plate tectonics in which the formation of supercontinents is not linked to global mantle dynamics. This possibility might include a simple 'bottom-up' system dominated by two fixed LLSVPs (e.g. Torsvik *et al.* 2010), in which case supercontinents would be kinematic accidents of plate tectonics.

Supercontinent and Earth's magnetism: true polar wander and superchrons

Although the details are unclear, it is generally accepted that movement within the liquid outer core produces the geomagnetic field, which approximates a dipole with minor non-polar contributions. When averaged over thousands of years the locations of the mean geomagnetic poles coincide with Earth's spin axis, forming a geocentric axial dipole (GAD) (Hospers 1954). Although it is not known whether the GAD has always operated (Meert 2009), there is considerable palaeomagnetic evidence supporting it. Palaeomagnetists generally assume GAD when developing apparent polar wander paths, which were central to the development of plate tectonics as they currently are to tectonic reconstructions of the past (e.g. Torsvik *et al.* 2012).

True polar wander

In contrast to apparent polar wander, true polar wander (TPW) is caused by the rotation of the crust and mantle with respect to its spin axis (Fig. 9). In a stable state of rotation, the maximum moment of inertia axis is aligned with the spin axis, with the intermediate and minimum moment of inertia along the equatorial plane. When this is not the case, TPW occurs and the crust and mantle will rotate as a rigid body in an attempt to realign the maximum moment of inertia axis with the spin axis.

TPW has been described over a wide range of time scales, from millions of years in connection with major geodynamic processes, to quasi-instantaneous earthquake-triggered crustal advection (Mitchell 2014). At present, TPW is occurring at a rate of *c.* $10\ \mathrm{cm\ a^{-1}}$, probably owing to Holocene reduction in the size of Earth's ice caps (Evans 2003; Mitchell 2014 and references therein). The mass distribution of Earth is inherently uneven and potential TPW events can be triggered by plate motions that result in supercontinent amalgamation. Several authors have linked major TPW events to the supercontinent cycle. For a long-lived superplate ringed by subduction zones (as would be the case for a supercontinent), the slabs descend as a conical 'curtain', circumscribing the superplate (Anderson 1994; Evans 2003). This scenario, in turn, produces major mantle density discontinuities, resulting in excess inertia. To compensate, the full Earth undergoes TPW to centre the supercontinent and subduction ring on the equator (Zhong *et al.* 2007).

TPW has been used to constrain pre-Pangaean supercontinents (Mitchell *et al.* 2012; Evans 2013). Pangaea, however, formed roughly on the equator and the estimated TPW is very small (Steinberger & Torsvik 2008), making it difficult to test the link between TPW and supercontinent formation. Unravelling TPW effects also requires a robust palaeomagnetic dataset from independent continents and time coverage to test whether all continents move in unison, a requirement of TPW events. Without a sufficient dataset including different landmasses, TPW can be misinterpreted as the rapid movement of a single continent. This poses difficulties for pre-

