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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

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, Δl , divided by its original length
38 prior to a discrete deformation episode, l_0 . Alternatively, one may calculate
39 natural strain, ϵ , 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} 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} 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):

$$\Delta\sigma = BD^{-p} \tag{1}$$

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 μm to about 40 μm 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} 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} 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} s^{-1} in low strain areas, and greater

138 than 10^{-13} 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 ≤ 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} 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 monominerallic 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} s^{-1} to 10^{-8} 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} 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} 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} s^{-1} compared to
214 10^{-15} 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} 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} 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 ± 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} 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 ≥ 10 s⁻¹, 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 10⁻³ yr⁻¹
508 or 10⁻¹¹ s⁻¹ (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} s^{-1} to 10^{-14} 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} 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 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} 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} 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} s^{-1} , separating lower strain blocks where transient

611 strain rate increases may occur, but average strain rate is less than 10^{-14} 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} s^{-1} to 10^{-8} s^{-1} (e.g. Gueydan et al., 2005; Boutonnet et al., 2013;
616 Viegas et al., 2016). Thus, while 10^{-14} 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} 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} 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} 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} 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} 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° 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} 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). 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 | $1.5 - 3 \times 10^{21}$ | 2×10^{-15} | D'Agostino et al. (2014) |

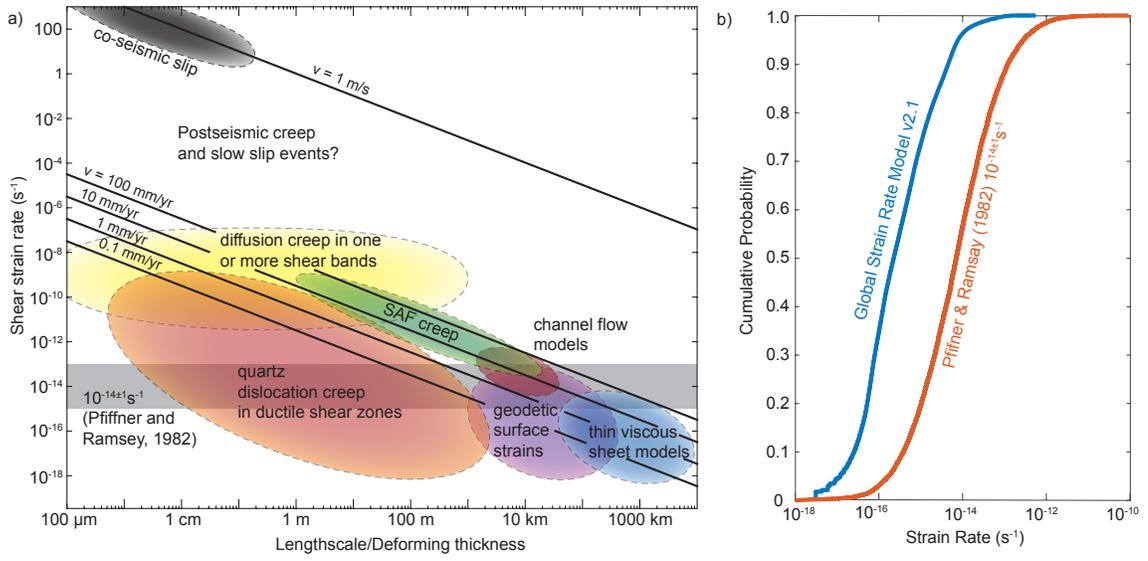


Figure 1:

