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1 **Extending Full-Plate Tectonic Models into Deep Time: Linking the Neoproterozoic and the**
2 **Phanerozoic**

3
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17
18 **Abstract**

19
20 Recent progress in plate tectonic reconstructions has seen models move beyond the classical idea of
21 continental drift by attempting to reconstruct the full evolving configuration of tectonic plates and plate
22 boundaries. A particular problem for the Neoproterozoic and Cambrian is that many existing
23 interpretations of geological and palaeomagnetic data have remained disconnected from younger, better-
24 constrained periods in Earth history. An important test of deep time reconstructions is therefore to
25 demonstrate the continuous kinematic viability of tectonic motions across multiple supercontinent cycles.
26 We present, for the first time, a continuous full-plate model spanning 1 Ga to the present-day, that
27 includes a revised and improved model for the Neoproterozoic–Cambrian (1000–520 Ma) that connects
28 with models of the Phanerozoic, thereby opening up pre-Gondwana times for quantitative analysis and
29 further regional refinements. In this contribution, we first summarise methodological approaches to full-
30 plate modelling and review the existing full-plate models in order to select appropriate models that
31 produce a single continuous model. Our model is presented in a palaeomagnetic reference frame, with a
32 newly-derived apparent polar wander path for Gondwana from 540 to 320 Ma, and a global apparent
33 polar wander path from 320 to 0 Ma. We stress, though while we have used palaeomagnetic data when
34 available, the model is also geologically constrained, based on preserved data from past-plate boundaries.
35 This study is intended as a first step in the direction of a detailed and self-consistent tectonic

36 reconstruction for the last billion years of Earth history, and our model files are released to facilitate
37 community development.

38

39 **1 Introduction**

40

41 Plate tectonics is a unifying theory of modern geology, explicitly connecting the evolution and processes
42 that bridge the mantle, lithosphere, hydrosphere and atmosphere. Tectonic forces control the rates of
43 uplift and erosion where continents collide or separate (England and Molnar, 1990) and modulate the flow
44 of energy between oceans, lithosphere and mantle as continental configurations evolve (Bebout, 1995;
45 Karlsen et al., 2019; Müller et al., 2008). Evolving plate tectonic configurations also determine changes in
46 how species are distributed across different landmasses (McKenzie et al., 2014; Meert and Lieberman,
47 2008) and infer the rates of chemical flux between the Earth's surface and the deep interior (Gernon et al.,
48 2016; Jarrard, 2003).

49

50 Global reconstructions have traditionally focussed on the positions of the major continents and geological
51 terranes preserved within them. Data acquired from modern oceans provide a powerful constraint on the
52 breakup of the supercontinent Pangea over the last ca. 200 Ma, and form the basis of continuous models
53 of plate configurations from the Mesozoic to present (e.g. Müller et al., 2016; Seton et al., 2012). These
54 'full-plate' reconstructions use geological and geophysical data to determine the configurations and
55 motions of both continental and oceanic lithosphere, and the nature of the plate boundaries that separate
56 neighbouring plates. Together with the development of free software tools (Boyden et al., 2011; Müller et
57 al., 2018), full-plate reconstructions permit quantitative estimates of tectonic processes through time
58 within a continuous, consistent kinematic framework, opening up portions of Earth's history to
59 quantitative analysis (e.g. Bower et al., 2013; Brune et al., 2017; Dutkiewicz et al., 2019; Hounslow et al.,
60 2018; Karlsen et al., 2019; Meredith et al., 2019a).

61

62 Plate tectonic processes are thought to have been the dominant control on Earth's paleogeography
63 possibly since 3.2 Ga (Brenner et al., 2020; Brown et al., 2020a; Cawood et al., 2018a; Gerya, 2014; Palin
64 et al., 2020). Studies of the pre-Pangean Earth have led to the proposal that Pangea was preceded by the
65 Proterozoic supercontinents Rodinia (Dalziel, 1991; Hoffman, 1991; Moores, 1991) and Nuna/Columbia
66 (Meert, 2002; Rogers and Santosh, 2002; Zhao et al., 2002) and earlier Archaean 'supercratons' (e.g.
67 Bleeker, 2003; Pehrsson et al., 2013; Smirnov et al., 2013), reflecting transient aggregations of
68 continental blocks interspersed between other phases of Earth's history when the continents were more
69 dispersed. The absence of a pre-Mesozoic ocean floor record necessitates that reconstructing the pre-

70 Pangean Earth relies on the fragmented geological record preserved within the continents. Early studies of
71 Proterozoic supercontinents provide individual snapshots of continental configurations; though there are
72 differences between competing interpretations. More recently, attempts have been made to reconcile
73 Neoproterozoic continental motions within a continuous kinematic framework (Cawood et al., 2020;
74 Collins and Pisarevsky, 2005; Li et al., 2008). To further infer the extent and nature of tectonic
75 boundaries covering all of Earth's surface in the Proterozoic requires methodical extrapolation of
76 available observations and is subject to major uncertainties. Despite this, these reconstructions are
77 valuable in that they make testable predictions about regions and time periods where observations are
78 lacking.

79

80 Full-plate models published over the last decade collectively span the last 1 Ga. However, each of these
81 models cover different time periods or areas of the world and each model is based on different
82 assumptions and hypotheses, and place differing emphases on subsets of the geological record. Thus,
83 although continental motions and plate boundary evolution have been categorised in some manner for the
84 past 1 Ga, there is no fully continuous model defining Earth's tectonic history for this time. A
85 fundamental test of any tectonic reconstruction for the Precambrian is that the configurations of
86 continents, terranes and plate boundaries can evolve continuously as to seamlessly merge with
87 reconstructed configurations for more recent times that are better constrained and ultimately tied to the
88 present-day Earth. The absence of such continuous reconstructions highlights a critical uncertainty for
89 assessing interpretations of Neoproterozoic palaeogeography, tectonics and geodynamics.

90

91 Our key motivations for this study are three-fold. Firstly, a 1 Ga model will permit, for the first time,
92 Neoproterozoic and Cambrian quantitative analysis that constrains (bio)geochemical and volatile fluxes,
93 palaeoclimatic studies and the nature of earth systems, during times of biological evolution and extreme
94 climate change (Gernon et al., 2016; Godd ris et al., 2017; Mills et al., 2011, 2019). Second, a full-plate
95 model would be a starting point for future studies to constrain both the tectonic (e.g. supercontinent cycle
96 (Li et al., 2019; Merdith et al., 2019b)) and geodynamic (e.g. core-lithosphere-mantle connection (Heron
97 et al., 2020; Tetley et al., 2019)) nature and evolution of the Earth. Third, a consistent model for the
98 Neoproterozoic and Cambrian that coherently links with younger models can be used as a framework to
99 support future regional studies that test and enhance the resolution of the model or spawn alternative
100 models that can be used for hypothesis testing. We stress that our reconstruction is intended to capture the
101 main aspects of global tectonics across the last billion years and consequently lacks many details that
102 could be incorporated for individual regions. Just as the earliest full-plate models for the Cenozoic and
103 Mesozoic (Seton et al., 2012) and late Palaeozoic (Domeier and Torsvik, 2014) have provided valuable

104 open-access resources for numerous other studies to test and improve, we intend that the global
105 framework provided by our reconstruction will form the basis for future studies that will generate
106 improved reconstructions by incorporating new or different observations and ideas.

107

108 This paper is organised as follows: first, we provide a review of the concepts behind full-plate
109 reconstructions, including the types of observations and assumptions on which they are based. We then
110 summarise the previously published reconstruction models and justify which elements of these existing
111 studies we have chosen to include in our reconstruction. Finally, we present the main outcome of our
112 study: the first continuous and self-consistent full-plate model from 1 Ga to present day with a single set
113 of polygons, Euler rotations and plate boundaries.

114

115 **2 Full-plate reconstruction models**

116 There are two broad categories of models that can be constructed to describe Earth's tectonic or
117 palaeogeographic history. The first category we refer to as 'continental drift' type models (Fig. 1a), as
118 they model the motion of continents drifting across the Earth's surface and tend to explicitly reflect
119 palaeogeography rather than tectonic evolution. The second type we refer to as 'full-plate models' (Fig.
120 1b), which, in addition to tracking the motion of continents, trace the evolution of plate boundaries and by
121 implication, the evolution of tectonic plates themselves (Gurnis et al., 2012). In effect, continental drift
122 type models are the precursor to full-plate models, but rather than supersede continental drift models, both
123 types of models complement each other and provide different avenues for research. Continental drift type
124 models are useful for analysing palaeomagnetic data, for contextualising regional studies or as a ground-
125 breaking study where there is little preserved data on plate boundaries. Comparably, full-plate models are
126 more encompassing, but are also much harder to iterate over and generate alternative models from. If one
127 simply requires the distribution of continental crust and not of plate boundaries, then it is much easier and
128 simpler to build a continental drift style model than categorically describe and model plate boundaries
129 through time. However, both types of models use the same reconstruction framework.

130

131 **2.1 Reconstruction Framework**

132

133 The essential characteristic of any plate reconstruction is the reconstruction framework (or network
134 (Domeier and Torsvik, 2017)), which is the organisation of data used to describe the motion of rigid
135 objects on a sphere using Euler's rotation theorem (McKenzie and Parker, 1967; Morgan, 1968). In plate
136 reconstructions, the rigid objects in motion are the plates themselves and, in addition to requiring the
137 temporal and spatial components of moving plates (i.e. the time period of motion, latitude, longitude and

138 angle of rotation) the theorem also requires that each rotation be defined relative to another object. This
139 rotation of one object relative to another forms the basis of a 'relative plate motion model'. It is also
140 desirable for these relative rotations to be tied to something (relatively) immutable within, or around, the
141 Earth (e.g. the core, mantle, or the spin-axis) thus, transforming the model into an 'absolute plate motion
142 model'. For the Mesozoic and Cenozoic, plates are described in a relative framework due to the
143 preservation of oceanic lithosphere within oceans formed since the breakup of Pangea. In this way each
144 plate's motion history is described as moving relative to another plate, with the African plate typically at
145 the top of the hierarchy (e.g. (Ross and Scotese, 1988; Torsvik et al., 2008); Fig. 2a, b). The motion of the
146 African plate can then be defined absolutely (though using observations from many or all plates, and not
147 just Africa) to the deep Earth through alternative methods such as hotspot chains (Müller et al., 1993;
148 O'Neill et al., 2005), seismic imaging of subducted slabs (van der Meer et al., 2010), palaeomagnetic data
149 (with or without true polar wander corrections, Torsvik et al., 2012) or methods jointly evaluating the
150 characteristics of multiple constraints including plate velocities, hotspot chains and subduction trench
151 migration (Tetley et al., 2019). Regardless of which method is chosen, the result is a global reference
152 frame defined as a sequence of absolute motions of the African plate, which, together with the relative
153 motions between plate-pairs arranged within a hierarchy (Fig. 2), define the absolute motions of all plates.
154 An exception to all plates being tied to Africa occurs for the Mesozoic Pacific Ocean. Before 83 Ma (and
155 the opening of the West Antarctic spreading centre) the motion of the Pacific plate is preserved and
156 reconstructed absolutely to the spin axis through hotspot motion, rather than through Africa. Plate models
157 can also be described in a purely relative framework, in which case a single continent (or plate) is fixed to
158 its present-day position and all other continents or plates are rotated relative to the fixed plate. This
159 approach is commonly used for localised studies or to easily highlight the difference between two
160 contrasting models (Fig. 2c).

161

162 Before the Mesozoic era, it is not possible to use preserved, *in situ* oceanic lithosphere, hotspot motions
163 and seismic imaging of slabs. Therefore, the logical arrangement of connections within the rotational
164 framework changes (Domeier and Torsvik, 2017), however the general principles of plate reconstructions
165 remain the same. For these times, the only quantitative information on the positions of plates is through
166 palaeomagnetism, which describes the palaeolatitude of continents with respect to the Earth's spin axis. In
167 these cases, where the absolute motions are more directly constrained than relative motions, rotation
168 models traditionally favour a simpler hierarchy, rather than the complex hierarchies used for post-Pangea
169 times (e.g Fig. 3a, b). The motion of major continents relative to the spin axis is determined using their
170 own palaeomagnetic data (Fig. 3a, b). Within this data set, continents can be grouped together in localised

171 hierarchies where there is evidence that they have remained together or close to one another or, in
172 instances where paleomagnetic data are lacking from some of the blocks if geological constraints permit.

173

174 Finally, although Euler's rotation theorem is based on the motion of rigid bodies, this is a simplification
175 because the lithosphere is deformable. Recent advancements in plate modelling (Gurnis et al., 2018) have
176 allowed for the development of deforming plate models where rigid plates are able to deform along their
177 edges.

178

179 **2.2 Palaeolatititude**

180

181 Palaeolatititude is determined through the study of palaeomagnetic data and is the only method to
182 quantitatively constrain the absolute latitudinal position of a continent for pre-Jurassic times. If sufficient
183 data are present from a single continent, an apparent polar wander path (APWP) can be constructed that
184 describes the motion of that continent through time (Torsvik et al., 2012). If there are very good
185 geological constraints on the relative positioning between multiple continents, then poles from these
186 continents can be merged to form composite apparent polar wander paths. This merging is done by
187 rotating the poles of multiple continents into the coordinate frame of a single continent, typically found
188 higher in the framework (e.g. Africa, Fig. 2a). If this process is done using all global data it is known as a
189 global apparent polar wander path (GAPWaP). GAPWaPs provide the potential for a more rigorous
190 description of the evolution of a suite of continents as more data are available.

191

192 There are, however, many caveats and uncertainties associated with GAPWaPs and APWPs that contain
193 data sourced from more than one continent (e.g. APWP for Gondwana). In particular, they are strongly
194 dependent on the relative position of continents, as even minor changes in these relative positions can
195 result in large differences in the resulting wander paths. They also are directly dependent on the quality
196 and abundance of palaeomagnetic data, thus a degree of subjectivity can be introduced by what criteria
197 are used for selecting and filtering poles (Van der Voo, 1990; see a new approach in Wu et al., 2020). For
198 example, in the Gondwana APWP of Torsvik et al. (2012) (and also for the APWP we construct in this
199 paper), there is one pole for Gondwana between 440 and 400 Ma that constrains the motion of over half
200 of all known continental crust at the time. In the Precambrian, the geological uncertainty of exactly how
201 two cratons (or continents) fit together limits the usefulness of GAPWaPs and APWPs that are defined by
202 multiple continents in conjunction with one another. Instead individual APWPs can be constructed using
203 available palaeomagnetic data for continents and then these are all balanced together globally to produce

204 a coherent kinematic continental model (e.g. Li et al., 2008; Pisarevsky et al., 2014). In this manner, it is
205 possible to build continental drift models purely from palaeomagnetic data.

206

207 **2.3 True Polar Wander**

208

209 True polar wander (TPW) is the motion of the entire solid Earth (mantle-lithosphere) with respect to the
210 spin axis due to centrifugal forces from Earth's orbit acting on mass anomalies in the upper mantle,
211 wherein positive anomalies are driven towards the equator and negative anomalies towards the poles
212 (Evans, 2003). Since TPW is inherent in palaeomagnetic data, all APWP are a composite of both plate
213 motions and some component of TPW. Thus, in the strictest sense, to properly use a plate model for
214 geodynamic modelling, a correction that removes any component of detected TPW should be applied (i.e.
215 the mantle reference frame). Further, as the mass anomalies in the mantle are thought to arise from the
216 flow induced by subducted oceanic lithosphere and the associated return flows, TPW excursions are
217 closely linked to the supercontinent cycle (Zhong et al., 2007).

218

219 Raub et al. (2007) identified three types of TPW summarised briefly below, though only the latter two are
220 relevant here. Type 0 TPW operates on short ($< 10^3$ a) timescales as a response to elastic deformation
221 within the lithosphere arising from seismic events (Soldati et al., 2001) and so has a negligible effect on
222 plate motions on the timescales pertinent here (Evans, 2003; Raub et al., 2007). Type 1 TPW is the most
223 important to consider, and is broadly defined as the slow motion of the solid Earth (mantle and
224 lithosphere) readjusting to mass anomalies in the mantle. Type 1 TPW is what is commonly detected and
225 corrected in plate motion models (Steinberger and Torsvik, 2008; Torsvik et al., 2012) in order to
226 constrain geodynamic relationships between deep Earth processes and tectonics (Mitchell et al., 2012).
227 Type 2 TPW (Inertial-Interchange True Polar Wander, IITPW) was originally described by Kirschvink
228 (1997) and is a hypothesis where the mantle and crust are rapidly displaced over large distances relative
229 to the spin axis as the moment of minimum inertia (I_{min}) approaches the maximum moment of inertia
230 (I_{max}) resulting in an interchange between the two (i.e. I_{min} becomes I_{max} and vice versa). IITPW has been
231 linked to supercontinent breakup as continental lithosphere and subduction zones move away from the
232 upwelling, known as a superswell, developing beneath the supercontinent, which is assumed to remain
233 quasi-stable (Li et al., 2004). In the Li et al. (2004) model, the superswell is likely to maintain the I_{max} , but
234 as all the continents move away I_{min} approaches I_{max} , thus they are speculated to interchange with each
235 other.

236

237 The primary challenge of subtracting the effects of TPW from APW paths is identifying and separating
238 TPW components from continental motion. Methods of detecting TPW vary, either isolating it directly
239 from palaeomagnetic data (Mitchell et al., 2012) or deconstructing the APWP of continents by comparing
240 the kinematic motions of all continental lithosphere on the globe to isolate TPW (Steinberger and Torsvik,
241 2008). Both approaches require *a priori* assumptions stemming from the choice of plate model being
242 analysed, which makes it impossible to apply a TPW correction from one model to another. For example,
243 fitting a great circle to palaeomagnetic data following the approach of Mitchell et al. (2012) is dependent
244 on knowing the relative continental configuration from which the palaeomagnetic data are sampled.
245 Alternatively, deconstructing the motions of the continents after Steinberger and Torsvik (2008) is
246 dependent on constraining absolute palaeolongitude to separate TPW from apparent polar wander. In
247 order to support their arguments, both methods are dependent on having some form of absolute
248 palaeolongitudinal control.

249

250 **2.4 Palaeolongitude**

251

252 There is no well-established method to compute the absolute palaeolongitude of any given plate that is
253 applicable to pre-Jurassic times. However, two hypotheses have been proposed for establishing absolute
254 palaeolongitude for pre-Pangea reconstructions (Torsvik and Cocks, 2017): (i) the plume generation zone
255 method (PGZ) (Torsvik et al., 2014) and (ii) the orthoversion model of the supercontinent cycle (Mitchell
256 et al., 2012).

257

258 The PGZ method is based on reconstructing the surface locations of kimberlites and large igneous
259 provinces (LIPs) to the margins of the large low shear velocity provinces (LLSVPs) (Burke and Torsvik,
260 2004; Torsvik et al., 2010a) situated on the core-mantle boundary (CMB) (Garnero et al., 2007; Li and
261 McNamara, 2013). The foundation for this method is the observation that kimberlites and LIPs, when
262 restored to a mantle reference frame at the time of eruption, are positioned preferentially above the
263 margins of the LLSVPs (Burke and Torsvik, 2004; Torsvik et al., 2010a). LLSVPs are regions of
264 anomalously slow (slower than ambient mantle) seismic velocities, close to the core–mantle boundary.
265 Since large-scale mantle structure is intimately related to the supercontinent cycle, LLSVPs (Li and
266 Zhong, 2009) are thought to exist for (at least) a similar time frame as the supercontinent itself (i.e.
267 present-day LLSVPs are thought to have existed at least back to 320 Ma) (Li and Zhong, 2009), but not
268 necessarily maintaining their present-day geometry (Flament et al., 2017; Zhang et al., 2010; Zhong and
269 Rudolph, 2015).

270

271 The PGZ method to establish absolute palaeolongitude therefore assumes: (i) the geometric stability of
272 the present-day positions of LLSVPs back to the time of interest, and (ii) that there is a positive statistical
273 correlation with margins of LLSVPs and the extrusion of LIPs and kimberlites at the Earth's surface.
274 Adopting the PGZ method allows one to position continents longitudinally by reconstructing the positions
275 of LIPs and kimberlites to overlie the margins of LLSVPs while simultaneously utilising TPW-corrected
276 palaeomagnetic data (Torsvik et al., 2014). In this way, models with explicitly defined and reproducible
277 absolute plate motions can be created (Domeier and Torsvik, 2014). However, a number of recent studies
278 have raised questions about the assumptions implicit in the PGZ approach. For example Flament et al.
279 (2017) and correspondence by Torsvik and Domeier (2017), Doucet et al. (2020), Zhong and Rudolph
280 (2015) and Zhong and Liu (2016) on long term LLSVP stability and Austermann et al. (2014), Davies et
281 al. (2015), with response by Doubrovine et al. (2016) on the statistical correlation. Consequently, while
282 utilising the PGZ method for plate reconstructions in the Palaeozoic is currently an open area of research,
283 the theoretical and practical application for Precambrian times remains untested.

284
285 The other proposed method of determining absolute palaeolongitude for deep time reconstructions is
286 known as the 'orthoversion model' and suggests that successive supercontinents coalesce orthogonally
287 (90° longitude) above the downwelling formed by subduction at the margin of the previous
288 supercontinent (Mitchell et al., 2012). Mitchell et al. (2012) test their model by first determining the
289 minimum moment of inertia (I_{min}) during each phase of supercontinent assembly. This is done by rotating
290 the available paleomagnetic data into a relative reference frame of a fixed continent (Africa for the
291 Phanerozoic, Laurentia for the Neoproterozoic) and then fitting a great circle to the resulting poles. I_{min} is
292 defined as the orthogonal axes (or pole) of the great circle and is taken to approximate the TPW axis.
293 Mitchell et al. (2012) fitted great circles to palaeomagnetic data for a selection of time periods (1165–
294 1015 Ma, 805–790 Ma, 550–490 Ma and 220–90 Ma) and calculated that the angle between each
295 successive I_{min} was $\sim 90^\circ$; as expected in the orthoversion model. The initial (palaeolongitudinal)
296 placement of Pangea is constrained from its position at ~ 90 Ma (centroid of I_{min} at 0°N , 10°E) when the
297 most recent TPW episode finished (Steinberger and Torsvik, 2008). Therefore, during Rodinia and
298 Gondwana, when I_{min} was 90° from present day, Mitchell et al. (2012) proposed that the centre of mass of
299 both was positioned at 100°E .

300
301 As with the PGZ, a number of limitations are apparent with the orthoversion model. In particular, the
302 method of Mitchell et al. (2012) does not separate continental motion from true polar wander when
303 calculating I_{min} , thus it assumes that during these times TPW is the primary signal recovered from
304 palaeomagnetic data and not continental motion. Secondly, the orthoversion model is inherently

305 dependent on both the quality and abundance of palaeomagnetic data (e.g. the Rodinia I_{min} in Mitchell et
306 al. (2012) is based on only three poles), as well as the continental configuration of the time (as the
307 continental configurations can determine the relative position of palaeomagnetic data when rotated into a
308 specific reference frame). Torsvik and Cocks (2017) highlight this succinctly by using slightly different
309 palaeomagnetic data, and a slightly different Gondwana configuration to produce an I_{min} between 550 and
310 490 Ma of 50°S, 64°E (compared to the estimate of Mitchell et al. (2012) of 30°S, 75°E). Finally, the
311 calculation of I_{min} at each time step occurs within a fixed relative reference frame (i.e. Fig. 2c), meaning
312 that the I_{min} itself cannot be restored to an absolute palaeogeographic position. Thus, the longitudinal
313 centre of Rodinia at 100°E is not explicitly proven by Mitchell et al. (2012). Instead, since the calculated
314 successive I_{min} 's are 90° apart, it is inferred to be in this location.

315

316 **2.5 Geology**

317

318 Geology is unfortunately silent on the absolute positioning of continents in time, except in circumstances
319 such as the PGZ method discussed above. However, it contains a wealth of information of relative plate
320 motions in the Mesozoic and Cenozoic (e.g. seafloor spreading), and in deeper time through the temporal
321 evolution of sedimentary basins and facies, and tectonic affinities inferred from geochemistry, zircon
322 arrays, magmatism and metamorphism. However, geology does have an advantage over palaeomagnetic
323 data in that there are many different types of data available, especially from small and minor terranes in
324 the Precambrian that are otherwise unconstrained palaeomagnetically. Given the wealth of geological
325 data, especially in the Neoproterozoic, there is a general hierarchy of use that is a reflection of the scale of
326 the problem. Our approach is to start by building a global framework and work progressively to finer
327 resolution to inform the localised nature of that framework.

328

329 The first and most important geological data to gather is evidence of rifts and arcs, as they describe
330 separation (usually leading to seafloor spreading) and convergence, respectively and can therefore put in
331 place the framework for plate motions and subsequent interpretation of geological data. Secondly,
332 identifying piercing points where geological boundaries can be matched on now separate continents (e.g.
333 Appalachians and Gondwana forming orogenies), or ways of fitting two continents together in a stable
334 configuration. These piercing points are important because reconstructing continental configurations with
335 high confidence allows the generation of more rigorous APWPs and they help to constrain the location
336 and orientation of both rifts and arcs on the periphery of continents. Unfortunately, due to deformation
337 and progressive alteration of continental crust (e.g. changes of continent-ocean boundaries (COBs)), in
338 pre-Gondwana times it is difficult to be more precise than matching the margins of large continents and

339 for smaller cratons and terranes it is almost impossible. A pertinent example of this are the four different
340 proposed configurations of Australia and Laurentia during Rodinia (see reviews by Li et al., 2008;
341 Merdith et al., 2017b), which all broadly match the same margins against each other (east coast of
342 Australia with the west coast Laurentia, with or without an intervening continent) but place them in
343 different relative positions.

344

345 After arcs and rifts, we can loosely (but not exhaustively) group geological features and data based on
346 their applications. Sedimentary basins, dyke swarms, detrital minerals, geochemical signatures and fossils
347 are typically used to determine provenance or latitudinal band and align once contiguous regions. The
348 time-scales and conditions of metamorphism together with structural data can be used to infer the tectonic
349 setting and polarity of collisional events or help constrain the nature of indeterminate plate margins such
350 as transform boundaries. These types of data assist in increasing the resolution of a plate model by
351 understanding the geology at smaller scales within the framework of a specific tectonic setting, such as an
352 arc or rift. We stress here that the relationship between detailed regional geology and the broader
353 framework of a plate model is not a ‘one-way street’ but is highly iterative. If, for example, a detailed
354 geological study determines that an interval of magmatism and sedimentation that was originally
355 interpreted as a failed rift, in fact led to seafloor spreading, then the broader scale tectonic framework and
356 plate model must be re-evaluated. Finally, due to the qualitative nature of most geological data, the
357 iteration and implementation of these data into the plate model typically necessitates qualitative decisions
358 that others may disagree with. Iteration over the model continues until we approach tectonic congruency
359 within the model (i.e. data-based iterations in one part of the world do not nullify data in other areas of the
360 world). We stress that this does not mean our model is ‘correct’ or ‘true’, just that it is internally
361 consistent with as much data as possible. Consequently, we consider the model presented here a viable,
362 but non-unique interpretation of Neoproterozoic data.

363

364 A specific example of the importance of interpreting geology within a full-plate framework is given by
365 recent work on the Stenian–Cryogenian evolution of the East African Orogen. Although arc-related
366 magmatism has been recognised in the northern East African Orogen for a number of decades (e.g. Stern,
367 1994), the recognition of similar-aged arc magmatism in the higher-grade southern East African Orogen
368 of Madagascar, Southern India and East Antarctica has been more controversial and under-appreciated
369 until recently (Archibald et al., 2018, 2017; Armistead et al., 2019; Plavsa et al., 2015; Ruppel et al.,
370 2018). In one particular example, work over the last decade on Western Dronning Maud Land in East
371 Antarctica has identified an extensive Stenian–Tonian juvenile arc system (named TOAST; (Elburg et al.,
372 2015; Jacobs et al., 2015; Ruppel et al., 2018)). This discovery has gone hand-in-hand with the

373 recognition of a similar region in western Madagascar, known as the Dabolava Arc (Archibald et al.,
374 2017; Tucker et al., 2011) (e.g. Fig. 8). These arcs are now separated by a considerable distance, but their
375 reconstructed position in the Neoproterozoic and their similarity has led to us interpreting them as part of
376 one continuous subduction system that was active for the Stenian and Tonian. In this manner, we now
377 include TOAST as another part of Azania and have reworked a number of the plate boundaries in Merdith
378 et al. (2017a) to reflect these new geological data and tectono-geographic interpretations.

379

380 **2.6 Kinematic considerations**

381

382 The final line of reasoning used to create full plate models are plate kinematic constraints. These are not
383 defined explicitly through geological or geophysical data of the types outlined above, but rather come
384 from the idea that the evolution of plate motions through time must follow the broad principles of plate
385 tectonics in a way that would seem reasonable; for example, by equivalency with more recent and well-
386 constrained plate motions. The most basic requirement is that continental blocks cannot pass through or
387 significantly overlap other continents and we must be able to describe the position and motion of each
388 continental block for as long as the crust within that block is thought to have existed. While this may
389 seem obvious, these considerations present a powerful method for discriminating between competing
390 reconstruction scenarios. Models constructed for deep time that cannot evolve towards more recent and
391 present-day configurations of continents cannot be considered correct. Similarly, models requiring an
392 implausible kinematic evolution in order to meet present-day configurations cannot be correct. A tangible
393 example is Rodinia, for which a range of configurations could be permissible based on available
394 paleomagnetic and geological data (see reviews in Evans, 2013; Li et al., 2008; Merdith et al., 2017b).
395 However, analysing the sequence of plate motions required to translate each continent to their (better
396 constrained) positions during the Palaeozoic is more plausible in some of these scenarios than others, such
397 as not requiring individual terranes or blocks to cross multiple ocean basins or navigate their way around
398 a stable continent (Merdith et al., 2017b). Further examples where kinematic constraints add useful
399 insights are when constructing models that explicitly trace the evolution of plate boundaries and tectonic
400 plates. An example is expressed in the logic of Domeier (2018) who inferred that the longitudinal position
401 of Tarim, North China and South China during the late Cambrian–Devonian must have remained stable
402 relative to one another, because palaeolatitudes from palaeomagnetic data overlapped and therefore must
403 be consistent with their end position in the more well constrained Devonian–Triassic.

404

405 Finally, we take a uniformitarian view of tectonic evolution, in that we assume that plate tectonics and
406 relative plate motions were operating on similar principles in the Neoproterozoic to what we can observe

407 in the Mesozoic and Cenozoic. During more recent times, the motion of plates remains relatively constant
408 for time lengths on the order of 10–100 Ma, with changes in motion occurring comparatively quickly (< 3
409 Ma) and tied to an event further afield, such as terrane collision in a subduction zone, subduction onset,
410 rifting onset or ridge subduction (e.g. Austermann et al., 2011; Cawood et al., 2016; Cawood and Buchan,
411 2007; Knesel et al., 2008). The representation of this within a conceptual framework is that a single
412 continent or plate may move for time lengths on the order of 10–100 Ma around a single Euler pole,
413 before a plate re-organisation event triggers a change in direction and velocity of the plate (Gordon et al.,
414 1984). Within palaeomagnetic data, this approach is exemplified in Torsvik et al. (2008), where filtered
415 compilations of data result in smooth APWP segments punctuated by cusps in motion and velocity, and
416 more recently by Wu et al. (2020). For the Neoproterozoic, which has much more sparse palaeomagnetic
417 data coverage than the Phanerozoic, the logic can be applied by linking changes in plate direction (as
418 suggested or necessitated by palaeomagnetic data) directly to geological evidence of a change in tectonic
419 regime within the region of interest (e.g. Merdith et al., 2017b).

420

421 **2.7 Synthetic Ocean Plates**

422

423 A complete full-plate model by definition includes a representation of the evolution of ocean basins
424 through time. This is the most uncertain part of any full-plate model. With a few exceptions (e.g. Granot,
425 2016), no pre-Jurassic ocean crust is preserved *in situ*, so that even where we can infer the presence of
426 divergent plate boundaries (for example, following continental breakup), the precise geometry and
427 spreading rates at these boundaries are conjectural. Consequently, there are no unique solutions to the
428 definition of these boundaries and the synthetic ocean plates constructed from them. The main aim in the
429 reconstruction of ancient ocean basins is to ensure that the synthetic plates and plate boundaries are at
430 least consistent with sparse observations preserved on the continents.

431

432 Firstly, although we have little data on the creation of oceanic lithosphere for pre-Jurassic times, the
433 geological record does preserve data on the consumption of oceanic lithosphere at subduction zones. The
434 most important criteria that therefore must be met are that (i) (sub-) orthogonal divergence occurs at mid-
435 ocean ridges and that (ii) convergence occurs at subduction zones (Domeier and Torsvik, 2014). The
436 former of these two criteria typically occurs between continents during supercontinent breakup or when
437 small terranes rift from a continent (Dalziel, 1997). The latter of these includes subduction along the
438 margins of continents as well as within intra-oceanic arcs. Thus, these two criteria necessitate the
439 extrapolation of known plate boundary positions (e.g. preserved continental arcs, rift zones) into larger
440 areas to ensure the tectonic congruency of the model (in this case, that convergence or divergence at one

441 location doesn't nullify the same criteria in another location). It is this step, in particular, that requires
442 significant iteration when constructing a model. Maintaining tectonic congruency for a model is best
443 achieved when the extrapolation of plate boundaries is done as simply as possible. For example, in
444 reconstructing an ocean basin without any continental crust (e.g. Pacific Ocean, Panthalassa Ocean), a
445 triple junction is usually the simplest expression of a ridge system that ensures divergence in all directions
446 and convergence at its margins (Domeier and Torsvik, 2014). The evolution of such a triple junction
447 could also be seen in the Ediacaran opening of the Iapetus Ocean (e.g. Pisarevsky et al., 2008; Robert et
448 al., 2020),

449

450 **3 Model selection and justification**

451

452 **3.1 Existing plate models**

453

454 The selection of which full-plate model to assist in solving a problem is dependent on the nature of the
455 problem, as each published plate model is constructed using a different approach and has a different
456 reconstruction framework. For example, a study looking at absolute plate motions of the Cenozoic has
457 little use for models connected to the Palaeozoic or Neoproterozoic. Instead, such studies typically
458 include a comparison with previously published models as well as a rigorous mantle reference frame.
459 Similarly, a study that traces the latitudinal distribution of continental arcs through the Phanerozoic has no
460 need for a mantle reference frame and a study investigating the changes in net rotation through time
461 would be concerned with small, localised improvements from regional models but more focussed on
462 capturing the large scale changes that occur when continents breakup. Although newer plate models
463 typically address the shortcomings of previous models or implement more refined updates of regional
464 areas, this does not necessarily make them better for all applications. Older plate models have been more
465 rigorously tested and used by the community and as one travels further back in geological time, the data
466 are more ambiguous and can invite alternative interpretations. For the purpose of this study, the existing
467 plate models we consider are only those that are publicly released with fully self-consistent with coherent
468 plate motions and plate boundaries. Thus, we omit many models that provide only continental motions
469 (Scotese, 2016; Torsvik and Cocks, 2016), alternative or regional Rodinia configurations (Evans, 2013) or
470 regional refinements of global models for the Mesozoic and Cenozoic (e.g. Vaes et al., 2019).

