

# Exploring the theory of plate tectonics: the role of mantle lithosphere structure

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**Abstract:** This review of the role of the mantle lithosphere in plate tectonic processes collates a wide range of recent studies from seismology and numerical modelling. A continually growing catalogue of deep geophysical imaging has illuminated the mantle lithosphere, and with it generated new interpretations of how the lithosphere evolves. Here, we present a review of the current ideas about the role of continental mantle lithosphere in plate tectonic processes. Evidence seems to be growing that scarring in continental mantle lithosphere is rather ubiquitous, which implies a reassessment of the widely-held view that it is inheritance of crustal structure only (rather than the lithosphere as a whole) that is most important in the conventional theory of plate tectonics (e.g., the Wilson Cycle). Recent studies have interpreted mantle lithosphere heterogeneities to be pre-existing structures, and as such linked to the Wilson Cycle and inheritance. We consider the current fundamental questions in the role of the mantle lithosphere in causing tectonic deformation, reviewing recent results alongside highlighting the potential of the deep lithosphere in infiltrating every aspect of plate tectonics processes.

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24 The reactivation of features formed through previous collisional or rifting events (i.e.,  
25 inheritance) is a tenet of plate tectonic theory (e.g., Wilson, 1966). Reactivation events occurring  
26 along well-defined, pre-existing features such as faults, shear zones or lithological contacts  
27 (Holdsworth et al., 1997) are well understood in that they form in preference to new structures  
28 (e.g. Sutton and Watson 1986; Butler et al. 1997 and references therein) during continental  
29 lithosphere deformation (e.g., major transcurrent fault systems, orogenic belts, and rifted basins  
30 in both intracontinental and continental margin settings (White et al., 1986; Handy, 1989;  
31 Tommasi et al., 1994, Holdsworth et al., 1997, 2001; Vauchez et al., 1998; Handy et al., 2001;  
32 Thomas, 2006)). Furthermore, the migration of hydrous fluids and magmas in continental  
33 regions are often through channelways defined by long-lived inherited structures (e.g. see  
34 Kerrich 1986; Hutton 1988; McCaig 1997), adding to the importance of pre-existing features in  
35 the continental lithosphere. Although discussion of inheritance in the mantle lithosphere has been  
36 conducted (e.g., Holdsworth et al., 2001), most research into this topic has focussed on crustal  
37 tectonics rather than any deeper structures (e.g., D’Lemos et al., 1997; Holdsworth, 2004;  
38 Thomas, 2006).

39

40 Compared to the overlying crust, the evolution of the mantle lithosphere is poorly understood;  
41 yet, as the main constituent of the lithosphere, this region is fundamental to controlling the  
42 tectonic behaviour of the Earth. Although the crust and the mantle lithosphere differ in their  
43 chemical compositions, the mantle lithosphere can be distinguished from the sub-lithosphere  
44 through mechanical properties related to flow regime. The rheology of the lithospheric layers  
45 governs deformation driven by interior forces (Bürgmann and Dresen, 2008), with elastic, plastic

46 (brittle), or viscous (ductile) properties exhibited (Burov, 2011). This layering of the lithosphere  
47 is complex, and often unique to the local environment. However, it is important to understand in  
48 the context of plate tectonics.

49  
50 Evidence is growing that heterogeneities within the mantle lithosphere are ubiquitous (e.g.,  
51 Rawlinson and Fishwick, 2011; Bastow et al., 2013; Schiffer et al., 2014, 2015, 2016; Schaeffer  
52 and Lebedev, 2014; Rasendra et al., 2014; Bao et al., 2014; Kahraman et al., 2015; Hopper and  
53 Fischer, 2015; Tauzin et al., 2016; Park and Levin, 2016a, 2016b; Biryol et al., 2016; Boyce et  
54 al., 2016; Dave et al., 2016). The first-order principles of what this means for past and future  
55 tectonic processes are still not clear. However, there are a number of studies offering theories as  
56 to what these structures can mean in terms of the wider Wilson Cycle process. Below, we  
57 outline broad descriptions of lithosphere rheology to contextualize the arena of study. In the  
58 following sections, we highlight the processes involved in the Wilson Cycle (focussing on  
59 inherited structures), followed by a discussion on imaging structures in the mantle lithosphere  
60 and the difficulty in unravelling the processes required to generate them, culminating in an  
61 analysis of recent numerical models and seismic studies that add to the understanding of the role  
62 of the mantle lithosphere in the Wilson Cycle. The main focus of the review is to bring together  
63 thoughts on the mantle lithosphere and, to begin, we need to understand how the layer behaves  
64 rheologically.

65

## 66 **Lithosphere rheology**

67

68 Layering is present within tectonic plates due to the modifying effects of depth-dependent  
69 temperature and pressure on rheology. Through the extrapolation of experimental rock  
70 mechanics data, yield-strength envelopes can predict the maximum differential stress supported  
71 by rock as a function of depth (Goetze and Evans, 1979). By integrating the plastic and ductile  
72 conditions of the material within each layer as a function of temperature and pressure, the flow  
73 regime of the lithosphere can be estimated. As a result, yield-strength envelopes offer an insight  
74 into the mechanical behaviour of lithospheric plates (Burov, 2011).

75

76 Bürgmann and Dresen (2008) outlined three food-based analogies to the strength of continental  
77 tectonic plates: jelly sandwich; crème brûlée; and banana split (Figure 1). A ‘jelly sandwich’  
78 strength profile is characterized by a weak lower crust (jelly) between a strong upper crust and  
79 mantle lithosphere (bread), as shown in Figure 1a. Relatively cool temperatures in continental  
80 interiors generate a strong upper crust (Rutter and Brodie 2004a,b; Rybacki et al. 2006),  
81 governed by Mohr-Coulomb theory to produce frictional plastic deformation. The lower crust  
82 transitions to viscous flow as temperature and pressure increase, producing a weak ductile layer  
83 (Bürgmann and Dresen, 2008). The strength of the jelly sandwich profile lies in the ultramafic  
84 mantle (Hirth and Kohlstedt 2003). A ‘crème brûlée’ profile describes a lithosphere where the  
85 strength resides within the crust (Figure 1b), with high temperatures and/or water content  
86 weakening the material strength below the crust (Jackson, 2002). The brittle crust produces a  
87 deformation regime which acts as the lid to the crème brûlée profile.

88

89 Jelly sandwich and crème brûlée can describe the profile within a continental interior (the third  
90 profile – banana split – predominately describes plate boundaries and will be discussed below)

91 and have generated some discussion as to the preferred model to be used in geodynamic analysis.  
92 Studies into earthquake distribution suggest that continental mantle lithosphere could behave in a  
93 ductile manner, with most of the strength of the lithosphere residing in the upper crust (i.e., a  
94 crème brûlée rheology) (Déverchère et al., 2001; Jackson, 2002; Maggi et al., 2000). However,  
95 laboratory flow laws indicate that the mantle lithosphere would have a complex layering of  
96 brittle and ductile material (e.g., Brace and Kohlstedt, 1980; Sawyer, 1985; Gueydan et al.,  
97 2014), with a broad consensus in the literature indicating that the mantle lithosphere would be  
98 strong enough to support high stresses. Old stable intraplate lithosphere has been interpreted to  
99 not have a crème brûlée rheology as it would not maintain the strength and stability to support a  
100 craton over long-timescales (Burov and Watts, 2006; Burov, 2010).

101  
102 The final model is described as a ‘banana split’ and refers to the changing strength profile across  
103 a plate boundary (Bürgmann and Dresen, 2008). Thermal, fluid, and strain-rate processes can  
104 combine at tectonic boundaries to weaken the overall strength of the lithosphere (Figure 1c).  
105 Major crustal fault zones are taken into consideration with this strength profile, with zones of  
106 weakness being generated throughout the thickness of the lithosphere (Bürgmann and Dresen,  
107 2008). Previous studies on mature fault zones (e.g., the San Andreas) have suggested a  
108 frictionally weak crust, with weakened shear zones within the viscous regime (Zoback et al.,  
109 1987). There are a number of mechanisms that can produce weakening at plate boundaries, such  
110 as grain-size reduction (Bercovici and Ricard, 2014; Krajcinovic, 1996; Skemer et al., 2010;  
111 Warren and Hirth, 2006; Linckens et al., 2015), that occur through plate tectonic processes  
112 related to the Wilson Cycle.

