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Episodic kinematics in continental rifts, modulated by changes in mantle-melt fraction in the underlying diverging mantle flow: evidence from GPS data

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Abstract. Mantle-melt production is widely ascribed to decompression in an upwelling mantle flow¹⁴. Direct surface manifestations for this mantle flow are, however, difficult to find. Here, we show from >10 years of campaign and continuous GPS measurements that the back-arc continental rift system in North Island, New Zealand accommodates a far field extension of 6 - 15 mm/yr, yet a ~70 km long segment of the rift axis is associated with a strong horizontal contraction and rapid subsidence. Both flanking and along strike of the contractional axis are coeval extension and uplift. These features fit a simple model of flexure of an elastic upper crust pulled downwards or pushed upwards along the rift axis by a driving force in the viscous lower crust or uppermost mantle at a depth >15 km. We propose that flexure of the crust is a response to a transient relaxation flow after a rapid change in melt fraction, and hence mantle viscosity, of the diverging mantle beneath the rift axis. An increase in viscosity due to melt extraction leads to subsidence, and a decrease in viscosity after melt has accumulated permits uplift, a process that is also likely to occur in oceanic spreading centres. This way, we show that it is now possible to monitor the time and length scales of mantle melt production in continental rift settings with modern GPS methods.

Introduction

Along oceanic spreading centres, melt production is well modelled by decompression in an upwelling mantle flow¹⁴. However, the process is difficult to observe directly in the oceans. Continental rifts, however, are amenable to detailed measurements of the short-term kinematics. We show these data can provide evidence for an upwelling mantle flow, as well as the length and timescale of mantle melting.

In New Zealand, the Havre back-arc basin, which extends for several thousand kilometres behind the Tonga-Kermadec subduction zone, terminates in the northern part of North Island. Here, the mid ocean ridge transitions into a continental rift system in a region of prolific and active basaltic to rhyolitic, and mainly andesitic volcanism since ~2 Ma, referred to as the Taupo Volcanic Zone²⁸ (TVZ, Fig. 1). Since 1995, a series of campaign measurements of position have been made using the Global Positioning System (GPS) for a dense network³ (~5 – 10 km scale) spanning the TVZ, with a sparser network (~20 km scale) of continuous GPS stations³⁰ since 2000 (Extended Data Fig. S1a, b).

Short term deformation

Recent geodetic analyses of the TVZ have focused on the vertical motions, which have been interpreted in terms of a mix of diverse shallow level magmatic processes such as crystallisation and/or dewatering driving subsidence^{6,7,19}, and intrusion driving uplift¹⁰. We integrate the vertical *and* horizontal motions, and we show that these can be explained by a single process of ductile flow in the lower crust and uppermost mantle.

The rate of extension across the TVZ increases along its length, towards the NE, from ~6 - ~15 mm/yr (Fig. 3b and c), with a component of dextral strike-slip and clockwise rotation^{3,13,24,26} (Extended Data Fig. S1a,b). However, the horizontal dilatation field, defined as rate of change of surface area per unit surface area, derived from 1995 – 2011 GPS campaigns³ and continuous GPS data²⁹ up to 2015 (Fig. 2, Methods and Extended Data Figs. S1, S2) shows pronounced symmetry with an axis of horizontal contraction (-ve dilatation) in the central part of the TVZ between Lakes Taupo and Rotorua, noted also by ref 6, with a width of ~20 km and length of ~70 km, surrounded by zones of strong expansion (+ve dilatation, Fig. 2, 3).

The pattern of vertical velocities for both the campaign⁸ and continuous GPS networks²⁹ is also symmetrical and co-axial with the zones of horizontal contraction and expansion, with peak subsidence at 15 – 20 mm/yr (Figs. 2, 3, Extended Data Fig. S2), but flanked by a narrow axis of uplift at 0.5 - 2 mm/yr to the northwest and

southeast, and strong uplift (up to 4 mm/yr) to the northeast and southwest where we also find strong dilatation ($>2 \times 10^{-7}$ /yr).

Elastic models of deformation

Given the close spacing of active normal faults in the TVZ⁷, we would not anticipate strain accumulation due to deeper slip on faults to fit the marked variations in strain rate across the rift. Even in two extreme models¹³, where strain accumulates on either one or two antithetic faults locked down to ~ 7 km, the data are not well matched (Fig. 3a, Extended Data Fig. S3a, b), and does not explain the variation in uplift and strain rate along the length of the TVZ.

An alternative approach is to consider horizontal contraction or expansion as a consequence of elastic flexure, superimposed on far field extension (Methods and Extended Data Fig. S4a, b). The subsidence due to this extension must be negligible (< 1 mm/yr) because there is uplift along the length of the TVZ. We model the flexure as downward and upward axial loads, with a specified width, applied to a uniform infinite elastic plate. The shape and magnitudes of the fields of dilatation and vertical motion can be well matched by flexure of a ~ 11 km thick plate on a viscous substratum, with a 2.75 km thick elastic core²⁷ (T_e); the region of contraction and subsidence is pulled down at 13 mm/yr by a 60 km x 8 - 15 km load along a 60 km segment of the rift axis (Fig. 3a, b) – note the pattern of observed and modelled flanking uplift. Uplift and expansion near the coastal end of the TVZ are well modelled by an adjacent upward axial force, pushing up at ~ 4 mm/yr (Fig. 3a, c). The widths of the loads are determined by the assumption of a uniform traction, and there will be an equally good fit, given the spread in the observations, for a more realistic broader and symmetrically variable traction²¹.

We note that elastic half space models, used previously^{8,10} to explain short term vertical deformation in the TVZ, *cannot* explain the coeval flanking axes of uplift, either side of the axial region of subsidence, or flanking subsidence either side of the axial regions of uplift, which are distinctive characteristics of flexure of an elastic plate on a viscous substratum (Fig. 2, Methods and Extended Data Figs. S6 and S7), or viscous flow in a confined layer (Extended Data Fig. S8). However, 2-D elastic

half space velocity models – which are similar to 2-D half space viscous flow models if gravitational forces play no role - do provide support for a deep driving force (>15 km deep) to explain the symmetrical surface deformation (see Methods and Extended Data). The observed rate of flexure is the response to either time varying changes of mass within the flexed plate, or an underlying deep viscous flow – note that density changes due to cooling or magma crystallization within the plate⁹ cannot drive flexure because they do not result in any change in mass, and dewatering would drive upward flexure rather than subsidence.

A ~11 km thick flexed plate is consistent with the depth of seismicity along the rift axis (Fig. 1b), and seismic velocity and magnetotelluric studies which require any ductile flow or widespread partial melting to be deeper than 10 km, below an upper crust (5 – 10 km depth) with microseismicity, high Q, high Vp/Vs and high resistivity (Fig. 1b), except for very localized shallow plumes of low resistivity beneath major geothermal systems^{2,4,5,10,11,21} (Fig. 2, Extended Data Fig. S6). The surface expressions of the major geothermal systems appear more closely associated with uplift and expansion than subsidence and contraction^{5,20,21} (Extended Data, Fig. S7).

Viscous coupling with underlying mantle flow

In the absence of rapid changes in suitable plausible surface loading along the TVZ, we consider a flexural model with an axial viscous flow forces *beneath* the flexed plate. For a ductile flow at the rates and wavelength of flexure, the force must be deeper than the base of the plate by an amount comparable to the width of the equivalent uniform axial load, or 8 – 15 km, placing the force in the uppermost mantle at a depth of ~20 km. This is comparable to the depths suggested by the best-fitting half space elastic or viscous models (Extended Data Fig. S6).

We propose that temporal changes in a basal vertical ‘suction’ force along the axis of an upwelling mantle flow, which has been invoked to explain 1 – 2 km deep mid ocean-ridge axial grabens^{16,18,22}, leads to symmetrical regional patterns of downward and upward flexure along the TVZ. The origin of the vertical basal force is rapid extensional strain rates at the point where the upwelling flow diverges, with a magnitude determined by the strain rate and viscosity of the flow^{16,18}. The force is

transmitted to the brittle upper crust by flow in the intervening viscous lower crust, causing uplift or subsidence at a rate determined by both lower crustal and mantle viscosities (Figs 4, Methods and Extended Data Figures S4c, S8).

Over time, flexural stresses, together with gravitational potential energy contrasts due to topography in the rift, will provide a resistance to the vertical basal force proportional to displacement, giving rise to an exponential decay $e^{-t/\tau}$ with time t of subsidence or uplift rates, where τ is the decay constant, as the system evolves to equilibrium²³ (see Methods and Extended Data, Figure S9b). Thus, any process that perturbs the vertical basal force, such as a change in the mantle flow viscosity, will trigger a transient and exponentially decaying flow in the lower crust, leading to either subsidence or uplift as the rift adjusts to a new state that can be supported by the perturbed vertical basal force (Methods and Extended Data, Figure S9b). Mantle viscosity is sensitive to the melt fraction, with an exponential relation¹² (Methods section 5), and so relatively small changes in this along the axis of the upwelling mantle flow would be a powerful mechanism to cause viscosity, and hence basal force, perturbations (Fig. 4).

If the effective mantle viscosity undergoes an increase, because previous melt extraction has created a low melt fraction residue, then this will drive a transient flow in the viscous lower crust as the topographic relief in the rift and flexural stresses readjust by deepening the rift axis in response to the new basal force, resulting in a broad symmetrical pattern of subsidence, horizontal contraction and flanking extension, at an initial rate determined by the ratio of the viscosity of the lower crust to change in viscosity of the mantle in the axis of the upwelling mantle flow (see Methods and Extended Data Fig. S4c, S9). However, if an increase in melt fraction near the base of the crust – either along grain boundaries¹², or in the form of distributed pockets¹² or local pooling - has reduced the effective viscosity of the underlying mantle, then the lower crustal transient viscous flow will drive uplift and a reduction in topographic relief in response to the decrease in the basal force. Note that even if the finite amplitude of these changes is quite small, the rate of change of surface elevation (>1 mm/yr) may be high enough to be measured geodetically.

Recent basaltic to andesitic volcanism²⁸, or evidence for melt accumulation at depth⁹ indicate mantle melt at depth beneath the northern and southern parts of the TVZ. But beneath the axis of the intervening ~70 km long zone of downward flexure, there is no active basaltic volcanism²⁸. Thus, temporal and spatial changes in mantle melt fraction and transport along the TVZ could explain synchronous rift axis uplift and subsidence, modulated by flexure of the brittle crust.

Finite element modelling of the transient viscous flow due to a change in vertical basal force given the current rate of subsidence in the central TVZ, and assuming that it initiated relatively recently - last few hundred years – requires the ratio of lower crustal viscosity to ‘perturbation’ in mantle viscosity to be in the range $\pm 0.3 - 0.9$, for a plausible depth to width ratio >1 of the surface over which the vertical basal force acts beneath the TVZ – note the flanking upward flow in the viscous layer (see Methods and Extended Data Fig. S8, S9). The length of the subsiding zone gives the horizontal spacing of melt extraction and accumulation, of the order of 50 – 100 km, which is comparable globally to typical active volcano spacing²². The relatively small topographic relief (<200 m) within the central depression of the TVZ implies that the current maximum subsidence rates here cannot have been maintained for >10 ka; long term episodic subsidence is also indicated by the maximum ~3 km depth to basement²¹, which would be reached in <0.2 Ma at current rates compared to the much older ~2 Ma age of the TVZ²⁷. This and the maximum rate of subsidence are consistent with a lower crustal viscosity of $0.5 - 1 \times 10^{19}$ Pa s (ref 17) and mantle viscosity that changes by $\pm (0.15 \text{ to } 0.9) \times 10^{19}$ Pa s or 15% – 90% of a reference mantle viscosity of 10^{19} Pa s equivalent to a melt fraction change of 0.3 – 3% (ref 12,17), or a subsidence/uplift decay constant τ of ~1000 years after melt extraction or accumulation (see Methods and Extended Data, Section 4). Finally, extraction and migration of melt may be a 3-D process, with both horizontal and vertical focusing from a large volume, and a lateral transfer of melt accumulated over periods >10 ka – this may also be part of the explanation for synchronous uplift/subsidence along the length of the TVZ.

We suggest this is part of the active process that creates the oceanic crust at spreading ridges. However, in the TVZ, a deep axial graben has not formed as a long term topographic feature, partly because it is being continually filled by volcanics, but also because the viscous lower crust and underlying mantle flow are too weak to support it. But the sensitivity of mantle viscosity to the presence of melt means that perturbations in mantle melt fraction can influence the kinematics of the overlying continental rift, opening up the possibility of using surface geodetic data to monitor mantle melt production in the upwelling mantle flow.

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Figures

Figure 1. (a) Tectonic setting of Taupo Volcanic Zone (TVZ, outlined in gray box) in North Island, New Zealand, at the southern termination of back-arc spreading in the Havre Trough behind the Hikurangi Margin. Black crosses show earthquake epicentres. Projections of earthquakes, topography and subsidence rates in a swath along Profile AA' shown in (b) and (c). Axis of rift is seismically active, located ~100 km above subducted Pacific Plate.

Figure 2. Map of central and northern part of Taupo Volcanic Zone (box in Fig. 1a) showing dilatation strain rate (rate of change of surface area per unit surface area with +ve expansion, -ve contraction) calculated in this study from combining ~16 years of campaign GPS velocities³ with 3 to 15 years of continuous GPS measurements²⁸ (Methods and Extended Data Fig. S2), defining a symmetrical zone of strong horizontal contraction up to 20 km wide and ~70 km long, following the axis of the back-arc rift, surrounded by zones of extension. Contours of subsidence/uplift rate (based on both the campaign⁸ and continuous²⁹ GPS data) are also symmetrical and co-axial with pattern of dilatation (solid black circles are campaign GPS sites, white circles are continuous GPS sites). Thick grey dashed line shows axis of uplift (0 – 2 mm/yr) ringing the region of subsidence. Thin red lines show major active faults⁷. Thick dashed red boxes show profile swaths projected along AA' and BB' (Fig. 3, thick solid red lines).

Figure 3. (a) Modelled horizontal dilatation (colour coded) and contours of vertical motion for a flexed 11 km thick plate with an effective elastic thickness T_e of 2.75 km, pulled down in the central part at 13 mm/yr by a rectangular load that is 60 km long and 8 km wide, with additional upwardly directed loads either end of the downward force. A dilatation has been added to the model between $x = \pm 50$ km to account for rifting in the TVZ increasing linearly in rate from ~6 to ~14 mm/yr towards the NE (y direction of model flexed plate), from 0.6×10^{-7} /yr at $y = -80$ km to 1.4×10^{-7} /yr at $y = +80$ km across a width of 100 km (See Methods and Extended Data Fig. S5). Compare the model with the observed axes of uplift (heavy dashed grey lines) and 2 mm/yr subsidence contour (thinner grey line) - see Fig. 2. (b) and (c) Comparison between horizontal (green crosses) and vertical (black crosses) modelled

velocities and dilatation (red crosses) for Profiles AA' and BB' in (a), and horizontal (red squares) and vertical (blue circles) components of GPS velocities with 1-sigma uncertainties projected onto Profile AA' and BB' in Fig. 2 – horizontal velocities are resolved parallel to the line of profiles.

Figure 4. (a) Diagram illustrating a patch of basal normal deviatoric stress along the axis of the underlying upwelling mantle flow beneath the back-arc rift of the Taupo Volcanic Zone^{10,21}, which creates a vertical force on an overlying elastic plate depending on the viscosity of the diverging part of the mantle flow. (b) If the mantle viscosity undergoes an increase, because melt extraction has created a low melt fraction residue, then this will drive a transient flow in the viscous lower crust as the topographic relief in the rift and flexural stresses readjust by deepening the rift axis in response to the new basal force, resulting in a broad symmetrical pattern of subsidence, horizontal contraction and flanking extension, at an initial rate determined by the ratio of the viscosity of the lower crust to change in viscosity of the mantle in the axis of the upwelling mantle flow (see Methods and Extended Data Fig. S4c, S9). (c) However, if melt accumulation near the base of the crust has reduced the viscosity of the underlying mantle, then the lower crustal transient viscous flow will drive uplift and a reduction in topographic relief in response to the decrease in the basal force.

Fig. 1

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Methods

1. Calculating velocities from continuous GPS time series (Figs. 3, Extended Data Fig. S1, S2).

We use the daily solutions reported by Geonet (www.geonet.org.nz) for continuous GPS stations in the vicinity of the TVZ (Fig. S1) for individual components of displacements (east, north, vertical). These are given in the ITRF 2008 reference frame. Velocities are calculated using the methodology in Lamb and Smith (2013), where time series for the individual stations are first ‘detrended’ over the entire period of operation, then the ‘detrended’ series from different stations throughout the network are stacked to bring out network wide noise and artifacts. The stacked ‘detrended’ series are then subtracted individually from each station time series, resulting in significant noise reduction and removal of network-wide artifacts.

Velocities are determined by finding the best fitting slope, with uncertainties, for the entire period of operation of each time series, and only velocities for stations in operation for >3 years are used in this study. The longest period of operation is ~15 years, with an average duration of ~8 years. Velocities are reported in the ITRF 2008 Australian Plate reference frame using the data in Altamimi et al. 2012, but leaving out the Auckland cGPS site which may be affected by elastic strain accumulation in the new Zealand plate boundary zone. Details of methodology in processing cGPS time series are available from Geonet (www.geonet.org.nz) for the raw data, and Euan Smith (Euan.Smith@vuw.ac.nz) for calculation of long term velocities and uncertainties.

2. Calculating strain rate fields (Figs. 2, Extended Data Fig. S2).

Strain rates are calculated using the methodology of Lamb (2015). We construct the Delauney network for GPS sites (Fig. S1) using the TRIANGULATE routine in GMT (Wessel et al. 2013), and then assume constant velocity gradients within each triangle. The strain rates are then gridded at ~2 km spacing, and smoothed with a Gaussian Filter with a 3σ radius ~ 18 km (GRDFILTER in GMT, Wessel et al. 2013), based on

a ~ 12 km spacing for centres of the individual triangles for the campaign data. Given ~ 1 mm/yr uncertainties in individual velocities, we anticipate uncertainties in strain rates $< 5 \times 10^{-8}$ /yr. However maximum strain rates are determined by both the baseline length of the triangles and the degree of smoothing. Smoothed strain rates have a maximum value $\sim 5 \times 10^{-7}$ /yr, compared to peak strain rate ‘spikes’ $\sim 10^{-6}$ /yr for small triangles (sides < 5 km) in the unsmoothed data (Fig. 3).

For orthogonal horizontal axes x, y , with components of velocity u, v parallel to these axes respectively, the dilatation Δ is defined:

$$\Delta = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}$$

Note that this combination of velocity gradients is independent of the choice of reference frame and defines the rate of change of surface area per unit surface area. Positive dilatation denotes expansion, and negative dilatation denotes contraction. When combining campaign and continuous GPS velocities (Fig. 2 and Extended Data Fig. S2c), 1 mm/yr was added to each component of the continuous GPS horizontal velocities to harmonise reference frames, as suggested by velocities projected into Profile AA’ (see Fig. 3) – this correction has negligible effect on the resultant strain rate field.

3. Modelling both far field extension and elastic flexure (Extended Data Fig. S4b and S5)

We model the far field extension as a constant strain rate over Profiles AA’ and BB’, such that the end velocities match the observed ~ 10 mm/yr across 100 km for AA’ (a strain rate of $\sim 10^{-7}$ /yr) and ~ 14 mm/yr across 100 km for BB’ (a strain rate of $\sim 1.4 \times 10^{-7}$ /yr). This corresponds to a background dilatation gradient for the flexed plate in the y direction for $x = \pm 50$ km of 0.6×10^{-7} /yr at $y = -80$ km to 1.4×10^{-7} /yr at $y = +80$ km (Fig. 3a, S5). We then add to this linear velocity field the velocity field due to elastic flexure (see below and Extended Data Fig. S4b, S5).

4. Flexure of a continuous plate by a rectangular load (Extended Data Fig. S4, S5)

We model the flexure in 3-D of a continuous plate floating on an inviscid fluid, infilled with air, by defining a load of specified width and length (load at $x = \pm \text{half width}$, $y = \pm \text{half length}$) on a horizontal plate with constant thickness. The horizontal plate size is large (400 km x 400 km) compared to the flexed region. The model is reformulated in terms of displacement rate V_z , for a maximum vertical velocity $V_{z_{\max}}$ along the axis of the load, and $V_{\text{horizontal}x}$ is the x component of horizontal velocity, $V_{\text{horizontal}y}$ is the y component of horizontal velocity, and $\epsilon_{\text{horizontal}x}$ is the gradient of velocity (i.e. $\epsilon_{\text{horizontal}x} = dV_{\text{horizontal}x}/dx$). The thickness of the plate is made up of an elastic core thickness T_e , which determines the flexural behavior, and outer parts that have undergone failure, with a total plate thickness of $2\Delta R$, where the neutral plane is at a depth ΔR (Extended Data Fig. Fig. S4a).

$$V_{\text{horizontal}x} = \Delta R \cdot dV_z/dx$$

$$V_{\text{horizontal}y} = \Delta R \cdot dV_z/dy$$

$$\epsilon_{\text{horizontal}x} = \Delta R \cdot d^2V_z/dx^2$$

$$\epsilon_{\text{horizontal}y} = \Delta R \cdot d^2V_z/dy^2$$

Calculations are carried out using the Fortran code written by Garry Quinlan, February 1990.

5. Viscous coupling between crust and upwelling mantle flow (Extended Data Fig. S4c)

Here we compare dynamic pressures induced in an upwelling mantle flow with those require to produce viscous flow. We consider a Newtonian viscous mantle flow (viscosity η_{mantle}) with overlying layer (thickness T_v) of Newtonian viscous lower crust (viscosity η_{crust}) (Extended Data Fig. S4c). We assume that the upwelling mantle flow has a characteristic horizontal length scale W (Extended Data Fig. S4c).

We begin with a simple scaling analysis – see also Phipps Morgan et al. (1987) and Scott and Stevenson (1989). For a total horizontal extension rate of V_{mantle} across a width W , this will induce a negative vertical stress $\sigma_{\text{mantle}} \sim 2V_{\text{mantle}} \eta_{\text{mantle}}/W$. For a vertical velocity, V_{crust} , at the top of the overlying viscous lower crust, there must be flow stresses σ_{crust} in the lower crust $\sim 4V_{\text{crust}} \eta_{\text{crust}}/W$. Thus, assuming there is no restoring force, and the lower crustal flow is driven by the underlying mantle flow, $\sigma_{\text{mantle}} \cdot W \sim \sigma_{\text{crust}} \cdot T_v$ or $\eta_{\text{mantle}} / \eta_{\text{crust}} = 2 \cdot (T_v/W) \cdot V_{\text{crust}} / V_{\text{mantle}}$. For $T_v/W \sim 1$, and $V_{\text{crust}} / V_{\text{mantle}} \sim 1$, this suggests $\eta_{\text{mantle}} / \eta_{\text{crust}} \sim 2$. This simple scaling analysis shows the order of magnitude relation between lower crustal viscosities and mantle flow viscosity at the axis of upwelling to drive subsidence without a restoring force. Below we use a full finite element model of viscous flow.

We employ a finite element model to calculate the flow within a 2-D viscous layer of thickness T_v , representing the viscous lower crust, with a specified normal traction at the base of the layer on a patch with width W , zero Dirichlet basal boundary conditions either side of the patch, and with free sides, using code written by JM. The layer is 50 times wider than thick, with a free surface and no vertical surface displacement. We only consider the flow field at time $t = 0$, so that there is no contribution from gravitational stresses and density plays no role. Thus, the model is specified by the ratio W/T_v , the viscosity of the half space η_{crust} , and the basal normal stress σ_n . We consider extreme ratios of W/T_v in the range 0.1 – 2. Figure S8 shows flow fields for the model for W/T_v of 0.5, 1, and 2. We consider the results in non-dimensional form by comparing W/T_v with the ratio $V \eta / (\sigma_n \cdot T_v)$, where V is the maximum vertical surface velocity. Figure S9a shows a plot of $V \eta / (\sigma_n \cdot T_v)$ with W/T_v for different implementations of the model.

For time $t > 0$, vertical displacement of the surface will create gravitational and elastic restoring forces, analogous to isostatic or post glacial rebound (Turcott and Schubert 1982). We do not attempt to model this with the finite element code, but consider the general anticipated response from viscous rebound theory (Turcott and Schubert 1982) and finite element models with ‘sticky air’ (Crameri et al. 2012). Thus, the surface displacement will evolve until the restoring force balances the vertical basal force (width x basal normal traction). The restoring forces are proportional to

displacement S , and viscous flow forces are proportional to velocity (dS/dt) so evolution of the surface displacement involves a transient flow in the viscous lower crust that decays as an approximate exponential function (Crameri et al. 2012), analogous to isostatic rebound, with a decay constant τ (Turcotte and Schubert 1982, Ranalli 1995). For an initial maximum vertical velocity V_{initial} , then velocity V and vertical displacement D evolve with time t (see Extended Data Fig. S9c):

$$V = V_{\text{initial}} e^{\left(\frac{-t}{\tau}\right)}$$

$$D = V_{\text{initial}} \tau \left[1 - e^{\left(\frac{-t}{\tau}\right)} \right]$$

Effective equilibrium is achieved on a time scale of $\sim 3 \tau$. At this time, any perturbation in the vertical basal force will trigger flow in the viscous lower crust until a new equilibrium is achieved: an increase in the basal force will drive subsidence, and a decrease will drive uplift. The initial rate of uplift or subsidence can then be modelled by a basal force in our viscous flow model, where the traction is the ‘change’ in actual basal driving force (basal normal traction $\times$ width W). In terms of the upwelling mantle flow, the change in basal driving stress is:

$$\Delta \sigma_n \sim 2 \Delta \mu_{\text{mantle}} V_{\text{ext}} / W$$

where V_{ext} is the horizontal velocity of rifting and $\Delta \eta_{\text{mantle}}$ is the ‘change’ in mantle viscosity. Thus, using the scaling relation in Figure S9a for model runs with variable W/T_v :

$$\frac{\mu_{\text{crust}} V_{\text{vertical}} \left(\frac{W}{T_v} \right)}{2 \Delta \mu_{\text{mantle}} V_{\text{ext}}} = \alpha$$

where values of α are shown in Figure S9a plotted against W/T_v . Note that for $W/T_v > 2$, $\alpha \sim T_v/W$. Given that the thickness of the crust in the TVZ is 20 - 30 km, and the neutral surface of the flexed ‘plate’ is at ~ 6 km depth, then the wavelength of flexure requires $W/T_v > 1$ if the basal driving force is at the top of the upwelling mantle flow. For initial subsidence $V_{\text{vertical}} \sim 15$ mm/yr, and $V_{\text{ext}} \sim 10$ mm/yr, this requires

$\mu_{\text{crust}}/\Delta\mu_{\text{mantle}}$ to be in the range 0.3 - 0.9 (Figure S9b). For a crustal viscosity $\mu_{\text{crust}} \sim 10^{19}$ Pa s (Ranalli 1995).

this implies an increase in mantle viscosity of $0.3 - 0.9 \times 10^{19}$ Pa s. We can calculate the required change in melt fraction to cause this using the experimentally-derived exponential relation between viscosity η and melt fraction ϕ (Hirth and Kohlstedt 2003):

$$\eta(\phi) \propto \eta_{\text{ref}} e^{-\alpha\phi}$$

where α is a constant in the range 30 -45. Thus, a change in melt fraction leads to a change in viscosity:

$$\frac{\Delta\eta}{\eta} \sim -\alpha\Delta\phi$$

Taking a reference mantle viscosity of 10^{19} Pa s, the required changes in viscosity calculated above ($0.3 - 0.9 \times 10^{19}$ Pa s) implies a change in melt fraction $\Delta\phi$ in the range 0.3% – 3%.

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Extended Data Figures

Figure S1. 16 years of campaign and up to 10 years of continuous GPS velocities (Beavan et al. 2016, www.geonet.org.nz), this study) in the TVZ, showing the Delauney networks used to calculate the strain rate field.

Figure S2. (a) and (b) Contours of subsidence rate plotted with field of dilatation calculated from the GPS horizontal velocities (see Methods). Black filled circles show location of GPS sites. Campaign GPS 1995 – 2011 with horizontal velocities from Beavan et al. (2016) and vertical velocities from Hamling et al. (2015). Continuous GPS for 3 – 15 year time spans (www.geonet.org.nz, this study). (c) Principal axes of strain rate plotted on field of dilatation calculated from combined campaign and continuous GPS data (see Fig. 2, and Methods).

Figure S3. Models of elastic half-space dislocation models of interseismic strain accumulation for slip on a buried normal faults in the TVZ, using the twist-down method of Lamb & Smith (2013) which satisfies the far field vertical velocity boundary condition with no relative vertical motion between forearc and Australian Plate. Models are 2-D, fitted along Profile AA' in Fig. 2, using the horizontal velocity resolved parallel to the profile. Models with one or two faults result in the greatest perturbation of velocities from a simple linear slope, but still only provide poor fits to the observations (see Fig. 3).

Figure S4. (a) Diagram illustrating how velocities and strain rates are calculated for a simple model of downward flexure of a continuous elastic plate, pulled down by a line load. The shape of flexure is determined by the thickness T_e of the elastic core, whereas the strain rate and velocities are determined also by the half beam thickness, where elastic flexure has occurred but the flexural stresses in the upper and lower portions of the beam have exceeded the yield strength (see also Watts 2001). (b) Resultant velocities (V_{tot}) in models are obtained by adding the velocity solution for elastic models (V_{model}) to a linear increase in velocity across the TVZ to account for far field tectonic extension (V_{ext}). (c) Diagram illustrating geometry of a flexed elastic plate (brittle upper crust), overlying a viscous lower crust (viscosity η_{crust}) with effective thickness T_v pulled down by a basal normal traction σ_n over characteristic

width W above an upwelling mantle flow at velocity V_{mantle} (viscosity η_{mantle}). In this case, $\sigma_n \sim 2 \eta V_{\text{mantle}}/W$.

Figure S5. Flexure of a ~ 11 km thick continuous plate floating on an inviscid fluid with a 2.75 km thick elastic core (Te), pulled down at 13 mm/yr by an axial load (8 km wide and 60 km long). Horizontal dimensions of the plate are effectively infinite compared to dimensions of load (horizontal plate size for flexure calculation = 400 x 400 km) so that edge effects are negligible. (a) Dilatation and vertical motion (mm/yr) of plate. (b) Dilatation and vertical motion (mm/yr) of plate with far field extensional strains added (see Methods, section 3), with a dilatation gradient in the y direction for $x = \pm 50$ km of $0.6 \times 10^{-7}/\text{yr}$ at $y = -80$ km to $1.4 \times 10^{-7}/\text{yr}$ at $y = +80$ km.

Figure S6. (a) and (b) Best fitting elastic half-space model for surface distortion from a buried compacting horizontal planar source (15 km wide x 60 km long) at a depth of 17 km, using the Okada formulation for contractional dislocations (Okada 1985) orthogonal to a horizontal rectangle, for resolved horizontal velocities along Profile AA'. Rectangle in (a) shows horizontal extent of compacting source. (c) Vertical velocity field for elastic layer have compacting source – note that this is similar to flow in a Newtonian viscous half space with uniform density and viscosity, and no surface topography, responding to an internal planar normal traction (see Figures S7).

Figure S7. Map of the TVZ showing the pattern of horizontal dilatation in this study, together with the main Pleistocene and Holocene volcanic complexes (Wilson et al. 1995), geothermal systems (total heat loss ~ 4.3 GW over 6000 km², Bibby et al. 1995), and low resistivity patches at 13 km depth (5 Ωm contour), determined from MT probing (Heise et al. 2010). There is a broad coincidence between the axis of horizontal contraction and the zone of volcanic activity in the last 1 Ma, but not with the main geothermal systems which are offset 10 – 15 to the SE of the contraction axis. A similar lack of correlation is seen with the low resistivity patches, which may reflect deep partial melts or fluid rich zones (Heise et al. 2010).

Figure S8. Finite element models of instantaneous viscous flow in a layer, with a basal patch of normal traction σ_n , for three width to depth ratios W/T_v of the basal patch. The surface is flat, and viscosity is scaled with velocity V , using the scaling $V\eta/(\sigma_n T_v)$. Note streamlines of flow upper panel in each pair, and velocities in lower panel. (a) $W/T_v = 0.5$. (b) $W/T_v = 1$. (c) $W/T_v = 2$.

Figure S9. (a) Scaling relations in viscous flow models illustrated in Figure S7, for maximum surface vertical velocity V , viscosity η , basal normal traction σ_n , and width to depth ratio W/T_v of basal patch, shown by plotting $V\eta/(\sigma_n T_v)$ against W/T_v for model runs. For $W/T_v > 2$, $V\eta/\sigma_n T_v \sim T_v/W$. (b) Ratio of ‘change’ in mantle viscosity to crustal viscosity, normalized by the ratio of velocities of rift extension and maximum vertical velocity. (c) Normalised plots illustrating exponential decay with time t of surface subsidence/uplift rate and subsidence/uplift in transient lower crustal viscous flow for decay constant τ , and maximum initial vertical velocity V_{max} . Decay is driven by evolution of and flexural forces in elastic layer and gravitational potential energy in underlying viscous flow after a perturbation in a driving basal force (see Methods Section 5 and Figure S7).

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