471

472 The major step forward for producing full-plate models was the construction of open-source computer
473 software specifically designed to work with full-plate reconstructions (e.g. GPlates; (Gurnis et al., 2012).
474 Following their development, GPlates-compatible global models for the Early Jurassic to present (Seton

475 et al., 2012; Shephard et al., 2013) and a model for the Late Palaeozoic (Domeier and Torsvik, 2014) soon
476 followed (hereafter, SET12 and DT14, respectively). Subsequent work by Matthews et al. (2016)
477 (MAT16) bridged the gap between the Palaeozoic and Jurassic, linking a slightly modified version of
478 DT14 with an updated SET12 model, the Müller et al. (2016) model (MUL16). Further back in time, two
479 models for the Early Palaeozoic (500 to 410 Ma) now exist: Domeier (2016) (DOM16), which
480 encompasses the evolution of the Iapetus and Rheic Oceans, as well as the motion of Gondwana, and
481 Domeier (2018) (DOM18), which models the evolution of the first generations of Tethyan Oceans and
482 Central Asian blocks (Siberia, North and South China, Tarim). For the Neoproterozoic, Merdith et al.
483 (2017a) (MER17) produced a full-plate model from 1000 to 520 Ma, using the models of continental
484 motion presented by Li et al. (2013, 2008) as a base. An alternative reconstruction from the late
485 Palaeozoic to present-day has been presented by Young et al. (2019) (YOU19). YOU19 offers an
486 alternative full-plate model for the Palaeozoic to the DT14 and MAT16 models that does not rely on the
487 PGZ method. Finally, a deforming plate model was produced (Müller et al., 2019) (MUL19) that
488 modelled rift and convergence deformation from 250 to 0 Ma. Table 1 summaries the main features of
489 these models.

490

491 **3.2 Cenozoic and Mesozoic plate models**

492

493 We consider three plate models for the Cenozoic and Mesozoic: SET12, MUL16, and MUL19 (Fig. 4).
494 The SET12 model spans from present-day to 200 Ma, with MUL16 and MUL19 extending back to 230
495 and 240 Ma, respectively. The main geological constraint of these models are the magnetic lineations
496 preserved in oceanic crust that describe the relative movement between the pairs of continents breaking
497 apart during the fragmentation of Pangea. The evolution of the Atlantic, Indian, Southern, Arctic and
498 Cenozoic Pacific oceans are consistent, to the first order, across all three models. Larger differences
499 between the models arise in regions where oceanic crust has been subducted, upon which they then rely
500 on a combination of geometric (e.g. assumption of symmetrical spreading), geological, seismic and
501 palaeomagnetic data to constrain the motion and evolution of terranes that open and close ocean basins
502 (e.g. Liu et al., 2010, 2008; Sigloch and Mihalynuk, 2013). In these regions, the tectonic histories, even
503 for Cenozoic times, remain the subject of ongoing research and the scenarios embedded in the global
504 models used in this study represent one candidate amongst many competing models. For example, the
505 extent of subduction systems within the Tethyan domain and the nature of India-Eurasia collision is still
506 hotly debated (Hu et al., 2016; Parsons et al., 2020; van Hinsbergen et al., 2020, 2012). Similar
507 combinations of methods have been used to propose alternative interpretations for circum-Pacific regions,
508 especially for the Cretaceous and earlier times in the northwest Pacific (Domeier et al., 2017;

509 Konstantinovskaya, 2002; Shapiro and Solov'ev, 2009; Vaes et al., 2019; Wu et al., 2016), the southwest
510 Pacific (Hochmuth et al., 2015; Matthews et al., 2015; Schellart et al., 2006; Sutherland et al., 2020) and
511 the northeast Pacific (Clennett et al., 2020; Sigloch and Mihalynuk, 2013). All of these regions invite
512 competing models and alterations to existing global models and it is notable that many of the studies
513 mentioned above have benefited from using the resources made available by previous global studies,
514 beginning with SET12, as a starting point for detailed regional analysis.

515

516 All three models employ a relative hierarchy (Fig. 2a), in which a fully relative plate motion model is tied
517 to an absolute plate motion model through Africa. The relative hierarchies are similar across the models
518 because of the preserved oceanic crust. These global relative motion hierarchies are then linked to an
519 absolute reference frame tied to the mantle, though the details of these reference frames differ between
520 models. SET12 uses a hybrid reference frame, using a moving hotspot reference frame for 100–0 Ma
521 (O'Neill et al., 2005) and a true-polar wander corrected palaeomagnetic reference frame for 200–100 Ma
522 (Steinberger and Torsvik, 2008). MUL16 also adopts a hybrid absolute reference frame, but uses the
523 moving hotspot model of Torsvik et al. (2008) instead of that of O'Neill et al. (2005) based on an
524 assessment of the geodynamic plausibility of a range of alternative mantle reference frames by Williams
525 et al. (2015). MUL19 departs from both SET12 and MUL16 in that it uses an absolute reference frame
526 derived by Tetley et al. (2019). The method of Tetley et al. (2019) optimises absolute plate motions
527 through a joint inversion involving trench motion, fitting hotspot motion tracks and net rotation to
528 determine the motion of Africa (at the top of the relative hierarchy) that simultaneously best fits all three
529 criteria. Despite the emphasis on mantle reference frames in these previous global models, the same
530 relative plate motion hierarchies can also be tied to a pure paleomagnetic reference frame (e.g. Cao et al.,
531 2019; Torsvik et al., 2008).

532

533 In this study, we rely on palaeomagnetic data as the main basis for linking absolute plate configurations
534 continuously from the Cenozoic to the early Neoproterozoic, when tying plate configurations to the
535 mantle is far more problematic. Nonetheless, some aspects of the more recent plate motions still rely on
536 observations from hotspot trails—specifically, the motion of the Pacific Plate and other oceanic plates
537 that have bordered it. During the Early Cretaceous, these plates lay within the Panthalassa ocean basin
538 that was entirely surrounded by subduction zones, meaning that we cannot tie the motions of the oldest
539 crust of the Pacific Plate to the continents by seafloor spreading anomalies.

540

541 Finally, with regard to Mesozoic-Cenozoic global plate models, we note that MUL19, while containing
542 the same relative framework as SET12 and MUL16, also contains deforming plates. In MUL19,

543 deformation of rifts and collisional zones are modelled explicitly, making it the first plate model to step
544 away from the simplification of rigid plates that all other models assume. However, the reconstruction
545 here relies on the simpler, rigid approximation used in SET12, MUL16, and other previous studies.

546

547 **3.3 Mid-late Palaeozoic plate models**

548

549 The progression of the three Palaeozoic plate models (all modelling 410–250 Ma) is slightly more
550 nuanced than in the Mesozoic and Cenozoic because of the absence of preserved oceanic crust. DT14 is
551 the original model and both MAT16 and YOU19 used DT14 as the basis of their Palaeozoic models and
552 then connect to MUL16 for the Mesozoic and Cenozoic. In effect, MAT16 is the connection of DT14 (i.e.
553 minimal changes) to MUL16, while YOU19 is an alternate version of DT14 for the Palaeozoic.

554

555 DT14 is heavily based on work by Torsvik et al. (2012) and models the amalgamation of Pangea through
556 the collision of Laurussia and Gondwana and the evolution of the Palaeo-Tethys and opening of the
557 Meso-Tethys oceans. The model has a flat hierarchy (Fig. 3a) with APWPs defined for each individual
558 continent and each continent being tied directly to the spin axis. Domeier and Torsvik (2014) also use the
559 PGZ method to constrain absolute palaeolongitude. Therefore, their model assumes the stable, immutable
560 nature of the present-day LLSVPs within the mantle back to 410 Ma. DT14 is presented in both a mantle
561 and a palaeomagnetic reference frame, with the mantle reference frame being corrected for TPW after
562 Torsvik et al. (2014) .

563

564 MAT16 adopted the DT14 model, with minor amendments required to link it with MUL16 (see Matthews
565 et al. (2016) for details). A key difference between the two models is that MAT16 translated the flat
566 hierarchy of DT14 into a fully relative reference frame (i.e. converted the structure from Fig. 3a into Fig.
567 2a), where the motion of all plates is tied to Africa, which is then tied to an absolute plate motion model.
568 The absolute plate motion model of MAT16 is the same as in both MUL16 (i.e. a hybrid between
569 hotspots, slabs and palaeomagnetic data) and DT14 (i.e. absolute latitude and longitude, corrected for
570 TPW). Thus, MAT16 also invokes the PGZ method.

571

572 YOU19, while starting from MAT16 as a base, differs much more from MAT16 than MAT16 does from
573 DT14. This is because YOU19 uses a different base assumption, leading to notable changes in the actual
574 plate model itself. The most important difference is that YOU19 does not assume that LLSVPs were fixed
575 and stable back to 410 Ma. They abandon the PGZ method for constraining palaeolongitude and thus
576 argue that they can better accommodate geological and kinematic (plate speed and trench migration)

577 criteria more strongly than either DT14 or MAT16. The two key changes that YOU19 implemented
578 (relative to DT14 and MAT16) are shifting Laurussia in latitude and longitude to be closer to its position
579 in Pangea against Gondwana, thus removing a dextral motion between the Patagonian margin of South
580 America and Laurussia (e.g. Fig. 2c), and removing easterly drift of South China during the
581 Carboniferous–Permian. The implementation of both in DT14 (and then in MAT16) is a consequence of
582 the PGZ method, since the longitudinal position of Laurussia and South China is based on fitting
583 preserved eruptions to the edges of LLSVPs (Domeier and Torsvik, 2014). YOU19 argue that the
584 interpretations of DT14 introduce unrealistic kinematic scenarios: 8000 km of relative dextral motion
585 between Laurussia and the Patagonian margin of Gondwana (e.g. Fig 2c) at a relative plate velocity of 30
586 cm/a and South China moving at plate speeds of 40 cm/a between 260 and 250 Ma in MAT16 (Young et
587 al., 2019). The dextral motion between Gondwana and Laurussia of DT14 and MAT16 that that YOU19
588 removed is not a transition from the Pangea B to Pangea A configuration, which is explicitly defined as
589 dextral motion after Pangea formed (Domeier et al., 2012). Rather, all three models adopt a Pangea A
590 configuration and include *some* component of dextral motion between Laurussia and Columbian–
591 Mexican margin of Gondwana in the Devonian. However, the position of Laurussia in YOU19 is
592 unsupported palaeomagnetically by $\sim 30^\circ$ (Section 4.2 and 5.1), which is problematic as there is an
593 abundance of data from both Laurentia and Baltica to constrain its position at this time.

594

595 Recently, Wu et al. (2020) have also proposed an integrated geological and palaeomagnetic model for the
596 amalgamation of Pangea. Their study used a different selection of palaeomagnetic data (all three mid-late
597 Palaeozoic full-plate models discussed here used the compilation of Torsvik et al. (2012) as a base) and a
598 new method of APWP generation that weighted poles based on their quality and uncertainty. The model
599 of Wu et al. (2020) also suggested that the formation of Pangea was originally initiated by collision
600 between Laurussia and a promontory of Gondwana consisting of the Variscan Massifs at ca. 400 Ma. In
601 their model, the promontory formed by the scissor-like opening of the Palaeotethys Ocean off the
602 northern margin of Gondwana. In principle this model would be very compatible with the one we present
603 here, however as the model of Wu et al. (2020) is currently only palaeogeographic, we do not consider it
604 as an option for merging in this study.

605

606 YOU19, like MAT16, uses a relative plate hierarchy with Africa as the root of the hierarchy. Africa is
607 connected to an absolute plate motion model using the Torsvik and Van der Voo (2002) APWP for the
608 Palaeozoic, before the model transition to the absolute plate motion model of MUL16. Because the model
609 has no absolute palaeolongitude control, YOU19 does not constrain or correct for TPW in the Palaeozoic
610 and is therefore presented in a purely palaeomagnetic reference frame.

611

612 **3.4 Early Palaeozoic plate models**

613

614 Two separate models exist for part of the Early Palaeozoic between 500 and 410 Ma. Each of these two
615 models focus on a separate area of the Earth at the time. DOM16 focuses on the evolution of the Iapetus
616 and Rheic oceans and the amalgamation of Laurussia (Baltica, Laurentia and Avalonia). DOM18 models
617 the evolution of Siberia, Gondwana, the terranes that now make up the Central Asian Orogenic Belt and
618 the Chinese cratons (Tarim, North and South China). While each model focuses on a different area, the
619 overarching assumptions and framework of both models are identical. Both models follow the approach
620 of DT14, possessing a flat hierarchy with APWPs being defined for most continents, such that they all
621 move independently from each other. The model extends the assumptions of the PGZ method back to 500
622 Ma, using the location of LIPs and kimberlites to constrain absolute palaeolongitude. Both models
623 therefore have a TPW correction, and are presented in a mantle and palaeomagnetic reference frame.

624

625 **3.5 Neoproterozoic**

626

627 Only one full plate model exists for the Neoproterozoic (MER17, Merdith et al., 2017a). MER17 is based
628 on Li et al. (2013, 2008) and models the evolution of Rodinia; it's breakup and the amalgamation of
629 Gondwana (1000–520 Ma). There are, however, a number of important considerations that differentiate
630 MER17 from Phanerozoic full-plate models. Firstly, MER17 used a hybrid plate rotation hierarchy,
631 defining two separate nodes that move independently from each other using palaeomagnetic data tied
632 directly to the spin axis (India as one and Laurentia, as the centre of Rodinia, as the second) (Fig. 5a, b).
633 In this model, India and Laurentia both act as separate roots that then constrain a series of relative plate
634 rotations that collectively describe the rest of the world. Secondly, the continental drift model of Li et al.
635 (2013) invokes the orthoversion model of determining palaeolongitude, where they fix the I_{min} of Rodinia
636 to be at 100°E. As MER17 adopted these rotations as their base, there is an element of the orthoversion
637 model preserved between models. However, MER17 also drastically changed the configuration of
638 Rodinia, as well as the timing of breakup, compared to Li et al. (2013), relative rotations within the
639 Laurentian (i.e. Rodinian) node and absolute rotation of Laurentia itself in order to fit geological and
640 kinematic constraints. They did not recalculate TPW and the I_{min} of Rodinia, thus MER17 does not have a
641 strict absolute palaeolongitude control. Finally, there is no correction for TPW in MER17 and no mantle
642 reference frame; the model is presented purely in a palaeomagnetic reference frame.

643

644 **3.6 Model selection**

645

646 In order to produce a coherent global model, we must select from the models described above that best
647 align with our goals: (i) open the Neoproterozoic and Cambrian up for quantitative tectonic analysis; (ii)
648 create a framework that can support local or regional studies for the Neoproterozoic and (iii) a foundation
649 for future studies looking at long timescale (> 10–100 Ma) trends in either tectonics or geodynamics.

650

651 As our aim is to produce a model back to 1 Ga, three choices are already made for us because they are the
652 only models that exist for those time periods: MER17 for 1000–520 Ma, and DOM16 and DOM18 for
653 500–410 Ma. For the remaining time period, MAT16 (an extension of DT14) and YOU19 are viable
654 options. Both models link to the MUL16 model and choosing between them requires considering the
655 reconstruction framework (e.g. hierarchy, reference frame) of our model with respect to our intent. To
656 satisfy our goals, we need a model in a palaeomagnetic reference frame. We acknowledge the value and
657 potential in exploring hypotheses for constraining palaeolongitude. However, since tectonic
658 reconstruction models, such as the one we are presenting here, are a required starting point for exploring
659 the long-term evolution of mantle, we have kept the model conservative by not assuming the fixity of
660 LLSVPs or that TPW dictates supercontinent position. Therefore, we elect to leave palaeolongitude
661 unconstrained by either PGZ or orthoversion methods. Thus, we selected the YOU19 model, as it has
662 removed the absolute palaeolongitude controls adopted in MAT16. We stress that our approach is
663 deliberately conservative and requires the fewest *a priori* assumptions, but anticipate that future
664 developments will see improvements in the model presented and that a comparison of several end
665 member models would be useful for evaluating the long-term connection between lithosphere and mantle.
666 One promising way forward could be to optimise tectonic parameters, such as subduction zone migration
667 and plate velocity in the manner of Tetley et al. (2019) to define a mantle reference frame. The model we
668 present is an essential precursor for such techniques. Our model is constructed with a palaeomagnetic
669 reference frame, that is, even at younger times, there is no mantle reference frame. For studies needing
670 such a reference frame, which only exists since the Cretaceous, we suggest people use either SET12,
671 MUL16 or MUL19.

672

673 **3.6.1 Our approach to plate modelling**

674

675 Here we outline our approach to merging individual plate models into a coherent global plate model
676 spanning the past 1 Ga. As detailed above, our goal is a geologically constrained model within a
677 palaeomagnetic framework. In addition to reconstructing plate motions, we also model plate boundaries,

678 which requires us to also focus on the relative motion between plates. The evolution of plate boundaries is
679 commonly preserved in the geological record such as passive margins marking divergent boundaries and
680 magmatic arcs recording convergence. Palaeomagnetic data, although providing a quantitative absolute
681 constraint on the position of a craton, have varying uncertainty that allow some manipulation and
682 flexibility. For example, continental drift models (and APWPs) typically fit the data as tightly as possible.
683 However recent studies have analysed the effect of exploring both the temporal (i.e. age constraints) and
684 statistical uncertainty to create alternative APWPs (Tetley et al., 2019; Wu et al., 2020). Given this, we
685 use a hybrid approach that adopts parts of both the flat, palaeomagnetic hierarchy traditionally used for
686 Precambrian reconstructions and the relative framework used in more recent times.

687
688 Using this approach, our model has multiple clusters of related cratonic elements moving together (Fig.
689 5a). Whether a specific cratonic element moves relative to another is dependent on their geological
690 relationship. For example, if they are separated by an incipient ocean basin as indicated by geological data
691 then we suggest that it is easiest for them to be moving relative to each other (within the bounds of
692 whatever available palaeomagnetic data), because the relative relationship of divergence (expressed as a
693 mid-ocean ridge) is preserved. The hierarchy is defined by geological precedence, where terranes move
694 relative to blocks, blocks move relative to cratons and cratons move relative to 'supercontinents'. In this
695 manner, generally (but not necessarily exclusively), crust with more preserved data should always be
696 placed above crust with less preserved data in the hierarchy, because we have more confidence in the
697 geological evolution (and also likely palaeomagnetic constraints) from these continents. Alternatively, if
698 continents are separated by a large ocean basin, we form a new cluster. This cluster-approach has an
699 added benefit in the more uncertain Neoproterozoic, as it means that we can simply introduce a new
700 cluster for Rodinia, which has Laurentia as the root of the relative hierarchy, instead of maintaining
701 Africa or the Congo craton at the top of the cluster (which are the roots of Pangea and Gondwana,
702 respectively) and can also introduce a cluster for India and South China which, in our interpretation, move
703 separately to Rodinia on the other side of the globe.

704
705 A full-plate reconstruction models both continents (palaeomagnetic data) and plate boundaries (geological
706 data). Therefore, we use both data simultaneously to iterate towards a solution. Palaeomagnetic data are
707 used initially to build a continental drift framework (e.g. (Li et al., 2013, 2008)). If palaeomagnetic data
708 are abundant and of high quality, then either an APWP or GAPWaP are constructed, which we do for the
709 Phanerozoic. We then introduce geological data in the form of plate boundaries (e.g. compilation of rifts
710 and arcs (Merdith et al., 2017b)) and use the compilation to manipulate the model in a manner that still
711 fits the palaeomagnetic data, but also accommodates geological data. Structural and metamorphic

712 constraints are used here principally to infer (where possible): (i) polarity of subduction, (ii) collision
713 timing and (iii) orientation of rifting. Once the broader tectonic framework is implemented, we begin
714 introducing smaller blocks and terranes into this framework. This approach allows us to increase the
715 resolution of the model within key areas while maintaining the tectonic coherency of the model as a
716 whole. For example, where the broader tectonic framework models subduction leading to collision, the
717 introduction of terranes and smaller blocks that preserve evidence suggesting a two-stage collision or
718 accretion of an oceanic arc can be used to more finely model the plate boundary network. Other pieces of
719 geological data, such as faunal provinces (e.g. Burrett et al., 1990), isotopic signatures (e.g. Collins et al.,
720 2011) and detrital zircons (e.g. Cawood et al., 1999) are used here to assist with connecting disparate
721 terranes to larger blocks that they share affinity with or to infer the presence of a plate boundary not
722 directly preserved in the geological record (such as by a diverging fossil record).

723

724 Because our model contains geological data in the form of plate boundaries, we are particularly interested
725 in ensuring the forwards and backwards compatibility of any decision made around geological data,
726 especially for terranes or blocks that have limited palaeomagnetic data. For example, if the data support
727 two or three interpretations in the early Neoproterozoic, but only one of those is consistent with an
728 Ediacaran (or younger position), then we consider that position more reliable than the other two.
729 However, we argue that this logic also works in reverse; if a number of positions are deemed viable in the
730 Ediacaran for a terrane based on the data available, but only one of those also fits what data are available
731 in the early Neoproterozoic, then we will use the older data to force an interpretation of the younger data
732 (e.g. Evans, 2009). This argument is highlighted by the concept of ‘world uncertainty’ (the percentage of
733 total crust (oceanic and continental) on the earth at any one time that is also preserved at present-day) of
734 Torsvik et al. (2010b). Because time is asymmetrical, this means that the level of confidence we have at
735 present-day is 100%, but decreases linearly back in time to ~70% at 200 Ma (i.e. 70% of all crust at 200
736 Ma is no longer preserved). At 400 Ma, the world certainty is ~73% (Domeier and Torsvik, 2017), and
737 using estimates of continental crustal volumes, at 600 Ma it is between 75 and 80% (Cawood et al.,
738 2013a). Therefore, for the Mesozoic and Cenozoic, data from younger times are much more compelling to
739 force interpretations of older data because of how much more confident one can be in the last 20 Ma.
740 Comparably, the difference in this measure of uncertainty is much smaller between the Neoproterozoic
741 and Cambrian, thus we suggest that models in this time period should simultaneously use older and
742 younger data to iterate towards a solution.

743

744 **4 Palaeomagnetic Data**

745

746 A major problem in comparing the Phanerozoic models is that they all use different rotational
747 frameworks, including reference trees (i.e. flat vs hierarchical) and different absolute reference frames. In
748 order to properly synthesise the DOM16, DOM18 and YOU19 models into a single reconstruction, we
749 first need to lay a coherent groundwork in defining an absolute plate motion model for the largest
750 continents during this time in order to merge the individual models. To do this, we first derive a new
751 APWP for Gondwana (540–320 Ma) and GAPWaP for Pangea (320–0 Ma) using the palaeomagnetic
752 compilation of Torsvik et al. (2012) to provide an absolute palaeomagnetic reference frame for 540 to 0
753 Ma. We also apply the GAPWaP to the MUL16 portion (i.e. the Mesozoic and Cenozoic) of the YOU19
754 model. We do this to ensure compatibility between the Cenozoic and Palaeozoic, and also because our
755 goals for this model are broad scale (> 10–100 Ma) trends, mostly focussed on the Neoproterozoic. The
756 method we follow to calculate our APWP and GAPWaP is outlined below. For the Neoproterozoic, and
757 non-Gondwana constituents of the Palaeozoic, we use the compilations of palaeomagnetic data as
758 presented in MER17, DOM16 and DOM18, along with some other additions. Palaeomagnetically derived
759 alterations to the models are also discussed below.

760

761 **4.1 APWP and GAPWaP construction**

762

763 The absolute reference frames for Gondwana (540–320 Ma) and Pangea (320–0 Ma) used in this study
764 are derived using the method and velocity-optimised global palaeomagnetic data of Tetley (2018).
765 APWPs are routinely constructed using poles assigned averaged or nominal ages, which particularly for
766 older times, where palaeomagnetic constraint becomes increasingly limited, contribute to spurious
767 apparent polar wander behaviours. This method directly evaluate individual palaeomagnetic pole age and
768 associated uncertainty in combination with calculated pole A95 latitude and longitude uncertainties to
769 derive optimised APWPs that minimise predicted plate velocities and plate velocity gradients
770 (instantaneous accelerations). The resulting rate of apparent polar wander in optimised APWPs was
771 reduced globally by an average of 56% by comparison to existing APWPs, resulting in predicted
772 Phanerozoic plate motions displaying greater kinematic consistency with present-day plate motion
773 behaviours.

774

775 Optimised APWPs were produced for the 15 major continental blocks of Amazonia, Australia, Colorado,
776 East Antarctica, Greenland, India-Pakistan, Madagascar, Meseta, North America, Northeast Africa,
777 Northwest Africa, Panama, Patagonia, Somalia, and Stable Europe. Applying the method as described in
778 Torsvik et al. (2012, 2008) and the data provided in Torsvik et a. (2012), optimised continental pole data
779 from all 15 continents were rotated from their individual source coordinate frames into a consistent South

780 African coordinate frame using the rotation model from this study. Now in a consistent coordinate frame,
781 a GAPWaP for Pangea was produced using all poles aged 320-0 Ma (due to the collision of Laurussia and
782 Gondwana during the Late Carboniferous), with a second GAPWaP produced using poles aged 540-320
783 Ma associated with Gondwana (South America, Africa, India, Antarctica and Australia) only. For both
784 GAPWaP reference frames, the *GMAP* software was used (Torsvik et al., 2012, 2008; Torsvik and
785 Smethurst, 1989), applying a running mean method using a window size of 20 Ma and a step size of 10
786 Ma.

787

788 **4.2 Palaeomagnetic compilation**

789

790 A compilation of palaeomagnetic data was used to constrain the position of all continental blocks during
791 the Neoproterozoic and for non-Gondwanan continental blocks during the early and middle Palaeozoic.
792 The Neoproterozoic data are presented in Table 2 and a GPlates compatible file of the data is also
793 presented in the Supplementary Material. Figure 6a–c shows the great circle misfit of our model to the
794 selected poles. As the majority of the data have already been used in the MER17, DOM16 and DOM18
795 reconstructions, we point readers to those publications for in-depth discussion of the data. Here, we
796 discuss two alterations to the base models that we implemented based on palaeomagnetic data: Tarim at
797 ca. 700 Ma and Laurussia at ca. 420–400 Ma.

798

799 The MER17 model omitted Tarim prior to 700 Ma due to its pre-700 Ma palaeomagnetic data nullifying
800 the position that placed it outboard of Australia. We rectify this and include a robust, time-sensitive
801 geological argument for its position against India-South China during the Tonian (1000–720 Ma) and
802 Cryogenian (720–635 Ma) (Section 5.4.2). The rationale for placing Tarim in this position is to allow for
803 a significant 180° rotation required by paleomagnetic data. Two well-dated and high-quality poles; the
804 760–720 Ma Qiaoenbrak Formation (Wen et al., 2013) and the 770–717 Ma Baiyisi Volcanics (Huang et
805 al., 2005) are internally consistent and require Tarim to be inverted 180° from its present-day position
806 (i.e. northern margin facing south). A third pole, assumed to be pre 700 Ma, from the Aksu Dykes (Chen
807 et al., 2004) is rejected for poor age constraints and the possibility of remagnetisation (Wen et al., 2017).
808 Comparably, three younger poles: the 635–550 Ma Sugetbrak Formation (Zhan et al., 2007), the ca. 635
809 Ma Tereeken Cap Carbonate (Zhao et al., 2014) and the 621–609 Ma Zhamoketi Andesite (Zhao et al.,
810 2014) are all consistent with each other and indicate that Tarim was in its present-day orientation. A
811 recent pole by Wen et al. (2017) from the Lower Sugetbrak formation (640–615 Ma) stands in conflict
812 with these three poles, suggesting a ~50° rotational difference in the orientation of Tarim, while
813 maintaining the same palaeolatitude. Wen et al. (2017) dismiss the three earlier poles due to similarity to

814 Silurian–Devonian poles for the Sugetbrak Formation and possible remagnetisation for the latter two
815 poles, respectively.

816

817 Successfully fitting the two older poles with either the three younger poles or the single pole of Wen et al.
818 (2017) is not possible in a model in which Tarim is surrounded by continental lithosphere for the
819 Neoproterozoic or in a scenario where Tarim is attached to the north-western or northern margin of
820 Australia or the northern margin of India (e.g. instead of South China). An accommodation of the data can
821 be obtained by placing Tarim in a ‘Missing-Link’ position (between Australia and Laurentia) with
822 breakout and rotation of Tarim from ca. 700 Ma; as argued by Wen et al. (2018, 2017). However, beyond
823 the kinematic issues with Missing-Link style models (Merdith et al., 2017b), it is difficult to account for
824 the formation of the 760 Ma Aksu Blueschist if Tarim was located in the centre of an assembled Rodinia
825 (see comment by Song and Li (2019) and reply by Wen et al. (2019)). In the present model, we use the
826 pole of Wen et al. (2017) due to its greater reliability, though our model could easily be adapted to fit the
827 other three poles if necessary (Section 5.4.2.). We also note that although we argue that India-South China
828 were separate from the rest of Rodinia, our preferred Neoproterozoic position for Tarim is also
829 compatible with models in which India-South China formed part of Rodinia (e.g. Cawood et al., 2013b).

830

831 In the YOU19 model, Laurussia was moved $\sim 30^\circ$ further north in latitude relative to its position in DT14
832 and MAT16 (Section 3.3). The consequence of this decision by YOU19 is that Laurussia is not in a
833 palaeomagnetically constrained position in the late Silurian and early Devonian. For this model, in
834 addition to the compilation of palaeomagnetic data in Torsvik et al. (2012), we also used palaeomagnetic
835 data from mainland Baltica (Table 2) to constrain the late Silurian to early Devonian position of
836 Laurussia. We implement these alterations to produce a coherent motion of Laurentia (later Laurussia)
837 and Gondwana that fits palaeomagnetic data, while also ensuring that the relative motion between
838 Gondwana and Laurussia is convergent, rather than dextral transform (e.g. Fig. 2c).

839

840

841 **5 Alterations to models**

842

843 We describe the alterations made to individual models separately for clarity, however, we stress that no
844 single model was treated in isolation. Each alteration, whether during the Tonian or Devonian, was
845 evaluated both forwards and backwards in time in order to ensure continuity with both older and younger
846 geological and/or palaeomagnetic evidence. We begin our discussion with the alterations made to
847 YOU19, followed by alterations to DOM16, DOM18 and MER17. The alterations to DOM16 and

848 DOM18 are discussed together, as the two models essentially form a single global model between 500
849 and 410 Ma. The most significant changes, and the focus of much of the discussion, occurred in MER17.
850 This is because, firstly, connecting this model with younger models in order to validate Neoproterozoic
851 tectonic-geography is a primary goal of this study, and secondly, because many of the alterations are
852 completely new and have never been incorporated into a plate model before. Figure 7a–b displays a
853 latitudinal overview of the major and minor cratons in our model, with a comparison to the DOM-MAT
854 models for the Phanerozoic, and the model of Li et al. (2013, 2008) for the Neoproterozoic. Figures 8, 9
855 and 10 show overviews of our reconstruction at 1000 Ma, 500 Ma and present-day with relevant terranes
856 and blocks highlighted.

857

858 **5.1 YOU19**

859 The primary change we made to YOU19 was the implementation of a new APWP for Gondwana from
860 540 to 320 Ma and a GAPWaP for 320 to 0 Ma (Tetley, 2018). We also adjusted the position of Laurussia
861 during the Devonian in order to better fit palaeomagnetic data and implemented an alternative position for
862 Lhasa (against northwest Australia) that we consider to be more consistent with geological data in the
863 Neoproterozoic (Section 5.4.1).

864

865 In the YOU19 reconstruction, Laurussia is rotated roughly 40° counter-clockwise compared to the DT14
866 and MAT16 reconstructions (Fig. 11). While these changes improved the global kinematic integrity of the
867 model (Young et al., 2019), the position of Baltica in this configuration conflicts with palaeomagnetic
868 data, which indicates it lay at more southerly latitudes (Torsvik et al., 2012). We used the compilation of
869 Torsvik et al. (2012) as a base, however, as many of the poles in that dataset for the late Silurian to early
870 Devonian were taken from the British Isles (c.f. Jeleńska et al., 2015), we supplement it with some data
871 from cratonic, continental Europe (Table 2, S. Pisarevsky pers. comm.). We modify the position of
872 Laurussia from that used in YOU19 such that it fits the cluster of poles at this time (Fig. 11b). This
873 modification places it in a similar latitudinal position to DT14 and MAT16 (e.g. Fig. 7a), however, we
874 shift it further east relative to DT14, so that it is closer to its final position in Pangea. This position then
875 satisfies the kinematic and structural issues outlined in YOU19, while maintaining Baltica and Laurussia
876 at a palaeolatitude permitted by palaeomagnetic data.

877

878 The Lhasa block, currently preserved in the Tibetan Plateau between India and Tarim, is an E-W
879 elongated block consisting of Precambrian metamorphic basement, overlain by predominantly Palaeozoic
880 sedimentary rocks and Mesozoic and Cenozoic volcanic assemblages (Yin and Harrison, 2000; Zhu et al.,
881 2013). The Precambrian basement is established only in the southern and central terranes of Lhasa (Zhu et

882 al., 2013) and in the Amdo micro-block that is preserved in the Mesozoic northern Lhasa terrane (e.g. Fig.
883 10b). Lhasa is typically placed outboard of the Tethyan Himalayan terranes, off the northern margin of
884 India within Gondwana (e.g. Domeier and Torsvik, 2014). However, an alternative position outboard of
885 northwest Australia is also supported; a scenario consistent with tectonic affinities interpreted from zircon
886 age spectra and Hf isotopic signatures (Burrett et al., 2014; Zhu et al., 2011). We find a position off
887 northwest Australia more consistent with the Neoproterozoic geological record of Lhasa (Section 5.4.1)
888 that preserves magmatism interpreted to represent the existence of a subduction zone and back-arc
889 (Gynn et al., 2012; Hu et al., 2018), which would otherwise be impossible to produce if it was land-
890 locked between India and South China. We therefore alter the position of Lhasa from YOU19 to outboard
891 of NW Australia. For simplicity, we infer Lhasa's motion to follow the drift of the Cimmerian terranes,
892 but recognise that the palaeolatitudes need further refinement in the Jurassic and Cretaceous following
893 syntheses such as Li et al. (2017, 2016) and others.

894

895 **5.2 DOM16 and DOM18**

896

897 We sought to preserve, as closely as possible, the palaeolatitudes and the tectonic interpretations (i.e.
898 history of collisions, rifting, subduction and ocean basin evolution) of the DOM16 and DOM18
899 reconstructions. The position of Gondwana (both in Early Palaeozoic and the older times covered by
900 MER17) has been adjusted to inherit the Gondwana position at 410 Ma from the new Gondwana APWP
901 path described above. Relative rotations of smaller blocks to Gondwana were calculated from DOM16/18
902 and translated into the new position of Gondwana. Further adjustments to all continental polygons have
903 been implemented to smooth the transition from the late Palaeozoic configurations inherited from the
904 adjusted YOU19 model (adj-YOU19). The following sections detail adjustments to the DOM16/18
905 models made for the late Cambrian–Devonian.

906

907 Figure 11 shows a direct comparison between the DOM16/18 reconstructions and our adj-YOU19 model
908 at 411/410 Ma. The primary differences between the models are longitudinal, as our adj-YOU19 model
909 shifts Laurussia, Siberia, Kazak, Tairm and North China $\sim 30^\circ$ longitudinally to the east compared to
910 DOM16/18 (as well as when compared to DT14 and MAT16, which link closely to DOM16/18), in order
911 to better model the amalgamation of Laurussia and Gondwana. This results in a far narrower Rheic Ocean
912 at 410 Ma in our model and a much simpler collision between Laurussia and Gondwana (e.g. Wu et al.,
913 2020; Young et al., 2019). This change is discussed further in context in the following section.

914

915 The motions of Baltica, Laurentia, and Siberia are tied directly to the spin axis through their own
916 individual palaeomagnetic reference frame, in contrast to the younger parts of the reconstruction where
917 we retain the hierarchy inherited from YOU19. For minor cratons (Tarim, North China, South China) and
918 smaller blocks (e.g. Tianshan, Kara), we model their motion within a relative hierarchy for the practical
919 reason that this makes it easier to preserve the consistency of their geological history, as they share
920 multiple plate boundaries with Gondwana. For all these cratons and blocks, we sought to preserve the
921 palaeolatitudes from the parent studies to a reasonable degree allowing for data uncertainties (e.g. Fig. 6).
922 Figure 7 provides a quantitative comparison between the paleolatitudes of the DOM16/18 models and our
923 incorporation of them into our adjusted reconstruction.

924

925 *5.2.1 Deviations from DOM16*

926

927 The key deviation between DOM16 and our model is a difference in the orientation of Baltica in the late
928 Cambrian (Fig. 12a), where our model has Baltica rotated 90° counter clockwise relative to DOM16. As
929 the DOM16 model only begins at 500 Ma it does not have to explicitly consider the earlier
930 Neoproterozoic history of Baltica. In our opinion the position of Baltica in DOM16 is more congruent
931 with an inverted Baltica (relative to Laurentia) during the Neoproterozoic (Hartz and Torsvik, 2002),
932 where the southern-peri Urals are connected to Greenland. In comparison, MER17 connects Baltica to
933 Rodinia through the Sveconorwegian margin in an upright position relative to the present-day (e.g.
934 Cawood et al., 2003; Dalziel, 1992; Weil et al., 1998). The inverted Neoproterozoic Baltica position of
935 Hartz and Torsvik (2002) results in simpler kinematic motions during the late Ediacaran and Early
936 Palaeozoic, whereas the ‘traditional’ Neoproterozoic coupling of Baltica-Laurentia requires a more
937 complex motion path in order to fit palaeomagnetic data. However, in our opinion the ‘upright’ coupling
938 of Baltica and Laurentia is far more geologically and palaeomagnetically robust (Cawood and Pisarevsky,
939 2006; c.f. Slagstad et al., 2019) in the Neoproterozoic than the alternative, and as such we adopt this
940 configuration at ca. 600 Ma (as in MER17) during the opening of the Iapetus Ocean; necessitating a more
941 complex kinematic evolution for Baltica between 600 and 470 Ma (Fig. 12a ,b). Our reconstruction of
942 Baltica is therefore quite different to that of DOM16 during the late Cambrian and early Ordovician as we
943 have to implement a ~90° rotation between 520 and 475 Ma of Baltica to fit the same series of
944 palaeomagnetic data at 470 Ma as DOM16.

945

946 Baltic palaeomagnetic data compiled by Meert et al. (2014), Torsvik et al. (2012) and Domeier (2016)
947 were used to ensure its latitudinal position remained valid between 550 and 470 Ma (Table 2).
948 Importantly, only two poles—one from the Andarum Shale (Torsvik and Rehnström, (2001) categorised

949 as C-grade quality by Meert (2014)) and the other from the Narva sediments, (Khramov and Iosifidi,
950 (2009), categorised as B-grade quality by Meert (2014))—are identified for Baltica between 530 and 480
951 Ma, both with a nominal age of 500 Ma. Despite coeval ages, the poles are $\sim 35^\circ$ apart from one another
952 and Meert (2014) identified unresolved issues, specifically: few samples constraining the pole and a
953 (possible) effect of inclination shallowing (even after correction for inclination shallowing a $\sim 25^\circ$
954 mismatch remains). A strict fitting of the Narva sediments pole would require Baltica to rotate by $\sim 2^\circ/\text{Ma}$
955 between 550 and 500 Ma; a situation we consider implausible and likely reflecting underlying issues in
956 the palaeomagnetic data. As such, our model does not fit either pole explicitly, though our reconstructed
957 position of Baltica is consistent with the latitude suggested by the inclination data of both poles. By 460–
958 440 Ma, our model closely resembles the DOM16 model, with similar sized Rheic oceans, latitudinal
959 positions and orientations of Gondwana, Baltica and Laurentia (Fig. 12c, d). At 410 Ma, when DOM16
960 finishes, the only difference is the relative position of Laurasia and Gondwana due to differences in
961 palaeolongitude (Fig. 2c; 11). In our model, Laurussia is much closer to Gondwana resulting in a far
962 narrower Rheic Ocean at 410 Ma than in DOM16/18, DT14 and MAT16 (e.g. Fig. 2c, see also Wu et al.
963 (2020)).

964

965 The width of the Rheic ocean is poorly constrained (c.f. Dalziel and Dewey, 2019; Domeier, 2018; Wu et
966 al., 2020) especially because of palaeolongitudinal uncertainty. The problem is compounded by the fact
967 that there is only one reliable palaeomagnetic pole constraining our Gondwana APWP between 430 and
968 400 Ma (Aïr intrusives in Niger, age at 410 Ma, (Hargraves et al., 1987)), meaning the early Devonian
969 portion of the APWP has large uncertainty. Because of this uncertainty, the methods used to create
970 APWPs tend to dampen the effect of this pole (which is true for our APWP). Nonetheless, even with the
971 effect of this pole being dampened, reconstructed palaeomagnetic data (without considering longitudinal
972 constraints) at 410 Ma allow placement of Laurussia and Gondwana to within a few thousand kilometres
973 of one another (e.g. Fig. 11a, see also Wu et al. (2020)). At first glance, this seems problematic, because
974 one might expect that the forces related to subduction zones modelled along the craton margins facing
975 each other would draw the cratons to each other. DOM16, DT14 and MAT16 (along with Torsvik et al.,
976 (2014)) solve this problem by changing the palaeolongitude of Laurussia to 90° east relative to
977 Gondwana, allowing for a much wider ocean basin. This position is justified and necessitated in these
978 models by their use of the PGZ method, where Laurussia is reconstructed over the eastern arm of the
979 present-day position of the Pacific LLSVP. However, this position then requires > 8000 km of dextral
980 motion between Laurussia and the Patagonian margin of Gondwana from 400 and 340 Ma (e.g. Fig. 3c).
981 As we do not adopt the PGZ method, our position of Laurussia at 410 Ma relative to Gondwana is much
982 closer to its final collision place and the resulting relative motion between Laurussia and Gondwana

983 between 410 and 340 Ma is of sub-orthogonal–orthogonal collision along the southern Appalachian zone
984 (Hopper et al., 2017), with some dextral transform motion between north-east Laurentia and the northwest
985 margin of Gondwana (e.g. Murphy et al., 2011) and between southern Baltica and north Gondwana (e.g.
986 Arthaud and Matte, 1977). From 340 to 320 Ma YOU19, MAT16 and DT14 have a similar configuration
987 between Laurussia and Gondwana. We stress that while our model this differs from the adopted DOM16
988 model, it is not a particularly novel interpretation and many continental reconstructions show a similar
989 scenario (e.g. McKerrow et al., 2000; Scotese, 2004; Stampfli and Borel, 2002). A more detailed
990 geological and kinematic justification of this interpretation is presented in Young et al. (2019) and a
991 discussion of the palaeomagnetic challenges in amalgamating Pangea are presented in Domeier et al.
992 (2020, 2012) which we encourage interested readers to.

993 994 *5.2.2 Deviations from DOM18*

995
996 Two salient points made by Domeier (2018) pertaining to relative longitude and his model are also of
997 interest here, and worth reiterating when merging DOM18 into the adj-YOU19 model and then extending
998 the adj-YOU19 model into the Neoproterozoic. Firstly, a long lived south-dipping subduction zone,
999 preserved in the northern Kazakhstan terranes of Urumbai, Selety and Erementau (Degtyarev, 2011;
1000 Domeier, 2018; Windley et al., 2007) was longitudinally distributed from Siberia through to the
1001 northernmost margin of Gondwana (i.e. North Australia, Papua New Guinea) and secondly, the broader
1002 framework of Baltica-Siberia-Gondwana provides geological, spatial and temporal limits on the evolution
1003 of this area. These two aspects of the model allowed DOM18 to infer with some certainty the relative
1004 palaeolongitude of many of the smaller terranes within the broader absolute palaeolongitudinal constraints
1005 imposed by (principally) Siberia, Baltica, Laurentia and Gondwana. Likewise, even though we do not
1006 adopt the absolute longitudinal framework of DOM18, we can use the positions of the major cratons to
1007 provide a relative control on the possible kinematic evolution of the Chinese cratons and terranes. We
1008 follow DOM18 by reconstructing a quasi-stable south-dipping subduction zone to delimit the northerly
1009 extent of the Palaeo-Tethys Ocean and this, coupled with the palaeomagnetic data from these cratons and
1010 terranes between 500 and 410 Ma, make longitudinal re-ordering of these blocks unlikely. The
1011 unlikelihood of longitudinal re-ordering is one of the main constraints and pieces of evidence that lead to
1012 significant revision of the positions of Tarim and North China in MER17 that are discussed in the
1013 following section (5.4).

1014
1015 The relative positions of terranes (Tianshan, Qaidam-Qilian, Kunlun etc.) can also be considered in a
1016 similar manner to how we conceive of the relative ordering of the Chinese cratons during the Early

1017 Palaeozoic. In particular, the ca. 470 Ma suturing of Tianshan-Chu Yili (Alexeiev et al., 2019) and the
1018 440–430 Ma suturing of Kunlun, Qaidam, Qilian and Alxa to Tarim (Xiao et al., 2009) both necessitate
1019 an internally consistent relative position between these terranes and Tarim and Siberia in order to be
1020 consistent with the geological record. That is, relative to present-day Tarim, Tianshan, Chu-Yili and other
1021 Kazakh terranes should be reconstructed somewhere north of the *northern* margin of Tarim during the
1022 Ediacaran–Ordovician, but south of Siberia. Similarly, Kunlun, Qaidam, Qilian and Alxa need to be
1023 reconstructed to the *south* and *east* of Tarim in the same time period. In a plate model framework, this
1024 means that the longitudinal structure of the Proto-Tethyan ocean basin should reconstruct (from west to
1025 east): Siberia—Chu-Yili-Kazakh-Tianshan—Tarim—Alxa-Qaidam-Qilian-Kunlun—North China—South
1026 China (e.g. Fig. 13f, g). Further back in time, we also maintain this same configuration to minimise any
1027 terrane re-organising, such that their relative positioning is broadly reminiscent of present-day (e.g. Figs.
1028 9; 16). In this manner, we use these relative longitudinal constraints to infer a configuration for the nuclei
1029 of these terranes in the Neoproterozoic, thereby connecting the present-day with their Precambrian
1030 history.

1031
1032 The fundamental difference between DOM18 and our implementation of the model is the longitudinal
1033 width of the Proto-Tethyan ocean basin, bounded by Siberia in the west, Gondwana in the east and south
1034 and the aforementioned south-dipping subduction zone in the north (e.g. Fig. 13b, c, f, g). The latitudinal
1035 extent of the ocean basin remains similar in both models (30–50°), constrained by palaeomagnetic data
1036 from Gondwana, Siberia and the Chinese cratons (e.g. Table 2, Fig. 13a, d, e, h). In our model, this ocean
1037 basin is much wider (longitudinally) in the late Cambrian and Ordovician than in DOM18, narrowing in
1038 size as it evolves due to our implemented easterly drift of Siberia from the Cambrian through to the
1039 Devonian. The size of the ocean basin is then similar at 410 Ma (Fig. 11), due to the adopted similarity of
1040 the YOU19 from the DT14 reconstruction (Young et al., 2019). The key reason for the difference in width
1041 at 500 Ma is because DOM18 places Laurussia further east at 500 Ma than we do (Section 5.2.1). This
1042 then forces a narrower ocean basin between 500 and 410 Ma in DOM18 than in our model. The following
1043 paragraphs will discuss the regional longitudinal constraints of this ocean basin by considering the
1044 position of Gondwana and Siberia.

1045
1046 The Early Palaeozoic position of Siberia in our model is a function of palaeomagnetic data and its
1047 position in, and breakout from, Rodinia. In isolation, the simplest explanation of Siberia’s journey is that
1048 sometime during the late Tonian–Cryogenian (750–700 Ma) Siberia rifted off the northern margin of
1049 Laurentia (somewhere near Greenland, see Pisarevsky and Natapov (2003) and Pisarevsky et al. (2013)).
1050 At this time Siberia was located equatorially and rotated 60° clockwise from its present-day orientation

1051 (Pisarevsky et al., 2013). Palaeomagnetic data are sparse for the remainder of the Neoproterozoic, with
1052 the few calculated poles having either poor age constraints or unresolved tectonic coherence with the
1053 Siberian craton (Pavlov et al., 2015), and are therefore typically omitted from syntheses (Cocks and
1054 Torsvik, 2007; Li et al., 2008; Merdith et al., 2017a). However, from the mid-Cambrian the
1055 palaeomagnetic record of Siberia is reasonable (Cocks and Torsvik, 2007) and broadly congruent with the
1056 palaeolatitude of the 750 Ma pole (that is, equatorial–sub-equatorial). From the mid-Cambrian, the data
1057 suggest a slow northward drift and counter-clockwise rotation (Cocks and Torsvik, 2007), with the
1058 orientation of Siberia at ca. 530 Ma inverted relative to present-day. This then requires a ~120° clockwise
1059 rotation between 720 and 530 Ma in order to fit its Neoproterozoic position (Metelkin et al., 2012). For
1060 the Cryogenian and Cambrian, Siberia's motion can be inferred indirectly using data from Baltica, as
1061 outlined in the next paragraph.

1062
1063 The position of Baltica is constrained by clusters of palaeomagnetic data during the Ediacaran and early
1064 Cambrian. Furthermore, its latitudinal position places limits on the possible position of Siberia. The
1065 equatorial excursion of Baltica in the latest Ediacaran places Baltica at a latitude similar to Siberia. As
1066 they cannot be positioned on the same longitude (Merdith et al., 2017a), Siberia must be located either
1067 longitudinally east or west of Baltica between 600 and 500 Ma (with Laurentia also occurring on a similar
1068 palaeolatitude but further east than either Siberia or Baltica). Domeier (2018) presented similar arguments
1069 for the second latitudinal excursion of Baltica (it returns to a high latitude during the Ordovician–Silurian)
1070 when it collided with Laurentia to form Laurussia at ca. 440–430 Ma. He argued that the overlapping
1071 palaeolatitudes of Siberia and Baltica at this time then requires Siberia to be located more easterly than
1072 Baltica by ca. 470 Ma (Domeier, 2018). We also adopt this logic, thus providing a rough relative
1073 longitudinal framework for Laurentia-Baltica-Siberia relations from 700 to 450 Ma.

1074
1075 Our model requires the motion of Siberia to be predominantly longitudinal between 700 and 450 Ma (the
1076 available palaeomagnetic data do not suggest more than 30° latitudinal movement). We constrain this to
1077 two broad phases of movement, defined by Baltica's two latitudinal excursions (at ca. 560 Ma and 450
1078 Ma). In the first excursion (750–550 Ma), we keep Siberia longitudinally between Laurentia and Baltica,
1079 because to move it further east than Baltica at 550 Ma would require relative plate motion greater than 30
1080 cm/a, which we deem unlikely. By 470 Ma palaeomagnetic data from Baltica suggest it has started
1081 drifting north again, so we therefore reconstruct Siberia to also be moving east longitudinally between
1082 550 and 470 Ma, such that by 470 Ma it is located along the same longitude as Baltica; by 450 Ma it is
1083 further east than Baltica, such that Laurussia can form by 430 Ma. Our model is therefore similar to that
1084 of DOM18 in concept and adherence to available observations, however the different absolute

1085 longitudinal positions create a notably different ocean basin in the late Cambrian and Ordovician, within
1086 which the Chinese cratons and terranes are then arranged.

1087

1088 Domeier (2018) does not explicitly consider the Neoproterozoic or Early Palaeozoic evolution of the
1089 Chinese cratons in his model. At 500 Ma, the DOM18 model places North China off the northern margin
1090 of India, and South China off northern Australia. However, retaining these relationships in the
1091 Neoproterozoic is invalidated by what Neoproterozoic palaeomagnetic data exists for each craton (e.g. Fu
1092 et al., 2015; Li et al., 2004), and also conflicts with interpreted geological histories (Cawood et al., 2018b,
1093 2013b). Consequently, our model, which balances both older and younger times, alters the kinematic
1094 evolution in order to fit older constraints (Fig. 13, Sections 5.4.1 and 5.4.2).

1095

1096 **5.4 MER17**

1097 *5.4.1 Australia, North China, Lhasa and Tasmania*

1098 Significant alterations to MER17 were made between 1000 and 900 Ma along the north-western, western
1099 and south-western margins of Laurentia, affecting the motions and positions of Australia, North China,
1100 Siberia, Tasmania and Lhasa. We adopt the model of Wen et al. (2018) in having a dextral shear zone
1101 between Australia-Antarctica (A-A) and Laurentia during the early Tonian. Wen et al. (2018) argued for
1102 placing Tarim against the eastern margin of Laurentia, separating Laurentia from Australia in a ‘Missing
1103 Link’ position (cf. (Li et al., 2008, 1995), with the dextral shear zone transecting Tarim. However, we
1104 consider this position for Tarim is incompatible both with geological data (the 760 Ma Aksu Blueschist
1105 (C.-L. Zhang et al., 2013)) and with the kinematic constraints that would be necessary to move Tarim
1106 from this position to its Palaeozoic position (e.g. Merdith et al., 2017b). We therefore use the alteration of
1107 Wen et al.’s (2018) model presented in Mulder et al. (2018b), who place a dextral boundary separating the
1108 Antarctic crust of Australian affinity exposed in Terre Adélie Land from the Antarctic crust of Laurentian
1109 affinity in the Nimrod Igneous Province (Fig. 14a, see also Goodge et al., (2017)).

1110

1111 North China and northern Australia share a similar Mesoproterozoic and early Tonian sedimentary record
1112 and both preserve contemporaneous ca. 1.33–1.31 Ga magmatism (Bodorkos et al., 2020; Yang et al.,
1113 2020; Zhang et al., 2017) that is interpreted as a large igneous province. Yang et al. (2019) also
1114 demonstrated the similarity in detrital zircon ages and hafnium isotope compositions between the Tonian
1115 strata of both areas. We find that this position of North China in the latest Mesoproterozoic is remarkably
1116 compatible with the few reliable palaeomagnetic data for North China in the Neoproterozoic (e.g. Fig 6,
1117 (e.g. Fig. 6, Fu et al., 2015) and places North China in a position readily compatible with its Palaeozoic
1118 constraints where the same species of distinctive tommotiid fossils have recently been reported (Pan et al.,

1119 2018). A distinct Sino-Australian Cambro–Ordovician faunal province was identified by Burrett et al.
1120 (1990) that suggests some proximity in the early Palaeozoic. Cambrian–Ordovician rifting in the Arafura
1121 Basin north of northern Australia may represent the initial separation of North China from this margin of
1122 Gondwana (Ahmad and Munson, 2013). Palaeomagnetic data necessitate some relative motion between
1123 Australia and North China in the Early Palaeozoic from its inferred Mesoproterozoic–Neoproterozoic
1124 position (e.g. Domeier, 2018). Given there is no evidence of orogenesis between North China and
1125 northern Australia in the Phanerozoic, we infer that North China slowly drifts off this margin from the
1126 Cambrian, remaining in close enough proximity to share the identified Cambro–Ordovician faunal
1127 provinces (e.g. Fig. 13).

1128

1129 Dong and Santosh (2016) and Dong et al. (2014) describe a 1000 to 900 Ma suture between the Qinling
1130 Terrane and North China, preserved as the Kuanping Ophiolite (Fig. 14a–c). As Siberia and Australia are
1131 reconstructed adjacent to each other in Rodinia (Pisarevsky et al., 2013), the position of North China
1132 along the northern margin of Australia suggests that the Qinling terrane could feasibly be an extension of
1133 the Central Angara terrane, where there is a similarly aged (but sparsely described) ophiolite, the
1134 Ribnaya-Panimba ophiolite (Vernikovskiy et al., 2004, 2003). In our model, the subduction zones
1135 represented by these two ophiolites consume the oceanic lithosphere between Australia, Siberia and
1136 Laurentia (the Kuanping Ocean) during the early Tonian. Mulder et al. (2018b) ceased motion at 900 Ma
1137 in their model but, we adjust this cessation to 930 Ma wherein the Qinling Terrane rotates to fit against
1138 the North China block. This is because we also reconstruct the Lhasa block along the western margin of
1139 Australia ((Zhu et al., 2011), see Section 5.1) and here magmatism is preserved from ca. 925 Ma (Guynn
1140 et al., 2012, 2006; Hu et al., 2018; Zeng et al., 2018) (Fig. 14d,e). Consequently, we suggest that this
1141 subduction initiated after the closure of the interior Kuanping Ocean and collision of North China-
1142 Australia-Antarctica with Siberia-Laurentia along the Qinling-Central Angaran Terrane. The subduction
1143 zone then connects northwards through to subduction preserved in Taimyr outboard of Siberia
1144 (Vernikovskiy et al., 2004; Vernikovskiy and Vernikovskaya, 2001) and southward into an oceanic arc
1145 outboard of the Mawson Craton of Antarctica, possibly preserved in the southernmost Tonian Oceanic
1146 Arc Super Terrane (TOAST) or between Indo-Antarctica and Australia-Antarctica). In our model,
1147 Australia sits in a typical SWEAT (South West United States, East Antarctica) configuration (Moore,
1148 1991). We made this change to better fit the arguments put forward by Mulder et al. (2018b), while still
1149 maintaining integrity of relative plate kinematics following the reasoning of Merdith et al. (2017b).

1150

1151 Our revised model also incorporates recent refinements to the Proterozoic and early Palaeozoic
1152 paleogeography of the Western Tasmania Terrane. The Western Tasmania Terrane, comprising the

1153 Proterozoic geology of Tasmania and the East and West South Tasman Rises (Berry et al., 2008),
1154 occupies an important position in deciphering the geological relationship between Laurentia and
1155 Australia-Antarctica in Rodinia, and also in understanding the transition between Rodinia to Gondwana.
1156 The Western Tasmania Terrane represents an exotic Proterozoic microcontinent that was accreted onto
1157 the Pacific margin of eastern Gondwana in the late Cambrian during the Ross-Delamerian orogenic cycle
1158 (Berry et al., 2008; Cayley, 2011). The terrane has geological affinities with the central Transantarctic
1159 Mountains of East Antarctica and the western margin of Laurentia, including overlapping
1160 Palaeoproterozoic basement ages, contemporaneous Mesoproterozoic magmatic and fluid-flow events,
1161 and correlated Mesoproterozoic sedimentary strata (Berry et al., 2008; Fioretti et al., 2005; Halpin et al.,
1162 2014; Mulder et al., 2015, 2018b). Based on these geological connections, the Western Tasmania Terrane
1163 was likely located between East Antarctica and western Laurentia within an assembled Rodinia. The
1164 breakout of the Western Tasmania Terrane from its central position within Rodinia is recorded by
1165 widespread Tonian–Ediacaran sedimentation and rift-related magmatism (Mulder et al., 2020). The onset
1166 of Neoproterozoic rifting of the Western Tasmania Terrane is marked by 780–750 Ma intraplate
1167 magmatism (Black, 1997; Calver et al., 2013) and latest Tonian (750–730 Ma) siliciclastic and carbonate
1168 sedimentation (Calver et al., 2014; Mulder et al., 2018a). Following deposition of Cryogenian rift-related
1169 strata and glaciogenic intervals (Calver, 2011; Calver et al., 2014) a final pulse of Neoproterozoic rifting
1170 is recorded by voluminous ca. 580 Ma rift-related basalts in northwest Tasmania (Direen and Crawford,
1171 2003; Meffre et al., 2004). Geological correlations permit the Western Tasmania Terrane to have
1172 remained attached to either the western margin of Laurentia or the eastern margin of Australia-Antarctica
1173 following the opening of the Pacific Ocean (Fig. 10f), prior to being isolated as a microcontinent during
1174 ca. 580 Ma rifting and accretion onto its present-day position along the margin of Gondwana by the late
1175 Cambrian (Fig. 10g; Berry et al., 2008; Mulder et al., 2020). We follow Mulder et al. (2020) in having
1176 Tasmania rift from the Antarctic margin (rather than the alternative scenario of Laurentia), thus implying
1177 that some further unknown micro-continents rifted from the western margin of Laurentia at the same time
1178 in order to account for that passive margin (e.g. Fig. 14f, g Macdonald et al., 2013) (see also Colpron et
1179 al., 2002; Cox et al., 2018; Eyster et al., 2019).

1180

1181 *5.4.2 India-South China Accretionary Belt*

1182 We preserve the MER17 interpretation of a tight India-South China connection (after Cawood et al.,
1183 2013b). This possible connection was suggested previously by Jiang et al. (2003) who noted the similarity
1184 between sequence stratigraphy in rift basins preserved in both South China and the Indian Lesser
1185 Himalaya (Fig. 15a), as well as by Hofmann et al. (2011) who suggested a geological similarity based on
1186 detrital zircon analysis. Arguments for this connection are succinctly summarised in Cawood et al.

1187 (2018b) and are not repeated here—instead we focus our discussion of this margin on the relative position
1188 of outboard terranes during the Neoproterozoic (Fig. 15), which are based predominantly on their
1189 Palaeozoic positions (Domeier, 2018). Here we have sought to preserve their relative internal positions in
1190 order to minimise reshuffling of terranes during the Neoproterozoic (e.g. Fig. 16). For example, Kunlun,
1191 which is currently preserved south of Tarim and west of Qaidam-Qilian, is always reconstructed with the
1192 same internal consistency. Although we note that this may not be an accurate reflection of the ordering
1193 and positioning of the terranes, it ensures consistency within the model and minimises terrane shuffling
1194 which can preclude unrealistic scenarios, where terranes have to kinematically skirt one another
1195 precariously. Figure 16 gives a schematic overview of our conceptualisation and implementation of this
1196 model.

1197

1198 Following Alessio et al. (2018) and Armistead et al. (2019), the northwest margin of India is here
1199 interpreted as an extensive Stenian-Tonian accretionary margin that extends as far as the Omani basement
1200 and northernmost Madagascar. The pre-Ediacaran basement rocks in Rajasthan and Pakistan share
1201 similarities with those of Oman. Granitoids have been dated from Rajasthan and from Nagar Parkar in
1202 eastern Sind (Pakistan) at ca. 1.1 Ga (Meert et al., 2013; Raza et al., 2012). There is no evidence of older
1203 crust occurs west of the Western Margin Fault of the Aravalli-Delhi Orogen, where the Marwar terrane
1204 accreted to India in the latest Mesoproterozoic (Meert et al., 2010) (Fig. 15b). Tonian granitoids and
1205 rhyolites occur in inliers through northwest India and Pakistan, where they cluster into crystallisation ages
1206 of ca. 990–970 Ma (Haldar and Deb, 2001; Pandit et al., 2003), ca. 860–820 Ma (Davies and Crawford,
1207 1971; Deb et al., 2001; Just et al., 2011; Van Lente et al., 2009) and ca. 775–760 Ma (Ashwal et al., 2013;
1208 Gregory et al., 2009; Meert et al., 2013; Van Lente et al., 2009). The latter magmatic and extrusive phase
1209 forms one of the largest felsic igneous provinces on the planet—the Malani Igneous Suite—which is also
1210 traced to the Seychelles (Fig. 15c–e) (Torsvik et al., 2001; Tucker et al., 2001). Arc accretion continued
1211 outboard to Oman where two main phases of subduction and arc magmatism occur, at ca. 850 Ma and ca.
1212 770 Ma (Blades et al., 2019a). The latter phase focussed in the southern Mirbat area and interpreted here
1213 as the arc that formed ocean-ward of the more back-arc Malani Igneous Suite. Further outboard still, and
1214 later accreting onto the Indian margin, the Bobakindro Terrane of northern Madagascar (Armistead et al.,
1215 2019) consists of juvenile magmatism that dates from ca. 750–705 Ma (Armistead et al., 2019; Collins,
1216 2006; Thomas et al., 2009).

1217

1218 Many terranes currently preserved north of South China and south of Siberia have Neoproterozoic or
1219 older cores. They have not been previously considered in global models (Li et al., 2008; Merdith et al.,
1220 2017a) due to the sparsity of data and small size of terranes, which invites many competing and

1221 conflicting interpretations of their history. However, in the construction of this continuous plate model,
1222 where spatial and temporal continuity is vital, the most compatible Tonian position for these terranes was
1223 outboard of the afore-discussed large accretionary subduction zone of South China and India. This
1224 position places them in the most favourable kinematic, palaeomagnetic and geologically plausible
1225 positions for their (more well constrained) Palaeozoic journeys (e.g. Charvet et al., 2011; Domeier, 2018;
1226 Xiao et al., 2013). Below we summarise some geological evidence for this, with particular reference to
1227 the Tarim Craton, as it is the only block that has multiple reliable palaeomagnetic data from the
1228 Neoproterozoic that act as a another line of evidence. We also note that Huang et al. (2019) recently
1229 proposed a location outboard of Greenland for the Yili-Tianshan Block on the basis of similar detrital
1230 zircon age spectra. However, more work would have to be done to determine whether this Tonian position
1231 is consistent with the kinematic evolution necessary for these blocks to fit their Palaeozoic constraints.

1232
1233 Early Tonian age (1000–900 Ma) magmatism and high-pressure metamorphism is preserved in the
1234 basements of the Qilian-Qaidam (Qi-Qa), Kunlun and Tianshan-Chu Yili terranes (Song et al., 2012;
1235 Tung et al., 2007; Wu et al., 2017). Importantly Song et al. (2012) identified an early Tonian event
1236 preserved in a high-pressure metamorphic belt in Qi-Qa. Here, a ~200 km linear belt of granitic gneisses,
1237 metamorphosed in the Palaeozoic, have crystallisation ages between ca. 1000 and 900 Ma (Song et al.,
1238 2012). Zircons recovered from pelitic and psammitic gneisses from the same belt possess multiple
1239 generations of growth, as suggested through cathodoluminescence imaging, and return ages of the (first
1240 generation) to between ca. 940 and 900 Ma (Song et al., 2012). These are interpreted to represent a period
1241 of granulite facies metamorphism from a continental arc indicating that subduction was active during the
1242 Early Tonian (Song et al., 2012). Song et al. (2012) and others (e.g. Zhang et al., 2008) suggested a link
1243 between these two blocks and South China on the basis of similar-aged magmatism and metamorphism.
1244 However, as the Qi-Qa preserves a different Palaeozoic tectonic history to South China, as opposed to
1245 fragments of older lithosphere preserved in the Panxi-Hannan Belt of the Yangtze Craton, we suggest that
1246 a subduction zone was located outboard of Qi-Qa while a secondary, smaller ocean was closing between
1247 Qi-Qa and the accretionary orogen of the Panxi-Hannan Belt (Fig. 11a–c). Upon the suturing of Yangtze
1248 with Cathaysia (ca. 900 Ma, Fig. 11b), subduction relocated outboard of South China and began to close
1249 the ocean between Qa-Qi and South China. Similar to the Qi-Qa, the Kunlun terrane preserves scattered
1250 magmatic ages of S-type granites and protoliths of orthogneiss and amphibolites ranging between ca.
1251 1000 and 900 Ma (Chen et al., 2008; He et al., 2018; and Chen et al., 2006a; 2006b—both cited in Chen
1252 et al., 2008; He et al., 2018). We interpret these rocks and ages as an extension of the same subduction
1253 zones outboard of Qi-Qa, and extend it further to the south where (again) similar-aged magmatism is also
1254 preserved in the North Tianshan and Chu-Yili (Degtyarev et al., 2017).

1255

1256 The Kazakh terranes (including Krygyz Tianshan and Chu-Yili, Fig. 15a) have poorly constrained
1257 Neoproterozoic histories, with only a handful of ages from outcropping magmatic rocks and sedimentary
1258 successions. We predominantly follow the summary of Degtyarev et al. (2017) in offering a possible
1259 tectonic interpretation of their Neoproterozoic geological history that is linked to the wider globe.
1260 Degtyarev et al. (2017) note that there are two broad categories of Precambrian terranes preserved in the
1261 Krygyz-Tianshan-Yuli area; the Issedonian and Ulutau-Moyunkum terranes. The Issedonian terranes,
1262 preserved in the northeast of the western Central Asian Orogenic Belt, include Chu-Yili and the Chinese
1263 Central Tianshan and are characterised by late Mesoproterozoic magmatism (e.g. Degtyarev et al., 2011),
1264 thick (> 1000 m) 1050–950 Ma quartzite-schist successions followed by ongoing magmatism from ca.
1265 960–890 Ma (e.g. Degtyarev et al., 2008; Gao et al., 2015; Huang et al., 2015). Comparably, the Ulutau-
1266 Moyunkum terranes are preserved only in the west (in Krygyz) within Krygyz Middle Tianshan and
1267 Krygyz North Tianshan (e.g. Fig. 10b), and consist of a Palaeoproterozoic basement, with predominantly
1268 sedimentary Mesoproterozoic and early Neoproterozoic rocks (Degtyarev et al., 2017). Magmatism,
1269 between 840 and 760 Ma (Kröner et al., 2012) and granulite facies metamorphism from 800–760 Ma
1270 (Degtyarev et al., 2017; Tretyakov et al., 2016) are recorded only in the late Tonian. Both sets of terranes
1271 preserve distinct differences in their Mesoproterozoic histories, minor differences in the Early
1272 Neoproterozoic histories but similar histories from the mid-Neoproterozoic (ca. 700 Ma) onwards,
1273 suggesting proximity sometime during the late Tonian (800 to 700 Ma?) (Degtyarev et al., 2017). We
1274 interpret the Issedonian terranes to be the southernmost extent of the subduction zone spanning Qi-Qa and
1275 Eastern Kunlun, as the magmatism preserved in the Issedonian terranes has a continental arc signature
1276 (Huang et al., 2014) and is broadly coeval. Comparably, the Ulutau-Moyunkum terranes, which share
1277 Mesoproterozoic similarities to Tarim and record no magmatism in the early Neoproterozoic, are located
1278 on the opposite side (lower plate) of a closing ocean basin. This culminates with the collision of Tarim
1279 (with the Ulutau-Moyunkum terranes) and the combined India-South China continent at ca. 800–760 Ma
1280 along the Issedonian margin (Fig. 11d, e). The sparse data and age constraints from these terranes means
1281 much of their Neoproterozoic history is conjectural. Although the specific orientation and positioning of
1282 the terranes along the margin is speculative our interpretation is that it places the Kazakh terranes in a
1283 favourable position for their Palaeozoic evolution which is, comparably, much better constrained.

1284

1285 Two clusters of Neoproterozoic-aged palaeomagnetic data from Tarim make it difficult to elucidate a
1286 consistent position with other palaeomagnetic data from Rodinian constituents. Three poles from the
1287 Tonian and Cryogenian require a 90° rotation of Tarim in order to fit the younger pole of Wen et al.
1288 (2017), or a 180° rotation to fit the cluster of three poles in the Ediacaran to Cambrian (see Merdith et al.

1289 (2017a) and Section 4.2 for a discussion). Within a self-consistent kinematic plate boundary framework,
1290 this motion is not permissible if Tarim is positioned either against north-western Australia (e.g. Zhang et
1291 al., 2012) or as an extension of the ‘Missing-Link’ model (Li et al., 2004 for the original proposal of the
1292 “Missing-Link” model; Wen et al., 2018, 2017). A plausible position where these palaeomagnetic criteria
1293 are met, along with satisfying key geological evidence, such as the 800–760 Ma Aksu Blueschist
1294 preserved on the northern margin of Tarim(C.-L. Zhang et al., 2013), is outboard of the India-South China
1295 accretionary belt, where it acts as the final piece of continental lithosphere accreted to the margin. In our
1296 model, we suggest the metamorphism recorded by the Aksu Blueschist marks the accretion of Tarim to
1297 Chu-Yili and the Tianshan (see Xia et al. (2017) for the most recent discussion, but see also data and
1298 discussion from Zhang et al. (2012, 2009). To accommodate the change in position suggested by the
1299 palaeomagnetic data, we introduce a $\sim 120^\circ$ rotation of Tarim and Southern Tianshan away from this
1300 margin, such that Tarim’s southern margin collides with the outboard margin of Alxa-Qaidam-Qilian, so
1301 its northern margin faces an open ocean basin, allowing it to rift northward facing as Gondwana forms
1302 towards its more well constrained Palaeozoic position (Fig. 16).

1303

1304 To support this interpretation of Tarim’s evolution we present the following geological observations in
1305 support of this model. Firstly, there is an absence of extensive magmatism on either the northern or
1306 southern margin of the Tarim craton between 1000 and 850 Ma, which makes it difficult to include as a
1307 part of the upper-plate circum-Rodinian subduction girdle (Cawood et al., 2016). Previous studies,
1308 including MER17, place Tarim on the margin of Rodinia typically also include the Chu Yili and Tianshan
1309 crust attached in a quasi-present-day configuration to the northern margin of Tarim. In such cases their
1310 record of Tonian magmatism supports the interpretation that they formed part of the circum-Rodinian
1311 subduction girdle (Ge et al., 2014) however, this location is inconsistent with available palaeomagnetic
1312 data (e.g. (Wen et al., 2018). Secondly, rift related granitoids preserved in the Southern Tianshan
1313 (Degtyarev et al., 2017; Gao et al., 2015) are here interpreted as evidence of the re-adjustment of Tarim
1314 between 730 and 680 Ma to account for the change in palaeolatitude and orientation inferred from
1315 palaeomagnetic data (Fig. 11e). In addition, extensive rifting in the southwest of Tarim (Wang et al.,
1316 2015) is interpreted to reflect the rearrangement of subduction after Tarim/India-South China
1317 amalgamation. This motion is similar to the adjustment of Baltica relative to Laurentia in the latest
1318 Mesoproterozoic proposed by Cawood et al. (2010) to account for the Valhalla Orogeny. Xiao et al.
1319 (2010) summarise the geology and geochronology of rocks found in the Beishan area of China which
1320 record a protracted and complex history of multiple arc development and accretion through the late
1321 Neoproterozoic and Early Palaeozoic. Rocks in the Beishan area (Alxa, Fig. 15, 16) range from low-grade
1322 sedimentary metamorphic assemblages to gneiss and eclogite complexes and intrusive granitic bodies that

1323 have late Neoproterozoic–Cambrian ages (Xiao et al., 2010). These rocks, inferred to represent an active
1324 subduction zone and accretionary complex, are unconformably overlain by Cambrian–Ordovician
1325 sediments. The earliest record of metamorphism in the area are from a series of SHRIMP U-Pb ages taken
1326 from the metamorphic rims of zircons of an eclogite unit at ca. 830–800 Ma (Yang et al., 2006).

1327

1328 With respect to the palaeomagnetic issues of Tarim we raised earlier (Section 4.2), we find that our
1329 conceptual model of this rotation of Tarim (e.g. Fig. 16) can fit either set of palaeomagnetic data equally
1330 as well, with the key factor being the time of subduction in the Beishan area. Under the cluster of three
1331 poles, the motion of Tarim occurs more quickly and peak subduction (possibly resulting in a collision?)
1332 would be earlier (ca. 650 Ma), while to fit the pole of Wen et al. (2017) it would occur later at between
1333 600 and 550 Ma. Based on the review of Xiao et al. (2010), we infer that the geological data support the
1334 later interpretation more strongly however, given the novelty of this scenario and the absence of
1335 identifiable piercing points, it could be revised in the future to fit the alternative scenario.

1336

1337 Late Neoproterozoic–Early Cambrian rifting events are inferred to have occurred within all the terranes
1338 (Kunlun, Qa-Qi, Chu Yili and Tianshan) that we have placed on this Indian-South Chinese accretionary
1339 margin, however, the high degree of reworking and suturing of crust from the terranes, coupled with the
1340 small size of these terranes makes it difficult to pin down precise rift times. We stress that our
1341 interpretation here is preliminary, especially when compared to specialised reviews of the tectonics of this
1342 area (Kroner et al., 2007; Wilhem et al., 2012; Windley et al., 2007; Xiao et al., 2013; Yakubchuk, 2017).
1343 We reiterate that our intention here is to provide a possible framework that connects and contextualises
1344 these terranes within a consistent kinematic and tectonic evolution between the Neoproterozoic and
1345 Palaeozoic (Fig. 16), which can be more tightly refined in the future. Ordovician–Silurian sutures
1346 between Qa-Qi-Kunlun and surrounding cratons preserve late Neoproterozoic–Cambrian ophiolites (Jian
1347 et al., 2014; Shi et al., 2018; Song et al., 2013, 2009) thus necessitating the existence of ocean basins, but
1348 there are few dates of ocean basin initiation. Xu et al. (2015) date Qilian-Qaidam continental rift basalts
1349 to 600–580 Ma, constraining ocean basin formation to the latest Ediacaran–earliest Cambrian, with the
1350 oldest ophiolite (the Yushigou Ophiolite) preserved in these terranes dated to 550 Ma (Shi et al., 2004).
1351 Evidence of rifting is more sparse in Kunlun, however, similar stratigraphy, ages and geochemistry of
1352 metavolcanic deposits between Kunlun and Qaidam-Qilian (Yuan et al., 2004) suggest coeval rifting is
1353 reasonable, though not definite. We follow some recent work (Peng et al., 2019; Zhao et al., 2018) in
1354 having separation of Tarim and these terranes in the late Ediacaran (550 Ma here, as a response to the
1355 closure of the Mozambique Ocean between India and Congo). Our model closely resembles the schematic
1356 framework outlined by Qiantao et al. (2001), while also following Domeier (2018) in maintaining a very

1357 close affinity between Kunlun, Q-Q and Alxa for the Early Palaeozoic, such that they conceptually form a
1358 single elongated terrane that rifts off Gondwana at ca. 550 Ma and collides along the southern margin of
1359 Tarim by 440 Ma.

1360

1361 Comparably, rifting on the northern margin of Tarim is easier to constrain. Thick, late Neoproterozoic
1362 sedimentary sequences capped with carbonate and an unconformity on the Ediacaran–Early Cambrian
1363 imply a prolonged rift phase with breakup at ca. 550–540 Ma (Zhao et al., 2014; Zhu et al., 2017).
1364 Similarly, sedimentary assemblages preserved in the South Tianshan orogen are dated from 540–520 Ma
1365 (Alexeiev et al., 2020; Safonova et al., 2016), suggesting that ocean basin formation between Tarim and
1366 Krygyz Middle Tianshan begins in the Early–Middle Cambrian. Further north, the relationship between
1367 the Kryguz North Tianshan and Chu-Yili is also reasonably well established (e.g. Windley et al., 2007;
1368 Xiao et al., 2013). Late Ediacaran–Early Cambrian magmatism preserved in Chu-Yili and North Tianshan
1369 (Alexeiev et al., 2011; Degtyarev et al., 2017; Kroner et al., 2007) is interpreted to represent development
1370 of multiple contemporaneous arcs (e.g. Alexeiev et al., 2020). The Ordovician-aged sutures between Chu-
1371 Yili and North and Central Tianshan (Windley et al., 2007) are defined by ophiolitic slithers, implying an
1372 ocean basin (or a back-arc) existed between these terranes. We infer that in the late Neoproterozoic to
1373 Early Cambrian, the Kazakh and Tianshan terranes that accreted outboard of India-South China were
1374 fragmented and rifted off this margin forming a collage (not dissimilar to modern southeast Asia, or the
1375 NE Pacific in the Mesozoic, e.g. Sigloch and Mihalynuk (2013) that eventually re-assembled in the
1376 Ordovician. We model the time of fragmentation at 550 Ma, because this is the time of collision between
1377 India and Congo along the East African Orogen, so we infer that subduction relocated outboard of the
1378 northern margin.

1379

1380 Palaeomagnetic data from South China do not permit a fixed fit between South China and Gondwana for
1381 the early Palaeozoic. The data do, however, permit a close spatial relationship between them (<1000 km).
1382 Palaeomagnetic data from South China suggest it moved from mid to lower latitudes between the
1383 Cambrian and Devonian (Domeier, 2018; Han et al., 2015). Furthermore, shallow marine faunal data and
1384 detrital zircon arrays suggest that between the Cambrian and Devonian, South China shifted from an
1385 Indian-Himalayan-Iran affinity (Burrett et al., 1990) to Sibumasu-Australian affinity (Cocks and Fortey,
1386 1997; Metcalfe, 2013, 2011), broadly consistent of the positions necessary to fit the palaeomagnetic data.
1387 Rift sequences in South China and in northern India (Himalayan terranes) are similar, but diverge strongly
1388 after the early Cambrian (Jiang et al., 2003), providing some kinematic support for invoking for relative
1389 motion between South China and Gondwana. The Sanya Block of Hainan Island (Fig. 15f) is linked
1390 through detrital zircon provenance and middle Cambrian trilobites to western Australia and Antarctica

1391 rather than South China in the Neoproterozoic (Cawood et al., 2018b; Xu et al., 2014). The presence of
1392 Early Ordovician trilobites in the Sanya block also common to South China and Australia (Torsvik and
1393 Cocks, 2009) support a close relationship between the three domains by this time (Cawood et al., 2018b).
1394 A late Cambrian–Ordovician (520–450 Ma) metabasaltic arc assemblage in Hainan Island (Xu et al.,
1395 2008, 2007) is interpreted to be the northerly extension of the Kunguan Orogen that sutured Australia-
1396 Antarctica and India (e.g. Xu et al., 2014). We introduce divergent motion between Gondwana and South
1397 China at 550 Ma, coinciding with the rift-to-drift sequences of Jiang et al. (2003). This divergent motion
1398 moves South China from a position outboard of Northern India at 550 Ma to one that is slightly further
1399 east and adjacent to western Australia at 500 Ma, accounting for the arc assemblage in Hainan as well as
1400 similar faunal patterns.

1401

1402 5.4.3 ANS-Azania-TOAST

1403 We suggest that, to a first order, the central Arabian-Nubian Shield (ANS) accreted on the kernel of
1404 Azania and formed a semi-continuous archipelago outboard of the eastern margins of the Congo Craton
1405 and Sahara Metacraton (SM) (Fig. 17; Collins and Pisarevsky, 2005; Merdith et al., 2017a). Geological
1406 details and a regional plate model of the accretion of the ANS and Azania are adopted from Collins et al.
1407 (in revision), Blades et al. (2019a) and Johnson et al. (2011), though here we extend Azania to the south
1408 by attaching portions of the Tonian Aged Ocean Arc Super Terrane (TOAST—Jacobs et al. (2017,
1409 2015)). The similarity in ages, petrology of rocks and $\delta^{18}\text{O}$ from zircons between the Dabolava Suite in
1410 Madagascar (Archibald et al., 2018) and TOAST (Jacobs et al., 2017, 2015; Wang et al., 2020) suggest a
1411 similar tectonic environment. As the southern tip of Azania is reconstructed to be adjacent to the location
1412 of the TOAST terrane in Gondwana, at the nexus between the East African Orogen and the
1413 Pinjarra/Kuunga Orogen, there is also a strong palaeogeographic argument for attaching TOAST to
1414 Azania, as their Rodinian reconstructed position requires no alteration for their position in Gondwana
1415 (Fig. 17b).

1416

1417 5.4.4 Hoggar, Borborema, Avalonia and Ganderia

1418 The Hoggar Block is preserved between the Sahara Metacraton (SM) and the West African Craton
1419 (WAC) in northwest Africa and records a long Neoproterozoic history of accretion of island arcs and
1420 continental ribbons. The model incorporated here is based on fieldwork by Caby et al. (1989), Black et al.
1421 (1994) and Liégeois et al. (1994) and involves three main constituents of present-day Hoggar (from west
1422 to east): IOGU-IGU, LATEA and the Air Block (Fig. 18a, b).

1423

1424 The broad tectonic framework of the Hoggar block is an accretionary margin consisting of at least 23
1425 individual terranes that were slowly compressed between two large tectonic units—the WAC and the
1426 SM—as Gondwana amalgamated. The Air block, preserved in the east, is an amalgamation of three
1427 closely related terranes: the Aouzegueur, Barghot and Assodé-Issalane terranes. The first two terranes
1428 accreted onto the margin of the SM by 650 Ma, with the Aouzegueur terrane preserving a tonalite-
1429 trondhjemite-granodiorite (TTG) suite dated at ca. 730 Ma and the Barghot terrane recording calc-alkaline
1430 granitoids from ca. 730 to 660 Ma, with a post nappe pluton preserving a U-Pb zircon age of 664 ± 8 Ma,
1431 interpreted to provide a minimum age for deformation (Liégeois et al., 1994). Both these terranes were
1432 metamorphosed to greenschist or amphibolite facies and were cut by east-verging thrusts. In contrast, the
1433 Assodé-Issalane terranes exhibit younger magmatism (ca. 640–580 Ma) and amphibolite-facies
1434 metamorphism. They are thrust east over the Barghot terrane. Both Black et al. (1994) and Liégeois et al.
1435 (1994) suggest that the Aouzegueur terrane collided first with the SM, followed by the Barghot terrane,
1436 which is positioned slightly further south than the former, through an east-dipping subduction zone
1437 underneath the two terranes (Fig. 18). Following collision, the Assodé-Issalane terrane (which until this
1438 time we position slightly west of the former terranes) was thrust above of the Barghot terrane in response
1439 to the closure of the ocean between the WAC and the SM. Our reconstruction implies that this motion
1440 was predominantly transpressive, along the dextral Raghane shear zone with plutons dating from ca. 630
1441 to 580 Ma (Liégeois et al., 1994).

1442
1443 Further west from the Air block in central Hoggar, the LATEA terranes (Laouni, Azrou-n-Fad, Tefedest
1444 and Egéré-Aleksod) all consist of Archaean to Palaeoproterozoic basement, but preserve no
1445 Mesoproterozoic or early Neoproterozoic rocks. LATEA was a passive cratonic unit for most of the
1446 Neoproterozoic and acted as a nucleus for the accretion of juvenile terranes. The earliest Neoproterozoic
1447 activity is the accretion of the ca. 900 Ma juvenile Iskel island arc to the western margin of LATEA, with
1448 subduction inferred to be west dipping away from LATEA (Liégeois et al., 2003). The protolith of an
1449 eclogitic unit, currently preserved along the shear zone delineating the Iskel arc and LATEA, is dated to
1450 ca. 870 to 850 Ma by U-Pb dating from zircons extracted from syn-to-late kinematic plutons (Caby et al.,
1451 1982). The combined In-Ouzzal and Iforas granulite units (IOGU/UGI), which are Palaeoproterozoic
1452 continental ribbons, preserve few Tonian rocks. From ca. 700 to 640 Ma magmatism is recorded
1453 throughout the entire region, suggesting that subduction occurred along both margins of the terranes
1454 (Caby, 2003) (Fig. 18d, e). At 630 Ma, collision between the IOGU/UGI terranes and the LATEA block
1455 occurred, forming the combined present-day central-western Hoggar region.

1456

1457 The final tectonic events of this area involve a two-step amalgamation process of Western and Central
1458 Hoggar (i.e. IOGU/UGI and LATEA) to the Air Shield and SM, and the collision between this landmass
1459 (Hoggar and the SM) and the WAC, where collisional deformation is preserved in the Pharusian and
1460 Dahoymede belts (Merdith et al., 2017a) (Fig. 18f, g). Here, subduction is inferred to have occurred away
1461 from the WAC (i.e. underneath Hoggar) due to the absence of magmatic rocks preserved on the WAC.
1462 Continual dextral deformation is preserved throughout Hoggar until ca. 530 Ma, suggesting that there was
1463 still relative motion until the Cambrian (Liégeois et al., 2003; Paquette et al., 1998). The major regime of
1464 the Hoggar block between the Ediacaran and early Cambrian was transpressive, resulting in extensive
1465 faulting and upwelling of the asthenosphere; causing partial melting of lower Archaean crust in some
1466 areas (Hadj-Kaddour et al., 1998). We interpret the final amalgamation of Hoggar to occur at ca. 580 Ma,
1467 as this is the age of the syntectonic, deformed plutons found among the shear zones that bind the SM and
1468 the WAC. Here, a suite of ages include: Rb-Sr whole rock ages from dykes affected by the transpressive
1469 event yield an age of 592 ± 6 Ma (Hadj-Kaddour et al., 1998); a 583 ± 7 Ma U-Pb age on zircons extracted
1470 from the syn-to-late tectonic, elongated Imezzarene pluton (Lapique et al., 1986); a 594 ± 4 Ma and
1471 593 ± 17 Ma U-Pb age of zircon extracted from the Ohergehém and Adaf plutons, respectively (Henry et
1472 al., 2009). Fezaa et al. (2010) identified younger (ca. 575–555 Ma) deformation in the Murzoq area of
1473 Hoggar, however, they suggested that it was unrelated to the main convergence between WAC and SM.

1474

1475 The Borborema block sits between Congo-SF and the Nigeria-Benin Shield in a reconstructed Gondwana
1476 (Fig. 18a). This small block consists of Archaean–Proterozoic basement that was strongly reworked and
1477 deformed during the Gondwana amalgamation events between Africa and Amazonia (dos Santos et al.,
1478 2010). Magmatism in the Transversal Domain of Central Borborema is thought to represent a local, early
1479 Tonian orogeny called the Cariri Velhos Orogeny (da Silva Filho et al., 2002; dos Santos et al., 2010).
1480 Late Stenian rift deposits are preserved in the Cariri Velhos belt, suggesting that the Pernambuco-Alagoas
1481 domain (PEAL) (the crystalline basement of Southern Borborema) originally rifted from this margin
1482 (Guimarães et al., 2012), before the ocean inverted and closed. Our model places the Northern and
1483 Transversal Domains of Borborema fixed to the Nigeria-Benin-SM blocks, while the PEAL closed the
1484 small Stenian aged relict ocean basin as it collided with the northern and central Borborema provinces by
1485 ca. 920 Ma (Caxito et al., 2014a), forming the Cariri Velhos Orogen. An ocean basin remained on the
1486 southern margin of PEAL until the Ediacaran, as the final collision between Borborema and SF did not
1487 occur until this time forming the Sergipano belt, which preserves relict oceanic crust (e.g. Caxito et al.,
1488 2014b; Ganade de Araujo et al., 2014) (e.g. Fig. 9). To accommodate the Tonian closure, while
1489 maintaining an open ocean basin to the south of PEAL, we follow Caxito et al. (2014a, 2016) who
1490 suggested that the synchronous rifting and aulacogen formation preserved in SF (Pedrosa-Soares et al.,

1491 2001), which in our model is reconstructed to be adjacent to PEAL, are relicts of the divergent motion
1492 necessary to achieve this.

1493

1494 The Avalonian terranes, currently preserved in the east coast of modern day North American and western
1495 Europe, have a well-documented Neoproterozoic history (e.g. Murphy and Nance, 1989). The Avalonian
1496 terrane is interpreted to be underlain by ca. 1.0 Ga juvenile basement on the basis of 1.3–0.8 Ga Sm-Nd
1497 depleted mantle model ages in in younger Neoproterozoic rocks (Murphy et al., 2000; Thorogood, 1990).
1498 These younger rocks consist of magmatic gneiss and plutonic complexes, along with tuffs, pelitic schists
1499 and quartzites. U-Pb ages of the complexes and tuffs range from ca. 750–650 Ma (Bevier et al., 1993;
1500 Doig et al., 1993; Keppie and Dostal, 1998; Krogh et al., 1988; O’Brien et al., 2001) and detrital zircons
1501 from the (meta-)sedimentary rocks suggest they were sourced from a juvenile arc (metapelites, Murphy,
1502 2002) and a cratonic source (quartzite), typically inferred to be Amazonia or the WAC (Nance et al.,
1503 2008). This magmatism is followed by amphibolite–granulite metamorphism from 660–630 Ma (Keppie
1504 et al., 1998; Strachan et al., 2007). Younger magmatism (ca. 640–550 Ma) is more voluminous and
1505 includes abundant arc derived volcanic, plutonic complexes and coeval volcanic-sedimentary successions
1506 (Bevier et al., 1993; Compston et al., 2002; Doig et al., 1993; Nance et al., 2008; O’Brien et al., 2001;
1507 White et al., 2020). Subduction does not continue into the Cambrian, instead a clastic platform and
1508 transition to rift environment begins to form, culminating in the opening of the Rheic Ocean (Domeier,
1509 2016; Nance et al., 2008; Nance and Linnemann, 2008).

1510

1511 To a first order, Ganderia and Carolina, both preserved in North America, record a similar Neoproterozoic
1512 history to Avalonia, differing predominantly in that key metamorphic and magmatic events are ca. 30–
1513 40 Ma younger than in Avalonia (Hibbard et al., 2007; Nance et al., 2008; van Staal et al., 2012).
1514 Depleted mantle model ages from Sm-Nd isotopes also hint at the presence of ca. 1.2–0.8 Ga juvenile
1515 crust in Carolina (Hibbard et al., 2007). However, the principal period of magmatism in both Ganderia
1516 and Carolina is preserved from 650 to 580 Ma and is inferred to have occurred in an ocean-arc
1517 environment (Hibbard et al., 2007). Metamorphism, up to eclogite facies, occurs at the end of this period
1518 and continues into the earliest Cambrian (580–540 Ma) (Barker et al., 1998; Shervais et al., 2003).
1519 Younger magmatism (to ca. 520 Ma, White et al., 2002) linked to the rifting and opening of the Rheic
1520 Ocean, is only preserved in Ganderia (Hibbard et al., 2007). Finally, the Suwannee terrane of Florida is
1521 linked tectonically to both the West African Craton and Amazonia throughout the Neoproterozoic
1522 (Dallmeyer, 1989), lacking the Neoproterozoic arc development preserved in Avalonia, Ganderia and
1523 Carolina. Instead, 550 Ma calc-alkaline volcanic rocks are inferred to represent the remnants of a
1524 continental arc (Heatherington et al., 1996).

1525

1526 We follow the model of Nance et al. (2008) and Murphy et al. (2004) for the Neoproterozoic evolution of
1527 these terranes. An early Tonian (1–0.8 Ga) oceanic arc outboard of Baltica-Ama-zonia-WAC dipping
1528 under the Rodinian plate, formed the earliest portions of crust preserved in these terranes (Fig. 18c–e).
1529 The relative positioning of the terranes follows that of DOM16, with East Avalonia most easterly (Fig.
1530 18b), then West Avalonia, and Ganderia and Carolina furthest west, with Ganderia sitting oceanward of
1531 Carolina. This arrangement follows the same logic outlined in Section 5.2.2, as by maintaining this
1532 relative positioning we avoid having to laterally re-organise the terranes during the Palaeozoic. This is
1533 slightly different from the positioning in Nance et al. (2008), who model West Avalonia more easterly
1534 (relative to a fixed WAC) than East Avalonia and invoke wrench-tectonics to laterally translate the
1535 terranes. Nonetheless, at ca. 750 Ma this subduction polarity reversed and the adjacent ocean basin
1536 between WAC-Baltica began to subduct underneath Avalonia. We model Ganderia and Carolina slightly
1537 further away from the active subduction front—behind Western Avalonia, to account for their lack of
1538 Cryogenian magmatism. This arc front collided with WAC at ca. 650 Ma when the subduction ceased and
1539 reset outboard of the now amalgamated Avalonian-WAC continent as a continental arc (Fig. 18f, g). This
1540 subduction continued until the Cambrian when the area transitioned into a rift environment when the
1541 Rheic Ocean opened.

1542

1543 This last phase our model is preliminary and needs further development, though we hope that it provides a
1544 framework that can assist with testing alternative scenarios. By fitting the latitude of the ca. 520–500 Ma
1545 poles of Baltica (Section 5.2.1), subduction must consume most of the relic ocean basin immediately
1546 north of the Avalonian margin of Gondwana between 550 and 520 Ma. The magmatism in Ganderia
1547 easily accounts for this, however Baltica needs to be further east at 520 Ma, otherwise it must undergo
1548 4000 km of dextral motion (relative to Gondwana) to allow for the initial stages of the Rheic Ocean
1549 opening at ca 500 Ma (e.g. Domeier, 2016; von Raumer and Stampfli, 2008), which we suggest is not a
1550 reasonable scenario. As such, our model places a subduction zone slightly outboard of Avalonia, but
1551 acknowledge that this is a simplification that needs further refinement.

1552

1553 **6 Plate Model**

1554

1555 Having discussed the motions of the evolution of continental configurations in previous sections, here we
1556 discuss the more speculative elements of the reconstruction—the oceanic plates and plate boundaries. For
1557 further in-depth discussion of the continental portions of the model, in particular the major Gondwana
1558 forming sutures and evolution of the post-Cambrian world, we point readers to the studies that produced

1559 the base models used here (MER17, DOM16/18, YOU19, DT14). We also provide in our supplementary
1560 material the associated plate model files (SM2), as well we a tectonic summary of seafloor production and
1561 consumption rates, mid-ocean ridge length and subduction zone length (SM3).

1562

1563 **6.1 Synthetic ocean plates**

1564

1565 The construction of synthetic ocean plates is required to maintain tectonic congruency (Section 2.7) but,
1566 with few exceptions (e.g. ophiolites), there is no direct evidence of the configuration or tectonic
1567 parameters (e.g. spreading rate, asymmetry) of oceanic crust for the pre-Mesozoic due to the constant
1568 subduction of oceanic lithosphere. However, we know that oceanic crust typical of present-day (i.e.
1569 MORB) did exist in the Palaeozoic and Neoproterozoic, as evidenced by ophiolitic remains preserved in
1570 orogens (Furnes et al., 2014). We therefore use one key assumption when constructing oceanic plates: we
1571 assume that the production (rate of motion, orthogonal spreading etc.) and subduction of oceanic crust in
1572 the Neoproterozoic was fundamentally similar to the Cenozoic. We note that this may not be a valid
1573 assumption for the early Neoproterozoic, since abundant ophiolite preservation only occurs after Rodinia
1574 breakup (Stern and Miller, 2018) and pre-1 Ga ophiolites suggest thicker oceanic crust (Moores, 2002)
1575 which, along with secular changes in Earth's heat loss (Brown et al., 2020b) could have an influence on
1576 spreading and subduction dynamics. Nonetheless, we maintain that if available palaeomagnetic and
1577 geological data can be reconciled within a uniformitarian framework of oceanic crust production and
1578 destruction, then our model becomes a useful reference model for future models that explore alternative
1579 hypotheses. Measurements of seafloor production, crustal consumption, ridge length and subduction zone
1580 length are provided in the supplementary material.

1581

1582 *6.1.1 Early Tonian until Rodinia breakup (1000-750 Ma)*

1583

1584 In our model, three prominent ocean basins existed in the early Tonian: the Mirovoi Ocean (McMenamin
1585 et al., 1990; Meert and Lieberman, 2008), the Mawson Sea (Meert, 2003; Meert and Lieberman, 2008)
1586 and the Mozambique Ocean (Fig. 8) (Collins et al., 2003; Collins and Pisarevsky, 2005). These ocean
1587 basins have been defined previously in the same context as they appear in our model, however, given
1588 differences between our model and the original publications, the geographical boundaries of each ocean
1589 are slightly different.

1590

1591 We define the Mirovoi Ocean as the large ocean bordering Rodinia in its west and India-South China and
1592 the Sahara Metacraton in the north east and south east respectively. The Mirovoi is the largest and most

1593 prominent ocean basin for the Neoproterozoic in our model, existing until ca. 520 Ma with the opening of
1594 the Proto-Tethyan Ocean and Ran Sea (Fig. 9, Hartz and Torsvik, 2002). It is (conceptually) equivalent to
1595 the external Panthalassic and Pacific oceans of the Phanerozoic, as it consists almost entirely of oceanic
1596 lithosphere and is ringed by subduction for the majority of its existence. At 1000 Ma, we model a triple
1597 junction spreading ridge located roughly in the centre of the ocean basin. The triple junction provides
1598 three directions of spreading to account for convergence in three areas: (i) the closure of the ocean basin
1599 separating India-South China from Tarim, Qaidam-Qilian, Lut, Afghanistan, Kunlun and Tarim (*this*
1600 *study*); (ii) the Taimyr subduction zone outboard of *northern* Siberia (Metelkin et al., 2012; Vernikovskiy
1601 et al., 2004) and the Valhalla Orogen outboard of Greenland (Cawood et al., 2010), and; (iii) the Proto-
1602 Avalonian-Cadomian subduction zone outboard of Baltica and Amazonia (Murphy et al., 2000) that
1603 extends northwards to Sao Francisco and further north where it is preserved in the Iskel Island Arc of
1604 Hoggar (Liégeois et al., 2003).

1605
1606 The spreading arms of the triple junction span north, east-southeast and southwest. The northern arm
1607 separates Siberia and the India-China accretionary zone and extends partway into the Mawson Sea. The
1608 east-southeastern arm extends towards India, where we connect it via a transform fault to the mid-ocean
1609 ridge in the Mozambique Ocean. The southwestern arm intersects the Proto-Avalonian subduction zone.
1610 We model this configuration (triple junction) as being stable through the early Tonian until ca. 870–850
1611 Ma, where a plate-reorganisation event occurs. We link this organisation to a change in kinematics of
1612 Rodinia, suggested through palaeomagnetic data. At this time, palaeomagnetic data from the Baltica
1613 (Walderhaug et al., 2007) suggest that Rodinia had drifted to southerly latitudes, before returning to
1614 equatorial latitudes by ca. 750 Ma (Eyster et al., 2019). MER17 modelled significant relative dextral
1615 motion between Congo-SF-Azania and Rodinia during this time (870–750 Ma) as well, as suggested by
1616 palaeomagnetic data (Evans et al., 2016). Geological data from Congo-Azania also supports the rotation,
1617 with onset of sedimentation interpreted to be a rift event in the Damara region (Armstrong et al., 2005;
1618 McGee et al., 2012) and the onset of a massive subduction system outboard of Azania (Archibald et al.,
1619 2017; Handke et al., 1999). During this transition (850–800 Ma), we model the Mirovoi Ocean as a single
1620 spreading ridge, orientated sub-parallel to the *north*-facing arm in the original triple junction, extending
1621 northwards toward Siberia and intersecting an oceanic arc outboard of Baltica.

1622
1623 For the Tonian, until 750 Ma, the South China-India continent moved to polar latitudes on the north-
1624 eastern side of the Mirovoi Ocean basin. Comparably, Siberia started dextral motion relative to Rodinia,
1625 from a position near Australia-North China to one near Greenland (Pisarevsky et al., 2013). To account
1626 for this motion, alongside ongoing subduction in the Taimyr region of Siberia, we have extended the

1627 spreading ridge from the Mozambique Ocean into the Mirovoi Ocean (running E-W) and have a northern
1628 arm accounting for divergence between Tarim and the Taimyr subduction zone. This interpretation
1629 necessitates the presence of a triple junction, with a third ridge arm intersecting the subduction zone
1630 outboard of Baltica, similar to the configuration at the start of the Tonian.