113

## 114 **The Wilson Cycle**

115

116 In 1966, based on evidence in the fossil record and the dating of vestiges of ancient volcanoes,  
117 Wilson (1966) proposed a cycle describing the opening and closing of oceanic basins. This cycle  
118 provided a method of amalgamating continental material (into a supercontinent) that would be  
119 subsequently dispersed (e.g., into a fragmented configuration like the present-day). Wilson  
120 (1966), building on previous studies (e.g., Hess, 1962; Vine and Matthews, 1963; Wilson, 1965),  
121 outlined a four-stage “Wilson Cycle” (as it was later named by Dewey and Burke (1974)): the  
122 dispersal (or rifting) of a continent; continental drift, seafloor spreading, and the formation of  
123 oceanic basins; new subduction initiation and the subsequent closure of oceanic basins through  
124 oceanic lithosphere subduction; and continent-continent collision and closure of the oceanic  
125 basin (Figure 2).

126

127 Over the past 50 years this conventional theory of plate tectonics has been at the forefront of  
128 geodynamics. However, many features of lithosphere evolution fall outside the realm of the  
129 Wilson Cycle: plate tectonics has progressed beyond plate boundaries as the sole locus of major  
130 deformation with the study of intraplate orogenesis (e.g., Sykes, 1972, 1978; Smith and Bruhn,  
131 1984; Sibson, 1992; Ziegler et al., 1995, 1998; Stein and Liu, 2009; Stephenson et al., 2009);  
132 mantle lithosphere processes generating lithospheric instabilities (in the form of viscous dripping  
133 and delamination) that represent a foundering and recycling of plate material (e.g., Bird, 1979;  
134 Houseman et al., 1981, 1997; Gögüs and Pysklywec, 2008; Bajolet et al., 2012; Gögüs et al.,  
135 2016) in situ mantle lithosphere inversion of Archean cratonic keels (Percival and Pysklywec,  
136 2007); and the interaction of subduction and large low shear velocity provinces in driving the

137 development of large igneous provinces at the surface (e.g., Ernst et al., 2005; McNamara and  
138 Zhong, 2005; Bull et al., 2009; Heron et al., 2015a; Mallard et al., 2016).

139

140 Among these, the study of intraplate orogenesis has generated several mechanisms for  
141 deformation within a plate interior (Figure 2). These mechanisms include pre-existing  
142 lithosphere structures, the presence of fluids, the burial of highly radiogenic material and other  
143 temperature anomalies, mantle lithosphere instability, compositional strengthening, and strain  
144 rate (e.g., Ziegler, 1987; Ziegler et al., 1995, 1998; Sandiford, 1999; Nielsen and Hansen, 2010;  
145 Hansen and Nielsen, 2002; Pysklywec and Beaumont, 2004; Sandiford et al., 2006; Stephenson  
146 et al., 2009; Heron and Pysklywec, 2016). If intraplate orogenesis can be influenced by similar  
147 mechanisms that generate other (established) plate tectonic processes (such as rifting), then it  
148 should be recognized as part of plate tectonic theory (e.g., Figure 2).

149

## 150 **Inheritance**

151

152 Experiments on rock properties find that deformation generates weak zones that, over time, can  
153 be dormant (or be reactivated) depending on how the material strength is affected by changes in  
154 ambient stresses. A reduction in grain size is a characteristic of this lithospheric damage  
155 (Bercovici and Ricard, 2014), which can be abundant at tectonic margins in the form of  
156 peridotite mylonites (Warren and Hirth, 2006; Skemer et al., 2010). The lithospheric strength of  
157 the banana split model (Figure 1c) could be indicative of this weakness at plate boundaries given  
158 the rheological impact of the reduced grain size.

159

160 The reactivation of structures within the crustal lithosphere has previously been well documented  
161 as being part of Wilson Cycle processes (Holdsworth et al., 2001; Holdsworth, 2004). In terms of  
162 rifted continents, brittle structures in the shallow crust inherited from previous tectonic events  
163 have been interpreted to define the shape of the margin (Thomas, 2006). Furthermore, crustal  
164 inheritance could also play a role in intraplate deformation. Stephenson et al. (2009) identified  
165 that thermal structures from previous tectonic events could also play an important role in  
166 deformation away from plate boundaries in southeastern Ukraine. The continuation of ancient  
167 tectonics to influence deformation, even away from active plate boundaries, is a strong indication  
168 of the role of inheritance in all forms of plate tectonics.

169

170 In discussing Laurentian-age rifting through Appalachian-Ouachita structures, Thomas (2006)  
171 interpreted that inheritance would be on a lithospheric scale. This notion that the mantle  
172 lithosphere would be susceptible to inherited structures, just as the crust would be, is in keeping  
173 with several studies highlighting the complete lithosphere as playing a part in deformation (e.g.,  
174 Vauchez et al., 1997, 1998; Holdsworth et al., 2001; Bendick and Flesch, 2013; Li et al., 2016).  
175 In studying why continents seem to break-up parallel to orogenic belts, Vauchez et al. (1997)  
176 proposed that a pervasive fabric exists in the mantle lithosphere from ancient collisional events  
177 that can guide the propagation of continental rifts. Although the mantle lithosphere has been  
178 inferred to control rifting within the Wilson Cycle, the region has not had the same attention as  
179 the crust in terms of the evolution of the lithosphere. This is due, in part, to the difficult nature of  
180 studying the mantle lithosphere through imaging methods. However, recent advances have seen a  
181 substantial increase in research into the sub-crustal lithosphere.

182



## 183 **Imaging the mantle lithosphere**

184

185 Afonso et al. (2016) described the range of approaches used to study the lithosphere and upper  
186 mantle: teleseismic tomography (e.g., see Evans and Achauer (1993), Granet et al. (1995),  
187 Rawlinson et al. (2006)); surface-wave tomography (e.g., see Pasyanos and Nyblade (2007),  
188 Yang et al. (2008), Fishwick et al. (2008), Agius and Lebedev (2013)); gravity modelling (e.g.,  
189 see Zeyen and Fernández (1994), Torne et al. (2000), Ebbing et al. (2006), Chapell and Kuszniir  
190 (2008), Tašárová et al. (2009)); electromagnetic methods (e.g., see Heinson (1999), Jones  
191 (1999), Jones et al. (2009), Evans et al. (2005), Evans et al. (2011), and Meqbel et al. (2014));  
192 local earthquake tomography (e.g., Aki and Lee (1976), Eberhart-Phillips (1990), and Kissling et  
193 al. (1994)); and receiver function studies (e.g., Yuan et al. (2006), Kawakatsu et al. (2009),  
194 Rychert and Shearer (2011), Kind et al. (2012)).

195

196 The increase in the number of high-resolution large-scale seismic arrays used in studies across  
197 the world has allowed for a clearer image of the deep lithosphere. The successful Lithoprobe  
198 project lasted from 1984 to 2005 and produced over 1500 publications on the evolution of the  
199 northern North American lithosphere. EarthScope initiated a 15-year programme of USArray,  
200 which consisted of the deployment of temporary and permanent seismic stations across the  
201 United States (comprising a Transportable Array, a Flexible Array, a (permanent) reference  
202 network and a magnetotelluric facility). The dense, moving network allowed for an  
203 unprecedented increase of image resolution of the North American lithosphere (e.g., Schaeffer  
204 and Lebedev, 2014). Other recent high resolution networks include (but are by no means limited  
205 to): the AFRICA Array (e.g., O'Donnell et al. 2016); the WOMBAT seismic array (e.g.,

206 Rawlinson and Fishwick, 2011); the M.A.G.I.C. array studying the crust and upper mantle of the  
207 Appalachian mountains; the ocean-based MERMAID project (Mobile Earthquake Recorder in  
208 Marine Areas by Independent Divers) uses floating receivers to image the deep earth (e.g., Hello  
209 et al., 2011); DANA (Dense Array in Northern Anatolia), imaging northern Turkey tectonics  
210 (e.g., Kahraman et al., 2015); the POLARIS (Portable Observatories for Lithospheric Analysis  
211 and Research Investigating Seismicity) array in Canada (e.g., Bastow et al., 2013); and the China  
212 National Digital Seismic Network (CNDSN) (e.g., Niu and Li, 2011; Bao et al., 2013).

