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ABSTRACT

The status of Pannotia as an Ediacaran supercontinent, or even its mere existence as a coherent large landmass, is controversial. The effect of its hypothesized amalgamation is generally ignored in mantle convection models claiming the transition from Rodinia to Pangaea represents a single supercontinent cycle. We apply three geodynamic scenarios to Pannotia amalgamation that are tested using regional geology. Scenarios involving quasi-stationary mantle convection patterns are not supported by the geological record. A scenario involving feedback between the supercontinent cycle and global mantle convection patterns predicts upwellings beneath the Gondwanan portion of Pannotia and the arrival of plumes along the entire Gondwanan (but not Laurentian) margin beginning at c. 0.6 Ga. Such a scenario is compatible with regional geology, but the candidates for plume magmatism we propose require testing by detailed geochemical and isotopic studies. If verified, this scenario could provide

geodynamic explanations for the origins of the late Neoproterozoic and Early Palaeozoic Iapetus and Rheic oceans and the terranes that were repeatedly detached from their margins.

INTRODUCTION

Over the past three decades, a consensus has emerged that Pangaea is only the most recent in a series of supercontinents that have punctuated Earth history for billions of years (Nance et al., 1986, 1988; 2014; Zhao et al., 2002). Episodic recurrence of supercontinent assembly and breakup, i.e., the supercontinent cycle, has been linked to global-scale orogenesis, crustal growth, rapid climate swings, the evolution of life, biogeochemical cycles, sea-level change, large igneous provinces (LIPs), biasing of the geological record, and whole mantle convection patterns (Nance et al., 1986, 2014; Condie, 2011; Bradley 2008, 2011; Mitchell et al 2012; Yoshida and Santosh 2011; Cawood et al., 2013; Ernst, 2014). Best known of the pre-Pangaean supercontinents is Rodinia (McMenamin and McMenamin, 1990; Li et al., 2008; Merdith et al., 2017), which amalgamated during global-scale *c.* 1.1–0.9 Ga orogenesis, and reputedly rifted apart in two stages (*c.* 0.85-0.70 Ga and *c.* 0.62-0.54 Ga; Cawood et al., 2001; Hoffman, 1991; Li et al. 2008). Hypothesized earlier supercontinents include *c.* 1.6 Ga Nuna (aka Columbia), *c.* 2.5 Ga Kenorland, *c.* 3.0 Ga Ur, and *c.* 3.6-2.8 Ga Vaalbara, although Ur and Vaalbara, if they existed, might be better described as supercratons (i.e. composed of large Archean landmasses of "with a stabilized core that on break-up spawned several independently drifting cratons". Bleeker, 2003).

The hypothesized late Neoproterozoic supercontinent Pannotia (*c.* 0.65-0.54 Ga; Gondwana+Laurentia+Baltica ± Siberia; Stump 1987, 1992; Powell 1995; Dalziel 1997) is the most controversial of the hypothesized pre-Pangaean supercontinents, the very existence of which is debated (Evans et al., 2016; Nance and Murphy, 2019). Its hypothesized existence spans a pivotal time interval in Earth's evolution characterized by global-scale orogeny, rapid continental growth, profound changes in the chemistry of the oceans and atmosphere, the last of the Snowball glaciations, and the evolution of multicellular animals (Hoffman 1991; Murphy and Nance, 1991; Hoffman et al 1998; Dalziel, 1991, 1997; Knoll 2013; Maruyama and Santosh, 2008; Narbonne, 2010; Nance and

Murphy, 2019). Pannotia's configuration, size, and duration relative to Pangaea—i.e., its legitimacy as a supercontinent—is therefore controversial (Fig. 1). Its existence as a large landmass, or even as a collection of nearby but unsutured continents, has been called into question.

Concerning Pannotia's putative size and configuration, enduring controversies exist with regard to the 0.6-0.5 Ga global palaeomagnetic database in terms of its reliability and palaeogeographic implications (Evans, 2000, Evans et al., 2016; Scotese 2009; Pisarevsky et al. 2008; Levashova et al., 2013; Abrajevitch and Van der Voo 2010; Mitchell et al., 2011; Meert, 2014). Nonetheless, palaeomagnetic data are permissive of its existence (e.g., Scotese, 2009; Scotese and Elling, 2017; Robert et al., 2018). It is questionable whether Pannotia would have fulfilled the criterion that a supercontinent should "consist of at least 75 % of the preserved continental crust prior to initial breakup" (Meert, 2012). Nonetheless, its existence as a collage of closely located large landmasses is difficult to refute.

The Pannotia hypothesis may have issues with timing and duration as well as size. Geochronological data suggest that its breakup (the 605-520 Ma separation of Laurentia, Baltica, and Amazonia during opening of the Iapetus Ocean; Cawood et al., 2001) may have started before its Gondwanan portion was fully assembled (620-530 Ma collisional orogenesis; Collins and Pisarevsky, 2005; Cawood and Buchan, 2007; Li et al., 2008; Merdith et al., 2017; Oriolo et al. 2017, Fig. 2). Nonetheless, the breakup (and assembly) of supercontinents is a diachronous process such that supercontinents may begin to rift internally before all the continental fragments have assembled along their margins. Indeed, the breakup of Pangaea (opening of the Atlantic Ocean) initiated before much of southeastern Asia and China had accreted to its eastern flank (e.g., Scotese 2001; Stampfli & Borel 2002). In addition, as Nance and Murphy (2019) point out, it is the onset, rather than the termination, of collisional orogenesis that likely provides the best record of the timing of continental collision, after which orogenesis becomes more Mediterranean in style (i.e., localized within oceanic re-entrants between continental promontories). If so, much of the Gondwanan portion of Pannotia may have been assembled by *c*. 620 Ma. Nevertheless, these data collectively indicate that, if Pannotia existed, it did so only fleetingly between *c*. 620 Ma and 605 Ma. If longevity (≥ 100 Myr) is part of the definition of a supercontinent, then Pannotia clearly does not qualify. Indeed, syntheses that interpret the evolution from Rodinia to Pangaea as one supercontinent cycle infer the opening of the Iapetus Ocean as the final stage in the protracted (0.8-0.6 Ga) breakup of Rodinia (Hoffman, 1991; Li et al., 2008) rather than Pannotia.

For the purposes of this paper, we consider the size and longevity of a supercontinent to be less important criteria than a supercontinent's effect on global mantle convection. From this perspective, a supercontinent assembly must influence mantle circulation in a manner that triggers the next stage in the cycle (i.e., breakup), thereby allowing the cycle to repeat (Nance and Murphy, 2019). In this context, if the amalgamation of Pannotia did affect global-scale mantle circulation, its mantle legacy must be factored into models for Pangaea amalgamation.

An over-arching question critical to the legitimacy of Pannotia is whether or not its assembly influenced global-scale mantle convection patterns (Pastor-Galán et al., 2019; Nance and Murphy, 2019). In this volume, for example, Heron et al. present 3-D global convection models showing that convergence leading to the amalgamation of Pannotia resulted in increased core-mantle heat flux similar in magnitude to that of Rodinia and Pangaea, implying that convergence leading to its assembly may indeed have had the ability to impact mantle circulation patterns. Although the details are controversial, there is a general consensus that supercontinent assembly involves a coupling between the lithosphere and mantle convection (e.g., Zhong et al., 2007; Mitchell et al., 2019). For example, subduction zones, involving oceanic lithosphere descending to the core-mantle boundary (e.g., Dziewonski et al., 2010), can form a global-scale mantle downwelling zone where a supercontinent eventually assembles (Anderson, 1994). More controversially (see Anderson and Natland, 2007; Anderson, 2013; Hole and Natland, 2019) supercontinent dispersal has been linked to upwelling from the core-mantle boundary in the form of mantle plumes (e.g., Nance at al., 1986, 1988, 2014; Gurnis, 1988; Condie, 2004; Zhong et al., 2007; Li and Zhong, 2009; Mitchell et al 2012; Yoshida and Santosh 2011; Ernst, 2014; Keppie 2016; Heron, 2019; Doucet et al., 2020).

The potential influence of Pannotia assembly on mantle convection patterns is therefore of fundamental importance and has been implicitly ignored in models claiming that the transition from Rodinia to Pangaea represents a single supercontinent cycle. For example, Li et al. (2019) propose that Pangaea formed by "the closure of the external ocean that surrounded Rodinia" (i.e., extroversion), whereas Murphy et al. (2009) maintain that while this was the case for the assembly of Gondwana (Pannotia), Pangaea was formed by closure of the interior Iapetus and Rheic oceans (i.e., introversion). In the Li et al. (2019) model, any mantle legacy at ca. 600 Ma could be interpreted as a lingering mantle structure of Rodinia, with the assembly of Gondwana viewed as an early stage in the assembly of Pangaea and widespread late Neoproterozoic-early Cambrian rifting as the final stage in Rodinia breakup. Deducing the mantle legacy of Pannotia by investigating the effect of its assembly on late Neoproterozoic-early Palaeozoic (0.6-0.5 Ga) mantle convection patterns is consequently fundamental to understanding both the supercontinent cycle and, more specifically, the formation of Pangaea. Indeed, if Pannotia is viewed as a supercontinent, then its fleeting tenure relative to Rodinia and Pangaea would challenge many pre-conceived views about the very cyclicity of the supercontinent "cycle", or at least introduce an additional higher frequency harmonic than the typically assumed singular ~600 Myr time interval (Evans et al., 2016).

Pannotia's hypothesized existence spans a pivotal time interval in Earth's evolution characterized by global-scale orogeny, rapid continental growth, profound changes in the chemistry of the oceans and atmosphere, rapid and dramatic climate swings, as well as an explosion in biological activity (Hoffman 1991; Murphy and Nance, 1991, 2003; Hoffman et al 1998; Dalziel, 1991, 1997; Knoll 2013; Maruyama and Santosh, 2008). Nance and Murphy (2019) examined tectonic, geochemical, climatic, and biological proxies of these events and, from that perspective, concluded that the evidence for the assembly and dispersal of a late Neoproterozoic–Cambrian supercontinent was "unmistakable". This evidence includes the timing of orogenic events, secular trends of $\varepsilon Hf_{(t)}$ and average $\delta^{18}O$ of detrital zircon ages, passive margin development and epeirogenic uplift, $^{87}Sr/^{86}Sr$ and stable isotope excursions in seawater, atmospheric oxygen levels, and biological extinctions and radiations. These proxies are interpreted to reflect the geological record of global-scale (Pan African-

Cadomian-Baikalian) orogeny that heralds Pannotia assembly. Then the occurrences of widespread mafic dyke swarms and passive margin development, as well as major sea-level change, both point to important mantle involvement in rifting and ocean basin volume, respectively. Models that maintain Pannotia did not exist, or if it did exist, was not a supercontinent, interpret this rifting event as the final breakup of Rodinia (e.g., Hoffman, 1991; Li et al., 2008; Evans et al. 2016).

In this paper, we complement the proxy approach by highlighting aspects of the geological record that can be used to test whether the amalgamation of Pannotia affected global-scale mantle circulation patterns, and propose a particular methodology that could be applied to specific field areas. From the outset, it is important to realize that, at the time of writing, the co-authors of this paper disagree about the status of Pannotia as a supercontinent. Indeed, the published opinions of some have vacillated in recent years. We do agree, however, that resolution of this controversy is fundamental to understanding the dynamics of the supercontinent cycle and has first-order implications for models of the origin of Pangaea. To that end, we present evidence (some admittedly circumstantial), which combines insights from recent advances in geodynamic modelling with geological constraints, to argue the case that Pannotia may indeed have had a mantle legacy that should be factored into geodynamic models for supercontinent formation. Resolving this issue may take years of research in which cooperation among international geoscientists will be paramount. But we also propose specific regions where the ideas presented can be tested. Some of these regions have been identified as LIPs in recent compilations (Ernst at al., 2020), including areas where modern geochemical modelling techniques have identified plume basalts and elevated mantle potential temperatures (100-250˚C higher than ambient mantle) consistent with a plume origin and mantle upwelling (e.g., Lee et al., 2009; Tegner et al., 2019). In other regions, the role of mantle plumes has been independently interpreted by researchers, although modern geochemical investigations, as exemplified in Tegner et al. (2019), have yet to be employed. Such investigations recognize that an ascending plume will likely entrain upper asthenospheric mantle and so generate basaltic magmas with trace element and isotopic compositions that range from OIB to MORB. In such regions, mantle potential temperatures could be calculated for Mg-rich lavas (MgO > 8.5 wt %) whose compositions are mainly controlled by olivine fractionation (Lee et al., 2009; Lee and Chin, 2014).

In regions where the appropriate data are not currently available, it is hoped that this contribution stimulates the research necessary to test the model we propose. More generally, our approach shows that advances in geodynamic modelling provide the opportunity for testing these models with detailed fieldwork complemented by laboratory analyses of samples selected from key localities. We provide three examples showing how geodynamic models can be tested by using regional geology combined with geochemical and isotopic studies of specific igneous suites. These examples predict contrasting locations for large low-shear-velocity provinces (LLSVPs), and hence for the plume-related magmatism that would have emanated from their margins.