1631

1632 The Mawson Sea is defined as the ocean basin between Australia and India-South China that closed with
1633 the amalgamation of Gondwana (Meert, 2003). In the early Tonian, this basin is large in our model,
1634 necessitated by MER17s removal of India-South China from Rodinia. The large size is because relative
1635 longitude prevents India-South China (at this point in time travelling north from the equator) from being
1636 any closer to Australia, as Azania occupied the same latitude as India and lay between India and
1637 Australia. We model a single spreading ridge in the centre of the ocean basin, accounting for the
1638 subduction on the India-South China margin as suggested in the Eastern Ghats of India and accretion of
1639 the Ruker Terrane to Indo-Antarctica (e.g. Corvino et al., 2008; Liu et al., 2017). In South China, this
1640 subduction is more sparse, but recent work has suggested that part of present-day Vietnam is associated
1641 with the southwestern Yangtze craton (Minh et al., 2020) and could record the late Tonian portion of a
1642 subduction system and be the focus of future work. At 930 Ma the spreading direction of the ridge
1643 changes, to compensate for the southerly drift of Rodinia. The change in spreading direction coincides
1644 with the docking of North China-Australia-Antarctica with Laurentia as the final amalgamation event of
1645 Rodinia (Mulder et al., 2018b). At this time we model subduction initiating outboard of Australia (against
1646 Lhasa, (e.g. Guynn et al., 2006) and into the ocean outboard of Antarctica further *south*. Preserved
1647 evidence of subduction here is sparse due to ice cover in Antarctica. However, this area of the ocean (just
1648 outboard of the western margin of the Mawson craton) is positioned at the centre of Antarctica in the
1649 nexus between the TOAST terrane and Mawson craton, so it could be possible some arcs are preserved in
1650 Antarctica. Further south, this subduction zone transitions into a transform boundary and separates
1651 relative motion between Azania-Congo and Rodinia between 930 Ma and 850 Ma. We model continual
1652 spreading in the Mawson Sea until 850 Ma, at which time the ocean basin begins to close and we do not
1653 model an active ridge. By 750 Ma the Mawson Sea is extremely narrow, with only 2500–3500 km of
1654 ocean basin separating Australia and India. A narrow ocean basin is supported geologically by the strong
1655 evidence suggesting that a large sinistral shear zone was present outboard of western Australia and
1656 Antarctica during the Cryogenian and Ediacaran, suggesting that there was close proximity without
1657 collision between Australia-Antarctica and another continent (Collins, 2003; Fitzsimons, 2003; Halpin et
1658 al., 2017; Merdith et al., 2017b; Powell and Pisarevsky, 2002).

1659

1660 The Mozambique Ocean is described as the ocean that closed with the collision of India and Congo along
1661 the East African Orogen reacted to Gondwana amalgamation during the Cryogenian and Ediacaran (e.g.
1662 Collins and Pisarevsky, 2005). For the sake of continuity, we refer to this ocean as the Mozambique in the
1663 Tonian as well. A small ocean (in our model, roughly equivalent in size to the Tasman Sea between
1664 Australia and New Zealand), termed ‘Neomozambique Ocean’ also closed with the formation of the East
1665 African Orogen, however this ocean was located between ANS-Azania-TOAST and Congo (Fig. 8).

1666

1667 Geographically, the location of the Mozambique ocean is difficult to determine in the early Tonian due to
1668 overlapping latitudes between Azania and India (they are separated longitudinally by $\sim 120^\circ$ in our
1669 model). In our model there is no spreading in this ocean basin at this time, because Congo-Azania was
1670 latitudinally stable while India-South China moves towards the North Pole on a different longitude,
1671 accounted for by spreading in the Mirovoi Ocean. The spreading ridge in the Mawson Sea at ca. 900 Ma
1672 extends sufficiently south so that the ocean lithosphere generated here is subducted during closure of the
1673 Mozambique Ocean. We model active spreading in a clearly defined Mozambique Ocean beginning at
1674 850 Ma, by which time India-South China had a similar longitude to Azania-Congo, making it easier to
1675 delineate the geographical extent of the ocean distinct from the Mawson Sea. Between 820 and 750 Ma
1676 the ocean basin closes rapidly, in order to fit palaeomagnetic constraints at 750 Ma that show India at
1677 $\sim 60^\circ\text{N}$ (Gregory et al., 2009; Torsvik et al., 2001) and Congo at 15°S (Meert et al., 1995) (placing Azania
1678 at the equator).

1679

1680 At 800 Ma, we model the birth of the Pacific/Panthalassic Ocean, here defined as the ocean basin
1681 separating Laurentia from North China, Australia and Antarctica, which most likely opened as Rodinia
1682 broke-up (although terrane migration across this ocean basin cannot be ruled out, e.g. Mulder et al.
1683 (2020)). We note here that the pre-Mesozoic Pacific Ocean is already universally referred to as the
1684 Panthalassic Ocean. Consequently, we refer to the ocean separating North China-Australia-Antarctica and
1685 Laurentia between Rodinia and Gondwana times (ca. 800–520 Ma) as the proto-Pacific Ocean, the ocean
1686 surrounding Gondwana as the Panthalassic Ocean (‘Panthalassa’, ca. 520–200 Ma) and the ocean that has
1687 existed from 200 Ma to present-day as the Pacific Ocean, noting that all three of these oceans essentially
1688 refer to the same ocean basin (in a geographical sense) that existed between North China-Australia-
1689 Antarctica and Laurentia. From 800 to 720 Ma we have a single ridge system separating Australia-
1690 Antarctica and Laurentia. This ridge produces a highly angular divergence, with spreading rates faster
1691 towards northern Australia and Canada than in Antarctica, resulting in a wider ocean basin in the north
1692 and narrower in the south. These variable spreading rates are required fit the ca. 750 Ma palaeomagnetic

1693 data, which require Australia to be ‘upright’ (same orientation as present-day) and perpendicular to
1694 Laurentia (Wingate and Giddings, 2000).

1695

1696 *6.1.2 late Tonian–Cambrian (Rodinia breakup–Gondwana Assembly, 720–520 Ma)*

1697

1698 By 720 Ma, continental motions around the Mirovoi Ocean are latitudinally stable with no polar
1699 excursions thus, the simplest way to account for the necessary subduction is with a stable triple junction.

1700 At this time, we show subduction along the western margin of the Mirovoi Ocean, outboard of Siberia in
1701 the Taimyr region, as well further south outboard of the WAC and Baltica where the Avalonian and

1702 Cadomian terranes were coalescing (Murphy et al., 1999; Murphy and Nance, 1989) and along the
1703 *northern* margin of the Sahara Metacraton in the east. This latter subduction zone is speculative since

1704 there are no known rocks of this age in the Sahara Metacraton (Blades et al., submitted). The Mirovoi
1705 ocean basin does not grow appreciably during the Ediacaran due to onset of subduction around its

1706 periphery, however, at 590 Ma Baltica begins moving north, resulting in the consumption of the Mirovoi
1707 Ocean basin and necessitating a change in mid-ocean ridge configuration. At this point we model a single

1708 ridge parallel to the Gondwanan margin, from Amazonia in the west towards India in the northeast. This
1709 ridge also accommodates the final motion of India as it collides with Congo. It is difficult to identify

1710 when the Mirovoi Ocean ceased to exist, but by ca. 490 Ma, expansion in the Proto-Tethyan ocean basin
1711 outboard of northwest Gondwana (e.g. Fig. 12), as well as the easterly drift of Siberia amalgamates the

1712 Siberian subduction zones outboard of North China and Chu-Yili-Tianshan, probably indicating that the
1713 majority of Mirovoi crust produced during the Cryogenian and Ediacaran has been consumed.

1714

1715 An obvious issue with this discussion on the size of the Mirovoi is the longitudinal uncertainty between
1716 the position of Congo and Laurentia during the Ediacaran. This uncertainty is because the distance

1717 between these two cratons dictates the size of the Adamastor Ocean (between Kalahari and Congo),
1718 which grew at expense of the Mirovoi Ocean. Earliest evidence of subduction in the Damara belt exists

1719 from 650 Ma, with final closure occurring at 550 Ma. A conservative convergence rate of 40–60 km/Ma
1720 (roughly equivalent to present-day Pacific convergence rates) would make the ocean basin at 4000–6000

1721 km wide, roughly equivalent to the present-day Atlantic Ocean (this width is similar in our model to that
1722 of the Adamastor Ocean). However, faster divergence during Rodinia breakup would increase the size of

1723 this ocean basin, in turn reducing the size of the Mirovoi Ocean during the Ediacaran, similar to how the
1724 size of the modern Pacific Ocean would become smaller or larger depending on the changing size of the

1725 Atlantic Ocean.

1726

1727 The Mawson Sea remains very small in size during the Cryogenian due to the close proximity between
1728 India and Australia-Antarctica. There is no active ridge, instead a transform fault separates the two
1729 continents. There is little evidence of subduction-related magmatism on either cratonic Australia or India
1730 or this time (e.g. Halpin et al., 2017), suggesting the intervening lithosphere that eventually closed with
1731 Gondwana amalgamation could have also involved a more complex scenario of terrane accretion. The
1732 veracity of that statement is strongly dependent on the configuration and relative positioning of other
1733 terranes that are typically reconstructed to the north-western margin of Australia in the late Palaeozoic,
1734 such as Sibumasu and Indochina (Metcalf, 2011) or other terranes preserved in Antarctica that are
1735 speculated to have rifted off the Indo-Antarctica and accreted to the western margin of the Mawson
1736 Craton (Daczko et al., 2018; Mulder et al., 2019). Neither set of terranes are yet considered explicitly in
1737 our model. By 520 Ma Australia-Antarctica is sutured with India, closing any remnants of the Mawson
1738 Sea.

1739
1740 Both the Mozambique and NeoMozambique oceans close orthogonally from 720 to 550 Ma due to the
1741 continual southward motion of India towards Congo-SM. We do not model an active spreading ridge
1742 during this time, as subduction is only preserved on the African side of the collision (Collins and
1743 Pisarevsky, 2005). The presence of an earlier ridge does, however, suggest that at least two ridges (in our
1744 model they are extinct) were subducted during the East African Orogeny.

1745
1746 The Proto-Pacific Ocean grows predominantly longitudinally during the Cryogenian–Cambrian. In our
1747 model, we show separation between Australia-Antarctica-North China and Laurentia using a single ridge
1748 system that propagates southwards, around *southern* Laurentia to eventually separate the Kalahari Craton
1749 and DML at ca. 700 Ma. We maintain a single ridge system until the Ediacaran, although when Kalahari
1750 begins drifting from Laurentia, we follow Merdith et al. (2017b) in inferring a ridge jump to re-align
1751 spreading between Australia and Laurentia with the incipient ridge separating Kalahari and Laurentia.
1752 This ridge jump also assists with reconciling the necessary motion of Australia to Ediacaran
1753 palaeomagnetic data, which require a $\sim 35\text{--}45^\circ$ counter-clockwise rotation from its present-day orientation
1754 (Schmidt and Williams, 2010). At 580 Ma we model a triple junction in the Proto-Pacific Ocean that
1755 coincides with the equatorial excursion of Baltica and cessation of triple junction spreading in the
1756 Mirovoi Ocean. The arms of this triple junction intersect sub-perpendicular to subduction outboard of
1757 Laurentia, a transform fault outboard of North China and another transform boundary outboard of Baltica.

1758
1759 Between 500 and 410 Ma, we refer to the ocean as the Panthalassic Ocean and for this time interval we
1760 extended the triple junction of YOU19 backwards through DOM16 and DOM18. We found at 500 Ma

1761 when Cuyania rifts off the promontory of Laurentia (Domeier, 2016), that the position of the triple
1762 junction and orientation of the ridges extended backwards from YOU19 was parallel to the direction of
1763 spreading separating Cuyania from Laurentia. While coincidental, given the arbitrary nature of pre-
1764 Mesozoic ocean plates, we find it useful to use the configuration at this time as a transition from the
1765 Proto-Pacific Ocean to the Panthalassic Ocean.

1766

1767 **7 Conclusions**

1768

1769 We present here the first continuous full-plate model from 1 Ga to present-day. The model traces the
1770 kinematic evolution of all cratonic crust and links the Neoproterozoic to the Phanerozoic, encompassing
1771 an entire supercontinent cycle, and enabling quantitative analysis of tectonic features for deep time. We
1772 present the model in a palaeomagnetic reference frame, including a new APWP for Gondwana from 540
1773 to 320 Ma and a GAPWaP from 320 to 0 Ma. For the Neoproterozoic, the model uses a hybrid hierarchy,
1774 where relative plate motions are tied to a key plate, forming distinct nodes. This cluster-approach allows
1775 for the model to be iterated, constructed and modified in the future more easily in light of sparse
1776 palaeomagnetic data, but abundant geological data. Our revised Neoproterozoic model incorporates a late
1777 amalgamation of Rodinia with a novel configuration, in particular through the removal of India, South
1778 China and Tarim from the supercontinent. We incorporate major plate re-organisation events at ca. 850
1779 Ma and again at ca. 750–700 Ma, corresponding to the counter-clockwise rotation of Congo-Sao
1780 Francisco-Sahara Metacraton relative to Rodinia and the initial closing of the Mozambique Ocean and
1781 coeval opening of the Proto-Pacific Ocean, respectively. Our model also includes preliminary
1782 interpretations of the Neoproterozoic history of many regional areas, such as terrane amalgamation
1783 outboard of India-South China, Hoggar and Avalonia, that then link coherently with their more
1784 established Phanerozoic histories. We reiterate that this model is a non-unique solution of global
1785 palaeogeography and tectonics for the Neoproterozoic but we hope it can provide a framework on which
1786 future studies can build upon. To facilitate this, we include in our supplementary material (SM3) the
1787 tectonic parameters of seafloor production and consumption as extracted from the model.

1788

1789 Because our model has continuous plate boundaries, it enables a range of new scientific experiments such
1790 as those seeking to link plate boundary processes to other aspects of the Earth system. This includes
1791 experiments related to the biosphere, hydrosphere and atmosphere investigating events surrounding
1792 oxygenation of Earth's atmosphere, Snowball Earth and animal radiation (e.g. (Gernon et al., 2016;
1793 Godd ris et al., 2017; He et al., 2019; Hoffman et al., 1998; Hoffman and Schrag, 2002; Lenton et al.,
1794 2016; Mills et al., 2019) and those studying the deep Earth (e.g. Heron et al., 2020). There are a number

1795 of limitations of this study, in particular, we do not address TPW in this model. Most previous studies
1796 looked at TPW purely from a palaeomagnetic framework, however the incorporation of geological data in
1797 the form of plate boundaries in this model (and others like it) open up opportunities for to analyse whole-
1798 lithospheric motion from other directions (e.g. Tetley et al., 2019).

1799

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1801

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1816

1817

1818 Table 1

Model	Time (Ma)	Scope	Reference Frame			LLSVPs	Hierarchy Structure
			Palaeomagnetic	Mantle	Details		
SET12	200–0	Global	No	Yes	TPW corrected palaeomagnetic data (200–100 Ma). Hotspot motion (100–0 Ma).	No assumption.	Branching.
DT14	410–250	Global	Yes	Yes	TPW corrected mantle reference from 410–250 Ma.	Assumes stable and long-lived.	Flat.
MUL16	230–200	Global	No	Yes	TPW corrected palaeomagnetic data (200–100 Ma). Hotspot motion (100–0 Ma).	No assumption.	Branching.
MAT16	410–0	Global	Yes	Yes	TPW corrected mantle reference from 410–0 Ma.	Assumes stable and long-lived.	Branching.
<i>DOM16</i>	500–410	Gondwana-Laurentia-Baltica	Yes	Yes	TPW corrected mantle reference from 500–410 Ma.	Assumes stable and long-lived.	Flat.
<i>MER17</i>	1000–520	Global	Yes	No		No assumption.	Nodal.
<i>DOM18</i>	500–410	Gondwana-Siberia-China	Yes	Yes	TPW corrected mantle reference from 500–410 Ma.	Assumes stable and long-lived.	Flat.
<i>YOU19</i>	410–0	Global	Yes	Yes	TPW corrected mantle reference frame from 250–0 Ma.	No assumption.	Branching.
MUL19	250–0	Global	No	Yes	Optimisation method after Tetley et al. (2019).	No assumption.	Branching.
This model	1000–0	Global	Yes	No		No assumption.	Nodal from 1000–410 Ma. Branching from 410–0 Ma.

1819

1820 **Table 2**

Key	Rockunit	OldAge	YoungAge	Glat	Glon	Plat	Plon	A95	Reference
<i>Laurentia</i>									
L1	Gunbarrel Intrusions combined	780	776	45	-110	14.6	127	3.2	Harlan et al. (2008).
L2	Uinta Formation	800	750	41	-110	0.8	161.3	4.7	Weil et al. (2006).
L3	Galeros - Carbon Canyon	764	750	35.15	-111.8	-0.5	166	9.7	Weil et al. (2004); Eyster et al. (2019).
L4	Kwagunt Formation	759	743	36.15	-112	14.2	163.8	3.5	Eyster et al. (2019).
L5	Kwagunt Formation 2	748	736	36.15	-112	18.2	166	7	Weil et al. (2004).
L6	Franklin Dykes	727	712	75	-82	8.4	163.8	2.8	Denyszyn et al. (2009).
L7	Long Range Dykes	617	613	53.5	-57.5	-19	175.3	17.4	Murthy et al. (1992); Hodych et al. (2004); Age: Kamo and Gower (1994).
L8	Skinner Cove Formation	554	548	50	-60	15	157	9	McCausland and Hodych (1998).
L9**	Andres Red Beds	423	393	41	-74	13	105	9	Miller and Kent (1988).
L10**	Wabash Reef	423	415	40.85	-85.7	17	125	5.3	McCabe et al. (1985); Torsvik et al. (1996).
L11**	Rose Hill formation	433	427	39	-79	19.1	128.3	5.8	French and Van der Voo (1979).
L12**	Ringgold Gap Sediments	456	433	34.51	-85.06	24	146.6	7.7	Morrison and Ellwood (1986).
L13**	Tablehead Group Limestone mean	470	456	48.33	-58.43	13.4	149.3	3.9	Hodych (1989); Hall and Evans (1988); Deutsch and Prasad (1987); Torsvik et al. (1996); Torsvik et al. (2012).
L14**	St George Group Limestone	485	456	48.3	-59	17.5	152.4	4.3	Deutsch and Prasad (1987).
L15**	Oneota Dolomite	485	470	43.41	-91.23	10.3	166.5	11.9	Jackson and Van der Voo (1985).

L16**	Moore Hollow sediments	500	490	31	-99	-0.6	163	8.5	Far and Gose (1991).
L17**	Morgan Creek	497	470	30.25	-98.5	10.6	158	9.7	Loucks an Elmore (1986).
L18**	Point Peak	497	485	30.5	-99	5.2	165.8	6	Van der Voo et al. (1976).
L19**	Taum Sauk Limestone	497	485	37.55	-90.31	-3.4	175.1	7.1	Dunn and Elmore (1985).
L20**	Roywe dolomite	497	485	34.25	-97.11	12.6	157.3	4.3	Nick and Elmore (1990).
L21**	Florida Mountains	497	485	32.05	-107.37	-5.4	168.7	10	Geissman et al. (1991).
L22**	Tapeats Sandstone	520	497	36.11	-113.99	2.3	162.6	3.3	Elston and Bressler (1977).
L23**	Mount Rigaud and Chatham-Grenville	534	530	45.28	24.2	-11.9	184.5	6.2	McClausland et al. (2007).
<i>Baltica</i>									
B1	Southern Sweden Dykes	946	935	59	16	-0.9	240.7	6.7	Elming et al. (2014); Pisarevsky and Bylund (1998).
B2	Branton-Algo Anorthosite	927	905	58.5	6.5	5	249	3.9	Stearn and Piper, (1984); Age: Scherstén et al. (2000).
B3	Rogaland Igneous Complex	883	855	58.5	6	-46	238	18.1	Walderhaug et al. (2007).
B4	Hunnedalen Dykes	875	821	59	6.75	-41	222	10	Walderhaug et al. (1999).
B5	Egersund Dykes	619	613	58.5	6	-31.4	224.1	15.6	Walderhaug et al. (2007).
B6	Kurgashlya Formation	570	560	53.3	57.5	-51	135	4.9	Lubnina et al. (2014).
B7	Bakeevo Formation	570	560	54.5	58.2	-42	119	5.3	Lubnina et al. (2014).
B8	Winter coast sediments	558	552	65.5	40	-25.3	132.5	2.8	Popov et al. (2002); Age: Martin et al. (2000).
B9	Zolotitsa sediments I, Russia	560	550	65.5	40	-31.7	112.9	2.4	Popov et al. (2005).

B10	Verkhotina sediments	560	550	65.5	40	-32.2	107.1	2	Popov et al. (2005).
B11	Zolotitsa sediments II	560	550	65.5	40	-28.3	109.9	3.8	Iglesia Llanos et al. (2005).
B12	Zigan Formation	552	544	53.7	56.7	-16	138	3.7	Levashova et al. (2013).
B13	Swedish Limestone (1N)	467	458	58.3	13.9	3	35	13.4	Torsvik & Trench (1991).
B14	Swedish Limestone	480	470	58	13	30	55	9	Torsvik & Trench (1991).
B15	Swedish Limestone (1R)	480	470	59	15	18	46	5.1	Torsvik & Trench (1991).
B16	Narva Limestones	485	470	59	31	18	55	4	Khramov and Iosifidi (2009).
B17	St Petersburg Limestone	480	470	58	30	33	58	3.6	Smethurst et al., (1998).
B18+	Gotland Follingbo Limestone	430	420	57.5	18.5	21	164	6	Claesson (1979).
B19+	Dniestr Silurian Lmst.	428	416	48.6	27	14.3	169.3	7.4	Jelenska et al. (2005).
B20+	Gotland Medby Limest.	427	417	57.5	18.5	23	171	8	Claesson (1979).
B21+	Ringerike Sandst. Norway	426	410	60	10.2	19	164	6.7	Douglass. (1988).
B22+	Tiverskaya Series	419	411	48.6	27	0	149	13.3	Jelenska et al. (2015).
B23+	Ivaniev and Dniestr Sediments	419	393	48.7	26	-1	175	9.6	Lubnina et al (2007).
B24+	Devonian Seds. Podolia	416	406	48.7	26	-3.7	145.5	6.7	Smethurst and Khramov (1992).
B25+	Dniestrovskaya Series	416	407	48.6	27	2.3	158.4	7.4	Jelenska et al. (2005).
B26+	Zilair Sediments, Russia	411	375	54	59	-2	161	3.1	Danukalov et al. (1983).
B27+	Eifelian sedimentary rocks, Russia	398	392	50	5	19	173	2.9	Minibaev and Sulutdinov (1991).

B28+	Bashkirea Sediments, Russia	398	385	54	59	-7	162	4.8	Danukalov et al. (1983).
B29+	Kola Dykes, Russia	390	370	68	33	11	147.6	11.1	Veselovsky and Arzamastsev (2011).
<i>Siberia</i>									
S1	Ust-Kirba Formation	960	930	60	137.2	8.1	2.6	10.4	Popov et al. (2002).
S2	Kitoi Dykes	762	754	52	103	0.4	21.8	6.1	Pisarevsky et al. (2013).
S3	Kesyussa Formation	542	535	71	122.5	-37.6	165	5.2	Pisarevsky et al. (1997).
S4-	Moyero River sediments	459	439	68	104	-14	124	8	Gallet and pavlov (1996).
S5-	Angara River sediments*	460	450	58.5	99.8	-29.5	140.2	6.4	Pavlov et al. (2012).
S6-	Kulumbe section	466	456	68	88.8	-24.1	152.4	3.3	Pavlov et al. (2008).
S7-	Stolobovaya section	466	456	62.1	92.5	-22	158	4	Pavlov et al. (2008).
S8-	Moyero River sediments	468	458	68	104	-23	158	4	Gallet and Pavlov (1996).
S9-	Angara River sediments	473	463	58.5	99.8	-35.2	153.2	3.6	Pavlov et al. (2012).
S10-	Moyero River sediments	474	464	68	104	-30	157	4	Gallet and Pavlov (1996).
S11-	Guragir Formation	475	465	68	88.8	-30.9	152.7	3.2	Pavlov and Gallet(1998).
S12-	Angara River sediments	480	470	58.5	99.8	-36.4	158.2	6.5	Pavlov et al. (2012).
S13-	Moyero River sediments	483	473	67.5	104	-33.9	151.7	2.2	Gallet and Pavlov (1996).
S14-	Uigur and Nizhneiltyk Formations	488	478	68	88.8	-35.2	127.2	4.9	Pavlov and Gallet (1998).

S15-	Moyero River sediments	488	478	68	104	-40	138	9	Gallet and Pavlov (1996).
S16-	Kulumbinskaya Formation	505	495	68	88.8	-36.1	130.7	6	Pavlov and Gallet (1998).
S17-	Moyero River sediments	505	495	68	104	-37	138	9	Gallet and Pavlov (1996).
S18	Yunkyulyabit-Yuryakh Formation	512	502	70.9	122.6	-36.4	139.6	4.6	Pisarevsky et al. (1997).
S19-	Kulumbe River section	512	502	68	88.4	-41.9	135.8	2.3	Pavlov and Gallet (1998).
S20-	Khorbusuonka Amgan and Mayan seds.	512	502	71.5	124	-43.7	140.5	2.6	Gallet et al. (2003).
S21-	Nyuya and Lena River sediments	437	427	60.6	116.3	-17.6	102	3.2	Powerman et al. (2013).
S22-	Nyuya and Lena River sediments	438	428	60.7	116.3	-18.6	101.9	4.6	Shatsillo et al. (2007).
S23-	Lena River sediments	454	444	60.5	116.4	-21	109	17.3	Torsvik et al. (1995).
S24-	Nyuya River sediments	456	446	60.6	116.3	-31.3	129.5	3.6	Powerman et al. (2013).
S25-	Kudrino section	466	456	57.7	107.99	-21.1	143.4	5	Pavlov et al. (2008).
S26-	Krivaya Luka Formation	469	459	59.7	118.1	-25.6	117.9	5.1	Iosifide et al. (1999).
S27-	Krivaya Luka Formation	469	459	59.7	118.1	-28.2	127.1	2.5	Iosifide et al. (1999).
S28-	Krivolutsky Suite	470	460	57.6	107.8	-32.6	137	8.3	Rodionov et al. (2003).
S29-	Lena River sediments	473	463	59.8	118.1	-32	139	3.1	Torsvik et al. (1995).
S30-	Lena River redbeds	475	465	60	114	-25	137	9	Rodionov et al. (1966).
S31-	Surinsk Formation	485	475	58.3	109.61	-42.2	128.1	5.8	Surkis et al. (1999).

S32-	Verkholensk Formation	506	496	58.5	109.8	-37.69	124	4.5	Rodionov et al. (1998).
S33-	Maya River sediments	515	505	60	132	-45.8	115	5	Pavlovi et al. (2008).
<i>West Australia</i>									
WA1	Browne Formation	830	730	-26	126	44.5	141.7	6.8	Pisarevsky et al. (2007).
WA2	Hussar Formation	800	730	-26	126	62.2	85.8	10.3	Pisarevsky et al. (2007).
WA3	Mundine Dykes	758	752	-23	115.8	45.3	135.4	4.1	Wingate and Giddings (2000).
<i>North Australia</i>									
NA1	Johnny's Creek Member	780	660	-24	133	15.8	83	13.5	Swanson-Hysell et al. (2012).
<i>South Australia</i>									
SA1	Angepena Formation	660	640	-32	138	47.1	176.6	5.3	Williams and Schmidt (2015).
SA2	Yaltipena Formation	650	635	-31.5	139	44.2	172.7	8.2	Sohl et al. (1999).
SA3	Elatina Formation, MEAN	645	635	-32	137.5	49.9	164.4	13.5	Embleton and Williams (1986); Schmidt et al. (1991); Schmidt and Williams (1995); Sohl et al. (1999).
SA4	Nuccaleena Formation	635	610	-31	139	32.3	170.8	2.9	Schmidt et al. (2009).
SA5	Brachina Formation	620	590	-32.2	138	46	135.4	3.3	Schmidt and Williams (2010).
SA6	Bunyeroo Formation	590	570	-31.6	138.4	18.1	196.3	8.8	Schmidt and Williams (1996).
SA7	Wonoka Formation	575	555	-31.3	138.6	5.2	210.5	4.9	Schmidt and Williams (2010).
<i>North China</i>									

NC1	Huaibei Sills 890 Ma	913	876	34	117	-52.3	149.3	3.5	Fu et al. (2015).
NC2	Wudaotang and Xinji Fm	541	521	35.6	110.5	18.5	341.9	6.5	Huang et al. (1999).
NC3	Hebei and Shandong Sediments	541	501	36	118	21.2	335.2	12.4	Zhao et al. (1992).
NC4	NE Sino-Korean Massif	541	485	35.6	110.5	26.8	334.5	8.9	Gao et al. (1983) (recalculated by Zhao et al. (1992)).
NC5-	Zhangxia and Xuzhuang Fms	510	496	35.6	110.5	37	326.7	5.5	Huang et al. (1999).
NC6-	Zhaogezhuang area carbonates	490	467	39.7	118.5	32.9	294.6	4.7	Yang et al. (2002).
NC7-	Changshan and Gushan Fms	496	485	35.6	110.5	31.7	329.6	5.4	Huang et al. (1999).
NC8-	Hebei and Shandong Sediments	499	461	36	118	28.8	310.9	12.3	Zhao et al. (1992).
NC9-	Liangjiashan and Lower Majiagou Fm	485	470	35.6	110.5	37.4	324.3	8.5	Huang et al. (1999).
NC10-	Jinghe	470	456	35.6	110.5	31.5	327.7	7	Huang et al. (1999).
NC11	Upper Majiagou Formation	470	456	35.6	110.5	27.9	310.4	9.2	Yang et al. (1996).
NC12-	Tianjinshan and Miboshan Formations	480	464	37.2	105.5	31.8	326.5	9.5	Huang et al. (1999).
<i>South China</i>									
SC1	Yanbian Dykes A	830	818	26.5	101.5	45.1	130.4	19	Niu et al. (2016).
SC2	Xiaofeng Dykes	812	792	31	111	13.5	91	10.9	Li et al. (2004).
SC3*	Yanbian Dykes B	814	798	26.5	101.5	14.1	32.5	20.4	Niu et al. (2016).
SC4	Liantuo Formation	735	705	30.8	111	9.9	160.3	4.6	Jing et al. (2015) and Evans et al.

									(2000), combined.
SC5	Nantuo Formation	641	631	28.5	110	7.5	161.6	5.9	Zhang et al. (2013); Zhang and Piper (1997).
SC6	Doushantuo Formation	614	590	30.8	111	25.9	185.5	6.7	Zhang et al. (2015).
SC7-	Douposi Formation (Wangcang)	510	496	32.1	106.2	-39.5	185.1	8.3	Lixin et al. (1998).
SC8-	Douposi Formation (Guangyuan)	510	496	32.4	106.3	-51.3	166	8.3	Yang et al. (2004).
SC9-	Shiqian Redbeds	440	425	27.5	108	4.9	194.7	5.6	Opdyke et al. (1987).
SC10-	Shiqian Redbeds	440	425	27.5	108	14.9	196.1	5.1	Huang et al. (2000).
SC11-	Pagoda Formation	458	445	32.4	106.3	-45.8	191.3	3.6	Han et al. (2015).
<i>Congo</i>									
C1	Luakela Volcanics A	770	757	-11.5	24.25	-40.2	122	14.1	Wingate et al. (2010).
C2	Mbozi Complex	773	713	-9.2	32.8	-46	145	6.7	Meert et al. (1995).
<i>Sao Francisco</i>									
SF1	Bahia Dykes (N+R)	928	912	-14	-39	7.3	106.4	6.2	Evans et al. (2016).
<i>Tarim</i>									
T1*	Sugetbrak Formation	635	550	40.9	79.4	19.1	149.7	9.3	Zhan et al. (2007).
T2*	Zhamoketi Andesite	621	609	41.5	87.8	-4.9	146.7	3.9	Zhao et al. (2014).
T3	Lower Sugetbrak Formation	640	615	41	79.5	-21.1	87.4	7	Wen et al. (2017).

T4*	Tereeken Cap Carbonate	640	630	41.5	87.8	27.6	140.4	9.9	Zhao et al. (2014).
T5	Qiaoenbrak Formation, Aksu	760	720	40.8	79.5	-6.3	17.5	9.1	Wen et al. (2013).
T6*	Aksu Dykes	819	795	41.15	80.1	19	128	4.5	Chen et al. (2004).
T7	Baiyisi Volcanics	770	717	41.6	86.54	-17.7	14.2	6	Huang et al. (2005).
T8-	Ordovician Limestones	485	470	41.3	83.4	-20.4	180.6	8.5	Fang et al. (1996).
T9-	Aksu-Kalpin-Bachu area sediments	455	445	40.6	78.9	-40.7	183.3	4.8	Sun and Huang (2009).
T10	Red Sandstone	433	427	40.6	79.4	12.8	159.8	7.3	Zhao et al. (2014)(2014) average of three poles from Fang et al. (1996); Li et al. (1990); Fang et al. (1998).
<i>India</i>									
I1	Malani Igneous Suite Grand Mean	770	734	26	72	69.4	75.7	6.5	Meert et al. (2013).
I2	Bhander and Rewa formations	650	530	26	78	47.3	212.7	5.8	McElhinny et al. (1978).
I3*	Jodphur Group	570	520	27	73	1	164	6.7	Davis et al. (2014).
<i>Seychelles</i>									
SE1	Mahe Dykes	753	747	-4.7	55.5	54.8	57.6	12.1	Torsvik et al. (2001).
<i>Rio de la Plata</i>									
RP1	Sierra de las Animas Complex	582	574	-34	-55.3	12.2	78.9	14.9	Rapalini et al. (2015).
RP2	Sierra de los Barrientos Redbeds	600	500	-37.8	-59	15.1	72.6	12.4	Rapalini et al. (2006).

<i>West African Craton</i>									
WAC1	Djebel Boho Volc.	547	526	30.4	-6.7	-27.3	207.1	14.9	Robert et al. (2017).
WAC2	Fajjoud and Tadoughast Volc.	572	551	30.2	-7.8	-21.9	211	15.6	Robert et al. (2017).
WAC3*	Adrar-n-takoucht Volc	577	564	30.4	-7.8	57.6	115.6	15.7	Robert et al. (2017).
<i>East Avalonia</i>									
EAV1#1	Treffgarne volcanics	482	472	52	-5	-56	126	5.5	Trench et al. (1992).
EAV2#	Stapeley volcanics	471	463	52.6	-3	-26.6	216.1	4.9	McCabe and Channell (1990).
EAV3#	Builth igneous and sediments	468	460	52.1	-3.3	-11	198	10	McCabe et al. (1992).
EAV4#	Tramore volcanics	461	449	52.1	-7.4	-11	198	8.5	Deutsch (1980); Trench and Torsvik (1991).
EAV5#	Borrowdale volcanics	448	438	54.4	-3.2	-8.1	186.2	6.9	Millward and Evans (2003); Channell and McCabe (1992).
EAV6#2	Midlands Minor Intrusives	442	432	52.5	-1.5	-52.5	181.5	10.4	Vizan et al. (2003).
EAV7#2	Browgill redbeds	439	433	54.3	-2.5	-13.6	133.9	12.4	Channell et al. (1993).
EAV8#2	Tortworth volcanics	437	431	51.7	-2.5	7	124	4.7	Torsvik et al. (1994).
EAV9#2	Mendips volcanics	435	429	51.2	-2.5	-13	91	8.8	Torsvik et al. (1993).
<i>Ganderia</i>									
GAN1#	Bourinot Group	510	496	46.1	-60.4	-21	160	8.1	Johnson and Van der Vood (1985).

