Four-dimensional context of Earth's supercontinents

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Abstract: The supercontinent-cycle hypothesis attributes planetary-scale episodic tectonic events to an intrinsic self-organizing mode of mantle convection, governed by the buoyancy of continental lithosphere that resists subduction during closure of old ocean basins, and consequent reorganization of mantle convection cells leading to opening of new ocean basins. Characteristic timescales of the cycle are typically 500–700 myr. Proposed spatial patterns of cyclicity range from hemispheric (introversion) to antipodal (extroversion), to precisely between those end-members (orthoversion). Advances in our understanding can arise from theoretical or numerical modelling, primary data acquisition relevant to continental reconstructions, and spatiotemporal correlations between plate kinematics, geodynamic events and palaeoenvironmental history. The palaeogeographic record of supercontinental tectonics on Earth is still under development. The contributions in this special publication provide snap-shots in time of these investigations and indicate that Earth's palaeogeographic record incorporates elements of all three endmember spatial patterns.

Introduction: how do we recognize ancient supercontinents?

Alfred Wegener (1912, 1929) considered Pangaea to be the *Urkontinent* or 'ancestral continent'. Despite his remarkable insights into other aspects of Earth dynamics, he always referred to his Pangaea configuration as the 'origin of the continents'. Available knowledge at the time limited the understanding of possible pre-Pangaean oceans and landmasses (aside from anecdotal references by van der Gracht, see Holmes 1929), but decades later Wilson (1966) elucidated them on the basis of earliest Palaeozoic biogeographic provinces. Geochronological quantification of the vastness of Precambrian time through the twentieth century also began to allow us, in principle, to explore the rich history of pre-Pangaea continental motions and interactions (Fig. 1). As the first global compendia of Precambrian isotopic ages were assembled (Gastil 1960), temporal peaks and troughs of age frequencies emerged, conveying a *c.* 500 myr episodicity in global tectonic processes. Following the plate-tectonic revolution of the 1960s that provided a new uniformitarian paradigm for continuous (rather than convulsive) evolution of global tectonics,

> many [geoscientists] expected that the long-term episodicity displayed by mineral age frequencies would degrade as the data base expanded and its reliability improved. It did not. (Hoffman 1989)

Continental lithosphere is buoyant, aided by not only a silica-rich crust but also a magnesian lithospheric keel (Lee *et al.* 2011). Some continental lithosphere can be subducted to depths of hundreds of kilometres (van Roermund 2009), but continents generally resist subduction owing to their buoyancy. Assemblies of large landmasses via continental collisions are thus an inevitable consequence of plate tectonics, and impart a stochastic component to supercontinent amalgamation. After assembly over subduction downwellings, however, why should supercontinents then break apart? Holmes (1929) suggested that thermal insulation by radiogenic continental crust would cause a reversal of mantle convection cell geometry, a concept that was much later linked to post-Pangaea geophysical legacies (Anderson 1982) and verified by simple numerical models (Gurnis 1988). The models, now quite sophisticated, contain two physical consequences of supercontinental assembly that may invoke subsequent fragmentation: the aforementioned thermal insulation (Phillips & Coltice 2010) and global reorganization of the bulk planform of mantle convection (Zhong *et al.* 2007).

At no time did Pangaea contain all extant continental lithosphere – the Chinese blocks, in particular, were excluded from the Pangaea landmass for most if not all of its tenure (Torsvik & Cocks 2004; Stampfli *et al.* 2013; Fig. 1). Definition of a supercontinent cannot, therefore, require 100% of

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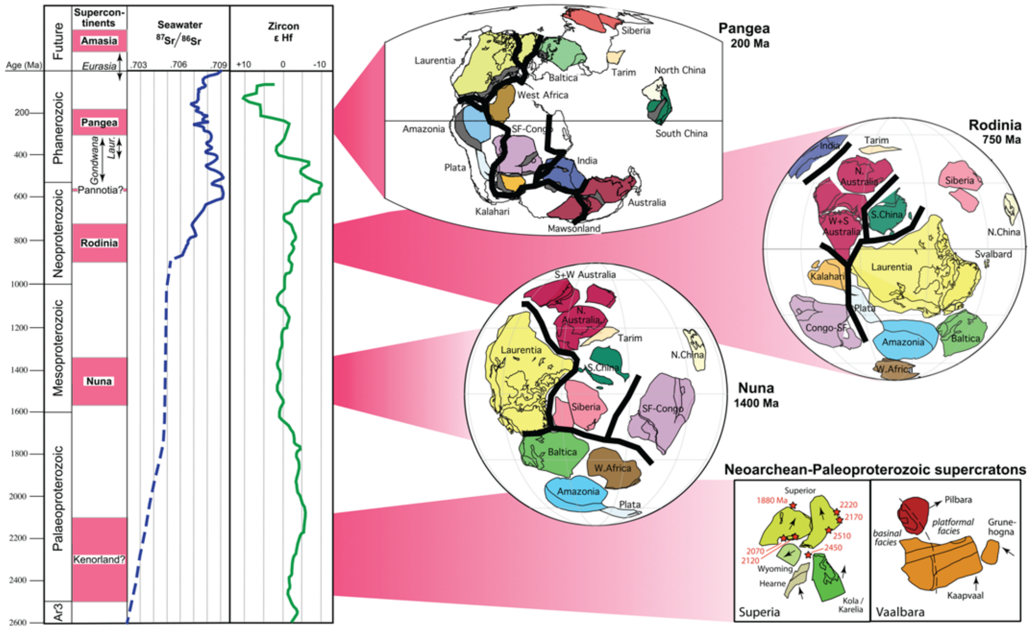

