Do flexural stresses explain the mantle fault zone beneath Kilauea volcano?

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SUMMARY
Recent relocation and focal mechanism analyses of deep earthquakes beneath Kilauea volcano, Hawaii indicate that seismicity is concentrated on a horizontal fault zone at a depth of 30 km, with seaward slip of the upper block on a low-angle plane. We discuss whether the observed localization of the earthquakes can be explained primarily by stresses induced by flexure of the Pacific Plate beneath the Hawaiian load. We find that flexural stresses are consistent with the observed fault plane orientation, and the direction and rate of slip. The mechanisms of mantle earthquakes in other regions of Hawaii are also consistent with the flexural calculation. However, the model has four shortcomings: (1) the fault zone is displaced 15–20 km to the NW of the region of predicted maximum shear stress; (2) the maximum shear stress on preferred fault planes in the vicinity of the fault zone seems too low to overcome Coulomb friction (by about a factor of 2, assuming hydrostatic pore pressure); (3) the fault zone is much more localized laterally than is the region of large flexural stresses and stressing rates and (4) the fault zone is more localized vertically than might be inferred from the calculation as well. Simple and plausible extensions of the plate flexure model that account for spatial variations in the location of pore fluids, and/or the possible existence of a passive low shear stress magma transport system can overcome most of these shortcomings. Several magma pipes would be necessary to explain the observed earthquake locations, and simple thermal arguments indicate that such pipes could be conduits for porous flow if they are a few kilometres in radius.

Key words: earthquakes, flexure of the lithosphere, magma flow, mantle.

1 INTRODUCTION
In volcanic regions such as Hawaii, earthquakes are sometimes thought to surround areas being stressed by magma movement. For example, earthquake locations have been used to infer a 36 km vertically continuous magma plumbing system beneath Kilauea (e.g. Ryan 1988). However, because the stress perturbation from actively moving magma is generally small, earthquakes are thought to occur only where the ambient stress is near to failure (Rubin & Gillard 1998). In fact, recent relocation of deep Hawaiian earthquakes and analysis of their mechanisms indicates no seismological evidence for a vertically continuous pathway, but that instead earthquakes are localized on a series of horizontal planes, with the most extensive at 30 km depth with seaward slip (hereafter called the mantle fault zone) (Wolfe et al. 2003, 2004, Fig. 1). In fact, there are several quasi-horizontal fault planes beneath Kilauea and the south flank area (figs 9 and 14 in Wolfe et al. 2004).

In this paper, we discuss whether this deep planar fault zone can be explained by primarily tectonic stresses due to flexure of the Pacific lithosphere by the Hawaiian load. Our hypothesis is that earthquakes occur in regions that are close to failure because of the regional stress field, and that then experience a small stress perturbation (relative to the ambient stress field) that pushes the area to failure. These additional small stresses could have several possible causes including magma movement, temporal changes in the flexural stress field itself, and large decollement earthquakes (between the volcanic pile and the oceanic crust, Fig. 1) which have been shown to affect the rate of deep earthquakes (Klein et al. 1987). It is not obvious that there are pre-existing horizontal planes of weakness 20 km below the oceanic crust, although Wright & Klein (2006) propose that the mantle fault zone is localized along the diffuse phase boundary between plagioclase- and spinel-peridotite or a region of neutral buoyancy. In particular, we seek to test whether a plate flexure model can explain the location and rate of seismicity, the orientation of the fault plane, and the direction of slip, or if the observations require additional sources of stress. We focus on the mantle fault zone beneath Kilauea because it dominates the mantle seismicity (42 per cent of all events greater than $M_L > 3$ between 1959 and 2000, Fig. 2b, Wolfe et al. 2003), although we also mention the relation between flexural stress.
and other areas of significant mantle seismicity which are shown in Fig. 2.

2 BACKGROUND

Plate flexure has been suggested to be the dominant source of stress at mantle depths beneath Hawaii (e.g. Rogers & Endo 1977; Klein et al. 1987), and a possible cause for the mantle fault zone (Wolfe et al. 2003, 2004). Other sources of stress are certainly important at less than 10 km (the depth of contact between the Pacific Plate and the Hawaiian extrusive load, e.g. Delaney et al. 1990). However, to reach failure by frictional slip using the Coulomb failure criteria at the relevant depth requires differential stresses of the order of hundreds of MPa with hydrostatic pore pressure. Of proposed mechanisms, only flexure appears to be capable of generating an ambient stress state of this magnitude.

Many models for the deflection of the Pacific Plate under the Hawaiian load have been proposed, encompassing a range of rheological properties (see review and references in, Watts 2001). The most recent results give estimates of the effective elastic thickness of the plate that ranges from 25 to 44 km, depending on the model assumptions (e.g. Watts & ten Brink 1989; Wessel 1993). All of
these models have constrained some parameters using near-surface observations (gravity, bathymetry, depth to subsurface interfaces and rate of subsidence/uplift), usually along single profiles across Oahu. The location and mechanisms of deep earthquakes provide information on the orientation and absolute value of stress at depth, and therefore, are an additional data set that can be used to assess the validity of the flexural models.

3 METHOD

We model the plate flexure assuming an isotropic, homogeneous, unbroken elastic plate of uniform thickness on a sphere, although the curvature effects are small (see list of model parameters in Table 1). We use a lithospheric thickness of 40 km, because in a uniform thickness model this value most closely matches the gravity profile in the region of interest near the Big Island. We consider only surface loading of constant density with a slightly lower density of sediments and volcanic deposits filling topographic depressions (Table 1). This model is obviously an oversimplification—for example, it predicts failure of the plate in several locations, and the solution does not include the effects of this (e.g. Melosh 1976). However, as yet, there is no model for flexure at Hawaii with a realistic constitutive law for the lithosphere that is consistent with all available data (e.g. even recent models are not consistent with the subsidence history recorded at the Hawaiian Drill Hole, e.g. (Zhong & Watts 2002)). Therefore, this model is a first-order test of the feasibility of plate flexure as an explanation of the characteristics of the mantle fault zone.

We use the semi-analytic solution of Melosh for the equations of flexure for a compressible thick elastic plate over an inviscid substratum (Melosh 1976, 1978; Cyr & Melosh 1993). The densities used are slightly modified to allow for seamount loads—the load density is reduced by the water density and the mantle density is reduced by the infill density (e.g. Comer 1983), although these modifications do not change our conclusions (parameters in Table 1). The thin-plate equations are not appropriate for subsurface stress calculations, because they assume that principal stresses are always horizontal and vertical. While Melosh’s equations account for buoyancy restoring forces from displacements of density interfaces, they neglect internal body forces which may have a second-order effect upon the stress calculations (e.g. Ward 1984).

We use the actual topography and bathymetry of Hawaii (ETOPO2, www.ngdc.noaa.gov/mgg/image/2minrelief.html) to load the plate, after removing a model of the long-wavelength swell, calculated using a series of super-Gaussian functions (Wessel 1993) (Fig. 3). Each gridpoint in the topography above a minimum height (100 m) is converted to a cylinder (Cyr & Melosh 1993) of the appropriate height and density (depending on if it is subaerial or subaqueous, Fig. 4), and accounting for the difference in volume of the cylinder compared to the rectangular blocks. For the load from each gridpoint, we calculate the resultant stress tensors for all other gridpoints using Melosh’s axisymmetric equations, integrate over the wavenumber domain in Matlab and then resolve the tensors into Cartesian coordinates. We repeat the procedure for all gridpoints tracking the cumulative sum of the stress tensors. We assume a lithostatic stress state, with a single density for the elastic plate (Table 1), which is added to the diagonal components of the stress tensor. This model assumes zero shear stress at the top of the elastic plate, which may be appropriate for a weak decollement interface between the plate and the topographic load of Hawaii. On the other hand, if the volcanic load is mechanically welded to the lithosphere, the shear stresses in the upper lithosphere are reduced (e.g. McGovern & Solomon 1993), but the impact is less at the depth of the mantle fault zone. We have benchmarked our method with thin-plate analytic solutions for an equivalent large cylinder.
Figure 4. The load from Fig. 3(b) is separated into subaqueous (a) and subaerial (b) loads, shown as height above the reference seafloor and sea surface, respectively. The loads are summed to calculate the stresses and deflection of the surface of the plate. (c) The deflection of the ocean floor (negative equals subsidence) is shown in colour.

Figure 5. Results from plate flexure model shown along the profile from Fig. 1(a), where the black line is the topography, and the white dots are the earthquakes from that profile. (a) We show the ratio of shear stress on favourably oriented planes (assuming a frictional coefficient of 0.6) to the normal stress on those same planes (assuming no pore pressure). The dotted line shows sea level. (b) Angle (in degrees) between the most compressive stress ($\sigma_1$) and a horizontal plane (i.e. the dip of the mantle fault zone). (c) The pore pressure necessary to induce slip on horizontal planes as a fraction of lithostatic pressure. A friction coefficient of 0.6 is assumed.

4 RESULTS

We present cross-sections and a map view of our flexural model calculations in Fig. 5. As in most previous plate flexure models, we predict compressional failure near the surface and extensional failure directly beneath the load, while stresses are far from failure in the middle of the plate—the so-called ‘neutral plane’ (Fig. 5a). In Hawaii, the vast majority of earthquakes appear to be located within the upper 10 km of the crust. Most of these are thought to occur within the volcanic edifice or at the contact with the pre-Mauna Loa seafloor sediments (i.e. above the ‘plate’ in Fig. 5a) (Klein et al. 1987). However, an unknown fraction may occur within the region of predicted compressional failure in the uppermost portion of the plate (Got & Okubo 2003). There are many fewer earthquakes in the vicinity of the neutral plane, with the important exception of the mantle fault zone at 30 km and a few other mantle earthquakes.

Although the stresses predicted by the flexure model are not by themselves large enough to cause earthquakes in the mantle fault zone, the ratio between shear and normal stress on favourably oriented planes is of the proper order of magnitude (Fig. 5a). Moreover, the model predicts approximately the correct fault orientation and slip direction. Fig. 5(b) shows the angle of the maximum compressive stress from horizontal. If we assume faulting is occurring on a horizontal plane and use Byerlee’s law, the angle between the maximum compressive stress and the horizontal should be about 30°, while the model predicts angles between 40° and 60° in the vicinity of the mantle fault zone. It has been observed that some earthquakes occur on faults closer to 45° from the inferred maximum compressive stress (e.g. Thatcher & Hill 1991), although it is not clear that
the explanations Thatcher & Hill (1991) proposed would be relevant to Hawaii. A horizontal fault is also the only geometry that is potentially ‘self-perpetuating’ in a steady-state sense, as the Pacific Plate sweeps to the NW beneath the island load (assumed stationary with respect to the mantle source region). That is, once formed, the stress concentration at the leading edge of the mantle fault zone would facilitate its propagation into unfractured rock moving in from the SE. The direction of slip on the mantle fault zone, as indicated by the focal mechanism in Fig. 1(a), is also similar to the direction predicted by the model in Fig. 6(b), as first noted by Wolfe et al. (2004).

Despite these successes, there are some shortcomings of the model presented, as follows:

1. The mantle fault zone is displaced 15–20 km to the NW of the region of predicted maximum shear stress on horizontal planes (Fig. 6a), which is also the region closest to failure at these depths (Fig. 5a), although this region is not very close to failure (see next point).

2. The maximum shear stress in the vicinity of the mantle fault zone, ~100 MPa, seems too low to overcome Coulomb frictional failure (for friction coefficients > about 0.2 and hydrostatic pore pressure).

3. The mantle fault zone is much more localized laterally than is the region of large flexural stresses (Fig. 6b).

4. The mantle fault zone is more localized vertically than might be inferred from the calculation as well (Fig. 6a).

The limited vertical extent (point 4) need not be a serious shortcoming, since slip within the fault zone reduces to some extent the shear stress on shallower and deeper fault-parallel planes, and because the depth of the fault zone does coincide with that of the maximum predicted shear stress on horizontal planes (Fig. 6a). In fact, the relocations of Wolfe et al. (2003) show some secondary, quasi-horizontal fault planes deeper than the primary fault zone by roughly 5 and 10 km, and shifted slightly to the southeast. Regarding points 1 and 2, both the offset in map view from the location of maximum shear stress on horizontal planes, and the insufficient magnitude of the computed shear stress, might be artefacts of an inadequate constitutive model (e.g. the neglect of both viscoelasticity and brittle failure elsewhere in the plate). However, both these processes seem likely to perturb the stress field on a length scale that is too large to explain the limited lateral extent of the mantle fault zone (point 3). Therefore, we search for an explanation of how to bring the fault zone to failure that also can explain the localization in map view.

5 DISCUSSION

5.1 Other stressing mechanisms

5.1.1 Pore pressure

Perhaps the simplest additional mechanism to invoke for bringing the fault zone to failure would be large pore pressures. Fig. 5(c) shows the pore pressure (as a fraction of the lithostatic load, defined as \( p_g z \), the density times the gravitational acceleration and depth) needed to cause failure on a horizontal plane, assuming a coefficient of friction of 0.6. At the depth of the mantle fault zone, a pore pressure that is 50–80 per cent of lithostatic is required. If these elevated pore pressures were also localized in map view, they could explain the localization as well as the occurrence of failure. As the mantle fault zone lies directly beneath the active volcanoes of Kilauea and Mauna Loa, magmatic volatiles are the obvious candidate for localized pore fluid pressures, although the distribution of such fluids with depth is not well constrained. Primary Kilauea magmas have been estimated to contain roughly 0.35 wt per cent \( \text{H}_2\text{O} \) (Clague et al. 1991) and 0.7 wt per cent \( \text{CO}_2 \) (Gerlach et al. 2002). From fig. 6(c) of Dixon (1997), such melts would be expected to become saturated at roughly 900 MPa, close to 30 km depth. As these figures pertain to some average melt, given a compositionally heterogeneous source region individual magma batches could become vapour-saturated at still greater depth. If magmas are not saturated at the depth of the mantle fault zone, then an elevated pore pressure here would likely require ponding and freezing of magma below.

The existence of long-period earthquakes and harmonic tremor at depths of 30–50 km is indicative of the presence of magmatic fluids, and very likely of volatiles. These events appear to be concentrated to the north and south of the mantle fault zone beneath Kilauea (Klein et al. 1987; Koyanagi et al. 1987); however, the numerous earthquakes that occur on subhorizontal fault zones along the south coast, southwest of the main mantle fault zone (figs 2 and 14 in Wolfe et al. 2004) are closely located in space with the long-period events (Klein et al. 1987). Long-period earthquakes and harmonic tremor at shallow depth are thought to be produced by the vibrations of cracks hosting bubbly magmas, with the bubbles providing the impedance contrast necessary for the crack to ‘ring’ (Chouet 1996). At mantle depths an exsolved vapour phase would be highly compressible, so this mechanism becomes unlikely (although a variant that involves collapsing bubbles might be feasible). Julian (1994) suggested that tremor might result from the dynamic flow of incompressible magma through an elastic crack, but a more complete assessment of this mechanism by Balmforth et al. (2005) suggests that this would require a combination of large flow speeds (>10 m s\(^{-1}\)) and small crack apertures (<1 cm) that seem unreasonable for a silicate liquid. Thus, current models of tremor all seem to require a free volatile phase (although it should be borne in mind that Balmforth et al. (2005) did not consider irregular crack geometries).
5.1.2 Magma transport system–dykes

A second possible explanation for both the occurrence and localization of failure is the perturbation to the flexural stress field due to the magma transport system. On the ∼10-km length scale of the mantle fault zone, time-dependent stresses due to dyke propagation or porous flow are expected to be relatively small, perhaps 1–10 MPa (e.g. Rubin 1998). Here we are more concerned simply with the existence of the transport system. Whether magma rises through the Hawaiian lithosphere by porous flow or dyke propagation is unknown (see Appendix A), but in either case we expect a local stress perturbation of the same order of magnitude as the flexural stress field, as we illustrate now.

The geometry of the deep (>10 km) magma plumbing system beneath Hawaii is not well constrained, but will affect the stress state in several ways. First, for a purely flexural stress state below the Hawaiian load, with compression at the top of the plate and extension at the bottom, dykes would not be able to reach the surface. Within the shallow, compressive regions, dyke ascent requires the intrusion and freezing of dykes at greater depth, which increases the horizontal tensional stress at shallower levels. This process is analogous to the mechanism by which a single fracture can propagate through a bent beam, which, prior to the approach of the crack tip, has a compressive stress state on the side opposite to that of crack initiation. Furthermore, it is an underestimated point that even the tensional stress at the base of the flexed plate is not conducive to dyke ascent. Beneath Kilauea, this stress increases with depth at a rate of ∼20 MPa km−1 (Fig. 7). Superimposed on a lithostatic background, this gives rise to a least compressive stress that increases at only ∼10 MPa km−1. As the static pressure gradient in a column of magma increases at ∼25 MPa km−1 (Fig. 7), this makes the magma effectively much ‘denser’ than the host rock in the lower half of the plate, and dykes rising from the source that encounter this stress field will be driven laterally rather than vertically (e.g. Lister & Kerr 1991; Rubin 1995). If the stress field were computed for a more realistic elastic–Coulomb–viscous rheology, the same conclusions would apply within the elastic–brittle portion of the plate, with the dykes spreading laterally at the level of the long-term brittle–viscous transition as shown in the viscoelastic model of McGovern (2006). The final lateral and vertical extent in these cases would depend upon the magma supply characteristics and freezing. Because each frozen dyke increases the adjacent dyke-perpendicular stress essentially to the level of the magma pressure, while decreasing the dyke-perpendicular stress at shallower levels, after several generations of intrusion dykes might be able to penetrate the plate. This superposition of several generations of dykes will cause a stress perturbation of the same order of magnitude as the flexural stress field (perhaps greater very near the edge of the swarm), but with an unknown, presumably complex geometry.

5.1.3 Magma transport system – low shear stress pipe

While such a deep dyke network could have a large impact on the stress state within the mantle fault zone, we explore another component of the magma plumbing system, more amenable to simple analysis, that will also sufficiently perturb the mantle stress field. If there has been enough prior advection of heat (presumably via dyke propagation, at least initially, see Appendix A) that a partially molten “pipe” exists through a portion of the lithosphere, then the differential stress within that pipe should be negligible in comparison to the flexural stresses. A low shear stress pipe could represent a locus of porous flow from the source region to the surface, or could consist of a network of dykes, which could relax shear stresses either by thermally increasing the ductility of the rock or through the relaxation of shear stresses within fluid-filled cracks. We note that there is no unequivocal evidence of lithospheric dyke propagation in the form of deep migration of seismicity or inflationary episodes of Kilauea’s shallow magma reservoir that mirror the dramatic deflations associated with dykes exiting the reservoir, although deep harmonic tremor may be indicative of magma-filled cracks. In addition, the mantle xenoliths associated with pre-shield alkalic lavas (e.g. at Loihi) and post-shield rejuvenated lavas that are indicative of rapid dyke ascent through the lithosphere, are absent from the large-volume flux shield-building stage of Hawaiian volcanoes (Clague 1987). This could indicate that lithospheric dyke propagation does not occur during the shield-building stage, or that such dykes exist but are intercepted by magma reservoirs at crustal depths. A partially molten magma pathway a few kilometres in radius is thermally viable and could plausibly deliver the observed-magma flux at Kilauea (see Appendix A).

For a pipe with a diameter that is small compared to the plate thickness, a rough estimate of the stress perturbation can be obtained by imagining the plate, with its flexural stress field, to be cut by a vertical, shear stress-free tunnel (with the normal stress on the tunnel wall equal to pgz) (e.g. Hiramatsu & Oka 1962). The resulting shear stress on horizontal planes, for an elastic body with a vertical tunnel subjected to a uniform remote stress state (taken from the appropriate location within the flexure calculation for the mantle fault zone), is shown in Fig. 8. While Fig. 8 is in Cartesian coordinates (the x-direction is east), for the sake of discussion we define the local x’-direction to be parallel to the flexure-induced shear traction on a horizontal plane (approximately along the direction with the lowest stresses in Fig. 8). At the margins of the tunnel

![Figure 7](image-url)  
**Figure 7.** Compressive stress (compression positive) as a function of depth below sea level from the flexure calculation (bending stress), compared to the lithostatic load (which has approximately the same slope as a static column of magma), for a location beneath Kilauea. Within the lower half of the plate, the bending stress is approximately the least compressive stress, and is oriented horizontally. This orientation is favourable for dyke formation. However, because the magma driving pressure at Z₁ (defined as the pressure in the static column minus the ambient stress, or the length of the thick line at Z₁) is smaller than the driving pressure at Z₂, small dykes within the lower half of the plate would descend rather than ascend, even though the plate is undergoing tension in this region and the host rock is more dense than the magma.
along the diameter parallel to $\gamma$, the horizontal stress $\sigma_{\gamma\gamma}$ is reduced to zero, as it must be to remain in equilibrium with the boundary condition $\sigma_{\gamma\gamma} = 0$ at the edge of the tunnel. However, at the margins along the orthogonal diameter, $\sigma_{\gamma\gamma}$ is increased to about $2\sigma_{\gamma\gamma}^0$. Although the flexural stresses beneath Hawaii are not uniform, to the extent that they vary slowly on the scale of the pipe diameter this simple analogue problem suggests that a partially molten pipe could nearly double the shear stress on horizontal planes locally. This would reduce the pore pressures required for failure from 0.5 to 0.8 times lithostatic to only 0.5 times lithostatic. In addition, the model predicts that earthquakes should be concentrated in particular azimuths around the pipe, a distribution that is not easily reconciled with the earthquake locations in Fig. 1. Therefore, if such pipes are important in localizing the earthquakes, one might have to invoke a network of them with radii of order a few kilometres. If the magma pipe extended into the upper lithosphere it might promote faulting at certain azimuths. The predicted pattern is not observed in the seismicity and this may indicate complexities in the magma geometry or stress field at shallow depths or shortcomings in the model.

5.2 Large stresses or large stressing rates?

Thus far we have interpreted earthquake locations only in terms of the computed stress magnitudes, without considering what controls the seismicity rate. Significant seismicity requires not only that the regional stress lie on the appropriate failure envelope, but that, if one imagines locking the faults momentarily, the stressing rate would carry the region further outside that envelope. To the extent that earthquake stress drops are independent of stressing rate, a simple assumption is that seismicity rate and stressing rate within a given volume are proportional (e.g. Dieterich 1994). A simple method for approximating the stressing rate is to idealize the stress field as time-invariant in the reference frame of the mantle source region. Then the time derivative of some component of the stress field $\partial \sigma_{ij}/\partial t$ is just $-v(\partial \sigma_{ij}/\partial x)$, where $v$ is the plate velocity.

As we have not computed the 3-D stress field using a model that includes inelastic deformation, we use spatial derivatives from the flexural calculations as proxies for the actual $\partial \sigma_{ij}/\partial x$. In Fig. 9(a), we show the ratio of shear to normal stress on optimal planes in a vertical section beneath the centre of an idealized Hawaiian load (Fig. 9e), and the shear stress on a horizontal plane at 30 km (Fig. 9c).

In Figs 9(b) and (d), we show the derivatives of these quantities in the approximate direction of plate motion. Note, for example, that in Fig. 9(c) there is a U-shaped region with large shear stresses that extends around the island chain, but that significant seismicity would not be expected along the ‘arms’ of the U, because the stressing rate there is much less than that at the leading edge.

Within regions that are close to failure, seismicity should be highest in the regions with the highest stressing rates (red regions in Fig. 9). Two such regions are restricted to shallow depths. The first is shallower than 15 km, ~100 km seaward of the leading edge of the load, where the extensional bending stresses are large and increasing. As there is no permanent seismic network in this region, small earthquakes here may go unrecorded (the threshold for detection may be $M_L > 4$).

The second region shallower than 15 km is beneath the point where the load reaches its maximum elevation, where the compressive stresses are large and increasing. It is unclear if such earthquakes have been detected. Most shallow Hawaiian earthquakes occur above the decollement, shallower than the domain of our flexural model. However, by relocating earthquakes Got & Okubo (2003) identify several steep thrust faults, dipping at $50^\circ-60^\circ$, that are located below what they interpret as the basal decollement beneath the south flank of Kilauea (although, as they point out, the $20^\circ$ dip of the shallow ‘decollement’ illuminated by the microseismicity is too steep to reflect the large-scale dip the decollement). The steep reverse faults are located at ~8–10 km below sea level, shallower than the 10 km depth of the decollement estimated from flexural models (Thurber & Gripp 1988), but given the uncertainty in both the earthquake depths and flexural models, the interpretation of Got & Okubo (2003) is plausible. Thus, these earthquakes might be indicative of radially directed horizontal compression beneath the decollement. In between the two regions shallower than 15 km with large stressing rate, the stress field is near failure, but the stressing rate actually decreases because the stress orientation is changing rapidly—from extensional faulting favoured in the forebulge, to strike-slip, to reverse faulting beneath the maximum load elevation.

There are also two regions with large stressing rate located well within the mantle. One is at the base of the plate,
The second deep region of high-stressing rate extends vertically through the middle of the plate near, to just seaward of the leading edge of the load. This is the region that corresponds most closely to the mantle fault zone, as the increasing stress occurs on gently dipping fault planes and the maximum occurs near the appropriate depth. However, as with the maximum stress occurring on horizontal planes, the maximum stressing rate is displaced to the northwest of the mantle fault zone, in this case by 25–30 km. As was stated previously, it might be possible to adopt a more complicated flexure model that could move the region of largest stress and stressing rate 20 km (or more) to the northwest, but it seems unlikely that such a model would also explain the localization of the mantle fault zone in map view (hence the need to consider the stress concentration due to the magma transport system, as in the previous section).

To the extent that the stress concentration due to the magma transport system is important in generating seismicity, one can legitimately question the steady-state assumption in the computation of stressing rate. For example, strictly speaking this would require the plumbing system to migrate continuously through the moving plate, whereas the volcanoes on top of that plumbing system appear to ‘jump’ in a more discrete fashion. Although migration of the plumbing system is impossible to constrain, one can model the flexural stresses due to an idealized Hawaiian load, see (e). (a) The ratio of shear stress on favourably oriented planes (assuming a frictional coefficient of 0.6), corresponding to the profile in (e). (b) Horizontal derivative of (a), which is a proxy for the stressing rate in a steady-state model. The units are change in the ratio (a) per year. (c) Map view of shear stress on a horizontal plane at the depth of the mantle fault zone (roughly 30 km) in units of MPa. The arrows show the direction of shear stresses. (d) Derivative of the shear stress in (c) along lines with the same orientation as the profile from (e). The units are MPa change per year. (e) Load geometry where colours indicate the height of the load in kilometres. The white solid line shows the profiles in (a) and (b), and the black dashed line shows the regions in (c) and (d).

5.3 Seismic moment rate

As a final test of the flexural model, we compare the observed seismic moment rate to a hypothetical steady-state rate based on the computed flexural stresses. To estimate an upper bound on the observed seismic moment rate \( \dot{M} \), we will assume that all earthquakes in the mantle fault zone have the same mechanism. We include earthquakes between depths 20 and 40 km, latitudes 19.15°–19.5°N, longitudes 155.45°–155.1°W (Fig. 1c), and 1959–2004 from the Hawaii Volcanos Observatory catalogue (quake.geo.berkeley.edu/cnss/catalog-search.html), and use the relation between local magnitude and moment given by Savage & Meyer (1985). We find \( \dot{M} = 2.5 \times 10^{6} \text{ N m s}^{-1} \) (about one \( M_w \) 4.5 yr\(^{-1}\)). The moment rate has not been dominated by a single earthquake (the largest event has a \( M_w \) ~ 5), and in the 45 yr we considered, there were about 60 events with \( M_L > 4 \).

To estimate a theoretical seismic moment rate, we imagine that slip on the mantle fault zone relaxes some fraction of the computed shear traction (\( \Delta \tau \)) on a horizontal plane at 30 km depth. Given a fault area determined from the above coordinates (roughly 40 by 40 km), the total moment is \( \sim \Delta \tau (40 \text{ km})^3 \), or \( \sim 6 \times 10^{19} \text{ N m} \) per MPa of stress drop. If we assume a steady-state process whereby the total slip accumulates in the time required for a point to move from the leading edge to the trailing edge of the fault zone where the fault becomes locked, then the moment rate is determined by dividing the total moment by (40 km)/(10 cm yr\(^{-1}\)), yielding \( \sim 6 \times 10^6 \text{ Nm s}^{-1} \) per MPa of stress drop. To reach the current seiz-
mic moment rate of $2.5 \times 10^6 \text{Nm s}^{-1}$ would then require a stress drop of $\sim 50 \text{MPa}$, or roughly half the computed shear stress of $\sim 100 \text{MPa}$ (Fig. 6a). Fractional stress drops this large probably require some mechanism of substantial fault weakness at large slip speed (Rice & Cocco 2005), or very high pore pressure. Given the order-of-magnitude nature of this estimate, we conclude that the observed seismicity rate is at least consistent, within large error bounds, with steady-state behaviour and the computed flexural stresses.

5.4 Other mantle earthquakes

Besides the quasi-horizontal fault zones beneath Kilauea and the south flank, there are at least six distinct zones of mantle earthquakes beneath the Big Island (Fig. 2) composed of either a large number of small earthquakes or one or more events with $M_w \sim 5$: (1) a ring fault beneath Mauna Kea (Wolfe et al. 2004); (2) a swarm starting in 2002 beneath Mauna Loa (Okubo et al. 2005); (3) the 1973 $M_f 6.2$ Hononu earthquake (Unger & Ward 1979; Butler 1982; Chen et al. 1990); (4) a thrust earthquake with $M_w 5.5$ on 1991 May 8 west of the Big Island (Wolfe et al. 2004); (5) an oblique strike-slip earthquake with $M_s 4.9$ on 1991 December 9 southwest of the Big Island; (6) two earthquakes on 2005 May 13 ($M_w 4.7$) and 2005 August 17 ($M_w 5.1$) near Loihi (M. Nettles, personal communication, 2006). Outside of these regions, there is also a generally uniform, diffuse mantle seismicity which may indicate real earthquakes outside of these regions or be an artefact of poor locations of small, poor signal-to-noise earthquakes. For example, earthquakes in the region labelled ‘Probably not in mantle’ in Fig. 2(c), are probably mislocated because of poor network geometry and do not occur in the mantle (Wolfe et al. 2004). We now compare each of these zones of mantle earthquakes to the predictions of the flexural stresses.

The mantle earthquakes beneath Mauna Kea occur in a ring around the volcano, but offset from its centre (Fig. 2). The earthquakes are not uniformly distributed around the volcano, but occur mostly in the E and W–NW. One possible explanation for this non-uniform distribution in azimuth is that the earthquakes are caused by a low shear stress tunnel with a radius of about 5 km that concentrates the ambient stress field. In Fig. 8(b) (as in Fig. 8a) we have calculated the azimuthal distribution of the stress concentration surrounding a vertical shear stress-free tunnel cut through a locally uniform stress field taken from the flexural calculations, in this case in the vicinity of the ring structure beneath Mauna Kea. We assume that the faults strike north–south and dip at 45°, although east–west strikes and horizontal faults yield similar patterns. The largest stress concentrations are in approximately the same locations as the observed mantle earthquakes. The mechanisms that have been calculated for earthquakes beneath Mauna Kea (Fig. 2c) indicate a compressive stress field consistent with the flexural models, but without a simple, single orientation.

Based on the presence of seismic swarms and long-period earthquakes, fluids and magma movements are clearly associated with the seismicity at Mauna Loa, source 2 (Okubo et al. 2005). Mechanisms have not been calculated for these events, so direct comparison with the flexural model is not possible.

There is no obvious magma plumbing system or long-period earthquake activity associated with sources 3–5. All of these events occur in a ring around the main island where flexural stresses on horizontal planes are maximum at the depth of the neutral plane. The fact that events 3–5 are northeast and southwest of the Big Island and outside the zones with the highest computed flexural stressing rates, which are southeast of the Big Island, Fig. 10, indicates that non-flexural stresses may be the trigger for these earthquakes. For the Honomu earthquake (source 3), there is no obvious relation between the earthquake and volcanic activity or surface geological features (Unger & Ward 1979), although the preferred fault plane is parallel to the Mauna Kea east rift zone (Butler 1982). One possible trigger for this and other earthquakes is viscoelastic readjustments of the stress field from large landslides.

Butler (1982) first noted that the stress orientation inferred from the focal mechanism for the Honomu earthquake (45° E of N, inclined 25°) is consistent with flexural stresses. Klein et al. (1987) also noted that unpublished mechanisms from deep mantle earthquakes beneath Hawaii suggest a radially oriented stress field consistent with the flexural calculations (Fig. 6b). The mechanisms of two earthquakes in 1991 are also broadly consistent with the flexural models (predicted strike and dip within 20°–30° of that observed). Like sources 3–5, the earthquakes in region 6 occur in the ring of high flexural stresses at the depth of the neutral plane, but these events occur in the area where the flexural stressing rate is the highest. The mechanisms of these earthquakes is generally consistent with the Kilauea fault zone (seaward slip on low-angle faults). These events are also located near Loihi and may also be influenced by its magma plumbing system.

6 SUMMARY

While previous workers proposed that earthquake locations delimit the magma pathway in the mantle beneath Hawaii (Ryan 1988), recent analysis of the characteristics of these earthquakes, including their mechanisms, main shock/aftershock behaviour, as well as their planar distribution, support the hypothesis that earthquakes in this region occur along mantle fault zones and do not represent a vertical chain of seismicity surrounding a magma pathway (Wolfe et al. 2004). In conclusion, recent analysis of the characteristics of these earthquakes, including their mechanisms, main shock/aftershock behaviour, as well as their planar distribution, support the hypothesis that earthquakes in this region occur along mantle fault zones and do not represent a vertical chain of seismicity surrounding a magma pathway (Wolfe et al. 2004).
et al. 2003). Such a pathway may exist, but our interpretation is that magma movements are not the primary cause of the mantle earthquakes (e.g. Klein et al. 1987). It is most likely that plate flexure is the dominant source of stress at the depths of the mantle fault zone, and the sense of slip on this fault as well as (approximately) the fault orientation and location are all consistent with this. In addition, the rate of seismicity is consistent, within a substantial uncertainty, with the flexure model and a steady-state assumption. However, the existence of stresses of sufficient magnitude, as well as the localization of the fault zone, particularly in map view, cannot be explained solely with a simple flexural model. Plausible additions to the flexural model can increase the magnitude of the stress and its localization—the existence of high pore pressures and/or a stress concentrating magma pathway.

Many other regions of the lithosphere are flexed (e.g. other ocean islands, subduction zone outer rise and unbounding, postglacial unloading), and precise earthquake locations from these regions could help better constrain the flexural models. While failure on horizontal planes will be predicted in certain regions, our models from Hawaii suggest that earthquakes will only occur on horizontal planes when there is some additional stress concentrator at depth.

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REFERENCES

APPENDIX A: THERMAL CONSTRAINTS ON MAGMA PATHWAY SIZE

The lack of obvious evidence for dyke propagation, such as seismicity and rapid reservoir-inflating events that might herald the arrival of large dykes, leads us to consider the possibility that porous flow is the dominant mechanism of magma transport beneath Kilauea. Even if this is the case currently, a series of dyke intrusions is necessary initially in order to provide a viable (i.e. near-solidus) thermal environment for porous flow. And, as we discuss in Section 5.1.2, such dykes are also necessary to provide a stress environment that is permissive of dyke ascent through the lithosphere. Hieronymus & Ber covici (2001) also address the apparent paradox of having dykes that are required to overcome the unfavourable flexural stress field can be obtained by assuming that their net effect is to increase the least compressive stress from the value of the flexural calculation up to $\rho_m g z$. Then from elasticity

$$W = 2 \frac{\Delta P}{G} \min[l, h],$$

(A1)

where $\Delta P$ is the final minus the initial pressure, $G$ is the elastic modulus, $l$ is half the cumulative strike length of the dykes, and $h$ is the cumulative dyke height (not the half-height because on long timescales the base of the plate is effectively a free surface). If $l > h$ and $h$ approaches the lithospheric thickness this estimate might be several-fold too low because it does not account for the Earth's surface. If we conservatively take $h = 20$ km (from the base of the plate to the neutral plane) and assume $l > h$ km, then an average $\Delta P$ is $(1/2)(20 \text{ MPa})$. From (A1) this leads to a cumulative average thickness of at least 200 m, a number that will increase with $h$ more rapidly than $h^2$. Provided $l > h$. (Structural discontinuities along strike could limit $l$, but for a homogeneous stress field akin to Fig. 7, dykes at the base of the lithosphere would propagate laterally to a much greater extent than vertically, and the characteristic length scale for lateral stress variations in the flexural model is considerably greater than the plate thickness, so $l > h$ seems reasonable.) For a dyke swarm reaching through the lithosphere a cumulative elastic thickness in excess of 1 km seems likely. If the strike of such a swarm parallels the island chain then the plate is essentially ‘broken’ on a large scale and it seems likely that even greater aggregate elastic thicknesses would be required for dykes to traverse the lithosphere. A ‘broken’ plate would likely change the orientation of the stress field, and more sophisticated modelling would be required to assess the magnitude of this effect. However, the effect may not be so important if there are structural features that limit the strike length of the dykes, or if emplacement of a large-scale dyke swarm is a transient stage before a cylindrical porous flow channel is established (see below).

As an example estimate of the timescale for such intrusions we assume a total volume of $40 \text{ km} \times 80 \text{ km} \times 1 \text{ km} \approx 3 \times 10^3 \text{ m}^3$. For an intrusion rate equal to the current extrusion rate of $0.1 \text{ km}^3 \text{ yr}^{-1}$ this would take $\sim 3 \times 10^8 \text{ yr}$, a plausibly small value as it is roughly 10 per cent of the volcano separation divided by the plate motion rate ($\sim 10 \text{ km} \text{ Myr}^{-1} \approx 300 \text{ 000 yr}$). As this magma freezes it releases roughly $500 \text{ kJ kg}^{-1}$ of latent heat, plus the heat capacity of roughly $1.2 \text{ kJ kg}^{-1} \text{ K}$ times a lithosphere-averaged drop in the solidus temperature of perhaps 100 K. If we add this to the latent heat and assume an average host rock temperature of 700 K below the solidus, then $\sim 3 \times 10^8 \text{ yr}$ is sufficient to freeze a cumulative thickness of $\sim 1.4 \text{ km}$ (from each surface, Turcotte & Schubert 1982). As this value is three times the assumed cumulative dyke thickness, a dyke swarm of this size could not generate and maintain near-solidus temperatures over this timescale.

There are several options for making the results of this calculation more favourable for maintaining solidus temperatures. If the cumulative dyke thickness increases linearly with time, then at some point the intrusions will cease to freeze completely because the diffus-
Figure A1. Comparison of the heat input into a cylindrical pipe (dashed line—equal to the heat associated with the change in solidus between the mantle fault zone and the surface from the observed-magma flux) and the heat lost by conduction for pipes of different diameters (solid lines—equal to the heat flux times the surface area). The dashed line labelled 30 000 yr is the approximate order-of-magnitude time necessary to create a magma pathway (see text). After 30 000 yr, pipes with a diameter smaller than 2 km have no trouble remaining at the solidus temperature if the observed-magma flux is maintained.

Whether the space for such a conduit is initially generated thermally (with a cylindrical geometry) or mechanically (with a planar geometry), we imagine that the geometry quickly evolves to cylindrical. If they can form, conduits with diameters of a few kilometres or less can be maintained because the heat input by the decrease in the leading edge needs to be lengthened, or because of fault slip that simultaneously limits the intrusion length and permits greater intrusion thicknesses, widening rates could increase by another factor of 2 or more. A final possibility is that the requisite intrusion thickness could be reached via thermal erosion surrounding a cylindrical conduit that nucleates somewhere along the intrusive complex, much as is observed to occur at shallow levels in Hawaii and elsewhere (e.g. the still-active Pu’u O’o conduit at Kilauea that grew from the 1983 dyke intrusion).

The assumed 80 km strike length for the dyke complex is also woefully inefficient thermally. If this number could be reduced, either because the dyke complex is aligned with the island chain and only the leading edge needs to be lengthened, or because of fault slip that simultaneously limits the intrusion length and permits greater intrusion thicknesses, widening rates could increase by another factor of 2 or more. A final possibility is that the requisite intrusion thickness could be reached via thermal erosion surrounding a cylindrical conduit that nucleates somewhere along the intrusive complex, much as is observed to occur at shallow levels in Hawaii and elsewhere (e.g. the still-active Pu’u O’o conduit at Kilauea that grew from the 1983 dyke intrusion).

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