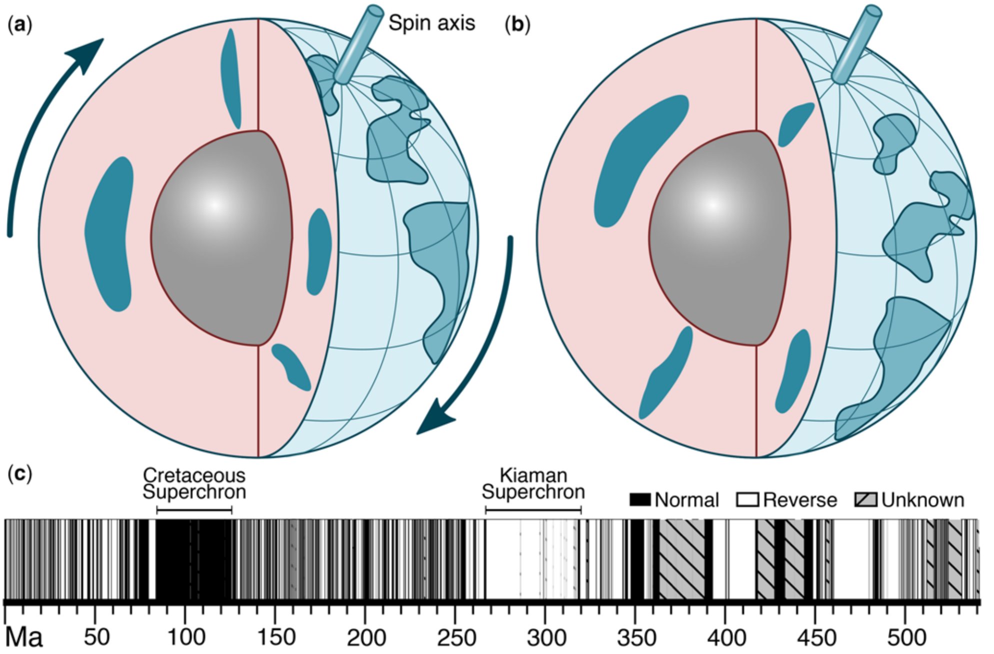

Fig. 9. Figure showing how, when the mass distribution of the Earth is not aligned with the largest moment of inertia, (**a**) the crust and mantle will rotate with respect to the spin axis and (**b**) realign the largest moment of inertia axis with the spin axis in a process called TPW. (**c**) Earth's magnetostratigraph with two major superchrons in the Permian and Cretaceous, broadly coincident with the amalgamation and breakup of Pangaea.

Cretaceous datasets and is very challenging for the pre-Permian (Torsvik *et al.* 2012). Interpretations of large TPW in the pre-Mesozoic should consequently be treated with caution, especially in cases where the palaeomagnetic coverage is limited in space and time.

Superchrons

The pattern of reversals of Earth's magnetic field appears to be random with the time span of chrons usually varying from 0.1 to 1 myr, although Earth has recorded superchrons (>10 myr with one polarity) as long as 50 myr (e.g. Permian superchron; Langereis *et al.* 2010).

The occurrence of two superchrons following the amalgamation of Pangaea – in the Late Carboniferous (Kiaman) and Permian – and another in the Cretaceous during its breakup, have stimulated debate about the relationship between superchrons and supercontinents (e.g. Eide & Torsvik 1996). However, in the absence of sufficient palaeomagnetic information for other supercontinents to establish a longer reference frame, such a genetic connection is difficult to test. Numerical simulations of the geodynamo suggest that superchrons may occur after periods of rapid polarity reversals, which may have been triggered by a decrease in core–mantle boundary heat flow (Biggin *et al.* 2012). These changes in heat flow have been additionally linked to differential activity in LLSVPs, perhaps related to supercontinent amalgamation and breakup (Zhong *et al.* 2007), to reduced mantle plume-head production at the core–mantle boundary, and to episodes of TPW (Biggin *et al.* 2012). However, in their numerical simulations, Driscoll & Evans (2016) found no obvious relationship between superchron occurrence and phases of the supercontinent cycle.

Supercontinents' cold shoulders: connections with climate

Sea-level changes and glaciations recorded in the sedimentary and isotopic record were amongst the first proxies used to support cyclic changes in global climate (e.g. Umbgrove 1947; Fischer 1984; Worsley *et al.* 1984, 1991; Nance *et al.* 1986; Miller 2005). Along with the well-studied climatic changes caused by short-term (20 to 400 kyr) orbital forcing (Milanković 1930; Imbrie *et al.* 1992), long-term climate cycles (e.g. Umbgrove 1947) have been linked to tectonics and the supercontinent cycle (Nance *et al.* 1986, 2014; van der Meer *et al.* 2017 and references therein).