Fig. 1. Graphical summary of Earth's supercontinents, adapted from elements in Evans (2013). Abbreviations: Ar3, Neoarchean; Laur., Laurussia. Semi-supercontinents are named in italics. Seawater Sr-isotope curves from Halverson *et al.* (2007) and Shields (2007). Hf-isotopic curve (note that negative values are to the right, to facilitate direct comparison with the Sr-isotope curve) from Spencer *et al.* (2013). Reconstructions show supercontinents at the ages of incipient breakup (rifts drawn as heavy black curves): Pangaea at 200 Ma in present South American reference frame, Rodinia at 750 Ma in the palaeomagnetic reference frame (Li *et al.* 2013) and Nuna at 1400 Ma in the palaeomagnetic reference frame (Pehrsson *et al.* 2015). Supercratons Superia (Ernst & Bleeker 2010) and Vaalbara (de Kock *et al.* 2009) are shown with arrows indicating present local north for each constituent block.

continental area. How large must a landmass be, to be considered 'supercontinental'? Meert (2012) suggested a threshold value of 75% of extant continental area from any given age, a definition that is simple, unambiguous and includes Pangea (85–90%), but excludes large landmasses like Gondwana (*c.* 60%) and Eurasia (35%). The former entity endured for about 400 myr and may have had profound influences on both geodynamics and palaeoenvironments, but as a hierarchical subset of Pangaea, it should not share the title, 'supercontinent'. The latter entity, Eurasia, is also enormous in size and no doubt influential on broader Earth-system dynamics, but it is still growing *en route* to the hypothetical future supercontinental assemblage of Amasia (Hoffman 1997). We suggest the term 'semi-supercontinent' for the hierarchical level attained by Gondwana, Eurasia and similarly large and long-lived landmasses that were subsets of true supercontinents as defined herein. *Note that for many years, some of the authors of this introduction referred to 'Gondwanaland' or 'Gondwana-Land' as the large continent recognized by the characteristic Gondwana volcano-sedimentary sequence across India and the present southern continents (eliminating ambiguity and honouring the naming priority of E. Suess; Sengör 1983), but usage of the shorthand version in this paper is inspired by a recent personal communication to the first author by Paul Hoffman: 'If Gondwana [referring to the semi-supercontinent] was good enough for Alex Du Toit, it should be good enough for us'.*

Following conceptual models of varying erosional preservation of continental crust through collision and rifting (Hawkesworth *et al.* 2010), Bradley (2011) noted that durations of supercontinental aggregation correspond to global *minima* of zircon distributions that can be used to identify pre-Pangaea supercontinents. Proceeding backwards in time, such minima occur at *c.* 0.3 Ga (Pangaea), 0.9 Ga (Rodinia), 1.5 Ga (Nuna) and 2.3 Ga (Kenorland?). The oldest of these intervals has been suggested as a time of global tectonic shutdown or quiescence (Condie *et al.* 2009), although tectonomagmatic activity is being increasingly recognized from that age among a broad sampling of cratons (Pehrsson *et al.* 2013). Four putative supercontinent cycles (Fig. 1) are discussed in turn, below.

Contributions to this volume: a history of Earth's supercontinents

First cycle: Kenorland(?) to Nuna

The hypothetical supercontinent at the Archean–Palaeoproterozoic transition has been given the name Kenorland, after the 2.7 Ga Kenoran orogeny consolidating the Superior craton in North America (Williams *et al.* 1991). If a Superior-centred view of the putative supercontinent may be exported to other cratons, then development of the rifted Huronian margin at 2.45 Ga (Ketchum *et al.* 2013) would correspond to the age of Kenorland's initial fragmentation. However, a subsequent definition of Kenorland entails assembly as late as *c.* 2.5 Ga, with breakup at *c.* 2.1 Ga (Barley *et al.* 2005). With such ambiguous defining features, mere existence of the alleged supercontinent is difficult to test, let alone its cratonic configuration. The alternative to a united Kenorland supercontinent is a palaeogeographic model of distinct, freely drifting, continent-sized supercratons (Fig. 1), including Superia, Sclavia and Vaalbara – each containing several modern cratons with characteristic ages of amalgamation (*c.* 2.7, 2.6 and 2.9 Ga, respectively; Bleeker 2003). Vaalbara has begun to take form in a reconstruction guided by both stratigraphic and palaeomagnetic data (de Kock *et al.* 2009). Sclavia may include the Dharwar craton of southern India (French & Heaman 2010). Superia has long been considered to include several North American and Fennoscandian cratons (Ernst & Bleeker 2010), including Wyoming along Superior's southern margin (Roscoe & Card 1993). **Kilian *et al.* (2015)** provide an expanded palaeomagnetic database for mafic dyke swarms in Wyoming that generally conform to that reconstruction of Wyoming against southern Superior, in a palaeogeographically inverted orientation. Endurance of the Superior–Wyoming connection as late as 2.16 Ga would correspond better to the younger age-definition of Kenorland (cf. Barley *et al.* 2005), if such an entity existed.

As either Kenorland or separate Archean–Palaeoproterozoic supercratons fragmented during early Palaeoproterozoic time, orogenesis continued among another 'clans' of cratons, those bearing the record of *c.* 2.3 Ga tectonism (Pehrsson *et al.* 2013). Convergence from 2.1 to 1.9 Ga can be found among many of the world's cratons (Zhao *et al.* 2002), leading towards the eventual consolidation of the supercontinent Nuna (Hoffman 1997). Nuna is named after the North American orogenic record of craton amalgamation, representative of a global orogenic peak within the aptly named 'Orosirian' Period of Earth history (Plumb & James 1986; Plumb 1991), and its name precedes that of synonymous 'Columbia' (Rogers & Santosh 2002) by half a decade, thus 'Nuna' is retained herein (see Meert 2012, and Evans 2013, for discussion). **Betts *et al.* (2015)** provide a detailed description of Australian terranes assembling with each other and with the western margin of newly consolidated Laurentia, as part of the core of Nuna supercontinent. **Pehrsson *et al.* (2015)** present the first compendium of cratonic terranes and orogens in a palaeogeographic framework that is consistent with both global tectonostratigraphic databases (Eglington *et al.* 2009), including mineral deposits, and palaeomagnetic data. The approach of Pehrsson *et al.* is to split modern cratons into as many indivisible elements as possible, so that polygons of younger terranes can be sequentially added moving forward in geological time, as dictated by the geochronological databases. The method is distinct from many previous models that consider modern cratonic areas as whole entities moving backward in time, until splits into smaller terranes are necessitated by discordant palaeomagnetic data (Pisarevsky *et al.* 2014). Despite the complementary nature of these two palaeogeographic strategies, the overall similarity of many aspects of their resulting models (Table 1), perhaps lends a degree of confidence to their veracity.

Second cycle: Nuna to Rodinia

Recent success in deciphering Nuna's palaeogeography stems mainly from integrated U–Pb geochronology and palaeomagnetism on Palaeo-Mesoproterozoic large igneous provinces (Buchan 2013). At the core of Nuna reconstructions is the NENA (Northern Europe–North America) juxtaposition of Baltica in an 'upside-down' orientation with its Russian margin adjacent to the east-Greenland margin of Laurentia (Gower *et al.* 1990; subsequent references reviewed by Evans 2013). **Salminen *et al.* (2015)** add to the high-quality palaeomagnetic pole database of Baltica with their study of *c.* 1570 Ma mafic dykes in the Åland archipelago of the Baltic Sea, supporting the NENA fit (Fig. 1). Beyond Baltica, palaeomagnetically quantified Nuna reconstructions in recent years have incorporated most of the world's Mesoproterozoic cratons (Table 1), but data from African blocks have been particularly poorly constrained through that time interval.

The Kalahari craton in southern Africa contains a Palaeoproterozoic marginal foldbelt (Moen 1999) and subsequent passive margin (Van Niekerk 2006),

Table 1. *Euler parameters (total reconstruction poles to Laurentia) for various Nuna supercontinent reconstructions. Highlighted craton names indicate substantial differences among the models*

Craton	Zhang *et al.* (2012)	Pisarevsky *et al.* (2014)	Xu *et al.* (2014)	Pehrsson *et al.* (2015)
Baltica	45.0, 007.5, 44.9	45.0, 007.5, 44.9	47.5, 001.5, 49.0	47.5, 001.5, 49.0
Siberia–Anabar	78.0, 099.0, 147.0	70.0, 133.2, 127.0	(Same as Aldan)	76.8, 106.2, 146.6
Siberia–Aldan	78.0, 118.3, 170.7	(Same as Anabar)	78.0, 099.0, 147.0	76.2, 121.6, 170.6
West Africa	30.0, 266.0, 70.7	Independent block	*Not illustrated*	58.3, 304.9, 85.8
Amazon	53.0, 293.0, 127.0	Independent block	58.0, 290.9, 138.0	53.9, 291.1, 122.2
Plata	*Not illustrated*	*Not illustrated*	*Not illustrated*	(Same as Amazon)
Congo	*Not illustrated*	−16.9, 236.3, 135.3	*Not illustrated*	35.1, 287.2, 103.8
Kalahari	*Not illustrated*	Independent block	*Not illustrated*	Independent block
Australia (N)	31.5, 098.0, 102.5	35.0, 103.3, 98.5	34.6, 093.5, 103.2	45.5, 089.2, 100.1
Australia (S)	29.0, 125.8, 133.0	29.9, 128.1, 137.5	32.7, 124.2, 130.5	27.7, 150.5, 137.5
Australia (W)	29.7, 121.2, 124.6	33.6, 108.0, 118.9	33.4, 119.0, 122.6	41.9, 117.1, 110.3
Tarim	*Not illustrated*	*Not illustrated*	*Not illustrated*	74.0, 117.9, 141.9
North China	45.7, 330.8, 33.4	31.1, 331.9, 40.5	45.0, 002.0, 34.0	36.5, 271.2, 34.1
Yangtze	*Not illustrated*	*Not illustrated*	*Not illustrated*	46.5, 137.5, 196.4
Cathaysia	*Qualitative illustration*	65.3, 159.9, 157.7	*Not illustrated*	48.3, 138.5, 196.6
India (N)	56.0, 075.4, 123.0	42.4, 064.8, −150.3	67.0, 103.0, 107.6	Independent block
India (S)	(Same as India-N)	(Same as India-N)	(Same as India-N)	Independent block
Nuna duration	*c.* 1780–1400 Ma	*c.* 1500–1400 Ma	*c.* 1780–1380 Ma	*c.* 1540–1420 Ma