GEOLOGICAL AND CONCEPTUAL FRAMEWORK

Global plate reconstruction models for the Neoproterozoic are poorly constrained because of the limited number of reliable palaeopoles and because of evidence of inferred episodes of true polar wander (e.g. Kirschvink et al., 1997; Evans, 2003; Li et al., 2004, 2008; Mitchell, 2014; Robert et al., 2018) and the possibility of an anomalous behavior of the geomagnetic field (e.g. Abrajevich and Van der Voo, 2010). Given the uncertainties inherent in continental reconstructions, for the purposes of this study we employ a working definition of a supercontinent as a continental configuration "with a size capable of influencing mantle convection patterns and core–mantle boundary processes" (Pastor-Galán et al., 2019; Nance and Murphy, 2019). In this view, supercontinent assembly is a phase of the supercontinent cycle and must have a *legacy in mantle circulation*, the consequence of which is to *initiate the next phase* (breakup), thereby "enabling the cycle to repeat". This view is implicit in numerical models (Zhong et al. 2007; Li and Zhong 2009; Fig. 4) showing a supercontinent assembling above region of mantle downwelling, which evolves to become an upwelling beneath the supercontinent, ultimately leading to its breakup. In this paper, we adopt the view (e.g., Nance and Murphy, 2019) that the establishment of the mantle legacy accompanying supercontinent assembly is a more useful criterion for defining a supercontinent than one employing an arbitrary percentage of

Earth's continental lithosphere. Any legacy of mantle circulation must be factored into models purporting to explain how the next supercontinent formed, just as Pannotia's mantle legacy, if one was produced, would need to be factored into models for Pangaea amalgamation.

Although the precise timing of Rodinia breakup is controversial, with estimates ranging from 0.85 to 0.70 Ga, most syntheses agree that it began with the opening of the Adamastor Ocean between Laurentia-Baltica-Amazonia-West Africa-Rio de la Plata and Australia-East Antarctica-India-Congo-Kalahari (Wingate and Giddings 2000; Li et al., 2008; Li and Evans, 2011; Merdith et al., 2017), and culminated at c. 0.62-0.55 Ga with the opening of the Iapetus Ocean between Laurentia and Baltica to the east, and Laurentia and South America to the west (e.g. Cawood et al., 2001; Pisarevsky et al., 2008; Fig. 2). Collectively, these two rifting events resulted in Laurentia being bordered by passive margins by 0.54 Ga (Bond et al., 1984; Hoffman, 1991). The opening of the Iapetus Ocean coincided with the emplacement of a large igneous province known as the Central Iapetus Magmatic Province (CIMP) at 615–530 Ma (Ernst et al., 2013, 2020).

The 0.75-0.55 Ga time interval is also characterized by widespread Pan-African collisional orogenesis brought about by the closure of the Braziliano, Adamastor, and Mozambique oceans and the amalgamation of the Gondwanan portion of Pannotia (e.g. Kröner and Stern, 2004; Cawood and Buchan, 2007; Li et al., 2008; Merdith et al., 2017 and references therein; Fig. 2). As subduction between the colliding continents terminated, renewed subduction occurred along much of Gondwana's periphery as exemplified by the regionally extensive arc magmatism that typifies the Avalonian-Cadomian and Terra Australis orogens (e.g. Murphy and Nance, 1991; Nance et al., 2002; Cawood and Buchan 2007).

We present three examples (here termed scenarios A, B and C) of how geodynamic models, when combined with regional geological constraints, can provide the conceptual framework to test the existence of Pannotia's mantle legacy (Figs. 5, 6 and 7). These scenarios have contrasting implications for Neoproterozoic global-scale mantle circulation patterns. One view (scenario A, Fig. 5) interprets the mantle to have been dominated since at least the Neoproterozoic (Torsvik et al, 2006) by spherical harmonic degree-2 planform convection, that is, a pattern of convection with two quasi-stationary antipodal regions of upwelling bisected by a subduction girdle of downwelling (Zhong et al., 2008; Fig. 4d). Today these regions of upwelling coincide with two Large Low Shear Velocity Provinces (LLSVPs), known as Tuzo and Jason (or African and Pacific) respectively (Fig. 3; Torsvik et al., 2006, 2014), located near the core–mantle boundary beneath Africa and the central Pacific Ocean (Dziewonski, 1984; Williams et al. 1996). Both features have equatorial positions that reflect the optimal moment of inertia on a spinning Earth (Niu, 2018), and are bisected by a north-south girdle of downwelling (Torsvik et al. 2006; Burke et al. 2008, Burke, 2011; Steinberger & Torsvik, 2010) that constitutes the subduction systems of the circum-Pacific. Adherents of the quasi-stationary model disagree on the origin of the LLSVPs, which are variously interpreted as the long-term accumulation of subducted slabs (e.g., Niu, 2018), or as long-lived structures that formed during the very early stages of planetary differentiation (e.g., Dziewonski et al., 2010; Trønnes et al., 2019).

In contrast to scenario A, scenarios B and C envisage changing mantle convection patterns in response to global tectonics. For instance, the circular distribution of the 0.6-0.5 Ga palaeomagnetic data in the palaeomagnetic reference frame (i.e. reconstructed continents) has been interpreted to be associated with true polar wander (TPW), thereby constraining relative palaeolongitudes of the continents (e.g., Raub et al., 2007; Mitchell et al., 2012, Robert et al., 2018). Episodes of TPW during the late Neoproterozoic, thought to have resulted from mass heterogeneities in the mantle introduced by reactivation of a subduction girdle surrounding the continents, permit Robert et al. (2018) to deduce areas of upwelling (LLSVPs) in a degree-2 planform mantle convection scenario (scenario B, Fig. 6).

In scenario C (Fig. 7a-d), LLSVPs are not fixed and mantle convection patterns alternate between degree-2 planform and a spherical harmonic degree-1 structure in response to the supercontinent cycle (Zhong et al., 2007; Li & Zhong 2009). Degree-1 structure is where major mantle upwelling occurs in one hemisphere whereas major downwelling occurs in the other (Fig. 4a-c). For example Li et al. (2019) proposed that Nuna amalgamation at *c.*

1.6 Ga was the outcome of a degree-1 mantle structure that transitioned to a degree-2 mantle structure resulting in the breakup of Nuna and the assembly of Rodinia at 0.9 Ga. Supercontinent amalgamation is a consequence of subduction of oceanic lithosphere between the converging continents in a region of mantle downwelling (Fig. 4) and the subducted oceanic lithosphere congregates in the lower mantle to form "slab graveyards" (e.g., Richards and Engebretson, 1992; Van der Voo et al., 1999; Tan et al., 2002; Tackley, 2011; Wu et al., 2017).

Several processes, the relative importance of which is a matter of current debate, are thought to cause these downwellings to evolve into upwellings of return mantle flow (Fig. 4c,d). For example, the role of increased insulation along the base of the lithosphere (e.g. Anderson, 1982; Gurnis, 1988; Nance et al., 1988) has been questioned by some studies (Lenardic et al., 2005; Li and Zhong, 2009; Heron and Lowman, 2011) but supported by others (e.g. Yoshida, 2013). Other processes that may play a significant role include the exothermic conversion of perovskite to post-perovskite in the downgoing slab, the bulldozing and displacement of existing thermochemical piles at the core-mantle boundary (possibly including radiogenically enriched melts trapped within the slab graveyard), increased insulation across the core-mantle boundary, and the thermal effects of subduction around the periphery of the supercontinent (e.g., Anderson & Dziewonski, 1984; van der Hilst & Kárason, 1999; Tan et al., 2002; Zhong et al., 2007; Maruyama et al 2007; Li et al., 2008; Li & Zhong 2009; Yoshida & Santosh 2011; Mitchell et al., 2012; Heron et al., 2015, this volume; Niu, 2018; Heron, 2019).

TEST FOR PANNOTIA

Geodynamic models indicate that deep mantle plumes are preferentially produced along the edges of LLSVPs (e.g. Tan et al., 2002; Steinberger & Torsvik, 2010), a conclusion supported by the correlation between the reconstructed positions of Mesozoic-Cenozoic hot spots, large igneous provinces and kimberlites around the margins of Tuzo and Jason (e.g., Torsvik et al. 2006; Burke et al. 2008; Li and Zhong 2009; Fig. 3). The three scenarios discussed above predict different locations of LLSVPs relative to the continents in the Late Neoproterozoic-Cambrian and hence different locations of plume magmatism. Therefore, they provide a conceptual framework to test their veracity and to investigate the potential existence of Pannotia's mantle legacy. Plume magmatism can be identified in the geological record by the geochemical and isotopic fingerprints of mafic igneous rocks, which inherit the compositions of their mantle sources (Hawkesworth & Kemp 2006; Hawkesworth & Scherstén 2007; Lee et al., 2009; White, 2010; Herzberg & Gazel, 2009; Brown & Lesher, 2014; Niu, 2018; Tegner et al., 2019).

Scenario A, which assumes that the Tuzo and Jason LLSVPs have remained quasi-stationary since the Neoproterozoic (e.g., Torsvik, 2019), can be tested in various continental configurations proposed for the Ediacaran-Early Cambrian. For example, in the Ediacaran-Early Cambrian reconstructions of Merdith et al. (2017) the assumption of a quasi-stationary LLSVP scenario places Jason beneath Laurentia (Fig 5a,b). Alternatively, Tegner et al. (2019) position CIMP along the eastern edge of Jason, thereby providing a means to reconstruct the palaeogeography of Baltica and Laurentia 615 million years ago. In their reconstruction, one of the two LLSVPs (Jason) would have been located beneath the Laurentian (rather than the Gondwanan) portion of Pannotia (Fig. 5a, dashed line), both before Pannotia assembled and after it broke up.

Scenario B deduces the locations of a subduction girdle and two antipodal LLSVPs assuming degree-2 planform convection (Robert et al. 2018). In this scenario, one of their LLSVPs is also located beneath Laurentia from 635-525 Ma (Fig. 6a, b).

By contrast, scenario C is based on the proposition that the supercontinent cycle is modulated by an alternation between degree-1 and degree-2 planform mantle convection (Zhong et al., 2007; Li & Zhong 2009, Li et al., 2019). Scenario C views Pannotia amalgamation in the Late Neoproterozoic (0.65-0.55 Ga) to be the product of continental collisions (Pan-African orogenesis) resulting from the subduction of oceanic lithosphere between the converging continents in a region of mantle downwelling located beneath its Gondwanan portion. Assuming a mantle transit time (the time it takes for a particle to migrate from the surface to the base of the mantle) of ~0.06 Ga (Davies, 2011; van der Meer et al., 2018), or ~ 0.12 Ga (Hounslow et al., 2018) the evolution from mantle downwelling to mantle upwelling at the core-mantle boundary would have begun at c. 0.6-0.54 Ga. The absence of late Neoproterozoic collisional orogenies in either Laurentia or Baltica would predict that this evolution was restricted to a location beneath the Gondwanan portion of Pannotia (Fig. 7).

Scenarios A-C have some features in common. Assuming that deep mantle plumes are preferentially produced along the edges of LLSVPs, all three scenarios require that mafic complexes along the Laurentia-Gondwana boundary have a plume signature, so the absence of this signature would either falsify them all, or brings the validity of the used reconstructions into question. Indeed, widespread plume activity associated with Laurentia-Amazonia-Baltica rifting satisfies the location of LLSVPs predicted by all three scenarios. The early stages of the opening of the Iapetus Ocean and Tornquist Sea are commonly interpreted to be the result of a plume centred on the triple junction between these three continents (Higgins and van Breemen, 1998; Cawood et al., 2001; Pisarevsky et al., 2008; Mitchell et al., 2011; Ernst, 2014; Tegner et al. 2019; Fig. 8a,b). This plume activity produced the Central Iapetus Magmatic Province (CIMP) and occurred in several pulses, at 615 Ma, 590 Ma, 563 Ma and peaking at 550 Ma in Laurentia-Gondwana (Ernst, 2014). CIMP comprises several LIPs and produced several mafic dyke swarms, basalts, bimodal basalt-rhyolite complexes, lamprophyres and carbonatites (Higgins & van Breemen 1998; Abdel-Rahman & Kumarapeli 1999; Puffer 2002; Kamo et al 1995, Ernst & Bell 2010; Mitchell et al., 2011; Ernst 2014; Keppie et al. 2006; Weber et al. 2019; Youbi et al. 2020; St. Seymour and Kumarapeli 1995; Cawood et al 2001; Thomas, 2011, 2012; Pisarevsky et al 2008; Tegner et al. 2019).

According to Ernst and Buchan (2001, 2004), the c. 615 Ma Long Range (Laurentia)-Egersund (Baltica) and c. 590 Ma Grenville-Rideau (Laurentia) mafic dyke swarms belong to LIP provinces associated with two CIMP plume centres about 800 km apart (Fig. 8a) located along the continental margins that herald rifting of Laurentia from Baltica and Amazonia, respectively. These plume centres are the likely source of intrusions along northern and eastern Laurentia, as well as West Africa and Scandinavia. Carbonatite and related ultramafic intrusions in Laurentia/Greenland at c. 600 Ma (Sarfartoq, Greenland; Larsen and Rex 1992) and c. 574 Ma (St. Honore; Doig and Barton 1968), in Baltica at c. 589 Ma and 584 Ma (Fen and Alnø complexes; Meert et al. 1998; Walderhaug et al. 2003), and the c. 560 Ma Sept-Îles–Catoctin LIP in Laurentia may constitute additional manifestations of plume activity (e.g., Pisarevsky et al., 2008). Likewise, the *c.* 619 Ma mafic dyke swarm in Oaxaquia, Mexico, has been attributed to initial Amazonia-Baltica rifting (Weber et al. 2019).

Despite their similar predictions for the evolution of the Laurentia-Gondwana margin, scenarios A-C differ profoundly in the inferred locations of Ediacaran-early Palaeozoic LLSVPs, and so can be distinguished by interrogating the geological record for the distribution of contemporary plume-related magmatism around Laurentia and Gondwana, respectively. The recent study of Doucet et al. (2020) suggests they also could be distinguished by the evolution of their respective mantle domains in responses to subduction of oceanic and continental material. Documented LIPs along the western and northern margins of Laurentia (a requirement of scenarios A and B in the chosen reconstruction models) are sparse. According to Ernst and Buchan (2001, 2004), late Neoproterozoic magmatism in northern Laurentia (Fig. 8a) was sourced from plume centres associated with Laurentia-Gondwana-Baltica rifting. The c. 780 Ma Gunbarrel and c. 723-715 Ma Franklin dyke swarms (Ernst and Buchan, 2001, 2004; Ernst et al., 2013) pre-date Gondwana amalgamation. Deposition of thick 0.65-0.54 Ga sedimentary successions with subordinate mafic rocks (e.g. 570 Ma Hamill Group and correlatives) occur on both Laurentia's eastern and western margins (Prave, 1999; Colpron et al., 2002; Keeley et al., 2013; Macdonald et al., 2013; Spencer et al. 2014; Fig. 9a,b). Although the basalts of the Hamill Group and correlatives are within-plate and thought to be rift-related, the overall dominance of thick sedimentary successions is not compatible with the stratigraphy associated with a LIP and so is inconsistent with scenarios A and B.