<i>West Avalonia</i>									
WAV1#	Nahant intrusives	490	488	42.4	-70.9	-34	140	3.9	Thompson et al. (2010).
WAV2#	Dunn Point volcanics	463	457	45.8	-62.1	2	130	4.1	Johnson and Van Der Voo (1990); Hamilton and Murphy (2004).
WAV3#	Cape St. Mary sills	443	439	46.8	-54	10	140	9	Hodych and Buchan (1998).
<i>Carolinia</i>									
CAR1# ₂	Cid Formation metasediments	450	445	35	-80.2	29.6	122.1	5.1	Vick et al. (1987).
CAR2# ₂	Uwharrie and Cid Formation metaseds	450	445	35.5	-80	20	80	14.2	Noel et al. (1988).
<i>Cuyania</i>									
CUY1#	Pavon Formation sediments	458	452	-34.6	-68.6	-3.6	166.4	6.6	Rapalini and Cingolani (2004).
<p>* pole not fit by model # from compilation of Domeier (2016) + from compilation of S. Pisarevsky (<i>Pers. Comm.</i>) - from compilation of Domeier (2018) ** from compilation of Torsvik et al. (2012) 1 inclination only used 2 no upper age constraint (fits inclination data) Glat, sample site latitude; Glon, sample site longitude; Plat, pole latitude; Plon, pole longitude; A95, 95% confidence ellipse.</p>									

1821

1822

1823 **Figure and Table Captions**

1824 Table 1

1825 Summary of main reconstruction features of the full-plate reconstructions discussed in text. Italicised
1826 models are the models used in construction of the model presented in this study.

1827

1828 Table 2

1829 Summary of palaeomagnetic data

1830

1831 Figure 1

1832 Schematic comparison of evolution of plate tectonic modelling. (a) ‘Continental drift’/palaeogeographic
1833 type models and (b) full-plate models; (i)–(iv) identify separate plates. Palaeomagnetic data are the
1834 primary constraint of the movement of continents in both (a) and (b) however, the inclusion of geological
1835 data into the model in (b) preserves the relative type of motion between two continents (divergence,
1836 convergence or transform) and allows for the construction of plate boundaries.

1837

1838 Figure 2

1839 Overview of the rotational framework of a relative plate motion model. (a) Schematic of the ‘tree’ like
1840 hierarchy (e.g. Ross and Scotese, 1988) from the Matthews et al. (2016) reconstruction, where the motion
1841 of all plates are defined relative to the African plate at 50 Ma. (b) A simplified map view of the hierarchy
1842 in (a). The African stage pole (blue hexagon) is at the top of the hierarchy and connected to the absolute
1843 plate motion model that describes the movement of the African plate (in this case, as a proxy for the entire
1844 globe) to the deep earth. (c) Example of plate motion with a continent (Africa) fixed in its present-day
1845 position. Fixed plate presentations enable observations of relative motions between the fixed continent or
1846 plate and another. Hence, they are useful to constrain relative plate motions and also compare two or
1847 more different plate models. In this instance, a key difference between the model of Young et al. (2019)
1848 and Matthews et al. (2016) is highlighted in the spatial relationship between Laurussia and Gondwana at
1849 410 Ma. In Domeier and Torsvik (2014) (which is preserved in Matthews et al. (2016)) Laurussia is
1850 proximal to southwest Gondwana while in Young et al. (2019) it is positioned to the northwest. ADR,
1851 Adria; AFR, Africa; ARB, Arabia; AUS, Australia; BAL, Baltica (cratonic Europe); BM, Burma; CAR,
1852 Caribbean; E.QT, East Qiangtang; EUR, Europe; GRN, Greenland; IC, Indochina; IND, India; JAP,
1853 Japan; KOH, Kohistan arc; LHA, Lhasa; LHR, Lord Howe Rise; MAD, Madagascar; MBL, Marie Byrd
1854 Land; MEX, Mexico; NAM, North America; NC, North China; NEA, Northeast Africa; NWA, Northwest
1855 Africa; NZ, New Zealand; PAC, Pacific; PHL, Philippines; PNG, Papua New Guinea; SAL, Salado
1856 Microplate; SAM, South America; SC, South China; SEA, Southeast Asia; SIB, Siberia; SOM, Somalia;
1857 SPN, Spain; TAU, Taurides; W.QT, West Qiangtang.

1858

1859 Figure 3

1860 Overview of the rotational framework of an absolute plate motion model. (a) Schematic of a flat hierarchy
1861 where the motion of all plates is defined relative to Africa. (b) Map view of the schematic at 450 Ma from
1862 Domeier (2018, 2016). Many different plates (blue hexagons) sit at the top of the hierarchy and connect
1863 to the deep earth. AFR, Africa; ANT, Antarctica; AUS, Australia; BAL, Baltica; C, Carolina; CY, ; EA,
1864 East Avalonia; G, Ganderia; IC, Indochina; IND, India; K-T, Kazakh-Tianshan; LHA, Lhasa; MAD,
1865 Madagascar; NAM, North America; NC, North China; NEA, Northeast Africa; NWA, Northwest Africa;
1866 SAM, South America; SC, South China; SIB, Siberia; SOM, Somalia; T, Tarim; WA, West Avalonia.

1867

1868 Figure 4

1869 ‘Family-tree’ of full-plate models and some of the key underlying continental drift type models leading to
1870 their development. We stress that more recent models are not necessarily better nor replace older models,
1871 rather typically present some alternative interpretation of key data or are based on different assumptions.

1872 See text for more details. The model presented here bridges the MER17, DOM16, DOM18 and YOU19
1873 models into a single continuous plate model. X-axis not to scale. Cr, Cryogenian, and; Ed, Ediacaran.

1874

1875 Figure 5

1876 Overview of the rotational framework of a hybrid plate motion hierarchy. (a) In this case, two different
1877 nodes or clusters sit at the top of their respective branches of the hierarchy and ground regional relative
1878 plate motions. (b) Map view of the schematic at 900 Ma from this study. Two different plates (India and
1879 Laurentia; blue hexagons) sit at the top of the hierarchy and connect to the deep earth, with regional plate
1880 motions being defined relative to each.

1881

1882 *Figure 6*

1883 Summary of the fit of palaeomagnetic data to our model for 1000 to 400 Ma (omitting poles used to
1884 construct the Gondwana APWP). Misfit is the minimum great circle distance (in degrees) within the valid
1885 time range between the palaeomagnetic pole and the geographic north pole. The error bar on each point is
1886 the pole 95% confidence limit ($A95$). Solid line is the mean misfit of all poles, with the dashed lines
1887 representing the standard deviation of the mean. Poles marked as ‘not used’ or ‘inclination only’ in Table
1888 2 are not included in this figure. (a) Poles from the constituent cratons of western Gondwana and the
1889 Avalonian terranes; (b) poles from the constituent cratons of eastern Gondwana and present-day Asia,
1890 and; (c) poles from Laurussia and Siberia.

1891

1892 Figure 7

1893 Comparison of palaeolatitude of major cratons (a) and Chinese cratons (b) from 1000–0 Ma between
1894 three models; (i) This model, M2020; (ii) DOM16/18 (500–410 Ma) and MAT16 (410–0 Ma), and; (iii)
1895 Li et al. (2013, 2008) (1000–550 Ma). The notable excursion at 800 Ma in the Li et al. (2013, 2008)
1896 model is due to their adoption of IITPW at this time (Li et al., 2004).

1897

1898 Figure 8

1899 Distribution of continental crust, ocean basins and plate boundaries in our plate model at 0 Ma, based on
1900 MUL16. Tan polygons are areas of continental lithosphere in the Neoproterozoic that we model, blue
1901 polygons are areas of present-day continental lithosphere. Abbreviations as Figures 8 and 9. A,
1902 Amazonia; Ant, Antarctica; AUS, Australia; BP, Borborema Province; CCT, Chinese Central Tianshan;
1903 Cu, Cuyania; CY, Chu Yili; K, Kalahari; KMT, Krygyz Middle Tianshan; KNT, Krygyz North Tianshan;
1904 KST, Krygyz South Tianshan; N-B, Nigeria-Benin; NC, North China; P, Paranapanema; ANS, Arabian
1905 Nubian Shield; Qa, Qaidam; Qi, Qilian; RDLP, Rio de la Plata; SC, South China; SM, Sahara
1906 Metacraton; T, Tarim; WAC, West African Craton.

1907

1908 Figure 9

1909 Distribution of continental crust, ocean basins and plate boundaries in this plate model at 500 Ma, based
1910 on DOM16 and DOM18 (e.g. Figs. 12 and 13). Tan polygons are areas of continental lithosphere in the
1911 Neoproterozoic that we model, blue polygons are areas of present-day continental lithosphere that are
1912 inferred to exist during the Neoproterozoic, but without having firm geological evidence or that have been
1913 effected by subsequent deformation (e.g. the distance between a present-day coastline and COB). Green
1914 polygons represent a schematic interpretation of congruent continental lithosphere, with intervening crust
1915 being subsequently deformed during future tectonic cycles. Abbreviations as Figure 8, in addition to; A-
1916 C, Avalonia-Cadomia; Af, Afghanistan; B, Baltica; C, Congo; CMT, Central Mongolian Terrane; H,
1917 Hoggar; I, India; S, Siberia.

1918

1919 Figure 10

1920 Distribution of continental crust, ocean basins and plate boundaries in our plate model at 1 Ga. Tan
1921 polygons are areas of continental lithosphere in the Neoproterozoic that we model, blue polygons are
1922 areas of present-day continental lithosphere that are inferred to exist during the Neoproterozoic, but

1923 without having firm geological evidence or that have been affected by subsequent deformation (e.g. the
1924 distance between a present-day coastline and COB). Green polygons represent a schematic interpretation
1925 of congruent continental lithosphere, with intervening crust being subsequently deformed during future
1926 tectonic cycles. A-Ant, Austral-Antarctica; Az, Azania; BP, Borborema Province; CA, Central Angara
1927 Terrane; CLB, Coats Land Block; Dabo. Arc, Dabolava Arc; DML, Dronning Maud Land; eSTR, Eastern
1928 South Tasman Rise; IOGU, In Ouzal Granulite Unit; IGU, Iforas Granulite Unit; KMT(I), Krygyz Middle
1929 Tianshan (Issedonian); KMT(UM), Krygyz Middle Tianshan (Ulutau-Moyunkum); KNT(I), Krygyz
1930 North Tianshan (Issedonian); Mad, Madagascar; NAC, North Australian Craton; NeoMoz. Ocean,
1931 NeoMozambique Ocean; NC, North China; p-ANS, proto-Arabian Nubian Shield; PH, Panxi-Hannan
1932 Belt; Ra, Rayner Province; Ru, Ruker Block; SAC, South Australian Craton; SB, Southern Borborema;
1933 SL, Sri Lanka; Tas, Tasmania; wSTR, Western South Tasman Rise.

1934

1935 *Figure 11*

1936 Comparison of our model (a, b) with Domeier (2018, 2016) (c, d) at 411–410 Ma to highlight the
1937 latitudinal changes implemented for Laurussia in our model. Dark outline of Laurussia and Lhasa
1938 (italicised labels) in (a) are their original positions in YOU19. Palaeomagnetic data are shown at time of
1939 best fit for both models (i.e. map projection view of Fig. 6). Tan polygons are areas of continental
1940 lithosphere in the Neoproterozoic that we model, blue polygons are areas of present-day continental
1941 lithosphere that are inferred to exist during the Neoproterozoic, but without having firm geological
1942 evidence or that have been affected by subsequent deformation (e.g. the distance between a present-day
1943 coastline and COB). Green polygons represent a schematic interpretation of congruent continental
1944 lithosphere, with intervening crust being subsequently deformed during future tectonic cycles. The
1945 DOM16/18 models extend until 410.1 Ma (so we display it at 411 Ma), while YOU19 begins at 410 Ma.
1946 NC, North China; SC, South China; T-K-Qi-Qa, Tarim-Kunlun-Qilian-Qaidam.

1947

1948 *Figure 12*

1949 Comparison of the evolution of the Rheic Ocean between our model (a, 500 Ma; c, 440 Ma) and DOM16
1950 (b, 500 Ma; d, 440 Ma) during the early Palaeozoic. We point out the rotational difference (but similar
1951 latitude) in Baltica at 500 Ma in our model relative to DOM16, discussed in text. Palaeolongitude is
1952 constrained absolutely in DOM16 but not in our model. Subduction polarities are not provided in the
1953 GPlates files of DOM16. Tan polygons are areas of continental lithosphere in the Neoproterozoic that we
1954 model, blue polygons are areas of present-day continental lithosphere that are inferred to exist during the
1955 Neoproterozoic, but without having firm geological evidence or that have been affected by subsequent
1956 deformation (e.g. the distance between a present-day coastline and COB). Green polygons represent a
1957 schematic interpretation of congruent continental lithosphere, with intervening crust being subsequently
1958 deformed during future tectonic cycles. IO, Iapetus Ocean.

1959

1960 *Figure 13*

1961 Comparison of our model (a, c) with DOM18 (b, d) to highlight changes made. Annamia (Indochina,
1962 Sibumasu) is not explicitly modelled in our reconstruction. L, Lut (Iran); F, Farah (Afghanistan); T,
1963 Tarim. Tan polygons are areas of continental lithosphere in the Neoproterozoic that we model, blue
1964 polygons are areas of present-day continental lithosphere that are inferred to exist during the
1965 Neoproterozoic, but without having firm geological evidence or that have been affected by subsequent
1966 deformation (e.g. the distance between a present-day coastline and COB). Green polygons represent a
1967 schematic interpretation of congruent continental lithosphere, with intervening crust being subsequently
1968 deformed during future tectonic cycles. Al, Alxa; An, Annamia; A-S, Altai-Sinai; SC, South China; K,
1969 Kunlun; KMT, Krygyz Middle Tianshan; KNT, Krygyz North Tianshan; NC, North China; Qa-Qi,
1970 Qaidam-Qilian; Ql, Qinling; T, Tarim.

1971

1972 *Figure 14*

1973 Snapshots of plate reconstructions showing our updated model for Australia-Laurentia at key time
1974 intervals, along with palaeomagnetic data. (a) 1000 Ma; (b) 975 Ma; (c) 950 Ma; (e) 925 Ma; (f) 900 Ma;
1975 (h) 580 Ma and (i) 560 Ma. The Tonian model (a–f) follows arguments laid out in Mulder et al. (2018b),
1976 while the Ediacaran-Cambrian evolution is after Mulder et al. (2020). Times in these panels reflex the
1977 nominal time of best fit for each pole. Tan polygons are areas of continental lithosphere in the
1978 Neoproterozoic that we model, blue polygons are areas of present-day continental lithosphere that are
1979 inferred to exist during the Neoproterozoic, but without having firm geological evidence or that have been
1980 affected by subsequent deformation (e.g. the distance between a present-day coastline and COB). Green
1981 polygons represent a schematic interpretation of congruent continental lithosphere, with intervening crust
1982 being subsequently deformed during future tectonic cycles. CLB: Coats Land Block; DML, Dronning
1983 Maud Land; eSTR: Eastern South Tasman Rise; MO, Mirovoi Ocean; MS, Mawson Sea; Tas: Tasmania;
1984 wSTR: Western South Tasman Rise; RDLP: Río de la Plata; WAC, West African Craton.

1985

1986 *Figure 15*

1987 (a) Regional map at present day of Central Asian Orogenic Belt with modelled terranes highlighted in tan.
1988 Light tan terranes are not explicitly modelled but are referred to in the main text. Black areas represent the
1989 cratonic components. For the Kazakh area, we use polygons consistent with their Palaeozoic structure.
1990 The size, orientation and distribution of crust in these terranes in the Neoproterozoic is unknown due to
1991 the subsequent reworking of the terranes. Therefore, the precise position (along a margin), orientation,
1992 size and shape of these terrane polygons in the reconstruction figures are speculative and should be
1993 treated cautiously. (b–f) Evolution of the India-South China system during the Tonian–Cryogenian at key
1994 times. Tan polygons are areas of continental lithosphere in the Neoproterozoic that we model, blue
1995 polygons are areas of present-day continental lithosphere that are inferred to exist during the
1996 Neoproterozoic, but without having firm geological evidence or that have been affected by subsequent
1997 deformation (e.g. the distance between a present-day coastline and COB). Green polygons represent a
1998 schematic interpretation of congruent continental lithosphere, with intervening crust being subsequently
1999 deformed during future tectonic cycles. ADO = Aravalli-Delhi Orogen; Af, Afghanistan; Aus-Ant,
2000 Australia-Antarctica; Ca, Cathaysia; DML-CLN, Dronning Maud Land-Coats Land Block; I, India; I-A,
2001 Indo-Antarctica; KNT (I/UM), Krygyz North Tianshan (Issendonian/Ulutau-Moyunkum); KMT (I/UM),
2002 Krygyz Middle Tianshan (Issendonian/Ulutau-Moyunkum); MIP, Malani Igneous Province; MO, Mirovoi
2003 Ocean; MzO, Mozambique Ocean; NC, North China; NeoMzO, Neomozambique Ocean; Qa, Qaidam;
2004 Qi, Qilian; Ru, Ruker; SC, South China; SM, Sahara Metacraton; T, Tarim; WAC, West African Craton;
2005 Y, Yangtze.

2006

2007 *Figure 16*

2008 Schematic of our model for India-South China at key time steps, showing accretion of the Yangtze Craton
2009 and numerous smaller terranes and blocks to a large, Tonian subduction zone outboard of the north-
2010 western India and the northern margin of South China and their subsequent fragmentation and rifting off
2011 during the Ediacaran–Early Cambrian. (a) early Tonian; (b) late Tonian; (c) Cryogenian–Ediacaran, note
2012 if the model were adopted to fit the cluster of three palaeomagnetic poles from Tarim then collision would
2013 be at ca. 650 Ma, and; (d) Ediacaran–Cambrian. Af, Afghanistan; Al-Qi-Qa, Alxa, Qilian, Qaidam; ANS,
2014 Arabian Nubian Shield; CY, Chu Yili; DML-CLB, Dronning Maud Land-Coats Land Block; EAO, East
2015 African Orogen; I, India; I-A, Indo-Antarctica; KMT (I/UM), Krygyz Middle Tianshan
2016 (Issendonian/Ulutau-Moyunkum); KNT (I/UM), Krygyz North Tianshan (Issendonian/Ulutau-
2017 Moyunkum); SC, South China; T, Tarim.

2018

2019 *Figure 17*

2020 Amalgamation of the Arabian-Nubian Shield (ANS) after Blades et al. (2019b, 2015), Armistead et al.
2021 (2019) and Collins et al. (in revision) and the incorporation of the ANS, Azania and TOAST into
2022 Gondwana along the East African Orogen at: (a) 800 Ma and (b) 520 Ma. Tan polygons are areas of

2023 continental lithosphere in the Neoproterozoic that we model, blue polygons are areas of present-day
2024 continental lithosphere that are inferred to exist during the Neoproterozoic, but without having firm
2025 geological evidence or that have been affected by subsequent deformation (e.g. the distance between a
2026 present-day coastline and COB). Green polygons represent a schematic interpretation of congruent
2027 continental lithosphere, with intervening crust being subsequently deformed during future tectonic cycles.
2028 Thick black lines follow the suture of the East African Orogen and Kuunga Orogen. ANS, Arabian-
2029 Nubian Shield; EAO, East African Orogen; SF, Sao Francisco; SM, Sahara Metacraton.

2030

2031 *Figure 18*

2032 Amalgamation of Hoggar after Caby et al. (1989), Black et al. (1994) and Liégeois et al. (1994),
2033 Borborema and Peri-Gondwanan terranes after Nance et al. (2008) at key time slices. (a) map of key
2034 terranes in a reconstructed Gondwana (dark blue terranes are oceanic terranes); (b) map of reconstructed
2035 Peri-Gondwanan terranes at 650 Ma; (c) 1000 Ma; (d) 900 Ma; (e) 860 Ma; (f) 700 Ma; (g) 640 Ma; (h)
2036 580 Ma; (i) 550 Ma and (j) 520 Ma. Tan polygons are areas of continental lithosphere in the
2037 Neoproterozoic that we model, blue polygons are areas of present-day continental lithosphere that are
2038 inferred to exist during the Neoproterozoic, but without having firm geological evidence or that have been
2039 affected by subsequent deformation (e.g. the distance between a present-day coastline and COB). Green
2040 polygons represent a schematic interpretation of congruent continental lithosphere, with intervening crust
2041 being subsequently deformed during future tectonic cycles A, Aouzegueur; A-I, Assodé-Issalane; Am,
2042 Amazonia; ANS, Arabian-Nubian Shield; B, Baltica; BP, Borborema Province; C, Congo; C-B, Central
2043 Borborema; IGU, Iforas granulite unit; IOGU, In Ouzzal granulite unit; IO, Iapetus Ocean; LATEA,
2044 Laouni, Azrou-n-Fad, Tefedest and Egéré-Aleksod terranes; MO, Mirovoi Ocean; Moz. O, Mozambique
2045 ocean; N-B, Niger-Benin Block; P, Paranapanema; PG, Peri-Gondwanan terranes; S-B, Southern
2046 Borborema; SF, Sao Francisco; SM, Sahara Metacraton; WAC, West African Craton.

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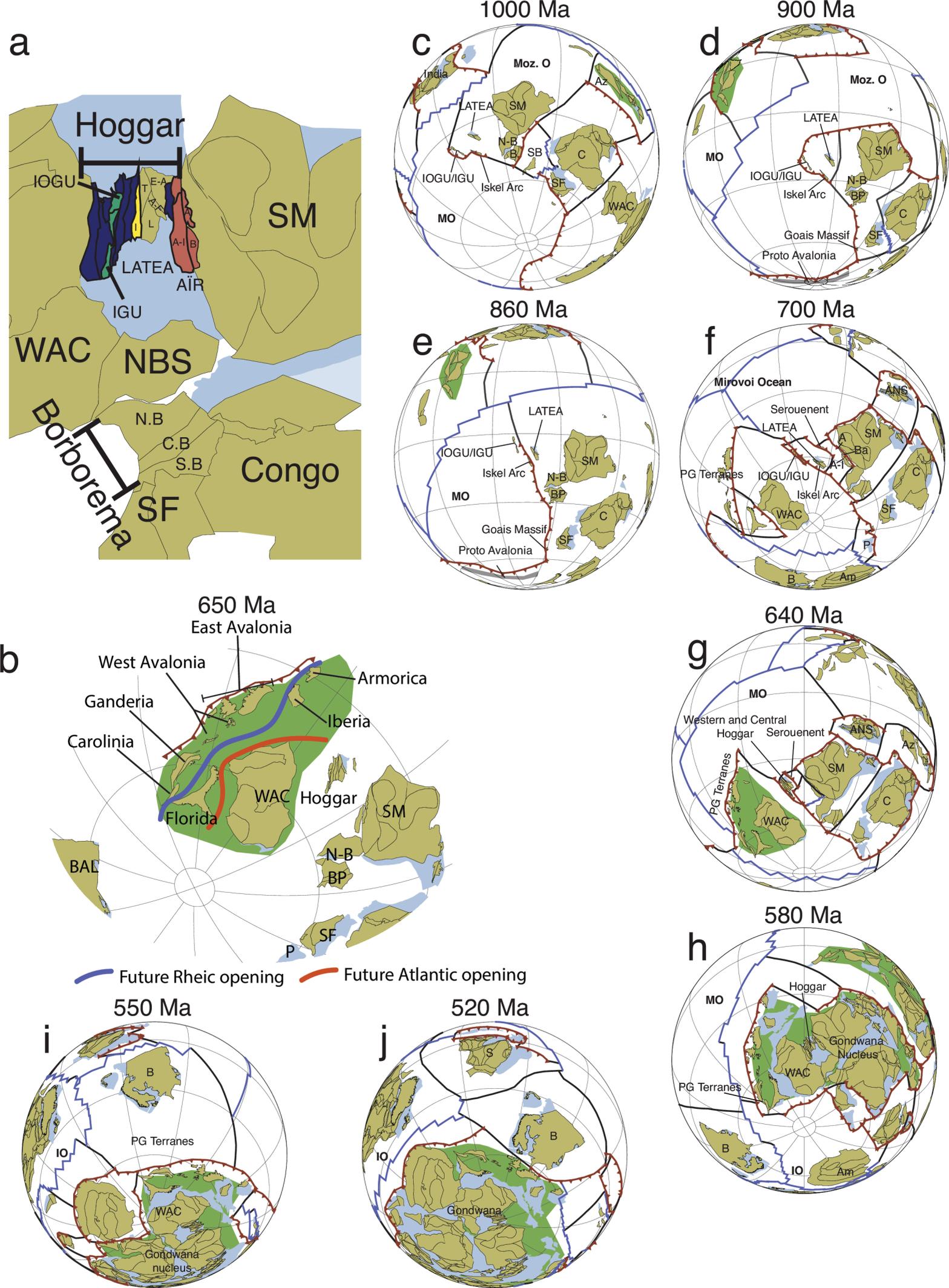
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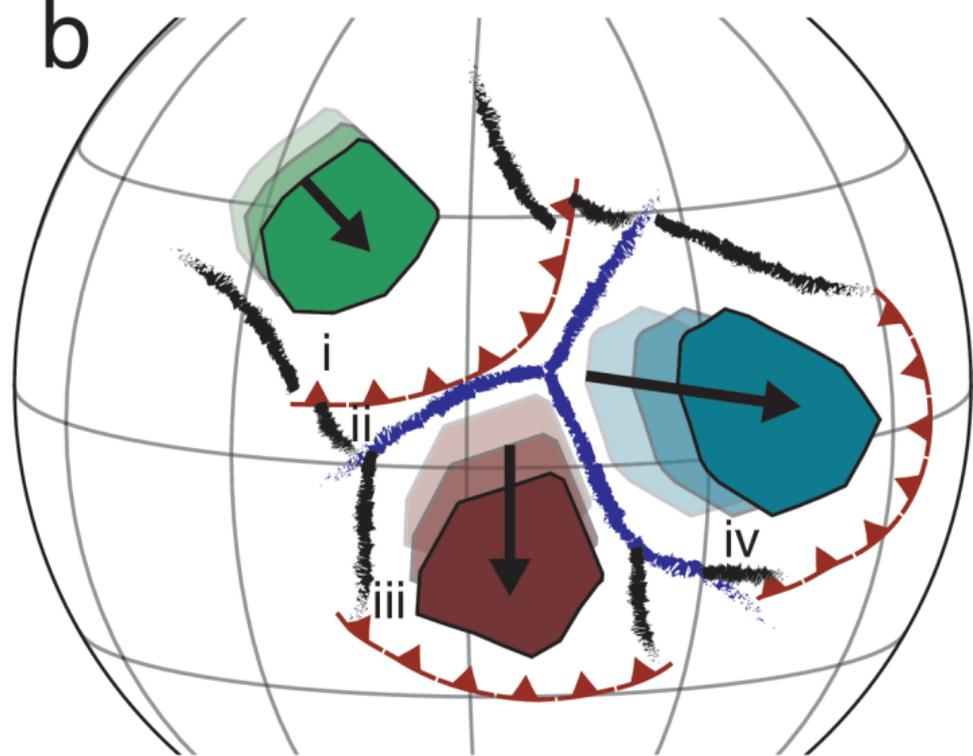
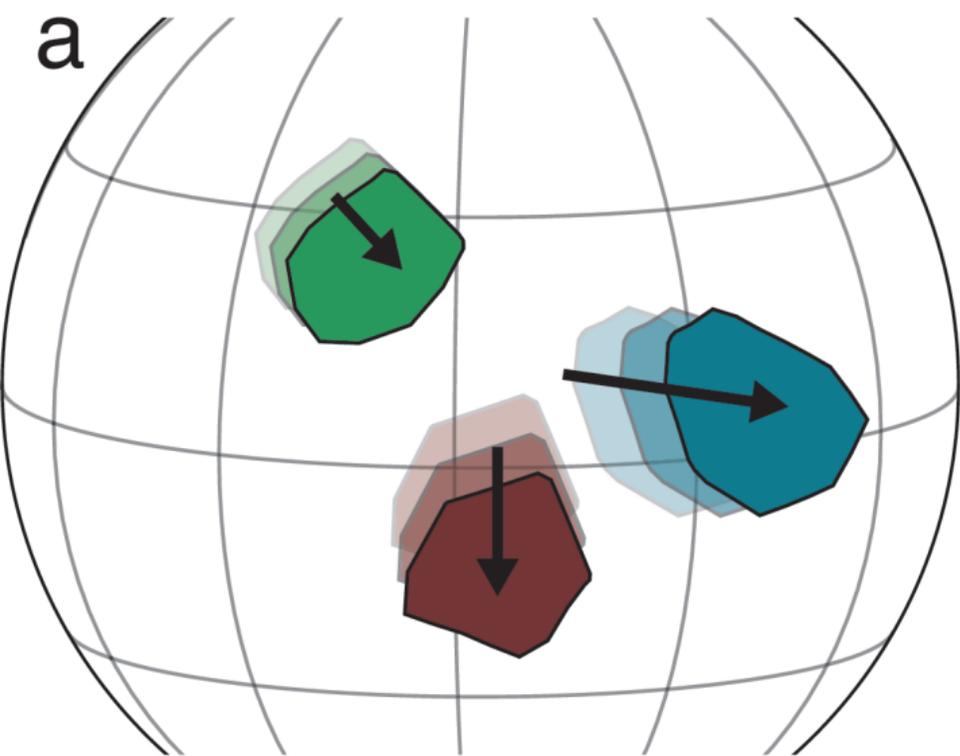
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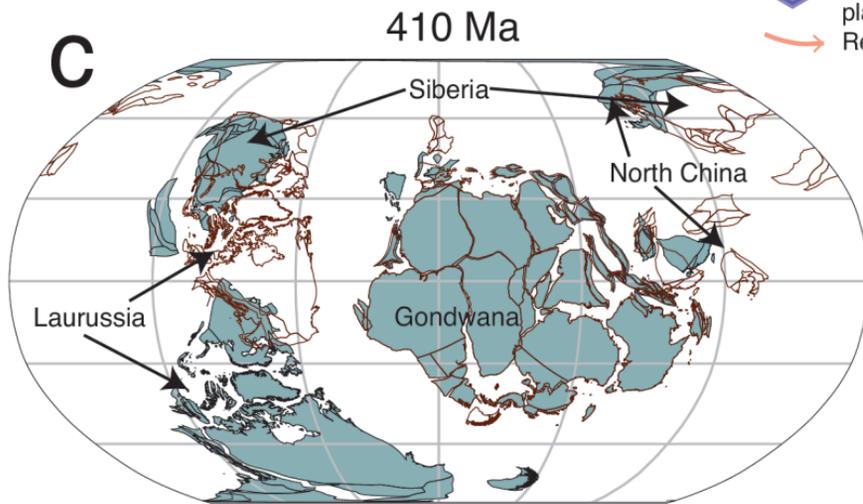
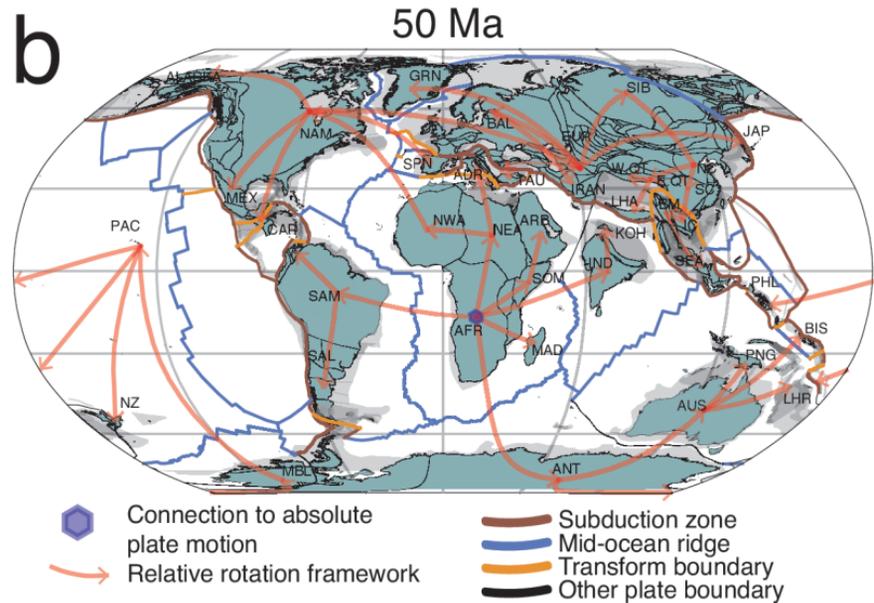
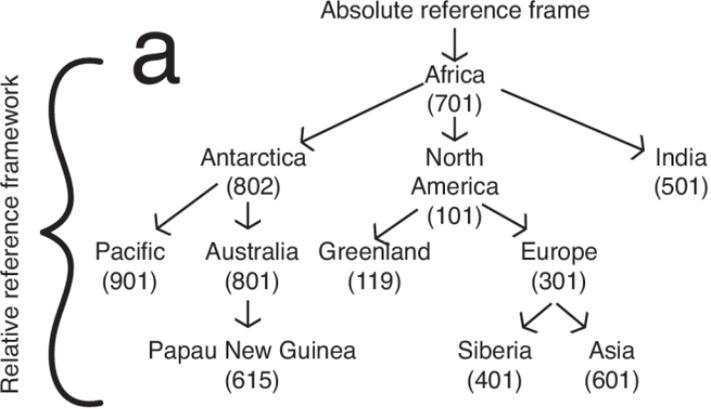
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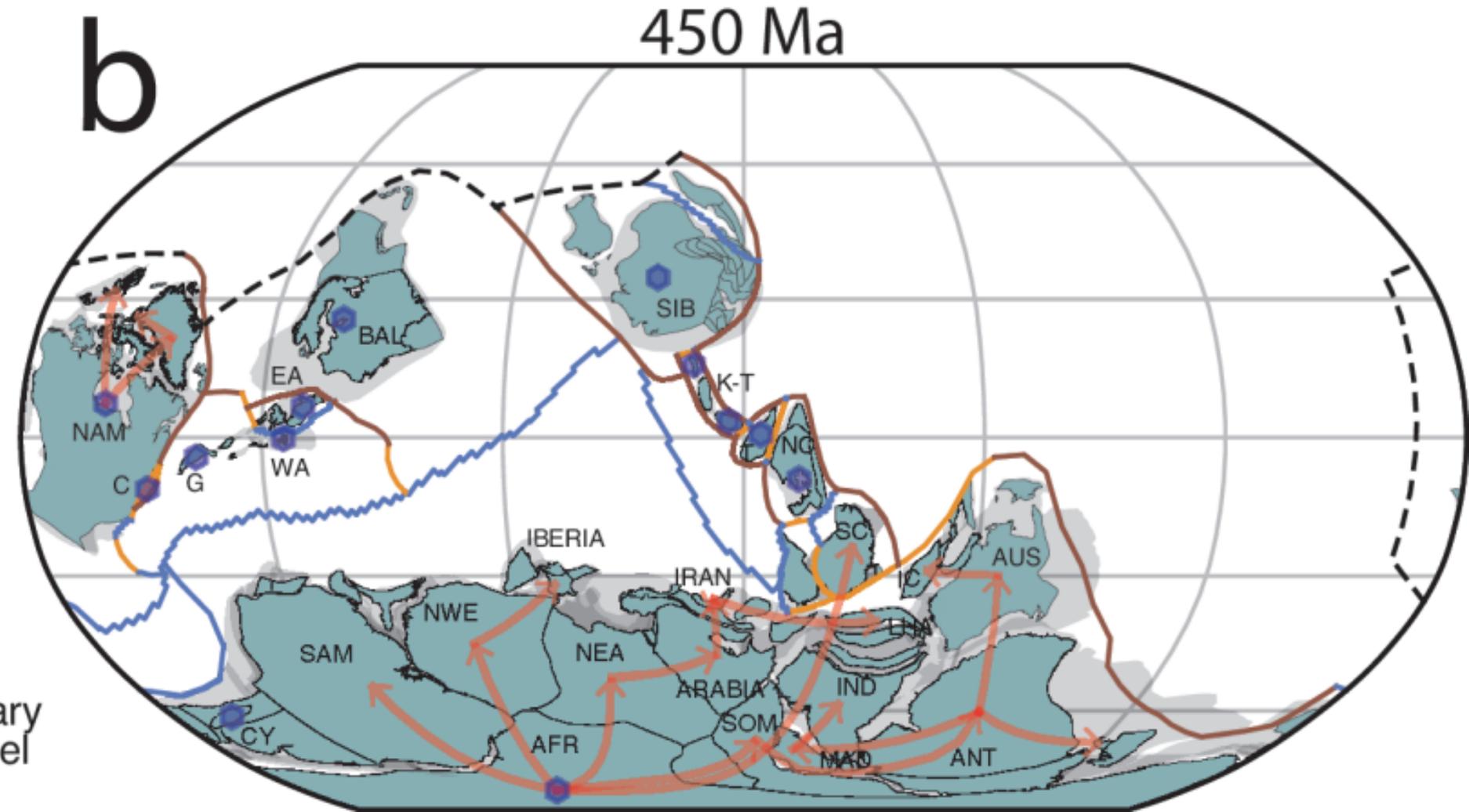
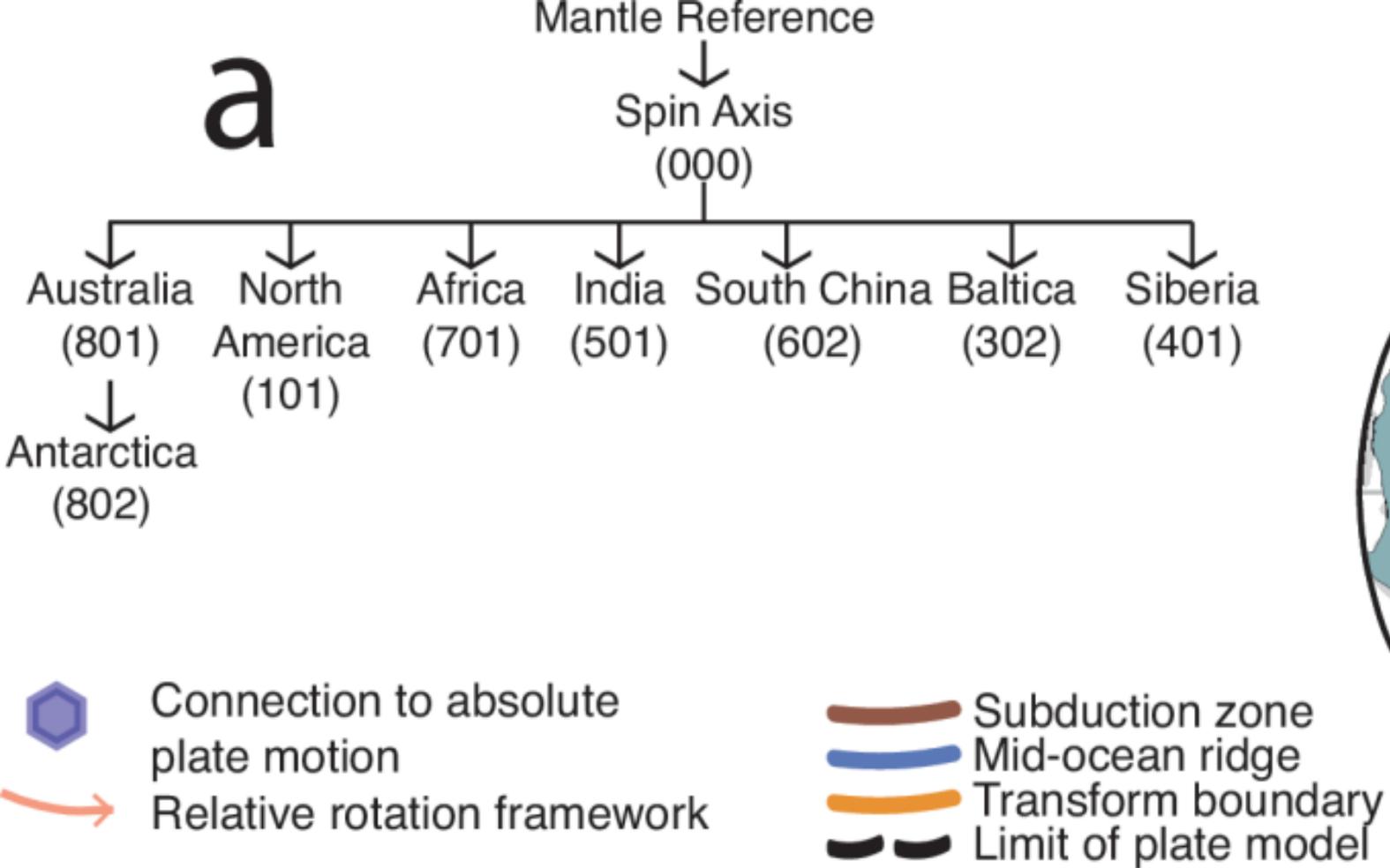


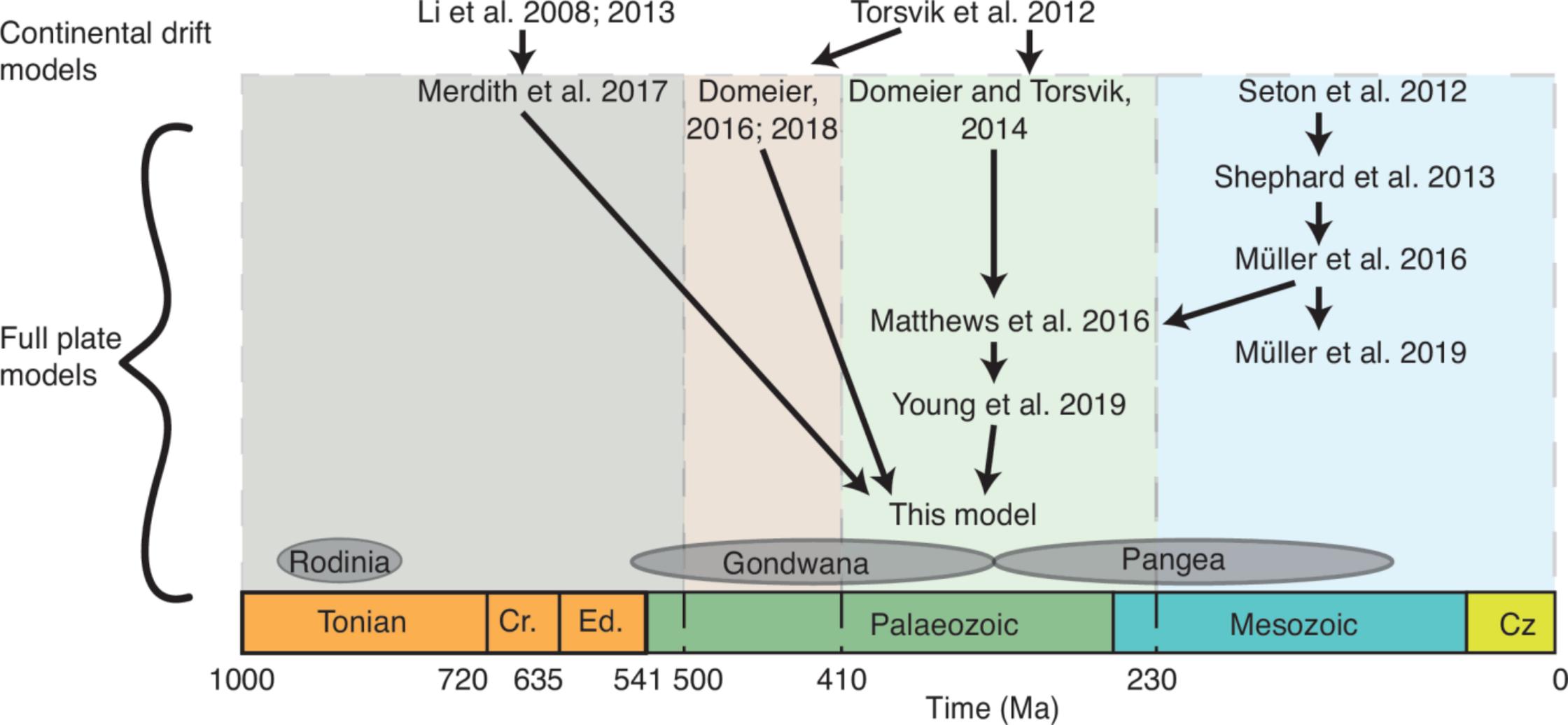
 Transform boundary
 Mid-ocean ridge
 Subduction zone

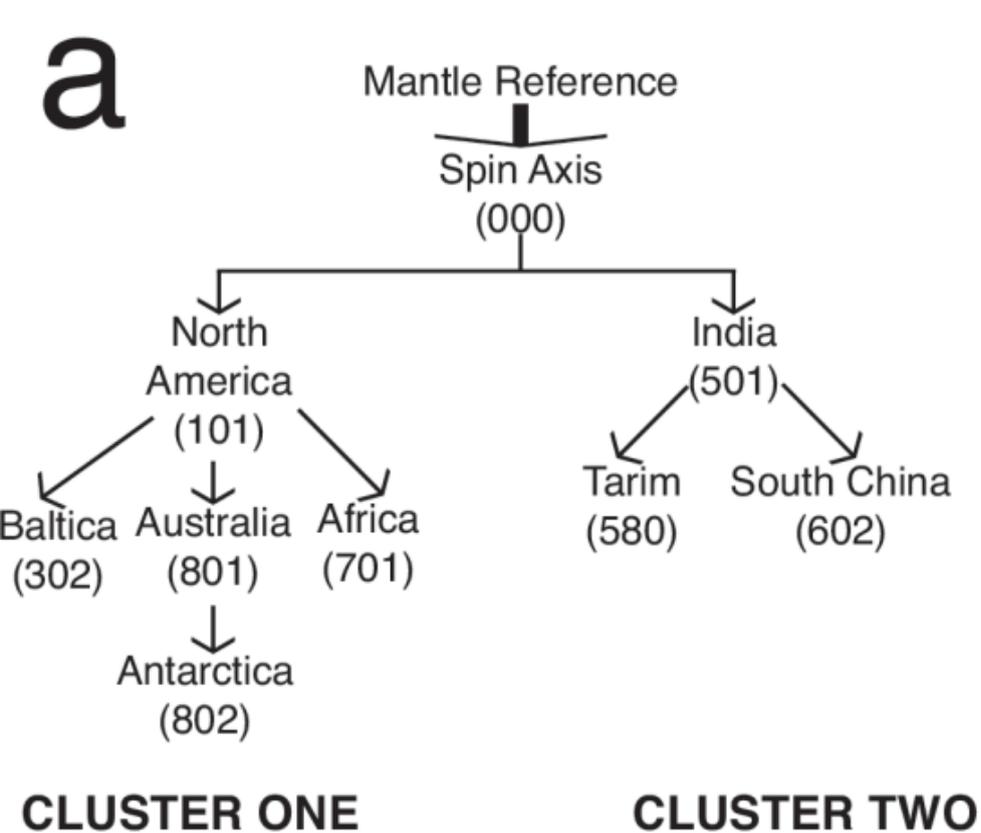
 Direction of movement
 Old
 Young

Young et al. (2019)
 Matthews et al. (2016)

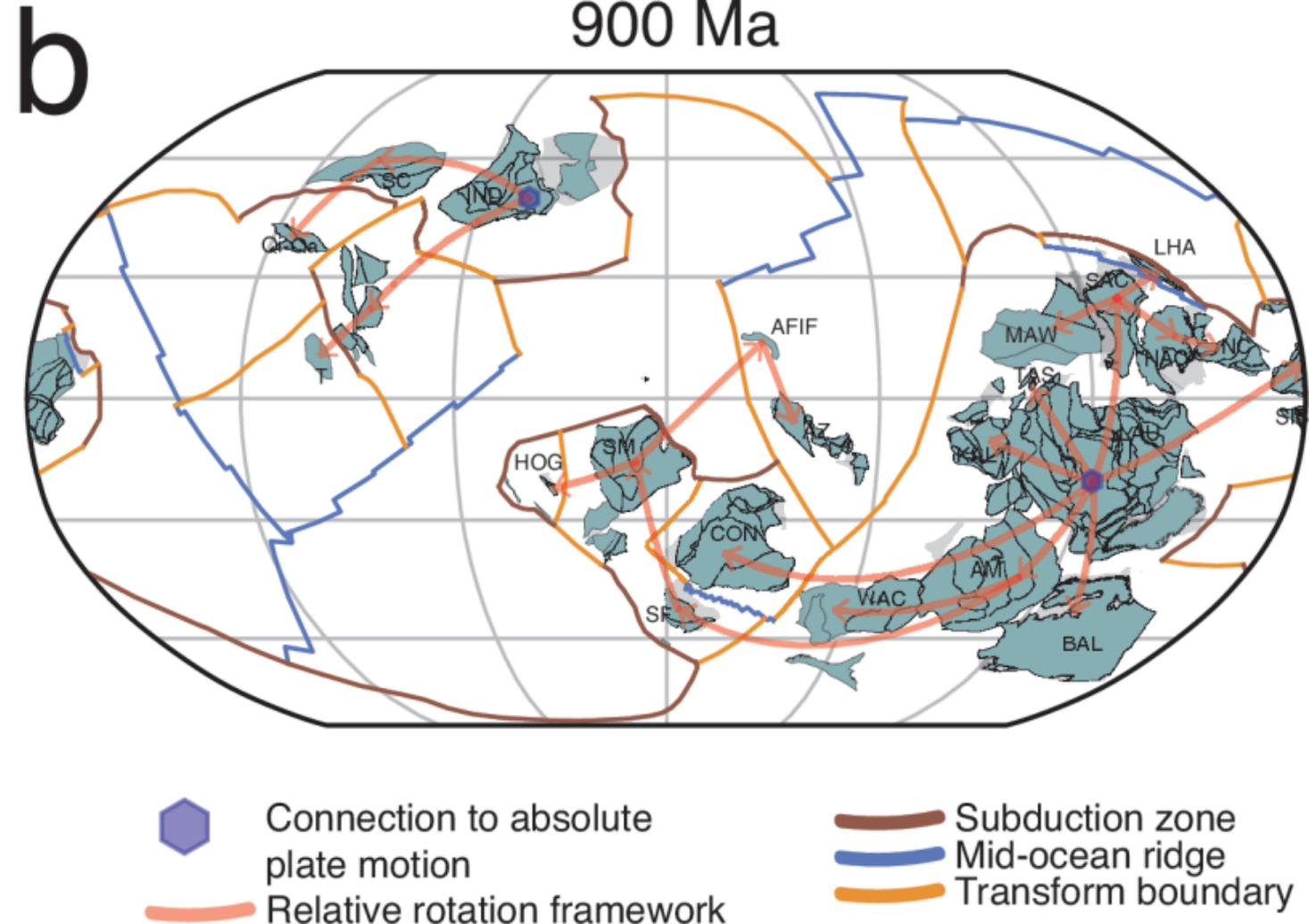


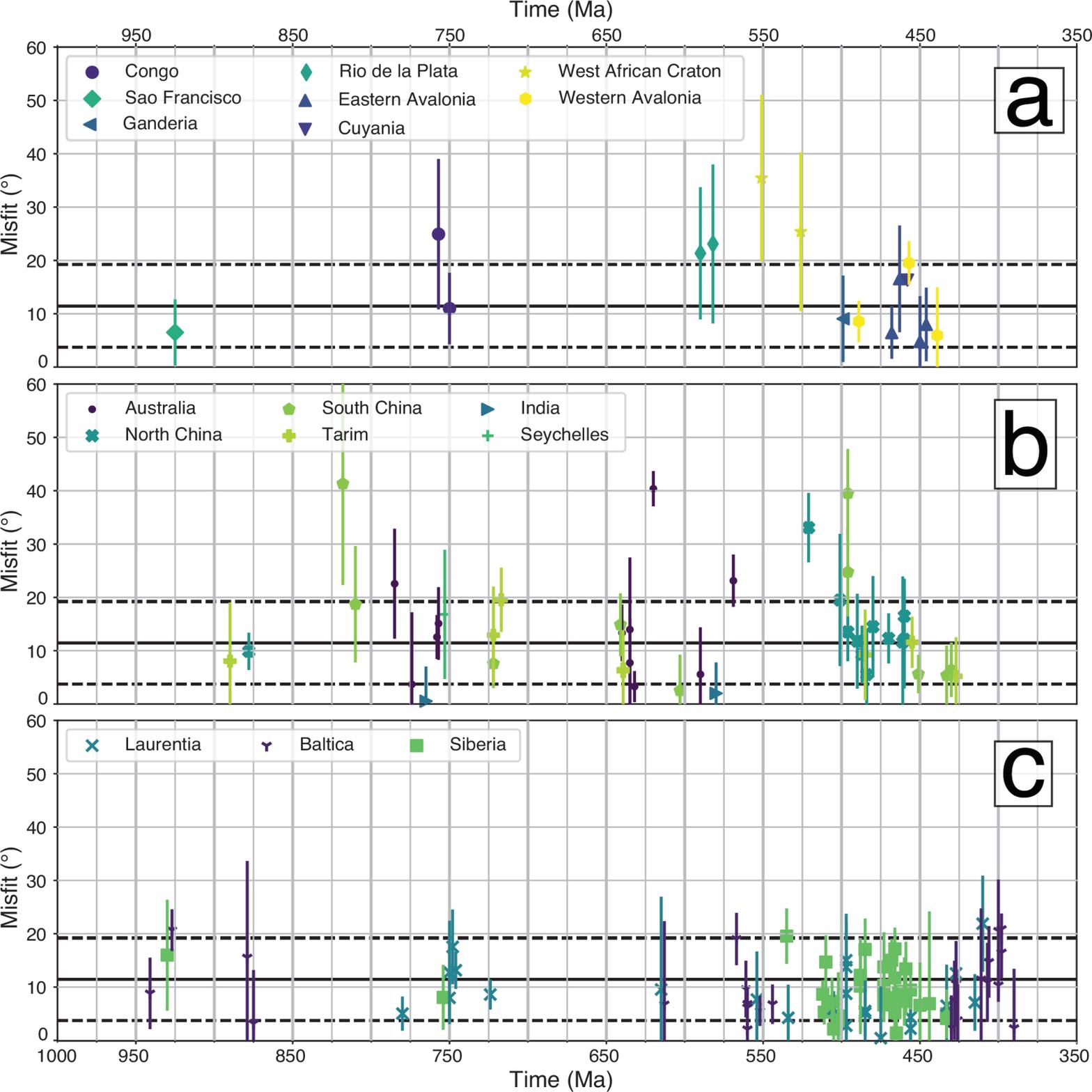


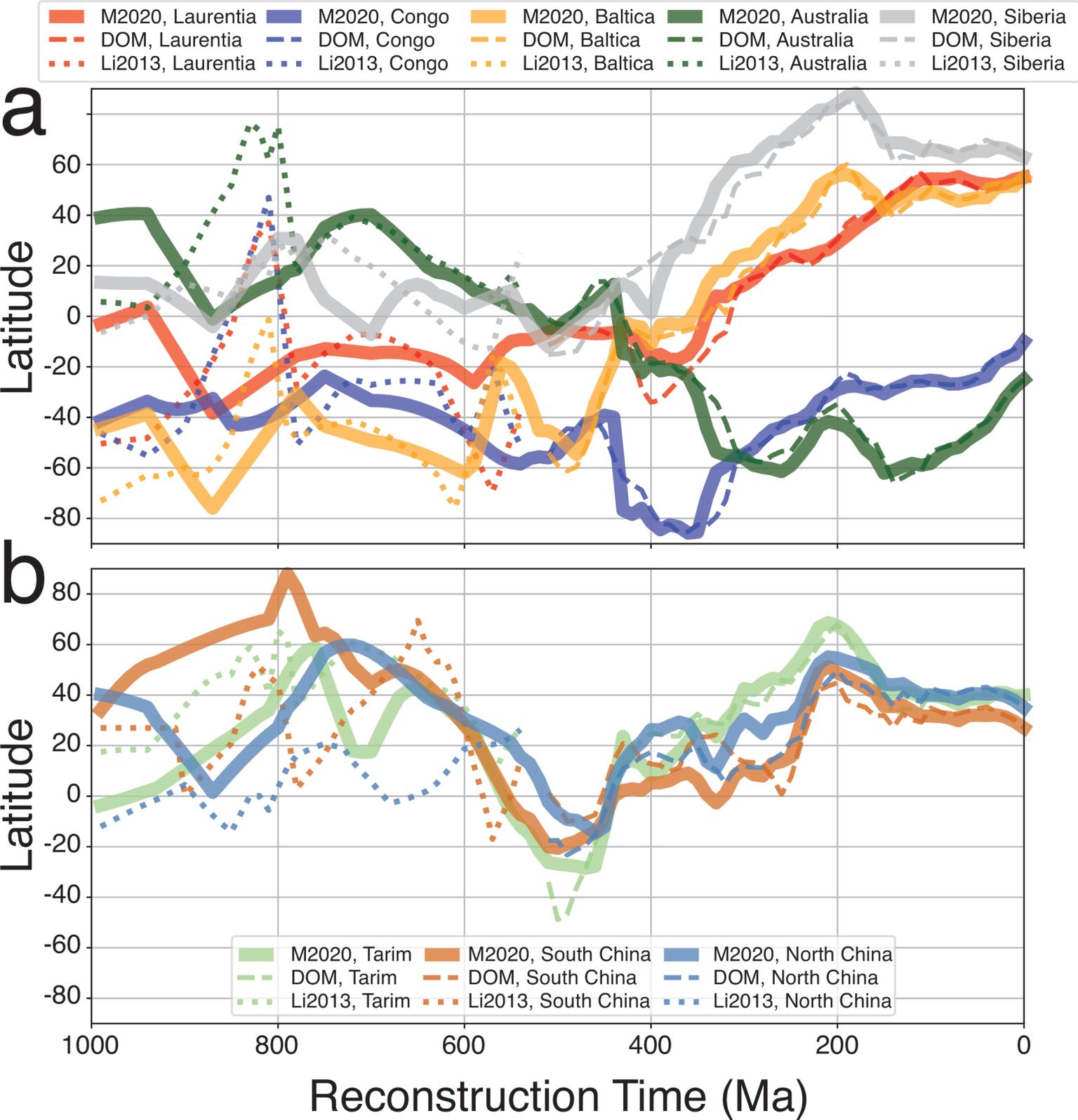


CLUSTER ONE

CLUSTER TWO

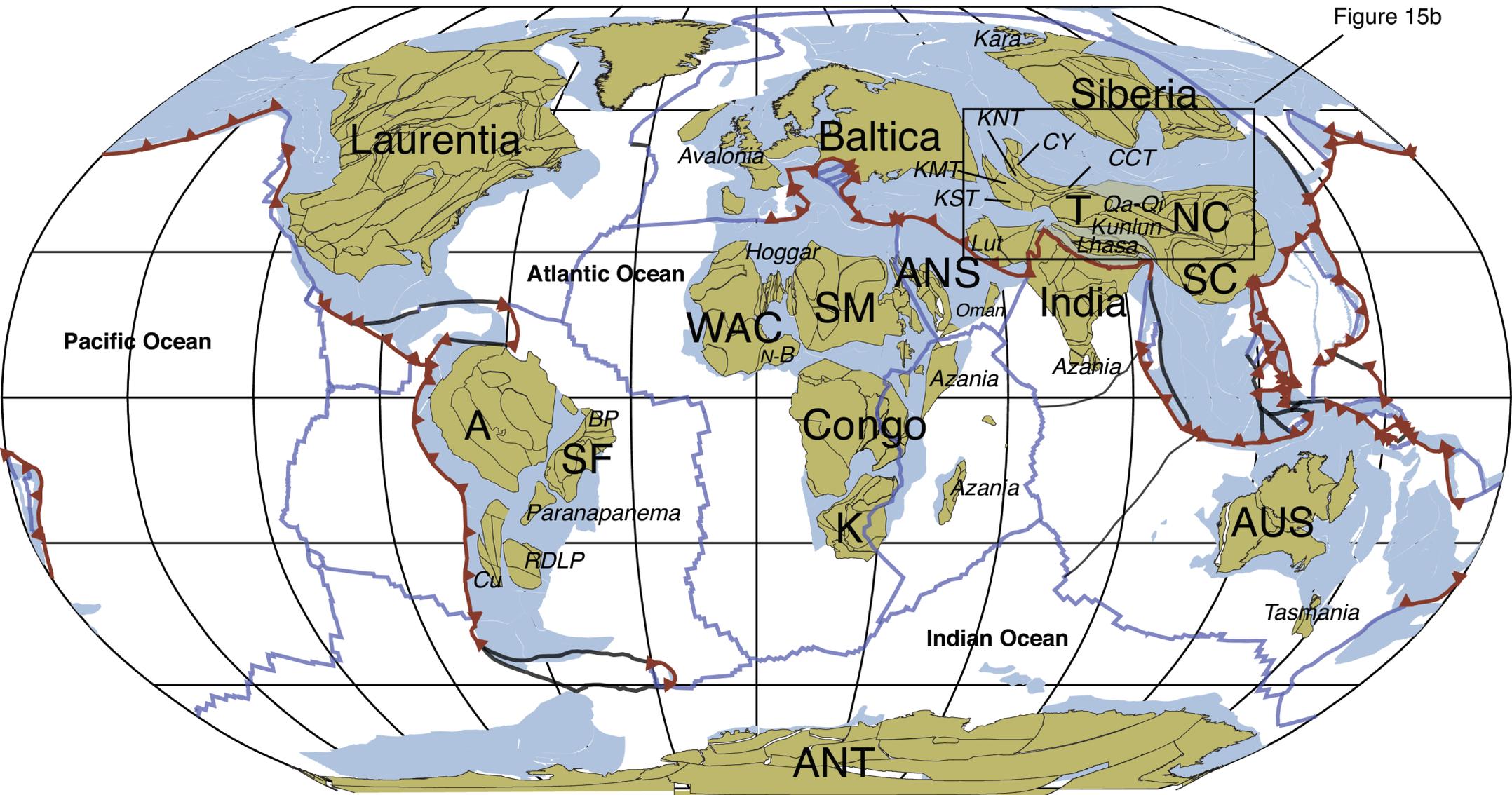






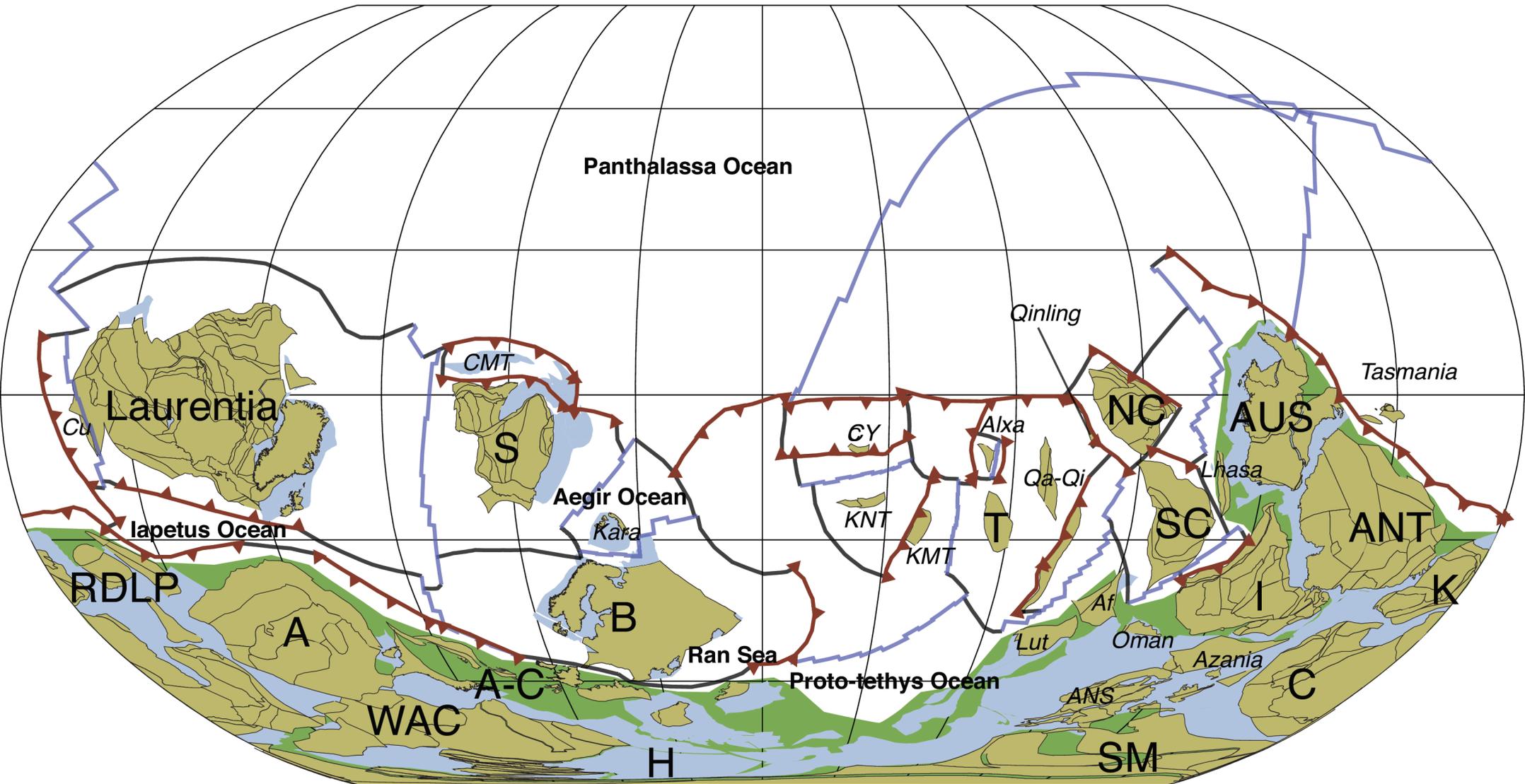
0 Ma

Figure 15b



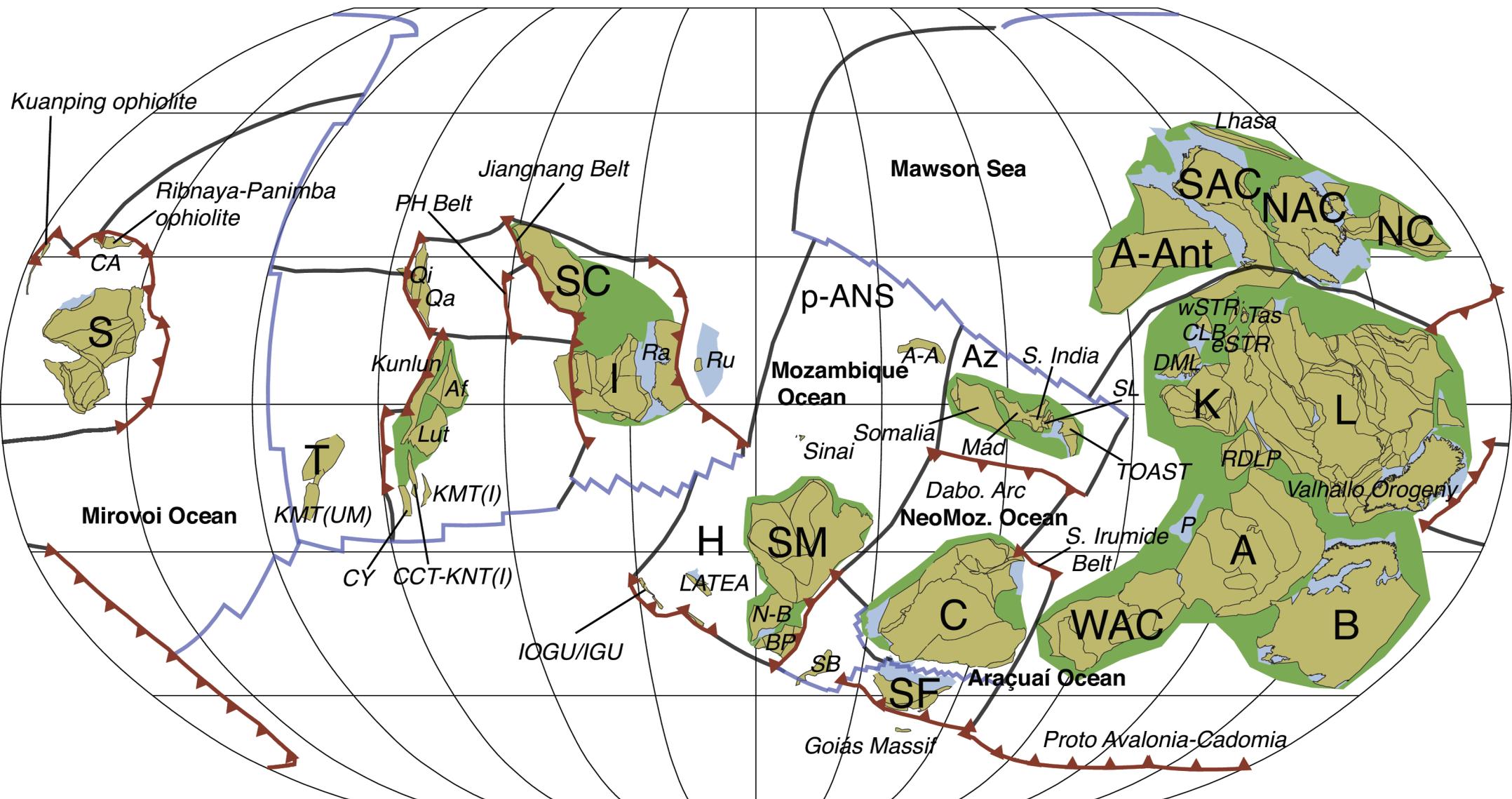
 Convergent boundary
  Divergent boundary
  Transform boundary