thus rendering likely its incorporation into, and separation from, the Nuna landmass. Kalahari and neighbouring terranes also play a significant role in Rodinia reconstructions (e.g. Loewy *et al.* 2011). Nonetheless, despite a robust palaeomagnetic pole from the Umkondo large igneous province at *c.* 1110 Ma (Gose *et al.* 2006), data from preceding periods are of a preliminary nature at best (Gose *et al.* 2013). The Sinclair region along the western margin of Kalahari, in south-central Namibia, offers an opportunity for substantial augmentation of the Mesoproterozoic palaeomagnetic database from southern Africa. Two papers herein report new palaeomagnetic results of high quality from the youngest components of Sinclair stratigraphy: **Panzik *et al.* (2015)** produce a pole that is precisely dated by U–Pb on zircons at 1105 Ma, and **Kasbohm *et al.* (2015)** generate a pole from unconformably overlying redbeds, constrained in age by detrital zircons and comparison to the Kalahari apparent polar wander path. Both results conform to that Umkondo-anchored apparent polar wander path and point the way toward studies of older Sinclair strata in extending that path backward in time to envelop the Nuna–Rodinia transition. More integrated geochronological–palaeomagnetic studies such as these, from the world's Mesoproterozoic cratons, promise to yield great advances in our understanding of that supercontinental transition in the immediate future.

Third cycle: Rodinia to Pannotia/Gondwana

The effort to reconstruct Rodinia experienced some degree of tribulation in the early 2000s, when several cratons were excluded from the landmass by various assessments: India (Pisarevsky *et al.* 2003), Congo-São Francisco, Plata and Kalahari (Kröner & Cordani 2003). A radical revision to its palaeogeography was proposed (Evans 2009). All the while, Piper (2000, 2007) continued to espouse his alternative 'Palaeopangaea' model, largely unchanged since its original inception in the 1970s but beginning to incorporate modest mobility to accommodate discordant palaeomagnetic data (Piper 2010). Exclusion of Gondwanan cratons was partly based on a consistently high-palaeolatitude location of Congo–São Francisco during the interval 1080–1010 Ma, in contrast to the low palaeolatitudes of cratons bordering Laurentia through the same period (Cordani *et al.* 2003). **Evans *et al.* (2015)** show that the mafic dykes of São Francisco craton, yielding those high-palaeolatitude poles that were previously dated to within geon 10 by Ar/Ar methods, are actually *c.* 920 Ma as constrained by concordant U–Pb analyses on baddeleyite. The new ages and reproducible palaeomagnetic data reported in that study allow for a 'return to Rodinia' of that craton.

All-inclusive Rodinia models are regaining popularity (Li *et al.* 2008) and being incorporated into holistic palaeogeographic models of tectonics and stratigraphy (Li *et al.* 2013). Remaining items of debate include whether South China formed a 'missing link' between Australia and Laurentia (Li *et al.* 1995, Fig. 1) or whether it lay elsewhere, permitting a direct SWEAT connection between the two larger landmasses (Goodge *et al.* 2008, 2010; Dalziel 2010; Loewy *et al.* 2011). This issue is central to the timing of Rodinia amalgamation, because the Sibao orogen in South China is as young as *c.* 900 Ma; if it represents a Rodinia-forming suture, then the supercontinent only existed for about 150 myr prior to initial stages of breakup. As Li & Zhong (2009) noted, such brevity of supercontinental existence and the central location of final suturing would have remarkable parallels to the Palaeozoic amalgamation of Pangaea.

One of the challenges presented by a short-lived Rodinia model, according to the 'missing-link' hypothesis, is how to constrain the motion of the supercontinent by palaeomagnetism, when globally available rocks of sufficiently low metamorphic grade are few in that relatively brief time span. Compounding this problem is the possible occurrence of moderate-magnitude true polar wander (TPW) events at about 800 Ma, coincident with the onset of continental rifting leading to Rodinia breakup. A single TPW event was hypothesized by Li *et al.* (2004) on the basis of new palaeomagnetic data from South China; the hypothesis was expanded to an oscillation of TPW episodes by Maloof *et al.* (2006) based on data from Svalbard terranes reconstructing to Laurentia's NE margin. That model encompasses a pair of carbon-isotope anomalies and sequence-stratigraphic boundaries in shallow marine strata of several cratons, referred to as the 'Bitter Springs Event' (Swanson-Hysell *et al.* 2012). Oscillations of TPW are to be expected in a world dominated by mobile subduction zones confined within a 'ring-of-fire'-like girdle surrounding a stable equatorial axis of large low shearwave velocity provinces (LLSVPs) as recognized in the post-Pangaea world (Richards & Engebretson 1992; Steinberger & Torsvik 2008) and hypothesized for Precambrian supercontinental cycles (Evans 1998, 2003; Mitchell *et al.* 2012). TPW oscillations also may be expected if the lithosphere is stiff enough to retain a 'memory' of previous stresses (Creveling *et al.* 2012). Nonetheless, testing of the 'Bitter Springs Event' TPW hypothesis requires palaeomagnetic data from additional cratons at *c.* 800 Ma. **Niu *et al.* (2015)** provide new palaeomagnetic and geochronological data from the western margin of South China to confirm a large discordance between palaeomagnetic poles of 824 and 806 Ma. Whether TPW can be invoked to explain this discrepancy depends on the location of South China in Rodinia. This issue, of major geodynamic significance, invites integrative studies of high stratigraphic resolution through this time interval from other cratonic regions (e.g. Swanson-Hysell *et al.* 2015).

Identifying the breakup configuration of Rodinia will depend in large part on understanding the rifted margins of Laurentia, which developed between 800 and 500 Ma. Along all three margins of the quasi-triangular Laurentian landmass, rift-related magmatism occurred in geon 7, but thermal subsidence did not develop until Cambrian time (Bond *et al.* 1984). As noted above, the western margin of Laurentia – whether adjacent to Australia–Antarctica or South China – is of central importance to Rodinia's breakup history. Several recent stratigraphic studies and syntheses have contributed to refined correlations across the southwestern Laurentian margin through the interval of its development (Macdonald *et al.* 2013; Mahon *et al.* 2014; Yonkee *et al.* 2014). **Smith *et al.* (2015)** present integrated stratigraphy and detrital-zircon geochronology to demonstrate that the pre-Cryogenian Beck Springs Dolomite and adjacent units were not deposited on a broad, coherent continental margin as had been characterized by some previous workers.

Following Rodinia's breakup, orogenic activity in the Late Neoproterozoic and Early Cambrian was dominated by the amalgamation of Gondwana. East Gondwana was assembled in at least two stages, between 750 and 620 Ma (East African Orogen) and at 570–500 Ma (Kuunga orogen; Meert 2003; Collins & Pisarevsky 2005). West Gondwana, comprising the cratons of South America and Africa, was amalgamated by numerous continental collisions between 650 and 600 Ma (e.g. Pankhurst *et al.* 2008) although some orogenesis continued into the Cambrian (e.g. Schmitt *et al.* 2008; Tohver *et al.* 2010). Such tectonism may have been Mediterranean in style, that is, occupying confined basins following the major continental collisions (Cordani *et al.* 2013), or it may have involved final collision of Amazon, closing a Clymene Ocean tract that remained wide in late Ediacaran time (Tohver & Trindade 2014). Mafic complexes within these collisional orogenic belts are characterized by crust-formation ages (*c.* 1.2–0.75 Ga) that predate the ≤750 Ma breakup of Rodinia. Hence, these complexes must have formed from 'exterior' oceanic lithosphere that surrounded Rodinia, implying that this ocean was preferentially consumed, implying Gondwana amalgamated in a style similar to extroversion (Murphy & Nance 2003).

With each successive collision, the locus of subduction moved from locations between the converging blocks of Gondwana, to the periphery of that landmass (Murphy & Nance 1989). This process generated the peripheral orogens of the peri-Gondwanan terranes, now located in eastern Laurentia and western Europe (e.g. Avalonia, Meguma, Carolinia, Ganderia, Cadomia) along the Amazonian–West African margin of West Gondwana, as well as the oldest components of the Terra Australis Orogen, an 18 000 km-long belt that includes Palaeozoic accretionary orogens in eastern Australia, New Zealand, Antarctica, southern Africa and the South American Andes (Cawood & Buchan 2007).

Most reconstructions show a connection between Amazonia, West Africa and Laurentia in the Neoproterozoic, leading to the hypothesis that the collision of East Gondwana with West Gondwana–Laurentia produced the supercontinent Pannotia (Powell 1995; Dalziel 1997). However, if Amazonia (perhaps still connected to Laurentia) did not collide with the remainder of West Gondwana until Cambrian time, then the so-called Clymene Ocean (Tohver & Trindade 2014) would intervene between South American blocks, precluding Pannotia's existence. The possibility of Cambrian suturing within eastern Gondwana (Boger *et al.* 2001) also casts some doubt on the concept of Pannotia.

Although rift-related magmatism and sedimentation commenced along Laurentia's eastern margin at *c.* 750 Ma, passive-margin subsidence did not develop until Ediacaran–Cambrian time (Cawood *et al.* 2001), ultimately opening the Iapetus Ocean. This event has been interpreted variously, as 'the final breakup of Rodinia' (e.g. Li *et al.* 2008; Bradley 2011), as the breakup of Pannotia (Dalziel 1997) or as merely the separation of a ribbon-like continental fragment akin to either the Lomonosov Ridge or the Lord Howe Rise (Cawood *et al.* 2001; Evans 2009). In either case, if Pannotia existed at all, then the temporal overlap of this rifting event with Gondwana-assembling orogenic events suggests that such existence would have been very brief, certainly not long enough for the mantle thermal insulation model to explain its breakup. Interaction between Laurentia's eastern margin, first with oceanic-arc and then with continental terranes, characterized the Late Cambrian to Devonian timespan, culminating in terminal collision with West Africa in the Late Carboniferous to amalgamate Pangaea (van Staal *et al.* 2009; Pollock *et al.* 2012).