Testing Scenario C: As our interpretation of the available literature does not support scenarios A and B in the chosen reconstructions, the remainder of the text will focus on the viability of scenario C. According to scenario C, one of the LLVSPs should be located beneath the Gondwanan portion of Pannotia where Neoproterozoic subduction culminated in Pan-African collisional orogenesis. Furthermore, since collisional orogenesis is not recognized at this time in either Laurentia or Baltica, the exterior (ocean-facing) margins of these two continents should be unaffected by plumes emanating from this sub-Gondwanan LLSVP. Instead, scenario C predicts the arrival of plumes along the entire Gondwanan margin beginning at c. 0.6 Ga as evidence of an LLSVP anomaly beneath Gondwana that produced mafic magma characterized by a mantle source potentially enriched in continental material assimilated during Pan African subduction (see Doucet et al., 2020).

Plume-related magmatism (see Tegner et al., 2019) occurs around the margins of Gondwana (Fig. 7). Although the details may differ, this overall relationship is not particularly sensitive to our choice of reconstruction (e.g. Fig. 7d, after Dalziel, 1997). In most of these localities shown, magmatism has been independently attributed to a mantle plume by the authors of the study cited, but modern geochemical modelling techniques have not been employed. By drawing attention to the temporal-spatial distribution of this magmatism we emphasize its importance in validating a fundamental aspect of the supercontinent cycle, and we hope to stimulate research that marries modern geodynamic models with detailed fieldwork complemented by geochemical and geochronological studies.

Through the late Neoproterozoic to early Palaeozoic, most of Gondwana was surrounded by subduction zones as evidenced by the dominance of voluminous arc magmatism along its margins (e.g., Avalonian-Cadomian belt: Nance et al., 2002, Murphy et al., 2019; Terra Australis orogen: Cawood, 2005; Cawood and Buchan, 2007, Merdith et al., 2017; Anatolia-Iran-Arabia, Stern, 1994). Recent laboratory studies, which suggest subduction zones are a barrier that may distort or even destroy ascending plumes (e.g., Kincaid et al., 2013; Druken et al., 2014), pose significant challenges to identifying regions where mantle plumes may have penetrated into the upper plate (Fig. 10). However, 3D numerical models (Betts et al., 2012, 2015) show that the transfer of plume material into the upper plate is facilitated by gaps or tears in the slab (e.g. slab windows, slab breakoff, delamination), or where roll-back results in the generation of back-arc basins. Such processes result in rejuvenation of the continental lithospheric mantle (CLM) that should be recorded in the isotopic signature of mafic magmas derived from it (Foden et al., 2002; Murphy and Dostal, 2007; Gutiérrez-Alonso et al., 2011; Murphy, 2016; Dostal et al., 2019).

Taken together, this analysis suggests that the late Neoproterozoic-early Palaeozoic evolution of the Gondwanan margin, although dominated by well-documented arc-related rocks, may preserve specific localities where igneous complexes have a plume component, reflecting the influx of plume material into the upper plate. The late Neoproterozoic northern margin of Gondwana is dominated by ensialic arc-related magmatism of the Avalonian-Cadomian belt, the main phase of which began at *c*. 640 and terminated diachronously along the margin from c. 610 Ma to *c*. 530 Ma as the result of the propagation of a San Andreas-style transform fault, to be followed by an early Palaeozoic platformal succession (e.g. Nance et al., 2002; Pollock et al, 2009, Fig. 11). Such environments commonly develop slab windows (e.g., Thorkelson, 1996) that would allow plume material to invade the upper plate (Betts et al., 2012, 2015). Indeed, several features of the belt's late Neoproterozoic-Early Cambrian magmatism may reflect the development of such a window, including the occurrence of *c*. 610-540 Ma appinite complexes (compositionally diverse suites dominated by hornblende-rich mafic gabbros and lamprophyres that crystallized from anomalously water-rich magmas). Appinite complexes typically reflect the product of asthenospheric upwelling in the aftermath of subduction cessation (e.g. Atherton and Ghani, 1992) and occur within this time window in Atlantic Canada, the Channel Islands and Brittany (Fig. 7; Murphy, 2013, 2020; D'Lemos et al., 1992, 2001; Inglis et al., 2005). Such an origin is supported by O isotopes in hornblendes from lamprophyres and Mg-rich mafic dykes in the Greendale Complex, Nova Scotia (Cawood et al., 2019), which indicates that the water accompanying crystallization had a strong juvenile mantle component.

Conversion of the Cadomian margin of Gondwana from a convergent to transform margin by the early Cambrian (Fig. 11; Nance et al., 2002) would have eliminated any potential barrier to ascending plumes, enabling their emplacement into the crust. In support of this, Cambrian alkalic basalts in Atlantic Canada were derived from an enriched mantle source attributed to a mantle plume (e.g. Greenough and Papezik, 1985; Murphy et al., 1985). By the middle Cambrian, the transform fault had propagated across eastern Avalonia (western Europe) as far as the Iberian margin of Gondwana, where Sanchez-Garcia et al (2003, 2008) invoke a plume model to explain voluminous middle Cambrian-early Ordovician alkalic mafic magmatism in the Ossa Morena Zone that ultimately led to the opening of the Rheic Ocean. Volcanic rocks in this region include Mg-rich basalts interpreted to reflect high degrees of partial melting of primitive mantle, consistent with their plume model. The anomalously large volume of magma emplaced is thought by these authors to reflect a thermal anomaly associated with the development of a slab window.

Along the northern Gondwanan margin, the Ollo de Sapo "formation" (Fig. 7) has long been recognized as an enigmatic Late Cambrian-Early Ordovician (495-470 Ma) magmatic event, the type area of which is in the Central Iberian Zone, but whose manifestations are expressed in every Variscan massif in continental Europe extending some 2000 km along the north African margin of Gondwana (Montero et al., 2009a,b; Talavera et al., 2015; García-Arias et al 2018; Casas and Murphy, 2018). Typical Ollo de Sapo rocks consist of a complex assemblage of predominantly foliated volcanic, subvolcanic and megacrystic plutonic igneous rocks that are largely felsic in composition. In the Iberian massif, the peak of magmatic activity occurred at *c*. 477 Ma and is classified (Gutiérrez-Alonso et al., 2016) as a "super-eruption" because lava exceeding 450 km^3 in volume erupted in "a relatively short period of time" (Self, 2006). Regional mapping shows that the

Ollo de Sapo eruption occurred before the deposition of the rift-to-drift Armorican quartzite (Montero et al., 2009a; Díez Montes et al., 2010) that heralds the opening of the Rheic Ocean by rifting and separation of Avalonia from the northern margin of Gondwana (Murphy et al., 2006; Sánchez-García et al., 2008; Pollock et al., 2009; Linnemann et al., 2010; Nance et al, 2010, 2012; Gutiérrez-Alonso et al., 2016). Penecontemporaneous ironstone deposits in Iberia and in several locations along the northern margin of the Rheic Ocean have been interpreted to record a change in ocean water chemistry associated with increased volcanic activity and input of hydrothermal Fe in this expanding seaway (Todd et al., 2019; Pufahl et al., 2020). The large amount of Late Cambrian-Early Ordovician magmatic rocks associated with the European Variscan belt has led some authors to propose the existence of a siliceous Large Igneous Province (SLIP) at this time (Díez Montes et al. 2010; Gutiérrez-Alonso et al., 2016; García-Arias et al., 2018). SLIPs are commonly interpreted as crustal expressions of hidden LIPs (Ernst, 2013), the large volumes of felsic magma they comprise requiring extensive crustal melting thought to be triggered by heat and fluids associated with underplating by a mantle plume (Bryan and Ferrari, 2013).

The late Neoproterozoic-Early Cambrian Blovice accretionary complex of the Bohemian Massif (Fig. 7) may preserve an example of mantle plume activity within an arc-backarc system along the northern Gondwanan margin (Ackerman et al., 2019). The blocks have both tholeiitic and alkalic affinities interpreted to have been derived from different mantle sources (MORB and OIB respectively). The complex is interpreted by Ackerman et al. (2019) as the vestige of intra-oceanic arcs, with mantle plume located at a spreading centre within a back-arc basins. This collage was accreted during protracted subduction along the Gondwanan margin, in a manner analogous to the Izu Bonin-Mariana arc-Philippine Sea in the western Pacific.

Further east, an example of plume magmatism similar to CIMP may be provided by the LIP of rift-related *c*. 580 Ma Volyn continental flood basalts (Fig. 7) overlain by a syn-rift to post-rift siliciclastic sequence (Poprawa, 2006). This sequence is thought to have been deposited in a failed rift and is exposed over 80,000 km^2 in western Ukraine, eastern Poland and southwestern Belarus (e.g. Krzywiec et al., 2018). These volcanic rocks include both high-Ti and low Ti-Nb tholeiitic basalts as well as high Al basalts, olivine basalts and picrites (Shumlyanskyy, 2016) and their emplacement has been attributed to a triple junction between Baltica, the Eastern European craton, and the Amazonian margin of Gondwana, associated with the opening of the eastern Tornquist Sea (Poprawa 2006; Krzywiec et al., 2018).

During the late Neoproterozoic and early Palaeozoic, most of southeastern Asia, including the Tarim, North China and South China blocks, and intervening microcontinents were either adjacent, or accreted, to northeastern Gondwana (Fig. 7; Zhao et al., 2018; Cawood et al, 2018). These cratons and microcontinents are characterized by 600-430 Ma rift-related magmatism. In the North Qilian orogen between the Qilian block and South China, for example, *c*. 600-580 Ma continental rift basalts have high potential temperatures (T_p ~1500 ^{o}C) attributed to a mantle plume coeval with rifting of the Qilian-Qaidam block and the *c*. 550 Ma opening of the Qilian Ocean (Xu et al., 2015). Similarly, the *c*. 525-500 Ma Lajishan-Yongjing ophiolitic blocks in the South Qilian orogen consist mainly of thick, E-MORB and OIB-type basalts and picrites, the last generated by high degrees (18-21 %) of partial melting at potential temperatures of ~1500-1600 °C (Zhang Y. et al., 2017). The ophiolitic blocks are interpreted to represent an oceanic plateau associated with a mantle plume. In the West Qinling orogen, located some 600 km to the east, ophiolites of similar age structurally overlain by marble are also interpreted to represent an oceanic plateau or seamount (Yang et al., 2018). Rift-related mafic rocks also occur in the NW Tarim Craton where they have been attributed to an initial (failed) attempt to break up Gondwana at *c*. 520 Ma (Turner, 2010; Lu et al., 2018).

Early Palaeozoic igneous rocks are also widely distributed in the Himalaya, Lhasa, Southern Qiangtang, Baoshan, Sibumasu and Tengchong terranes. Collectively they range in age from *c*. 530 to *c*. 430 Ma (Gao et al., 2019; Liu et al., 2020). Although bimodal in composition, they are dominated by granitoid rocks distributed over an area of >2500 km by 900 km and so constitute a typical SLIP (Dan et al., 2020). Almost all of the granites were derived from partial melting of sedimentary rocks, but a few show A-type characteristics (Ding et al., 2015; Gao et al., 2019). Coeval amphibolite facies metamorphic rocks

(Bhimpedian orogeny, Cawood et al 2007) yield ages of 490-465 Ma (Gehrels et al., 2006; Zhang et al., 2012; Palin et al., 2018). A sedimentary hiatus marked by either a disconformity or an angular unconformity coeval with the major magmatic flare-up is evident in all terranes (Myrow et al., 2009; Liu et al., 2020 and references therein). Although previous interpretations favour either an Andean-type subduction of proto–Tethys oceanic lithosphere beneath the northern Gondwanan margin, or post-collision extension associated with the collapse of the Pan-African orogen in NE Gondwana, the tectonic setting for this magmatic province may equally reflect a plume in a far-field subduction zone (Dan et al., 2020).

In northern, central, and Western Australia, geochemical and geochronological data indicate the existence of the *c*. 511 Ma Kalkarindji LIP (Fig. 7; Glass and Phillips, 2006; Evins et al., 2009; Jourdan et al., 2014; Ware et al., 2018). This LIP consists of *c*. 1500 m thick succession of basaltic andesitic flows, as well as sills and dykes that can be found across an area of 2.1 × 10^6 km^2 (Evins et al., 2009; Pirajno and Hoatson, 2012; Ware et al., 2018), and is broadly synchronous with the Early-Middle Cambrian (Stage 4) extinction recording the demise of the Archaeocyatha (Jourdan et al., 2014). Individual flows are commonly between 20 m and 60 m thick, with an eruptive volume estimated at 1.5 x $10^5 km^3$, similar to that of the Columbia River Basalt (Evins et al., 2009). Their geochemistry is typical of continental flood basalts with enriched Sr, Nd, Pb isotopic compositions that have been attributed to a mantle plume (Pirajno and Hoatson, 2012) and/or decompression melting and mantle warming focused by edge driven convection (Ware et al., 2018).