500 Ma



—▲— Convergent boundary > Divergent boundary — Transform boundary

1000 Ma



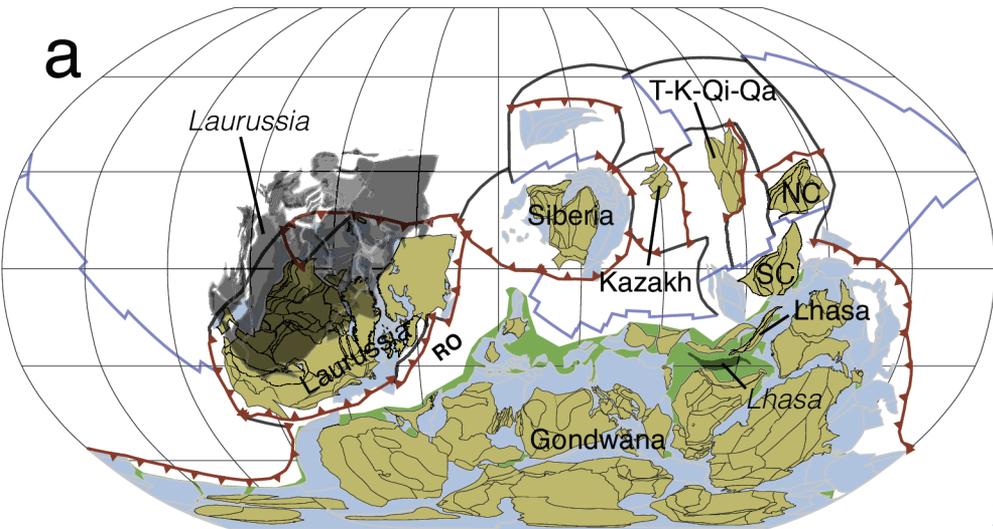
Convergent boundary

Divergent boundary

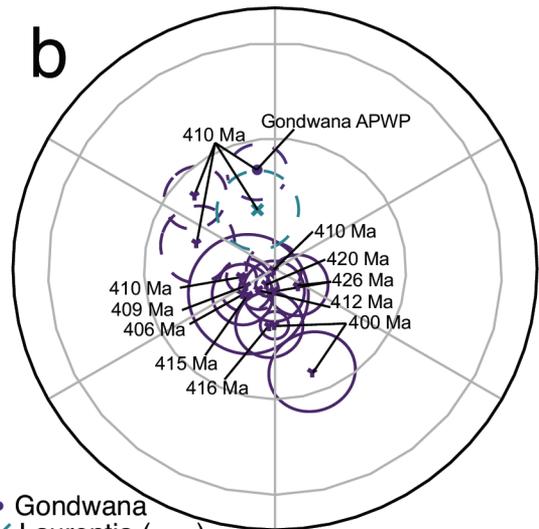
Transform boundary

MER20, 410 Ma

a



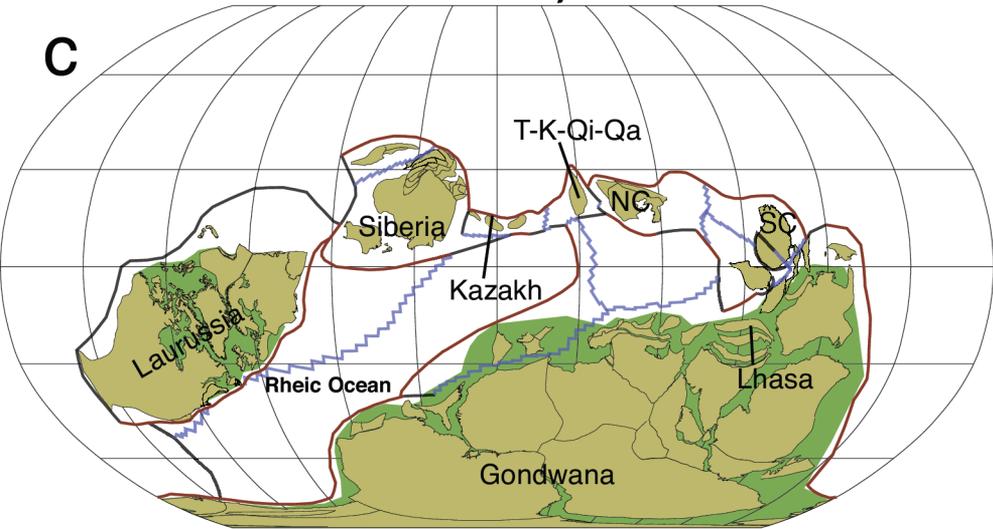
b



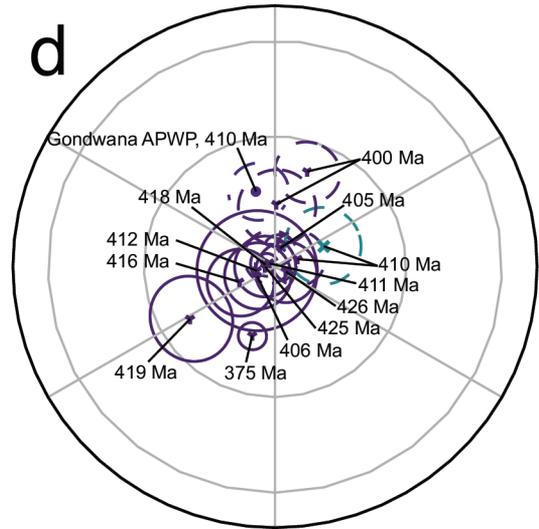
- Gondwana
- × Laurentia (— ·)
- ∨ Baltica, British Isles (—)
- · cratonic Europe (— ·)

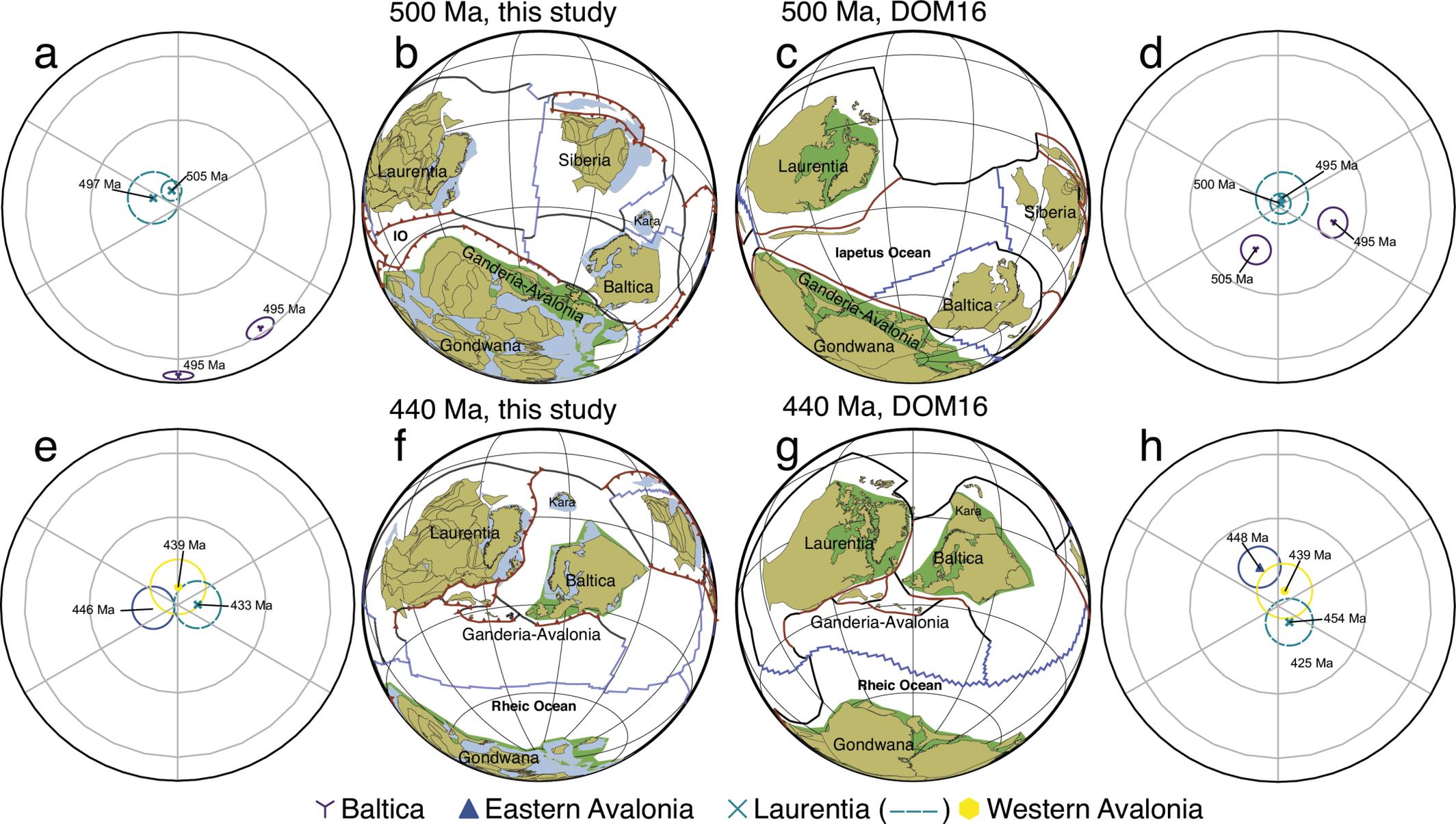
DOM16/18, 411 Ma

c



d

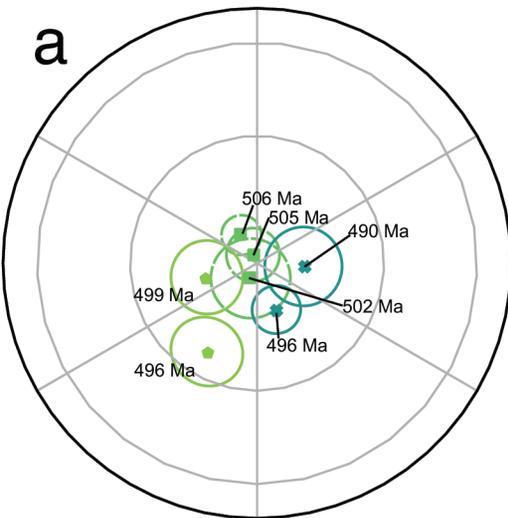




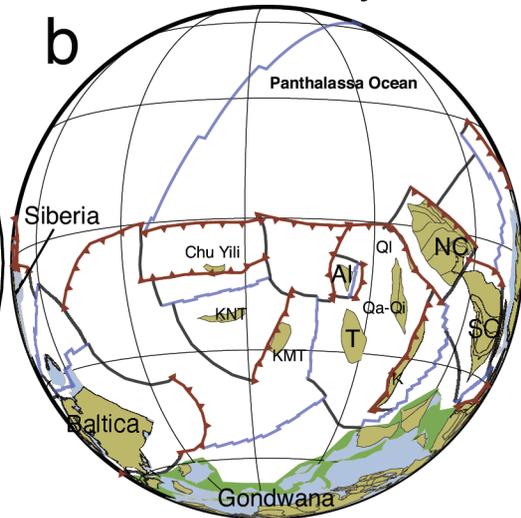
500 Ma, this study

500 Ma, DOM18

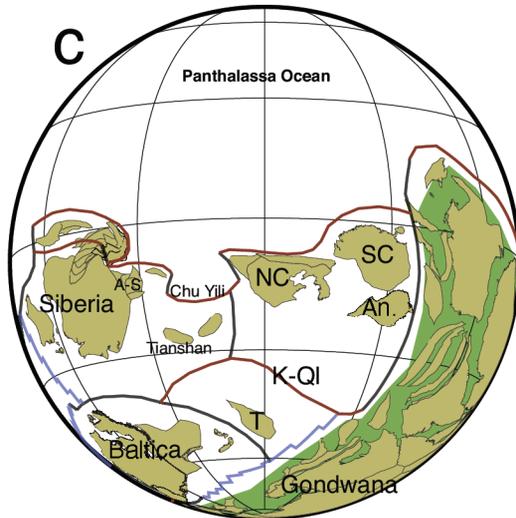
a



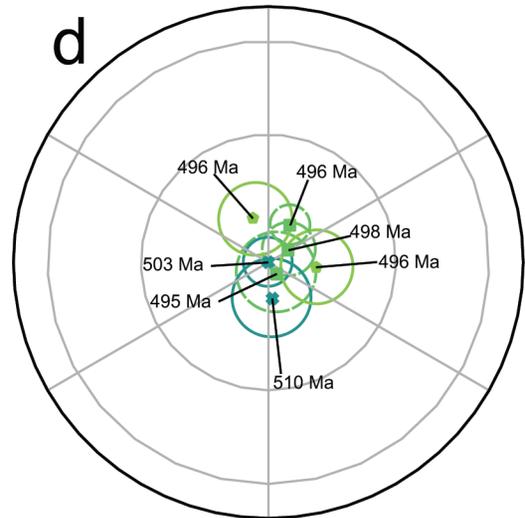
b



c



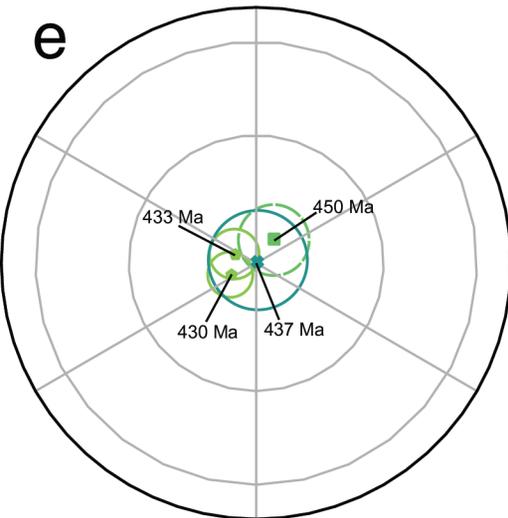
d



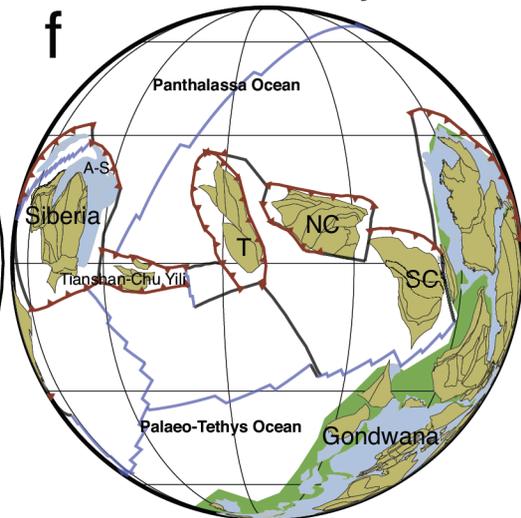
440 Ma, this study

440 Ma, DOM18

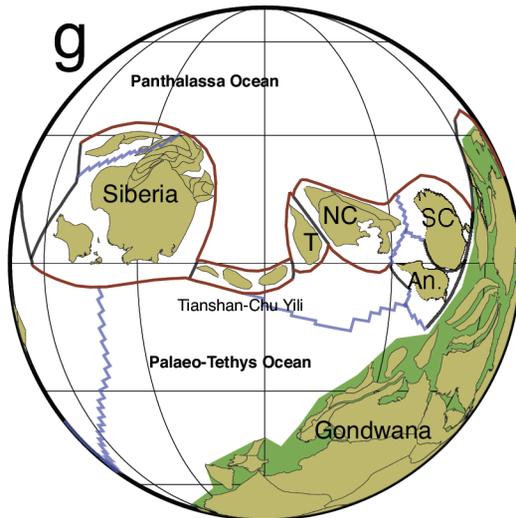
e



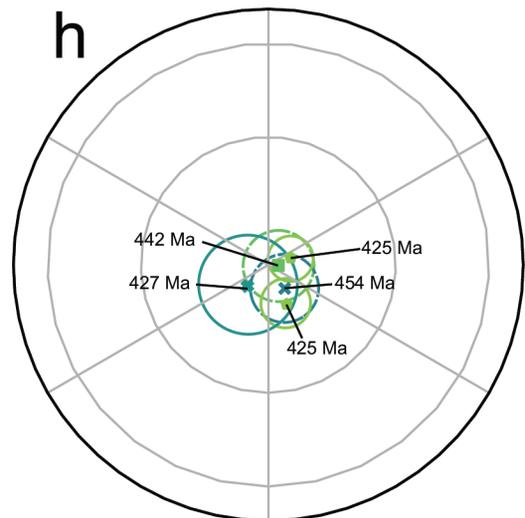
f



g

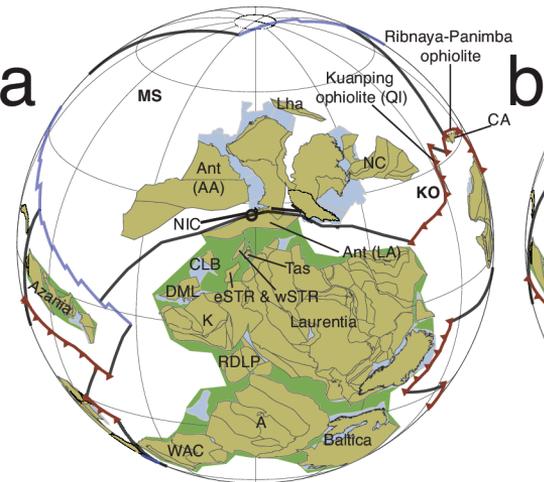


h

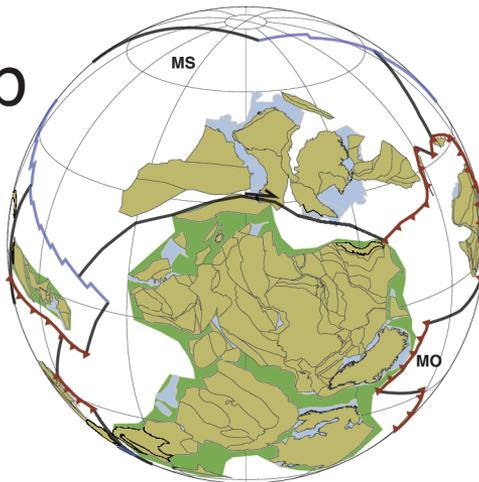


◆ North China
 ■ Siberia
 ◆ South China

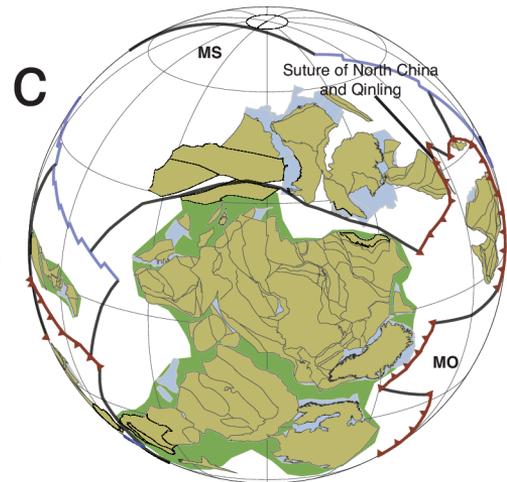
1000 Ma



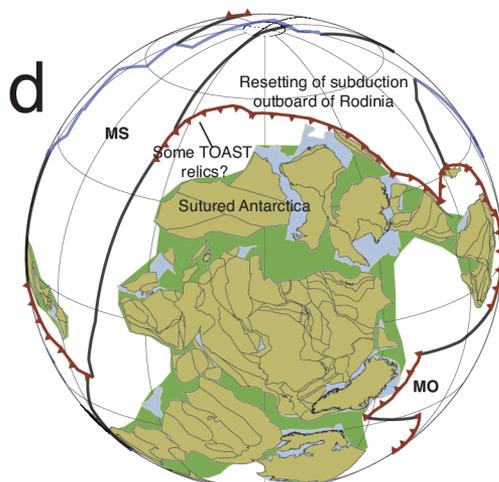
975 Ma



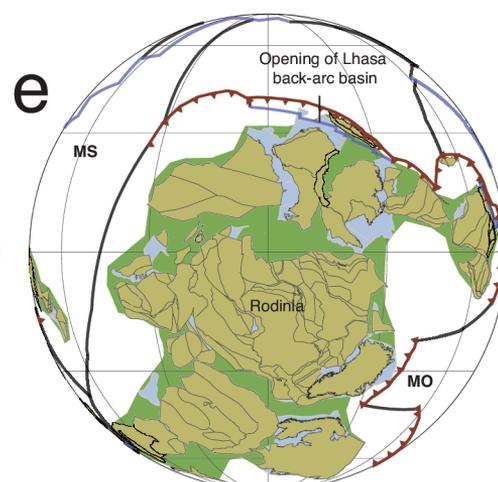
950 Ma



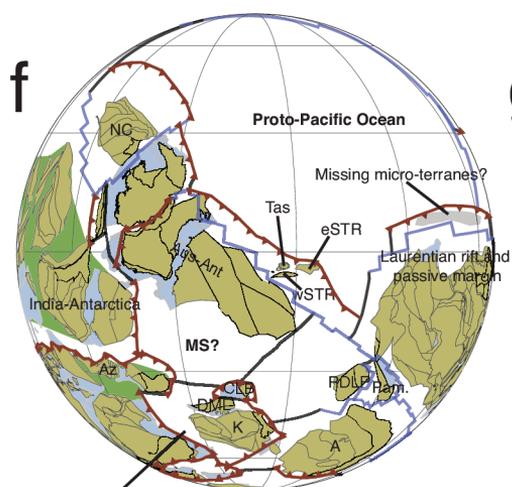
925 Ma



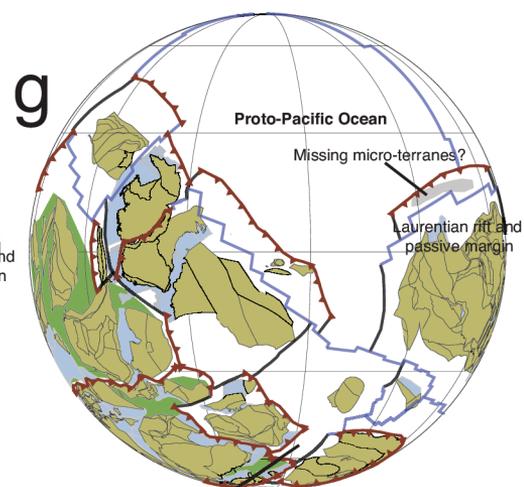
900 Ma

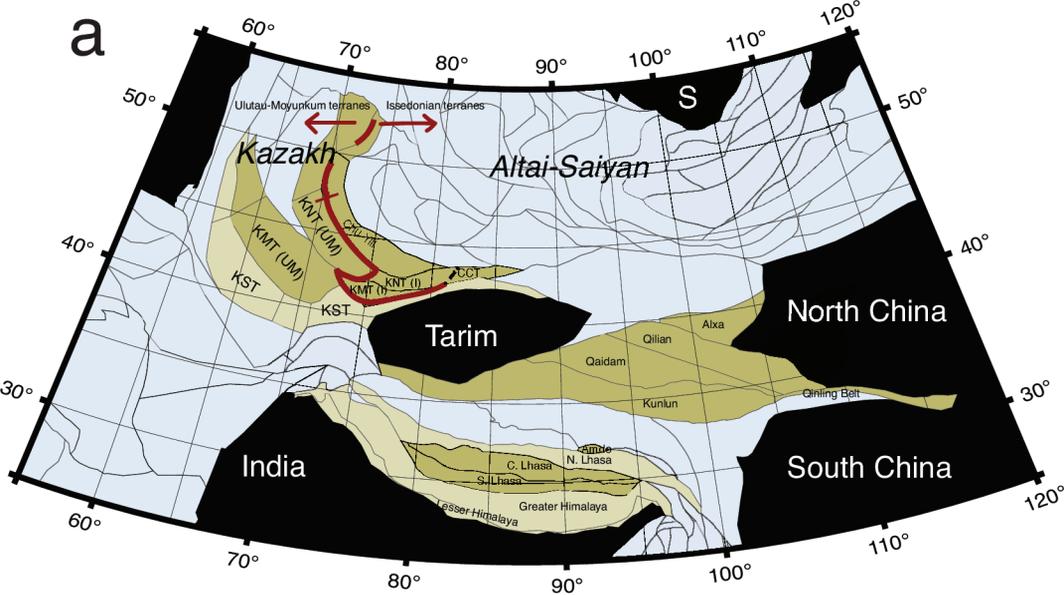


580 Ma



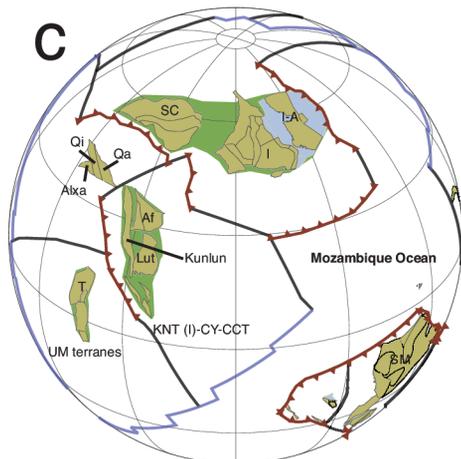
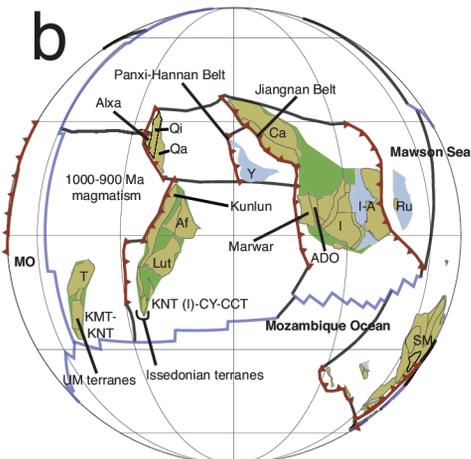
560 Ma





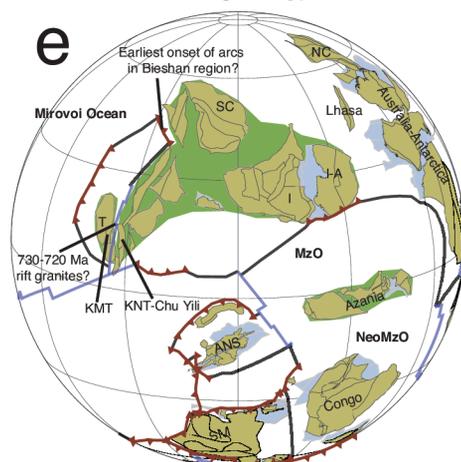
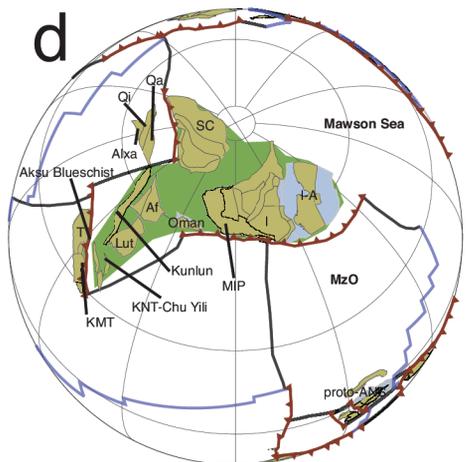
1000 Ma

900 Ma

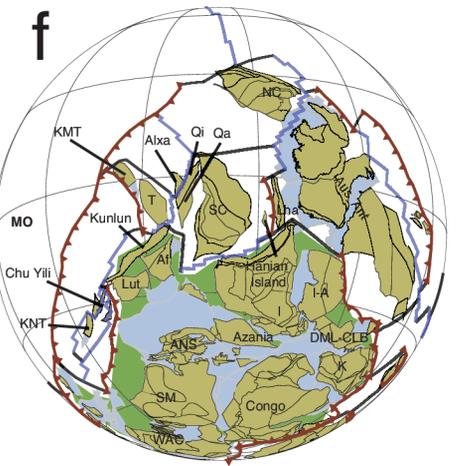


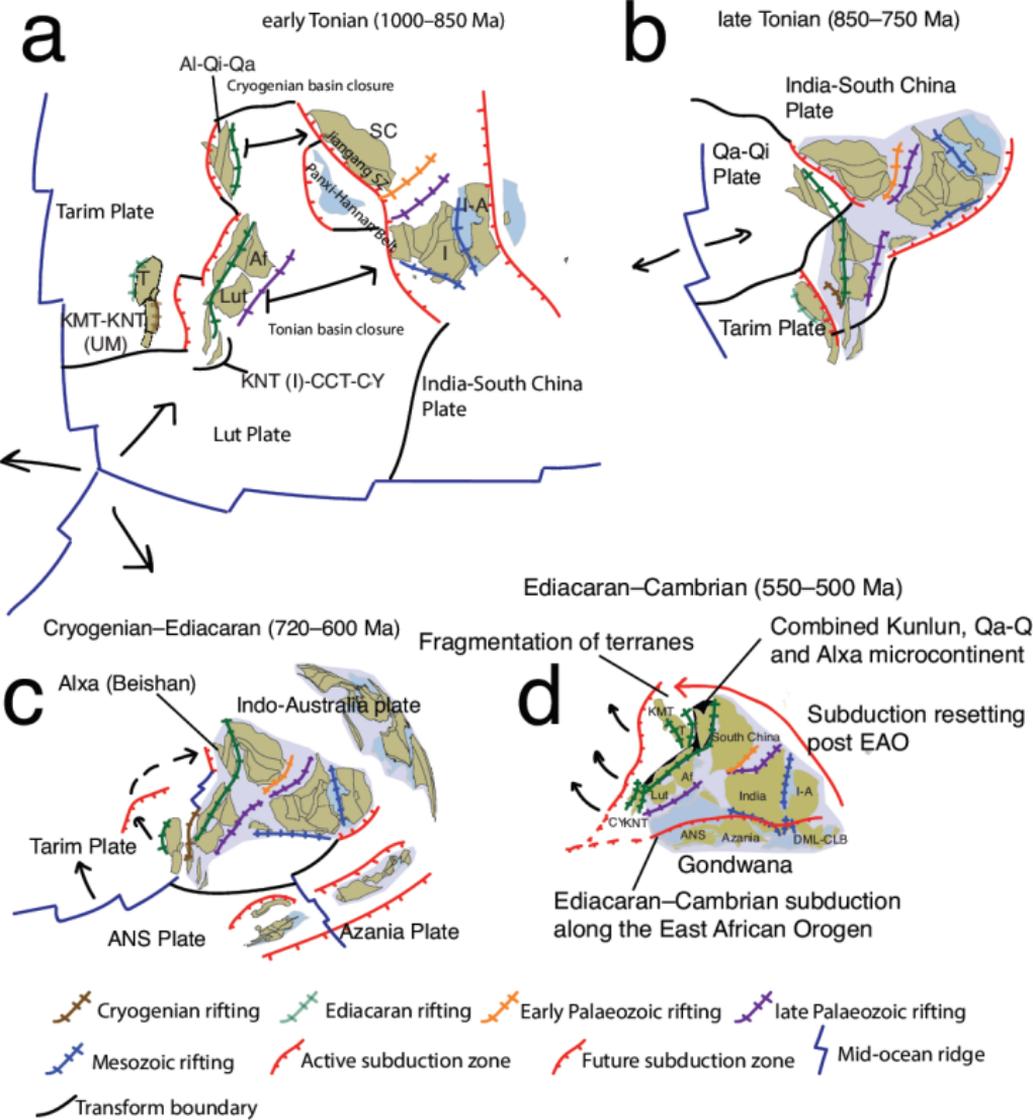
800 Ma

720 Ma



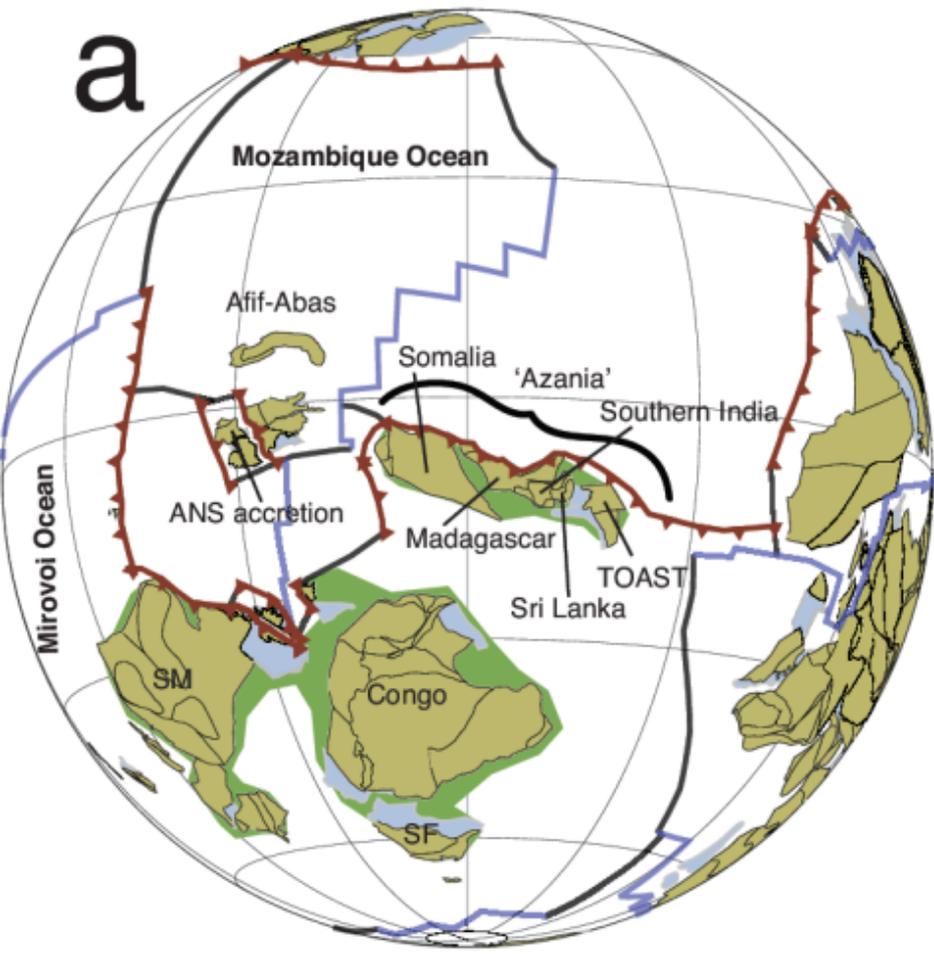
550 Ma





800 Ma

a



520 Ma

b

