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Moore, James Daniel Paul; Lamb, Simon; Arnold, Richard

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**Locking on a megathrust as a cause of distributed faulting and ‘fault jumping’
earthquakes**

Simon Lamb¹, Richard Arnold¹, James D.P. Moore²

*1. Institute of Geophysics, Victoria University of Wellington, Wellington, New
Zealand.*

2. Earth Observatory of Singapore, Nanyang Technological University, Singapore.

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Seismicity on large faults is often characterised in terms of an independent recurrence time and magnitude distribution, forming the basis for calculating future earthquake probabilities. The underlying assumption is that the driving mechanism for earthquakes on any particular fault is uniquely linked to that fault, determined by the rate of long term creep on its deeper extension. However, our modelling of nearly 20 years of Global Position System (GPS) data along the obliquely converging plate boundary in New Zealand shows that interseismic stress accumulation can be independent of the properties of the numerous crustal faults, and controlled by locking on the megathrust. This way, the interseismic driving mechanism for large crustal earthquakes is *not* directly linked to the individual major faults that rupture. This scenario predicts large magnitude earthquakes with complex multi-fault ruptures in broad zones that ‘jump’ from fault to fault, following the contours of stress/strain loading. This can explain the November 2016 M7.8 Kaikoura earthquake that shattered the plate boundary in central New

26 Zealand. Repeated episodes of this would create the observed complex array of
27 active faults with the *appearance* of coherent slip. Our analysis opens up the
28 possibility of identifying this type of earthquake behaviour from long-term GPS data.

29

30 In the last 20 years, precise GPS measurements of position have made it possible to
31 constrain the decadal velocity field of actively deforming regions on a spatial scale of
32 10 km or less, with a precision <1 mm/yr, forming a pattern of flow. In New Zealand,
33 there is now a recently published rich database of both campaign GPS¹ and
34 continuous GPS (www.geonet.co.nz) measurements, at a much higher spatial
35 resolution (~ 10 km) and longer time span (~ 18 years) than previously available. Here,
36 the boundary between the Australian and Pacific plates forms a broad zone of active
37 faulting⁴ in continental crust up to 250 km wide, including subduction along the
38 Hikurangi margin, which together accommodate ~ 40 mm/yr of oblique plate
39 convergence^{2,3} (Fig. 1). Estimates of slip rates based on Holocene geological offsets
40 range from 20-30 mm/yr for the Alpine, Hope and Kekerengu Faults, and 3-6 mm/yr
41 for the Clarence, Awatere and Wairau Faults (ref 4).

42

43 Previous models of GPS-derived velocities in active plate boundary zones, both
44 worldwide⁵⁻⁷ and in New Zealand⁸⁻¹⁰, explain the motion as being due to the complex
45 interaction of an array of crustal blocks. The assumption in these models is that each
46 major fault has a geodetic ‘identity’, involving aseismic slip or creep at depth that
47 drives a shallower elastic distortion and eventual earthquake rupture on individual
48 faults^{5-7,11-14}. If this is true, then GPS data would provide information on both the
49 long-term fault slip rate and the depth to the creeping zone and also seismic hazard for
50 individual faults. We show, however, that a model with fewer parameters and

statistically better fit to the GPS-derived velocities in New Zealand requires *no* specific information about crustal faults or blocks.

The decadal GPS-derived horizontal velocity field¹ for the New Zealand plate boundary zone, and the strain rate field calculated from it, are strongly two-dimensional (Fig. 1b,c, see Methods and Supplementary Fig. 2). This is clearly seen in the remarkably tight spread (<5 mm/yr) when *all* the campaign GPS velocities in central and northern South Island, extending for ~ 500 km along the length of the plate-boundary zone, are projected on to a single profile (azimuth 146° , with 330 velocities resolved into shear and shortening components) orthogonal to the Alpine Fault (Fig. 1c). Likewise, contours of strain rate define a straight zone along the length of the plate-boundary, 100 -150 km wide where strain rates $>10^{-7}$ /yr; the axis of maximum strain rate is oblique by $\sim 25^\circ$ to many of the major dextral strike-slip faults (Fig. 1b, see also Supplementary Fig. 2). In an isotropic elastic medium, the contours of the strain rate will be parallel to those of stress loading.

Local Elastic Models

We were motivated by the strong linearity of the strain/stress rate fields to analyse the GPS-derived velocities in terms of a series of local high resolution (100s m) 2-D dislocation models for profiles (forming swaths up to 120 km wide) that run orthogonal to strain rate contours, rather than individual active faults (see Methods for further justification of local 2-D modeling) – note that we assessed the results of this modelling with a coarser resolution (1 – 5 km) full 3-D model for the southern Hikurangi Margin (see below). For each 2-D profile, we projected and resolved the velocities into parallel and perpendicular components (Figs. 1, 2, Supplementary Fig.

1). We used the campaign GPS measurements for horizontal velocities, and the continuous GPS measurements, where available, for vertical velocities (see Methods and Supplementary Fig. 1).

We begin by analysing Profile 3, in the vicinity of the 2016 Kaikoura Earthquake, using a 2-D forward model of elastic deformation¹⁵, where interseismic stress transmission between the two plates is driven by deep slip (at close to the relative plate convergence) along the subduction megathrust (i.e. plate interface) terminating up dip at the ‘locking’ point; above this there is distributed elastic straining without slip on faults (see Methods). The megathrust is defined from seismicity to be approximately at the top of the Benioff Zone¹⁶ (Fig. 2, www.geonet.org.nz) – we then refined its location by allowing small vertical adjustments (± 2 km) in order to obtain the fits in this study. As a first assumption, we only considered slip or no slip on the megathrust – i.e. ‘coupling factor’ = 0 or 1, which in the simplest case slips at the point of locking at the ITRF 2008 GPS Australian/Pacific relative plate motion (derived from ref 3).

We progressively increased the complexity of the model by considering, in addition to a locking point on the megathrust (specified by 3 parameters for the horizontal position, depth and slope of the megathrust at the locking point), the effect of deep creep along individual major active faults or fault zones in the overlying crust, following the approach of block models⁸⁻¹⁰. We searched for the best fit (see Methods for description of misfit criteria), using between 3 and 13 parameters for a range of models, depending on how many faults we included and what parameters we allowed to vary, in order to model 70 data points (i.e. 57 – 67 degrees of freedom, Methods

101 and Supplementary Fig. 3, 4). We restricted ourselves to fault slip rates estimated
102 from the observed Holocene record⁴ – we note, however, that a wider range of slip
103 rates can be made to fit the GPS data equally well (see Methods). Because the faulting
104 in the overlying crust is not always orthogonal to the line of the profiles, models that
105 include deep creep on individual faults should only be considered as a 2-D
106 approximation to the 3-D pattern of faulting. However, adding more free parameters
107 to create a more complex model does *not* improve the goodness of fit (see Methods
108 and Supplementary Figs. 3, 4).

109

110 We assessed the effect of increased complexity in modelling with the Akaike
111 Information Criterion¹⁷ (AIC, see Methods), in which there is a statistical penalty for
112 increasing the number of free parameters. In the absence of additional information, a
113 model with a higher AIC value should not be favoured over one with a lower AIC¹⁷
114 (see Methods). Out of a range of possible models for Profile 3 (see Supplementary
115 Fig. 3), the one with the lowest AIC only fixes a single locking point on the
116 megathrust, together with minor strain accumulation about 250 km back from the
117 trench (accommodating <5% of plate convergence). This passed a chi-squared
118 goodness of fit test with p-value ~0.9 (Fig. 2, reduced chi-squared = 0.77, degrees of
119 freedom = 63, see Methods and Supplementary Table 1) – we consider this our ‘best’
120 model, although we acknowledge that further geophysical research is needed to
121 determine if it is in fact the ‘correct’ model.

122

123 **Deep Megathrust Locking Line**

124 We analysed Profiles 1 and 2 across the southern Hikurangi margin (Fig. 1b and see
125 Supplementary Fig 1, 5) in the same way to the ‘best’ model for Profile 3, with

126 excellent fits (p-values of 0.68 and 0.77 respectively, Supplementary Table 1,
127 Supplementary Fig. 5). This included for Profile 1 the minor residual effect in the
128 longer term GPS velocities of known slow-slip events, adding 1 parameter for the
129 effective coupling factor in the known slow-slip patch¹⁸⁻²⁰, following the approach of
130 ref 15 (see Supplementary Fig. 5a). The family of locking points on the megathrust
131 for successive profiles defines the locking line, which gradually deepens from 23 km
132 in the southern North Island (Profile 1) to 33 km in the northeastern South Island
133 (Profile 3), conforming with the depth of the seismically imaged plate interface (Figs.
134 2, 3, Supplementary Figs. 5, 7).

135
136 We assessed our 2-D models with a simple 3-D model of the megathrust along the
137 southern Hikurangi Margin (see Methods). However, we found it was only practicable
138 to implement a minimization scheme for a 3-D model involving a small number of
139 free parameters, treating the whole plate-boundary as a single rotating elastic block
140 and defining the locking line at least an order of magnitude coarser spatial resolution
141 (1 - 5 km) than that required for the excellent fits in our 2-D models. None-the-less,
142 the 3-D model replicated the main features of our 2-D models, and still achieved
143 acceptable fits with an average velocity misfit of ~0.5 mm/yr (reduced chi-squared =
144 1.13, p-value ~0.05 for 14 parameters and 372 datapoints), further demonstrating that
145 the GPS velocity field is well modelled by a simple locking line along the megathrust
146 (see Methods and Supplementary Table 1, Supplementary Figs. 8, 9).

147
148 We extended our modelling farther south, into the region of oblique continental
149 collision (Profiles 4 and 5, Fig. 1b, Supplementary Table 1, Supplementary Figs. 1, 6).
150 Here, the strong gradients in shear require a shallower locking depth (12 – 17 km)

151 with a low angle detachment that slips at the plate convergence rate and extends
152 northwest towards the Alpine Fault from the point of deeper lithospheric subduction
153 or peel-back, with similarities to the model geometry of ref 21 (Profiles 4 and 5,
154 Methods and Supplementary Figs. 1, 6). This way, oblique continental collision
155 involves underthrusting of the negatively buoyant lithospheric mantle, and an upthrust
156 of the overlying buoyant continental crust, resulting in splitting of the Pacific Plate
157 continental lithosphere at the Moho (Fig. 4, Supplementary Fig. 6). These features are
158 consistent with kinematics inferred from thermochronological studies together with
159 seismic imaging^{22,23}.

160

161 Our combined model for 618 data points has 42 - 44 parameters and passes a Chi-
162 squared goodness of fit test with p-value >0.36 (Supplementary Table 1). We
163 emphasize that this model contains *no* specific information about surface crustal
164 faults. This is in contrast to the large number of individual fault parameters involved
165 in fitting fault block models⁸⁻¹⁰ to the GPS data in the New Zealand plate boundary
166 zone (138 - 217 parameters for 736 – 1872 data points, ref 8 - 10), yet these models
167 fail Chi-squared goodness of fit tests with very small p-values less than 0.0002 (see
168 Methods and Supplementary Table 1 for full details). Furthermore, ref 10 found that
169 the block model solutions for fault parameters were unstable when a priori active fault
170 slip data were *not* included, supporting our conclusion that the GPS data does not
171 contain information about individual crustal fault slip behaviour, but instead
172 constrains the geometry and locking of the underlying megathrust.

173

174 We propose that free slip on the megathrust is controlled by the rheological contrast at
175 the Moho, where the locking line lies for the profiles analysed in this study (Fig. 1,

176 Supplementary Figs. 5, 6), focused at the point at which the subducted slab is in
177 contact with a weaker (and possibly serpentinised) mantle lithosphere of the
178 Australian Plate. The relatively thick crust above the locked part of the megathrust
179 behaves purely elastically, with no internal creep, as a consequence of the ‘cool’
180 thermal regime in the subduction zone²⁴. This dictates the long wavelength
181 deformation in the overlying brittle crust - the latter breaks up along faults with bulk
182 slip rates that relieve the stress loading in the underlying flow^{25,26}. In this case, the
183 zone of potential crustal faulting is determined by the depth at which the megathrust
184 becomes locked – as a rule of thumb, this width is roughly 8 times the locking depth,
185 or 240 km for 30 km deep locking (Fig. 2 and Supplementary Fig. 5).

186

187 **A Regional Stress Driver for ‘Fault Jumping’ Earthquakes**

188 We conclude that large earthquakes in central New Zealand are a response to
189 interseismic stress loading in the brittle crust on a horizontal length scale much
190 greater than the vertical extent of the brittle parts of individual faults. This differs
191 from the common assumption in dynamical generators of synthetic seismicity^{27,28},
192 where the main stress driving mechanism *is* linked to individual faults in order to
193 produce characteristic big earthquakes on these faults and defined long term slip rates.
194 Our results indicate another approach, where the usual far field stress loading
195 assumption for micro-seismic ruptures²⁹ on small faults (m to 100 m scale) is
196 effectively scaled up to large faults (km to 100 km scale).

197

198 The accepted picture of micro-seismicity is that it ‘lights up’ a volume of stressed
199 crust, activating a wide range of available fractures that constrain the orientation of
200 the regional stress field³⁰. The less frequent ‘macro-seismic’ equivalent of this would

201 be the rupture of a wide range of kilometre to tens of kilometre-scale faults which
202 have been brought to the verge of failure by a regional interseismic stress loading that
203 is not linked to any of the faults. We contend that this is exactly what happened
204 during the 2016 M7.8 Kaikoura Earthquake, which was the largest magnitude crustal
205 earthquake in New Zealand since 1855. Rather than rupturing along a single major
206 fault, the region was effectively shattered by at least 21 diverse, and sometimes
207 unconnected, ruptures with widely varying senses and displacements ($<1 - 20$ m) in a
208 volume of crust about 30 km wide and up to 30 km deep, and extending for 170 km
209 oblique to, and sometimes skipping, many of the major faults^{4,31-40}, but, importantly in
210 our view, parallel to the contours of regional interseismic stress/strain loading (Fig. 2,
211 3c). Interestingly, the interseismic locking line determined in this study appears to
212 have been briefly illuminated by the earthquake, defining the up-dip limit of the
213 strongest afterslip on the megathrust (Fig. 3c) in the following four months⁴¹. This
214 was presumably caused by coseismic stress loading on the freely slipping part of the
215 megathrust, although some afterslip extended further up dip to depths of ~ 15 ⁴¹.

216
217 We argue that the ultimate causation for this large scale shattering, with its ‘fault
218 jumping’ behaviour, is the locking on the megathrust, because this creates the
219 necessary conditions prior to failure: (1) purely elastic behaviour (i.e. no creep on
220 individual faults) in a relatively thick wedge of crust, because of the subduction
221 related ‘cool’ thermal regime above the locked part of the megathrust²⁴; and (2) an
222 external mechanism of interseismic stress loading, driven by free slip on the
223 megathrust down-dip of the locked part (i.e. our best fit elastic model). This way,
224 multiple fault planes in the locked crustal wedge are brought very close to failure, as
225 can also be observed with the pattern of micro-seismicity¹⁶ (Fig. 2, Supplementary

226 Fig. 5). If rupture initiates in the crustal wedge as a ‘macro-seismic’ event, then the
227 rupture process itself, involving static and dynamic stress changes, will set off a series
228 of internal ‘fault jumping’ earthquakes that find all the other fractures available for
229 failure³¹⁻⁴⁰ (Fig. 3c). The obliquity of many of the major faults in the overlying crust
230 to the contours of stress loading contours will be a further factor, leading to rupture of
231 numerous cross-linking faults.

232

233 Thus, if a significant proportion of the stress relief is accommodated by slip on faults
234 *other* than the megathrust itself³¹⁻⁴⁰, then large magnitude ‘fault jumping’ earthquakes
235 will be the usual mode of rupture, although we acknowledge that there are other
236 situations in which ‘fault jumping’ behaviour may occur (eg ref 42). Over geological
237 time, repeated episodes like this at different depths in the crust will create a complex
238 3-D array of active faults with time dependent slip behaviour, but the appearance of a
239 coherent slipping fault system⁴³ accommodating the relative plate motion, particularly
240 if faults are relatively immature⁴⁴. However the inferred characteristics of these
241 apparently coherently slipping faults are just short term (i.e. multi-millennial)
242 averages of ruptures induced by repeated shattering. This view is consistent with a
243 growing body of evidence that inferred slip rates on faults both in New Zealand and
244 worldwide, strongly depend on the timescale of observation⁴⁵⁻⁴⁷.

245

246 Many current methods to calculate future earthquake probabilities attempt to ‘reverse
247 engineer’ or synthesize the earthquake behaviour of a deforming zone, either
248 kinematically or dynamically in computationally intensive models, by defining all
249 potential faults and their slip rates and interactions^{27,28,48}. Our analysis suggests the
250 possibility of an alternative approach to earthquake forecasting based on a regional

251 interseismic stress loading independent of the putative characteristics or slip rates
252 of individual faults. This could also apply in tectonic settings without subduction
253 zones, where the regional interseismic stress loading is along the edges of the rigid
254 plates, and may be more realistic and have better predictive power than current
255 methods.

256

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263 manuscript.

264

265 **Author contribution**

266 SL carried out the strain rate analysis and 2-D elastic modelling, wrote the manuscript
267 and created the main figures. RA guided the statistical analysis, and JDPM wrote the
268 computer code for the 3-D modelling; both made comments on the manuscript.

269

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449

450 **Figure Captions**

451

452 **Figure 1. Topography/bathymetry, tectonic setting and strain rate field in New** 453 **Zealand along the boundary between the Pacific and Australian Plates. (a)**

454 Obliquely convergent plate margin cut by complex array of active faults (blue lines).
455 (b) Maximum shear strain rate, smoothed with Gaussian filter (1σ radius = 15 km),
456 with contours trending $\sim 056^\circ$ (2×10^{-7} /yr contour as thick red line), oblique to major
457 active faults (blue lines, ref 4). Depth contours⁴⁸ in km of the subduction megathrust
458 shown as thick green lines. Profile swaths analysed in this study are numbered P1 –
459 P5 (see Supplementary Table 1, Supplementary Fig. 1). (c) Shear and shortening
460 components of 330 campaign GPS velocities projected on to a single profile trending
461 orthogonal to the Alpine Fault (see projection swath in (a)).

462

463 **Figure 2. 2-D elastic model for interseismic deformation along profile across New**

464 **Zealand plate-boundary (Profile 3 in Fig. 1b).** Model involves free slip at relative
465 plate motion along megathrust (thick black line), defined by top of Benioff Zone (ref
466 16), terminating up dip at locking point. Coloured lines show components of model
467 surface velocities; observed decadal GPS velocities denoted by symbols with 1 sigma
468 error bars. Moho after ref 49; pink lines show major crustal faults. The model fit is
469 excellent (see Methods and Supplementary Table 1). Note rupture area of the 2016
470 M7.8 Kaikōura Earthquake.

471

472 **Figure 3. Trace of megathrust locking line, GPS-derived strain rate field, and**
473 **M7.8 2016 Kaikōura Earthquake ruptures.** Model locking line and depth with (a)
474 shortening ($\partial v/\partial y$) and (b) shear ($\partial u/\partial y$) components of the strain rate, relative to
475 Alpine Fault (056°), smoothed with a Gaussian filter ($1\sigma = 10$ km). Kaikōura
476 Earthquake surface ruptures shown as thick red lines in (a). Note region of slow-slip
477 events (SSE, ref 19, 20). (c) Zone of surface ruptures (thick blue lines) for Kaikōura
478 Earthquake is subparallel to locking line (black line) and contours of shear strain rate
479 (red lines). Note region of strong afterslip immediately down-dip of locking line.

480

481 **Figure 4. Three dimensional perspective view of modelled geometry of deep**
482 **slipping surfaces that drive interseismic deformation (see Figs 2 and 3).** Locking
483 line (thick red line) separates freely slipping (light brown region) from locked (light
484 blue region) parts of megathrust, bifurcating at the southwestern edge of subducted
485 oceanic plate (transition between subducted oceanic and continental Pacific plate
486 shaded light green). Both locking line and rupture zone of 2016 M7.8 Kaikōura
487 Earthquake cross-cut major crustal faults. Oblique continental collision also involves
488 slip on a mid-crustal detachment beneath the Southern Alps (see Supplementary Fig.
489 6).

490

491

492 **Methods**

493

494 **1. Calculating velocities from continuous GPS time series (Supplementary Fig.** 495 **1), and estimating uncertainties in campaign GPS data**

496

497 We use the daily solutions reported by Geonet (www.geonet.org.nz) for continuous
498 GPS stations (Supplementary Fig. 1) for individual components of displacements
499 (east, north, vertical). These are given in the ITRF 2008 reference frame. Velocities
500 are calculated using the methodology in ref 15, where time series for the individual
501 stations are first ‘detrended’ over the entire period of operation, then the ‘detrended’
502 series from different stations throughout the network are stacked to bring out network
503 wide noise and artefacts. The stacked ‘detrended’ series are then subtracted
504 individually from each station time series, resulting in significant noise reduction and
505 removal of network-wide artefacts.

506

507 Velocities are determined by finding the best fitting slope, with uncertainties, for the
508 entire period of operation of each time series, and only velocities for stations in
509 operation for >3 years are used in this study. The longest period of operation is ~15
510 years, with an average duration of ~8 years. Velocities are reported in the ITRF 2008
511 Australian Plate reference frame using the data in ref 3, but leaving out the Auckland
512 cGPS site which may be affected by elastic strain accumulation in the New Zealand
513 plate boundary zone. Details of methodology in processing cGPS time series are
514 available from Geonet (www.geonet.org.nz) for the raw data, and Euan Smith
515 (Euan.Smith@vuw.ac.nz) for calculation of long term velocities and uncertainties.

516

517

518 We used the quoted 1-sigma uncertainties in ref 1 for our statistical analysis of our
519 model fits to the data. In our analysis of model fits in 2-D swath profiles, we treated
520 the uncertainties in the east and north components of velocity as independent because
521 no correlation coefficients are given in ref 1 (ie parallel and perpendicular 1-sigma
522 uncertainties respectively are given by the radii of the error ellipse (with principal
523 axes E-W and N-S) in the directions parallel and perpendicular to the profile.

524

525 **2. Calculating strain rate fields (Figs. 1b, 2, Supplementary Fig. 2).**

526

527 Strain rates are calculated using the methodology of ref 50. We construct the
528 Delauney network for GPS sites (Fig. S1) using the TRIANGULATE routine in
529 Generic Mapping Tools⁵¹ (GMT), and then assume constant velocity gradients within
530 each triangle. The strain rates are then gridded at ~2 km spacing, and smoothed with a
531 Gaussian Filter with a 1σ radius ~ 10 or 15 km (GRDFILTER in GMT⁵¹), based on
532 the 10 - 20 km spacing for centres of the individual triangles for the campaign data.
533 Given 0.5 -1 mm/yr uncertainties in individual velocities, we anticipate uncertainties
534 in strain rates $< 5 \times 10^{-8}$ /yr. However maximum strain rates are determined by both the
535 baseline length of the triangles and the degree of smoothing. Smoothed strain rates
536 have a maximum value $\sim 5 \times 10^{-7}$ /yr, compared to peak strain rate ‘spikes’ $\sim 10^{-6}$ /yr for
537 small triangles (sides $\ll 5$ km) in the unsmoothed data.

538

539 Near the coast, at the edges of the onshore GPS data, strain rates are generally
540 extrapolated by allowing triangle sides to be long enough to include linking coastal
541 points. The effect of any bias here appears minimal because strain rate contours are

542 strongly two-dimensional onshore reflecting the two-dimensional nature of the
543 velocity field itself.

544

545 For orthogonal horizontal axes x, y , with components of velocity u, v parallel to these
546 axes respectively, the components of shear strain rate are defined²: $\gamma_1 = \frac{\partial u}{\partial x} - \frac{\partial v}{\partial y}$,

547 $\gamma_2 = \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x}$, and the maximum shear strain rate $\Gamma = (\gamma_1^2 + \gamma_2^2)^{0.5}$. Note that Γ is an

548 invariant of the strain rate tensor if vertical gradients of horizontal velocity, and
549 horizontal gradients of vertical velocity, are zero.

550

551 **3. Elastic Modelling of GPS velocities**

552

553 **3.1 Twist-down method**

554 For each profile, we model the GPS velocities in terms of a strike-orthogonal section
555 through the mid point of a dipping dislocation that extends for 1000 km along strike
556 in a homogeneous elastic half space with a flat surface.

557

558 We use the twist-down method of ref 15 in order to satisfy the constraint of no
559 relative vertical motion of the bounding plates (see also refs 52, 53). This is
560 essentially a forward model where stress transmission between the two plates is a
561 transfer of torque, but calculated using the conventional back-slip method⁵⁴ except for
562 the fact that the virtual ‘back-slip’ fault is now defined as the circular (in cross-
563 section) extension of the freely slipping portion to the surface, tangential to the freely
564 slipping part close to the locking point, but horizontal at the surface.

565

566 The elastic response to back-slip on the virtual fault is calculated using the Okada
 567 formulation for slip on a rectangular dipping dislocation element^{55,56}, and we
 568 approximate a curved fault by constructing the curvature out of numerous small
 569 adjacent elements. The twist-down method was extensively tested in ref 15 by
 570 comparing it with conventional back-slip approaches¹⁸, and was found to give the
 571 best-fit to the GPS data, including vertical velocities that conventional methods
 572 cannot easily fit (see Fig. 7 in ref 15 for a comparison with the conventional back-slip
 573 approach). For Profile 4 at the transition between subduction and continental collision
 574 (see Fig. 1, and Supplementary Fig 1), we found that the twist-up method¹⁵ gave a
 575 better fit (Supplementary Table 1), where we model an upward torque for the Pacific
 576 Plate (as opposed to a downward torque in the twist-down method) in the crust.

577

578 The exceptional fits in our models in this region strongly suggest that any effect of
 579 variable elastic constants⁵⁷ is small.

580

581 **3.2 Goodness of fit criteria**

582 We follow the standard definition of goodness of fit between our fitted model
 583 (predicted values Mod_i , $i=1, \dots, n$) and the corresponding observations (Obs_i), using
 584 the reduced Chi-squared value $= (1/\nu) \sum[(\text{Obs}_i - \text{Mod}_i)^2/\text{var}_i]$, where $\nu = n-p$ = degrees
 585 of freedom (df, number of data points – number of parameters), and var_i is the model
 586 based observational error variance predicted for observation Obs_i . The smaller the
 587 reduced Chi-squared value, the better the fit, and a perfect model would have a
 588 reduced Chi-squared close to 1. We can test the goodness of the model fit using the
 589 Chi-squared test. We compute the p-value, which is the probability that the reduced
 590 Chi-squared statistic will exceed its observed value, assuming the fitted model is true.

591 If the p-value is large, where ‘large’ is conventionally taken to be in excess of 0.05
592 (5% probability), then the model is said to pass the goodness of fit test. We note that
593 the models in this study have p-values in the range 0.05 – 0.9, with an overall value of
594 0.34 – 0.5, for the 2-D models indicating very good fits, and 0.05 in the 3-D model
595 indicating a good fit (see Supplementary Table 1). In contrast the published block
596 models of the GPS velocity field have p-values less than 0.0002 or 0.02%
597 probability⁸⁻¹⁰, indicating poor agreement between these models and the data (see
598 Supplementary Table 1).

599

600 We tabulate the reduced chi-squared values, degrees of freedom and p-values for the
601 Wallace et al. (2004, 2007, 2012) models, as well as our own, in Supplementary Table
602 1. We note that the reduced chi-squared value cannot be interpreted without reference
603 to the degrees of freedom, or computing the accompanying p-value. If the degrees of
604 freedom of the model are ν , then the reduced chi-squared statistic has a limiting
605 Normal distribution with mean 1 and standard deviation $\sqrt{2/\nu}$. Consequently as ν
606 increases, then tolerable deviations from 1 (in the sense of being consistent with the
607 proposed model) become smaller and smaller.

608

609 **3.3 Akaike Information Criterion (AIC)**

610 The Chi-squared value can always be reduced, if only slightly, by including additional
611 parameters in a model. In order to select a model which balances model goodness fit
612 (measured by the Chi-squared value) against unwarranted complexity (large numbers
613 of parameters) we use the Akaike information criterion¹⁷. AIC is defined as: -2
614 $\ln(\text{maximised likelihood}) + 2p$ (where p is the number of model parameters) and is
615 derived from information theoretic considerations. Improvements in model fit

616 increase the value of the maximised likelihood, thus reducing AIC, however where
617 these come at the cost of additional parameters the second term increases the AIC
618 value. The trade-off between the two terms in AIC enables its use to select the
619 simplest model which best describes the data among a set of competing models.

620

621 The value of AIC contains no information about goodness of fit, but it is instead a
622 relative measure of model adequacy to be used to compare competing models. A
623 model with a smaller AIC is preferred over one with a larger AIC. An AIC model
624 comparison does not have the status of a formal statistical test with a p-value, and
625 models with AIC values within two or three units of each other on the AIC scale can
626 be considered equally good.

627

628 The models we fit in this paper assume fully modelled Gaussian errors, and as such
629 the maximized log likelihood is equal to $(-1/2) \sum[(\text{Obs}_i - \text{Mod}_i)^2/\text{var}_i]$ so that the AIC
630 has the simple form $\text{AIC} = \text{constant} + \text{Chi-squared} + 2p$. We evaluate the AIC
631 criterion for models of varying complexity in this study in order to select the ‘best’
632 model (Supplementary Fig. 3).

633

634 **3.4 Complex elastic models with creep at depth on crustal faults**

635 We use Profile 3 (Figs. 1, 2, Supplementary Figs. 4, 5) to assess a suite of models
636 with varying complexity. For this, we approximate the major crustal faults as trending
637 orthogonal to the profile essentially to infinity. The surface position of each fault is
638 constrained by its surface trace where it intersects the profile, but the dips are less
639 certain, but assumed to be between vertical to 60°. In addition there is slip on the
640 megathrust, which has a geometry given by the top envelope of Benioff Zone

641 seismicity, but adjusted within ± 2 km to achieve the best fit (Fig. 2, Supplementary
642 Figs 4, 5). The megathrust is locked at the locking point, with, in the simplest model
643 (3 free parameters defining the horizontal position, depth and slope of the megathrust
644 at the locking point), free slip down-dip of the locking point at a rate given by the
645 relative plate convergence rate³ for the horizontal projection of this point.

646

647 The first step in increasing complexity is in essence a simple ‘offset’ correction. In
648 order to conform with block type models, we interpret this in terms of a small
649 component of distributed strain rate about 250 km from the trench, in NW Nelson that
650 is not associated with subduction (defined here by 4 adjustable parameters: location,
651 locking depth, and two components of slip – see Fig. 2 and Supplementary Figs. 4 and
652 5). It could also be in part an artefact of assuming simple elasticity with homogeneous
653 elastic constants (see section 3.1, ref 57), reference frame problems, or 3-D effects.
654 Thereafter, the complexity of the model is progressively increased by adding addition
655 blocks/faults involving in each step a further 4 free parameters, although we adjusted
656 just two of these (i.e. two of either locking depth, fault dip, or slip rate within
657 Holocene geological estimates for any particular fault), up to a total of 13 free
658 parameters. For each model we calculated the AIC value as defined in Methods
659 section 3.3 (Supplementary Fig. 3). It is clear that allowing only for a very small
660 amount of strain rate (accommodating <5% of total plate convergence) towards the
661 northwestern end of the profile yields both the lowest Chi-squared value, and AIC
662 value, and is therefore considered the ‘best’ model. Increasing model complexity fails
663 to improve to the goodness of fit, but also incurs an AIC penalty of twice the increase
664 in parameters.

665

666 For the purposes of AIC-based model selection, we fix slip rates on crustal faults
667 close to those indicated by geological studies of Holocene fault displacements⁴.
668 However, we found that a wide range of slip rates will also provide the same fits.
669 Thus, if deep slip on several faults *is* explicitly included in the elastic models (i.e.
670 block type models), it is clear from our simple 2-D modelling of Profile 3 (see
671 Supplementary Fig. 3) that there is no unique combination of slip rates or locking
672 depths that will provide the best fit to the data, because for any particular crustal fault,
673 there is a trade off between slip rate, fault locking depth and dip in term of its
674 potential contribution to the GPS-derived velocity field. We emphasize that models
675 which include the major crustal faults are not required, and are *not* the ‘best’ models
676 in terms of AIC or goodness of fit.

677

678 **3.5 Local 2-D versus full 3-D modelling**

679 Ref 15 showed that 2-D models of elastic dislocations are very insensitive to effects
680 that are more than twice the maximum depth of the dislocation out of the plane of the
681 section. For example, modelling Profile 3 assuming that the megathrust only extends
682 for 100 km either side of the profile, with no slip beyond that only changes the fit by
683 ~1 mm/yr. However, it is clear that the velocity field in the plate-boundary is strongly
684 two dimensional for several hundred kilometres along its length, and the megathrust
685 extends for thousands of kilometres NE of the study area. The profile that is likely to
686 be most affected by 3-D effects is Profile 4, at the transition between subduction and
687 oblique continental collision, and we note that this also shows the worst fits in our
688 analysis (see Supplementary Table 1).

689

690 For example, GPS velocities¹ extending for 500 km along strike from central South

691 Island to southernmost North Island, for 660 data (330 velocities), when projected on
692 to a single profile trending 146° , only show a spread <5 mm/yr at any point in the
693 profile (Fig. 1b, c). Extending the length of the projected profile for 100 km further
694 into southern North Island only increases the maximum spread to ~ 10 mm/yr. Thus,
695 velocity gradients along the length of the plate-boundary zone in our study are small
696 (<2 mm/yr/100 km). We tested this conclusion with a full 3-D model for the southern
697 Hikurangi Margin, from southernmost North Island to northern South Island, in the
698 region underlain by the subducted Pacific plate.

699

700 We constructed the 3-D model in the following stages. (1) We used our best-fit 2-D
701 models to determine a series of cross-sections of the 3-D geometry of the megathrust
702 that closely conforms with the Benioff Zone¹⁶. (2) We then created a triangular mesh
703 surface for the megathrust, with typical 5 km lengths for the sides of the triangles. (3)
704 We defined a locking line in terms of 8 adjustable nodes that separates the freely
705 slipping part of the megathrust, down-dip of the locking line, and the up-dip fully
706 locked part. (4) We calculated the elastic response to back-slip in an elastic half space
707 for the locked part, calculating with a single Euler pole the horizontal projection of
708 megathrust slip vectors for the centroid of each triangular patch⁶¹ on the megathrust.
709 (5) We included the possibility, as in our 2-D models, of some residual strain
710 accumulation away from the subduction zone by defining a single plate-boundary
711 ‘block’, rotating about an Euler Pole E_{pbz} . Thus, the residual strain accumulation is
712 defined by the Euler Pole E_{res} , which is the difference between the ITRF 2008 relative
713 plate Euler pole for the Australian and Pacific plates $E_{Pac/Aus}$ and E_{pbz} (ie $E_{Pac/Aus} = E_{pbz}$
714 $+ E_{res}$). (6) We also included the possibility of residual effects from known slow-slip
715 events down-dip of the locking line in the vicinity of southern North Island by

716 defining an effective coupling factor here (0.1 – 0.2). (6) Finally, we could make a
717 single vertical adjustment (± 500 m) of the entire megathrust mesh to allow for a
718 systematic uncertainty in the depth of the megathrust. (7) We sought a minimization
719 solution between the observed and modeled GPS velocity vectors, given their
720 uncertainties, through an adjustment procedure of the locking line, E_{res} , $E_{\text{Pac/Aus}}$ within
721 its 1-sigma uncertainties, slow-slip effective coupling, and depth of megathrust. The
722 overall programme, even at the relatively low resolution of 1 – 5 km, was
723 computationally intensive, requiring 9 days of CPU time to find a best-fit solution
724 (see Supplementary Table 1 and Supplementary Figs. 8, 9). Out of 190 GPS-derived
725 velocities, 186 were used in the model (372 datapoints), with the remaining 4 rejected
726 as outliers (at the 3σ level), fitted with 14 free parameters (8 locking line nodes, 2
727 Euler poles (3 + 1 parameters), 1 SSE effective coupling factor, 1 depth correction).
728 In Supplementary Figs. 8 and 9 we note good agreement between the model and data
729 vectors, with a reasonably smooth locking line. The diversions in the locking line near
730 the SSE likely relate to the fractional coupling expected between partially coupled
731 regions (SSE) and the locking line due to stress considerations⁶² not captured in our
732 simple model.

733

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735

736 All data used in this paper are from publications fully cited in the text. Extensive use
737 of Generic Mapping Tools⁵¹ was made to process data and create the figures.

738

739 **Code Availability**

740

741 Details of the computer code used to calculate models of interseismic deformation can
742 be obtained from SL (2-D modeling in Fortran) and JDPM (3-D modeling in Matlab).

743

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745

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(c) Combined GPS velocity profile