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# New perspectives on ‘geological strain rates’ calculated from both naturally deformed and actively deforming rocks

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## Abstract

A value of  $\sim 10^{-14} \text{ s}^{-1}$  is commonly cited as an average geological strain rate. This value was first suggested for finite strain across an orogen, but based on more limited information than the combined geophysical, geological, and experimental data now available on active and ancient rock deformation. Thus, it is timely to review the data constraining strain rates in the continents, and to consider the quantifiable range of crustal strain rates. Here, where resolution allows, both spatial and temporal strain rate variations are explored. This review supports that a strain rate of  $10^{-14\pm 1} \text{ s}^{-1}$  arises from geological estimates of bulk finite strains. Microstructural arguments combining laboratory-derived piezometers and viscous flow laws, however, imply local rates that are orders of magnitude faster. Geodetic rates, in contrast, are typically  $\sim 10^{-15} \text{ s}^{-1}$  in actively deforming areas, about an order of magnitude slower than the bulk rates estimated from geological observations. This difference in estimated strain rates may arise from either low spatial resolution, or the fact that surface velocity fields can not capture strain localisation in the mid to lower crust. Integration of geological and geodetic

rates also shows that strain rates can vary in both space and time, over both single and multiple earthquake cycles. Overall, time-averaged geological strain rates are likely slower than the strain rates in faults and shear zones that traverse the crust or lithosphere.

*Keywords:* strain rate, rock deformation, geodesy, faults, shear zones

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## 1. Introduction

Pfiffner and Ramsay (1982) suggested a ‘conventional geological strain rate’ of  $10^{-14\pm 1} \text{ s}^{-1}$ . This estimate has been widely applied since the publication of their now classic paper, which was based on the finite strain record of orogenic belts. However, Pfiffner and Ramsay (1982) begin their article by stating that data on rates of natural rock deformation are rare. At the time of their writing, geodetic surveys of the San Andreas fault (Whitten, 1956) and measurements of glacial isostatic adjustment (Hicks and Shofnos, 1965) were the main sources of such data. Today, modern geodesy has hugely increased the data set on directly measured surface deformation. In addition, decades of rock deformation experiments and microstructural studies have led to new inferences regarding the mechanisms and rates of rock deformation based on the rock record. Collection and analysis of seismological data have also greatly increased knowledge of how this deformation is distributed in space and time. Huntington et al. (2018) raise the understanding of rheological variations through the lithosphere, for which strain rate distribution is a critical constraint, as a current Grand Challenge in tectonics research. This is therefore an appropriate time to revisit the outcrop record of rock deformation in light of new geodetic, seismic, and laboratory data, and to

20 discuss the calculation and interpretation of a ‘geological strain rate’. In  
21 particular, we consider the following three questions:

- 22 1. What is the observed, quantifiable, range of strain rates in nature?
- 23 2. How does strain rate vary in space, and to what degree is strain localised  
24 onto crustal-scale fault zones?
- 25 3. How does strain rate vary in time, not only through individual earth-  
26 quake cycles, but also across geological timescales?

27 We consider these questions from two distinct perspectives: first we dis-  
28 cuss continental strain over lengthscales greater than the lithospheric thick-  
29 ness and timescales of multiple earthquake cycles. We then consider how  
30 variations in strain with depth during different phases of the earthquake cy-  
31 cle (a) translate into surface strain and (b) are recorded within fault zone  
32 rocks.

## 33 **2. Definitions of Strain Rate**

34 Strain, and its derivative, strain rate, are formally described by a second  
35 order tensor, but for the purposes of discussion, we primarily use the scalar  
36 magnitude, which can be defined in a variety of ways. Longitudinal strain,  $e$ ,  
37 is the change in length of a linear element,  $\Delta l$ , divided by its original length  
38 prior to a discrete deformation episode,  $l_0$ . Alternatively, one may calculate  
39 natural strain,  $\epsilon$ , where strain is defined as having occurred over multiple  
40 infinitesimal increments, each deforming a linear element that includes all  
41 the previous deformation increments, i.e.  $\epsilon = \int_{l_0}^{l_f} \frac{dl}{l}$ , where  $l_f$  is the final  
42 length.

43 Shear strain rate in simple shear can be considered in terms of shear strain  
44 accumulated within an idealised shear zone of width,  $w$ , accommodating a  
45 finite displacement,  $d$ , parallel to its boundaries. In this case, shear strain is  
46 defined as  $\gamma = d/w$  and the shear strain rate is  $\dot{\gamma} = \gamma/t = s/w$  where  $s$  is the  
47 velocity difference across the shear zone. In simple shear, shear strain rate  
48 is therefore critically dependent on the deforming shear zone thickness (Fig.  
49 1a).

50 To express three dimensional strain, one can define principal strains as  
51 the longitudinal strains perpendicular to planes of zero shear strain. The  
52 strain ellipsoid represents strain relative to an originally undeformed sphere,  
53 and is defined by the principal strains  $X \geq Y \geq Z$ , where  $X = (1 + e_x)$  and  
54  $Z = (1 + e_z)$  represent the greatest and least stretch, respectively. Strain  
55 rate ( $\dot{\epsilon}$ ), is typically calculated by dividing longitudinal finite strain by the  
56 time taken to accumulate it. However, we note that Pfiffner and Ramsay  
57 (1982) explored the effect of strain path and found that among end-member  
58 strain histories and combinations thereof, pure shear is the most, and simple  
59 shear the least efficient at accumulating longitudinal strain after any given  
60 time period at a constant  $\dot{\epsilon}$ . Here, we will refer to  $\dot{\epsilon} = \dot{\epsilon}_x$  as the greatest  
61 longitudinal strain rate at a given location, comparable with what is typically  
62 measured in laboratory experiments, or shear strain rate,  $\dot{\gamma}$ .

### 63 **3. Crustal-scale strain over multiple earthquake cycles**

#### 64 *3.1. Geological Strain Rates*

65 Pfiffner and Ramsay (1982) arrived on a longitudinal, average, conven-  
66 tional geological strain rate of  $10^{-14} \text{ s}^{-1}$  by considering calculations of bulk

67 finite strain across orogens, a range of potential strain paths, and geochrono-  
68 logical constraints on the time taken to accumulate such strain. Updated  
69 constraints on such bulk strain accumulation rates have been obtained since.  
70 For example, in the Lachlan orogen, Australia, Foster and Gray (2007) esti-  
71 mate 67 % bulk shortening based on restored thrust sheets, and determine  
72 from  $^{40}\text{Ar}/^{39}\text{Ar}$  dating of white mica that deformation lasted approximately  
73 16 million years. This gives an average strain rate ( $\dot{\epsilon}$ ) on the order of  $10^{-15} \text{ s}^{-1}$   
74 assuming deformation was evenly distributed in space and time. The authors  
75 note, however, that deformation could have occurred in much shorter pulses,  
76 giving a bulk strain rate as fast as  $1 \times 10^{-14} \text{ s}^{-1}$ . These rates reflect bulk  
77 deformation within a km-scale volume of rock, but result from a combination  
78 of localised thrust displacements and distributed folding. The latter repre-  
79 sent zones of higher and lower strain, respectively, and thus record slower  
80 and faster strain rates embedded within the deformed volume (Fig. 1a).

81 Another approach to estimating strain rate in exhumed rocks is to infer  
82 paleostress from microstructures in viscously deformed rocks, constrain tem-  
83 perature of deformation through a geothermometer, and put resulting values  
84 into empirically derived flow laws to calculate strain rate. This methodology  
85 has the advantage of allowing spatial variations in strain rate to be explored.  
86 To this end, a number of authors have used quartz paleopiezometry to esti-  
87 mate stresses involved in quartz deformation by dislocation creep, based on  
88 the empirical relationship (Twiss, 1977):

$$\Delta\sigma = BD^{-p} \quad (1)$$

89 which relates steady-state differential stress,  $\Delta\sigma$ , to recrystallised grain size,

90  $D$ , through the empirical constants  $p$  and  $B$  that depend on the microscale  
91 dynamic recrystallisation mechanism. The steady state shear stress can then  
92 be related to the strain rate accommodated by dislocation creep through a  
93 flow law

$$\dot{\epsilon} = \Delta\sigma^n A \exp(-Q/RT) \quad (2)$$

94 where  $A$  is a material constant,  $Q$  is activation energy,  $T$  is temperature  
95 in Kelvin,  $R$  is the universal gas constant, and  $n$  is the stress exponent  
96 which depends on the active deformation mechanism. Assuming a constant  
97 temperature and steady flow at constant stress, strain rate can therefore be  
98 calculated from the recrystallised grain size by calculating flow stress in Eq.  
99 1 and extrapolating a laboratory flow law to this stress in Eq. 2.