Palaeomagnetic records of the Rodinia to Gondwana transition are unusually complex, even more so than some preceding intervals. Within each craton, there is a dispersion of palaeopoles that far exceeds typical variation through younger geological periods (Torsvik *et al.* 2012). Proposed explanations include rapid plate tectonics (Meert *et al.* 1993; Gurnis & Torsvik 1994), rapid TPW (Evans 1998, 2003; Raub *et al.* 2007) or unusual geomagnetic field configurations (Abrajevitch & Van der Voo 2010). A trivial solution, that is, bias by rock-magnetic recording effects and incomplete separation of magnetic components, may help explain some anomalous datasets (Bono & Tarduno 2015), but that explanation does not easily account for other datasets from other mid-Ediacaran mafic suites (Halls *et al.* 2015) and redbeds (Schmidt 2014). If any of the geophysical explanations is valid singularly, or in combination, then novel approaches will be needed to address terminal Neoproterozoic palaeogeography. Palaeoclimate indicators in the sedimentary record, such as carbonates and evaporites (Evans 2006), may provide some assistance, but the 'strange-bedfellow' paradox of those facies interbedded with Neoproterozoic glacial deposits (Hoffman 2011) prompts caution in their universal application to palaeogeography.

Fourth cycle: Gondwana/Pangaea to Amasia

Subduction along both the Laurentian and Baltican flanks of the Iapetus Ocean began in the Late Cambrian–Early Ordovician and culminated in arc–continent collisions and obduction of supra-subduction zone ophiolite complexes (Taconic and Penobscot orogenies, van Staal *et al.* 1998, 2009). At about the same time, the Rheic Ocean opened by separation of some peri-Gondwanan terranes (Ganderia, Avalonia, Meguma) from the Amazonian–West African margin of Gondwana (Cocks & Fortey 1990; Stampfli & Borel 2002). This opening is variously attributed to slab pull related to subduction in the Iapetus Ocean (Murphy *et al.* 2006; Nance *et al.* 2012) or to roll-back of a subduction zone beneath the Gondwanan margin (van Staal *et al.* 2009, 2012). During the Ordovician, these terranes intervened between the Iapetus Ocean (to the north) and the Rheic Ocean (to the south). The Iapetus Ocean was closed in the Early Devonian by the collision of Avalonia with Baltica, and then Avalonia–Baltica with Laurentia (thereby forming Laurussia; van Staal *et al.* 2009). The Rheic Ocean expanded as the Iapetus Ocean contracted, attaining a maximum width in the Early Devonian, followed by contraction until the Mississippian, which culminated in the collision of Gondwana with Laurussia to form the vast Ouachita–Appalachian–Variscan orogen that became stranded in the interior of Pangaea (Hatcher 2002; Murphy & Nance 2008). In contrast to the collisional orogenic belts associated with the amalgamation of Gondwana, Sm–Nd crust-formation ages of accretionary mafic complexes in the Iapetus and Rheic Oceans postdate their formation, suggesting Pangaean reassembly by introversion (Murphy & Nance 2003).

Murphy *et al.* (2015) point out that much of the tectonic evolution of the Iberian portion of the suture zone postdates the onset of collisional tectonics elsewhere in the Variscan orogen, and that its evolution was dominated by subduction of varying polarity and longevity in relatively narrow tracts of oceanic lithosphere. This scenario may be analogous to the Cenozoic evolution of the eastern Mediterranean oceanic tracts relative to the ongoing collision between the African, Eurasian and Arabian plates. Such tectonic features, which have a high preservation potential, may be more common in collisional belts than is currently appreciated. If their features are not recognized, they may be misinterpreted to reflect the main collisional event, thereby yielding a misleading age for the timing of continent–continent collision.

Understanding the forces that led to the breakup of Pangaea has first-order implications for understanding the breakup of older supercontinents. Buiter & Torsvik (2014) find that most of the conjugate margins within the Atlantic Ocean are developed concurrently with emplacement of the large igneous province and rifting began before the main phase of volcanism. They interpret these associations to indicate that rifting was initiated by tectonic forces and that magmatism was associated with asthenospheric upwelling from plumes that exploited the thinned lithosphere and helped to trigger final continental drift. Exactly what roles plate dynamics and mantle plumes/superplumes play in supercontinent breakup (e.g. Storey 1995) will continue to be a topic of debate in the coming decades, and will benefit from our better understanding of supercontinent and plume histories and 4D geodynamic modelling.

The Mesozoic dispersal of Pangaea with the opening of the Atlantic Ocean is thought to have been balanced by either (a) the convergence and subduction along the flanks of the Panthalassan Ocean or (b) convergence and closure of the palaeo-Tethys and Tethys oceans linked to subduction off the southern margins of Eurasia (e.g. Collins 2003). The latter scenario is attributed to the forces causing Pangaea breakup being transmitted from the Tethyan domain into the Atlantic domain by slab sinking forces associated with the Tethyan subduction zones (Keppie 2015a). **Keppie (2015b)** proposes that Pangaea moved along a non-rigid path in a mantle reference frame at the time of breakup and that subduction along the margins of Pangaea can explain both the motion and deformation of Pangaea whereas mantle forces applied to the base of Pangaea cannot. Keppie concludes that top-down tectonics caused the breakup of Pangaea.