The 18,000 km long Terra Australis orogen is a composite orogen that preserves a *c*. 570-230 Ma record of subduction along much of the southern margin of Gondwana (e.g. Cawood, 2005; Cawood and Buchan 2007). However, in some segments of the orogen, there are volcanic sequences that pre-date the onset of subduction. The Delamerian segment of eastern Australia and the correlative regions in Tasmania and Antarctica, contain Neoproterozoic and Early Cambrian sequences that were emplaced before the onset of Cambrian contractional orogenesis and a protracted history of accretionary orogenesis (Fig. 7; Foden et al., 2002, 2006). These sequences are characterized by passive margin successions with abundant mafic magmatism. Thus, widespread Neoproterozoic to Cambrian basalt occurs in a belt up to 6 km thick and with a strike length of *c*. 3000 km (Direen and Crawford, 2003). Their geochemistry has been described by Foden et al., (2002), who report voluminous Neoproterozoic tholeiitic basalts, similar to continental flood basalts, with trace element abundances that have E-MORB characteristics, overlain by Cambrian undersaturated alkalic nephelinite–basanite-lamproite volcanic rocks with high LREE and HFSE abundances and juvenile Nd isotopic values. This range in isotopic compositions is consistent with an ascending plume that generated basaltic magmas with trace element and isotopic compositions ranging from OIB to MORB (Griffiths and Campbell, 1990).

Similarly, in southern Africa, a series of *c*. 505 Ma alkaline, mafic to felsic plutons intrude a zone of crustal weakness known as the Kuboos-Bremen line. Their emplacement post-dates Pan-African collisional tectonics caused by the closure of the Adamastor Ocean between the Kalahari and the Rio de la Plata cratons, and is coeval with subduction-related magmatism along this margin of Gondwana. Frimmel et al., (2000) attribute the alkaline magmatism either to a mantle plume located beneath southern Africa or to a far-field effect of collision along the Gondwana margin.

DISCUSSION

If a supercontinent must be large, long-lived, and mantle-influencing, then Pannotia may not merit the status as a "true" supercontinent. But if a supercontinent is viewed as an assembly that influences mantle circulation in a manner that triggers breakup, thereby allowing the cycle to repeat (Pastor-Galán et al., 2019; Nance and Murphy, 2019) then this paper presents a scenario in which Pannotia's supercontinent status can be tested. Irrespective of its supercontinental status, we have presented circumstantial evidence that its existence may have an influence on mantle convection, which if verified should not be ignored in mantle convection models for this pivotal interval of Earth's evolution.

Although the palaeomagnetic and geochronological data for Pannotia are equivocal, proxy records have been interpreted to provide a clear signal of supercontinent amalgamation during the late

Neoproterozoic (Nance and Murphy, 2019). If Pannotia amalgamation affected mantle circulation, then the legacy of this circulation must be factored into models of Pangaea amalgamation. We provide three examples of how geodynamic models can be tested with the available geological evidence. We show that scenario A, based on the assumption of quasi-stationary LLSVPs, and scenario B, which imposes degree-2 convection certain Ediacaran-Cambrian reconstruction, find little support in the geological record.

Detailed geochemical and isotopic studies of specific igneous suites located along the Gondwanan margin of Pannotia are consistent with scenario C. This model can be tested by detailed geochemical and isotopic studies of specific igneous suites located along the Gondwanan margin of Pannotia. If verified, scenario C would require dynamic LLSVPs whose stability and locations are in concert with lithospheric tectonics. Scenario C predicts that mantle plumes associated with the amalgamation of Pannotia would have emanated from an enriched mantle domain along the margins of an LLSVP located beneath its Gondwanan portion at *c*. 600 Ma. Given that much of Gondwana's periphery was characterized by subduction zones at this time, the effects of these plumes would be best expressed where gaps or tears occurred within these peripheral subduction zones or in specific regions where the onset of subduction was anomalously late.

Most geodynamic models of supercontinent amalgamation and dispersal assume that LLSVPs are produced beneath the centre of the supercontinent and subduction zones are symmetrically disposed around the periphery. However, in all three scenarios, the locations of the LLSVPs are asymmetric with respect to the centre of Pannotia – beneath Laurentia in scenarios A and B, beneath Gondwana in scenario C. Each of these scenarios imply that, if Pannotia existed, the thermal evolution of the mantle beneath it may have been profoundly asymmetric (Fig. 5-7), perhaps explaining its fleeting tenure. This asymmetry is a potential explanation for why supercontinents may start to rift along one margin before all the continental fragments have assembled (thereby explaining the enigma of Rodinia's final breakup occurring as the Gondwanan portion of Pannotia was amalgamating). The analysis also suggests that, in defining a supercontinent, we should not be distracted by its size or the timing of its final amalgamation versus initial rifting.

The location of CIMP at the junction between Laurentia, Baltica and Amazonia lies outside the realm of Gondwanan subduction zones where barriers to the ascent of plumes would have existed, and most models for the early stages of the opening of the Iapetus Ocean and Tornquist Sea involve one or more plumes centred on triple junctions between these continents (Higgins and van Breemen, 1998; Cawood et al., 2001; Pisarevsky et al., 2008; Tegner et al., 2019). Likewise, the triple point between Baltica, the Eastern European craton, and Amazonia has been linked to the plume-related opening of the eastern Tornquist Sea (Poprawa 2006; Krzywiec et al., 2018). Although the openings of the Iapetus Ocean and Tornquist Sea are well documented in the geological record, a geodynamic explanation of *why* they opened is lacking. Our approach suggests that plumes emanating from the margins of an LLSVP located beneath Gondwana can explain (i) the origin of the rift-rift-rift triple junctions that accompanied their opening and (ii) why CIMP emplacement occurred in several stages.

Plume magmatism along the northern margin of Gondwana also provides a potential explanation for the opening of the Rheic Ocean and the *origin* of as the so-called peri-Gondwanan terranes (Ganderia, Avalonia, Carolinia, Meguma etc), which were detached from this margin in piecemeal fashion beginning in the Late Cambrian (Murphy et al., 2006; Pollock et al., 2009; Nance et al., 2010, 2012; Linnemann et al., 2010; van Staal et al., 2012; Waldron et al., 2014; White et al., 2018). Most models for the Rheic Ocean favour an origin by subduction zone rollback, but the mechanism of subduction initiation is unclear (see Waldron et al., 2014, 2019). Geodynamic models suggest that plumes may assist subduction initiation by generating a zone of focused magmatic weakening and thinning of lithosphere above it (Gerya et al., 2015). More generally, plumes preferentially ascending beneath continental margins may explain why the same continental margin can undergo repeated episodes of rifting.

Our analysis suggests that Pannotia's assembly may have influenced global-scale mantle convection patterns in a manner similar to that associated with the amalgamation of Rodinia and Pangaea. Such influences have first-order implications for geodynamic models purported to explain the amalgamation of Pangaea, but are implicitly ignored in models claiming that the transition from Rodinia to Pangaea represents a single supercontinent cycle. Deducing the mantle legacy of Pannotia is therefore fundamental to understanding the supercontinent cycle itself, and cautions that taking a short-cut from Rodinia to Pangaea invites the possibility of inadequately considering early Palaeozoic mantle circulation patterns at a crucial stage in the assembly of Pangaea.

SUMMARY AND CONCLUSIONS

Models claiming the transition from Rodinia to Pangaea represents a single supercontinent cycle do not consider the potential mantle legacy of the Pannotia amalgamation. The proxy signals of assembly and breakup in the Ediacaran may be connected to major changes in Earth systems and possibly global mantle circulation.

We provide three examples of how geodynamic models can be tested by using regional geology. These examples (A, B and C) predict contrasting locations for LLSVPs and, hence, for the plume-related magmatism that emanates from their margins. In scenario A, mantle convection since at least the late Neoproterozoic is dominated by two quasi-stationary, antipodal, equatorial upwellings bisected by a north-south girdle of downwelling (degree-2 convection). Scenario B assumes degree-two convection to constrain the locations of LLSVPs relative to the continent. Although A and B differ in methodology, they both predict an LLSVP beneath Laurentia, which is not supported by the geological record.

Scenario C, however, is permissible although several candidates for plume magmatism along the Gondwanan margin require further testing. Pannotia amalgamation at 0.65-0.54 Ga requires subduction of oceanic lithosphere between the converging continents in a region of mantle downwelling located beneath its Gondwanan portion. A combination of processes then causes this region of downwelling to evolve into an LLSVP zone of mantle upwelling that results in plume-related magmatism in the lithosphere located above the margins of the nascent LLSVP.

We provide candidates for Neoproterozoic-Early Palaeozoic magmatism around the Gondwanan margin for which plume activity has either been interpreted or inferred. If these candidates are verified by detailed geochemical and isotopic studies, they would provide further support for Pannotia's mantle legacy as well as a geodynamic explanation for the origins of the late Neoproterozoic and Early Palaeozoic Iapetus and Rheic oceans and the terranes that repeatedly detached from their margins. If verified, this legacy would need to be factored into models for the origin of Pangaea.

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FIGURE CAPTIONS

Figure 1. Various continental reconstructions for the Ediacaran-early Cambrian. (a) c. 545 Ma (Dalziel 1997); (b) c. 600–580 Ma (Cordani et al. 2003a); (c) c. 580 Ma (Meert & Lieberman 2004); (d) c. 600 Ma (Scotese, 2009). (a) A, Arequipa; AM, Amazonia; B, Baltica; C, Congo; D-R-A, Delamarian–Ross arc; ESMT, hypothetical Ellsworth–Sonora–Mojave transform; F/MP, Falkland–Malvinas Plateau; K, Kalahari; MAOT, hypothetical Malvinas–Alabama–Oklahoma transform; R, Rockall; RP, Rio de la Plata; S, Siberia; SF, São Francisco; SV, Svalbard; TxP, hypothetical Texas plateau; WA, West African Craton (Cadomian arc: EA, East Avalonia; WA, West Avalonia). (b) Au, Australia; Ama, Amazonia; Ant, Antarctica; B, Baltica; BTS, Borborema–Trans-Sahara; C-SF, Congo–São Francisco; I, India; K, Kalahari; Lau, Laurentia; M, Madagascar; PA, Pampea; PR, Paraná; RA, Rio Apa; RP, Rio de la Plata; WA, West Africa. (c) Ama, Amazonia; Ant, Antarctica; Ara, Arabia; Arm, Armorica; Au, Australia; Ava, Avalonia; B, Baltica; C, Congo; I, India; K, Kalahari; Lau, Laurentia; RP, Rio de la Plata; SF, São Francisco; S, Siberia; Waf, West Africa. (d) Ama, Amazonia, Afr, Africa; Ant, Antarctica; Ara, Arabia; Au, Australia; B, Baltica; Cc, Cadomia;

Cm, Cimmeria; In, Indochina; Ind, India; NC, North China; S, Siberia; Sam, South America, SC, South China.

Figure 2: Late Neoproterozoic-Early Palaeozoic reconstructions of Merdith et al., 2017. In these reconstructions, final Gondwana amalgamation post-dates rifting, CIMP and opening of the Iapetus Ocean. Shaded grey region in (b) shows their "guide" to the extent of Gondwana at 520 Ma.

Figure 3. Shear wave velocity map at the core-mantle boundary (CMB) (from Becker & Boschi 2002) showing the modern Relationship between LLSVPs and plumes has existed since the amalgamation of Pangaea (after Torsvik et al. 2006). These two LLSVPs are bisected by a subduction girdle of downwelling (blue), as is typical of degree two convection. Plume generation zones are located adjacent to the edges two Large Low Shear wave Velocity Provinces (LLSVPs) along the CMB. Diagram from Torsvik et al. 2006; Burke, 2011).

Figure 4. Schematic diagram adapted from the numerical models of Zhong et al. (2007) and Li and Zhong (2009) showing the transitions from (a) degree-1 planform (supercontinent assembly), (b) stable degree-1 in which a supercontinent assembles above region of mantle downwelling, (c) onset of superplume (manifest as an LLSVP) beneath the supercontinent and degree-2 planform convection with subduction girdle, and (d) mature degree-2 planform antipodal superplumes and breakup of the supercontinent. True polar wander occurs between (c) and (d), restoring superplumes to a more stable equatorial position. Upwelling in yellow, downwelling in blue, continents in black.

Figure 5. Scenario A, using the reconstructions (A= 600 Ma; B = 550 Ma) of Merdith et al. (2017) to showing predicted locations of LLSVPs relative to continents assuming they were quasi-stationary since at least since the Neoproterozoic. The red contour corresponds to the present-day positions of Large Low Shear Velocity Provinces. This low velocity contour is generated from runs where 9 out of 14 seismic tomography models agree that there are slow anomalies (Shephard et al., 2017). On A, the dashed black line corresponds with the location of the LLSVP inferred by Tegner et al. (2019) who fixed the Jason LLSVP to match the location of CIMP.

Figure 6. Scenario B, based on the palaeogeographic reconstruction of Robert et al., (2018) which shows Laurentia in a fixed mantle reference frame as well as inferred subduction zones and LLSVPs assuming degree-2 planform convection. India (I), south China (SC), Siberia (S) and north China (NC)

Figure 7. Scenario C (a-c) evaluated using the reconstructions of Merdith et al. (2017) and (d) from Dalziel (1997) as examples to show the locations of possible plume magmatism around the Gondwanan portion of Pannotia following Pan-African collisional orogenesis. In (a-c) their locations closely match the periphery of Gondwana as shown in the Merdith et al. reconstructions (shaded grey area). (d) Dalziel (1997) is an example of an alternative reconstruction which differs in detail, but shows the same overall characteristic location of plumes around the Gondwanan portion of Pannotia.

1*=CIMP (Laurentia-Baltica-Amazonia; 615-580 Ma); 2*=Voldyn (580-570 Ma); 3=Greendale (607 Ma), 4*=CIMP, West Africa (580 Ma), 5*=Zhulongguan (600 Ma); 6*=CIMP, Wichita (540 Ma); 7=Avalonian basalts (530 Ma); 8=Ossa Morena (510 Ma); 9=Ollo de Sapo (SLIP, 480 Ma); 10=Blovice (530 Ma); 11=Soltan Maiden (450 Ma); 12=La Jishan (520 Ma); 13=SLIP India (500 Ma); 14=Dalamerian (570 Ma), New England Seamounts (480-460 Ma); 15*=Kalkarindji (510 Ma); 16=Kuboos (510 Ma); 17*=Paraupebas-Piranhas (Amazonia, 535 Ma). *=listed in LIP compilation of Ernst et al. (2020).