100 This method takes advantage of advances in laboratory rock deformation  
101 experiments since the work of Pfiffner and Ramsay (1982), but involves un-  
102 certainties in extrapolating flow laws from laboratory to nature, estimating  
103 temperature of deformation to calculate strain rate from driving stress, in  
104 addition to the inherent error in the laboratory piezometer and flow law cal-  
105 ibrations. These uncertainties are difficult to quantify, but could exceed an  
106 order of magnitude in the final absolute strain rate estimate (cf. Hacker et al.,  
107 1990). To minimise the effect of absolute uncertainty on our conclusions, we  
108 will emphasise relative strain rate variations within a region. In the studies  
109 we discuss, the authors measured grain size in monominerallic domains to  
110 avoid grains whose growth was limited by pinning. However, in multiphase  
111 rocks there is additional uncertainty arising because grain size may deviate  
112 from the equilibrium state inferred by laboratory piezometer calibrations.

113 Gueydan et al. (2005) studied spatial variation in strain rate within the  
114 exhumed Tinos metamorphic core complex, Greece. They report recrystallised  
115 quartz grain size ranging from 160  $\mu\text{m}$  to about 40  $\mu\text{m}$  in distributed  
116 and localised ductile deformation zones respectively. Using the  
117 quartz piezometer of Stipp and Tullis (2003) and the dislocation creep flow  
118 law of Luan and Paterson (1992), these grain sizes imply ductile flow at  
119 strain rates of  $1.5 \times 10^{-15} \text{ s}^{-1}$  and  $2.6 \times 10^{-14} \text{ s}^{-1}$ , for penetrative and localised  
120 ductile flow, respectively (Gueydan et al., 2005). However, scatter in  
121 the data implies that within the penetrative ductile flow regime, local strain  
122 rate variations are over an order of magnitude faster and slower than the  
123 mean inferred strain rate, and within shear zones, strain rate may locally  
124 be close to  $10^{-13} \text{ s}^{-1}$  (Gueydan et al., 2005). Adjacent to the main brittle  
125 detachment, ductilely deformed quartz shows a strain rate increase to  
126  $2 \times 10^{-12} \text{ s}^{-1}$ .

127 Similarly, strain rates locally elevated to faster than  $10^{-14} \text{ s}^{-1}$  have been  
128 reported from mylonitic gneisses in extended middle crust in the Whipple  
129 Mountains, California (Hacker et al., 1992). Behr and Platt (2011), however,  
130 suggest that this local increase in strain rate is a result of progressive strain  
131 localisation during exhumation along the Whipple Mountain detachment.

132 Spatial variations in geologically determined strain rates have also been  
133 quantified in the Red River and Karakorum shear zones, which are strike-slip  
134 zones exhumed from the lower crust. Boutonnet et al. (2013) combined  
135 stress estimates from the quartz paleopiezometer of Shimizu (2008) and the  
136 laboratory-derived stress-strain rate relationship of Hirth et al. (2001) and  
137 calculated strain rates less than  $10^{-15} \text{ s}^{-1}$  in low strain areas, and greater

138 than  $10^{-13} \text{ s}^{-1}$  within localised high strain zones considered to have deformed  
139 at the same pressure-temperature conditions. The shear zones considered by  
140 Boutonnet et al. (2013) are a few kilometres wide, and represent a 1000-fold  
141 increase in shear strain rate relative to the surrounding low strain blocks.

142 In the exhumed mylonitic hanging wall of the transpressional Alpine  
143 Fault, New Zealand, finite shear strains of  $\leq 300$  were calculated from duc-  
144 tility deformed pegmatites within a kilometre-wide mylonite-ultramylonite  
145 zone (Norris and Cooper, 2003). To our knowledge, these are the largest  
146 shear strains directly calculated from rock exposures. The strain distribu-  
147 tion across the Alpine fault, as determined from deformed pegmatites, is best  
148 explained if lower crustal deformation along the Alpine fault is localised in  
149 a 1 - 2 km wide zone (Norris and Cooper, 2003), implying elevated strain  
150 rates where strain is localised in the lower crust, here as well as in Tinos,  
151 Karakorum and Red River (described above). Uplift on the Alpine fault  
152 occurred over the last 5 Ma (Suther, 1995), such that a total, integrated  
153 shear strain as high as 300 implies an average shear strain rate of at least  
154  $2 \times 10^{-12} \text{ s}^{-1}$  in localised zones. Based on paleopiezometry and Ti-in-quartz  
155 geothermometry, Cross et al. (2015) determined a strain rate range for Alpine  
156 fault zone mylonites deformed at 450-500°C, and preferred a value on the or-  
157 der of  $10^{-13} \text{ s}^{-1}$ .

158 The method and examples above rely on the rock record of dislocation  
159 creep in quartz. It is, however, likely that other mineral scale deformation  
160 mechanisms, such as diffusion creep, also accommodate significant strain  
161 rates in the mid- to lower crust. For example, as recrystallisation in high  
162 strain zones leads to grain size reduction, a transition from dislocation creep

163 to a grain-size sensitive flow mechanism can occur (e.g. Platt, 2015). The  
164 strain rate in shear zones accommodating flow by grain-size-sensitive creep  
165 cannot be directly obtained from a paleopiezometer, as the proportionality  
166 between stress and grain size no longer applies. However, for the strike-slip  
167 Pernambuco shear zone in Brazil, Viegas et al. (2016) identified deformed  
168 quartz ribbons and monomineralic quartz veins within a polyphase ultra-  
169 mylonite dominated by fine-grained feldspar. Based on microstructures and  
170 EBSD analyses, the authors infer the dominant deformation mechanism to  
171 be diffusion creep in feldspar, and dislocation creep in quartz ribbons. Vie-  
172 gas et al. (2016) therefore determined flow stresses from the quartz veins and  
173 ribbons, and through flow laws for dislocation creep in quartz and diffusion  
174 creep in feldspar estimated strain rates ranging from  $10^{-10} \text{ s}^{-1}$  to  $10^{-8} \text{ s}^{-1}$ .  
175 These estimates, if correct, imply at least local and transient increases in  
176 shear zone strain rate, accommodated by viscous mechanisms, to  $10^{-10} \text{ s}^{-1}$   
177 or greater.

178 We have now listed a number of examples where geological constraints  
179 indicate that strain is focused into relatively narrow zones. In most of these  
180 examples, the narrow zones are interpreted as established at mid- to lower  
181 crustal depths, but note that there are also examples where strain localisation  
182 results from progressive deformation during exhumation to lower tempera-  
183 tures and pressures in an extensional tectonic regime (Behr and Platt, 2011).  
184 On the crustal scale, localisation of strain into plate boundary zones weak-  
185 ened by grain size reduction, increased temperature, or elevated fluid content,  
186 was discussed by Bürgmann and Dresen (2008). These authors suggested the  
187 ‘banana split’ model for lateral strength reduction between stronger conti-

188 nental interiors; this model is consistent with the above-average strain rates  
189 locally recorded within the high strain zones described above.

### 190 *3.2. Geodetic strain rate estimates*

191 Whereas geological strain rate estimates are typically based on observa-  
192 tions of deformation accumulated over millions of years, geodetic techniques,  
193 such as GPS and InSAR, measure current and ongoing surface displacements.  
194 By considering the lithosphere to deform as a continuum, surface velocity es-  
195 timates can be used to calculate surface strain (e.g. Haines and Holt, 1993).  
196 This approach is valid when considering horizontal lengthscales several times  
197 the brittle, elastic thickness of the lithosphere, and also at shorter length-  
198 scales if faults are considered locked. The Global Strain Rate Map (GSRM  
199 v2.1), interpolates horizontal velocities from 18,000 GPS sites to calculate  
200 the 2nd invariant of the strain rate tensor  $\sqrt{\dot{\epsilon}_1^2 + \dot{\epsilon}_3^2}$  (Kreemer et al., 2014),  
201 equivalent to the maximum strain rate reported in the geological estimates  
202 previously discussed. The highest strain rates occur on narrow plate bound-  
203 aries, particularly at fast-spreading ridges where new crust is created, in  
204 which estimated strain rates are as high as  $1.4 \times 10^{-13} \text{ s}^{-1}$ . Figure 1b shows  
205 the distribution of strain rates within the nodes defined as deforming in  
206 GSRM 2.1, the majority of which lie in the range  $5 \times 10^{-17} - 10^{-14} \text{ s}^{-1}$ .  
207 Examining the distribution of strain rates shows that these values are an  
208 order of magnitude lower than the earlier geological estimates of  $10^{-14 \pm 1} \text{ s}^{-1}$   
209 (Pfiffner and Ramsay, 1982), but that the variance is very similar (Fig. 1b).