213

214 This increase in research using large-scale imaging studies, alongside new techniques in  
215 acquisition and data processing (cf. Romanowicz, 2003; Artemieva et al., 2006; Rawlinson et al.,  
216 2010; Liu and Gu, 2012; Kuvshinov and Semenov, 2012) has also allowed structures below the  
217 Moho to be seen, with a multi-observable approach often built into the studies permitting  
218 corroboration of findings (e.g., deploying seismic and magnetotelluric stations). Results from  
219 new post-processing techniques of receiver function data have been encouraging (e.g., Rasendra  
220 et al., 2014; Tauzin et al., 2016; Park and Levin, 2016a; 2016b). The combination of receiver  
221 function and shear-wave splitting analysis on dense cross-fault arrays, as described in Rasendra  
222 et al. (2014), has been able to better characterize and understand the mechanics of large-scale  
223 strike-slip faults from the surface to the bottom of the lithosphere. When there is high-resolution  
224 imaging below the Moho, heterogeneities in the mantle lithosphere are ubiquitous (e.g.,  
225 Rawlinson and Fishwick, 2011; Bastow et al., 2013; Schiffer et al., 2014, 2015, 2016; Schaeffer  
226 and Lebedev, 2014; Rasendra et al., 2014; Bao et al., 2014; Kahraman et al., 2015; Hopper and  
227 Fischer, 2015; Tauzin et al., 2016; Park and Levin, 2016a, 2016b; Biryol et al., 2016; Boyce et  
228 al., 2016; Dave et al., 2016). The relevance of these structures is currently being debated, but

229 ultimately an understanding of them will help determine the role of the mantle lithosphere in the  
230 theory of plate tectonics.

231

### 232 **Unravelling the tectonic impact of the mantle lithosphere**

233

234 Through seismic imaging and geochemical analysis, the mantle lithosphere has been known to be  
235 disturbed or “scarred” for many years (e.g., Wendlandt et al., 1993; Lee et al., 2001; Yuan and  
236 Romanowicz, 2010; Lee et al., 2011), with deep inherited structures often interpreted to be the  
237 result of closure of ocean basins and continental collisions (e.g., Flack and Warner, 1990;  
238 Klemperer and Hobbs, 1991; Lie and Husebye, 1994; Morgan et al., 1994; Guellec et al., 1990;  
239 Pfiffner, 1992; Calvert et al., 1995; Calvert and Ludden, 1999; Cook et al., 1999; van der Velden  
240 and Cook, 2002; Cook, 2002; Cook and Vasudevan, 2003; White et al., 2003; Cook et al., 2004;  
241 van der Velden and Cook, 2005; Schiffer et al., 2014, 2015, 2016). The ages of these mantle  
242 lithosphere damage structures vary, with some features (Figure 3) thought to be of Archaean age  
243 (e.g., Calvert et al., 1995).

244

245 Although subduction scars have often been highlighted as a reason for the seismic visualization  
246 of mantle lithosphere reflectivity (e.g., Calvert et al., 1995; van der Velden and Cook, 2002;  
247 Cook, 2002), other processes exist that could create structures within the lithosphere. Van der  
248 Velden and Cook (2005) outline a number of other possibilities, including: mafic intrusions into  
249 the mantle (Steer et al., 1998); shear zones (Smythe et al., 1982; Warner and McGeary, 1987;  
250 Reston, 1990; McBride et al., 1995; Abramovitz et al., 1998); relict crustal fabrics and/or Moho

251 (Snyder, 1990; Cook and Vasudevan, 2003); and the lithosphere-asthenosphere boundary (Steer  
252 et al., 1998b).

253

254 The propensity of continents to break apart parallel to ancient orogenic belts also indicates a role  
255 of inherited structures in controlling tectonics, with rheological heterogeneity and mechanical  
256 anisotropy playing a factor (Vauchez et al., 1997, 1998). Furthermore, plate tectonic processes  
257 such as extensional stresses and plate bending prior to subduction have been suggested to  
258 weaken the rheology of oceanic lithosphere through the percolation of low-degree melts in  
259 metasomatic processes (Pilet et al., 2016). Taking such discussions into consideration, it is  
260 appropriate to interpret the seismic imaging of scarring to be regions of weakness in the  
261 continental mantle (e.g., Linckens et al., 2015; Heron et al., 2016a).

262

263 The role of grain damage in tectonic processes is also a method by which weakening could occur  
264 in the mantle lithosphere. In recent studies, Heron et al. (2016a, 2016b) interpret the seismic  
265 imaging of mantle lithosphere heterogeneities to be ancient deformation, with the reduction in  
266 grain size acting as a weak plane (Bercovici and Ricard, 2014). Lithospheric damage related to  
267 inheritance has been inferred to remain weak over very long timescales (Audet and Bürgmann,  
268 2011), allowing ancient processes related to Archean scarring to be considered in present-day  
269 tectonics. At present, further constraints from the geological history of a region are required to  
270 unravel the processes related to the generation of mantle lithosphere heterogeneities and their  
271 impact on crustal tectonics. Numerical modelling has been shown to be useful in adding to the  
272 discussion on this topic of mantle lithosphere processes, an example of which (Heron et al.,  
273 2016b) is discussed below. Heron et al. (2016b) presented 2-D numerical experiments of

274 continental convergence to generate intraplate deformation from inherited lithospheric structures  
275 (Figure 4a), exploring the limits of continental rheology to understand the dominant lithosphere  
276 layer across a broad range of geological settings.

277

### 278 **Constraints from numerical modelling**

279

280 The numerical experiments in Heron et al. (2016b), with some results shown here in Figure 4,  
281 were modelled using the two-dimensional, thermal-mechanical finite element numerical code  
282 SOPALE (Fallsack, 1995), which implements an Arbitrary Lagrangian-Eulerian (ALE) method  
283 to solve for the deformation of high Prandtl number incompressible viscous-plastic media. The  
284 models consider convergence in a stable (i.e., strong) (Burov and Watts, 2006) continental crust  
285 and mantle lithosphere setting (e.g., jelly sandwich rheology, Figure 1a) where the majority of  
286 mantle lithosphere scars are found (e.g., Steer et al., 1998a; Heron et al., 2016a). The model  
287 setup allows for a heterogeneous lithosphere, with a number of different weak zones in both the  
288 crust and mantle lithosphere (Figure 4a).

289

290 In Figures 4b–4e, crustal and mantle lithosphere inheritance is prescribed from Figure 4a as  
291 shown by the white scars and red heterogeneity, respectively. This configuration of the upper  
292 crust and lower crust weak zones permits easy identification of which layer is controlling  
293 deformation. After considerable shortening (in keeping with the extent of similar tectonic  
294 scenarios) (e.g., Cowgill et al., 2003), crustal thickening and faulting, key characteristics of  
295 intraplate orogenesis, are shown in models that feature upper crust (UC) or lower crust (LC)  
296 scars (Figures 4b and 4c). The implementation of a weak scar in the mantle lithosphere (overlain

297 by a heterogeneous crust) dominates tectonics for this jelly sandwich rheology (Figure 4d). The  
298 models suggest that the impact of crustal scars is minimal when in the presence of a mantle  
299 lithosphere (ML) scar, as shown by comparing Figure 4d, featuring UC, LC, and ML scars, with  
300 Figure 4e, one ML scar only.