Figure 8. (a) Location of the plume centre at the Laurentia-Amazonia-Baltica triple junction (from Pisarevsky et al., 2008), (b) Locations of magmatic rocks of the Central Iapetus Magmatic Province CIMP) from Ernst (2014), note the locations of the inferred plume centres lies close to the CIMP plumes shown in Fig. 7.

Figure 9a,**b**. Late Neoproterozoic-Early Palaeozoic stratigraphy of Laurentia (after Spencer et al., 2014; Keeley et al., 2013). Note the paucity of mafic magmatism.

Figure 10. Models for the penetration of plume material through gaps or windows in subducted slabs (after Obrebski et al., 2010; Coble and Mahood, 2010; Druken et al., 2014).

Figure 11. Model for the diachronous propagation of a San-Andreas style transform margin along the northern Gondwanan margin in the Late Neoproterozoic-Early Palaeozoic (after Nance et al, 2002).

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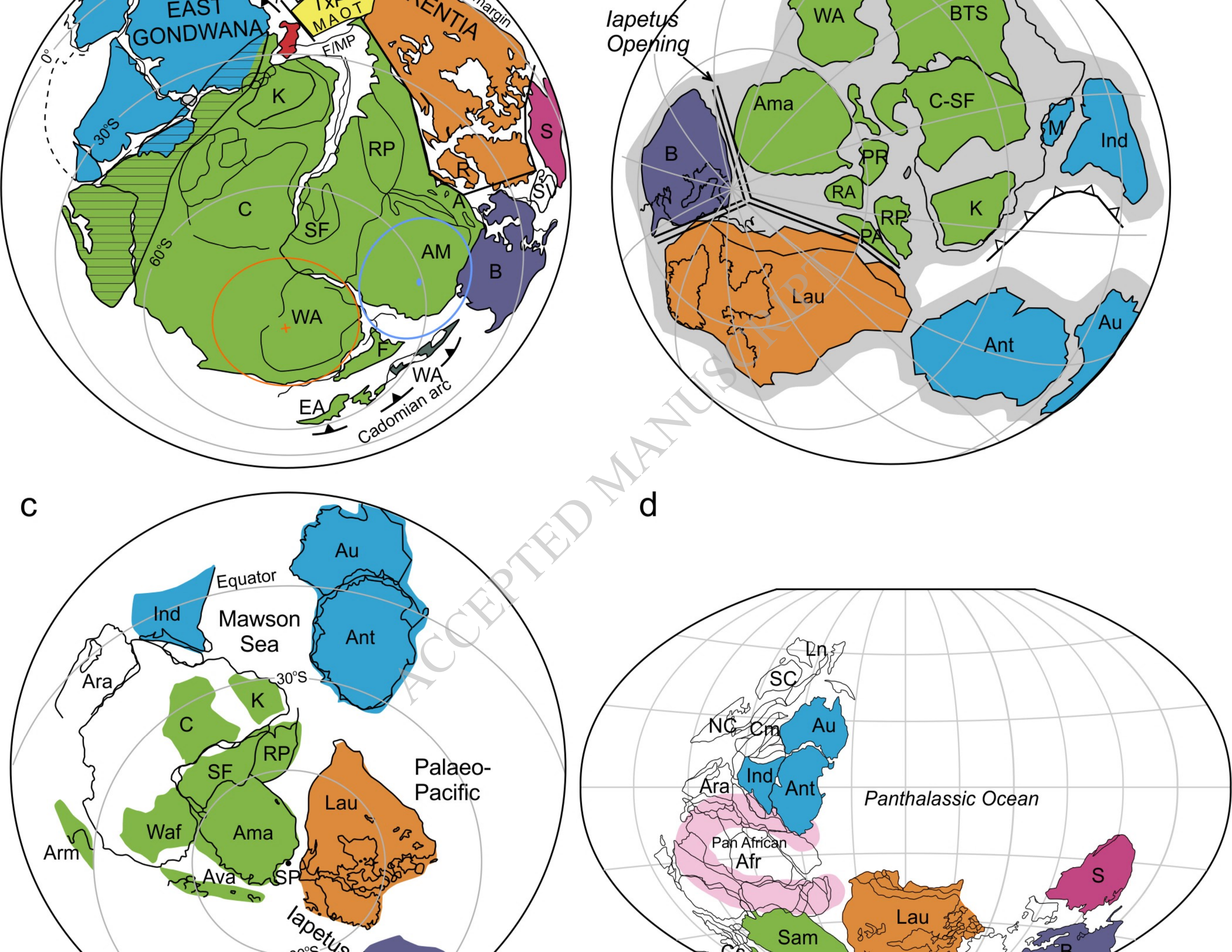
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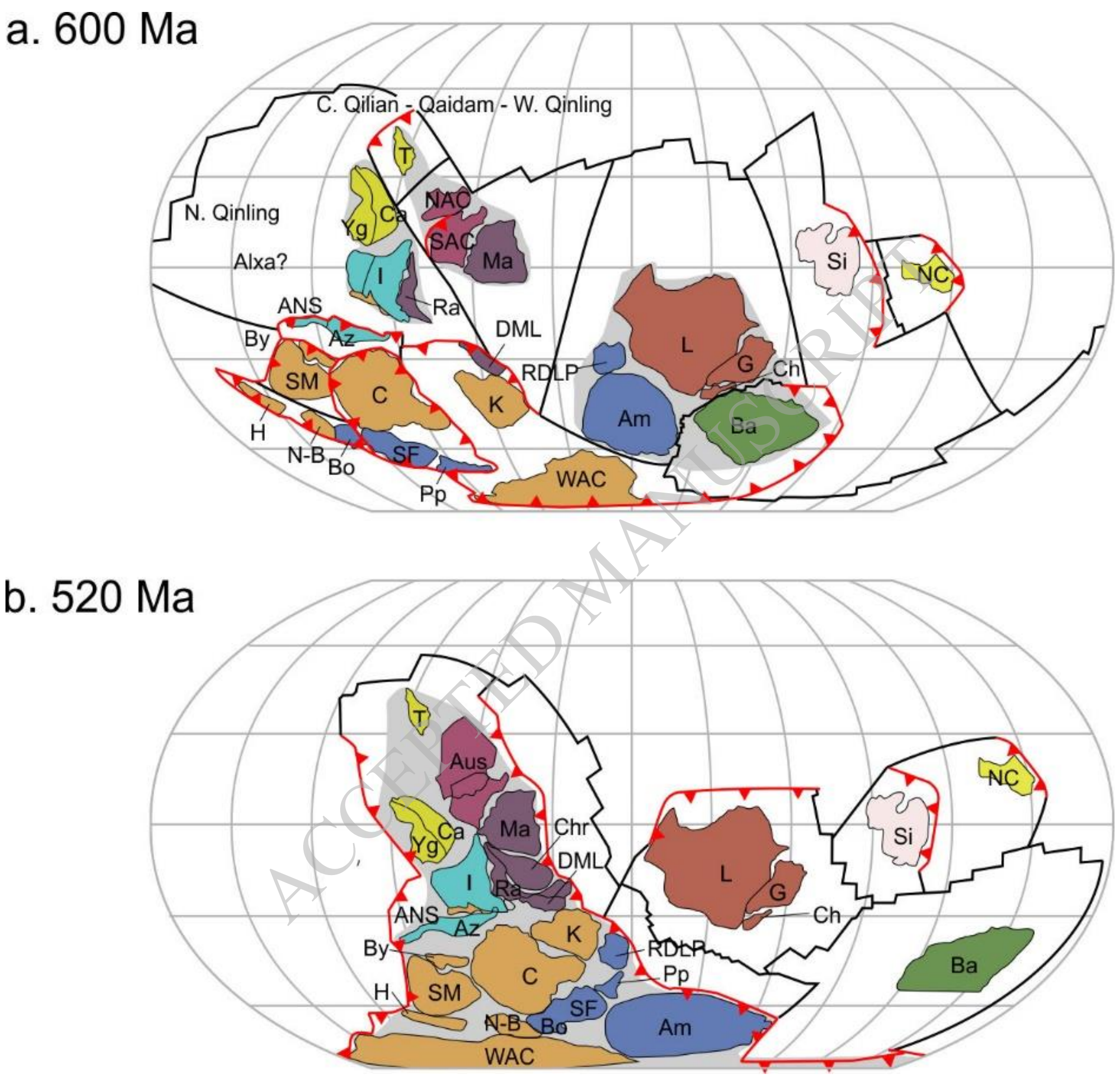
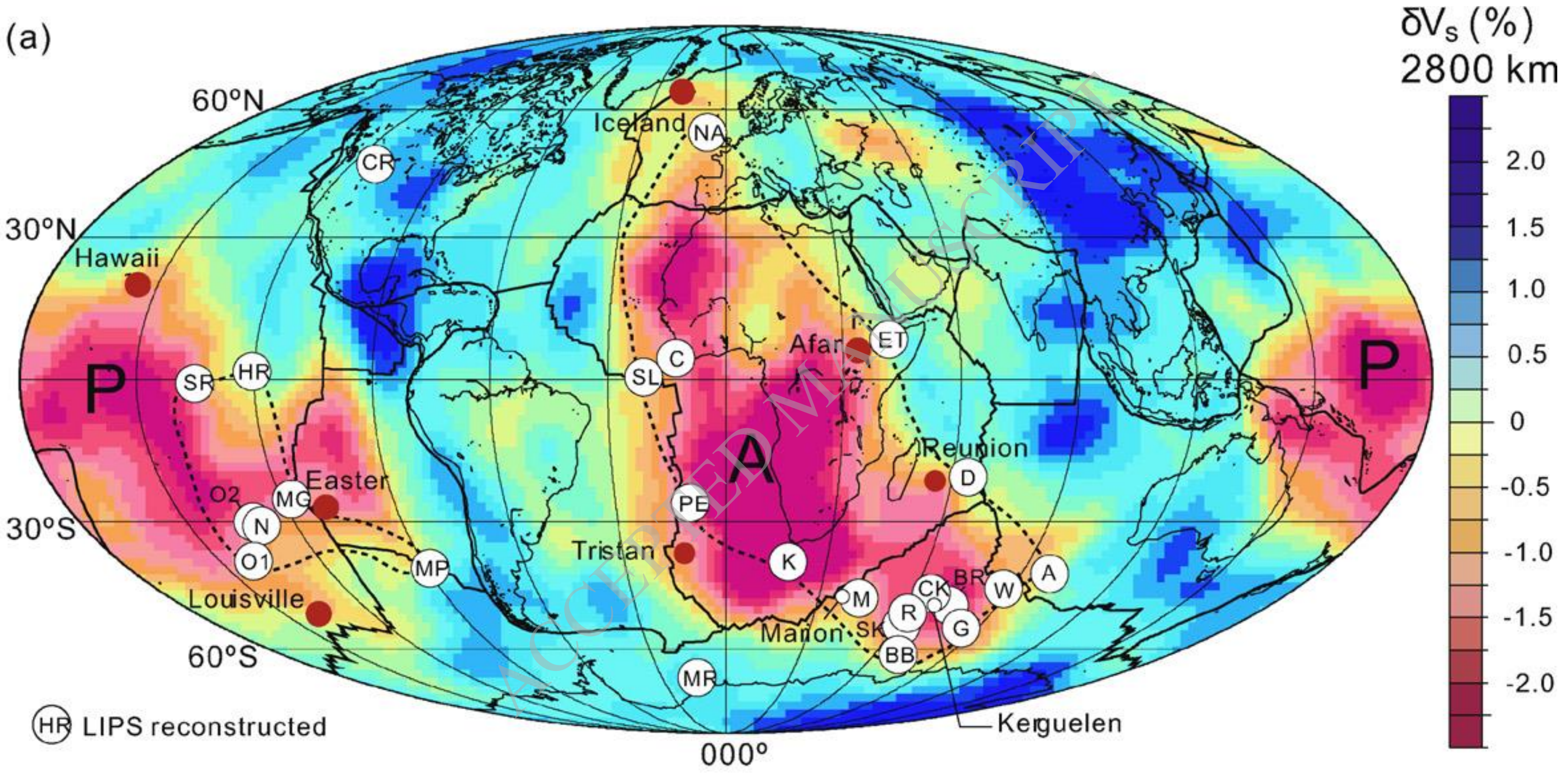
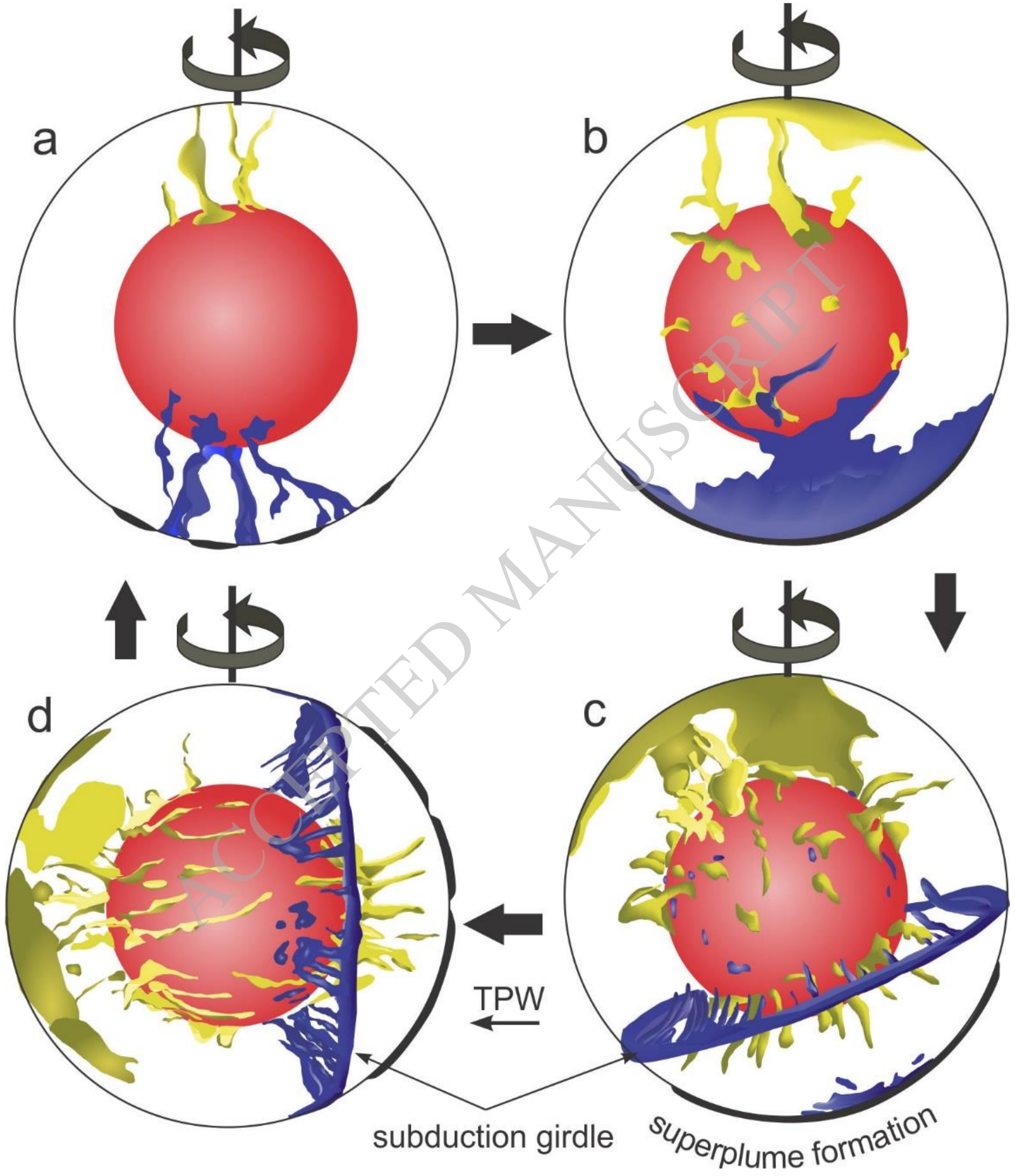

