SEISMIC STRUCTURE AND TECTONICS OF KILAUEA VOLCANO

By Clifford H. Thurber [1]

ABSTRACT

Local earthquake data have been used to study the internal structure and tectonics of Kilauea Volcano. A three-dimensional model of Kilauea's seismic velocity structure, determined using simultaneous inversion techniques, reveals features on a scale of one to a few kilometers. The core and rift zones of the volcano are marked by relatively high seismic velocity; a small, shallow, aseismic low-velocity zone beneath the summit coincides with the inferred locus of the summit magma reservoir. High-quality fault plane solutions have been derived for a suite of local earthquakes using this three-dimensional model. The spatial variations in focal mechanisms provide information on the complex tectonics of the volcano; the temporal variations in mechanisms observed for summit-area events may provide clues to help us understand and perhaps predict Kilauea's eruptive behavior.

INTRODUCTION

Kilauea Volcano is an excellent natural laboratory for investigating the relations among stress state, seismicity, and magma migration within an active volcano. Seismicity patterns have been used to outline the structure of Kilauea's magma-transport system (Ryan and others, 1981). Earthquake focal mechanisms from the upper east rift zone (UERZ) have been interpreted in terms of a model for magma transport (Hill, 1977). However, magma migration itself is a poorly understood process. Shaw (1980) suggests that seismic data, coordinated with eruption and deformation data, may provide critical tests of hypotheses related to magma transport. It is in this context that a detailed seismotectonic study of Kilauea Volcano has been initiated.

Routinely available local earthquake data, particularly body-wave arrival times and first-motion polarities, can be used as powerful probes of the structure and stress state of the Earth's interior on a local scale. Focal mechanisms derived from first-motion polarities are indicative of the principal stress directions. However the local Earth structure must be known to determine accurately both the source location (the hypocenter) and the direction from which seismic rays leave the source (focal azimuth and takeoff angles). Fortunately, the method of three-dimensional simultaneous inversion of body-wave arrival times is an effective method for determining a detailed model of earth structure on a scale appropriate for the investigation of focal mechanisms (Thurber, 1981). Obviously the velocity model itself also has important bearing on the understanding of the volcano's structure.

Thousands of instrumentally locatable earthquakes occur each year on the Island of Hawaii. These events are thought to be related primarily to the ascent of magma from the mantle, the injection of magma into crustal rift zones, the eruption of lava at the surface, and the mechanical failure of the crust due to volcanic loading. There are two key factors that permit the application of simultaneous inversion and related techniques to the study of volcanic/tectonic processes and of the volcano's internal structure. Firstly, the earthquake activity is broadly distributed laterally and with depth, and the earthquake focal mechanisms vary with location (and perhaps time) according to the nature of the volcanic and tectonic processes causing them. Secondly, the Hawaiian Volcano Observatory (HVO) has been continuously monitoring Kilauea's seismic and magmatic activity for many years, accumulating a substantial catalog of arrival-time data and allowing the possibility of observing temporal variations.

The Island of Hawaii has been the subject of a large number of investigations of crustal and upper-mantle structure (Eaton, 1962; Ryall and Bennett, 1968; Hill, 1969; Ellsworth and Koyanagi, 1977; Broyles and others, 1979; Crosson and Koyanagi, 1979; Zucca and Hill, 1980; Klein, 1981; Zucca and others, 1982) and earthquake focal mechanisms (Endo, 1971; Ando, 1979; Estill, 1979; Furumoto and Kovach, 1979; Unger and Ward, 1979; Crosson and Endo, 1981, 1982; Klein, 1981; Bosher and Duennebier, 1985). Few structure studies modeled laterally heterogeneous Earth structure, and all the focal-mechanism studies adopted simple one-dimensional velocity models for their analyses. Given the strong evidence for laterally varying structure and the somewhat ambiguous results of some of the focal-mechanism studies, a systematic investigation of three-dimensional crustal structure and earthquake source mechanisms is warranted.

ACKNOWLEDGMENTS

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METHODOLOGY

I use a method for simultaneously determining both the velocity structure of the Earth beneath a seismic network and the hypocenters of a set of local earthquakes. This simultaneous inversion method can be thought of as a generalization of the standard earthquake location procedure (Geiger's method) to include parameters describing the Earth model as unknowns along with the hypocentral parameters. The details of the method have been described in earlier papers (Thurber, 1983, 1984).

Two recent developments have made iterative three-dimensional simultaneous inversion practical. One is parameter separation, introduced independently by Pavlis and Booker (1980), Spencer and Gubbins (1980), and Rodi and others (1980). This is a mathematical device for maintaining a tractable computational problem size even when large numbers of earthquakes are included in the inversion. The second is approximate ray tracing (ART), originally introduced by Thurber and Ellsworth (1980), with variants developed by Thurber (1983) and Thurber and others (1984). These ART methods exploit Fermat's principle of stationarity to derive approximate ray paths and traveltimes in arbitrary velocity structures. The method of Thurber and others (1984) comes closest to being a true ray-tracing method.

One of the most important facets of the crustal-modeling approach is the manner in which the velocity structure is represented or parameterized. The ideal would be to specify the seismic velocity at any point within the volcano, but in practice this is not attainable. As a compromise, a reasonably dense three-dimensional grid of points is defined at which the velocity is to be estimated by the inversion. A simple interpolation function is used for intervening points. In each iteration, the inversion algorithm solves for the fractional change in velocity, using damped least squares. The F-test (De Groot, 1975) is used to terminate the iterations. The final velocity values derived by the inversion should represent an average velocity for the volume surrounding each grid point.

Once a three-dimensional model of crustal structure is available, it can be used in a straightforward way for the determination of focal mechanisms. The equivalent of Geiger's method and an appropriate ray-tracing method are used to locate the earthquake in the three-dimensional model, and the focal angles are obtained from the ray paths for the final hypocenter location. This study finds that fault-plane solutions are better constrained and contain fewer inconsistent polarities when a realistic three-dimensional crustal structure is employed.

In order to take best advantage of the distribution of the HVO seismic-network stations and the distribution of seismic activity, an area 20 km by 30 km encompassing Kilauea Volcano was chosen for study (fig. 38.1). The coordinate origin for the study area is at latitude 19° 13.5′ N., longitude 155° 14.0′ W., with the Y axis rotated 28° clockwise from north. The seismic-velocity model of Klein (1981), developed for Kilauea's south flank, was adopted as a reasonable starting model for the inversion. Grid-point spacings of 3–4 km in the X direction and 2–3 km in the Y and Z directions were adopted, reflecting the station density and dispersion of seismicity (figs. 38.1, 38.2). A set of 85 earthquakes from 1980 and 1981 (the years of data originally made available by HVO for this study) was selected for the inversions, on the basis of their having well-constrained epicenters, a large number of observing stations, and (most importantly) at least one station within a focal depth of the epicenter, so that the earthquake depths were well constrained. Arrival times from two man-made explosions near the summit in 1976 were also included.

A composite model for the P-wave velocity structure of Kilauea was constructed by averaging the results of a series of inversions for which the grid spacing was kept fixed but the grid as a whole was shifted systematically 1–2 km in the X, Y, and Z directions. The resulting velocity values varied smoothly among the set of inversion models. Typically, the inversions achieved an 80-percent reduction in the data variance after three or four iterations. Unweighted (RMS) residuals for individual events generally were reduced from about 0.11 to 0.04–0.05 s, close to the estimated reading error for P-wave arrival times. Estimated standard errors and model resolution (diagonal elements) for velocity parameters averaged 3 percent (fractional error) and 0.7, respectively. Resolution was poorest for grid points in the deepest layer, on the fringe of the model, and in the near-surface layer not close to a seismic station.

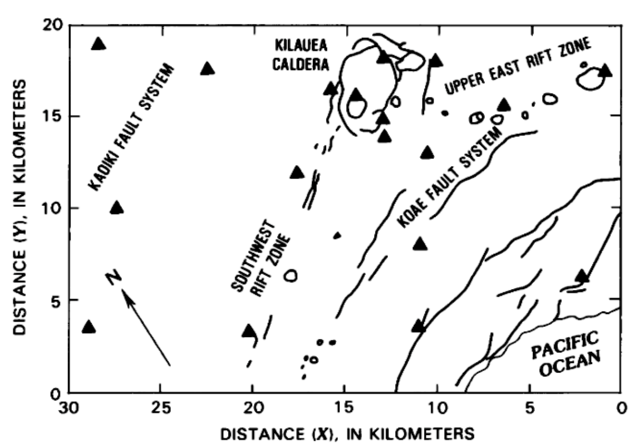

FIGURE 38.1.—Study area on Island of Hawaii, showing principal geologic features and HVO seismic stations (triangles). See figure 38.2 for regional location map. Modified from Thurber (1984, fig. 1), copyright 1984 by American Association for the Advancement of Science.