Discussion: styles of supercontinental tectonics

Knowledge about how the supercontinent cycles interact with the deep mantle is fundamental for understanding how the Earth's tectonic engine works. Chase (1979) first recognized the presence of broad residual geoid highs (superswells) in both the present-day southwestern Pacific and Atlantic–Africa region. Anderson (1982) correlated the African superswell with an area previously occupied by the Pangaea supercontinent in a fixed hotspot reference frame, and ascribed both superswells as the consequence of thermal isolation in the upper mantle by former continental aggregations. Larson (1991) first used the term 'superplume' to define a lower mantle thermal anomaly that drove the formation of a cluster of hotspots in the Pacific realm. Such a speculated lower mantle origin of large thermal anomalies that drive most of the hotspots is supported by the seismic tomographic discovery of two antipodal and equatorial LLSVPs (Dziewonski 1984; Romanowicz 2008). The two LLSVPs correspond to the two geoid superswells of Chase (1979), and are now often referred to as the African and Pacific superplumes.

The linkage between plume events and the LLSVPs/superplumes appears to have been widely accepted for the post-Palaeozoic time (Burke & Torsvik 2004), and geological observations have suggested temporal connections between episodic intense LIP occurrences (superplumes?) and supercontinents (Bleeker 2003; Li *et al.* 2003). The long-term stability of such first-order mantle features, and the question of whether they were connected to the supercontinent cycle (and if yes, then how) are currently topics of intense debate amongst the supercontinent and geodynamic communities. There are three major hypotheses regarding the formation and longevity of superplumes. One is to link superplumes with supercontinent formation through thermal insulation of the supercontinent continental lithosphere (Anderson 1982). However, this hypothesis suffers from the presence of the LLSVPs in the lower mantle instead of the upper mantle, and the lack of supercontinental insulation for the formation of the Pacific superplume that peaked during Cretaceous time (Larson 1991). Another view is that the present-day equatorial and antipodal superplume structure in the lower mantle is a long-lived feature from early Earth history (Dziewonski *et al.* 2010; Burke 2011), thus precluding any connection with the supercontinent cycle. A third model argues for the dynamic coupling of the supercontinent cycle and the formation of the antipodal upwelling zones or superplumes amid a circum-supercontinent subduction system (Evans 2003; Li *et al.* 2004; Li *et al.* 2008; Li & Zhong 2009). This model can best account for major TPW events during supercontinent cycles (Evans 1998; Li *et al.* 2004), and the mechanism is similar to what was first speculated by Maruyama (1994). A variety of this model (Zhong *et al.* 2007) incorporates a cyclic alternation of mantle convection planforms through the supercontinent cycle: between degree-1 mantle convection, where there is a major upwelling system in one hemisphere and a major downwelling system in the other hemisphere, and degree-2 convection, where there are two major, antipodal upwelling systems as are present today. There are a number of possible tests of these models. For instance, if we could prove that supercontinent formation is always followed by the formation of a superplume beneath the supercontinent even if the supercontinents are located far from the equator, it would disprove the long-lived static superplume model. Another way of testing these models is to see if palaeomagnetic data support the various types of TPW events as predicted by these models (Evans 1998; Li *et al.* 2004; Maloof *et al.* 2006; Niu *et al.* 2015).

Geochemical records may also shed light on styles of supercontinent transition. The very high $^{87}Sr/^{86}Sr$ initial ratios of seawater during Late Neoproterozoic–Early Cambrian time (Veizer *et al.* 1999) have been related to enhanced continental weathering during following Gondwanan collisions, and a sharp negative excursion in the εHf (zircon) data suggests that the weathered material was predominantly recycled ancient continental crust (Belousova *et al.* 2010; Collins *et al.* 2011). Conversely, development of the Iapetus Ocean is characterized by decreasing initial Sr and increasing εHf (zircon), consistent with enhanced ocean ridge activity, and a sharp rise in sea-level (e.g. Miller *et al.* 2005), consistent with models involving continental subsidence, the formation of youthful (and more elevated) ocean floor, and progressively less relative contribution from continental weathering.

It is interesting to note that the amalgamation of Rodinia is not clearly evident in these proxy geochemical records; initial Sr values started to decrease from *c.* 1.8 to 0.75 Ga, whereas εHf (zircon) values are close to CHUR from *c.* 1.6–0.7 Ga with the exception of modest, *c.* 20–50 myr excursions. These data suggest that relatively juvenile crust was preferentially recycled during Rodinia amalgamation (Spencer *et al.* 2013). Such contrasting signatures are interpreted to reflect ocean closure by way of single-sided (Gondwana) v. two-sided (Rodinia) subduction zones. The former leads to collision between a passive margin and juvenile continental arc, and the latter to collision between crust generated by two juvenile arcs.

