

Reconciling seismic and geodetic models of the 1989 Kilauea south flank earthquake

Andrew Hooper, Paul Segall, Kaj Johnson, and Justin Rubinstein

Department of Geophysics, Stanford University, Stanford, California, USA

Received 22 August 2002; revised 15 October 2002; accepted 17 October 2002; published 26 November 2002.

[1] The inconsistency in depth between geodetic and seismic models of the 1989 Kilauea earthquake has long been puzzling. Previous attempts to incorporate elastic heterogeneity have deepened geodetic models substantially, bringing them closer to the seismic depth. However, recent studies that have included heterogeneity for other faults indicate that the effect is not so great. We show here that elastic heterogeneity has a relatively minor effect on the geodetic model depth for this earthquake also. However, by combining three different sets of geodetic data we are able to get a more accurate estimate of the depth, which does in fact coincide with the hypocentral depth at the 95% confidence level. When we consider that static elastic parameter values are commonly less than dynamic values, the agreement is even better. Furthermore, the depth is consistent with the earthquake having occurred at the interface between the volcanic pile and the pre-volcanic seafloor. **INDEX TERMS:** 7280 Seismology: Volcano seismology (8419); 8429 Volcanology: Lava rheology and morphology; 9355 Information Related to Geographic Region: Pacific Ocean. **Citation:** Hooper, A., P. Segall, K. Johnson, and J. Rubinstein, Reconciling seismic and geodetic models of the 1989 Kilauea south flank earthquake, *Geophys. Res. Lett.*, 29(22), 2062, doi:10.1029/2002GL016156, 2002.

1. Introduction

[2] At 0327 UTC on June 26th, 1989 a magnitude M6.1 earthquake struck the south flank of Kilauea Volcano on the island of Hawaii. The hypocentral depth was determined from short-period seismic data to be at 9.2 ± 2.0 km [Bryan, 1992], and from teleseismic data to be at 13 km [Chen and Nábělek, 1990]. However, estimates from geodetic data using a homogeneous Earth model gave a sub-horizontal fault at an average depth of 4 ± 1 km [Arnadottir et al., 1991] and 6 ± 2 km [Dvorak, 1994] (see Table 1). Arnadottir et al. [1991] determined that adding elastic heterogeneity to the model would deepen the fault and adopting this approach Du et al. [1997] concluded that the depth of the fault could be as deep as 7 ± 0.5 km. However, the results of other recent studies suggest that heterogeneity is unlikely to have so large an effect [e.g. Cattin et al., 1999; Johnson et al., 2001; Cervelli et al., 2002]. In this study, we firstly re-evaluate the model of Du et al. [1997], and then introduce a more realistic model of elastic heterogeneity. In an attempt to reconcile seismic and geodetic models we also use all the geodetic data available

to us, including EDM data that has not been used in any published modeling of the earthquake.

[3] Knowing the actual depth of the earthquake is important in our understanding of the mechanics of Kilauea. The south flank of Kilauea is moving south-southeast relative to the Pacific Plate by about 8 cm/yr [Owen et al., 2000]. Many authors believe that this movement is accommodated by sliding along a basal decollement at the interface of the volcanic pile and the pre-volcanic sea floor [e.g., Nakamura, 1980; Dieterich, 1988; Thurber & Gripp, 1988; Delaney et al., 1993; Owen et al., 2000] at a depth of about 7 to 8 km. However, Cervelli et al. [2002] demonstrated that some slip at least does occur at a shallower depth.

2. Hypocenter Relocation

[4] We used the double-difference travel-time method of Waldhauser and Ellsworth [2000] to obtain more accurate locations for all the earthquakes within the eastern rift zone during the months of June and July of 1989 (see Figures 1 and 2). The velocity model consisted of 5 horizontal layers with the average velocity of each layer derived from Okubo et al. [1997]. Using this approach the uncertainty in the locations of the events was reduced by 84% relative to undifferenced locations.

[5] Although this method provides accurate relative positions of the earthquakes, the absolute positions are less well constrained, particularly in regards to the depths, which are dependent on the velocity model. However, our estimated depth of 7.4 km relative to the surface coincides with the decollement believed to be at 7 to 8 km [Thurber and Gripp, 1988].

3. Du Heterogeneous Model

[6] Du et al. [1997] used leveling data collected in early 1988 and after the earthquake in 1989 (see Figure 1 for benchmark locations) to solve for the best-fitting fault geometry, using a moduli perturbation method to include the lateral and vertical heterogeneity in elastic properties, as shown in Figure 3a. We essentially repeated this inversion, using the same leveling data and the same model of heterogeneity. We set up the problem to solve for the change with time in the height difference between each pair of leveling benchmarks. This is equivalent to solving for the change with time in the height of each benchmark, but has the advantage of making each data point independent and the covariance matrix is therefore purely diagonal. The standard error of each measurement was assumed to be $2.83 \text{ mm km}^{-1/2}$ as per Arnadottir et al. [1991]. Using the Arnadottir et al. [1991] homogeneous optimal model as our prior, we solved for 9 model parameters (length, width,

Table 1. Existing Geodetic Model Properties

Model	<i>Arnadottir et al.</i> , [1991]	<i>Dvorak</i> [1994]	<i>Du et al.</i> [1997]
Elastic Properties	Homogeneous halfspace	Homogeneous halfspace	Lateral and vertical heterogeneity
Data Used	Leveling	GPS	Leveling
Depth (km)	4 ± 1	6 ± 2	7 ± 0.5

depth, strike, dip, longitude, latitude, along-strike slip and down-dip slip) using the trust-region reflective Newton non-linear algorithm [Coleman and Li, 1994].

[7] We ran the inversion twice, once including the heterogeneity and once without for comparison. In the homogeneous case, our result differed slightly from that of *Arnadottir et al.* [1991], giving the average depth of the best-fitting fault plane to be at 3.5 km. In the heterogeneous case, the average depth increased by only 0.8 km to 4.3 km (model (a), Table 2). This is significantly different to the result of *Du et al.* [1997], whose best-fitting fault plane lay at 7.0 km. The minimum in the misfit with depth curve is not sharply defined in the homogeneous case and the introduction of a change in elastic properties at approximately the same depth as the fault adds numerical noise to the curve. Combined with the fact that computing efficiency restrictions at that time only allowed for one Newton step, makes it likely that the *Du et al.* [1997] solution is in fact a local minimum.

4. Extra Geodetic Data

[8] The leveling data provides only the vertical gradient along one line, which is not enough to constrain the fault

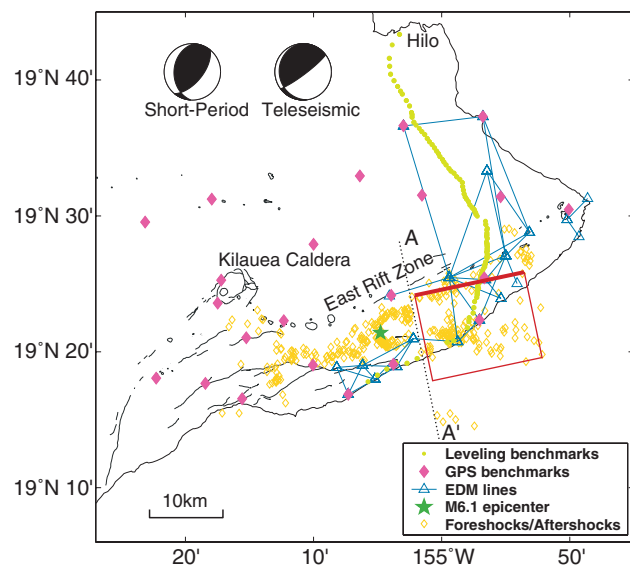


Figure 1. Map of southeastern Hawaii, showing geodetic benchmarks and earthquake epicenters for June and July 1989. The focal mechanism is given for the main shock, as calculated from short-period data [Bryan, 1992] and teleseismic data [Chen and Nábělek, 1990]. The rectangle shows the surface projection of model (b) in Table 2 (the thicker line represents the up-dip end).

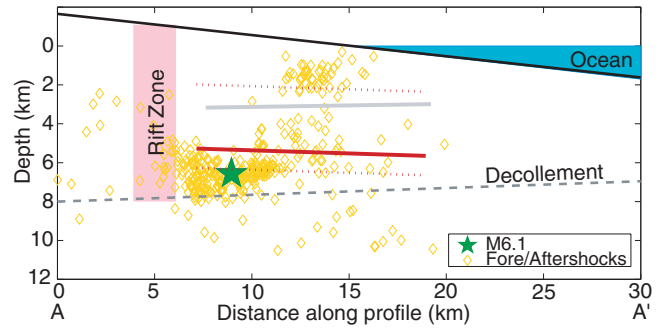


Figure 2. Earthquake hypocenters for June and July 1989 projected on to line A-A' in Figure 1. The gray sub-horizontal line is the projection of the best-fitting fault plane solution using leveling data only and modeled in a homogeneous halfspace as per *Arnadottir et al.* [1991]. The lower line represents model (b) in Table 2 and the dotted lines represent the 95% confidence bounds for this solution (derived from the 95% confidence bounds for the solution using all the data modeled in a homogeneous halfspace).

geometry well on its own. We therefore added the following data to our subsequent inversions:

4.1. GPS Data

[9] GPS surveys were carried out by HVO in August of 1988 and 1989 (see Figure 1 for benchmark locations) and we used the displacement solutions from *Dvorak et al.* [1994]. The covariance matrix was not available, so we scaled the NS and EW variances by the repeatability values and used these to construct a diagonal covariance matrix. We assumed that secular displacement was constant over this period and removed it using the model of *Owen et al.* [2000], which we also used to remove secular displacement from the leveling data.

[10] Hawaii lies outside the network of GPS-tracking stations in North America that were used to compute satellite orbits at this time. Additionally a key satellite was missing for much of both the 1988 and 1989 surveys, which meant that 4 satellites could only be tracked simultaneously for 20 minutes per observing session. For these reasons, the uncertainties are very high (see Figure 4) and therefore the GPS data contribute little to our geodetic fault solutions. However, the general pattern of displacement is useful in ruling out some solutions.

4.2. EDM Data

[11] We used EDM data that was collected by HVO on and around Kilauea from 1970 to 1995. To avoid deforma-

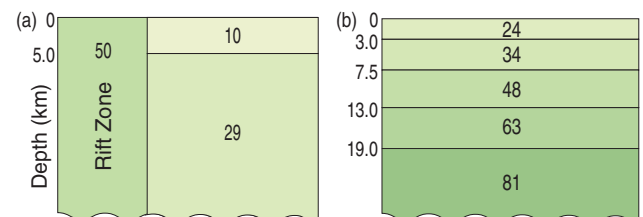


Figure 3. Models of shear modulus heterogeneity in GPa. (a) After *Du et al.* [1997]. (b) Derived from seismic P-wave velocities from *Okubo et al.* [1997].

Table 2. New Geodetic Model Properties

Model	(a)	(b)	(c)
Elastic Properties	Lateral and vertical heterogeneity ^a	Vertical heterogeneity	Vertical heterogeneity (top layer varies)
Data Used	Leveling	All	All
Heterogeneity Method	Perturbation	Propagator Matrix	Propagator Matrix
Depth (km)	4.3	6.0	7.4
Misfit ^b (mm)	2.7	3.8	3.8

^a*Du et al.*, [1997] model of heterogeneity.

^bThe weighted root mean square (rms) misfit [*Segall and Harris*, 1986].

tion associated with the caldera biasing our results, we only included lines that were wholly east of longitude $155^{\circ} 10'$ west. We further discarded lines that didn't span the 1989 earthquake and those surveyed less than 8 times, to allow reasonable error estimation (see Figure 1 for the lines we used). For each line we assumed a constant secular rate of change in line length over the entire period, with offsets at known events. We inverted for this rate of change and the offsets, and used a bootstrap percentile method [*Efron and Tibshirani*, 1993] to estimate the 95% confidence limits and standard error of the offset at the 1989 event for each line. Errors for each line were assumed to be uncorrelated and our covariance matrix was therefore simply a diagonal matrix of variances.

5. Layered Heterogeneity

[12] The P-wave velocity distribution derived by *Okubo et al.* [1997] from a tomographic inversion suggests that the model of heterogeneity used by *Du et al.* [1997] over-emphasizes the difference in elastic parameters between the rift zone and the adjacent crust. In fact, the lateral variation in velocity appears to be 2nd order compared to the vertical variation. We therefore ignored lateral contrasts and divided the crust into 5 horizontal layers of constant shear modulus derived from the average velocity of each layer (see Figure 3b), assuming the crust to be a Poisson solid (in fact *Cattin et al.* [1999] demonstrated that varying Poisson's ratio doesn't significantly alter the model depth).

[13] We included the heterogeneity using a propagator matrix method [*Ward*, 1985; *Johnson et al.*, 2001] and combining the leveling, GPS and EDM data we solved for the 9 model parameters using the Nelder-Mead simplex non-linear algorithm [*Nelder and Mead*, 1965] (the more efficient trust-region reflective Newton tended to converge on local minima). We approximated the 1st order effect of topography by fitting a plane to the surface of Kilauea and rotating our model and solution into this plane.

[14] The best-fitting solution (model (b), Table 2) is a sub-horizontal fault with an average depth of 6.0 km below the surface, striking 078° and dipping 2° SSE (see Figures 1 and 2). The slip on the fault has a 1.4 m normal dip-slip component and a 0.09 m left-lateral strike-slip component. The fit to the data is shown in Figures 4 and 5. Some of the EDM data points are not fit at the 95% confidence level and this may be partly because these measurements include a certain amount of inelastic deformation (surface ground cracks were noted around Kalapana by *Arnadottir et al.*, 1991). The weighted root mean square misfit [*Segall and Harris*, 1986] is 3.8 mm as compared to 2.7 mm for the *Du*

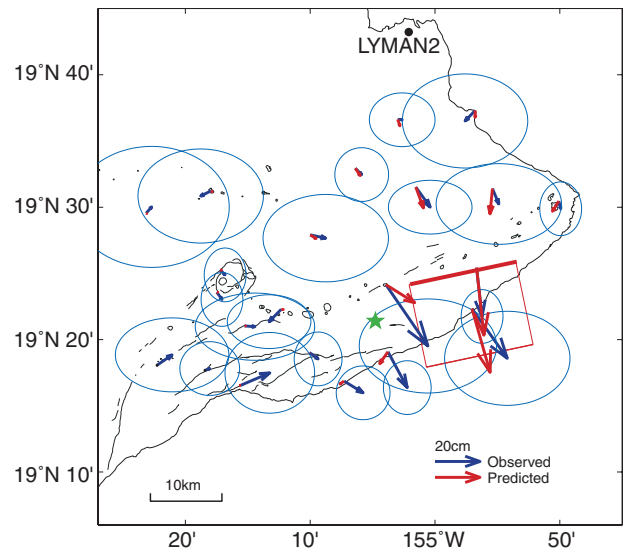


Figure 4. Predicted and observed GPS benchmark displacements relative to LYMAN2, with 95% confidence ellipses. The predicted values are for model (b) in Table 2. The surface projection of this solution is shown, with the thicker line representing the up-dip end.

et al. [1997] model of heterogeneity using leveling data only, but the addition of the EDM and GPS data also increases the measurement uncertainty (average standard error of the data increases from 1.2 mm to 1.6 mm).

[15] Using a bootstrap percentile method with 1000 iterations we constrained the 95% confidence bounds on the average depth when modeled in a homogeneous half-space. If we assume that the model variance of the heterogeneous model is reasonably approximated by the variance of the homogeneous model, we can apply the same confidence range to the heterogeneous model. This gives 95%

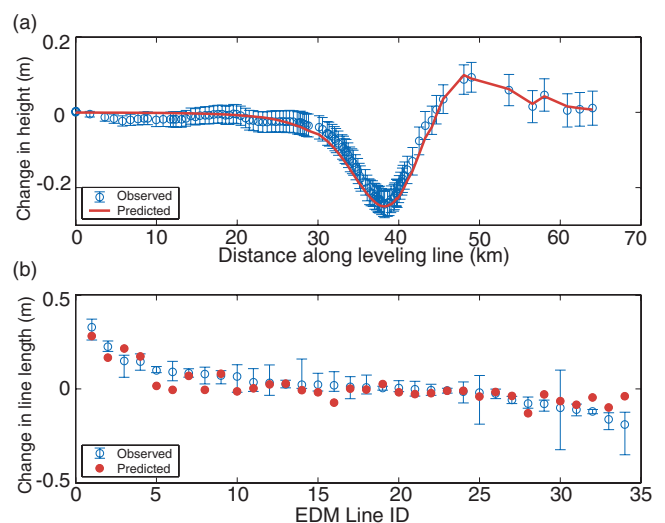


Figure 5. Predicted and observed values of (a) leveling and (b) EDM data. The predicted values are for model (b) in Table 2. In (a), distances along the line are from the Hilo tide gauge. In (b) the EDM lines are numbered in order of decreasing line length change.

confidence bounds of 3.2 and 7.5 km and the hypocentral depth lies within this interval.

[16] We note that the epicenter does not lie within the surface projection of our best-fitting fault plane (see Figure 1). This is likely an artifact of modeling the fault with uniform slip, which forces the slip to go to zero discontinuously. When we solved for the fault geometry in a homogeneous halfspace assuming distributed-slip, the slip went to zero gradually over a larger projected area, which included the epicenter.

[17] The strike and dip of our geodetic model is inconsistent with the focal mechanism determined from short-period seismic data [Bryan, 1992], but is consistent with that determined from long-period data [Chen and Nábělek, 1990] (see Figure 1). The total moment of our model is $7.7 \cdot 10^{18}$ Nm which is in reasonable agreement with the double-couple moment of $5.2 \cdot 10^{18}$ Nm estimated from the CMT solution [National Earthquake Information Center, 1989].

6. Static and Dynamic Elastic Parameters

[18] Static elastic parameter values are commonly less than dynamic values [Simmons and Brace, 1965; King, 1969; Cheng and Johnston, 1981]. This is thought to be due to differences in both strain amplitude and frequency. For rocks containing cracks, the effect is more pronounced at lower pressures, i.e. closer to the surface [Jizba, 1991]. The upper few km of Hawaiian crust likely contains many fractures and so would probably behave less stiffly in response to static displacement on a fault than would be predicted from the elastic parameters determined at seismic frequencies.

[19] To see how large this difference in elastic parameters would need to be to explain the difference in depth between our best-fitting fault plane and the hypocenter, we performed another inversion. This time we fixed the depth of the fault plane to be at our best estimate of the hypocentral depth (7.4 km) and allowed the shear modulus and depth of the upper layer to vary, as well as the other 8 fault parameters (model (c), Table 2). Our optimal solution is a 2.7 km thick layer with a shear modulus of 4.5 GPa, which is a factor of 5 lower than the value derived from the P-wave velocity. There is no laboratory data available for fractured basalts to compare this ratio to, but it lies at the upper end of the range determined for sandstones by Jizba [1991]. Considering that the sandstone samples did not contain any macroscopic fractures which would likely increase further the difference between static and dynamic values, a factor of 5 for Hawaiian crust certainly seems plausible.

7. Conclusions

[20] Adding heterogeneity to the fault model does not have as great an influence on the depth as previously concluded by Du et al., [1997]. Depth is only increased by 0.7 to 0.8 km, depending on model of heterogeneity, as opposed to 3 km.

[21] Including measurements of displacement in the horizontal does deepen the best-fitting solution considerably. When combined with the effects of heterogeneity, the geodetic model depth is indistinguishable at the 95% confidence level from the hypocentral depth determined from short-period seismicity. These depths permit that the fault plane

does in fact coincide with the decollement at the interface between the volcanic pile and the pre-volcano seafloor.

[22] **Acknowledgments.** We thank Thora Arnadottir and Valerie Cayol for their thoughtful reviews that led to significant clarification of the paper. This work was supported by the Robert and Marvel Kirby Stanford Graduate Fellowship and NSF grant EAR-9902875.

References

- Arnadottir, T., P. Segall, and P. Delaney, A fault model for the 1989 Kilauea south flank earthquake from leveling and seismic data, *Geophys. Res. Lett.*, *18*, 2217–2220, 1991.
- Bryan, C. J., A possible triggering mechanism for large Hawaiian earthquakes derived from analysis of the 26 June 1989 Kilauea south flank earthquake, *Bull. Seismol. Soc. Am.*, *82*, 2368–2390, 1992.
- Cattin, R., P. Briole, H. Lyon-Caen, P. Bernard, and P. Pinettes, Effects of superficial layers on coseismic displacements for a dip-slip fault and geophysical implications, *Geophys. J. Int.*, *137*, 149–158, 1999.
- Cervelli, P., P. Segall, K. Johnson, M. Lisowski, and A. Miklius, Sudden aseismic fault slip on the South Flank of Kilauea Volcano, Hawaii, *Nature*, *415*, 1014–1018, 2002.
- Chen, W. P., and J. Nábělek, Source parameters of the June 26, 1989 Hawaiian earthquake, *Eos Trans. AGU*, *71*, 562, 1990.
- Cheng, C. H., and D. H. Johnston, Dynamic and static moduli, *Geophys. Res. Lett.*, *8*, 39–42, 1981.
- Coleman, T. F., and Y. Li, On the convergence of reflective Newton methods for large-scale nonlinear minimization subject to bounds, *Math. Prog.*, *67*, 189–224, 1994.
- Delaney, P. T., A. Miklius, T. Arnadottir, A. T. Okamura, and M. Sako, Kinematic Motion of Kilauea Volcano during sustained eruption from the Pu'u O'o and Kupaianaha vents, 1983–1991, *J. Geophys. Res.*, *98*, 17,801–17,820, 1993.
- Dieterich, J. H., Growth and persistence of Hawaiian rift zones, *J. Geophys. Res.*, *93*, 4258–4270, 1988.
- Du, Y., P. Segall, and H. Gao, Quasi-static dislocations in three dimensional inhomogeneous media, *Geophys. Res. Lett.*, *24*, 2347–2350, 1997.
- Dvorak, J., A. T. Okamura, M. Lisowski, W. H. Prescott, and J. L. Svarc, Global Positioning System Measurements on the Island of Hawaii from 1987 to 1990, *U.S. Geol. Surv. Bull.*, *2092*, 33, 1994.
- Efron, B., and R. J. Tibshirani, *An Introduction to the Bootstrap*, Chapman and Hall, New York, 1993.
- Jizba, D. L., Mechanical and Acoustical Properties of Sandstones and Shales, Ph.D. thesis, Stanford Univ., 1991.
- Johnson, K. M., P. Segall, and P. Cervelli, Analytical methods for including vertical and horizontal heterogeneity of elastic properties in dislocation models and applications to two coseismic GPS datasets, *EOS Trans. AGU*, *82*, Fall Meet. Suppl., 296–297, 2001.
- King, M. S., Static and dynamic elastic moduli of rocks under pressure, in *Rocks Mechanics - Theory and Practice*, edited by W. H. Somerton, Proc. 11th Symp. Rock Mech., Univ. of Calif., 1969.
- Nakamura, K., Why do long rift zones develop in Hawaiian volcanoes - A possible role of thick oceanic sediments (in Japanese), *Bull. Volcanol. Soc. Jpn.*, *25*, 255–267, 1980.
- Nelder, J. A., and R. Mead, A simplex method for function minimization, *Comput. J.*, *7*, 308–313, 1965.
- Okubo, P. G., H. M. Benz, and B. A. Chouet, Imaging the crustal magma sources beneath Mauna Loa and Kilauea volcanoes, Hawaii, *Geology*, *25*, 867–870, 1997.
- Owen, S., P. Segall, M. Lisowski, A. Miklius, R. Denlinger, and M. Sako, Rapid deformation of Kilauea volcano: GPS measurements between 1990 and 1996, *J. Geophys. Res.*, *105*, 18,983–18,998, 2000.
- Segall, P., and R. Harris, Slip deficit on the San Andreas fault at Parkfield, California, as revealed by inversion of geodetic data, *Science*, *233*, 1409–1413, 1986.
- Simmons, G., and W. F. Brace, Comparison of static and dynamic measurements of compressibility of rocks, *Geophys. Res. Lett.*, *70*, 5649–5656, 1965.
- Thurber, C. H., and A. E. Gripp, Flexure and seismicity beneath the south flank of Kilauea Volcano and tectonic implications, *J. Geophys. Res.*, *93*, 4271–4278, 1988.
- Ward, S. N., Quasi-static propagator matrices: creep on strike-slip faults, *Tectonophysics*, *120*, 83–106, 1985.
- Waldhauser, F., and W. L. Ellsworth, A double-difference earthquake location algorithm; method and application to the northern Hayward Fault, California, *Bull. Seismol. Soc. Am.*, *90*, 1353–1368, 2000.