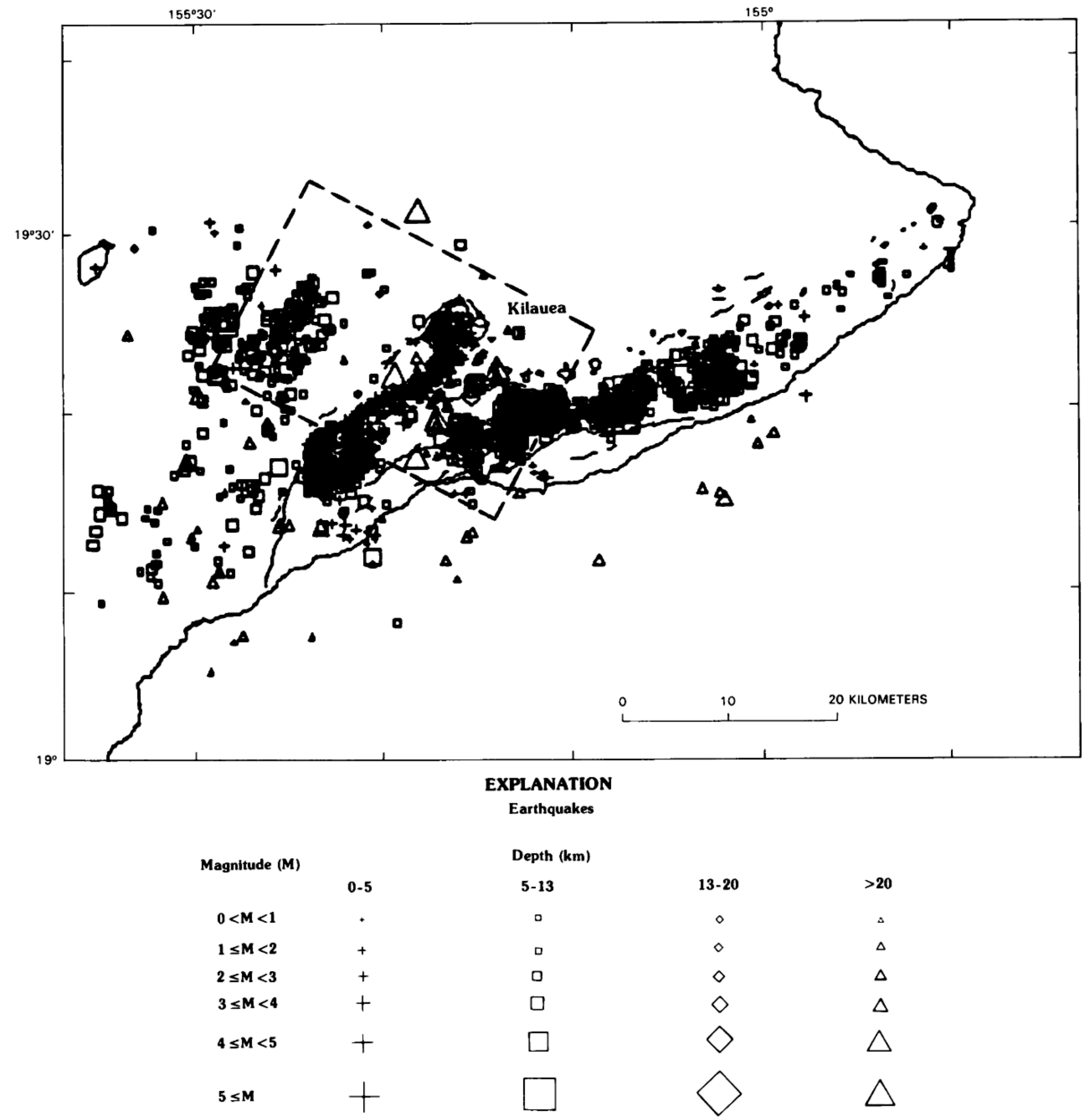

FIGURE 38.2.—Seismicity of southeastern Hawaii for 1981 (Nakata and others, 1982). Dashed lines indicate study area.

CRUSTAL STRUCTURE

Four grid layers of the composite P-wave velocity model are shown in map view in figure 38.3. They are positioned at 1 km above and 2, 5, and 8 km below sea level. In the top layer (fig. 38.3) a velocity high directly beneath the caldera corresponds to the roof of the summit magma complex, and velocity highs also are present along the upper east and southwest rift zones (fig. 38.1). A deeper look into the crust (fig. 38.3), shows a continuation of the velocity highs along the two rift zones to 2 km below sea level. The nonrift areas, particularly the Kaoiki region, have significantly lower seismic velocities.

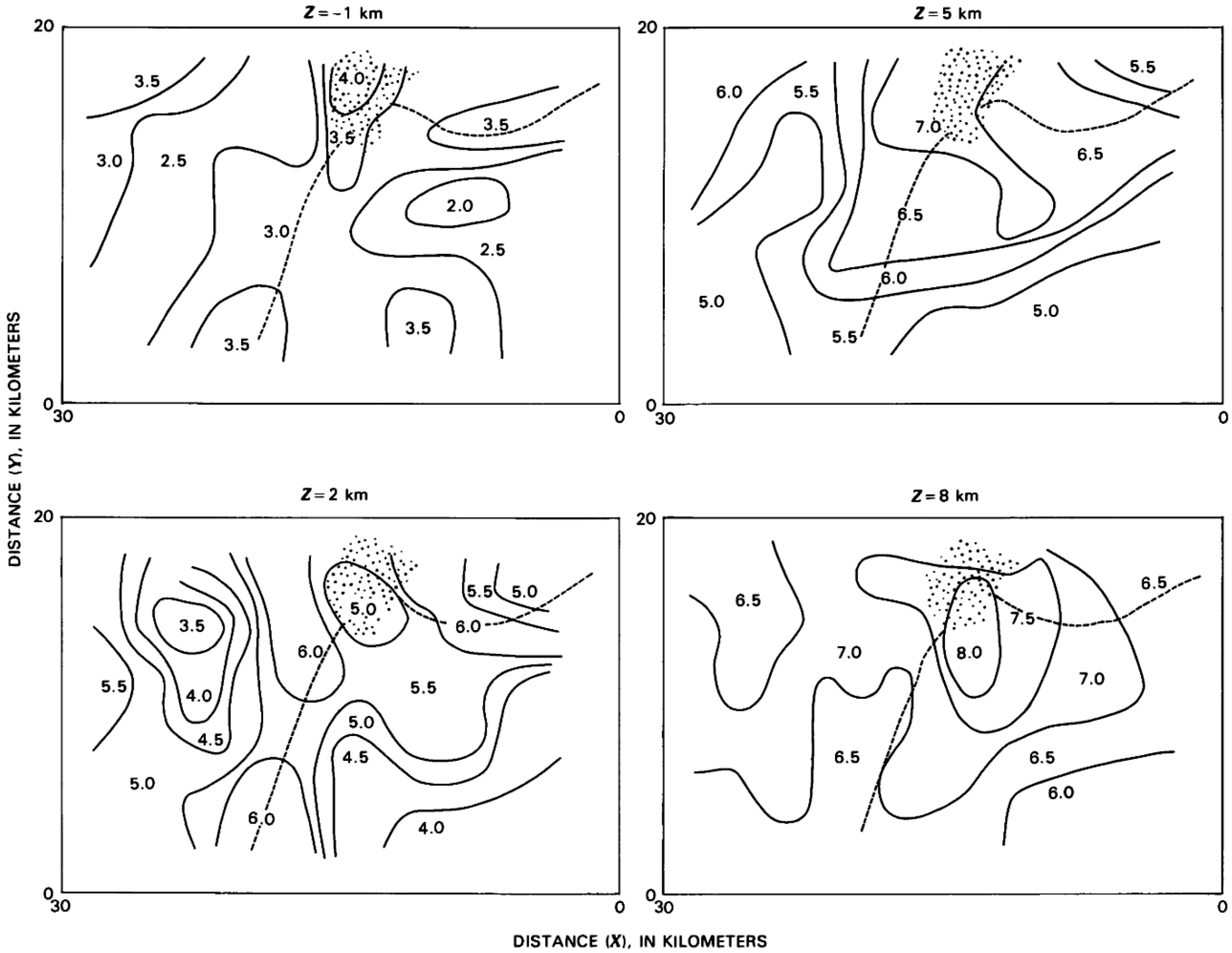

FIGURE 38.3.—Four horizontal slices through composite seismic-velocity model for Kilauea Volcano, positioned at 1 km above and 2, 5, and 8 km below sea level. Velocity is contoured at intervals of 0.5 km/s. Stippling and dotted lines indicate position of caldera and its rift zones, respectively. From Thurber (1984, fig. 2), copyright 1984 by American Association for the Advancement of Science.

Two other important features are apparent at this depth. A high-velocity area along the Koae fault zone indicates the presence of dense, probably competent rock. Recent episodes of intrusive activity in this area have been documented (Swanson and others, 1976), particularly at the northeastern and southwestern boundaries of the fault system, where it intersects the major rift zones. A distinct velocity low is found beneath the caldera, coinciding with the inferred location of Kilauea's summit magma reservoir. Tests with synthetic arrival-time data indicate that the data set is adequate to resolve a small (diameter 2–3 km) low-velocity zone at this point, increasing confidence in the validity of the model results. The actual low-velocity zone could be somewhat smaller, of course, as some degree of smoothing is inherent in the velocity inversion.