Kinematically, supercontinental transitions can be described using the terms extroversion, introversion and orthoversion. The first was proposed by Hartnady (1991), where 'extraversion' referred to the 'inside-out' model of Rodinia-to-Gondwana evolution of Hoffman's (1991) global Neoproterozoic tectonic synthesis: a new supercontinent assembles over the demise of the previous supercontinent-encircling global ocean. In contrast, 'introversion' refers to strict interpretations of the classic Wilson cycle, whereby relatively young, Atlantic-style ocean basins close to produce a new supercontinent in the same location, and with generally the same configuration, as the previous landmass. These contrasting endmember models were more recently expounded upon by Murphy & Nance (2003, 2005), who introduced possible tests using the Sm–Nd record of volcanic rocks within terranes caught up in the larger continental collisions. Introversion represents a geodynamic conundrum (Murphy & Nance 2008), because one would not expect relatively young oceanic lithosphere to subduct preferentially owing to its buoyancy. The conundrum may be answered in part by invasion of subduction zones from the exterior ocean into the young oceanic realms, as is happening today in the Caribbean and Southern Oceans, and may have happened within Iapetus during Palaeozoic time (Waldron *et al.* 2014). The third style of supercontinental transition appeals to the palaeomagnetic record of putative TPW oscillations using long-lived continental reference frames: Mitchell *et al.* (2012) proposed 'orthoversion' as an explanation for rapid shifts in the locations of the axes of those TPW oscillations as recorded in mid-Neoproterozoic Laurentia and late Palaeozoic Gondwana. In the orthoversion model, new pairs of LLSVPs are created upon each supercontinental assembly, located in the mantle precisely 90° away from the previous LLSVPs. Although the next future supercontinent's configuration cannot be tested realistically (one would need to wait about 100 myr or more for its final creation), we can imagine the future closure of present-day oceans as a thought experiment for contextualizing the three alternative transitional scenarios: extroversion closes the Pacific, introversion closes the Atlantic, and orthoversion closes the Arctic.

In such a tripartite context, we can summarize the four aforementioned palaeogeographic–kinematic cycles in the framework of the end-member geodynamic models. The Kenorland to Nuna transition is entirely unconstrained because it is still unknown whether Kenorland was a true supercontinent. The Nuna to Rodinia transition appears to include some elements of introversion, for example, the long-lived or repeated connections between proto-Australia, Laurentia,and Siberia (compare the Nuna configurations of Pisarevsky *et al.* 2014, and Pehrsson *et al.* 2015, with the Rodinia model of Li *et al.* 2008). Meert (2014) noted that many of the same 'strange attractor' cratonic connections appear and reappear between one proposed supercontinent and the next, which may be due to psychological bias or perhaps the introversion (accordion-like) phenomenon. According to popular Nuna and Rodinia models, the intervening transition also includes elements of extroversion; for example, the rotation of Baltica and perhaps Amazonia closed the long-lived accretionary margin that now extends from the Grenville Province, through the Sveconorwegian belt, into the Rondonia–Sunsas orogen (Johansson 2009, 2014). Nonetheless, if the Pisarevsky *et al.* (2014) model of Nuna is correct, then Amazonia (and West Africa) were not included in the Nuna landmass, and such closure of long-lived accretionary margins would be not be directly related to the supercontinental cycle. High-quality Mesoproterozoic palaeomagnetic data from South American and African cratons are acutely needed to distinguish among these alternatives. Following Hoffman's (1991) global Rodinia model and most subsequent variations (e.g. Dalziel 1997; Pisarevsky *et al.* 2003; Li *et al.* 2008, 2013), the Rodinia–Gondwana transition bears mainly the hallmarks of extroversion, whereby the so-called Mozambique Ocean (closed via Pan-African assembly) would have constituted merely one gulf of the global ocean Mirovia (McMenamin & McMenamin 1990). If Pannotia existed briefly as a true supercontinent, then it also would have formed by extroversion, but its transition to Pangaea would be characterized as tectonically introverted, because the younger landmass assembled via closure of the young Iapetus and Rheic ocean basins. As noted above, depending on how future supercontinent Amasia completes its assembly, the transition may be characterized by introversion (Atlantic Ocean re-closing), extroversion (Pacific Ocean closing) or orthoversion (Arctic Ocean closing).

Ten outstanding questions can be addressed for the imminent advancement of supercontinent research.

(1) How many supercontinents have existed on Earth (e.g. was there Kenorland, or were there separately drifting supercratons at 2.5 Ga)?
(2) How do semi-supercontinents (e.g. Gondwana, Laurasia, Eurasia) influence mantle structure and palaeoenvironments?
(3) What are the correct configurations of Nuna and Rodinia?
(4) Do supercontinents introvert, extrovert or orthovert – or do patterns alternate through supercycles?
(5) How long-lived are the present-day LLSVP structures, and how do plumes/superplumes relate to supercontinents?
(6) Is there a plate-tectonic speed limit?
(7) What is the solution to the Ediacaran palaeomagnetic enigma?
(8) What is the long-term behaviour of the geodynamo, and how did this affect the surface environment?
(9) How sensitive is global climate to the supercontinent cycle?
(10) Have pre-Pangean continental amalgamations affected the biosphere as significantly as post-Pangean palaeogeographies?

Although substantial progress has been made in supercontinental research within the last few decades, there is clearly a lot more work to be done.

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