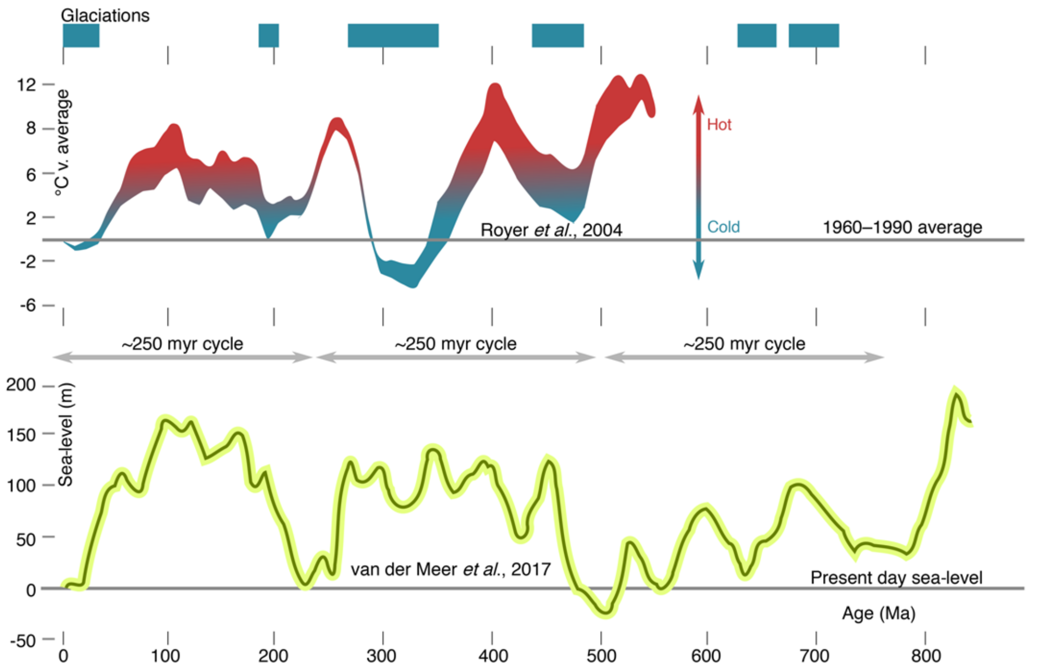

Fig. 10. Major Neoproterozoic and younger glaciations (Hoffman 2009), Phanerozoic global average temperatures (Royer *et al.* 2004), and Neoproterozoic to Recent sea-level fluctuations (van der Meer *et al.* 2017) showing a strong coupling between climate and sea-level with a cyclicity of *c.* 250 myr, or half the alleged duration for the supercontinent cycle.

Sea-level changes

Sea-level fluctuations may be influenced by many factors, including the loci of glaciations, the distribution of continents and oceans (large polar continents tending to accumulate ice), the average age of the ocean basins, and the distribution of geoid highs and lows (Worsley *et al.* 1984; Nance *et al.* 1986). The linkage between sea-level changes and supercontinents, therefore, requires a high-resolution record of fluctuations. The Phanerozoic record, especially from Triassic times, is of sufficient resolution to distinguish between sea-level changes driven by tectonics and those driven by other factors. However, whilst there are tantalizing indications, the Proterozoic record is too poorly known to investigate potential correlations with supercontinent cycles (Bradley 2011).

Umbgrove's (1947) proposal for a 250 myr cycle of sea-level maxima and minima for the last 750 Ma, and its revival in a plate tectonic context by Fischer (1984), is supported by many subsequent sedimentological studies (Bradley 2011 and references therein) and by the $^{87}Sr/^{86}Sr$ isotopic record used as a proxy for sea-level fluctuations (Nance *et al.* 1986; Van der Meer *et al.* 2017 and references therein). These 250 myr cycles (Fig. 10) are approximately half the length of the proposed duration of the quasi-periodic supercontinent cycle (*c.* 500 myr; Worsley *et al.* 1984; Nance *et al.* 1986, 2014; Bradley 2011; Bradley 2011; Condie *et al.* 2015; Van Kranendonk & Kirkland 2016). First-order sea-level changes also seem to correlate with the amalgamation and breakup of Pangaea. Sea-level was high in the early to middle Paleozoic, low in the late Paleozoic to early Mesozoic, high during the Cretaceous, and relatively low today (Fig. 10). The Late Paleozoic to Early Mesozoic low matches the tenure of Pangaea (Worsley *et al.* 1984; Nance *et al.* 1986), whereas the preceding major lowstand (at *c.* 550 Ma; Fig. 10) correlates with the tenure of Pannotia/Greater Gondwana (Condie *et al.* 2005).

The correlation of sea-level with major ice ages (Fig. 10) and the coincidence between the length of the sea-level cycles and the formation of Pangaea suggests a correlation with the supercontinent cycle. However, the correlation between Sr isotopes, which is a primary proxy for the sea-level curve, is counterintuitive to the formation of supercontinents (see section 'Seawater Sr and Nd'). Van der Meer *et al.* (2017) suggested that this 250 myr cyclicity may be the effect of subduction on heat flow at the core–mantle-boundary, or could instead be related to the survival time of deeply subducted slabs.

Glaciations

Glacial intervals during Earth's history (Hoffman 2009) can be directly correlated with sea-level changes, accumulation of ice on continents effectively removing water from the oceans. Multiple factors are responsible for glacial periods, but plate tectonics and related LIPs are one of the fundamental controlling influences. Plate tectonics appears to have controlled CO_2 levels, at least since the Triassic (van der Meer *et al.* 2014), since changes in the location of landmasses, even relatively small ones, may change global climate. For example, a minimal change like the closure of the Panama seaway at *c.* 4 Ma, altered global oceanic circulation, which resulted in northern hemisphere glaciation (e.g. Bartoli *et al.* 2005). TPW is another factor capable of changing global climatic zones. For example, a TPW of 30° will move a continent located in a subpolar area (60°) to a subtropical one (30°) (e.g. Muttoni & Kent 2016). The formation and disruption of supercontinents is also a candidate for triggering major changes in global climate since this both redistribute landmasses and probably causes TPW events.

Some authors have suggested that global glaciation may be caused by supercontinent amalgamation, especially if the supercontinent amalgamates at low latitudes (e.g. Worsley *et al.* 1986, 1991; Worsley & Kidder 1991; Young 1991; Li *et al.* 2004). The argument here is that low latitudes enhance chemical weathering that, in turn, causes climatic cooling by drawing down atmospheric CO_2 levels. However, the relation between supercontinents and glaciations is not unique since glacial episodes have taken place during times of assembly, tenure and breakup/dispersal in various supercontinent cycles (Fig. 10). For example, glacial periods occurred during Pangaea's tenure in the Permian but also occurred in the Quaternary, long after Pangaea broke up. Likewise, major glaciation occurred the late Cryogenian, coincident with the tenure of Greater Gondwana, but also occurred in the Ordovician, long after its breakup. The Snowball Earth glaciations, which may have frozen the entire Earth's surface during the Cryogenian (e.g. Hoffman *et al.* 1998), coincided with the breakup of Rodinia (e.g. Li *et al.* 2008; Fig. 10). Other major glacial periods have also been tentatively linked to the supercontinent cycle, such as the Paleoproterozoic Snowball Earth (e.g. Cox *et al.* 2016; Gumsley *et al.* 2017). These glaciations apparently ended abruptly after volcanism raised atmospheric CO_2 levels to about 350 times the modern level (Hoffman 2009). This rapid rise would have resulted in a warming of the snowball Earth to extreme greenhouse conditions.

Supercontinent cycles and life evolution

Following the Cambrian explosion, there have been several episodes of mass extinction and subsequent biotic radiation. Mass extinctions are less well known in the Precambrian (Santosh 2010*a*, *b*; Retallack *et al.* 2014 and references therein). Mechanisms held to be responsible for mass extinction events include major climatic fluctuations and/or fluctuations in sea-level (Hallam & Wignall 1999), global anoxia (Pálfy & Smith 2000), volcanic eruptions (LIPs) (Isozaki 2009), magnetic reversals (Wei *et al.* 2014), asteroid impacts (Alvarez *et al.* 1980; Renne *et al.* 2013) and enhanced dosages of gamma rays (Santosh 2010*a*, *b* and references therein). Hence, different extinction events may be the result of different mechanisms or even several mechanisms operating in concert (Richards *et al.* 2015). Some of these mechanisms are directly or indirectly related to plate tectonics, whereas others are extraterrestrial in origin.

The assembly and dispersal of supercontinents could be responsible for several of these mechanisms, either direct or indirectly (e.g. Worsley *et al.* 1991). However, the frequency of extinction events, and the evidence in some cases for extraterrestrial and orbital causes complicates any direct correlation with supercontinents or LIPs. It has been argued that Pangaea (or the LIP associated with its formation) influenced some of these extinction events (e.g. Wignall 2001; Grasby *et al.* 2015; Bond & Grasby 2017; Ernst & Youbi 2017), especially those that occurred during its amalgamation (e.g. the Devonian and Permian extinctions) and dispersal (e.g. the Jurassic extinction). However, whether evolution and extinction are closely linked to solid Earth processes remains controversial.

No prospect of an end

The increasing size of geochronological and isotopic databases, the advent of quantitative plate reconstruction software, and the existence of increasingly sophisticated numerical modelling has provided substantial advances in our understanding of plate tectonics, mantle dynamics and the interaction of Earth's geosphere, atmosphere, hydrosphere and biosphere. Taken together, these data suggest that Earth's history has been punctuated by quasi-cyclical events. In many cases, the trends may correspond to the amalgamation and breakup of supercontinents. We suggest that a 'supercontinent' should be defined as a single continental plate with a size capable of modifying or controlling mantle dynamics and core–mantle boundary processes, altering convection cells and enhancing thermal activity. Thus, in our view, if a large continent comprising most of the continental crust of Earth has little or no effect on mantle dynamics, it should not be considered a supercontinent. Conversely, a smaller continent that nevertheless has a significant effect on mantle dynamics should be considered a supercontinent. In this way, we see the supercontinent cycle as a geodynamic process, which may be, in whole or part, the mechanism controlling long-term inner Earth dynamics.

Following our definition, Pangaea and Rodinia clearly qualify as supercontinents. Kinematic and geological constraints for both landmasses support their association with enhanced thermal activity linked to major changes in mantle dynamics. At our present state of knowledge, Pannotia/Greater Gondwana only questionably qualifies as a supercontinent; however, the same database cannot dismiss its future membership of the supercontinent club. The database for Columbia/Nuna and Archaean supercontinents is still insufficient to judge whether they fulfil the criteria to be supercontinents. Nevertheless, the available data for Columbia/Nuna certainly points in this direction.

Many geological and isotopic features show long-term secular trends that are quasi-cyclical and generally coincident with the maxima in orogenesis including, amongst others, zircon production maxima, Lu–Hf and O isotopes in zircon, the development of metallogenic provinces, and Sr and Nd isotopes in seawater. Indeed, many authors have suggested that these secular trends act as proxies for the supercontinent cycle, linking orogenic maxima (at *c.* 2800–2500, *c.* 1900–1700, *c.* 1200–1000, *c.* 600–500 and *c.* 430–250 Ma) with the amalgamation of supercontinents.

Amongst the many uncertainties, it is crucial to resolve whether the supercontinent cycle and mantle dynamics are strongly coupled and, if so, how. Geological evidence links the occurrence of LIPs with supercontinent breakup, and the eruption sites of LIPs and kimberlites seem to correspond with present-day plume locations, which generally occur at the edges of the LLSVPs at the core–mantle boundary. These features suggest a relationship between supercontinents and mantle dynamics.

The relationship between supercontinents, core dynamics and the magnetic field are still vague. True polar wander effects would be expected if the supercontinent cycle is linked to mantle dynamics, especially when supercontinents form far from the equator. However, Pangaea formed in an equatorial position, so the amount of TPW was minimal. Data for other supercontinents suggest major TPW effects; however, neither the kinematics of the supercontinents nor the palaeomagnetic database are well constrained. In contrast, the formation and disruption of Pangaea evoke a cause–consequence relation between superchrons and supercontinents since two superchrons are coincident with its amalgamation and breakup. In this case, it is the incomplete palaeomagnetic and magnetostratigraphic database for pre-Permian times that hinders interpretation of this relationship in pre-Pangaean supercontinents.

The suggested relationship between supercontinents, LIPs, climate change and the evolution of life are complex since several additional external factors control the climatic conditions on Earth's surface. Amongst these, astronomic forcing and extraterrestrial impacts are known to have an important control on climate and life. However, until these different factors are fully understood, the influence of the supercontinent cycle on climate and life is only a suggestive one and, consequently, difficult to assess.

New developments that include increasingly precise radiometric dating, new kinematic plate reconstructions using quantitative software like GPlates, advances in geodynamic modelling and seismic tomography are increasingly revealing Earth history to have been punctuated by quasi-cyclical episodes of supercontinent assembly and breakup, at least from the Paleoproterozoic. Amalgamation and breakup of supercontinents may indeed be responsible for: (1) temperature increases in the underlying mantle as a result of thermal insulation, leading to degree-one (single cell) mantle convection and supercontinent breakup; (2) the formation of large igneous provinces and superplumes through subduction avalanches to the core–mantle boundary, fuelling LLSVPs and producing TPW events; and (3) changes in core dynamics and the behaviour of the magnetic field. In turn, the development of LIPs may be related to dramatic climate change and ocean anoxia that could trigger mass extinctions. Hence, a complete understanding of the supercontinent cycle and its relationship to mantle dynamics and Earth's history remains one of the most important challenges facing Earth science today.

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