At a depth of 5 km (fig. 38.3), the high-velocity core of Kilauea is clearly evident, extending along the UERZ and the southwest rift zone (SWRZ) and intruding into the Koae area. Kaoiki and other nonrift areas continue to exhibit rather low velocities. At 8 km depth (fig. 38.3) the model resolution has become somewhat poor, but the high velocities of the volcano core are still evident, apparently with a slight southward shift from the shallower levels. A similar southward shift with increasing depth is apparent in the estimation of inflation and deflation centers from surface deformation data (Dvorak and others, 1983). Unfortunately, our model does not extend to a sufficient depth to detect the low-velocity layer reported by Crosson and Koyanagi (1979).

These dramatic lateral variations in seismic velocity can be explained in terms of the difference in properties between magma intruded at depth and lava extruded at the surface. As magma nears the surface, the reduction in ambient pressure permits the exsolution of volatiles, creating gas bubbles within the magma. As a result, the extruded lava is vesicular and has high porosity, low density, and, hence, low seismic velocity. Since the flank areas of Kilauea are constructed predominantly from extruded lava, the low seismic velocities from the inversion are to be expected. In contrast, the rift zones at depth are formed principally from numerous dikes. Except for the very near surface portion of each dike, the intruded magma will not have undergone significant vesiculation. Thus the dikes are probably composed of fairly dense rock with lower porosity and therefore higher seismic velocity.

It is interesting to compare the inversion results with models derived by other seismological techniques. For example, the very high velocity (>8 km/s) at about 8 km depth directly beneath the caldera is consistent with the results of the refraction study of Hill (1969). Rift-zone velocities exceeding 6 km/s at shallow depth have been documented in several refraction studies, and flank velocities on the order of 5 km/s or less are also confirmed (Broyles and others, 1979; Zucca and Hill, 1980; Zucca and others, 1982). Perhaps the most significant comparison is with the teleseismic results of Ellsworth and Koyanagi (1977). The uppermost layer of their model spans the entire depth range of my model, but the correspondence between the results from the two disparate methods is surprisingly strong. The Ellsworth and Koyanagi model displays significantly high seismic velocities beneath Kilauea caldera and its rift zones and also indicates high velocity beneath the Koae area. The Kaoiki and south flank areas are marked by quite low velocities. This agreement strengthens the confidence in both studies.

A rather different comparison can be made between my velocity model and the pattern of seismicity beneath Kilauea caldera. Koyanagi and others (1976) inferred the location of Kilauea's magma chamber from the position of an aseismic zone directly beneath the caldera that spanned a depth range of 2–6 km. However, most evidence favors a smaller, shallow magma reservoir (interconnected sills?) rather than such a large zone of partial melt (Ryan and others, 1983). A nearly east-west cross-section (fig. 38.4) through Kilauea caldera (at $Y=16$ km in figure 38.1) illustrates the association between my model of the velocity structure and the seismicity in the immediate vicinity of the caldera. Earthquakes cluster around the low-velocity zone at shallow depth directly beneath the caldera. An aseismic zone coincides precisely with the zone of low P-wave velocity. Thus, the summit magma complex is clearly detectable in the seismic velocity structure of the volcano. The low velocity is most likely due to a combination of elevated temperature and the presence of partial melt and (or) pockets of magma.

EARTHQUAKE RELOCATIONS

Use of a three-dimensional model of the structure of Kilauea has a significant effect on the estimation of earthquake hypocenters (fig. 38.5*A*, *B*). I compare the computed location using the three-dimensional crustal model (3-D) with the HVO catalog location (Tanigawa and others, 1981; Nakata and others, 1982) and also with the location using the one-dimensional starting model (1-D). The 1-D crustal model is essentially the same as the HVO model (Klein, 1981) but does not include the ad-hoc station corrections used in the HVO solutions. One systematic difference is the general shallowing of all hypocenters in the one- and three-dimensional cases relative to the HVO solutions, because of the latter's referral of depths to the ground surface rather than sea level; an additional bias arises from HVO's neglect of station-elevation differences. I also suspect that the relative locations of events are more accurate from my results, as I use a uniform set of stations to locate the entire set of events. In contrast, HVO solutions for a given area are likely to incorporate additional observations from varying sets of stations in the surrounding region, introducing slightly different location biases for each event.

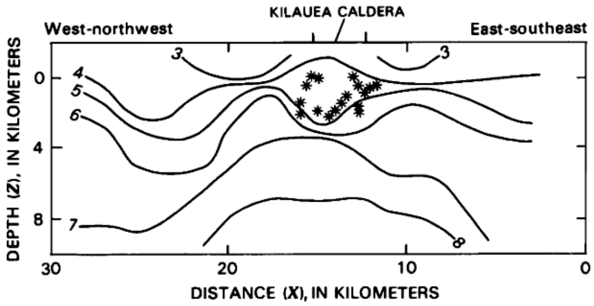

FIGURE 38.4.—Vertical cross section of seismic-velocity model through Kilauea caldera. Section is oriented west-northwest–east-southeast along line $Y=16$ km in figure 38.1. Contours of seismic velocity in kilometers per second. Note association of aseismic region (outlined by earthquake hypocenters indicated by asterisks) with zone of low seismic velocity. For clarity, only shallow earthquakes in immediate vicinity (within 2 km) of caldera are shown. From Thurber (1984, fig. 3), copyright 1984 by American Association for the Advancement of Science.

The epicentral shifts of figure 38.5*A* (1-D versus 3-D) indicate the general effect on hypocenters caused by lateral velocity variations, while in figure 38.5*B* (HVO versus 3D) they are indicative of possible location bias in the HVO catalog. The HVO and 3-D locations agree quite well in the UERZ and Kaoiki areas, although there appears to be a slight bias in the northeast Kaoiki region, with the HVO locations displaced 1–2 km southward. The 1-D locations are 1–2 km north of the 3-D locations in both the UERZ and northeast Kaoiki, but they are displaced 2 km eastward in southwest Kaoiki. In comparing the 1-D and 3-D solutions, it appears that accounting for the strong velocity low along the northeast Kaoiki area acts to pull epicenters in that area closer towards Kilauea, whereas the southwest Kaoiki events are pushed away because of the high velocity of the SWRZ. Similarly, the UERZ events are displaced southwards as a result of modeling the high UERZ and summit area velocities.

The comparisons for the summit and upper SWRZ are rather different. Here the 1-D and 3-D locations agree well, but the HVO solutions are displaced 1–2 km southward, most notably for the deeper summit events. This seems to indicate a significant bias in the HVO catalog for this area. Such a bias could have important implications for relating seismicity patterns beneath the summit, and the inferred magma conduit structure, to features visible at the surface (Ryan and others, 1981).

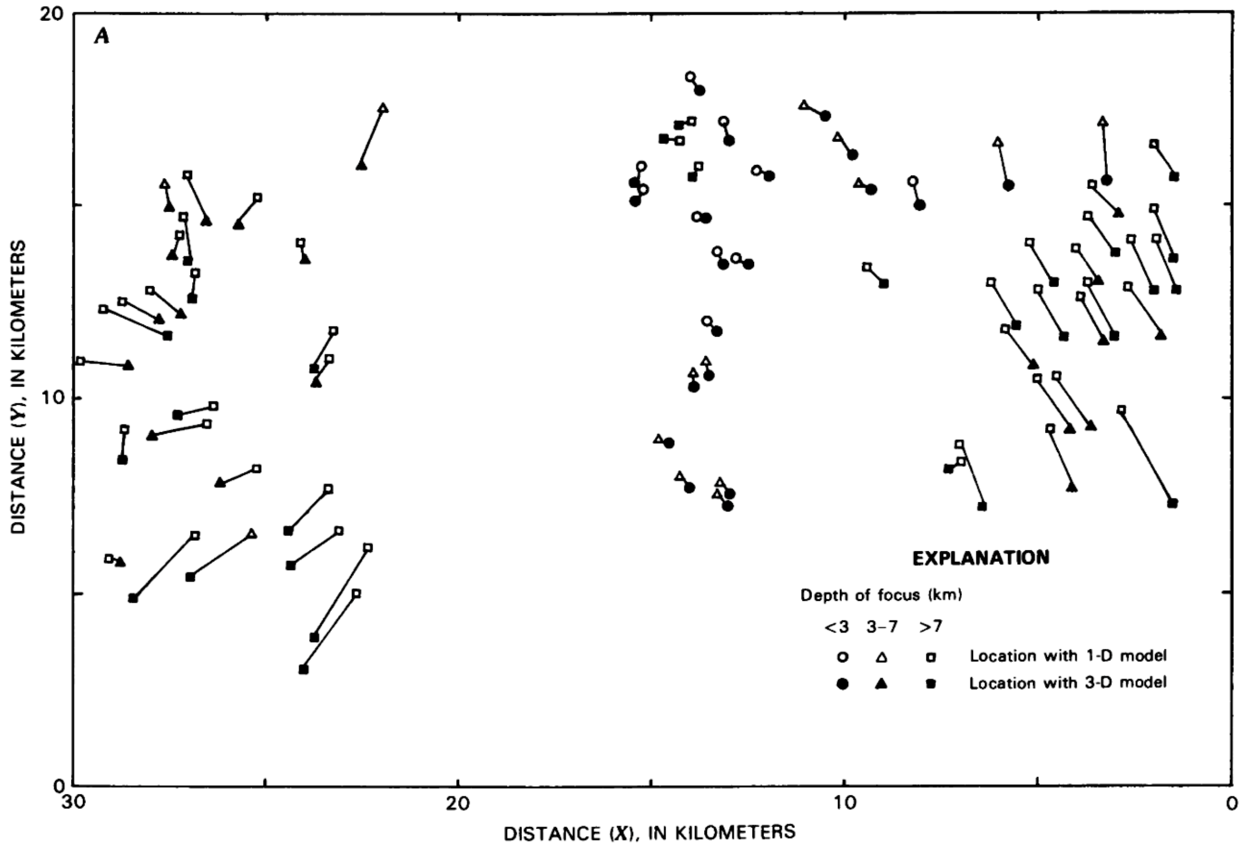

FIGURE 38.5.—Earthquake epicenters in study area relocated using 3-D model in figure 38.3. ***A***, Compared with locations using a 1-D model with no station corrections. ***B***, Compared with HVO catalog locations.

FINE-SCALE STRUCTURE

To derive a crustal model with significantly better spatial resolution (about 1 km) in a small zone directly beneath Kilauea caldera, I take as a starting point the fairly coarse three-dimensional model for Kilauea's velocity structure described above (which I shall call the regional model). A finer scale velocity grid is imbedded within this regional model, and an inversion is carried out to solve for perturbations to the velocity values on the fine grid only, keeping the values fixed on the coarse grid outside of the imbedded grid. Data from a large number of earthquakes within the imbedded region are included to aid in resolving the detailed structure, in addition to events distributed throughout the regional model.

Such a fine velocity grid, with five nodes by five nodes horizontally by three vertically and a uniform spacing of 1 km, was set up to model the shallow structure of the south caldera area of Kilauea (fig. 38.6). The quality of the fine-scale inversion model was limited by the tight clustering of the available events (from 1980 and 1981), as is clear in figure 38.7*A*. Many events in the south caldera fall within about 100 m of a planar surface 2 km in length and 500 m high dipping 60° to the east; presumably this is either a dike or a normal fault overlying a deflating sill. As a result of this clustering, the resolution of the model is only fair (diagonal elements average 0.4 for the 75 nodes, central nodes closer to 0.7). Nevertheless, consistent results were obtained from a set of inversions, typically achieving an additional 25 to 30 percent variance reduction compared to the regional 3-D model.

Slices through the model from one inversion at depths of 0, 1, and 2 km with respect to sea level are shown in figure 38.7*B–D*. Lateral velocity variations of roughly 1 km/s are present in the model at any given depth. There appear to be two separate zones of low *P*-wave velocity: one lies directly beneath Halemaumau Crater, and the other underlies the Ahua Kamokukolau area. The summit zone corresponds well with the location of the summit reservoir. The velocity low beneath Ahua is at the corner of the grid and therefore is less well resolved, but it may correspond to Ryan and others' (1981) UERZ pipe, which appears to lie about 1 km north of the velocity low. Additional earthquake data from more recent years may allow construction of a more definitive model for the fine-scale structure in this region.

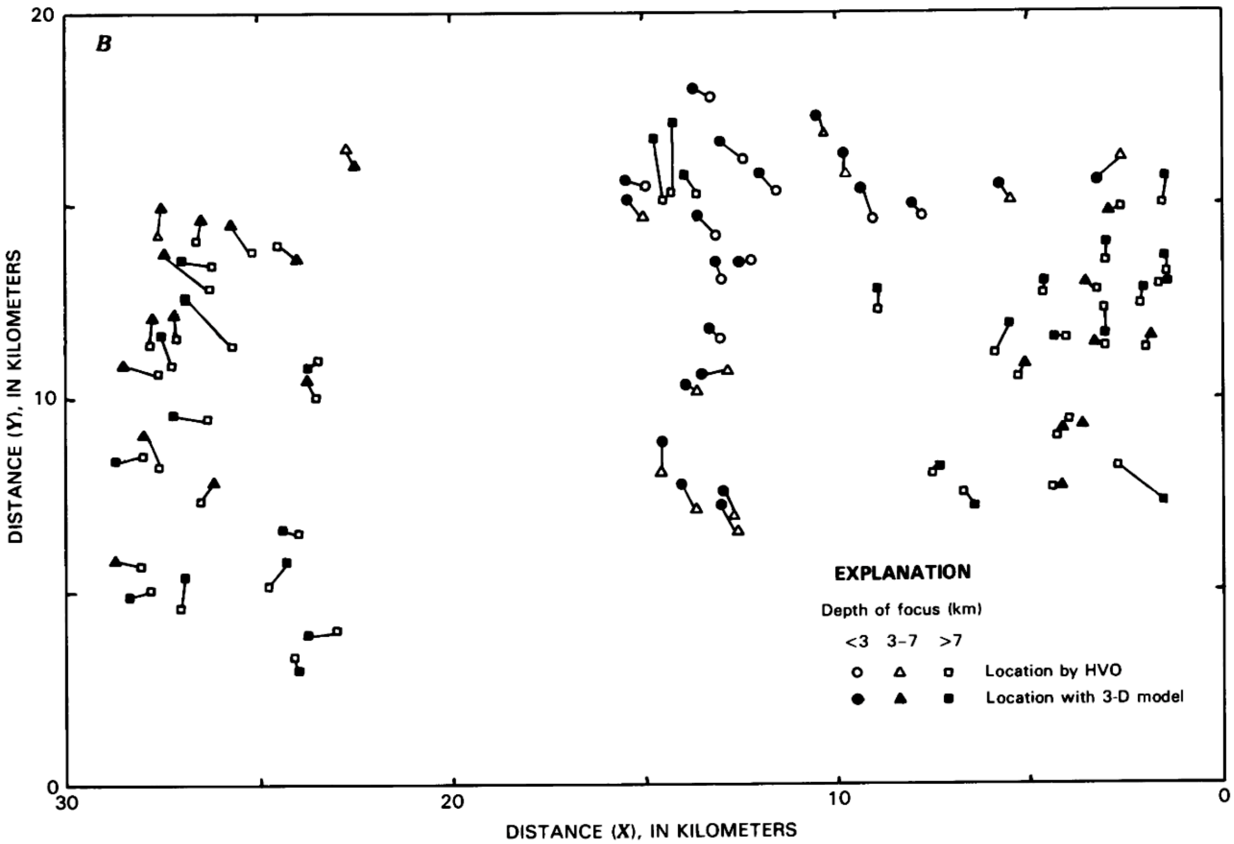

FIGURE 38.5.—Continued.

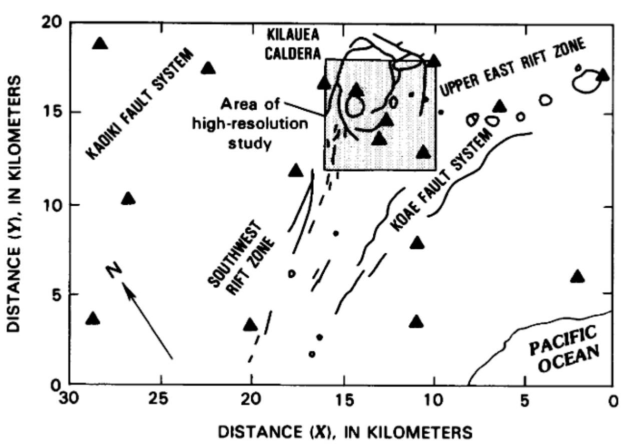

FIGURE 38.6.—Study area near Kilauea caldera (shaded) selected for high-resolution seismic imaging study.

FOCAL MECHANISMS

The state of stress within Kilauea Volcano arises from a complex interplay among crustal density variations, magma pressure, surface loading, and failure along zones of weakness. Fault-plane solutions of earthquakes are generally indicative of the deviatoric stresses in the earth, so they can be used to explore spatial and temporal variations in stress. Care must be taken in interpreting fault-plane orientations directly in terms of stress, however (Ellsworth, 1982; Gephart and Forsyth, 1984; Michael, 1984).

Using a realistic three-dimensional model for the seismic structure of Kilauea, a detailed study of the source mechanisms of earthquakes in the area has been initiated. The earthquake locations are computed using ray tracing in the three-dimensional model, and the focal angles of the rays for the final locations are taken to construct the fault-plane solutions. In general, markedly improved constraint and increased consistency are found in the 3-D solutions compared to the results obtained for the same events using a one-dimensional crustal model. Some of the more dramatic examples are shown in figure 38.8, but these are not atypical.

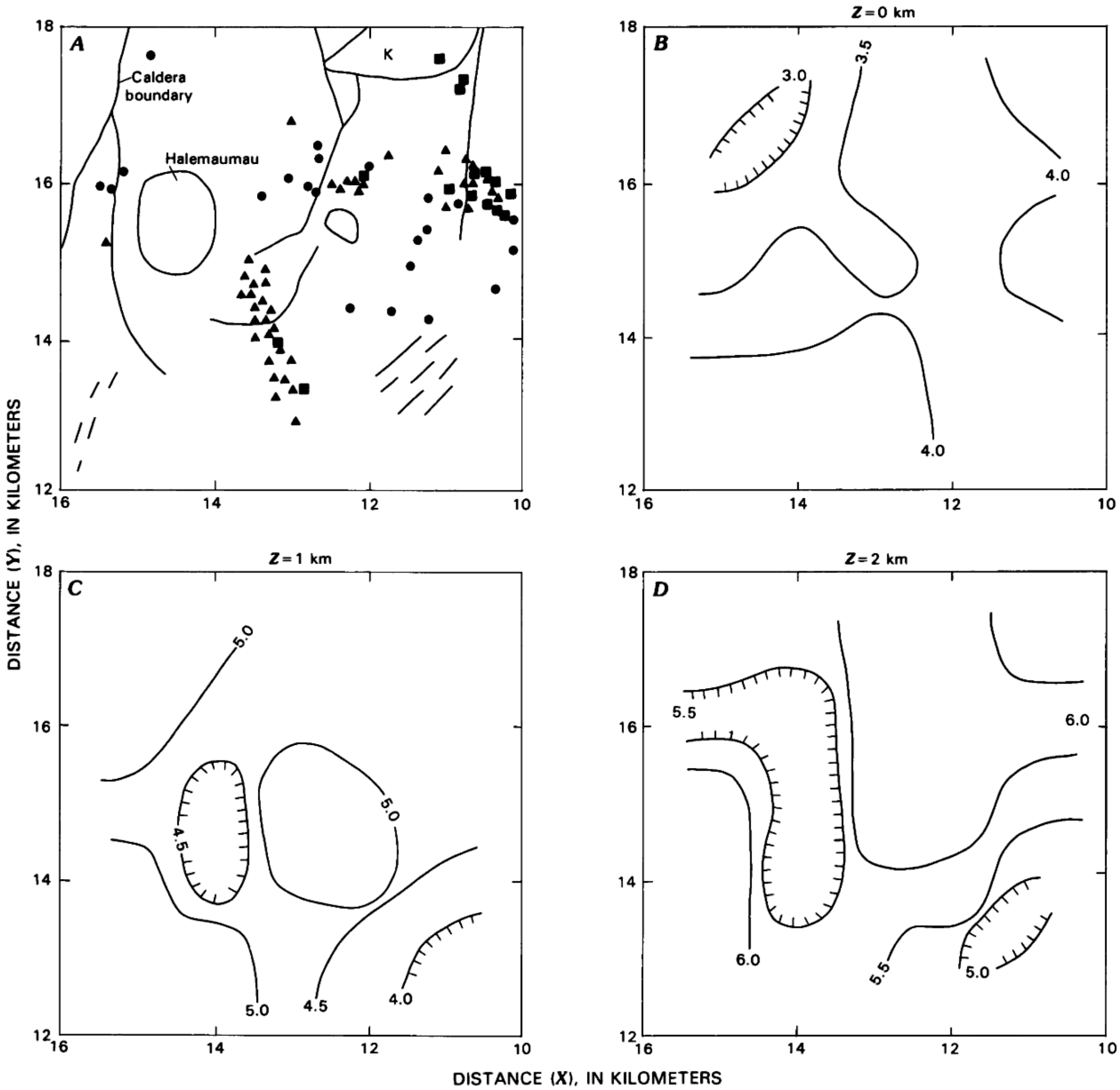

FIGURE 38.7.—Earthquake epicenters and seismic-velocity contours at three levels in high-resolution study area. ***A***, Surface geology and epicenters. Outlines of Kilauea caldera and Halemaumau and Kilauea Iki (K) shown for reference. Faults in lower right are in Ahua area. Focal depths: dots, <1 km; triangles, 1–2 km; squares, 2–3 km. ***B***, Seismic velocities along horizontal slice at sea level. ***C***, Velocities at depth of 1 km below sea level. ***D***, Velocities at depth of 2 km below sea level. Velocity contours in kilometers per second. Hachures indicate closed velocity lows. Two distinct low-velocity zones are evident, although model resolution is only fair.

SPATIAL VARIATIONS

Sets of earthquake focal mechanisms are first used to explore the tectonics of the major geologic areas surrounding Kilauea caldera: the Kaoiki fault system, the SWRZ, the UERZ, and the Koae fault system. A brief survey of source mechanisms from these regions clearly shows a general relation between the faulting in microearthquakes and the surface expression of the tectonics.

The Kaoiki fault system is thought to be a tectonic boundary between the shields of Kilauea and Mauna Loa. Mechanisms for three small events (magnitudes 2.5–3) are shown in figure 38.9*A*; the earthquake depths ranged from 5 km to 8 km. All three events are almost purely strike slip, presumably right lateral. The micro-earthquake data suggest east-west compressive stress, in agreement with the mechanism of the November 1983 Kaoiki earthquake of magnitude 6.7 (U.S. Geological Survey, 1983). These observations

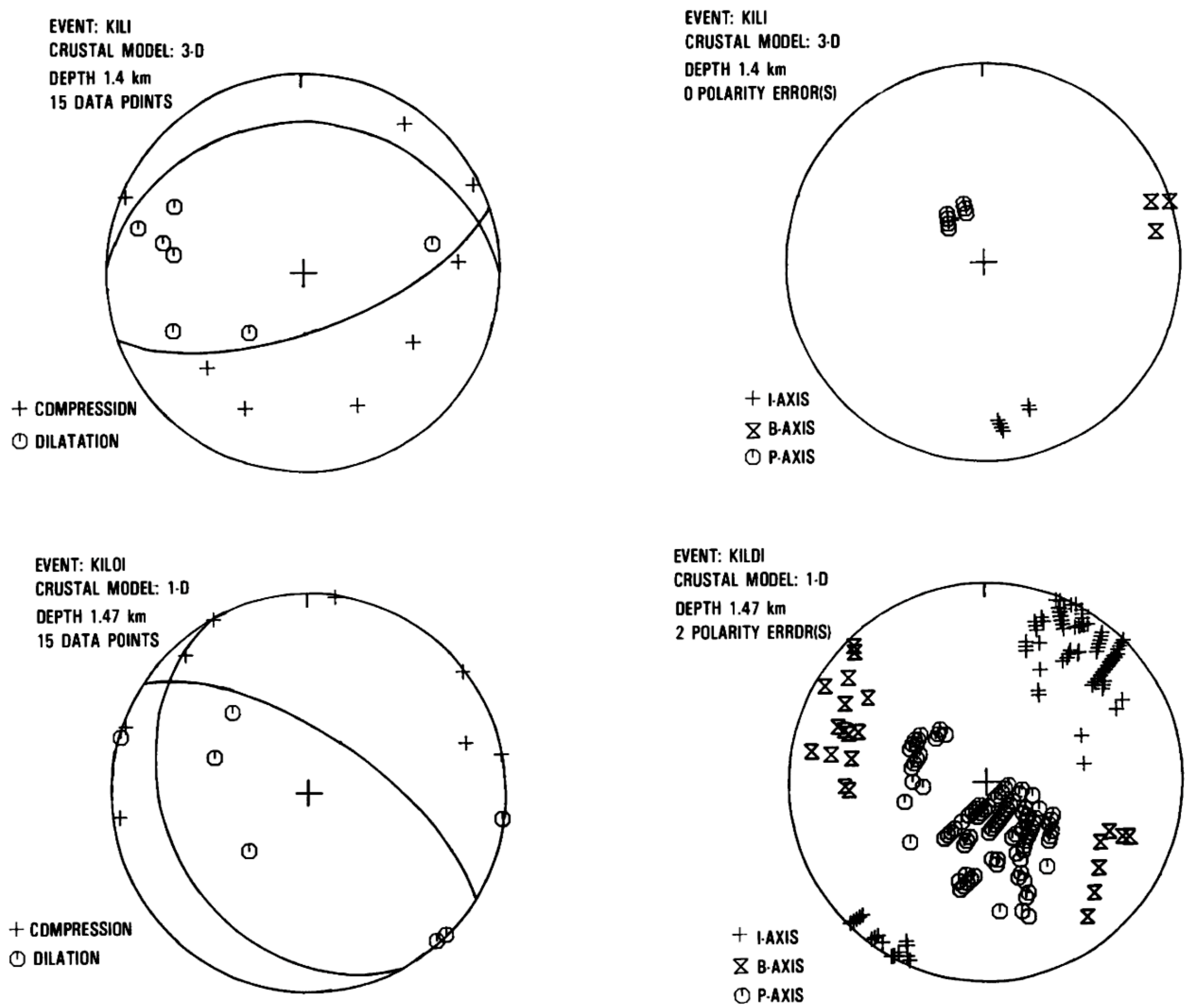

FIGURE 38.8.—Examples of fault plane solutions comparing results using 3-D model (above) to 1-D model (below). Diagrams at left display first motions and reasonable fault-plane solutions; those at right show orientations of permissible maximum, minimum, and intermediate compressive stress axes.

are generally consistent with the notion of stress caused by the nearby growing shields (Endo and others, 1986). Alternatively, it could be thought of as resulting from lateral density variations within Kilauea and Mauna Loa, arising from their dense central cores and in turn producing horizontal stresses oriented radially to their summits. Finally, note that the Kaoiki main shock was located slightly to the northwest and deeper than our events, just off the left side of figure 38.9*A*; if the small events are right lateral, then their northwest-dipping fault plane would dip toward the Kaoiki main shock.

Turning to the SWRZ (fig. 38.9*B*), mechanisms for five 1981 earthquakes of magnitudes 2–3 and depths of 1–3 km are shown. The year 1981 was atypical in that the SWRZ was active and the east rift zone (ERZ) was quiet. With one exception, the mechanisms are clearly consistent with northwest-southeast tension, favorable for intrusive activity on this northeast-southwest rift zone. The discrepant event, number 65, occurred on August 10, 1981, as part of a major intrusive episode, and it may perhaps reflect breaking of an asperity at a bend along the SWRZ.

Along the UERZ (fig. 38.9*C*) is a very clear pattern consistent with east-west compression, permitting intrusion on this east-west rift zone. Five events of magnitude 2–3 and depths of 1–3 km occurred in 1980, when the ERZ was active and the SWRZ was quiet. Except for number 21, all are strike-slip events, but in most cases neither possible fault plane parallels the rift-zone trend, indicating predominance of echelon faulting as suggested by Hill (1977). A similar relation was found by Smith (1984) for a shallow earthquake swarm in Long Valley Caldera, California.

Finally, in the area of the Koae fault system (fig. 38.9*D*) the pattern is as complicated as the surface geology. Five events from

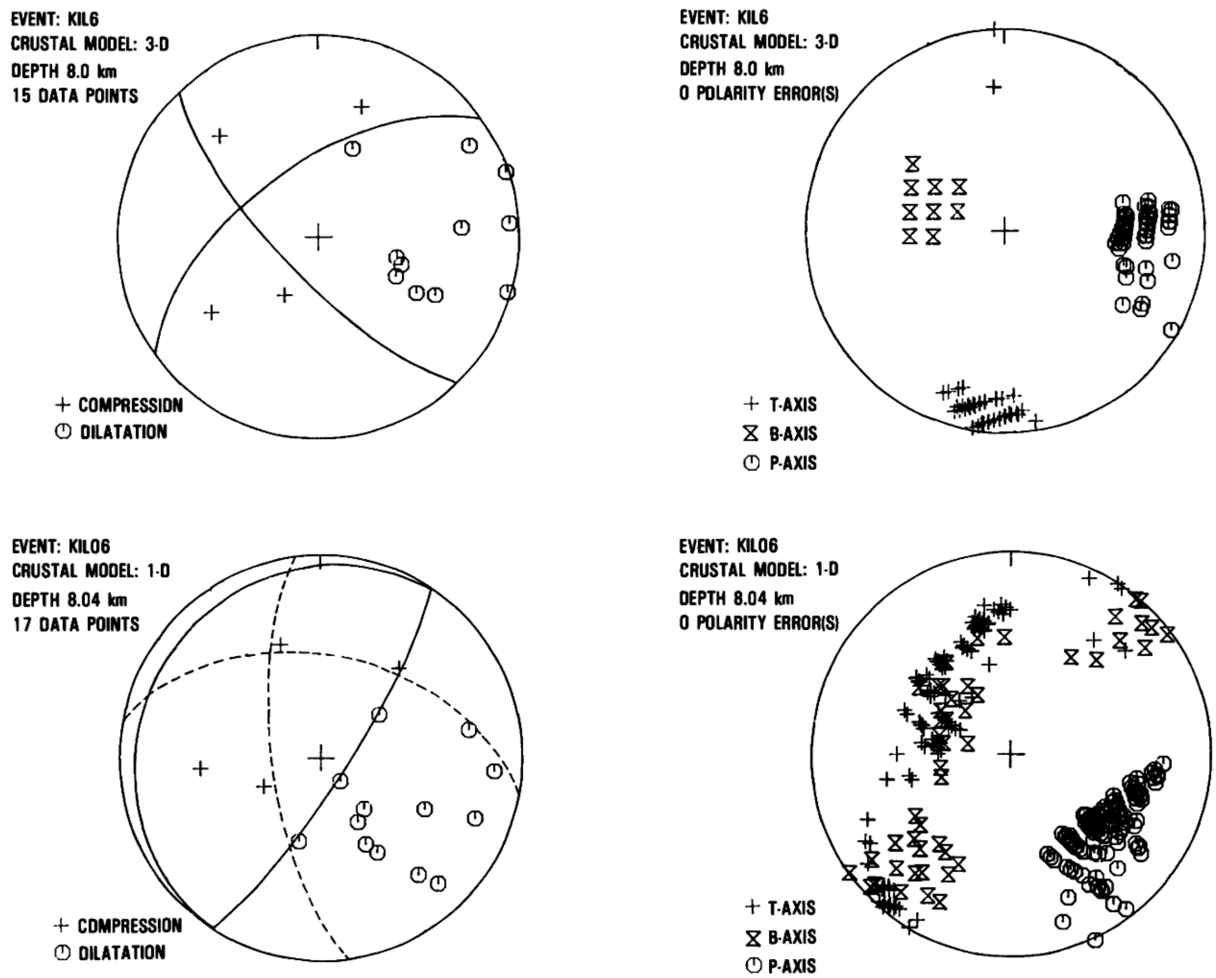

FIGURE 38.8.—Continued.

1980 and 1981 are shown, with magnitudes of 2.5–3 and depths of 4–8 km. Along the eastern section of this area, the earthquakes indicate north-south tension, but those in the western part indicate north-south compression. The eastern events are possibly related more closely to the magma conduit system of the UERZ than to the Koae faults. The western events occurred in 1981, when the SWRZ was experiencing major intrusive activity; the apparent compressive stress south of the SWRZ may be related to this intrusive activity. Mechanisms of events from other years need to be investigated to seek possible temporal variations.

Taken together, these results clearly demonstrate the spatial variability of the stress field surrounding Kilauea Volcano. We suspect these stresses are due in large part to the dramatic lateral density variations present inside the shield, as well as to uneven localized stresses introduced by dike formation. Combining gravity measurements with our model for the seismic velocity structure may enable the direct modeling of these forces.

TEMPORAL VARIATIONS

For a number of years preceding 1981, Kilauea's eruptive and intrusive activity took place exclusively along the ERZ or at the summit. In 1981, activity shifted abruptly to the SWRZ; subsequently, it has returned to the ERZ. I chose to investigate whether focal mechanisms of summit-area earthquakes might show a shift in fault-plane orientation because of a change in stress orientation related to this shift in direction of magmatic activity. For the ERZ to be active, the hypothesis would predict northwest-southeast compression at the summit, whereas a 90° shift to northeast-southwest compression would be required for the SWRZ to be active. It should be noted, however, that direct inferences of stress orientations cannot be made from earthquake focal mechanisms (MacKenzie, 1969), although efforts have been made recently to formally invert focal-mechanism data for the stress tensor (for