210 Roughly 5% of the area defined as deforming in GSRM 2.1 exhibits a  
211 strain rate exceeding  $10^{-14} \text{ s}^{-1}$ . These rates are concentrated in rapidly  
212 deforming zones with dense GPS networks such as the San Andreas fault

213 zone where GSRM reports strain rates exceeding  $10^{-14} \text{ s}^{-1}$  compared to  
214  $10^{-15} \text{ s}^{-1}$  or slower in the surrounding areas (Fig. 2a). These higher strain  
215 rate zones also correspond to areas of elevated seismic activity, attesting to  
216 localisation of deformation (Fig. 2b). However, comparison between the  
217 numerous strain models that have been produced for this well studied region  
218 demonstrates that the choice of interpolation scheme for GPS-derived models  
219 can lead to large near-fault discrepancies (Hearn et al., 2010). The inclusion  
220 of higher-resolution InSAR data is therefore critical to defining strain rates  
221 close to active structures (Fialko, 2006; Kaneko et al., 2013; Tong et al., 2013;  
222 Elliott et al., 2016). In particular, these InSAR data allow identification of  
223 structures that may accommodate locally higher strain rates (Elliott et al.,  
224 2016).

225 By approximating the lithosphere as a thin viscous sheet with vertically  
226 averaged forces and properties, continental-scale velocity fields can be used  
227 to investigate the rheology of the lithosphere (England and McKenzie, 1982).  
228 In such models, the horizontal gradients of the deviatoric stress associated  
229 with deformation are balanced by gradients of the gravitational potential  
230 energy (GPE). The models are capable of reproducing the first order patterns  
231 of deformation well, and typically return viscosities of  $10^{21} - 10^{22} \text{ Pas}$  for a  
232 viscous fluid with power law exponent  $n = 3$ , and strain rates up to  $10^{-15} \text{ s}^{-1}$   
233 (Table 1). The estimated average strain rate values are an order of magnitude  
234 lower than those derived by interpolating the velocity field, and averages  
235 from geological constraints, as the thin viscous sheet approach likely smooths  
236 out concentrations of strain over length-scales less than the thickness of the  
237 lithosphere. Some thin viscous sheet studies report large lateral variations in

238 rheological properties, for example, larger viscosities associated with semi-  
239 rigid microplates and lower values in rapidly deforming areas (Flesch et al.,  
240 2000, 2001). In other studies, however, such variations result in a negligible  
241 reduction in misfit compared to homogeneous models (England and Molnar,  
242 2015; Walters et al., 2017).

243 Because they vertically average rheological properties, thin viscous sheet  
244 models result in lower strain rates than obtained within models with vertical  
245 velocity gradients. Another end-member geodynamic model is the channel  
246 flow model, in which low viscosity channels accommodate high strain rate  
247 deformation driven by a lithostatic pressure gradient (Royden et al., 1997;  
248 Beaumont et al., 2001; Godin et al., 2006). This model has been invoked  
249 to explain both lack of shortening and presence of orogen-parallel extension  
250 within the Tibetan Plateau (Royden et al., 1997), and also a dynamic link  
251 between these two observations (Beaumont et al., 2001). Coupled to focused  
252 denudation (Beaumont et al., 2001), channel flow may lead to extrusion of  
253 mid-crustal rocks between bounding shear zones. Whereas the lower shear  
254 zone will be a thrust, the upper shear zone is either normal or reverse de-  
255 pending on the relative velocity of the channel versus its hanging wall (Godin  
256 et al., 2006, and references therein). A commonality for channel flow models  
257 is a low viscosity (typically  $\leq 10^{19}$  Pas, versus  $10^{21} - 10^{22}$  Pas typically re-  
258 turned by thin viscous sheet models) invoked based on weakening by partial  
259 melting under thickened crust (e.g. Jamieson et al., 2002). This local weak-  
260 ness will lead to higher strain rates than in depth-averaged thin viscous sheet  
261 models. For example, if channel thicknesses vary from 3 to 30 km (cf. Godin  
262 et al., 2006), and displacement is on the order of a centimeter per year, aver-

263 age  $\dot{\gamma}$  becomes  $10^{-14}$  to  $10^{-13}$  s $^{-1}$  (Fig. 1a). A range of geodynamic models  
264 employ strategies between the end member vertical strain rate average of the  
265 thin viscous sheet, and the significant vertical variation in strain rate of the  
266 channel flow model.

### 267 *3.3. Seismological strain rate estimates*

268 Whereas geodetic strain rates represent continuous deformation over some  
269 time period, seismic strain rates represent time-averaged slip along faults in  
270 earthquakes. By Kostrov summation (Kostrov, 1974; Jackson and McKenzie,  
271 1988), a seismic strain rate tensor can be obtained from earthquake moment  
272 tensors determined in a seismic volume over a given time period. Comparing  
273 geodetic and seismic strain rates allows comparison of aseismic and seismic  
274 deformation in a region. If seismic strain rates are low compared to geodetic  
275 strain rates, then either some deformation occurs aseismically, or the time of  
276 observation is shorter than the recurrence time of major earthquakes.

277 A comparison of seismic and aseismic strain rates for Iran, where the  
278 combined instrumental and historical earthquake catalogues go back over  
279 a millennium, has shown a large contrast in deformation style across the  
280 country (Masson et al., 2005). In Zagros, southern Iran, > 95% of strain  
281 is accommodated aseismically, although intensive microseismic activity is  
282 spatially correlated with this deformation. In contrast, northern Iran ex-  
283periences large earthquakes that account for 30 - 100% of the geodetically  
284determined strain. A reason for the largely aseismic strain accommodation in  
285southern Iran could be that a salt layer decouples an upper, 8 - 10 km thick,  
286aseismically deforming, sedimentary cover from underlying basement rocks,  
287leading to a thin seismogenic thickness (Jackson and McKenzie, 1988). In

288 northern Iran, few large earthquakes may accommodate the majority of the  
289 displacement because deformation occurs in characteristic earthquakes on a  
290 few, major strike-slip faults (Masson et al., 2005). Kreemer et al. (2002) have  
291 also argued that low seismicity rates, in regions of high geodetic strain rate  
292 along major strike-slip faults, can result from faults hosting few but large  
293 characteristic earthquakes. Such regions would lack small earthquakes rela-  
294 tive to predictions by a Gutenberg-Richter relationship (Wesnousky, 1994).

295 Although seismic strain rates may differ from geodetic and geological  
296 rates, they are particularly informative where other data are not available,  
297 such as for regions, depths, and time periods for which reliable geodetic data  
298 do not exist. Masson et al. (2005) found that although magnitudes of seismic  
299 and aseismic strain rates differ in places, orientations of principal strain axes  
300 are comparable. This observation was also made by Ekström and England  
301 (1989), who found that seismic strain rates were systematically smaller than  
302 expected from relative plate motions, but provided reliable estimates for the  
303 orientations of the principal horizontal strains. Therefore, summation of  
304 moment tensors may allow velocity fields to be calculated over time periods  
305 much longer than the geodetic record. For example, in deforming Asia the  
306 strain rate tensor based on instrumental and historical earthquakes show  
307 little difference from the velocity field indicated by paleomagnetic rotations in  
308 Cretaceous rocks (Holt and Haines, 1993). Furthermore, seismic strain rates  
309 can be estimated at depths where geodetic data are not available, and have  
310 for example been used to estimate a strain rate magnitude of  $\sim 1 \times 10^{-15} \text{ s}^{-1}$   
311 within slabs subducted to depths in excess of 75 km, implying significant  
312 internal deformation in these deeply subducted slabs of oceanic lithosphere

313 (Bevis, 1988; Holt, 1995).

