Earth and Planetary Science Letters 463 (2017) 333–343

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

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Detecting hazardous New Zealand faults at depth using seismic velocity gradients

S. Ellis [a,*], R. Van Dissen [a], D. Eberhart-Phillips [a,b], M. Reyners [a], J.F. Dolan [c], A. Nicol [d]

[a] GNS Science, PO Box 30-368, Lower Hutt, New Zealand
[b] Department of Earth and Planetary Sciences, University of California Davis, Davis, CA 95616, USA
[c] Department of Earth Sciences, University of Southern California, Los Angeles, CA, USA
[d] Department of Geological Sciences, University of Canterbury, Private Bag 4800, Christchurch, New Zealand

ARTICLE INFO

Article history:
Received 28 September 2016
Received in revised form 19 January 2017
Accepted 26 January 2017
Available online 22 February 2017
Editor: A. Yin

Keywords:
faults
earthquakes
seismic velocity
New Zealand
hazard

ABSTRACT

Many large damaging earthquakes occur along previously unmapped faults, because it is difficult to locate active faults that have slow average slip rates, long recurrence intervals, and weak surface expression. We use recently collected seismic wave velocity data from New Zealand to test whether there is a strong correlation between seismic velocity gradients deep in the earth's crust, known active faults, and large shallow historical earthquakes. The correlation with active faults is significant at the 99% confidence level, suggesting that seismic velocity gradients at depth can pinpoint active – and in some cases unmapped – faults that may reactivate and rupture in infrequent earthquakes. In addition, all eight of the post-1840 $M_w > 7$ upper crustal earthquakes in New Zealand within the region of good tomographic coverage are spatially correlated with mid-crustal seismic velocity gradients and ruptured faults that intersect them. Many of the seismic velocity gradients coincide with the faulted edges of strong blocks within basement rocks, consistent with these marking preferred sites for fault reactivation owing to inherited strength contrasts. We propose that seismic velocity gradients provide a means to map potentially hazardous undiscovered faults at mid-crustal depths, in advance of their activation in future damaging earthquakes.
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1. Introduction

Numerous recent large earthquakes causing significant casualties and economic loss have occurred along previously unknown faults (Scientists USGS, SCEC, 1994; Jackson et al., 2006; Quigley et al., 2010; England and Jackson, 2011). Detecting such faults is of critical importance for understanding seismic hazard and mitigating risk. Our ability to detect hazardous faults typically relies on evidence from tectonic geomorphology, historical seismicity, and geodetic estimation of strain-rates (Cornell, 1968; Allen, 1975; Matsuda, 1977; Wesnousky et al., 1984; Sagiya et al., 2001; Field et al., 2014; Stirling et al., 2012). Yet in most locations, the historic record is far too brief to provide a full picture of potentially hazardous faults, and assigning geodetically measured strain rates to individual faults is difficult at plate boundaries where faults are closely spaced. Moreover, many faults move only infrequently, and accumulate strain slowly between earthquakes. If the average displacement-rate is slow relative to the rate of surface processes, such as burial by active sedimentation or erosion, the surface expression of such faults can be obscured. Finally, an entire class of faults ("blind" thrust faults) can exhibit no surface expression if sediment accumulation rates are faster than fault-related folding uplift rates (Shaw and Suppe, 1996; Dolan et al., 2003; Leon et al., 2009). Additional strategies are required to identify such undetected faults.

The problem of unexpected, damaging earthquakes has been particularly pronounced in places with short historic records such as New Zealand. There, only half of the onshore $M_w > 7$ earthquakes since historic records began in 1840 showed prior evidence for recent surface rupture (Nicol et al., 2016); i.e., half of the 12 historic shallow, upper plate large earthquakes occurred along previously unknown and unmapped active faults (Fig. 1; Table S1). This includes the 2010 M_w 7.1 Darfield earthquake and its aftershocks that destroyed central Christchurch (Quigley et al., 2010). Moreover, it is not always clear whether shallow geomorphic features extend down into the deep crust, providing a large surface area for fault rupture; large earthquakes commonly nucleate at mid-crustal depths, near the brittle-ductile transition where the most elastic energy is stored.

Seismic velocity gradients in the earth's crust have been observed across many major active faults with large cumulative slip (e.g., McNally and McEvilly, 1977). Here we test whether seismic

* Corresponding author.
E-mail address: s.ellis@gns.cri.nz (S. Ellis).

http://dx.doi.org/10.1016/j.epsl.2017.01.038
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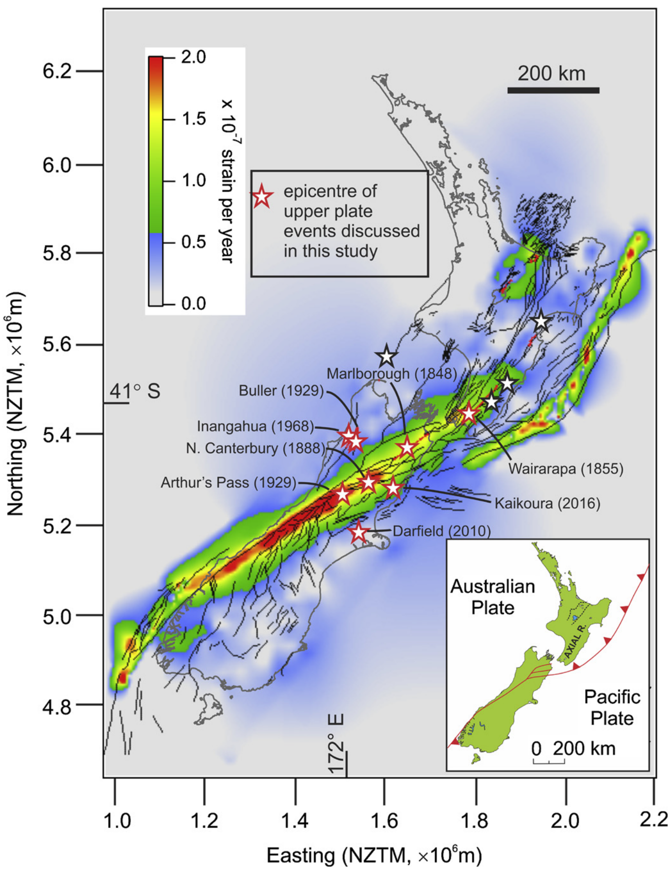

Fig. 1. Location of major historic upper plate New Zealand earthquakes relative to zones of plate boundary strain accumulation. Inset shows the tectonic setting of New Zealand. Colour contours are maximum shear-strain rate interpolated from a combination of GPS velocities observed from 1993–2008 and relative plate convergence placed onto Hikurangi and Fiordland subduction trenches (updated from Beavan et al., 2007). White bold stars show large historic upper plate earthquakes $>M_w$ 7 (Supplementary Table 1); lower plate and subduction interface events have been excluded. The stars outlined in red are located in regions with adequate tomographic coverage to be compared with mid-crustal discontinuities, labelled by earthquake, and discussed in this paper. Gray lines show coast outline and black lines are active fault traces (Litchfield et al., 2014). Axis coordinates are in New Zealand Transverse Mercator Easting and Northing (in meters).

velocity gradients can be used in New Zealand to systematically map rheological changes at earthquake nucleation depths, including low-slip as well as high-slip faults, and both strike-slip and dip-slip regimes. We demonstrate that there is a notable correlation between velocity gradients, locations of known large shallow historic earthquakes, and mapped active faults. Consequently, our data support the mapping of seismic velocity gradients as a prospective tool to pinpoint unknown and potentially hazardous fault zones at depth.

2. Data and method

2.1. Mapping seismic velocity variations at mid-crustal depths

A national seismic velocity model has recently been developed for New Zealand based on earthquake tomography (Eberhart-Phillips et al., 2010). To visualize changes in velocity structure in the mid-crust, we plot a 3D representation of this model by contouring the depth to a V_p isosurface that is typical for quartzofeldspathic rock ($V_p = 6$ km s^{-1}). Plotting a velocity isosurface reveals more details than a V_p depth slice because it represents a three-dimensional image of the transition from soft to hard rock at depth (Table 1; Fig. 2d vs 2e).

The New Zealand velocity model has been determined for an irregularly spaced grid where grid spacing takes into account the station and source distributions used to derive the velocities (Fig. 2a–c; Eberhart-Phillips et al., 2010; Eberhart-Phillips and Bannister, 2010; Reyners et al., 2014; Eberhart-Phillips and Reyners, 2012). To avoid effects of mesh distortion during contouring, we interpolate all the (x, y, z) velocity points from the New Zealand-wide data onto a new mesh with dimensions of $5 \times 5 \times 2$ km and x and y mesh axes aligned E–W and N–S. This is then plotted as a structured mesh using the Paraview software (Ahrens et al., 2005) which is also used to construct each V_p isosurface and to compute the horizontal gradients in isosurface depth (Fig. 2e, f).

We tested the effect that the New Zealand-wide velocity model mesh spacing has on results by constructing artificial examples of velocity gradient variations with depth in two and three dimensions (Fig. S1). The artificial examples have a mesh spacing of 1×1 km that is sub-sampled at the resolution of the New Zealand-wide model. These tests illustrate that, despite the coarse mesh spacing, the New Zealand-wide velocity model is sufficient to reproduce contours of constant V_p, although gradients may be misplaced by an amount up to the horizontal mesh spacing (2.5 to 5 km), and discontinuities are smeared and volumetrically averaged over a length-scale dependent on the grid resolution (e.g., Wagner et al., 2012).

Excessive smoothing or error from resolution issues can be estimated by using a spread function (Michelini and McEvilly, 1991) as calculated for the New Zealand-wide model to determine how strong and peaked the resolution is for each inversion node (Eberhart-Phillips and Bannister, 2010). The spread function is related to both grid and data resolution. We mask out areas with spread function >2.5 in Fig. 2d, e and f. For the New Zealand-wide model outside of the Canterbury region, the gradient in V_p isosurface we calculate relates to features within ∼2–5 km distance, since the locations of the velocity gradients are constrained within 2–5 km in the regions of adequate resolution. In areas with sparse

Table 1
Relevant rheological properties of typical New Zealand rocks.

Typical rock type	Sediment	Greywacke	Granite	Schist[a]	Basalt	Gabbro	Peridotite
Sample density kg m^{-3}	≤2500	2640 ± 100	2640 ± 80	2730 ± 115	2770 ± 160	2884 ± 150	3090 ± 225
Calculated $V_p^{(1)}$ km s^{-1}	≤4.3	5.5	5.3	6	6.1	6.5	7.1
Calculated $V_s^{(2)}$ km s^{-1}	≤2.6	3.3	3.2	3.6	3.7	3.9	4.3
Calculated shear modulus$^{(3)}$ GPa	≤17	29	27	35	38	44	57
C&M$^{(4)}$ V_p km s^{-1}	≤5	5.3	6.2	6.2	6.7	7.1	8.2
C&M density$^{(5)}$ kg m^{-3}	≤2600	2615	2650	2830	2990	2966	3310

Notes: The top row lists sample density ranges of New Zealand rock types (Tenzer et al., 2011) (not shown is mafic eclogite, Reyners et al., 2014, which typically has densities around 3500–3600 kg m^{-3} and V_p around 8.2 km s^{-1}, Christensen, 1996). Calculated V_p (1) use Gardner et al. (1974) for $V_p < 5.5$ km s^{-1} and Hill (1978) for $V_p > 5.5$ km s^{-1} (Eberhart-Phillips and Bannister, 2010). (2) The table assumes a constant V_p/V_s ratio of 1.65 (typical for most regions, however see Eberhart-Phillips and Bannister, 2010). (3) Shear modulus $= \rho V_s^2$. (4) For comparison, shows V_p from a global compilation, Christensen and Mooney (1995), and (5) densities. *C&M* refers to Christensen and Mooney (1995).

[a] Note that schist velocities are significantly anisotropic (Christensen and Mooney, 1995).

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