a. 600 Ma
C. Qilian - Qaidam - W. Qinling
N. Qinling
Alxa?
T
Ca
Yg
NAC
SAC
Ma
I
Ra
ANS
By
Az
DML
RDLP
L
G
Ch
SM
C
K
Am
Ba
Si
NC
H
N-B
Bo
SF
Pp
WAC
b. 520 Ma
T
Aus
Ca
Yg
Ma
Chr
DML
I
Ra
ANS
Az
K
By
C
RDLP
Pp
H
SM
N-B
Bo
SF
Am
WAC
L
G
Ch
Si
NC
Ba

Fig 2

Fig. 3


(a)
δVs (%)
2800 km
2.0
1.5
1.0
0.5
0
-0.5
-1.0
-1.5
-2.0
60ºN
30ºN
30ºS
60ºS
000º
Hawaii
Iceland
Afar
Reunion
Easter
Tristan
Louisville
Marion
Kerguelen
P
A
P
CR
NA
SR
HR
C
SL
ET
O2
MG
N
O1
MP
PE
K
D
M
CK
BR
W
A
R
SK
G
BB
MR
HR LIPS reconstructed


a
b
c
d
TPW
subduction girdle
superplume formation

5a

600 Ma

5b

550 Ma

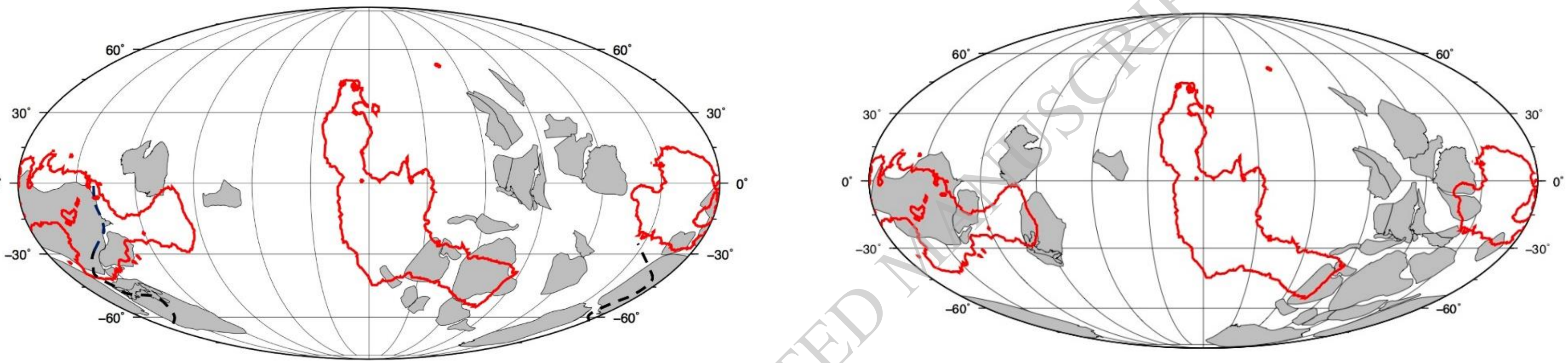


Fig. 5a,b

Model A: Reconstructions after Merdith et al. (2017); LLSVPs assuming quasi-stationary locations since the Neoproterozoic (Torsvik et al., 2016)

Fig 6: Model B
From Fig. 2 of
Robert et al. 2018

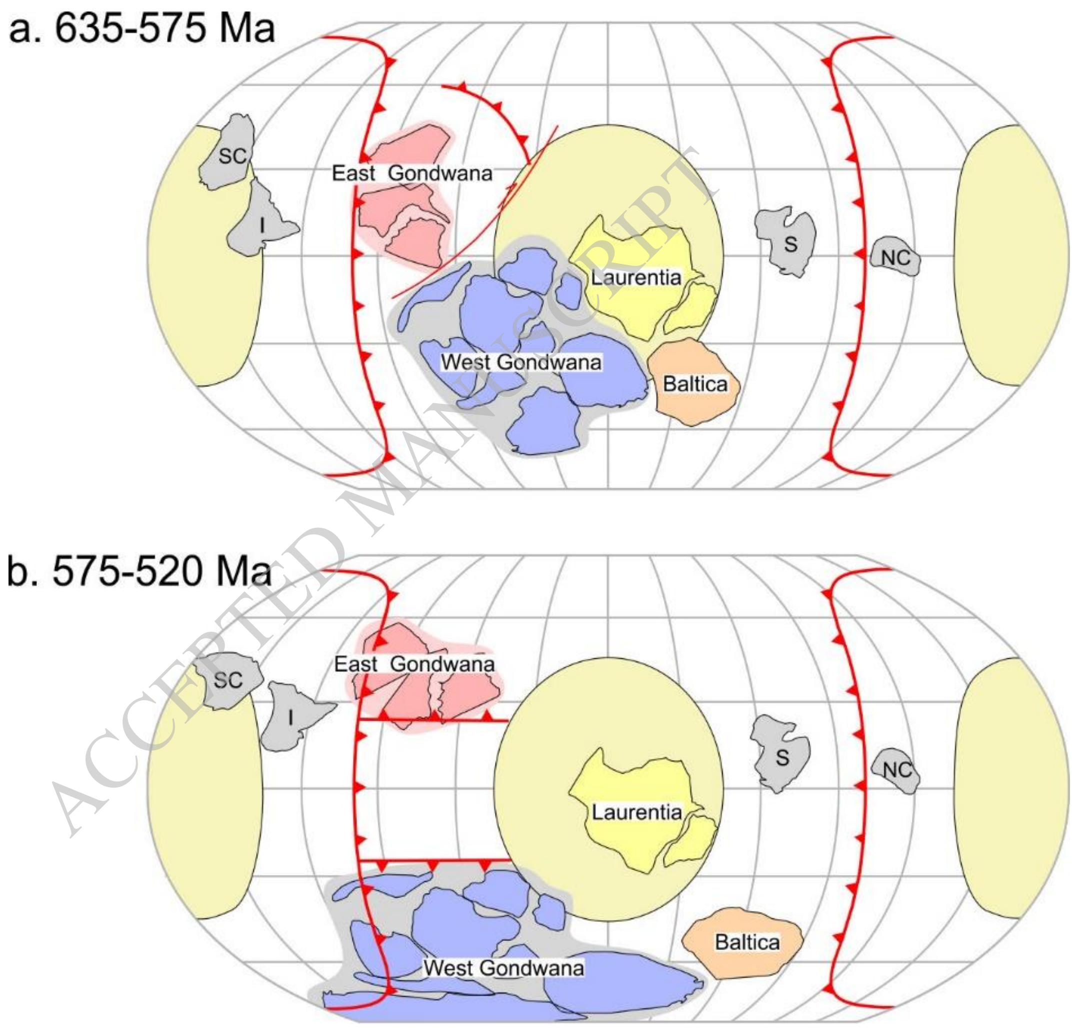
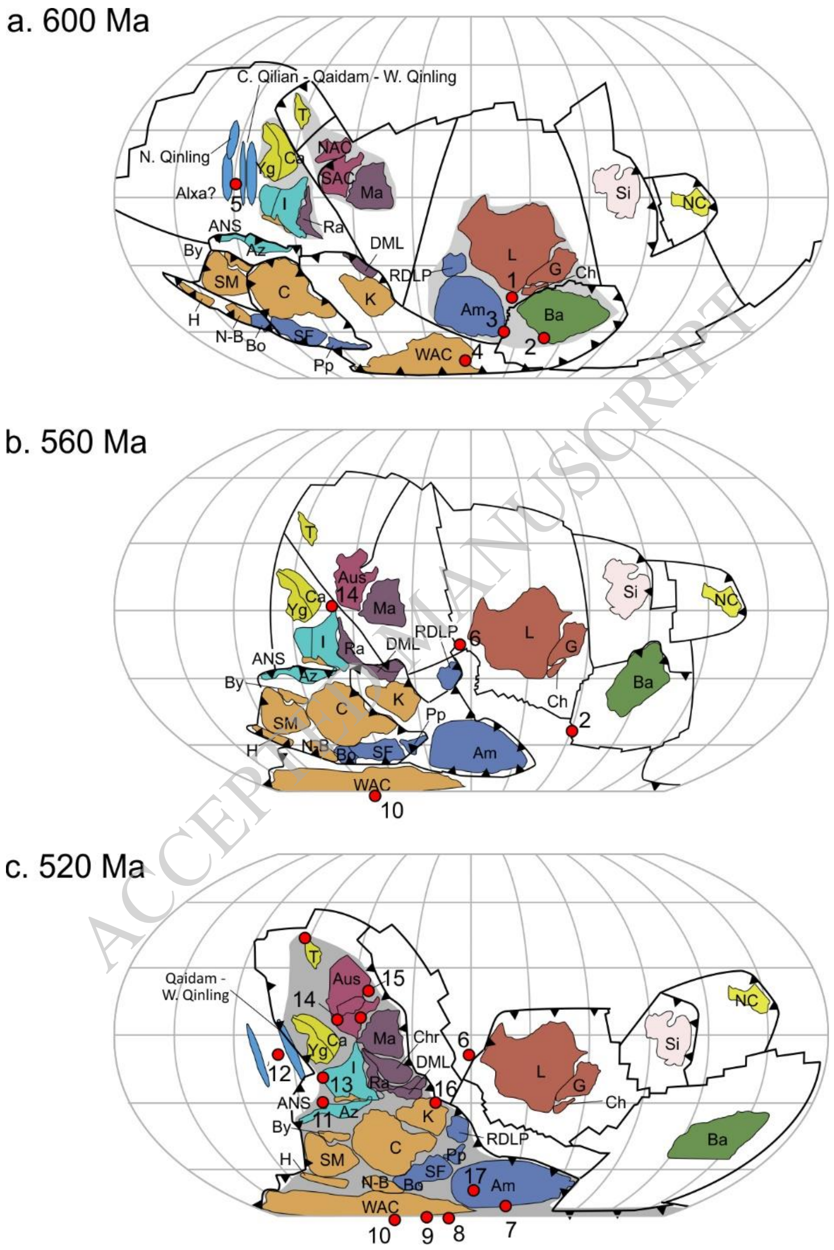


Fig 7a-c



a. 600 Ma
C. Qilian - Qaidam - W. Qinling
N. Qinling
Alxa?
ANS
By
Az
SM
C
H
N-B
Bo
SF
Pp
T
NAC
SAC
Yg
Ca
I
Ma
Ra
DML
RDLP
K
WAC
L
G
Ch
Am
Ba
Si
NC
1
2
3
4
5
b. 560 Ma
T
Aus
Ca
14
Yg
Ma
I
Ra
DML
RDLP
6
ANS
By
Az
C
K
SM
Pp
H
N-B
Bo
SF
WAC
10
Am
L
G
Ch
2
Ba
Si
NC
c. 520 Ma
Qaidam - W. Qinling
T
14
Aus
15
Ca
Yg
Ma
Chr
6
DML
12
13
I
Ra
16
ANS
Az
By
11
K
C
H
SM
Pp
RDLP
N-B
Bo
SF
17
Am
WAC
10
9
8
7
L
G
Ch
Si
NC
Ba

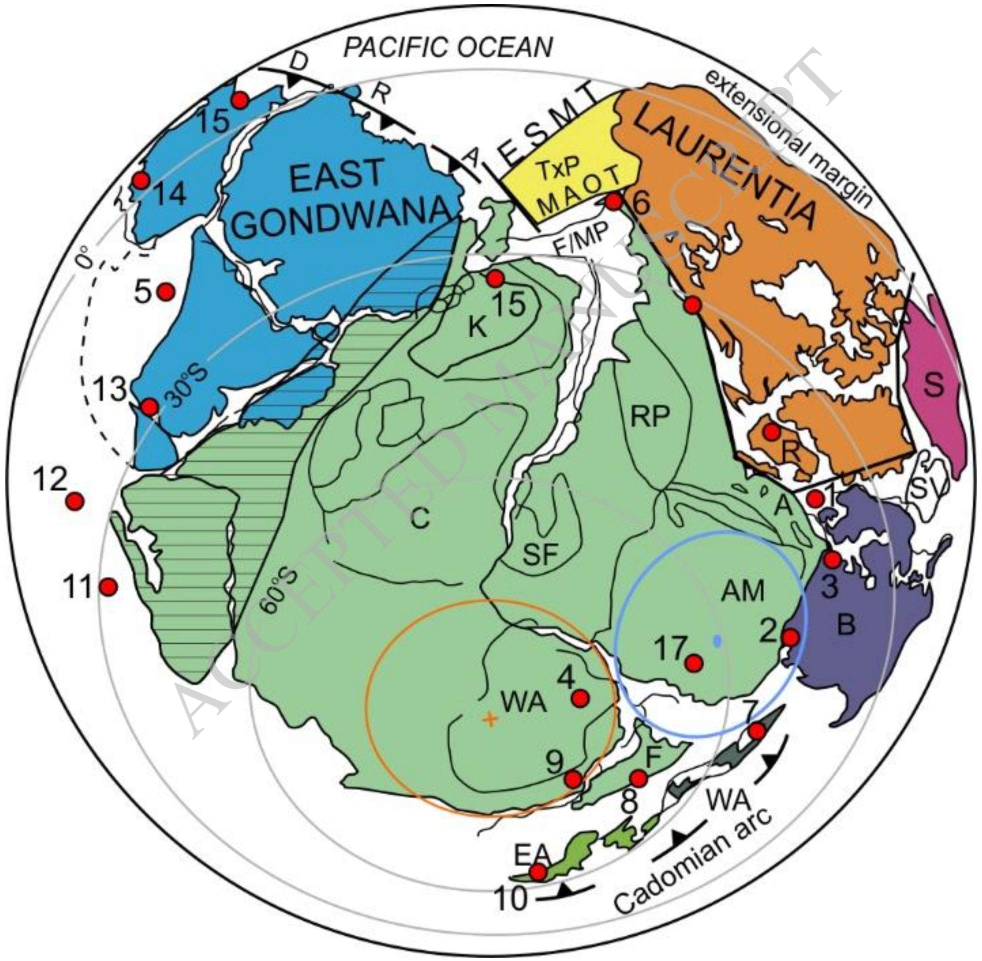
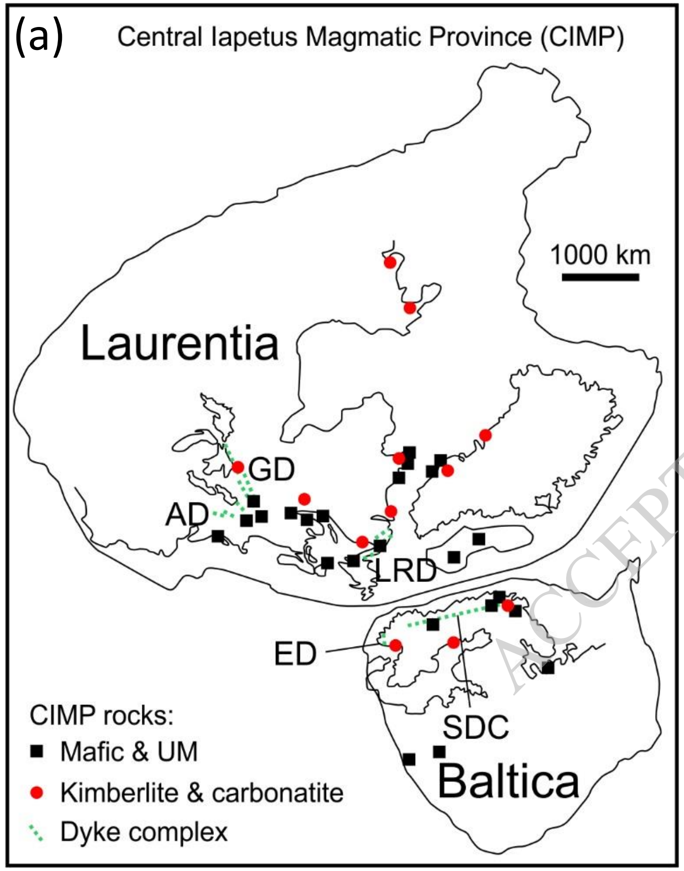

PACIFIC OCEAN
D
R
A
E S M T
TxP
M A O T
LAURENTIA
extensional margin
EAST
GONDWANA
F/MP
15
14
0°
5
15
K
13
30°S
RP
S
R
12
1
A
SV
C
11
SF
3
60°S
AM
B
2
17
4
WA
7
9
F
8
WA
EA
Cadomian arc
10

Fig. 7d


(a)
Central Iapetus Magmatic Province (CIMP)
1000 km
Laurentia
GD
AD
LRD
ED
SDC
Baltica
CIMP rocks:
Mafic & UM
Kimberlite & carbonatite
Dyke complex

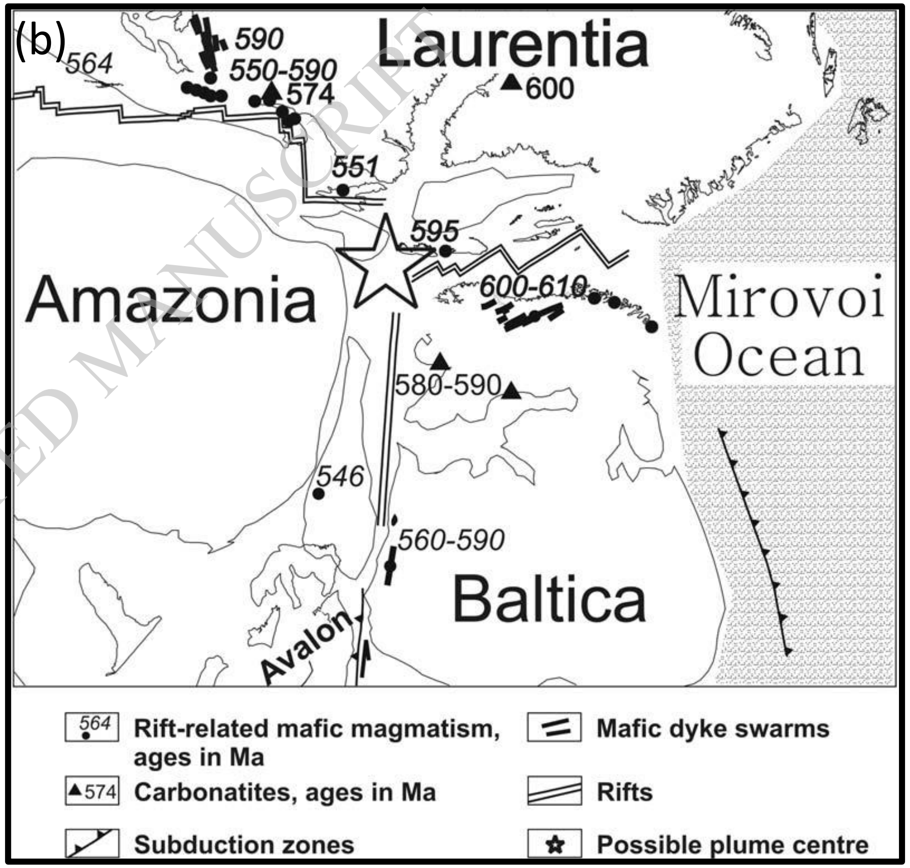

(b)
564
590
550-590
574
Laurentia
600
551
595
600-610
Amazonia
Mirovoi
Ocean
580-590
546
560-590
Baltica
Avalon
Rift-related mafic magmatism, ages in Ma
Carbonatites, ages in Ma
Subduction zones
Mafic dyke swarms
Rifts
Possible plume centre

Fig. 9A

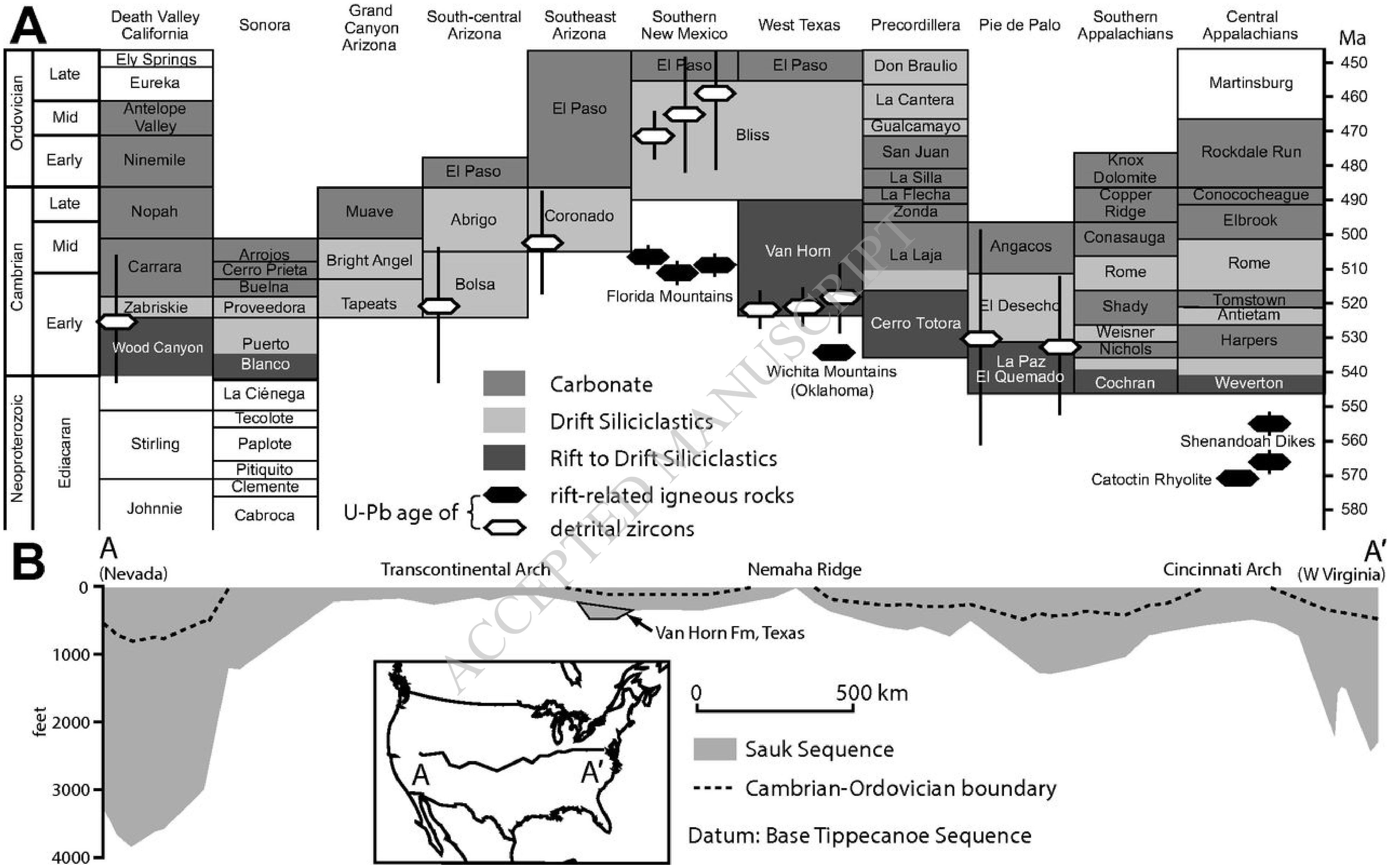
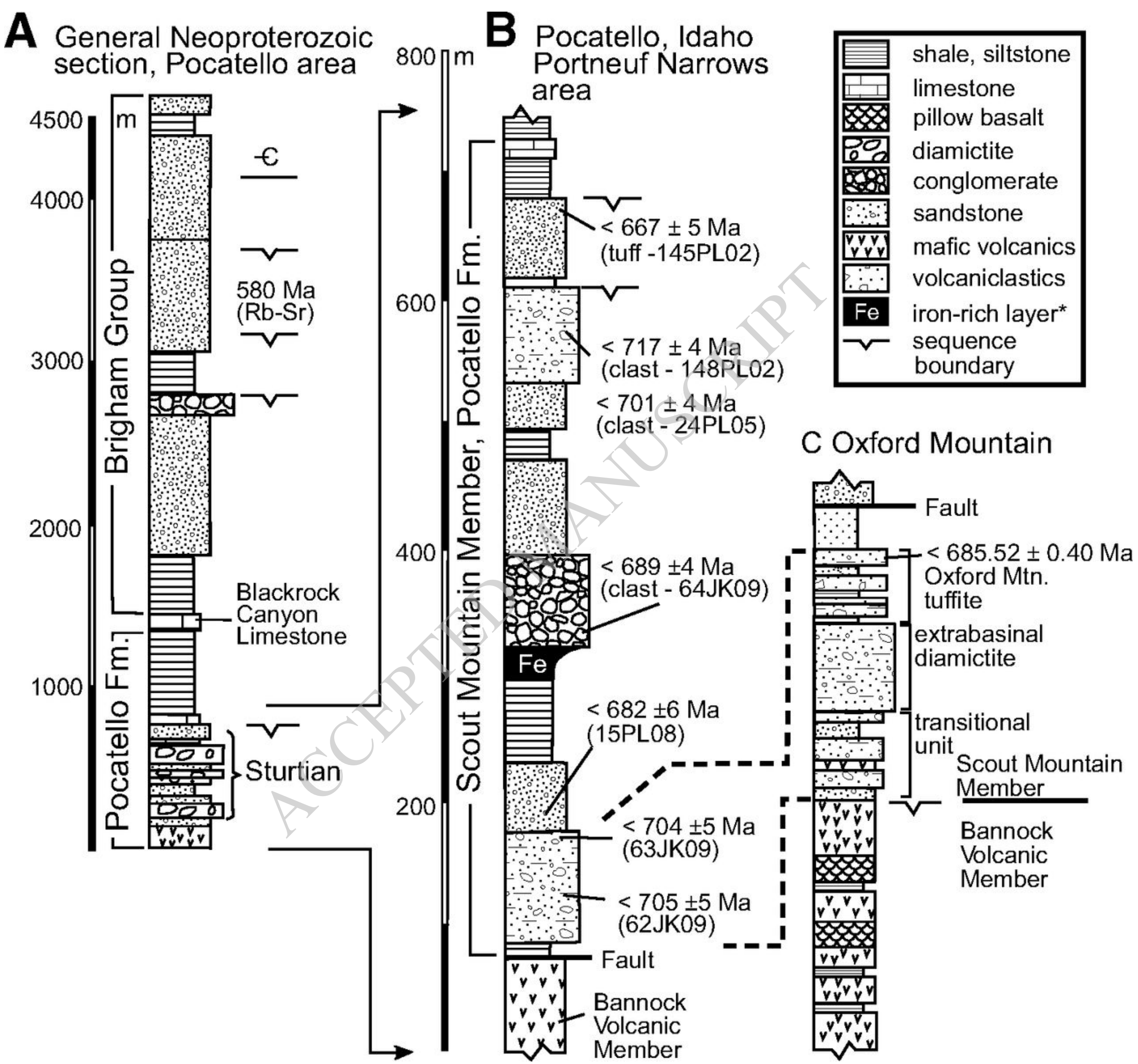

A
Death Valley California
Sonora
Grand Canyon Arizona
South-central Arizona
Southeast Arizona
Southern New Mexico
West Texas
Precordillera
Pie de Palo
Southern Appalachians
Central Appalachians
Ma
Ordovician
Late
Mid
Early
Cambrian
Late
Mid
Early
Neoproterozoic
Ediacaran
Ely Springs
Eureka
Antelope Valley
Ninemile
Nopah
Carrara
Zabriskie
Wood Canyon
Stirling
Johnnie
Arrojos
Cerro Prieta
Buelna
Proveedora
Puerto Blanco
La Ciénega
Tecolote
Paplote
Pitiquito
Clemente
Cabroca
Muave
Bright Angel
Tapeats
El Paso
Abrigo
Bolsa
El Paso
Coronado
El Paso
Bliss
Florida Mountains
El Paso
Van Horn
Wichita Mountains (Oklahoma)
Don Braulio
La Cantera
Gualcamayo
San Juan
La Silla
La Flecha
Zonda
La Laja
Cerro Totora
Angacos
El Desecho
La Paz El Quemado
Knox Dolomite
Copper Ridge
Conasauga
Rome
Shady
Weisner
Nichols
Cochran
Martinsburg
Rockdale Run
Conococheague
Elbrook
Rome
Tomstown
Antietam
Harpers
Weverton
Shenandoah Dikes
Catoctin Rhyolite
450
460
470
480
490
500
510
520
530
540
550
560
570
580
Carbonate
Drift Siliciclastics
Rift to Drift Siliciclastics
U-Pb age of
rift-related igneous rocks
detrital zircons
B
A (Nevada)
Transcontinental Arch
Nemaha Ridge
Cincinnati Arch
A′ (W Virginia)
0
1000
2000
3000
4000
feet
Van Horn Fm, Texas
A
A′
0
500 km
Sauk Sequence
Cambrian-Ordovician boundary
Datum: Base Tippecanoe Sequence

Fig. 9b



A General Neoproterozoic section, Pocatello area
m
4500
4000
3000
2000
1000
Brigham Group
Pocatello Fm.
-Є
580 Ma (Rb-Sr)
Blackrock Canyon Limestone
Sturtian
B Pocatello, Idaho Portneuf Narrows area
800 m
600
400
200
Scout Mountain Member, Pocatello Fm.
< 667 ± 5 Ma (tuff -145PL02)
< 717 ± 4 Ma (clast - 148PL02)
< 701 ± 4 Ma (clast - 24PL05)
< 689 ±4 Ma (clast - 64JK09)
Fe
< 682 ±6 Ma (15PL08)
< 704 ±5 Ma (63JK09)
< 705 ±5 Ma (62JK09)
Fault
Bannock Volcanic Member
shale, siltstone
limestone
pillow basalt
diamictite
conglomerate
sandstone
mafic volcanics
volcaniclastics
Fe iron-rich layer*
sequence boundary
C Oxford Mountain
Fault
< 685.52 ± 0.40 Ma Oxford Mtn. tuffite
extrabasinal diamictite
transitional unit
Scout Mountain Member
Bannock Volcanic Member

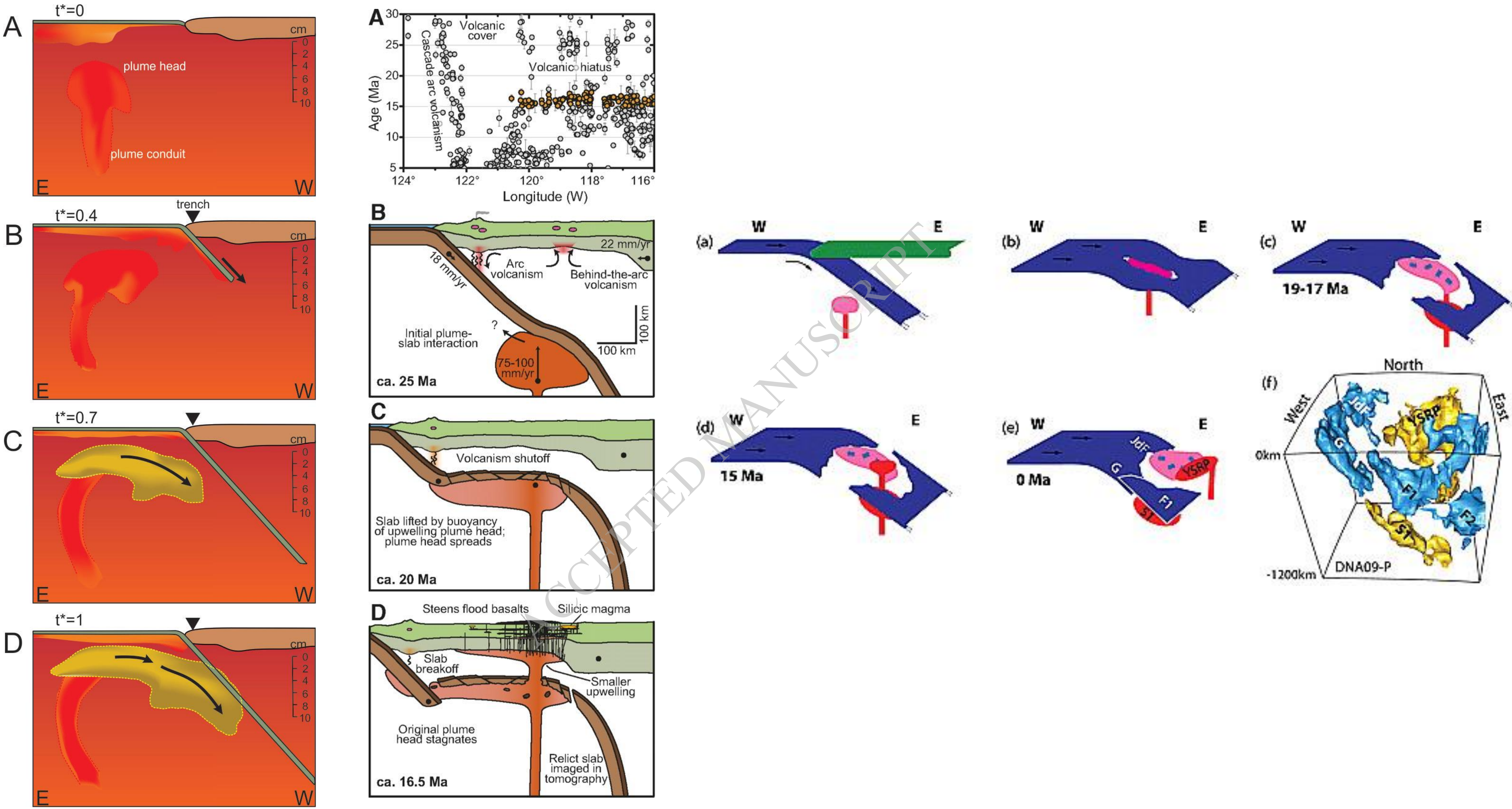
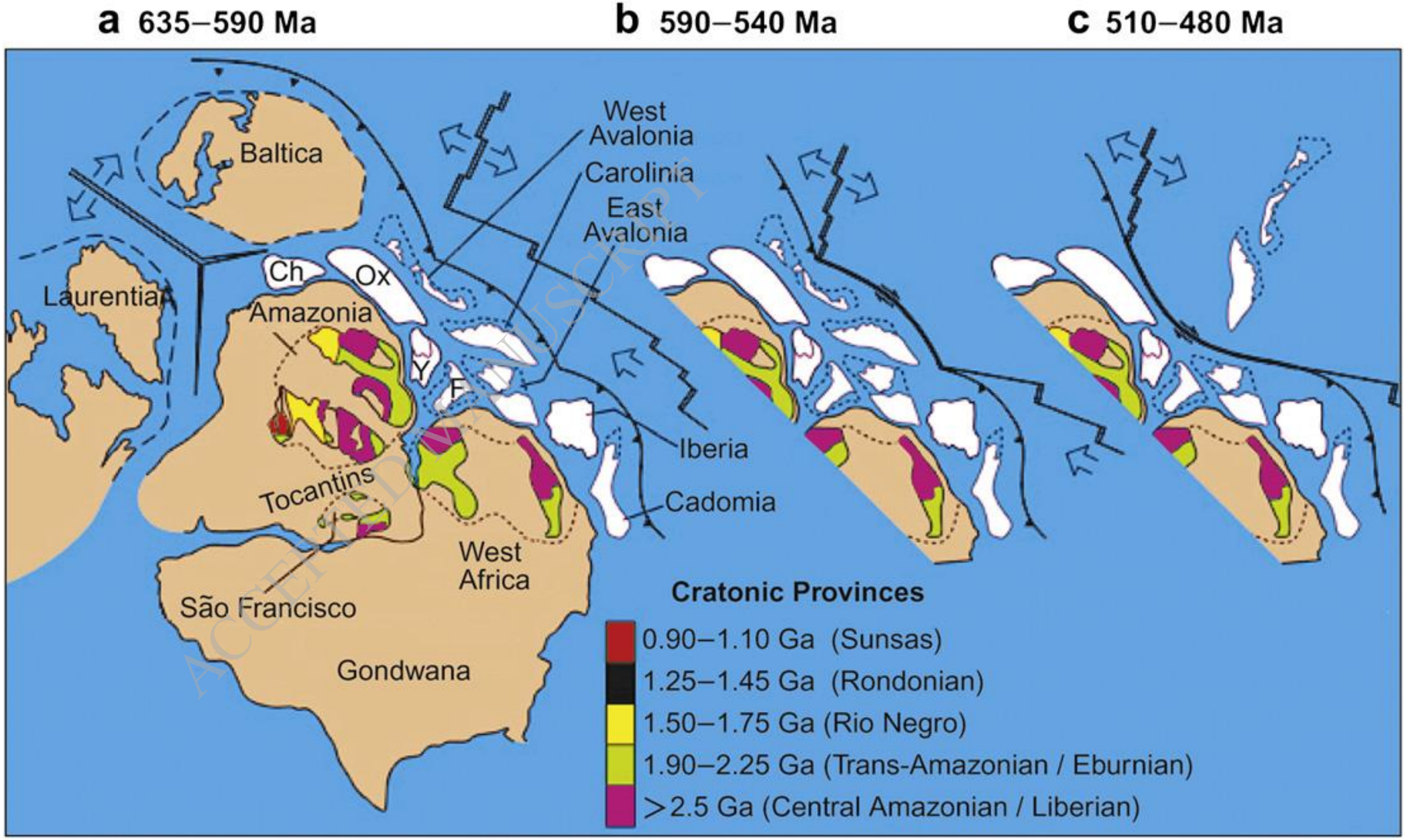

A
t*=0
cm
0
2
4
6
8
10
plume head
plume conduit
E
W
B
t*=0.4
trench
C
t*=0.7
D
t*=1
A
Age (Ma)
Longitude (W)
124°
122°
120°
118°
116°
Volcanic cover
Cascade arc volcanism
Volcanic hiatus
B
22 mm/yr
18 mm/yr
Arc volcanism
Behind-the-arc volcanism
100 km
Initial plume-slab interaction
75-100 mm/yr
ca. 25 Ma
C
Volcanism shutoff
Slab lifted by buoyancy of upwelling plume head; plume head spreads
ca. 20 Ma
D
Steens flood basalts
Silicic magma
Slab breakoff
Smaller upwelling
Original plume head stagnates
Relict slab imaged in tomography
ca. 16.5 Ma
(a)
(b)
(c)
19-17 Ma
(d)
15 Ma
(e)
0 Ma
JdF
G
F1
S1
YSRP
(f)
North
West
East
0km
-1200km
DNA09-P
F2

Fig. 10

Fig. 11

Peri-Gondwanan Terranes



Nance et al., 2002