#### 314 *3.4. Temporal Variations in Strain Rate*

315 Attempts to correlate decadal geodetic and seismic observations with  
316 much longer term geological estimates of strain rate have shed light on tem-  
317 poral strain rate variations at timescales of multiple seismic cycles. For ex-  
318 ample, tectonic reconstructions of the Hikurangi Margin, North Island, New  
319 Zealand, show approximately constant rates since 1.5 Ma (Nicol et al., 2007).  
320 These near-constant long-term rates are compatible with geodetic strain es-  
321 timates reflecting deformation in the last 10 - 15 years (Wallace et al., 2004).  
322 Thus, Nicol and Wallace (2007) concluded that on a million year timescale,  
323 strain rates can be essentially steady for a significant portion of the seismic  
324 cycle, with the corollary that GPS largely measures elastic strains that will  
325 be converted to permanent, localised deformation along faults in coseismic  
326 earthquake slip. Similar comparisons between decadal and million year strain  
327 rate estimates have been made elsewhere, including the Arabia-Eurasia col-  
328 lision zone (Allen et al., 2004), southwest United States (McCaffrey, 2005),  
329 and the Andes (Hindle et al., 2002). Like in New Zealand, these areas of well  
330 studied, regional crustal deformation show current geodetically determined  
331 strain rates within error of the geological strain rates estimated for the last  
332 few million years.

333 In contrast, the Tibetan Plateau has been an area of considerable contro-  
334 versy. Slip rates on major faults agree between geological and geodetic data;  
335 however, geomorphological data suggest more rapid motion over timescales  
336 of kyrs. Strain rate maps derived from InSAR and GPS demonstrate that  
337 at the present day, strain rates are relatively uniform within the Tibetan

338 Plateau at  $10^{-15} \text{ s}^{-1}$  (Wang and Wright, 2012; Garthwaite et al., 2013)(Fig.  
339 2c). Major Tibetan faults accumulate strain at rates generally less than  
340 1 cm/yr, resulting in near negligible increases in surface strain rate. In-  
341 terestingly, broad zones of slightly elevated strain rate are associated with  
342 faults that have experienced recent earthquakes (Wang and Wright, 2012;  
343 Garthwaite et al., 2013), for example the Kunlun fault (Garthwaite et al.,  
344 2013)(Fig. 2c). In addition, Daout et al. (2018) recently used InSAR data to  
345 highlight a wide zone of active strike-slip shear along the Jinsha suture, indi-  
346 cating reactivation of a lithospheric weakness that lacks expression of surface  
347 faulting. These observations highlight that long-term time-averaged strain  
348 rate estimates need to consider temporal variations within the earthquake cy-  
349 cle. Temporal strain rate variation is also seen in the Central Nevada Seismic  
350 Belt, where uplift detected by InSAR can be explained by postseismic mantle  
351 relaxation lasting several decades after major earthquakes (Gourmelen and  
352 Amelung, 2005).

353 Chatzaras et al. (2015) have provided a model for time-dependent inter-  
354 action between rheologically distinct mantle and crust. Their model is based  
355 on that low resolved shear stresses (less than 10 MPa) are recorded in both  
356 the frictional crust and viscous mantle of the San Andreas fault. They sug-  
357 gest an integrated crust-mantle system where distributed mantle deformation  
358 controls displacement, and loads the upper crust until its frictional failure  
359 strength is reached. This model implies that mantle deformation should ac-  
360 celerate as strain rate increases post-seismically, as seen for example after  
361 major earthquakes in southern California (Freed and Bürgmann, 2004), and  
362 that the next earthquake will occur where failure strength is first overcome

363 above a broad deforming zone in the mantle. Although designed for strike-  
364 slip faults (Chatzaras et al., 2015), this model may also explain the spatial  
365 and temporal strain rate variations cited above in collisional settings.

366 Geodetic strain rate estimates may be similar to strain rates inferred from  
367 the rock record of the last few million years of deformation. However, the  
368 geological records at several active zones of convergence show variation in the  
369 spatial distribution of strain rate on the multi-million year time scale. In the  
370 Himalayas, deformation can be interpreted to have gradually migrated onto  
371 the current locus at the orogenic front over a few tens of millions of years,  
372 as material accreted in the now  $> 100$  km wide zone of finite strain in the  
373 Himalayan arc (Fig. 2c)(Avouac, 2008). In the Central Andes, shortening  
374 currently accommodated by distributed strain in the foreland is faster than  
375 at 25 - 10 Ma, a time when convergence occurred at up to twice the cur-  
376 rent rate (Hindle et al., 2002). Hindle et al. (2002) interpreted this temporal  
377 change in strain rate partitioning to reflect a change in interseismic coupling,  
378 with convergence prior to 10 Ma dominantly accommodated by stable slid-  
379 ing localised along the megathrust, with little hanging wall shortening. This  
380 change from localised to distributed strain (and therefore strain rate) may  
381 reflect a change in the physical properties at the megathrust itself. Similarly,  
382 strain localised along many currently active faults in the Arabia-Eurasia col-  
383 lision zone occurs at strain rates that far exceed those calculated from their  
384 finite strain over the life time of the orogen (Allen et al., 2004). Allen et al.  
385 (2004) explain that currently active faults, located in areas of low elevation  
386 at the edges of the collision zone, initiated or took up increasing amounts of  
387 strain after 7 Ma. In earlier stages of collision, deformation occurred in what

388 is now uplifted regions with thickened crust. Similarly, shortening across the  
389 Himalayan mountain range does not occur on the high Tibetan Plateau, but  
390 has localised to the Main Himalayan Thrust Zone at the orogenic front in  
391 Nepal (Fig. 2c,d), for at least the last 20 Ma (Bilham et al., 1997; Bollinger  
392 et al., 2006; Avouac, 2008). These examples show that partitioning of defor-  
393 mation varies in time and space as convergent and collisional margins evolve,  
394 with deformation either slowing or accelerating in a given zone over time.  
395 Thus, a particular strain rate field is unlikely to be maintained for more than  
396 a few million years, substantially less than the lifetime of an orogen. Conse-  
397 quently, a bulk strain rate calculated from finite geological strain across an  
398 orogenic belt will not represent local, temporal strain rates that may control  
399 the bulk rheology at a given period of time.

#### 400 **4. Strain within and around faults**

401 The earthquake cycle includes high strain rate slip that lasts from seconds  
402 to minutes, associated with brittle failure of the upper, elastic layer, followed  
403 by slower postseismic transient creep that decays towards steady-state in-  
404 terseismic deformation rates driven by viscous creep at depth (e.g. Hetland  
405 and Hager, 2005; Handy et al., 2007; Wang et al., 2012). Postseismic tran-  
406 sients are attributed to viscoelastic relaxation of the lower crust and/or upper  
407 mantle, and/or afterslip caused by creep within the brittle fault zone (e.g.  
408 Wright et al., 2013). Variations in strain rates through the earthquake cycle  
409 are recorded as mutually crosscutting relationships between pseudotachylite  
410 and mylonites in the rock record (Fig. 3a)(e.g. Sibson, 1980a; Price et al.,  
411 2012; Menegon et al., 2017), and maybe also by mutually cross-cutting con-

412 tinuous and discontinuous deformation structures (Fig. 3b)(Fagereng and  
413 Sibson, 2010; Rowe and Griffith, 2015). It is possible, maybe even likely,  
414 that peak strain rates derived from quartz paleopiezometry (e.g. Boutonnet  
415 et al., 2013; Viegas et al., 2016) could be related to post-seismic afterslip. In  
416 the following section, we review strain rates associated with the earthquake  
417 cycle on individual fault zones from both geodetic and geological perspec-  
418 tives, since both records agree that strain rate is not constant in time.

#### 419 *4.1. Surface deformation during the interseismic period*

420 Geodetic observations record surface strain, and hence underestimate  
421 strain rates generated in the deep portions of fault zones. To illustrate,  
422 Savage and Burford (1973)'s widely used model of interseismic strain accu-  
423 mulation shows that surface velocity,  $u$ , at a distance  $x$  caused by slip rate  
424 of  $s$  on an infinitely long vertical, strike-slip fault with a locked elastic lid of  
425 thickness  $d$  is given by  $u(x) = \frac{s}{\pi} \arctan \frac{x}{d}$ . The shear strain rate is given by  
426 the derivative, such that  $\dot{\gamma}(x) = \frac{s}{\pi d} \frac{1}{(1+x^2/d^2)}$ , and the peak strain rate mea-  
427 sured at the surface,  $\dot{\gamma}_{max} = \frac{s}{\pi d}$ , depends not only on the slip rate across the  
428 fault, but also the locking depth. Thus a slip rate of 1 cm/yr with a locking  
429 depth of 20 km would produce a peak surface strain rate of  $5 \times 10^{-15} s^{-1}$ ,  
430 but  $2 \times 10^{-14} s^{-1}$  for a locking depth of 5 km (Fig 4).

431 Thus surface strain rate alone is not a direct indicator of strain rates  
432 within a fault zone itself. Locking depth must also be considered when inter-  
433 preting geodetic strain measurements. Locking depth is considered broadly  
434 equivalent to the frictional-viscous transition, and across the continents typi-  
435 cally lies within a range of  $14 \pm 7$  km (Wright et al., 2013). In contrast to  
436 oceanic crust, where locking depth varies smoothly as a function of temper-

437 ature, variations in continental locking depth do not correlate strongly with  
438 variations in crustal thickness, and it has therefore been suggested that vari-  
439 ations in lithology and strain rate can be responsible (Wright et al., 2013).  
440 However, heat flow also varies significantly throughout continents, partic-  
441 ularly as a function of tectonic regime, and long wavelength variations in  
442 thermal structure has successfully explained much of the depth variations  
443 in the seismologically determined locking depth (e.g. Sibson, 1984; Tse and  
444 Rice, 1986; McKenzie et al., 2005). Maggi et al. (2000) reviewed variations  
445 in earthquake focal depths, and suggested close correlation between elastic  
446 and seismogenic thickness, consistent with a first order dependence of lock-  
447 ing depth on temperature, and secondary variations caused by lithology and  
448 fluid content.

449 Relatively few faults exhibit creeping behaviour, with slip extending all  
450 the way to the surface (Burford and Harsh, 1980; Lee et al., 2001; Harris,  
451 2017). We expect the greatest rates of geodetic surface strain to be associated  
452 with these creeping faults. For example, the maximum rate of surface strain  
453 in California occurs on the creeping segment of the San Andreas fault, where  
454 slip rates up to 28 mm/yr generate surface strain rates that locally reach  
455  $2 \times 10^{-13} \text{ s}^{-1}$  (Tong et al., 2013)(Fig. 2a). Deformation associated with  
456 fluid flow within weakened fault rocks may well enhance shallow strain rate  
457 values, however, through alteration to frictionally weak minerals, or local  
458 elevation in fluid pressures (Rice, 1992; Wintsch et al., 1995). Ingleby and  
459 Wright (2017) have suggested that Omori-like decay of postseismic velocities  
460 is consistent with rate-and-state friction or power law shear zone models,  
461 implying that postseismic creep is also localised within a narrow tabular

462 zone. The fact that localised shear strain rate at depth is not fully recorded  
463 in the broad deformation field generated at the surface, may explain the order  
464 of magnitude difference between the Global Strain Rate Map (Kreemer et al.,  
465 2014), which considers the surface strain during interseismic periods, and the  
466 geological estimates of Pfiffner and Ramsay (1982), which consider the total  
467 intergrated strain.

#### 468 *4.2. Postseismic surface deformation*

469 Elevated rates of surface deformation have been detected following more  
470 than 20 earthquake sequences (Wright et al., 2013). Models of the earthquake  
471 cycle show that viscous postseismic transients occur when the earthquake  
472 return period is much longer than the relaxation time (Savage and Prescott,  
473 1978; Hetland and Hager, 2005). Models typically require Maxwell viscosities  
474 in the range  $10^{17} - 7 \times 10^{19}$  Pas to fit observational strain data (Wright et al.,  
475 2013), but the associated changes in velocity are on the order of mm/yr  
476 and occur over wavelengths of tens of kilometers, so the associated surface  
477 strain rates rarely exceed  $10^{-15} \text{ s}^{-1}$  (e.g. Wang and Wright, 2012). As argued  
478 above, however, even slightly elevated surface strain rate could translate into  
479 a much greater increase in subsurface strain rate if it reflected postseismic  
480 strain localised along the deep extension of crustal faults.

481 Afterslip within the brittle fault zone can amount to a significant portion  
482 of the coseismic slip and produce surface displacements (e.g. Reilinger et al.,  
483 2000; Lee et al., 2006; D'Agostino et al., 2012). Afterslip is associated with  
484 velocity-strengthening frictional properties and attempts have been made  
485 to model it with rate-and-state friction (e.g. Perfettini and Avouac, 2007).  
486 However, high resolution GPS and InSAR studies show short wavelength (less

487 than a few km) variations in afterslip that can only be attributed to along-  
488 strike variations in frictional properties that possibly relate to differences in  
489 lithology (Barbot et al., 2009; Floyd et al., 2016). Because fault geometry  
490 and material properties at depth cannot be determined from observations of  
491 surface deformation patterns alone, we return to the geological data set to  
492 discuss strain accommodation within localised structures.

#### 493 *4.3. Shear Strain within Fault Zones*

494 Geodetic models of strain accumulation cannot distinguish between slip  
495 on a single dislocation and that in a wider, tabular shear zone. Thus, esti-  
496 mates of strain rate within fault zones rely on geological observations of fault  
497 zone structure and dimensions. Sibson (2003) argued that the coseismic slip  
498 zone is commonly  $< 10$  cm, so that the  $\dot{\gamma}$  for seismic slip rates of 1 m/s  
499 becomes  $\geq 10$  s<sup>-1</sup>, assuming the coseismic slip zone behaves as a contin-  
500 uum (Fig. 1a). Such localised principal slip zones, commonly embedded in  
501 wider damage zones, are typical of faults in crystalline rocks, as described by  
502 Chester and Logan (1987) for the Punchbowl fault, and also seen in several  
503 other continental faults (Fig. 3c). In contrast, Burford and Harsh (1980) re-  
504 ported that aseismic distortion along a creeping segment of the San Andreas  
505 fault is accommodated within simple shear zones up to 15 metres wide. In  
506 these zones, taking the creep rate as 10s of millimetres per year (e.g. Titus  
507 et al., 2006),  $\dot{\gamma}$  can be approximated to an order of magnitude as 10<sup>-3</sup> yr<sup>-1</sup>  
508 or 10<sup>-11</sup> s<sup>-1</sup> (Fig. 1a), which is orders of magnitude faster than peak surface  
509 strain rates estimated at the resolution of the GSRM (Fig. 2a). While creep-  
510 ing faults in the upper crust are relatively unusual (Harris, 2017), mid- to  
511 lower crustal mylonites are typically inferred to accommodate steady creep,

512 or transient afterslip, over thicknesses of metres to kilometres. These shear  
513 zone widths imply strain rates ranging from  $10^{-10} \text{ s}^{-1}$  to  $10^{-14} \text{ s}^{-1}$  if slip rates  
514 are 1 - 10 mm/yr for shear zone width of 1 to 1000 m. Paleopiezometry re-  
515 sults obtained from monomineralic quartz layers in viscous shear zones reflect  
516 strain rates in this range (Fig. 1a)(Gueydan et al., 2005; Boutonnet et al.,  
517 2013; Cross et al., 2015). Although some mylonites record relatively homo-  
518 geneous strain (Fig. 3d), others have accumulated heterogeneous strain (Fig.  
519 3e), implying variable degrees of localisation, which by our logic implies het-  
520 erogeneous strain rate. An end-member example of such heterogeneity may  
521 be the discrete discontinuities observed within a zone of continuous defor-  
522 mation structures in *mélange* shear zones (Fagereng and Sibson, 2010; Ujiie  
523 et al., 2018)(Fig. 3f). In such *mélanges*, deformation occurs both in mm-  
524 cm wide principal slip zones, and distributed through matrix material over  
525 metres to hundreds of metres (Rowe et al., 2013). Thus, overall, localised  
526 deformation within high strain zones, which could be either steady or tran-  
527 sient, appears to occur at rates that range from  $< 10^{-10} \text{ s}^{-1}$  to  $> 10 \text{ s}^{-1}$ .  
528 Strain rates may be partitioned between individual, relatively homogeneous  
529 structures of different widths (Fig. 3c,d), or within a single, heterogeneous  
530 zone with variable degrees of strain localization (Fig. 3e,f).

531 We know that major shear zones typically contain thinner, anastomosing  
532 ultramylonites separating less deformed protomylonite to mylonite domains  
533 (e.g. Coward, 1990; Carreras, 2001; Rennie et al., 2013), meaning that strain  
534 rates within kilometre-scale shear zones are likely higher than the minimum  
535 estimated for their bulk. Evidence of strain localization, coupled with geo-  
536 metrical arguments of associated strain rate distribution over many orders

537 of magnitude (Fig. 1a), raise the question of how representative an average  
538 strain rate of  $10^{-14} \text{ s}^{-1}$  is in space. This point is emphasised by the range of  
539 strain rates inferred from calculations based on paleopiezometry (e.g. Guey-  
540 dan et al., 2005)(Fig. 1a).

541 An additional set of field observations is how structures crosscut each  
542 other. Pseudotachylytes, ‘fossilised’ and variably crystallised friction melt  
543 interpreted as unequivocal evidence for earthquake slip (cf. Cowan, 1999),  
544 are reported both crosscutting and locally overprinted by mylonitic fabric in  
545 a range of tectonic settings (Sibson, 1980b; Price et al., 2012; White, 2012;  
546 Menegon et al., 2017)(Fig. 3a). This mutually crosscutting relationship  
547 implies a strain rate cycling between spatially distributed, but temporally  
548 steady or transient, viscous flow in the mylonite, likely at  $\dot{\gamma} \leq 10^{-10} \text{ s}^{-1}$ , and  
549 seismic slip at rates exceeding  $10 \text{ s}^{-1}$ . Examples of this strain rate cycling  
550 are particularly abundant in places where shear zones were active within  
551 relatively dry, strong, middle to lower crust (Sibson, 1980b; Menegon et al.,  
552 2017; Hawemann et al., 2018).

553 Recently, Rowe and Griffith (2015) noted evidence for several other in-  
554 dicators, in the rock record, of frictional heating to temperatures too low  
555 to produce melting, but which also imply dynamic, elevated strain rates.  
556 Similarly, other mutually crosscutting structures implying different degrees  
557 of strain localisation, such as hydrothermal veins and synmetamorphic fo-  
558 liations in subduction-related thrust-sense *mélange* shear zones (Fig. 3b),  
559 may also reflect cycling between relatively steady and dynamic strain rates  
560 (Fagereng et al., 2011, 2018; Ujiie et al., 2018). Such temporal variations are  
561 not captured by bulk strain rate estimates.

562 A note of caution on when and where to invoke strain localisation, how-  
563 ever, is raised from observations of distributed strain in lower crustal and  
564 upper mantle rocks that lack signs of local high strain domains but record  
565 low differential stresses. For example, olivine grain size paleopiezometry in  
566 mantle xenoliths from the San Andreas transform fault system implies that  
567 increased mantle strain rates following crustal earthquakes can be accommo-  
568 dated by viscous dissipation of stress across a deforming zone much wider  
569 than in the overlying crust (Chatzaras et al., 2015). In another continental  
570 transform system, the Marlborough fault system of New Zealand’s South Is-  
571 land, lack of Moho displacement and pervasive seismic anisotropy below the  
572 faulted upper crust has also been interpreted to show strain distributed over  
573 a wide zone in the lower crust and upper mantle (Wilson et al., 2004).

574 Handy et al. (2007) reviewed the structure of continental faults below  
575 the transition from dominantly frictional deformation in the upper crust to  
576 dominantly thermally activated viscous deformation in the lower crust and  
577 upper mantle. They make the point that the structure and rheology of faults  
578 and shear zones depends on their strain and thermal histories. Pennacchioni  
579 and Mancktelow (2018) make the case that geometry of small scale shear  
580 zones is pre-determined by precursor heterogeneities such as fractures or low  
581 viscosity compositional layers. However, over time, additional mechanisms  
582 to develop and grow weak zones in the lower crust include networking of  
583 shear zones with increasing strain (Handy, 1994) and reaction weakening  
584 with increasing fluid-rock interaction (Wintsch et al., 1995). Handy et al.  
585 (2007) raise examples of faults that show fast post-seismic deformation that  
586 is well fitted to a localised low viscosity zone in the lower crust, such as the

587 North Anatolian transform fault of Turkey (Bürgmann et al., 2002) and the  
588 Chelungpu thrust fault in Taiwan (Hsu et al., 2002), and contrast these with  
589 faults where only minor surface displacement is recorded after major earth-  
590 quakes, including the 2001 Bhuj intraplate thrust event in India (Jade et al.,  
591 2002). In summary, it is likely that strain localisation in the lower crust re-  
592 quires some long-term thermal and/or kinematic weakening effects, although  
593 it is also promoted by stress increases down-dip of major earthquakes (Ellis  
594 and Stöckhert, 2004).

## 595 **5. Spatiotemporal strain rate distribution and average strain rate**

596 Overall, the observations we have collated show that where strain is not  
597 localised, strain rates are commonly  $10^{-15} \text{ s}^{-1}$  or slower, particularly if av-  
598 eraged over multiple earthquake cycles. Higher strain zones, in contrast,  
599 typically record strain rates of  $10^{-14} \text{ s}^{-1}$  or greater. Strain rates in high  
600 strain zones are likely underestimated, particularly where they are calculated  
601 from geodetic data. There are at least two reasons for this: (1) the spatial  
602 resolution of the data is not sufficient to identify high strain zones within  
603 anastomosing networks, which are known to exist from geological maps of  
604 shear zones (e.g. Carreras, 2001; Rennie et al., 2013); and (2) except along  
605 faults that creep steadily at the surface, surface strain rates underestimate  
606 strain rates on localised structures at depth (Fig. 4). We therefore highlight  
607 a need for care when comparing strain rates determined from geodetic data  
608 to those estimated from geological observations of rocks deformed at depth.

609 A picture arises of high strain zones accommodating strain rates faster  
610 than an average near  $10^{-14} \text{ s}^{-1}$ , separating lower strain blocks where transient

611 strain rate increases may occur, but average strain rate is less than  $10^{-14} \text{ s}^{-1}$ .  
612 Strain rate estimates based on a combination of microstructural observations  
613 and empirical stress-grain size and stress-strain rate relationships imply that  
614 the maximum strain rate within viscous high strain zones is in the range  
615 of  $10^{-13} \text{ s}^{-1}$  to  $10^{-8} \text{ s}^{-1}$  (e.g. Gueydan et al., 2005; Boutonnet et al., 2013;  
616 Viegas et al., 2016). Thus, while  $10^{-14} \text{ s}^{-1}$  may be a good estimate for the  
617 *time-averaged* bulk strain rate in an orogen, it does not represent the range  
618 of strain rates evidenced by the rock record. Low strain areas record slower  
619 strain rates. In contrast, localised high strain zones that are active for limited  
620 amounts of time accommodate strain rates higher than average (Fig. 1a).

621 On time scales comparable to the seismic cycle, seismological and geodetic  
622 networks in well instrumented, actively deforming areas record a spectrum of  
623 deformation rates (e.g. Peng and Gomberg, 2010). This spectrum ranges from  
624 plate tectonic displacement rates of mm/yr to earthquakes of m/s, through  
625 geodetically detected ‘slow slip’ of cm/week, to very low and low frequency  
626 earthquakes defined as seismic phenomena, with slip speeds slower than 1 m/s  
627 but sufficient to radiate seismic wave energy. Thus, in contrast to a paradigm  
628 where slip speeds are either steady or seismic, a range of values are allowed  
629 by the observations. This raises a question when interpreting strain rates  
630 that are elevated relative to a global average. Do they record steady viscous  
631 creep, transient slow slip, or post-seismic afterslip within a narrow zone or  
632 zones? This is a question to consider in future high resolution geophysical  
633 experiments, and highlights the point that strain rates are constant in neither  
634 space nor time.

635 In essence, any calculation of mid- to lower crustal rheology over multiple

636 earthquake cycles requires an estimate of strain rate. Pfiffner and Ramsay  
637 (1982)'s estimate of  $10^{-14} \text{ s}^{-1}$  is reasonable as a time averaged, bulk strain  
638 rate. However, strain rate is not steady in either time or space as the locus  
639 of deformation shifts in both time and space. The spatiotemporal variation  
640 in strain rate may, intriguingly, reflect changes in rheology with progressive  
641 strain. Another question with scope for additional future study is therefore  
642 what controls spatiotemporal variations in strain rate, particularly where  
643 geological and geodetic strain rates disagree, as in the India-Eurasia colli-  
644 sion zones (Wang and Wright, 2012; Garthwaite et al., 2013) and the Andes  
645 (Hindle et al., 2002).

## 646 **6. Conclusion and consequences**

647 High strain zones that traverse the lithosphere, which accommodate the  
648 bulk of continental deformation at any one time, typically deform at local  
649 and transient rates exceeding both the  $10^{-14} \text{ s}^{-1}$  estimated from bulk geo-  
650 logical reconstructions (Pfiffner and Ramsay, 1982), and absolute rates esti-  
651 mated from geodetically determined surface velocity fields (Kreemer et al.,  
652 2014). Two consequences of this conclusion are: (1) if higher strain rates  
653 are inserted in crustal strength curves, this implies either higher stresses  
654 or lower strengths within high strain zones, relative to predictions using a  
655  $10^{-14} \text{ s}^{-1}$  strain rate; and (2) in cases of spatiotemporal strain rate varia-  
656 tions on timescales of the earthquake cycle, there is a need for care in using  
657 time-averaged strain rates in estimating earthquake repeat times. The first  
658 of these consequences supports Bürgmann and Dresen (2008)'s banana split  
659 model for lithospheric strength distribution, with lateral strength and strain

660 gradients around weak, high strain, plate boundary zones.