301  
302 By implementing a ‘crème brûlée’ rheology (e.g., Figure 1b), featuring a weak mantle  
303 lithosphere and strong crust, it is found that heterogeneities within the mantle lithosphere become  
304 ineffective in controlling tectonics (Figure 4f). We posit that if the continental mantle is the  
305 strongest layer within the lithosphere, then such inheritance may have important implications for  
306 the development of tectonic processes in the Wilson Cycle (e.g., Holdsworth et al., 2001).  
307 Indeed, the rheological strength of the lithosphere may be imperative in analysing the cause and  
308 effect of large-scale tectonics (especially as scarring in the lithosphere is seen as ubiquitous).  
309 Furthermore, the models of Heron et al. (2016b) show that deformation driven by mantle  
310 lithosphere scarring can produce tectonic patterns related to intraplate orogenesis originating  
311 from crustal sources, making it difficult to unravel the cause of tectonic evolution while  
312 highlighting the need for a more formal discussion of the role of the mantle lithosphere in plate  
313 tectonics.

314  
315 The Altyn Tagh Fault (ATF) in China illustrates the difficulty in unravelling tectonic cause and  
316 effect within the lithosphere. The tectonic history of China provides one reference to understand  
317 plate tectonics beyond plate boundaries with regards to the studies of Heron et al. (2016a,  
318 2016b). Although there are many regions across the world where continents are subject to  
319 Wilson Cycle processes such as the continent accretion by closure of paleo-oceans between

320 micro-plates, China is a unique reference as the far-field convergent stress from the Indian–  
321 Eurasian collision is relatively recent and ongoing (Figure 5a). The Altyn Tagh Fault (ATF), on  
322 the northern margin of the Tibetan Plateau, has a distinct present-day ML heterogeneity linked to  
323 a continent–continent suture (Cowgill et al., 2003). The ATF accommodates some of the  
324 convergence between the Indian and Eurasian plates (Zhang et al., 2014) and is characterized by  
325 localized deformation that has produced  $\sim 475 \pm 70$  km of staggered displacement since the mid-  
326 Oligocene (Cowgill et al., 2003). Although focal mechanisms of earthquakes close to the ATF  
327 show strike–slip motion, compressional processes account for earthquakes to the south (Zhang et  
328 al., 2014), with numerous thrust faults also inhabiting the area (Figure 5b). Geophysical studies  
329 of the ATF show deformation that penetrates the entire crust to link to heterogeneous structures  
330 in the ML (Wittlinger et al., 1998; Zhao et al., 2006; Zhang et al., 2014) (Figure 5c).

331  
332 Could the ATF be interpreted as a ML scar originating as a continent–continent collision in the  
333 Palaeozoic (Sobel and Arnaud, 1999) that controls intraplate deformation during periods of  
334 compression (with the most recent episode starting in the Oligocene resulting from the India–  
335 Eurasia collision)? Or is it that the ML scar is a result of crustal deformation impinging on the  
336 deeper lithosphere? The ability of deep lithospheric heterogeneous structures to exist over long  
337 periods in stable continental settings allows for a new mechanism for intraplate evolution  
338 (following external forcing). If, as an example, the ATF has a long-lasting ML scar from a  
339 continental collision that is controlling the crustal evolution, then plate tectonics may indeed  
340 display timeless (‘perennial’) processes (e.g., Heron et al., 2016a) with plate boundaries never  
341 really disappearing. As such, an increase in intraplate orogenesis would be observed during

342 future (and past) periods of global compression and extension (that is, supercontinent formation  
343 and dispersal).

344

345 However, deep inheritance as a source of intraplate deformation (and as a process within the  
346 Wilson Cycle as a whole) is not a closed subject. One reason for this is the ambiguity in the  
347 rheological properties of the scars “frozen” into the lithosphere. Schiffer et al. (2016) interpret  
348 mantle lithosphere scarring on the continental margin of East Greenland to be of higher density  
349 than the surrounding mantle material, with Petersen and Schiffer (2016) providing modelling on  
350 the topic. However, a number of studies have discussed the weakening impact of tectonic  
351 processes on the lithosphere to facilitate continental rifting (Dunbar and Sawyer, 1988, 1989).  
352 Furthermore, the subduction of crustal material into the mantle through ancient processes could  
353 increase volatiles to the lower lithosphere, weakening the seismically imaged scarred material  
354 (Pollack, 1986).

355

356 Aside from numerical modelling, the wider discussion on what we can ‘see’ in the mantle  
357 lithosphere and what we can infer from structures has been bolstered by a great number of  
358 seismic studies in recent years.

359

### 360 **Constraints from seismic studies**

361

362 Figure 6a shows examples of regions where mantle lithosphere heterogeneities (yellow circles)  
363 have been inferred, compiled from a previous map by Steer et al. (1998a) and updated to include  
364 more recent studies (e.g., Cook et al., 1999; van der Velden and Cook, 2005; Yang et al., 2003;



365 Hopper and Fischer, 2015; Kahraman et al., 2015; Schiffer et al., 2016). As discussed, the  
366 increase in high resolution imaging studies has increased the discovery of such structures in  
367 recent years. For an interpretation of the 2D geometry of the heterogeneities, Figure 6b gives an  
368 estimation of diagonal length of a mantle lithosphere scar (from a 2D horizontal and vertical  
369 component), with accompanying angle from the horizontal, for eight examples of mantle  
370 lithosphere heterogeneities (from Heron et al., 2016b). Below we outline a number of studies  
371 indicating an increased ‘visibility’ into the mantle lithosphere.

372

373 For example, the high-density seismometer array on the North Anatolian fault (NADA) showed  
374 horizontal structural variations in the crust and upper mantle on scales of 10 km and 20 km,  
375 respectively (Kahraman et al., 2015). Using USArray data, Hopper and Fischer (2015) applied  
376 converted wave imaging to the northern US craton to reveal mid-lithospheric discontinuities  
377 within the thick, high-velocity mantle. Their findings show that volatile rich layers could become  
378 ‘frozen into’ the mantle lithosphere as the lithosphere cools.

379

380 A clear link between plate tectonics, inheritance, and intraplate tectonics has been highlighted in  
381 Biryol et al. (2016), which presents new tomographic images of the south-eastern United States,  
382 revealing large-scale structural variations in the upper mantle. The origin of these structures is  
383 inferred to be a product of earlier episodes of continental collision and breakup, suggesting that  
384 the Wilson Cycle can generate long-lasting features within the mantle. Biryol et al. (2016) also  
385 discuss that plate strength and pre-existed inherited structures are important mechanisms that  
386 may be controlling ongoing tectonism in the region, as well as the multiple zones of seismicity.

387

388 The WOMBAT transportable seismic array in southeast Australia has imaged multiple  
389 lithospheric structures, as described in Rawlinson and Fishwick (2011). The mantle lithosphere is  
390 shown to have a wealth of features related to the geology and tectonic history of the region. The  
391 discovery of structures in certain areas related to lithospheric thinning, as well as Paleozoic  
392 provinces at depth in other regions, may have profound implications for the break-up of  
393 Australia and Antarctica. Furthermore, the use of new P and S wave tomography has been able to  
394 constrain upper mantle structures beneath southeast Canada and the northeast USA, a region  
395 spanning three quarters of Earth's geological history (Boyce et al., 2016). The ability to  
396 differentiate wave speeds within a medium to a finer degree has allowed for better understanding  
397 of how stable cratonic keels may have formed (Boyce et al., 2016), as new interpretations can be  
398 made on the processes that could cause lateral strength variations within the mantle lithosphere  
399 under North America (based on the tectonic history). It is the high-resolution illumination of the  
400 sub-crust (e.g., Rawlinson and Fishwick, 2011; Boyce et al., 2016) that can generate discussion  
401 on Wilson Cycle processes (continental break-up, craton stabilization) that were never possible  
402 in the past.

403

404 An abrupt seismic velocity wave speed transition in the mantle lithosphere from craton to  
405 Cordillera in western Canada was recently documented by Bao et al. (2014). This transition was  
406 interpreted to be related to the modification of the mantle lithosphere through Wilson Cycle  
407 dynamics, namely subduction zone interaction (Bao et al., 2014). Their discussion highlighted  
408 the possibility of small-scale convection initiated by a zone of weakness between the craton and  
409 the thickened lithospheric margin. Another recent important paper is the work of Dave et al.  
410 (2016), which presents a three-dimensional shear wave velocity model beneath the Wyoming

411 craton constrained from Rayleigh wave data. Their model provides the first seismic evidence for  
412 complex small-scale mantle convection beneath the Wyoming craton, with a high-velocity  
413 anomaly having a dripping shape in central Wyoming extending to 200 - 250 km depth  
414 (indicating mantle downwelling and lithosphere erosion).

415

416 Chamberlain et al. (2014) studied the San Andreas Fault and analysed the strain history of the  
417 upper mantle. Through the comparison of the long-term finite strain field in the mantle and the  
418 surface strain-rate field, respectively inferred from fast polarization directions of seismic phases  
419 (SKS and SKKS) and GPS data, Chamberlain et al. (2014) inferred that the San Andreas Fault  
420 extends to depth, likely through the entire lithosphere, with the possibility of the asthenosphere  
421 and tectonic plate being coupled. Asthenosphere mantle flow generating dynamic topography  
422 through vertical motions has also been investigated as a cause of lithosphere tectonics. Becker et  
423 al. (2014) highlighted western US intermountain seismicity as being caused by changes in upper  
424 mantle flow. The study inferred that mantle flow plays a significant and quantifiable part in  
425 shaping topography, tectonics, and seismic hazard within intraplate settings. If intraplate  
426 tectonics can be added into the Wilson Cycle dynamics, as we consider is sensible (e.g., Heron et  
427 al., 2016b), then the influence of the mantle lithosphere and convecting mantle on long-term and  
428 short-term tectonics is an important factor that is becoming clearer in recent years.

429

## 430 **Discussion and Conclusions**

431

432 In this review, we have outlined the current research on the role of the mantle lithosphere in  
433 causing tectonic deformation, alongside highlighting the potential of the deep lithosphere in

434 infiltrating every aspect of plate tectonics processes. As such an endeavour often leaves more  
435 questions than answers, we have compiled open questions on the role of the mantle lithosphere in  
436 the Wilson Cycle:

437

438 - How pervasive is localized deformation within the mantle lithosphere? For example, are  
439 deeps scars abundant, but just not imaged; or is the imaging fairly accurate in  
440 showing lithosphere that is less scarred than the upper crust?

441

442 - Are the structures that are 'visible' in the continental mantle lithosphere of large-scale  
443 tectonic importance? Do they indicate zones of weakness (e.g., (Bercovici and  
444 Ricard, 2014) or strength (e.g., Schiffer et al., 2016)? Can they be treated as pathways  
445 of future plate tectonic deformation?

446

447 - Do all Wilson Cycle continent collision and break-up events generate major mantle  
448 lithosphere scale structures (e.g., Biryol et al., 2016)?

449

450 - How can we differentiate among the causes of lithosphere scale deformation? For  
451 example, can we differentiate between mantle lithosphere structures caused by  
452 deformation originating in the crust and crustal deformation caused by reactivating  
453 mantle lithosphere structures?

454

- 455 - What is the role of isolated mantle volatiles being ‘frozen’ into the mantle lithosphere  
456 (e.g., Hopper and Fischer, 2015)? Are non-continuous zones of volatiles widespread  
457 across the whole of continental mantle lithosphere or simply localized features?  
458
- 459 - Is the large-scale rheological layering of the lithosphere more important in permitting the  
460 initiation of tectonic deformation than features within the lithosphere (e.g., scarring  
461 and inherited structures)? Or is it that lithosphere rheology and small features must be  
462 considered as a coupled system (e.g., Heron et al., 2016b)?  
463  
464

465 At the centre of these questions is the rheological make-up of the mantle lithosphere and the  
466 layering of the lithosphere as a whole (as discussed in the introductory section). Future work is  
467 required to constrain the strength layering within the continental lithosphere, and to what spatial  
468 extent such an environment can be applied.  
469

470 The introduction of intraplate deformation to the Wilson Cycle is something that we put forth  
471 here and in a previous manuscript (Heron et al., 2016b). We would argue that the Wilson Cycle  
472 should be expanded to include intracontinental tectonics. Furthermore, we would highlight the  
473 notion that plate boundaries may never truly disappear through inherited structures. A tenet of  
474 the conventional theory of plate tectonics (and indeed the Wilson Cycle) is that crustal  
475 deformation is confined to near the boundaries of plates. Recent work on inheritance implies that  
476 this remains true for general planetary deformation as ML scars (that can control tectonic  
477 evolution) in a continent interior may originate from ancient plate boundary deformation (e.g.,

478 Heron et al., 2016a). In this way, ancient and present-day plate boundaries could be represented  
479 together as latent and active boundaries. A global map of perennial plate tectonics (Figure 6)  
480 presents a redefined illustration of tectonic activity and modifies the conventional theory of plate  
481 tectonics (in keeping with the recent findings of Vauchez et al., (1997), Rawlinson and Fishwick  
482 (2011), Bercovici and Ricard (2014), Leng and Gurnis (2015), Dave et al. (2016), Boyce et al.  
483 (2016)).

484

485 Although images of the sub-crustal lithosphere are becoming more commonplace, there are areas  
486 where such studies are not possible due to accessibility and expense. An interesting alternative is  
487 the work of Flesch and Bendick (2012) who consider the relationship between surface  
488 kinematics and deformation of the whole lithosphere. Flesch and Bendick (2012) used 3-D  
489 numerical models to find a relationship between tectonics at the surface and deformation  
490 throughout the crust and mantle lithosphere, through changing the lithosphere strength profile  
491 (e.g., Figure 1). Their study found that where viscosity is both discontinuous and differs by much  
492 more than an order of magnitude between the upper crust and mantle lithosphere, information  
493 about both force balance and rheology are absent from the surface deformation. It is therefore  
494 difficult to estimate either the dynamic or mechanical state of the lithosphere through surface  
495 observations (Flesch and Bendick, 2012).

496

497 The use of numerical modelling will help to understand further the complex nature of mantle  
498 lithosphere scarring, and this, as well as the interaction with the crust above, may be better  
499 understood in three dimensions (e.g., Chen and Gerya, 2016). Numerical modelling of a  
500 lithosphere with a 'lasting memory', following on from the work of Bercovici and Ricard (2014)

501 (and others), will become more commonplace in plate tectonic studies in order to meet the  
502 requirement of inherited structures. If inherited structures are to evolve and dictate lithosphere  
503 evolution, then numerical models will need to model long timescales to take into consideration  
504 past dynamics in order to understand present and future evolution (e.g., Bercovici and Ricard,  
505 2014).

506

507 As the imaging of the lithosphere becomes clearer, the assumed strength profile of tectonic plates  
508 is becoming more complex (e.g., Figure 1). At the same time, the inherent strength of the  
509 structures within the mantle lithosphere is not well known. Work is required to fully understand  
510 the nature of the mantle lithosphere heterogeneities, as mantle lithosphere scarring has been  
511 interpreted to be either areas of weakness (e.g., Dunbar and Sawyer, 1988, 1989; Pollack, 1986;  
512 Bercovici and Ricard, 2014; Linckens et al., 2015; Heron et al., 2016) or strength (e.g., Schiffer  
513 et al., 2016; Boyce et al., 2016), which may alter the deformation evolution (e.g., Heron et al.,  
514 2015b). The integration of mantle geochemistry into studies of lithosphere deformation will be  
515 important in this discussion, in particular the evolution of grain damage over time (e.g.,  
516 Bercovici and Ricard, 2014). The link between grain-damage hysteresis and plate tectonic states  
517 may allow for a new analysis on how our planet may evolve differently to other terrestrial bodies  
518 (Bercovici and Ricard, 2016).

519

520 As body of evidence grows for the importance of the mantle lithosphere in plate tectonic  
521 processes (e.g., Vauchez et al., 1997; Holdsworth et al., 2001; Rawlinson and Fishwick, 2011;  
522 Bercovici and Ricard, 2014; Leng and Gurnis, 2015; Dave et al., 2016; Boyce et al., 2016; Heron  
523 et al., 2016a), it would be prudent for future work to consider the global and/or local aspect of

524 their discoveries. The interpretation of the role of the mantle lithosphere should be considered as  
525 such: is the fundamental rheological composition of the mantle lithosphere important on a global  
526 scale, or does the evolution of the lithosphere in a given area present specific examples of mantle  
527 lithosphere importance? This distinction between a globally applicable discovery and local  
528 evolution may be important in the analysis of the role of the mantle lithosphere in the Wilson  
529 Cycle.

530

531 The Wilson Cycle (Figure 2) describes the closure and opening of oceanic basins (e.g., Wilson,  
532 1966; Dewey and Burke, 1974), where continental margins are deformed and weakened over  
533 time. The geological and geophysical mechanisms within the Wilson Cycle encapsulate our  
534 conventional theory of plate tectonics, with structural inheritance in the tectonic plates playing a  
535 strong role in the evolution of the lithosphere (e.g., Holdsworth et al., 2001). Heron et al. (2016a)  
536 argue that if intraplate deformation can be linked to inherited structures from ancient plate  
537 tectonic events, then deformation within continental margins should also be part of a wider  
538 Wilson Cycle (Figure 2). Furthermore, the role of the mantle lithosphere as a source of pre-  
539 existing structures that could influence tectonics is coming to the forefront of tectonic dynamics  
540 (e.g., Vauchez et al., 1997; Holdsworth et al., 2001; Rawlinson and Fishwick, 2011; Bercovici  
541 and Ricard, 2014; Leng and Gurnis, 2015; Dave et al., 2016; Boyce et al., 2016; Heron et al.,  
542 2016a), as well the role of the deep lithosphere (and sub-lithosphere mantle) in surface tectonics  
543 (e.g., Chamberlain et al., 2014; Becker et al., 2015; VanderBeek et al., 2016). High-resolution  
544 seismic imaging surveys over the past decade has found heterogeneous structures within the  
545 mantle lithosphere to be somewhat ubiquitous (e.g., Rawlinson and Fishwick, 2011; Bastow et  
546 al., 2013; Schiffer et al., 2014, 2015, 2016; Schaeffer and Lebedev, 2014; Rasendra et al., 2014;



547 Bao et al., 2014; Kahraman et al., 2015; Hopper and Fischer, 2015; Tauzin et al., 2016; Park and  
548 Levin, 2016a, 2016b; Biryol et al., 2016; Boyce et al., 2016; Dave et al., 2016). There is a strong  
549 case for the importance of the mantle lithosphere in Wilson Cycle processes, through inherited  
550 structures, with an incentive to look deeper at how tectonic plates evolve.

551

552

553

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568

569

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1088 **FIGURE CAPTIONS**

1089 **Figure 1.** Schematic view of alternative first-order models of strength through continental  
1090 lithosphere (from Bürgmann and Dresen, 2008). In the upper crust, frictional strength increases  
1091 with pressure and depth. In the two left panels a coefficient of friction following Byerlee's law  
1092 and hydrostatic fluid pressure (ratio of pore pressure to lithostatic pressure  $\lambda = 0.4$ ) are assumed  
1093 in a strike-slip tectonic regime. In the right panel, low friction due to high pore fluid pressure ( $\lambda$   
1094 = 0.9) is assumed. (a) A jelly sandwich strength envelope is characterized by a weak mid-to-  
1095 lower crust and a strong mantle composed dominantly of dry olivine (Hirth and Kohlstedt, 2003).  
1096 (b) The crème brûlée model posits that the mantle is weak (in the case shown resulting from a  
1097 higher geotherm, adding water would produce a dramatic further strength reduction). The dry  
1098 and brittle crust defines the strength of the lithosphere. (c) The banana split model considers the  
1099 weakness of major crustal fault zones throughout the thickness of the lithosphere, caused by  
1100 various strain weakening and feedback processes. Owing to small grain size in shear zones,  
1101 deformation in the lower crust and upper mantle is assumed to be accommodated by linear  
1102 diffusion creep (grain size of 50  $\mu\text{m}$ ).

1103

1104 **Figure 2.** The Wilson Cycle with the additional tectonic feature of intraplate deformation.  
1105 Rifting (B), continental collision (D), and/or intraplate deformation (i) can leave lasting  
1106 impressions on the crust and mantle. The importance of inherited crustal and mantle structures in  
1107 influencing the tectonic pathway of deformation is shown by purple arrows. The figure shows  
1108 that it is difficult to unravel the cause and effect on the lithosphere of Wilson Cycle processes.  
1109 The references for the established pathway tectonic influence are as follows: [1] e.g., Holdsworth

1110 et al. (2001); Holdsworth (2004); Thomas (2006); [2] e.g., Royden and Keen (1980), Davis and  
1111 Kuszniir (2004), Buiter et al. (2009), and Péron-Pinvidic et al. (2013); [3] e.g., Vauchez et al.  
1112 (1997); [4] e.g., Flack and Warner (1990), Morgan et al. (1994), Lie and Husebye (1994),  
1113 Calvert et al. (1995), Calvert and Ludden (1999), Ghazian and Buiter (2013), and Schiffer et al.  
1114 (2014, 2016); [5] e.g., Tapponnier and Molnar (1975); [6] e.g., Dèzes et al. (2004), Avouac et al.  
1115 (1993), Cowgill et al. (2003), Tapponnier and Molnar (1975), and Kahraman et al. (2015); [7]  
1116 e.g., Stephenson et al. (2009); [8] e.g., Heron et al. (2016a). This figure is modified from Heron  
1117 et al. (2016b).

1118

1119 **Figure 3.** An example of a mantle reflection from Calvert et al. (1995). Line migration results of  
1120 the Abitibi-Opatoca survey (a) with interpreted results (b). The most prominent feature of the  
1121 data is the band of mantle reflections that dip in the north to northwest direction beneath the  
1122 Opatoca belt. The mantle reflections intersect the Moho beneath the Abitibi-Opatoca boundary  
1123 mapped at the surface (Calvert et al., 1995).

1124

1125 **Figure 4.** Overview of numerical modelling results into continental intraplate deformation  
1126 related to far-field compression in the presence of upper crust (UC), lower crust (LC), and  
1127 mantle lithosphere (ML) heterogeneities. The full numerical simulation is performed with  
1128 SOPALE across 600 km depth and 1500 km across. Rheological parameters are given in Heron  
1129 et al. (2016b), with compression applied at 1 cm/yr. (a) Positions of scars used in the numerical  
1130 study of Heron et al. (2016b). The scar length and angle are given in Figure 6b. The weak zones  
1131 (scars) in the UC and LC (as shown in white) and ML (red). Panels (b) – (e) show deformation  
1132 patterns related to a ‘jelly sandwich’ rheology similar to that of Figure 1a. Material deformation

1133 (top) and visualization of the second invariant of the deviatoric strain rate tensor (bottom) after  
1134 shortening for (b) model with UC scar only, (c) model with LC scar only, (d) model with all  
1135 scars, and (e) model with a ML scar only. Top 100 km of the models are shown in a 3X vertical  
1136 exaggeration. Models show that heterogeneities within the mantle lithosphere can control  
1137 tectonics over shallower features in strong mantle lithosphere settings. Panel (f) shows the  
1138 deformation of a continental interior for a crème brûlée (CB) lithosphere strength profile  
1139 (generated through a hot Moho temperature). (f) shows the mantle lithosphere scar playing no  
1140 role in deformation, highlighting the importance of lithosphere strength in tectonic evolution  
1141 (e.g., Figure 1).

1142

1143 **Figure 5.** The suture zones of Chinese tectonics and the Altyn Tagh Fault (ATF) (from Heron et  
1144 al. (2016a). (a) A topographic map of the different tectonic blocks with paleo-suture zones (white  
1145 lines) of the India–Eurasia collision zone (suture zones from Watson et al., 1987). CAO, B,  
1146 Central Asia Orogenic Belt; L, Lhasa block; Q, Qaidam Basin; QI, Qiantang block; SQ,  
1147 Songpan–Ganzi complex; TB, Tarim Basin. (b) Grey boxed region in (a) showing the ATF with  
1148 strike-slip faulting denoted in black, with thrust faulting in white (Cowgill et al., 2003). NAF,  
1149 North Altyn Fault. (c) Schematic seismic model of ATF (Wittlinger et al., 1998) from Zhang et  
1150 al. (2015). Red and green regions indicate the crust and mantle, respectively. Regions that are  
1151 more yellow or red in the model are low-velocity zones. Seismic line A to A0 is marked on b.  
1152 This region may represent an instance of a mantle lithosphere heterogeneity controlling intraplate  
1153 crustal deformation through far-field compressional forcing (e.g., Heron et al., 2016a).

1154

1155 **Figure 6.** (a) A perennial plate tectonic map showing examples of regions where mantle  
1156 lithosphere heterogeneities (yellow circles) have been inferred, compiled from a previous map by  
1157 Steer et al. (1998a) and more recent studies (Cook et al., 1999; van der Velden and Cook, 2005;  
1158 Yang et al., 2003; Hopper and Fischer, 2015; Kahraman et al., 2015; Schiffer et al., 2016),  
1159 alongside some possible paleo-plate boundary locations (yellow lines) (as modified from Holt et  
1160 al., 2015). (b) Estimation of mantle lithosphere scar length and angle from horizontal for eight  
1161 examples of mantle lithosphere heterogeneities (from Heron et al., 2016b).

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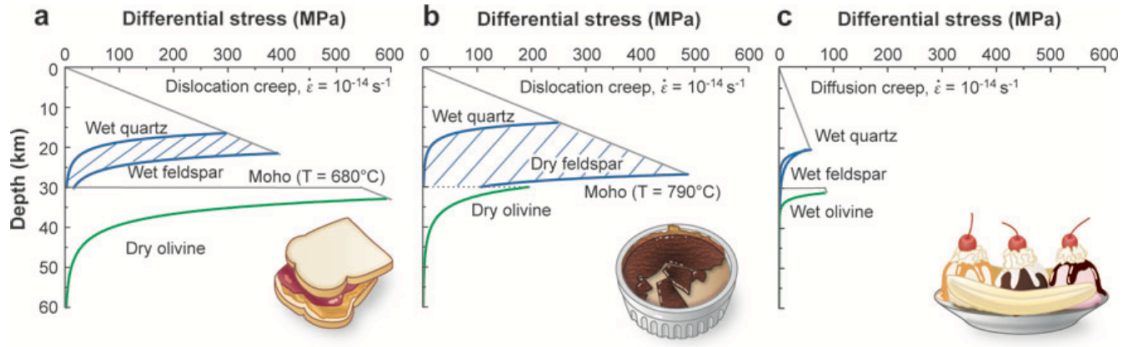
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1177 **Figure 1.**

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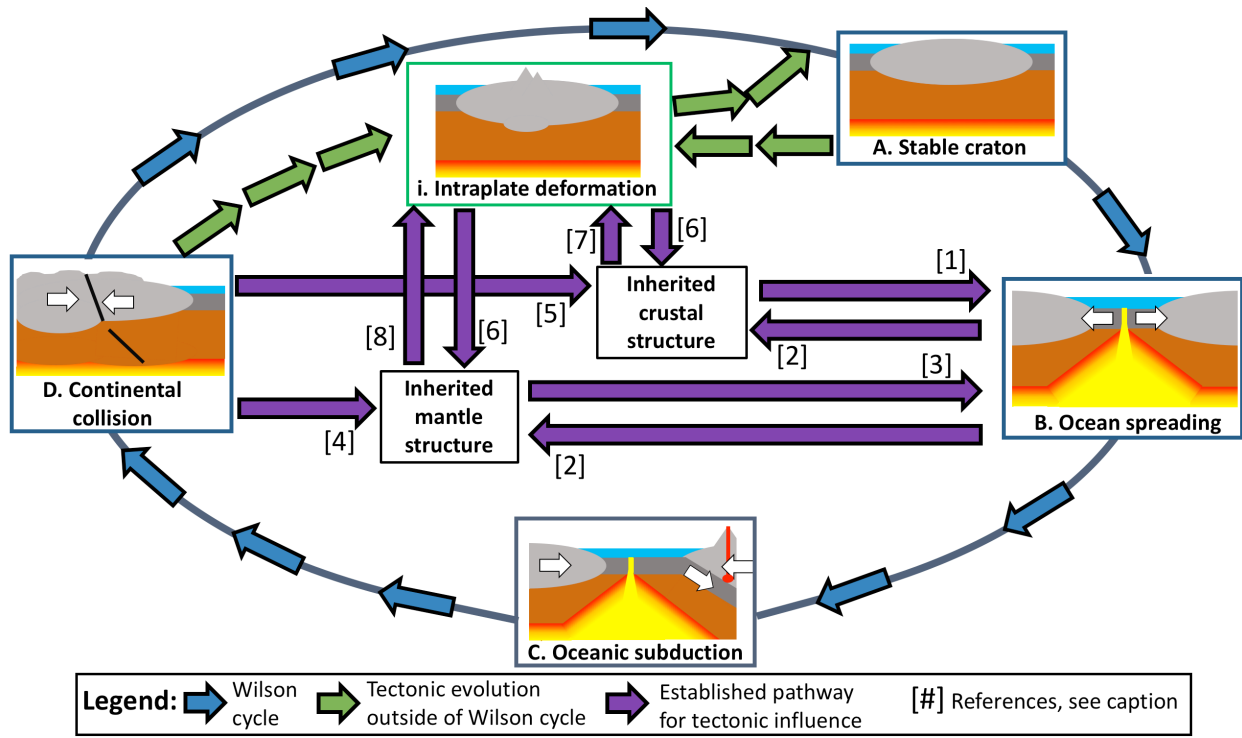
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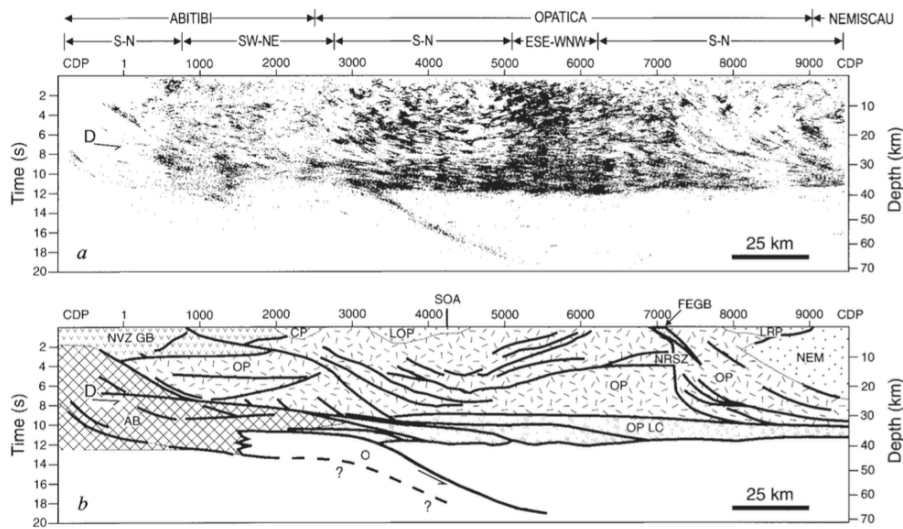
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1187 **Figure 2.**

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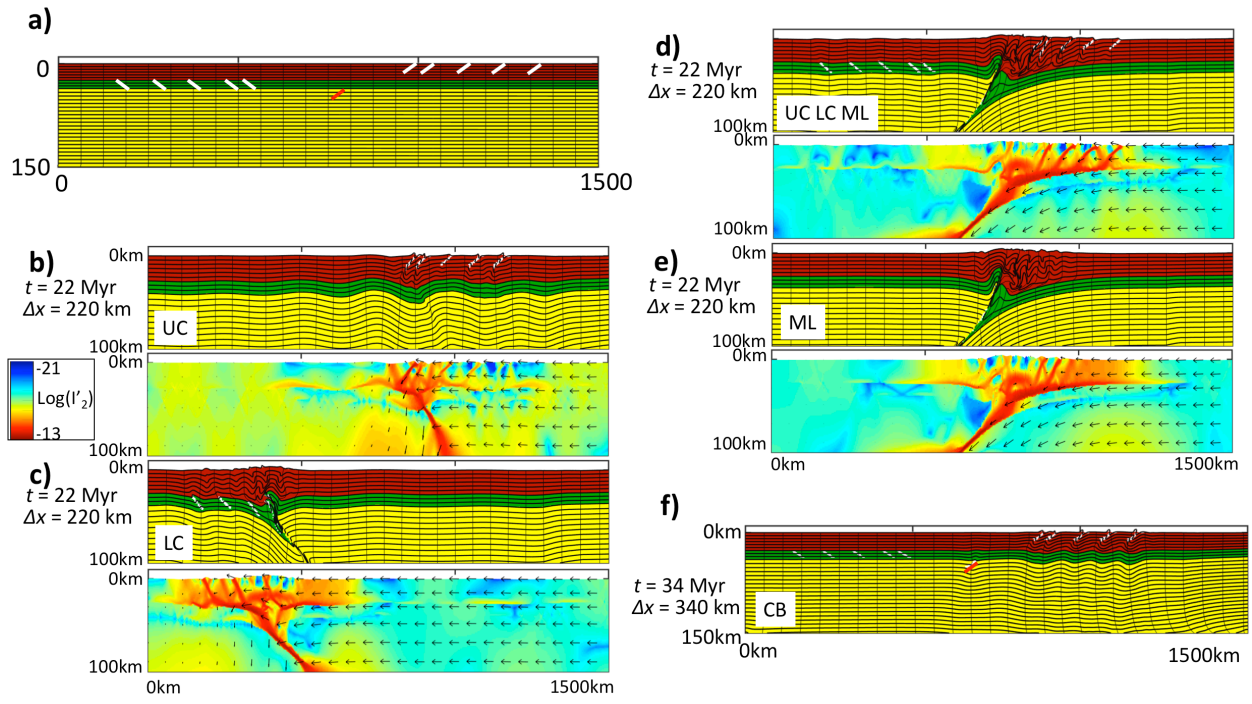


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1192 **Figure 3.**

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1196 **Figure 4.**

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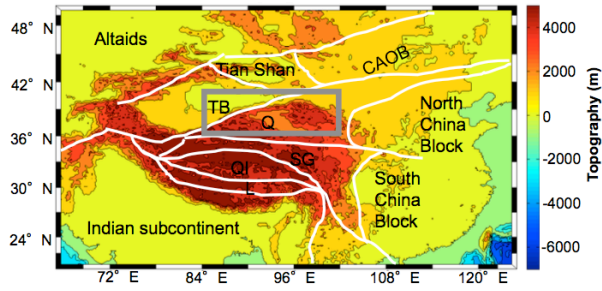
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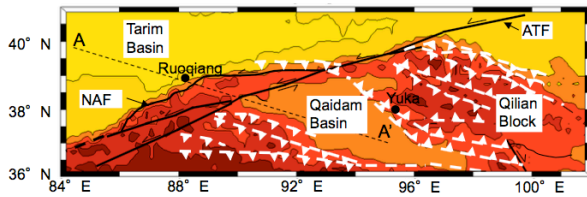
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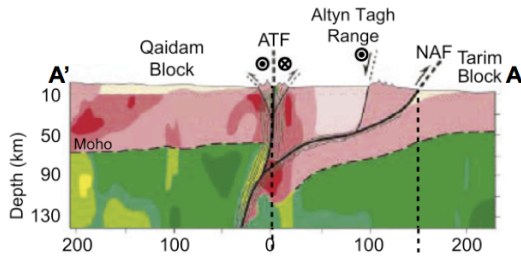
**a** India and Eurasia collision zone and ancient suture zones



**b** Altyn Tagh Fault (ATF)



**c** Seismic imaging of ATF (Wittlinger et al., 1998)



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1209 **Figure 5.**

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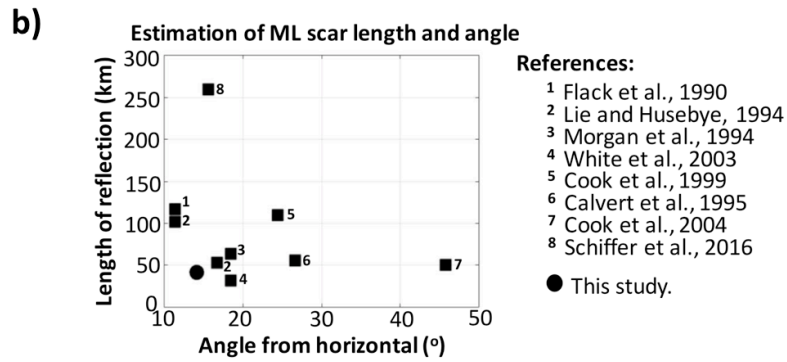
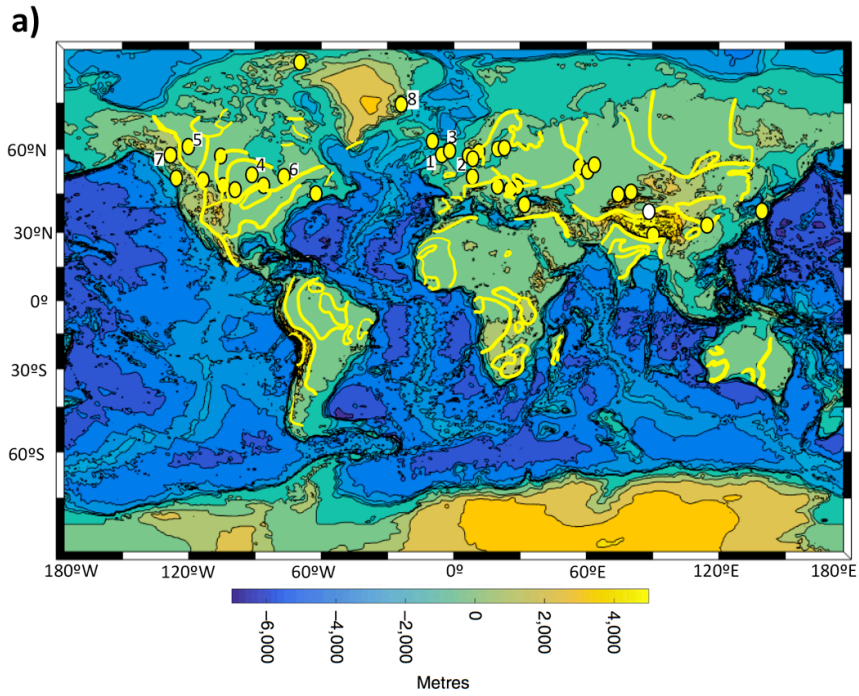
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1225 **Figure 6.**