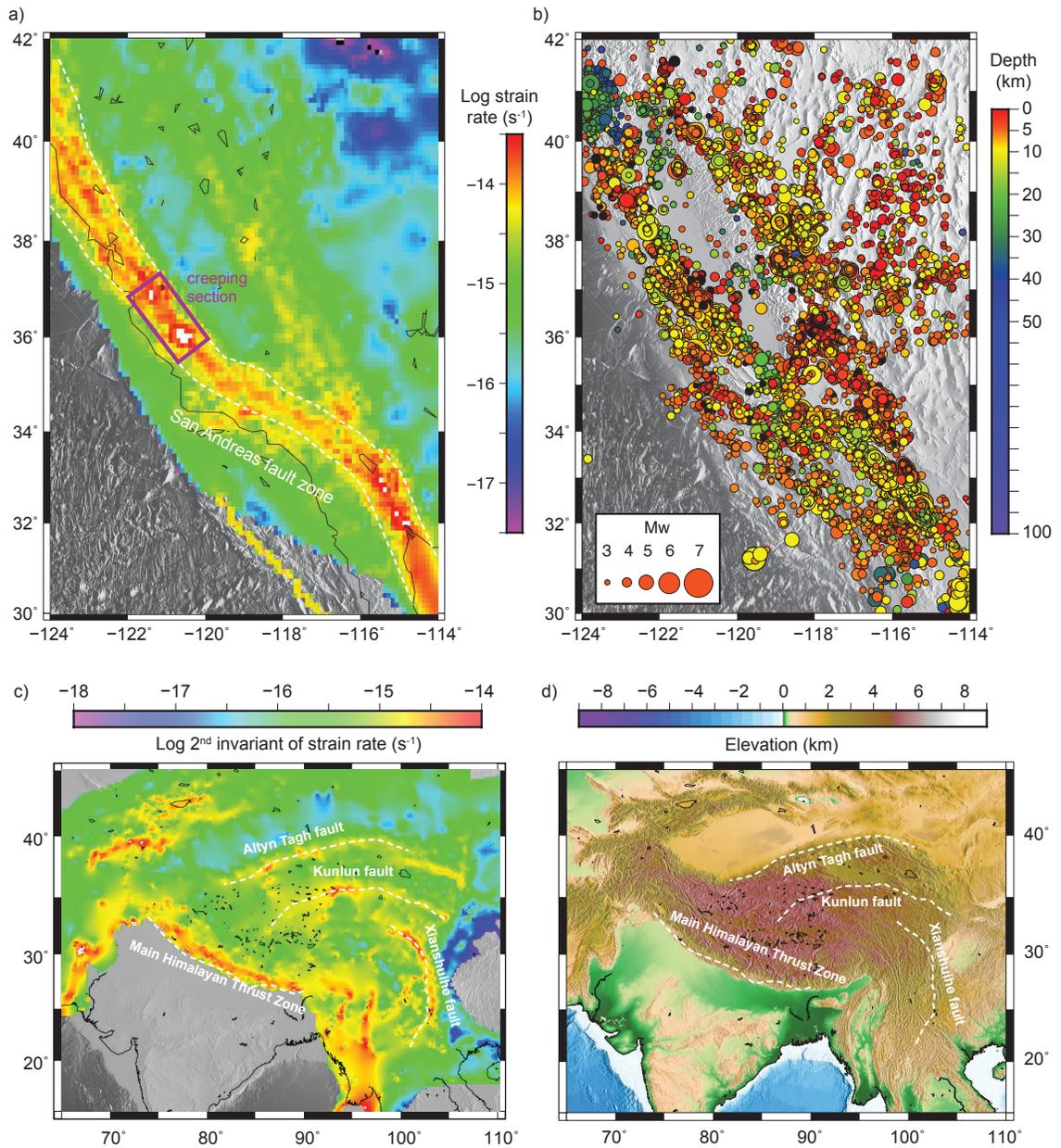


Figure 2:

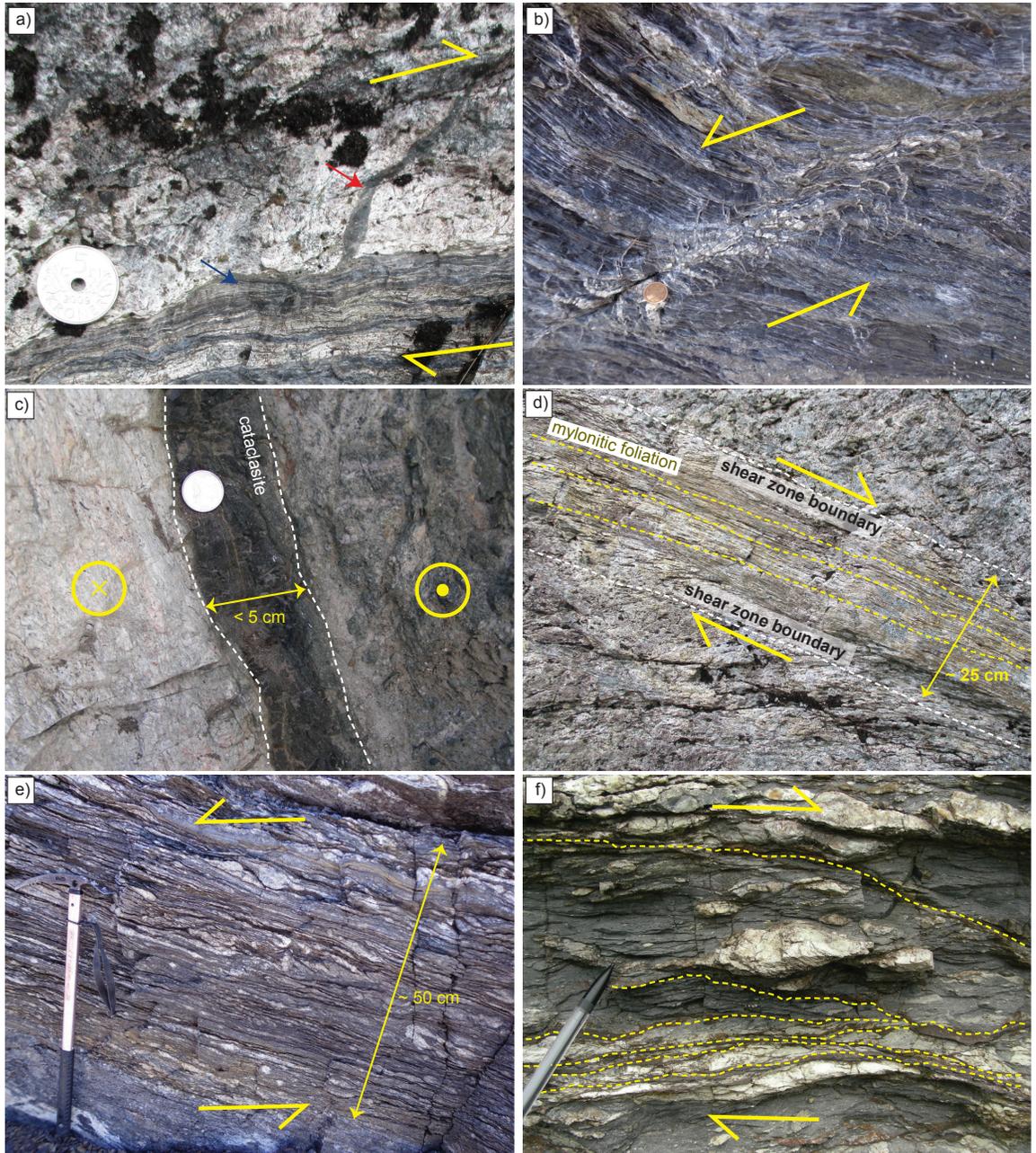


Figure 3:

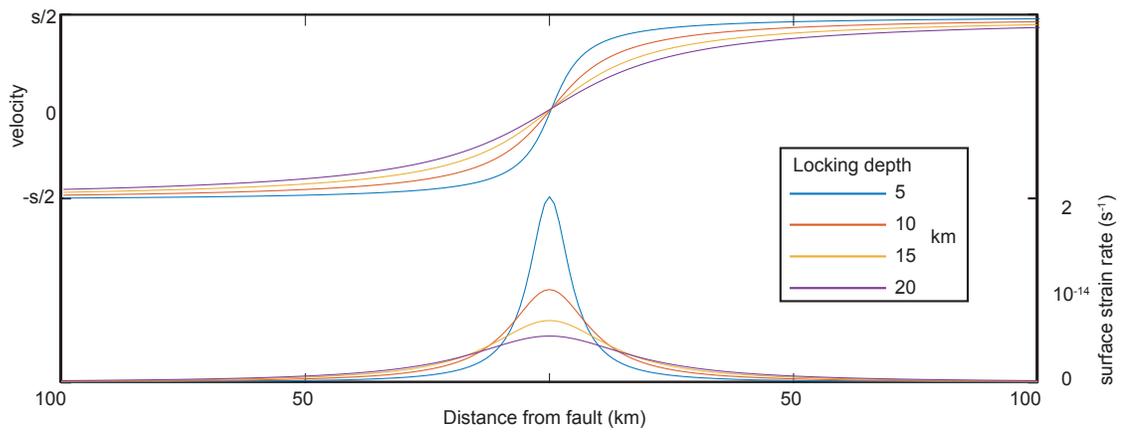


Figure 4: