1 Combining GPS and repeating earthquakes for a high 2 resolution analysis of subduction zone coupling

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5 **Abstract**

Increasingly complex spatiotemporal patterns of subduction zone coupling are being revealed by geodetic and seismic observations. Understanding the mechanisms which control it is useful for improving seismic hazard assessments. GPS and characteristically repeating earthquakes (CREs) are complementary datasets for monitoring aseismic slip. Here, both of them are combined to estimate the rate and distribution of creep on the northeast Japan subduction zone between 21st March 1996 - 24th September 2003. We find that the majority of the upper part at 0 - 30 km depth remains locked. There are three regions creeping at 7-8 cm/yr distributed along-strike at 40-70 km depth. We observe that these creeping regions occur in areas of low effective pressure and reduced porosity, which are inferred from V_p and V_s velocities. Moreover, an area of high clay content and high effective pressure coincides with the rupture area of the Tohoku-oki earthquake. We discuss these results in the context of potential mechanisms governing creep in northeast Japan. Our results highlight the benefits of combining GPS and CREs for advancing our understanding of the seismic cycle in subduction zones.

6 *Keywords:* subduction zone creep, repeating earthquakes, GPS,
7 earthquake cycle, interseismic coupling, northeast Japan

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1. Introduction

Geodetic and seismic observations at subduction zones have revealed a wide range of creep behaviours during the interseismic period. These include; persistent creep (Schmalzle et al., 2014), slow slip events (SSEs) (Wallace et al., 2012), and episodic tremor and slip (ETS) (Rogers and Dragert, 2003). More than one type of behaviour has been observed at the same subduction zone, for example persistent creep and ETS in Cascadia (Schmalzle et al., 2014; Rogers and Dragert, 2003), and SSEs of varying duration in Hikurangi (Wallace et al., 2012). More recently, SSEs (Kato et al., 2012) and accelerated creep (Schurr et al., 2014) have been suggested as precursors to large megathrust events. Robust observations of the complex spatial creep patterns are useful for understanding of the physical mechanisms driving creep, and consequently improving regional seismic hazard assessments.

GPS data have proven useful for monitoring creep phenomena in subduction zones, such as Cascadia (e.g., Dragert et al., 2001; Rogers and Dragert, 2003), and Japan (e.g., Hirose et al., 1999; Obara et al., 2004). Although for most subduction zones the nearest stations are onshore and far from the trench, thus the resolution of this data is reduced at shallow depths (e.g., Loveless and Meade, 2011). The large slip magnitudes at shallow depths in recent earthquakes, such as M_w 9.0, 11th March 2011, Tohoku-oki (e.g., Lee et al., 2011; Yamazaki et al., 2011), and M_w 7.7, Mentawi, 25th October 2010 (Yue et al., 2014), highlight the importance of the development of monitoring and modelling schemes to better constrain slip behaviour close to the trench.

The discovery of characteristically repeating earthquakes (CREs) has pro-

32 vided an alternative means for measuring creep, initially at transform settings
33 (e.g., San Andreas, Nadeau and McEvilly, 2004; Templeton et al., 2008), and
34 subsequently at subduction zones (e.g., northeast Japan, Uchida et al., 2009;
35 Uchida and Matsuzawa, 2013). CREs are believed to represent repeated
36 failure of the same fault patch, which is loaded by creep on the surrounding
37 fault surface (Nadeau and Johnson, 1998). As such they can act as creep
38 meters that are directly on the fault plane, but they cover less than 1% of
39 the plate interface. Thus, CREs can provide robust observations, but over
40 very limited areas. Conversely, GPS data monitor the whole subduction
41 zone from the trench to the bottom of the seismogenic zone but at variable
42 resolution. Therefore, combining both GPS and CRE datasets enables the
43 estimation of creep across the whole fault at high resolution, yet few studies
44 have combined the two (Shirzaei and Burgmann, 2013; Shirzaei et al., 2014).

45 The spatially heterogeneous distributions of subduction zone creep ob-
46 served by GPS (Nishimura et al., 2004; Ozawa et al., 2012) and CREs (Uchida
47 and Matsuzawa, 2013) have prompted a debate of what factors govern sub-
48 duction zone creep. Geometry of the plate interface (Gao and Wang, 2014;
49 Kopp, 2013; Wang and Bilek, 2011; Mochizuki et al., 2008; Scholz and Cam-
50 pos, 1995), material heterogeneities (e.g., Perfettini et al., 2010; Chlieh et al.,
51 2008), the presence of fluid (Peacock, 1990; Kulm et al., 1986), clay content
52 (Ujiie and Tsutsumi, 2010), permeability (Hilaret et al., 2007), and loca-
53 tions of past earthquakes (Burgmann et al., 2005) have all been suggested as
54 contributing factors.

55 There are a few studies which directly link these factors with observations
56 of creep. Direct measurements of these properties are difficult to obtain, in

57 part due to the inaccessible location of these tectonic regimes. However,
58 P and S wave velocities from local and regional tomography models have
59 been used in combination with empirical relations based on experimental
60 results to remotely investigate the mechanical properties of subduction zones.
61 Abnormally high pore fluid pressures have been suggested for the Nankai
62 accretionary prism, using this approach (Tsuji et al., 2008). At larger depths,
63 silent slip events in the Nankai Trough have been linked with high pore fluid
64 pressures due to the dehydration of a subducting ridge Kodaira et al. (2004).
65 However, it wasn't until recently that there was a direct link made between
66 the spatial distribution of both creep and effective pressure in subduction
67 zones. Moreno et al. (2014) located a region of low effective pressure, and
68 consequently high pore fluid pressure, which coincided with an area of creep
69 inferred from GPS data, in the northern Chile subduction zone.

70 These observations suggest that pore fluid pressures may play a role in
71 driving creep in subduction zone settings. However, it is also suggested that
72 it is the frictional properties controlling the distribution of effective pressure
73 and consequently pore fluid pressure, which in turn governs creep. Assuming
74 that the rate state framework holds, steady creep can seal the fault zone and
75 allow the development of high pore pressures which facilitate further sta-
76 ble creep (Sleep, 1995; Rubin and Ampuero, 2005; Perfettini and Ampuero,
77 2008). Both pore fluid pressure and frictional properties are discussed later
78 in relation to our results.

79 Comparisons between the distribution of effective pressure and creep have
80 yet to be carried out for an entire subduction zone, and existing studies rely
81 on creep distributions determined using either geodetic or seismic datasets.

82 It is also possible to estimate the clay content and porosity from seismic ve-
 83 locities yet distributions of these properties have not been directly compared
 84 with creep, in the context of effective pressure. Therefore, in this study we
 85 estimate the spatial pattern of creep rate between 1996 and 2003 using GPS
 86 and CRE data. This is compared with the distribution of effective pressure
 87 (P_e), clay content (C) and porosity (ϕ). Both the GPS and CRE datasets,
 88 and the P (V_p) and S (V_s) wave velocities are described in detail below. This
 89 is followed by an outline of the inversion scheme for the creep rate and the
 90 implementation of the empirical relation proposed by Eberhart-Philips et al.
 91 (1989), to infer the C , ϕ and P_e from V_p and V_s . The resulting patterns are
 92 then discussed in the context of the properties and mechanics of the northeast
 93 Japan subduction zone.

94 2. Data

95 Northeast Japan is perhaps the best monitored subduction zone in the
 96 world with over 1,200 GPS stations, and a dense seismic network. We con-
 97 sider the eight year period between 1996 and 2003, as it is believed to rep-
 98 resent a steady state in the region. This is due to the lower rate of observed
 99 deformation (Ozawa et al., 2012) and the absence of earthquakes greater than
 100 M_w 6.5. Specifically we use data between 21st March 1996 and 24th Septem-
 101 ber 2003 due to the availability of the data, and on the 25th September 2003
 102 the M_w 8.3 Hokkaido earthquake occurred north of the study region which
 103 resulted in a large coseismic and postseismic signal in the GPS data. Dis-
 104 placements measured by ~ 980 GPS stations (Ozawa et al., 2012) and CRE
 105 observations from ~ 100 sequences (Uchida and Matsuzawa, 2013) are used

106 in an inversion scheme (Shirzaei et al., 2014) to determine the spatial creep
107 rate distribution. Both datasets are shown in Fig. 1, and are described in
108 more detail in the following sections, as well as the mesh used to represent
109 the subducting plate and the regional tomography model used to infer the
110 C , P_e , and ϕ .

111 2.1. Plate Geometry

112 To represent the subducting plate interface we use a triangular mesh,
113 with variable strike and dip, and an average patch size of 20 km (Fig. S1). It
114 is based on the fault geometry from Slab 1.0, which is defined using a variety
115 of seismic datasets and bathymetry data (Hayes et al., 2012). This mesh is
116 used in the inversion, and to infer the spatial distribution of ϕ , C , and P_e .

117 2.2. GPS

118 The dataset consists of GPS displacements that were analysed daily us-
119 ing Bernese GPS software (version 5.0), for further details see Ozawa et al.
120 (2012). In this study only the horizontal components are considered due to
121 the low signal to noise ratio for the vertical component and the contribution
122 from subduction erosion that adds additional complexity to the vertical de-
123 formation data (Heki, 2004). Signals due to; nearby earthquakes, the effect
124 of the long term convergence rate, and seasonal components are all removed,
125 as well as any stations with significant noise. However, it should be noted
126 that we do not attempt to correct for any postseismic relaxation or afterslip,
127 therefore our model could include these effects.

128 Firstly, coseismic offsets due to the M_w 7.0 intra-slab thrust event on
129 26th May 2003 (e.g., Okada and Hasegawa, 2003) are removed from the time

130 series recorded by stations in the Miyagi prefecture, by subtracting the dis-
 131 placement on the day after from the day before. Next the effect of common
 132 mode noise, which is a linear combination of the entire network translation
 133 and rotation, is removed from the GPS network, following Wdowinski et al.
 134 (1997). The common mode noise is calculated by analysing the stations that
 135 are outside the Tohoku area, these fiducial stations are selected to ensure
 136 that the estimation of common mode noise is not affected by creep on the
 137 subduction zone (< 70 km depth). Finally, the long term velocity of each
 138 station is compared with the median long term velocity of surrounding sta-
 139 tions, within a 50 km radius, and if it is not within 15% of the median value
 140 then it is removed. This step minimises the effect of deformation signals due
 141 to local subsidence, volcanic activities, and in-land seismicity. The result-
 142 ing long term rates of displacement for 128 stations, which were consistently
 143 measuring throughout the entire period, is shown in Fig. 1a.

144 *2.3. P and S wave velocities*

145 We use the V_p and V_s velocities from the regional tomography model
 146 of Huang et al. (2011), which was determined using P and S wave arrivals
 147 from 4655 local earthquakes. This is one of the best models available that
 148 covers the majority of our study area. The model extends from 0 to 200
 149 km depth, with 10-30 km vertical resolution and ~ 30 km resolution in the
 150 horizontal direction, which is similar to the average patch size of the mesh
 151 used in the inversion. Errors of 0.2-0.3% for the V_p and V_s were estimated
 152 using checkerboard resolution tests (Huang et al., 2011).

Figure 1: a) Long term GPS displacement rates between 1996 - 2003 for 128 stations are shown by the yellow arrows. The V_p/V_s distribution (Huang et al., 2011) for the region is shown in red and blue. The central section which is the focus of the discussion later, is outlined by the grey dashed lines. b) Locations of the CREs occurring in the same period, are denoted by the black circles whereby the size corresponds to the magnitude of the cumulative slip. The coloured contour lines correspond to the spatial distribution of the creep rates determined from the CRE sequences, which is used as a constraint in the inversion, see text for further details. The contours of maximum coseismic rupture in the Tohoku-oki earthquake, based on the model from Shirzaei et al. (2014), are plotted in pink. Depth contours are denoted by the dashed black lines in both figures.

153 2.4. *Characteristically Repeating Earthquakes*

154 Uchida and Matsuzawa (2013) identified CREs on the northeast Japan
155 subduction zone between 1984 - 2012 using cross-spectral analysis and grouped
156 them into CRE sequences occurring at discrete locations on the subducting
157 plate interface, for further details see Uchida and Matsuzawa (2013).

158 In this study CRE sequences which occur between 21st March 1996 and
159 24th September 2003 are extracted from this dataset, and their location and
160 cumulative slip during this period are shown in Fig. 1b. The majority of the
161 CREs appear to be confined to depths of 40-70 km, whereas CRE activity is
162 reduced in areas close to the trench. As CREs are thought to represent the
163 failure of relatively small asperities that are loaded by aseismic creep on the
164 adjacent fault surface, the distribution of CRE sequences mark creeping and
165 locked areas on the Japan subduction zone (Uchida et al., 2005; Uchida and
166 Matsuzawa, 2013). The CRE distribution shown in Fig. 1b is significantly
167 different than existing coupling models based on GPS data (e.g., Ito et al.,
168 2000; Hashimoto et al., 2009; Loveless and Meade, 2010). Although in the
169 context of the 2011 Tohoku earthquake most of the CREs occur outside the
170 zone of maximum coseismic rupture (> 30 m, Fig. 1b).

171 To use the CRE data as bounds on the inverted creep, we identify CRE
172 sequences with more than three events and estimate the average creep rate
173 for each sequence using linear regression. Then a 50 km search radius centred
174 on each patch of the mesh representing the plate interface (Fig. S1) is used
175 to assign the creep rates and locations for these CRE sequences. The 50 km
176 radius was chosen so each patch is assigned a value, yet the distribution is
177 not oversmoothed. The resulting distribution of creep rates which are used

178 to define the creep bounds in the inversion are shown in Fig. 1b.

179 **3. Method**

180 *3.1. Inversion scheme*

181 To associate the rate of GPS displacements with the creep on the sub-
 182 ducting plate interface an inverse modelling scheme is implemented, which
 183 has previously been used to model afterslip following the M_w 9.0 Tohoku-oki
 184 event (Shirzaei et al., 2014). The GPS velocities are inverted to solve for a
 185 static creep rate distribution on the plate interface.

186 The creep rate, C_r , on each patch is related to the GPS velocities, L , as
 187 follows:

188

$$L + v = BC_r, P = \sigma_0^2 \sum_{l,l}^{-1}$$

$$l_b \leq C_r \leq u_b$$

(1)

189 where, v is the vector of observation residuals, and B is the Green's func-
 190 tions, which are calculated assuming purely dip-slip motion and a Poisson
 191 ratio of 0.25. The variance-covariance matrix of the observations is repre-
 192 sented by $\sum_{l,l}$, where the diagonal elements are proportional to the inverse of
 193 the variance of the observations, and σ_0^2 is the variance factor (Mikhail, 1976).
 194 This equation is solved subject to a re-weighted L2-norm minimization of the
 195 observation residuals, whereby following each iteration the observations are
 196 re-weighted based on the residuals from the previous iteration (see Shirzaei

197 and Burgmann (2013) for further details). This approach allows the minimi-
 198 sation of the effect of outliers and other observation artefacts (O’leary, 1990).
 199 The upper (u_b) and lower bounds (l_b) of the creep rate are defined using the
 200 creep rate on each patch observed by the CREs, as follows:

$$u_b = \alpha C_{CRE} + S_L$$

$$l_b = \max\{0, \alpha C_{CRE} - S_L\} \quad (2)$$

202 where, C_{CRE} is the mesh of creep estimated for each patch from the CRE
 203 data, described in section 2.1, S_L is the long term convergence rate (8 cm/yr),
 204 and α is a factor determined from iterative testing and the optimum value is
 205 200. To ensure that areas with no creep observed by CREs are still allowed
 206 to creep in the inversion during the inversion we check that optimum creep
 207 is not converging to either the upper or lower bounds and if it is the bounds
 208 are extended. The second order derivative of the creep is used to reduce
 209 the roughness of the creep, where λ is the smoothing factor and D is the
 210 Laplacian operator (Harris and Segall, 1987):

$$\lambda DC_r = 0 \quad (3)$$

212 The east component is given more weight than the north component as
 213 the convergence rate is predominantly east-west, consequently the signal in
 214 the north direction is smaller, and the signal to noise ratio is also much
 215 smaller. Following several tests using different ratios of weighting factors
 216 (Fig. S3), a ratio of 8:1 for the east and north components, respectively, was
 217 chosen. This gave the best fit to the data, and the most physically reasonable

218 and plausible creep rate distribution, particularly in terms of the magnitude
 219 and location of maximum creep rate (Fig. S3).

220 3.2. Creep model resolution

221 One popular approach to clarify the resolution of the inverted parameters
 222 is to estimated their associated standard deviations. The standard deviation
 223 is defined as the square root of the diagonal components of the variance
 224 covariance matrix (Mikhail, 1976). For the case of a GPS-only inversion, the
 225 variance-covariance matrix (Q_{CC}^G) is defined as:

$$226 \quad Q_{CC}^G = (B^T P B + \lambda^2 D^T D)^{-1} \quad (4)$$

227 where $()^{-1}$ and $()^T$ are inverse and transpose operators respectively. The
 228 standard deviations associated with equation 4 are provided in Fig. 2a.
 229 Similarly the model resolution associated with creep observed by CREs can
 230 be estimated as follows:

$$231 \quad Q_{CC}^E = (E^T E + \lambda^2 D^T D)^{-1} \quad (5)$$

232 Where, E is a square matrix whose size is equal to to the number of
 233 triangles representing the mesh, and contains ones at the diagonal elements
 234 and zeros everywhere else. Fig. 2b shows the distribution of the square root
 235 of the diagonal elements of Q_{CC}^E . Finally, the model resolution as a result of
 236 combining GPS and CREs is given by:

$$237 \quad Q_{CC}^{G+E} = (B^T P B + E^T E + \lambda^2 D^T D)^{-1} \quad (6)$$

238 As seen in Fig. 2c, the Q_{CC}^{G+E} provide the lowest overall standard devia-
 239 tions out of all three scenarios, namely a better resolution. This improvement
 240 can be quantified as:

$$241 \quad Iq = \frac{\sqrt{Q_{CC}^{G+E}} - \sqrt{Q_{CC}^G}}{Q_{CC}^{G+E}} \times 100 \quad (7)$$

242 The improvement (Iq) is shown in Fig. 2d.

243 3.3. *Effective pressure, porosity, and clay content estimation*

It is extremely hard to obtain “ground-truth” observations of ϕ , and C at subduction zones, mostly because they are so difficult to access. Therefore, empirical relations are often used to infer these values and the P_e , particularly when investigating such a large study area. We use an empirical relationship established between these parameters and V_p and V_s (Eberhart-Philips et al., 1989). It is based on data from experimental results involving velocity measurements of 64 different sandstones, under water-saturated conditions. Forward stepwise multiple regression is applied to this data to determine the best fitting relationships for V_p and V_s that are dependent on ϕ , P_e , and C. The relationships are as follows:

$$V_p = 5.77 - 6.94\phi - 1.73\sqrt{C} + 0.446(P_e - e^{-16.7P_e}) \quad (8)$$

$$V_s = 3.70 - 4.94\phi - 1.57\sqrt{C} + 0.361(P_e - e^{-16.7P_e}) \quad (9)$$

244 Given the distribution of the V_p and V_s on the subduction zone (Fig. 1,
 245 Huang et al., 2011) and equations (5) and (6), a Genetic Algorithm, (GA)
 246 (Shirzaei and Walter, 2009) is used to search for values of ϕ , P_e , and C for

247 each patch. In the basic form of GA (Holland, 1975) a key aspect is the
 248 binary encoding of the parameters. The search begins by initialising the GA
 249 parameters and defining the cost function:

250

$$\begin{aligned}
 Cost = & |V_p - [5.77 - 6.94\phi - 1.73\sqrt{C} + 0.446(P_e - e^{-16.7P_e})]| + \\
 & |V_s - [3.70 - 4.94\phi - 1.57\sqrt{C} + 0.361(P_e - e^{-16.7P_e})]| + \\
 & |\lambda D\phi| + \\
 & |\lambda DC| + \\
 & |\lambda DP_e| \quad (10)
 \end{aligned}$$

251 Where, $||$ is the absolute value operator, D is the Laplacian operator, and
 252 λ is the smoothing factor which determines the roughness of the estimate
 253 parameters.

254 The GA ends when a stopping criterion, such as the defined number of
 255 iterations, is achieved. The GA always deals with a set of solutions without
 256 any special emphasis on a particular solution. So, if the initial population
 257 consists of m -independent solutions, each solution includes n variables, and
 258 each variable can be presented by q bits, then the initial population is con-
 259 structed randomly by $n \times (m \times q)$ zeroes / ones Haupt and Haupt (2004).

260 As the algorithm progresses the initial population is subject to several
 261 random changes but systematically approaches the best solution among many
 262 investigated solutions. Selection, pairing, mating, and mutation are major
 263 GA operators that allow the modification of the population to explore the
 264 solution space, and are thoroughly explained in the literature (Holland, 1975;
 265 Michalewicz, 1994).

266 Although GA has remarkable success in solving very complex optimiza-
267 tion problems, it does not effectively provide information regarding the qual-
268 ity of the results. Thus, to evaluate the quality of our inversion results, we
269 also employ a randomly iterated search (Shirzaei and Walter, 2009), which
270 repeats the the optimisations and algorithm, and increases the freedom of
271 the search algorithm to explore the solution space. This iterated approach
272 allows the generation of an ensemble of solutions that can be used to estimate
273 the errors and the confidence level of the results.

274 It should be noted that “An increased value of V_p/V_s can indicate a de-
275 crease in σ_e , a decrease in porosity, or an increase in clay content or some
276 combination thereof ” (Eberhart-Philips et al., 1989). Using an empirical re-
277 lation has its disadvantages, mostly due to the data used to determine them.
278 For example, they are based on certain rock types. The Eberhart-Philips
279 et al. (1989) relation, used here, is based on experimental results from sand-
280 stones, which arguably as the subduction zone is believed to be comprised
281 of pelagic clay (Kameda et al., 2015; Chester et al., 2013) is only a first or-
282 der approximation. Although, Eberhart-Philips et al. (1989) established this
283 relation based on results from tests which covered a wide range of clay con-
284 tents (0-0.5), and porosities (0.2 - 0.3). The results from the Eberhart-Philips
285 model have been used to interpret coupling distributions on the Chile sub-
286 duction zone (Moreno et al., 2014). Here we investigate whether the results
287 from this model are interpretable in the context of creep rate and spatial
288 pattern on the northeast Japan subduction zone.

289 3.4. Cluster analysis

290 To establish the extent to which the spatial pattern of creep rate overlaps
291 with the distribution of ϕ , P_e , and C we apply cluster analysis to each of
292 the distributions to reduce noise. Cluster analysis is a means of grouping
293 together objects with similar properties into defined classes, using the metric
294 distance between them. It can be easily and effectively applied to large-scale
295 datasets, thus it has been adopted for a wide range of purposes (Houser et al.,
296 2013; Simpson et al., 2012; Shirzaei et al., 2013). We use the euclidean dis-
297 tance between the centroid of each cluster to quantify the similarity between
298 members of a class. A range of maximum number of clusters, from 1 to 50,
299 are tested. To determine the optimum number of clusters we plot the root
300 mean square (RMS) of the distance between clusters against the number of
301 clusters, and the minimum number of clusters which provides the minimum
302 RMS is selected (Fig. S4). In this study this we use between 8 to 10 clusters.
303 The optimum number of clusters is identified and applied for each spatial
304 pattern (creep rate, ϕ , C , and P_e). Then the clustered distribution of creep
305 rate is compared in turn with the C , ϕ , and P_e distributions to quantitatively
306 assess how much the distributions coincide.

307 4. Results

308 4.1. Creep Rate

309 We first test the resolution of the optimum creep rate model and estimate
310 the improvement achieved by using the location and rate from CREs as a
311 constraint in the inversion. For an inversion using just GPS data the model
312 resolution, in terms of the estimated creep standard deviation, decreases as

Figure 2: **a)** Creep rate model resolution in terms of the standard deviation (STD), for a model from an inversion using only GPS data. **b)** Same as **a)** except for an inversion using CRE data. **c)** GPS and CRE inversion. **d)** Percentage improvement in the model resolution when the CRE data are added to the GPS data.

313 you move towards the trench (Fig. 2a). Yet, when the CRE data are added
314 there is a reduction in the standard deviation, particularly for the patches
315 at the trench (Fig. 2b), and across the whole plate there is a substantial
316 improvement in the model resolution (40-70%, Fig. 2c).

317 The resulting long term (interseismic) creep rate from this joint inversion
318 of GPS and CRE data is highly heterogeneous along-strike and down-dip,
319 and fits the GPS data well (Fig. 3a) with an RMS of ± 0.15 cm/yr. The
320 upper part of the plate (0-40 km) is dominated by three regions exhibiting
321 little or no slip, and down-dip of these, at 40-70 km depth, there is substantial
322 creep with interseismic rates reaching 7-8 cm/yr (Fig. 3b).

323 The spatial creep rate pattern reported here is similar to that observed
324 by CRE studies. The long term rate of 8 cm/yr in the central and southern
325 regions is higher than the 4cm/yr reported by Igarashi et al. (2003); Uchida
326 and Matsuzawa (2013), but this could be due to a sparse and incomplete CRE
327 catalogue and the missing signal is augmented by GPS displacements in this
328 study. Furthermore the empirical relations used to estimate the slip and
329 seismic moment are based on observations in a strike-slip setting, therefore
330 they may differ for subduction zone settings.

331 Using the interseismic creep rates from our optimum model we calculate
332 the extent to which the plate interface is coupled, assuming a plate con-
333 vergence rate of 8 cm/yr (Fig. 4). The three creeping regions show low
334 to moderate coupling (0-0.2), whereas up-dip, close to the trench, the plate
335 interface is highly coupled. Rupture areas from historic earthquakes are plot-
336 ted on for comparison. We use the rupture areas reported in Shibazaki et al.
337 (2011) as it is one of the most comprehensive and widely cited. Five of the

Figure 3: **a)** Interseismic creep rate from the optimum model is shown as coloured contours. The black GPS vectors correspond to observations from 128 of the GPS stations used in the inversion, and the resulting displacement rates based on the optimum model are shown in green. **b)** Difference between the GPS data and the model.

Figure 4: The coupling distribution based on the interseismic creep rate model, where 0 corresponds to the plate interface moving freely and 1 means the interface is locked. Asperities from historical events, based on those reported in Shibazaki et al. (2011) are plotted in light blue, and are listed in the legend. The dark blue outline refers to the area of maximum slip in the M_w 9.0 Tohoku-oki earthquake (e.g., Ozawa et al., 2012).

338 events are located in areas with moderate coupling. However, the rupture
 339 area from the M 8.1, 1611 event and the area of maximum slip from the M_w
 340 9.0 Tohoku-oki earthquake coincide with a region of relatively high coupling
 341 (~ 0.8). This is discussed further in relation to the fault mechanics in section
 342 5.

343 4.2. C , P_e , and ϕ distribution

344 The distributions for all three parameters across the whole mesh are
 345 shown in Fig. 5, and the resulting patterns fit the V_p , V_s observations well
 346 (Fig. S5). Interestingly in the centre a region of high clay content and high
 347 effective pressure is also the location of the maximum slip in the Tohoku-oki
 348 earthquake, which are plotted on Fig. 5 as pink contours, for reference. This
 349 is discussed later in the context of the mechanical properties governing the
 350 rupture area.

351 In the central part of the study area (37-40° latitude), there is a clear
 352 transition from locked to creeping; little slip is observed from the trench to
 353 a depth of ~ 35 km and below this extending to a depth of 60 km there is a
 354 region of substantial creep (Fig. 6a). V_p and V_s are spatially heterogeneous
 355 in this region, with a patch of low P and S wave velocities close to the trench.
 356 P_e is high near the trench and decreases as you move down dip (Fig. 6a-
 357 b), whereas the clay content is relatively high and shows a homogeneous
 358 distribution across the region (Fig. 6c-d). The ϕ is relatively heterogeneous
 359 (Fig. 6e-f). In the context of creep rate, areas of high creep rate coincide with
 360 low P_e . A cluster analysis is used to quantitatively assess the relationship
 361 between the distribution of P_e and interseismic creep rate. Clusters are
 362 estimated based on the interseismic creep rate and this cluster distribution

Figure 5: The spatial distribution of P_e (a), clay content (b), and ϕ (c) for the whole mesh. The black contours with corresponding numbers illustrate the V_p/V_s velocities used in the inversion to estimate the three properties shown in a)-c) . The pink contours correspond to the location of maximum coseismic slip (> 30 m) in the Tohoku-oki earthquake, based on the model from Shirzaei et al. (2014).

Figure 7: **a)** Results of the cluster analysis based on clusters from interseismic creep rate, shown as the coloured background, and the average P_e values for the corresponding clusters are shown in the white ellipses. **b)** Results of the cluster analysis based on clusters from P_e , which is plotted as the background coloured distribution, and the average creep rates for example clusters are shown in the white ellipses.

363 is applied to the P_e pattern in order to calculate the average P_e for each
 364 cluster (Fig. 7a). The same approach is taken to calculate the average creep
 365 rate for clusters based on the P_e distribution (Fig. 7b). It is evident in both
 366 cases that for clusters with a high interseismic creep rate the P_e is low, and
 367 vice versa (correlation coefficient of $-0.94 \pm 10\%$). Moreover, in the area of
 368 substantial creep, ϕ is significantly reduced (≤ 0.1 , Fig. 6e-f) and in areas of
 369 little creep ϕ is higher (anti-correlation of -0.55).

370 **5. Discussion**

371 Multiple coupling models have been proposed for the northeast Japan
 372 subduction zone. In-land GPS studies spanning similar time periods (1996-
 373 2002) to the one investigated here suggest, substantial locking at 40 -70 km
 374 depth between 37-40 ° latitude, and a smaller creeping patch at the base of
 375 the seismogenic zone to the north at $> 40^\circ$ latitude (e.g. Ito et al., 2000;
 376 Hashimoto et al., 2009; Loveless and Meade, 2010, 2011). All of these models
 377 fit the GPS data equally well, yet there is a large difference between these
 378 models and the one presented here.

379 In most subduction zones, including northeast Japan, the resolution of
 380 the creep model obtained from in-land GPS data decreases as you move
 381 trenchward (Loveless and Meade, 2011). The high levels of coseismic slip
 382 observed at shallow depths in recent megathrust earthquakes, such as the M_w
 383 7.8, 25th October 2010, Mentawi (Yue et al., 2014) and the M_w 9.0, 11th March
 384 2011, Tohoku-oki (e.g., Lee et al., 2011; Yamazaki et al., 2011), highlight the
 385 importance of constraining slip behaviour close to the trench. It has been
 386 shown that the addition of offshore GPS stations improves the creep model

387 resolution near the trench (Nishimura et al., 2014). Cubas et al. (2015) added
388 the offshore GPS measurements to the existing onshore GEONET dataset
389 and used it to update the coupling model for the northeast Japan. The
390 resulting distribution reveals a large locked area close to the trench, which
391 coincides with the rupture area of the Tohoku-oki earthquake. Even though
392 this offshore GPS data is not included in our study, as it was not available for
393 the time period investigated here, we obtain a similar result. Note that our
394 model produces a fit to the GPS data which is comparable to that reported in
395 other studies (e.g., Loveless and Meade, 2011), suggesting that in-land GPS
396 data cannot uniquely resolve the distribution of fault coupling on subduction
397 zones. The addition of CRE data to the GPS data leads to substantial
398 improvement in the model resolution, particularly at shallow to intermediate
399 depths (Fig. 2c).

400 Thus, the addition of CRE data is the main factor responsible for the
401 large difference between the coupling model presented in this study (Fig. 4)
402 and those derived using only in-land GPS data (e.g. Nishimura et al., 2004;
403 Hashimoto et al., 2009; Loveless and Meade, 2010). Our coupling model
404 is also supported by the high interseismic velocities observed at multiple
405 offshore GPS stations between 2002 and 2011, which following the Tohoku-
406 oki earthquake were interpreted as an indicator of a large locked patch near
407 the trench (Nishimura et al., 2014; Sato et al., 2013). Therefore, the improved
408 spatial resolution and coverage due to the combination of GPS and CRE data
409 has the potential to provide a more realistic estimate of the state of coupling
410 on subduction zones, such as northeast Japan.

411 Robust coupling distributions can provide important information for seis-

mic hazard assessments, as they can be used to estimate the location and size of potential future earthquakes. The rate-state asperity model is commonly applied to explain the location of M 7 -8 earthquakes, and prior to the Tohoku-oki earthquake this was the approach taken in northeast Japan. It suggests that locked patches represent areas which repeatedly rupture in large earthquakes and are surrounded by areas of creep. For the coupling distribution estimated in this study, the location of maximum slip in the Tohoku-oki earthquake does coincide with a highly coupled region, in the centre offshore Sendai (Fig. 4). This is also the location of the 1611, M 8.1 Sanriku earthquake, as well as the 1981 M_w 7.1 Miyagi-oki earthquake, and the M 7.0, 31st October 2003 event, which occurs shortly after the end of the interseismic period being investigated here. A similar agreement between these rupture areas and the state of coupling is seen in Perfettini and Avouac (2014), which would be consistent with this rate-state asperity model. However, in our study five other asperities of $M \leq 7.5$ events are in areas with moderate coupling. Moreover, it has been found that the initial eight months of afterslip following Tohoku-oki, also overlaps with areas of historical earthquakes (Shirzaei et al., 2014). These observations suggest that the rate-state asperity model may not fully explain the mechanisms governing slip on the plate interface.

Numerous factors are involved in this widely debated topic of what controls creep and locking in a fault zone. The clear overlap between areas of low P_e and ϕ and high interseismic creep rates in Fig. 6 provides some of the first observational evidence regarding the link between pore pressure and slip behaviour, which we will now discuss in detail.

437 P_e is related to the confining pressure (P_c) and pore-fluid pressure (P_p),
 438 via the effective pressure coefficient, which is dependent on the porosity of
 439 the medium, ϕ , and the critical porosity, ϕ_c , (Biot and Willis, 1957; Nur
 440 et al., 1995):

$$P_e = P_c - \frac{\phi}{\phi_c} P_p \quad (11)$$

441 Assuming in this setting a constant P_c , then P_e is mainly controlled by
 442 ϕ and P_p . The low ϕ (≤ 0.1 , Fig. 6) in the creeping region implies that high
 443 P_p is responsible for the observed low P_e .

444 It is widely suggested that creep is accompanied by increased pore pres-
 445 sure distribution. However, the mechanism linking these two processes is the
 446 subject of debate. One hypothesis suggests that increased pore pressure can
 447 unclamp the fault and enhance the slip (e.g., Kodaira et al., 2004; Tsuji et al.,
 448 2008; Moreno et al., 2014). Whilst, the other argues that the fault frictional
 449 properties enables steady creep that compacts and seals the fault zone by
 450 reducing the permeability and thus causes an increase in pore pressure (e.g.,
 451 Perfettini and Ampuero, 2008; Rubin and Ampuero, 2005; Sleep, 1995). The
 452 data and model presented here are consistent with both hypotheses.

453 Regarding the first hypothesis, high pore pressures could be due to trapped
 454 fluid generated via the dehydration of the colder subducting plate and inter-
 455 action with the hot overlying mantle wedge (Peacock, 1990), which occurs
 456 at temperatures of ~ 200 - 300 ° C (Gao and Wang, 2014; Peacock and Hyn-
 457 dman, 1999), e.g., Fig. 8. A similar mechanism has recently has been sug-
 458 gested in northern Chile using the same empirical relations (Moreno et al.,
 459 2014). The fluid is trapped in this clay-rich environment, scaly clay has been
 460 detected within the plate interface in northeast Japan (Kameda et al., 2015;

461 Chester et al., 2013), which has reduced permeability due to pressure and
462 compaction (Sleep and Blanpied, 1992) and serpentinization (Hilaret et al.,
463 2007). Promotion of creep via pore-fluid pressurization has been observed
464 experimentally (Ujiie and Tsutsumi, 2010) and inferred from geological ob-
465 servations (Mizukami et al., 2014). Tomographic cross sections of this region
466 seem to suggest the presence of low velocity layers due to the presence of
467 fluid from the dehydration of the oceanic lithosphere (Zhi, 2014). Further-
468 more, numerical modelling of fluid flow, based on the northeast Japan sub-
469 duction regime results in a similar pore pressure distribution, whereby low
470 pore pressures persist at the trench, and subsequently increase with depth.
471 This mechanism is not only consistent with substantial creep at depth but is
472 also a possible explanation for the development of strong shallow asperities,
473 such as the one which ruptured in the M_w 9.0, 11th March 2011, Tohoku-oki,
474 earthquake (Yoshida and Kato, 2011).

475 Regarding the second hypothesis, it is argued that the frictional properties
476 govern the pore pressure distribution. Assuming a rate and state framework,
477 the scaling of the nucleation length of a rate weakening region is dependent
478 on the effective pressure. A reduction in effective pressure, due to an in-
479 crease pore pressure, based on equation 11, means the nucleation length will
480 increase. This will stabilise slip in this rate weakening region, as it decreases
481 the chance there is a patch large enough to rupture in an unstable manner
482 (Rubin and Ampuero, 2005). Conversely, in a rate strengthening region it
483 has been shown that if the patch exceeds this nucleation length then it can
484 promote bursts of creep (Perfettini and Ampuero, 2008). Therefore stable
485 slip, or creep can be promoted in rate weakening regions due to a reduction

Figure 8: A conceptual model of the creep mechanism in northeast Japan, based on that of Moreno et al. (2014). As the cooler oceanic lithosphere subducts it becomes dehydrated (blue arrows) and releases fluid which migrates along the fractures in the plate. The fluid becomes trapped due to reduced porosity and permeability (inset) and the clay acts as a weakening mechanism, thus facilitating creep. This part of the plate can also move in a stick-slip manner if the pore-fluid pressure drops. The upper part of the plate, where the pore-fluid pressure is lower, ruptures in large earthquakes.

486 in effective pressure or an increase in fluid pressure. Moreover, consistently
487 high creep rates continually seal the fault zone resulting in higher effective
488 pressures and thus promoting more creep. Pressure dissolution creep is one
489 such mechanism that promotes this behaviour (Sleep, 1995).

490 Interestingly, the area of maximum coseismic slip ($>30\text{m}$) in the Tohoku-
491 oki earthquake coincides with a region of high effective pressure, and high clay
492 content. Again assuming the rate state framework holds this high effective
493 pressure suggests a relatively low nucleation length is needed to promote
494 unstable slip, i.e., an earthquake. Therefore providing further verification for
495 our model as we observe coupling in this region of high clay content and high
496 effective pressure which is prone to unstable slip. Such behaviour has been
497 suggested for this region based on numerical modelling (Cubas et al., 2015).

498 Results from laboratory experiments and numerical models favour the
499 frictional properties governing the slip behaviour (e.g., Cubas et al., 2015;
500 Noda and Lapusta, 2013). Currently it is not possible to ascertain whether
501 the pore pressures or the frictional properties are more dominant, and per-
502 haps both are relevant. The data and model presented here provide strong
503 evidence that there is a link between effective pressure and fault slip, at least
504 on the northeast Japan subduction zone. This is illustrated by our con-
505 cept model in Fig. 8, which summarises the ongoing process and does not
506 distinguish between the relevant underlying mechanisms described above.

507 **6. Conclusions**

508 GPS and CRE observations are complementary datasets, and combining
509 the two allows the recovery of the spatial distribution of creep rate at high

510 resolution across the northeast Japan subduction zone. During the inter-
 511 seismic period (21st March 1996 - 24th September 2003) there is a spatially
 512 heterogeneous creep rate pattern. Three regions are distributed along-strike
 513 at 40 - 70 km depth, creeping at 7-8 cm/yr, whilst the majority of the up-
 514 per part of the plate (0 - 30 km) remains locked. The addition of CRE
 515 data to the GPS improves the model resolution by 40-70%, particularly at
 516 shallow depths. The resulting coupling pattern is not in agreement with var-
 517 ious rupture areas of historic earthquakes. This suggests that the rate-state
 518 asperity model does not fully explain the mechanisms driving slip in this tec-
 519 tonic setting. Patches of reduced porosity and low effective pressure, inferred
 520 from seismic velocities, coincide with the creeping regions. Consequently, we
 521 propose that high pore-fluid pressure and creep are linked on the northeast
 522 Japan subduction zone. Combining GPS and CREs enables us to signifi-
 523 cantly advance our understanding of subduction zone processes. Application
 524 of this approach to other subduction zones will enable the investigation of
 525 creep behaviour and can be used for improving seismic hazard assessments.

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