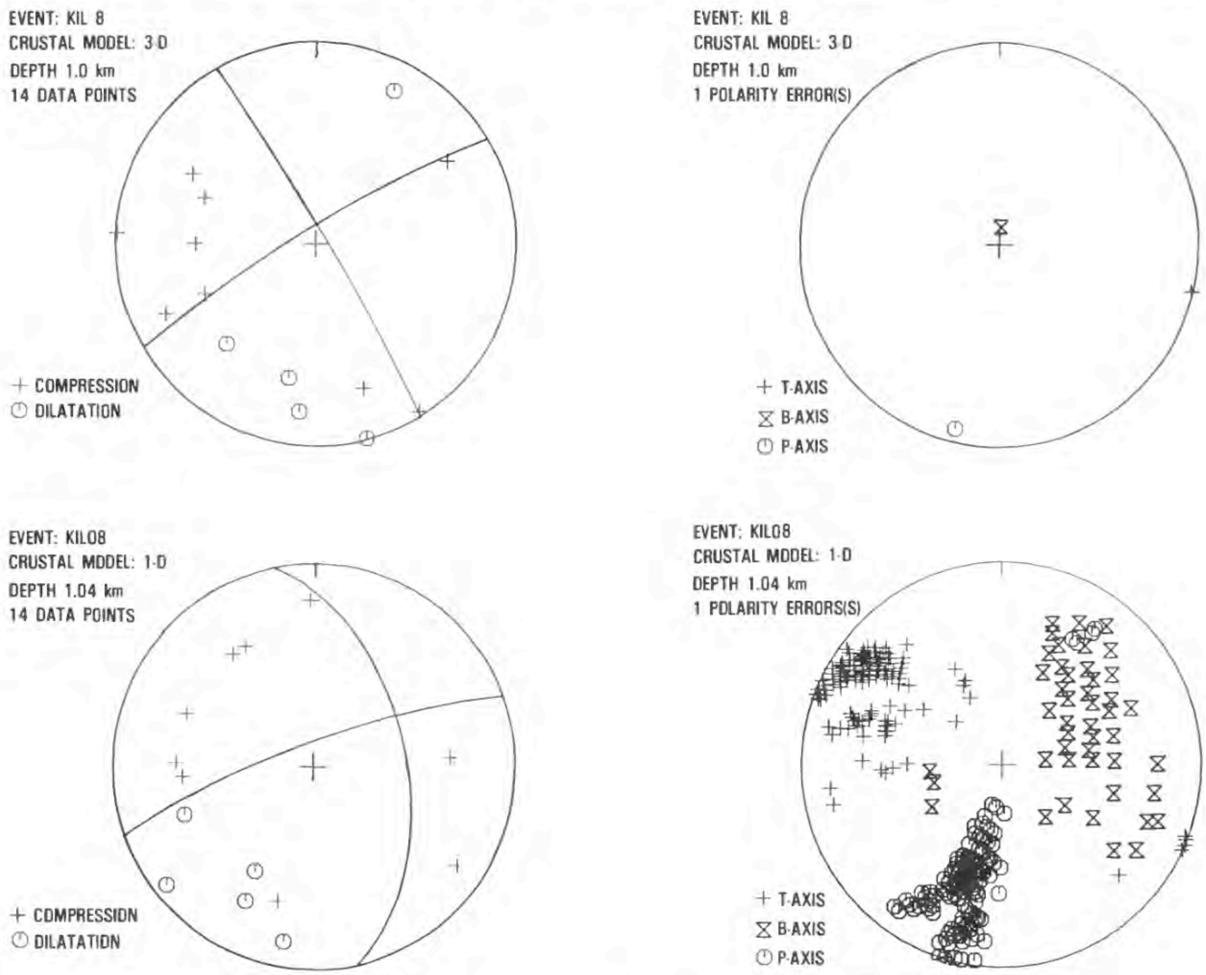

FIGURE 38.8.—Continued.

example, Ellsworth, 1982; Gephart and Forsyth, 1984; Michael, 1984).

Initially, four shallow summit-area earthquakes of magnitude 3 and above were examined (fig. 38.10), one pair from 1980 and a second pair from 1981. These events were the largest and best recorded from the area during this time period. As expected, the 1981 events indicate northeast-southwest compression, and the 1980 events are consistent with northwest-southeast compression. The second 1980 event occurred on December 24, less than one month before the beginning of SWRZ activity. A suite of events encompassing the largest summit earthquakes from January 1981 was therefore selected to seek possible precursory evidence for the impending change in the pattern of magmatic activity.

In order to minimize any confounding effect of spatial variations in stress, only shallow summit earthquakes were examined from two small areas of the caldera (fig. 38.11), all with magnitudes between 2 and 2.5. The range of focal depths was also kept quite small to further reduce the possibility of spatial effects. This series of predominantly strike-slip events, seven along the caldera's southeast rim and four from its west side, indicate a cyclic variation in P-axis orientation during January 1–18, with a very sharp change apparent from east-southeast to east-northeast over the last 6 days of the period (fig. 38.12). Unfortunately, no other earthquakes occurred in the vicinity during this time period that were large enough to yield a reliable mechanism.

What are possible causes of this apparent stress reorientation? One possibility would be a rapid change in the rate of magma flow into the summit storage complex. However, the records from tiltmeters around the caldera (fig. 38.13; A.T. Okamura, written commun., 1981) show only minor deviations from a smooth pattern of inflation. The beginning of a major tilt change is evident on January 19, well after the reorientation of the fault-plane solutions is apparent, but before the first SWRZ intrusive episode beginning January 20. Thus this reorientation could have provided early

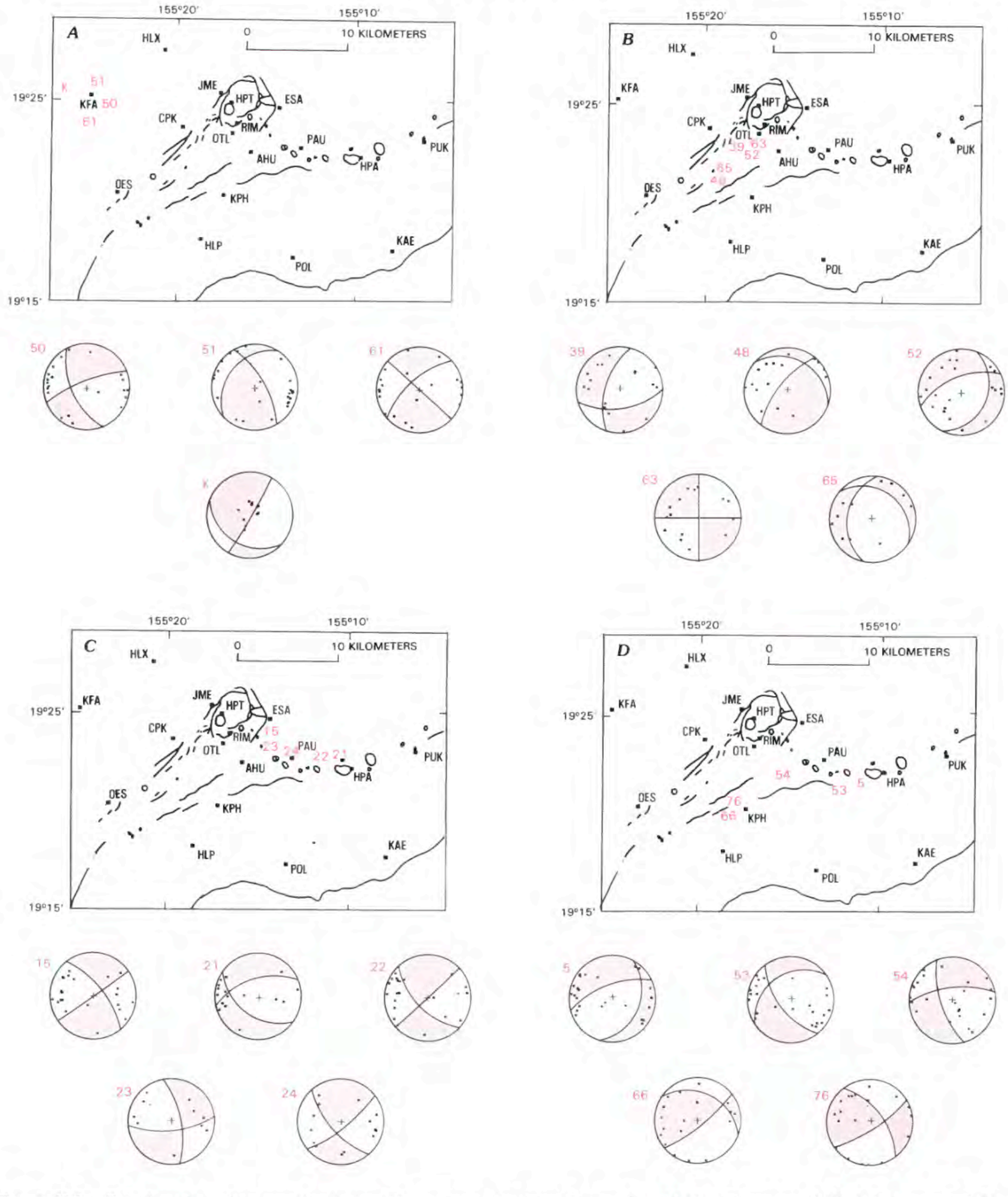

FIGURE 38.9.—Fault-plane solutions for microearthquakes from different sections of study area. Letters refer to HVO seismic stations. ***A***, Kaoiki fault system (including mechanism of 1983 Kaoiki earthquake; U.S. Geological Survey, 1983). ***B***, Southwest rift zone. ***C***, Upper east rift zone. ***D***, Koae fault system.

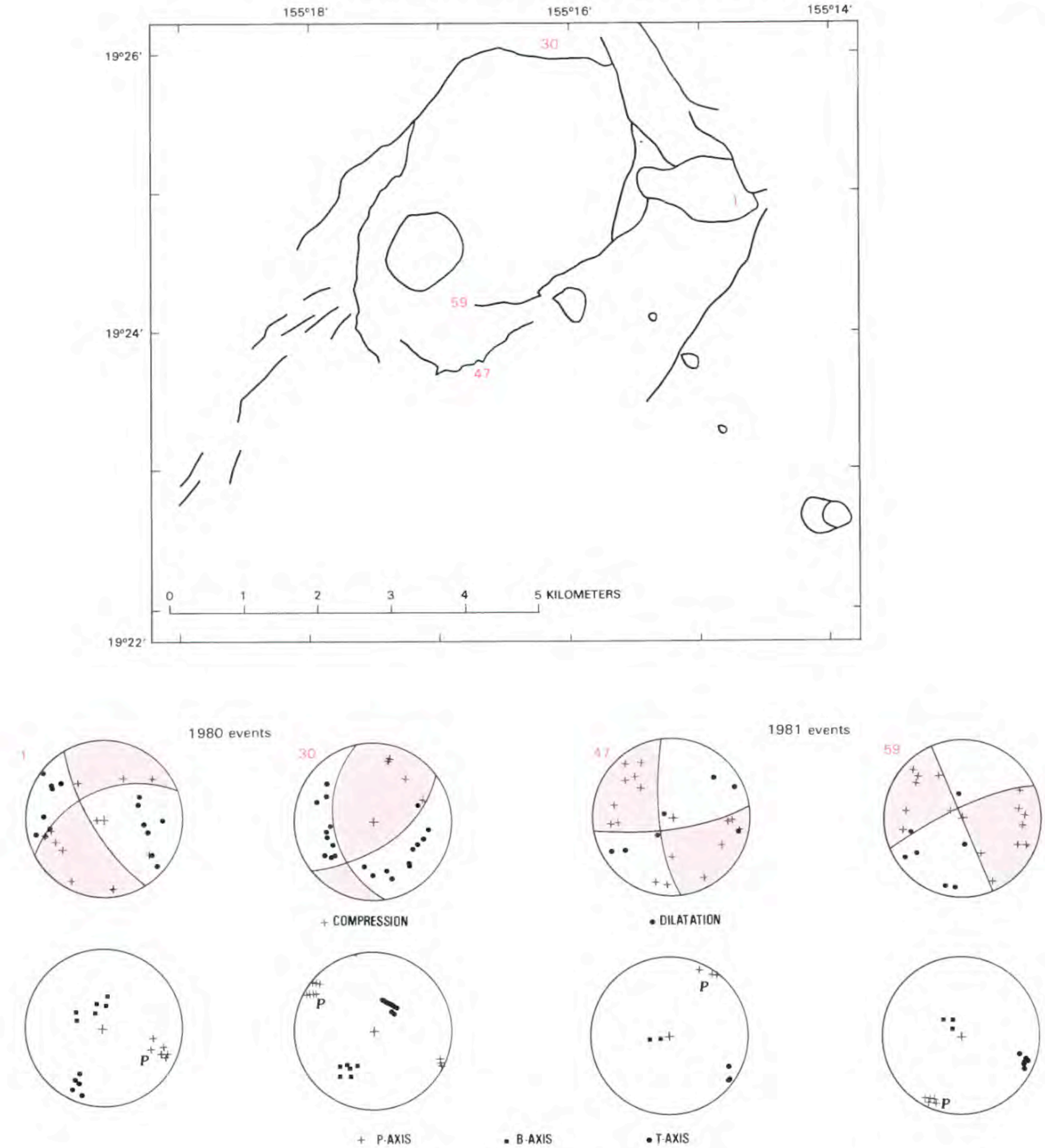

FIGURE 38.10.—Comparison of fault-plane solutions from two caldera-area events in 1980 and two in 1981; differences indicate area-wide change in P-axis orientation.

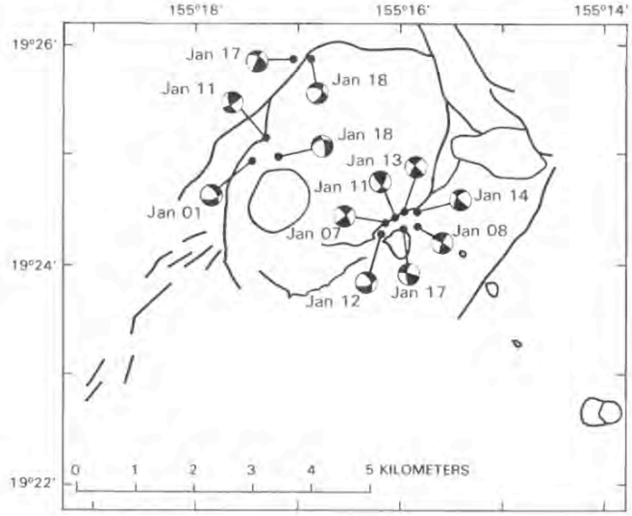

FIGURE 38.11.—Sequence of fault-plane solutions for Kilauea summit earthquakes in two small areas during January 1981 that demonstrates significant temporal variation in faulting orientation (see fig. 38.12).

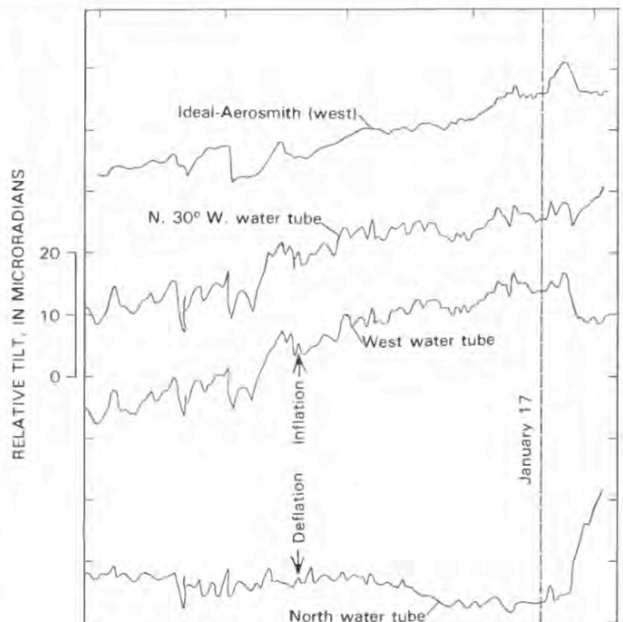

FIGURE 38.13.—Summit tiltmeter records preceding January 1981 southwest rift zone intrusion (A.T. Okamura, unpub. data, 1981), showing no anomalous features before January 17, when earthquake fault-plane solutions had already clearly shown an orientation change.

warning of the impending activity along the SWRZ. This also may be evidence that external forces control the pattern of magma intrusion, rather than the excess pressure of the magma itself.

An extension of this study is necessary to determine the trend of P-axis orientation before and after the time period considered. Some questions that need to be addressed include the following: Is this cyclic variation typical or unusual at Kilauea's summit? Are most major intrusions and eruptions preceded by a rapid change in apparent stress reorientation? If so, can this observation be used to help predict eruptions? Planned studies of summit earthquakes for the time period leading up to the 1983 ERZ eruption will seek confirmational evidence.

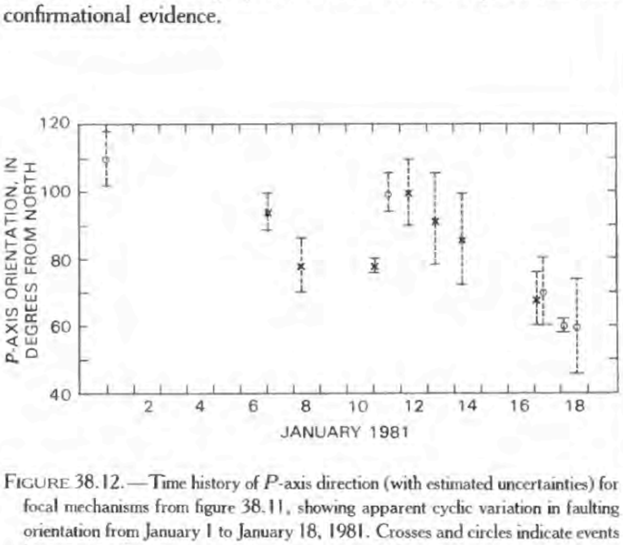

FIGURE 38.12.—Time history of P-axis direction (with estimated uncertainties) for focal mechanisms from figure 38.11, showing apparent cyclic variation in faulting orientation from January 1 to January 18, 1981. Crosses and circles indicate events in eastern and western parts of caldera, respectively.

CONCLUSION AND FUTURE RESEARCH DIRECTIONS

Seismological data clearly are valuable for exploring the structure and dynamics of an active volcano. The vast resource of local earthquake data for Kilauea is just beginning to be tapped; these investigations of volcano structure should be continued and extended both to larger and finer scales. Three-dimensional inversion techniques are perhaps the most promising tool for geophysical imaging of the interior of Kilauea. Fine-scale studies of seismically active areas could be achieved through the deployment of a large network of portable seismographs (for example, PASSCAL), or such an array could be used simultaneously for a larger scale local earthquake and for teleseismic investigation. Detailed focal-mechanism studies should constitute an integral part of the latter type of study.

It seems clear, though, that digital seismic data will hold the key to the analysis of the dynamics of magma transport within the volcano. Earthquake source parameters (moment, dislocation, stress drop, fault area) have already proven useful for the study of Usu Volcano, Japan (Takeo, 1983). Where practical, moment tensor analysis (Wu and Ioannidou, 1985) would provide a means to explore the relative degrees of shear versus tensile failure (that is, faulting versus magma injection) in local earthquakes associated with magmatic activity. Simply demonstrating the existence of significantly non double-couple events would be of considerable value. A detailed study of an earthquake swarm apparently related to dike formation, like that of 1981, might shed light on the mechanics of magma intrusion. Furthermore, shear-wave attenuation could be used to map the magma conduit system and to search for possible temporal variability. These tasks are difficult but realistic: accomplishing them would make a major contribution toward our understanding of how a volcano works.

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