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1028 **Figure Captions**

1029 Figure 1: Examples of geologically estimated and geodetically calculated  
1030 shear strain rates. a) Shear strain rate as a function of lengthscale, contoured  
1031 for displacement rate in ideal simple shear. See text for details, and note that  
1032 ellipses represent typical ranges but exceptions may occur. Note logarithmic  
1033 axes, and that localisation of strain in zones thinner than one kilometre im-  
1034 plies strain rates faster than  $10^{-14} \text{ s}^{-1}$  for displacement rates greater than  
1035 0.1 mm/yr, whereas estimates for deformation distributed over larger areas  
1036 produces strain rates less than  $10^{-15} \text{ s}^{-1}$ . b) The distribution of strain rates  
1037 taken from the deforming zones in the Global Strain Rate Model (Kreemer  
1038 et al., 2014) compared to those of Pfiffner and Ramsay (1982). ‘Deforming  
1039 zones’ are defined as plate boundaries and zones of diffuse deformation sepa-  
1040 rating rigid plates, amounting to about 14 % of the Earth’s surface (Kreemer  
1041 et al., 2014).

1042

1043 Figure 2: Strain rate and seismicity in California, USA, and strain rate and  
1044 topography for the Himalayan orogen. The strain rate maps show the 2nd  
1045 invariant of strain rate as determined by the Global Strain Rate Map project  
1046 (Kreemer et al., 2014) at  $0.1^\circ$  resolution. (a) Strain rate in California. Note  
1047 the localisation, by at least an order of magnitude in strain rate, into the San  
1048 Andreas fault system, which deforms at a strain rate greater than  $10^{-14} \text{ s}^{-1}$ .  
1049 (b) Earthquakes with magnitude 3.0 or greater recorded in the NEIC cata-  
1050 logue since 1970. (c) Strain rate in the Himalayan orogen. Note the increase  
1051 by at least an order of magnitude at the Himalayan front, as well as along a  
1052 few other localised (and potentially transient) active structures. (d) Eleva-

1053 tion from the GEBCO 2014 grid at 30 second resolution (The GEBCO\_2014  
1054 Grid, version 20150318, [www.gebco.net](http://www.gebco.net)). Figures created in Generic Map-  
1055 ping Tools (Wessel et al., 2013).

1056

1057 Figure 3: Examples of strain heterogeneity in the rock record, as shown by  
1058 brittle and ductile structures referring to mesoscopically discontinuous and  
1059 continuous deformation. Kinematics indicated by yellow arrows. (a) Duc-  
1060 tilely deformed pseudotachylyte (red arrow points to sheared injection vein)  
1061 that also crosscuts metamorphic tectonite (blue arrow), Nusfjord, Norway  
1062 (see Menegon et al., 2017, for more detail). (b) Hydrothermal veins cross-  
1063 cut metamorphic tectonite, but are also rotated and ductilely sheared. Both  
1064 veins and rotated foliation record normal shear sense. A later brittle fault  
1065 that is not ductilely deformed cuts through the centre of the veins implying  
1066 further brittle localisation with time. Makimine mélange, Kyushu, Japan  
1067 (Ujiie et al., 2018). (c) Localised brittle deformation in the core of the San  
1068 Gabriel strike-slip fault, California, produced cataclasite in a narrow princi-  
1069 pal slip zone. (d) Strain localisation within a relatively homogeneous ductile  
1070 shear zone, Nusfjord, Norway (see Menegon et al., 2017, for more detail). (e)  
1071 Quartz and felspar porphyroclasts behaving as relatively rigid bodies within  
1072 a lower viscosity biotite-rich matrix, Maud Belt, Antarctica. (f) A low com-  
1073 petency matrix enveloping sheared competent clasts in the Chrystalls Beach  
1074 Complex, New Zealand. Note thin cataclastic surfaces both parallel to, and  
1075 cross-cutting, the matrix cleavage (examples in dashed yellow lines).

1076

1077 Figure 4: Simple model of surface velocity and strain rate caused by inter-

1078 seismic slip on an infinitely long strike-slip fault (Savage and Burford, 1973).  
1079 Both parameters are controlled by locking depth, meaning geodetic mea-  
1080 surements of strain do not accurately record localised strain rates at depth,  
1081 particularly for regions with deep brittle-ductile transitions.

Table 1: Estimates of viscosity and strain rate from thin viscous sheet models of various continental regions. The quoted viscosities assume a power law exponent of  $n=3$ .

Region	Viscosity Pas	Strain Rate $s^{-1}$	Reference
Arabian-Eurasia	$1 - 5 \times 10^{22}$	$3 \times 10^{-16} - 3 \times 10^{-15}$	Walters et al. (2017)
Anatolia	$3 \times 10^{21} - 10^{22}$	$6 \times 10^{-17} - 6 \times 10^{-15}$	England et al. (2016)
Tibet	$10^{22}$	$10^{-16} - 10^{-15}$	England and Molnar (1997)
Tibet	$5 \times 10^{21} - 5 \times 10^{22}$	$< 5 \times 10^{-15}$	Flesch et al. (2001)
Tien Shan	$1 - 4 \times 10^{22}$	$10^{-15}$	England and Molnar (2015)
North America	$10^{21} - 10^{22}$	-	Flesch et al. (2000)
Appenines	$1.5 - 3 \times 10^{21}$	$2 \times 10^{-15}$	D'Agostino et al. (2014)

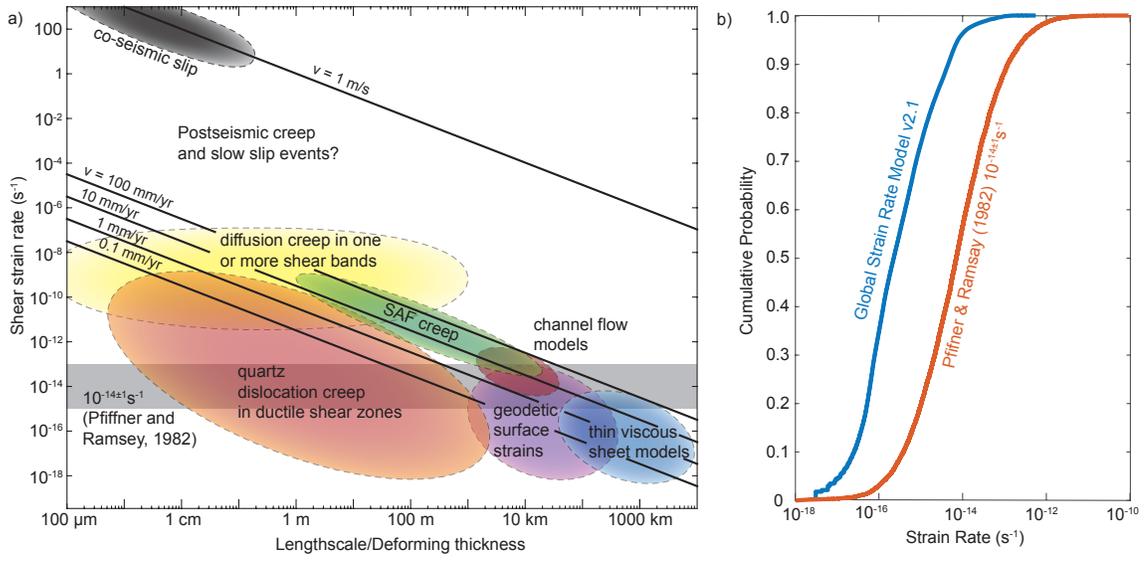


Figure 1:

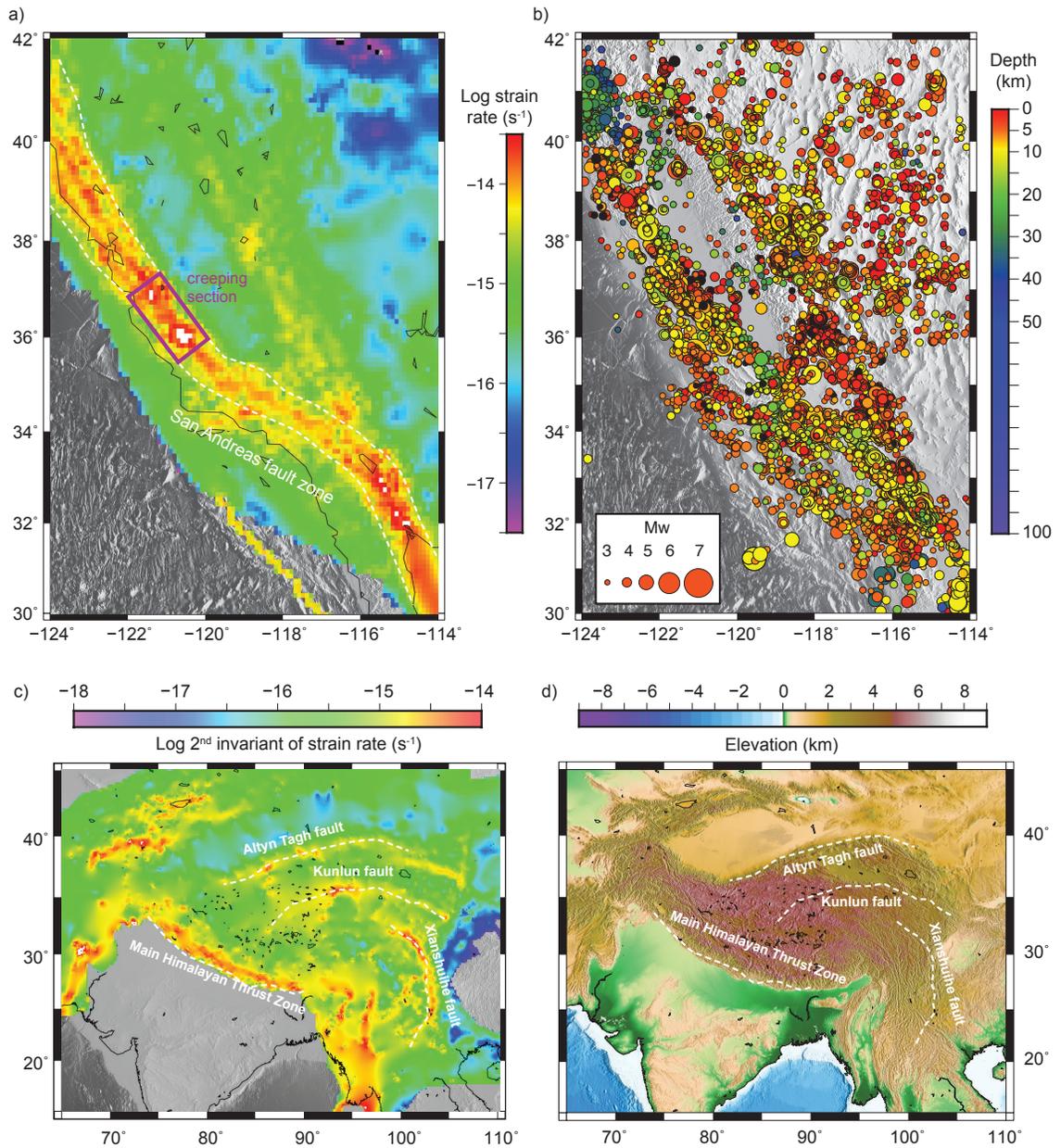


Figure 2:

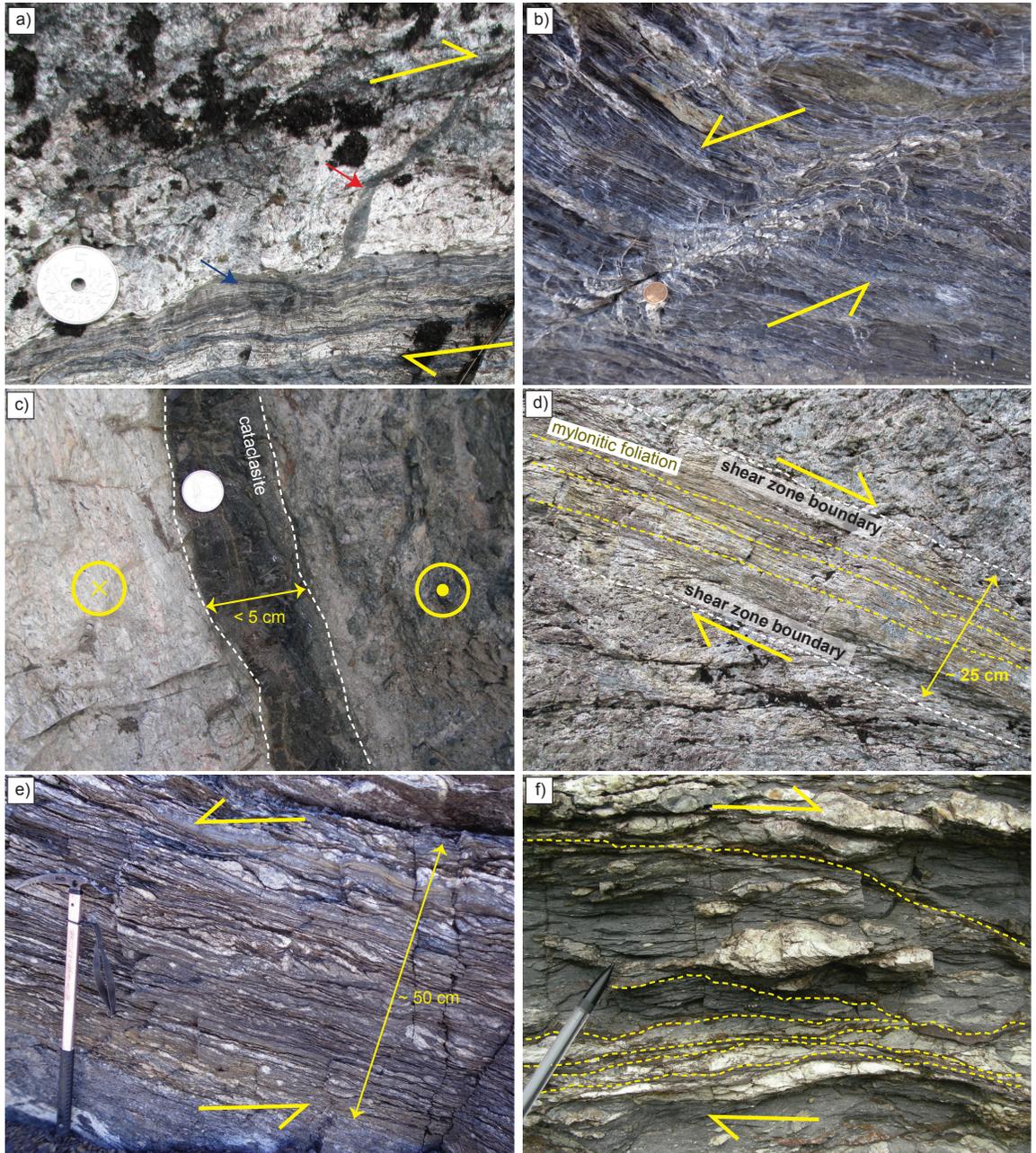


Figure 3:

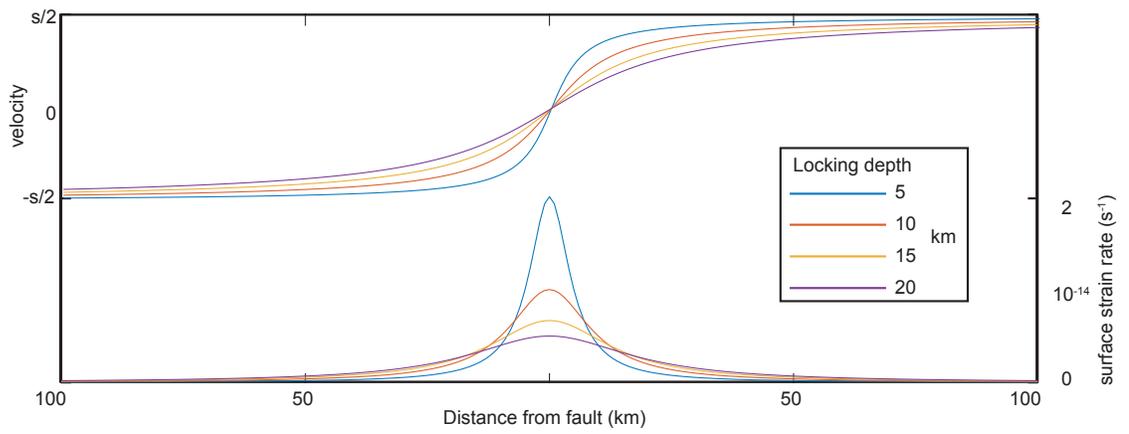


Figure 4: