Displacement of the South Flank of Kilauea Volcano: The Result of Forceful Intrusion of Magma Into the Rift Zones

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Interpretation of geodetic and geologic information leads to a new model for the structure of Kilauea Volcano

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DISPLACEMENT OF THE SOUTH FLANK OF KILAUEA VOLCANO: THE RESULT OF FORCEFUL INTRUSION OF MAGMA INTO THE RIFT ZONES

By Donald A. Swanson, Wendell A. Duffield, and Richard S. Fiske

ABSTRACT

Seismic evidence has long indicated that the south flank of Kilauea Volcano is mobile. Examination of triangulation, trilateration, and leveling data obtained throughout the 20th century shows that the south flank of Kilauea has been displaced upward and away from the rift zones by as much as several metres. The amount of horizontal displacement approximates the probable amount of dilation that accompanies the intrusion of magma as dikes in the rift zones and is greatest for periods of most intense intrusive activity, as evidenced by the frequency of eruptions. Displacement and seismic events on the south flank take place soon after intrusive activity, indicating that the displacement is the result of forceful intrusion in the rift zones, not the cause of relatively passive intrusion. The dikes are thought to be nearly vertical or to dip steeply southward, on the basis of both interpretation of seismic and displacement data for Kilauea itself and comparison with dikes exposed in older, eroded Hawaiian shield volcanoes.

In contrast to the south flank, seismic and geodetic data indicate that the north flank is virtually immobile. This contrast is believed to reflect the fact that Kilauea was built on the south slope of Mauna Loa and was consequently influenced by the gravitational stress system of Mauna Loa, which favors displacement away from the volcanic edifice. The north flank of Kilauea is effectively buttressed by Mauna Loa, whereas the south flank is unbuttressed and free to move away from the edifice when prompted by forceful intrusion of magma.

The active part of the east rift zone of Kilauea has apparently migrated several kilometres southward with time. This is shown by the location of recent vents and by the location of the axis of a positive gravity anomaly along the north edge of the active part of the rift zone. The southward migration helps explain several features of the geometry of the east rift zone, particularly its prominent bend near Kilauea Caldera. The southwest rift zone and the caldera also show some evidence of southward migration.

The Hilina fault system is considered to be a gravity-controlled system not directly related to the rift zones. Gravitational instability resulting from uplift and seaward displacement is eventually relieved by normal faulting along the seaward part of the south flank. The Hilina faults are thought to bottom at shallow depth without intersecting magma reservoirs, except possibly along part of the lower east rift zone, where the fault system impinges upon the rift zone. Strains have been accumulating within the Hilina system throughout this century, and a high level of instability may have been reached. We anticipate a subsidence event in the not too distant future, possibly similar to the damaging events of 1823 and 1868. (While this paper was in press, such an event occurred on November 29, 1975.)

INTRODUCTION

Kilauea Volcano is the southeasternmost shield volcano in the Hawaiian-Emperor chain of islands and seamounts that extends some 6,000 km across the central Pacific Ocean. Kilauea and its giant neighbor Mauna Loa (fig. 1) are among the most active volcanoes on earth, presently erupting tholeiitic basalt at an average rate of 0.1 km^3/yr (Shaw, 1973; Swanson, 1972). Kilauea is built on the southeast flank of Mauna Loa, and flows from the two volcanoes probably interfinger extensively at depth.

Much has been learned about the structure and magma-conduit system of Kilauea during the past 20 years. A wealth of geodetic, geologic, geophysical, and seismic data indicates that magma rises from the mantle through a nearly vertical conduit to a holding reservoir about 2 to 5 km beneath the summit caldera (fig. 2). This reservoir, almost certainly an intricate plexus of dikes, sills, and irregularly shaped chambers, swells as it fills, causing easily measurable tumescence at the surface. With continued magma supply, the strength of the reservoir rocks is eventually exceeded, and magma is leaked from the system, resulting in a shallow intrusion and usually, but not always, in an eruption.

Virtually all eruptions and shallow intrusions occur either in the summit area or along the east or southwest rift zones, which extend outward from the caldera (table 1). Summit eruptions, most of which take place in Halemaumau Crater, are accompanied by numerous earthquakes as magma works its way toward the surface. Local uplift is generally observed very near the site of summit eruption or intrusion, but the overall summit area may undergo net expansion (table 1, May 1970, August 1971, July and September 1974), contraction (table 1, 1924, 1938, 1959, 1967, December 1974), or virtually no change (table 1, 1907–24, 1927, 1929, 1930, 1931, 1934, 1952, 1961), depending upon complex interrelations among rate of eruption, rate of magma recharge from the mantle, and degree of shallow intrusion versus extrusion.

During rift eruptions, the summit area subsides (table 1, 1955, 1960, 1961, 1962, August and October 1963, March and December 1965, August and October

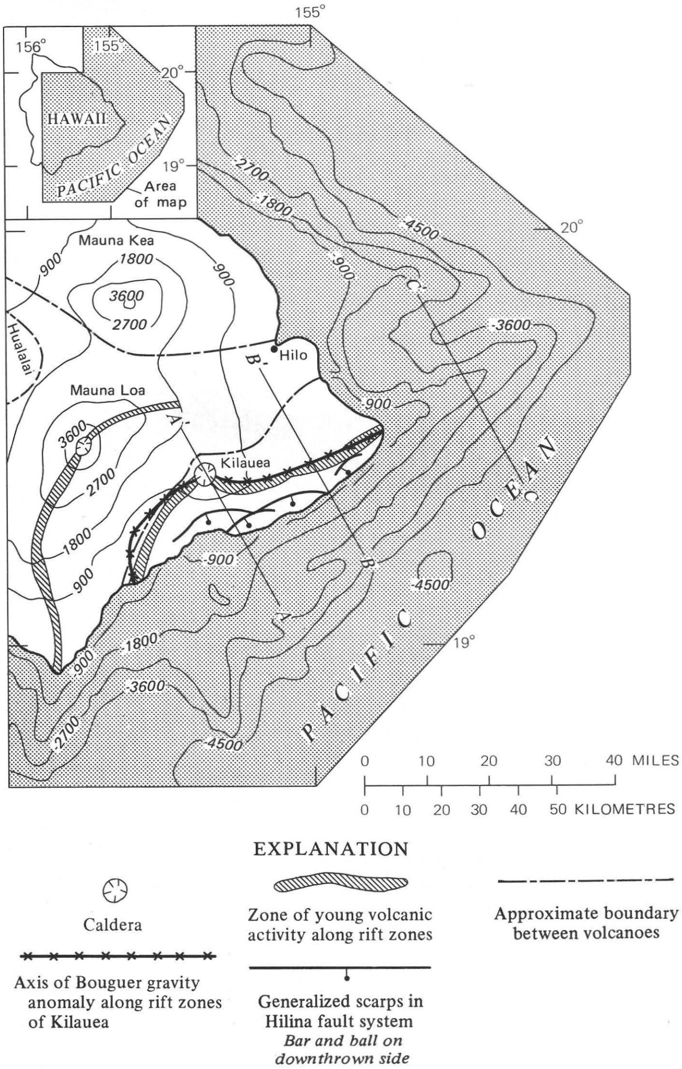


FIGURE 1.—Map of Kilauea Volcano and adjacent areas, showing generalized subaerial and submarine topography (from Moore and Fiske, 1969), zones of young volcanic activity along rift zones, axis of Bouguer gravity anomalies along rift zones (see also fig. 18), and generalized scarps in Hilina fault system. Cross sections *A–A′*, *B–B′*, and *C–C′* shown in figure 15. Land and bathymetric contour interval 900 m.

1968, and all rift eruptions in 1970–74), apparently because magma migrates outward from the central reservoir into a rift zone. During east rift eruptions, few earthquakes occur between the reservoir and a point near the eruption site, suggesting a virtually unobstructed interconnecting conduit. Many earthquakes, however, take place in the immediate vicinity of the eruption, as magma forcibly wedges its way toward the surface from the supply conduit or secondary holding reservoir. The site of rift eruption is further characterized by local uplift (table 1, 1919, 1955, 1960, 1961, 1962, March and December 1965, August and October 1968, and all rift eruptions in 1969–74), which forms a ridge commonly containing an axial graben.

The forceful intrusion of magma as dikes in the rift zones produces considerable dilation, which is manifested at the surface by gaping cracks and fissures whose "projections on one wall commonly fit into reentrants in the other wall* * *" (Macdonald, 1956, p. 278). The net long-term effect of such dike-induced dilation must be substantial, for hundreds or thousands of dikes are necessary to account for the gravity anomaly (Kinoshita and others, 1963) along the rift zones and the core of high-velocity material at depth (Hill, 1969). Not all of these forcible intrusive events are accompanied by eruptions. During this century, at least five inferred intrusive events with no eruption caused dilation of the rift zone, as evidenced by observed ground cracking (table 1, 1938, 1950, May and July 1963, May 1970).

We have studied the effects of rift dilation, principally through geodetic means, and find that the south flank of Kilauea has been uplifted and displaced southward as a result of forcible intrusion of magma as dikes. During this century, maximum net horizontal and vertical displacements of the mobile south flank have each been several metres, indicating that the volcano grows by ground deformation as well as by addition of lava flows, though to a lesser degree. Our results suggest that the south flank of the volcano has been displaced southward, away from the rest of the island, throughout the history of Kilauea, whereas the north flank has remained comparatively stable because of the gravitational and buttressing effects imposed by neighboring volcanoes, especially Mauna Loa. The results of this study help confirm the importance of the "edifice effect" (Fiske and Jackson, 1972) in determining the orientation of the stress field in clustered oceanic volcanoes.

ACKNOWLEDGMENTS

Most staff members of the Hawaiian Volcano Observatory took part in the 1970, 1971, and 1973 surveys; we thank them all, especially R. T. Okamura, M. K. Sako, J. B. Judd, and J. C. Forbes. Wilbur Griswold, helicopter pilot, skillfully ferried us into and out of very tight places during the 1970 trilateration. R. Y. Koyanagi contributed greatly through his continuing seismic investigation of Kilauea. B. K. Meade supplied data from the 1949 U.S. Coast and Geodetic Survey triangulation, and J. P. Church scoured the files for past U.S. Geological Survey leveling and triangulation data and gave much advice regarding their interpretation. T. Harano and A. C. Loo of the Hawaii Department of Transportation provided triangulation data for the 1955 eruption. R. W. Decker allowed the use of

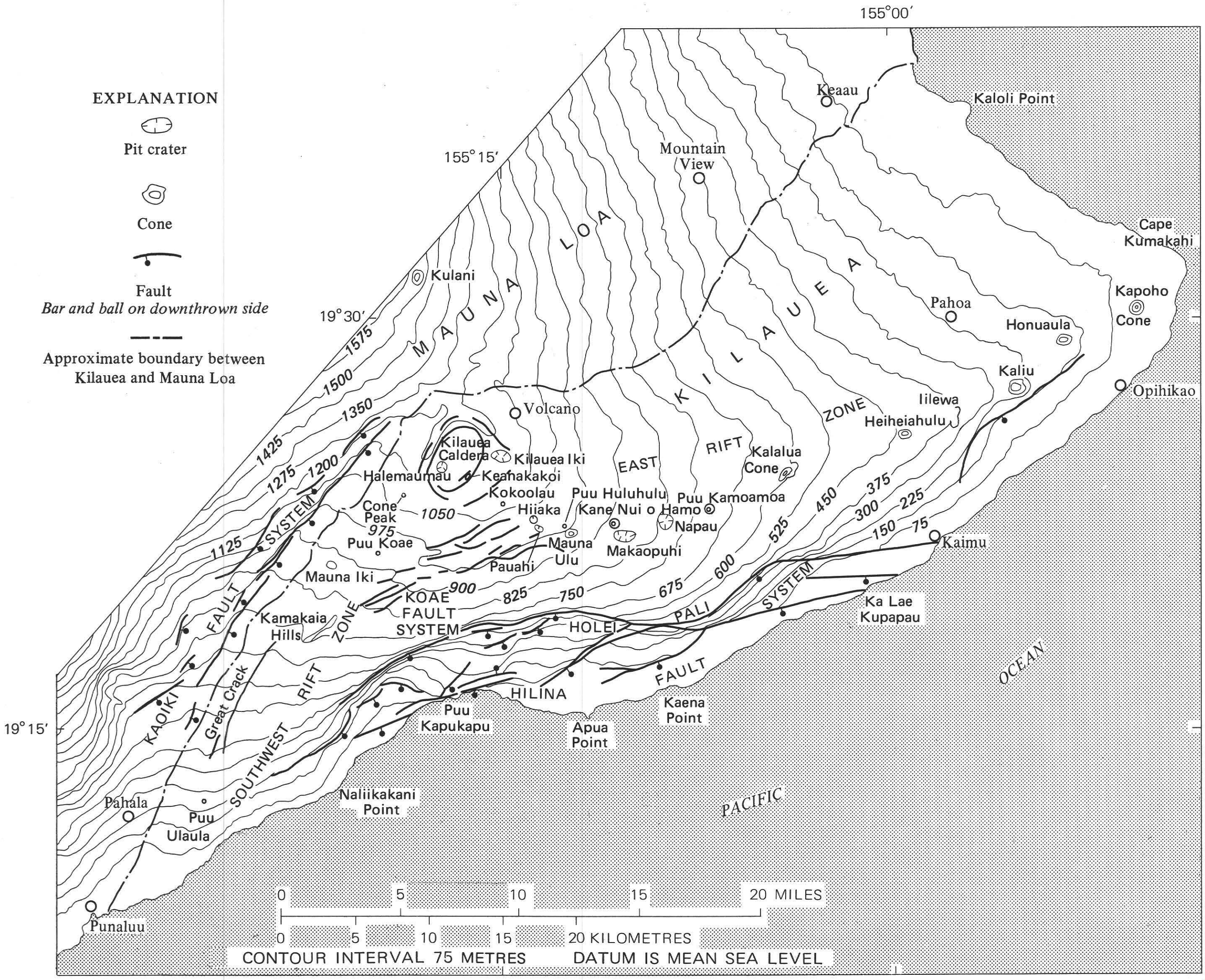


FIGURE 2.—Map of Kilauea and adjacent part of Mauna Loa, showing main structural features and locations of important faults, cones, and pit craters. Modified from Stearns and Macdonald (1946, pl. 1). Aloi Crater, mentioned in text, is buried beneath the southwest flank of Mauna Ulu; Alae Crater, beneath the southeast flank of Mauna Ulu.

unpublished leveling data, and W. T. Kinoshita did likewise with gravity data. R. W. Decker, J. P. Eaton, D. B. Jackson, W. T. Kinoshita, J. G. Moore, Kazuaki Nakamura, D. W. Peterson, T. L. Wright, and many others contributed stimulating ideas and willingly served as sounding boards and critics. R. W. Decker, E. T. Endo, R. C. Greene, E. D. Jackson, J. P. Lockwood, J. G. Moore, D. W. Peterson, and R. I. Tilling reviewed the manuscript at various stages, substantially improving it.

STRUCTURAL SUBDIVISIONS OF KILAUEA

For this paper, we can divide the subaerial part of Kilauea Volcano into six structural subdivisions (fig. 3) on the basis of the principal structures shown in figure 2:

1. The *summit area* is dominated by Kilauea Caldera and Halemaumau and Kilauea Iki Craters. The summit is the site of numerous eruptions and contains the center of surface deformation over the central magma reservoir.

2. The *east rift zone* is an elongate region as much as 5 km wide along the axis of the constructional ridge that forms the eastern part of the volcano. This zone is characterized by open cracks, pit craters, and small cones and shields and is the site of numerous eruptions. With increasing distance from the summit area, the subaerial part of the east rift zone may be loosely

TABLE 1.—*Eruptions and inferred intrusive events at Kilauea Volcano, 1900–1974*

[For volume and duration of eruptions, see Stearns and Macdonald (1946) and Wright (1971)]

Date[1]	Location of eruption	Deformation in summit area[2]	Deformation in areas outside summit area: Location	Deformation in areas outside summit area: Nature	References
1900–1907	Halemaumau, sporadically active.	None described	None described	None described	Macdonald and Abbott (1970).
1907–1924	Halemaumau, virtually continuously active.	Small reversible tilts of unknown significance.	do	do	Jaggar (1947).
December 1919	Southwest rift zone, creating Mauna Iki.	None described, but magma column in Halemaumau withdrew about 75 m.	From Halemaumau to Mauna Iki.	Ground cracks and uplift of at least 48 cm.	Jaggar (1947); Wilson (1935).
May 1922	Makaopuhi and Napau Craters.	None described, but magma column in Halemaumau withdrew at least 300 m.	Southwestward from Makaopuhi Crater for more than 1 km into Koae fault system.	Ground cracks	Bull. Hawaiian Volc. Obs., v. 10, p. 54, 1922.
August 1923	Upper east rift zone, between Alae and Makaopuhi Craters.	None described, but magma column in Halemaumau withdrew about 170 m.	Southwestward from the eruptive fissures for an unknown distance into Koae fault system.	do	Bull. Hawaiian Volc. Obs., v. 11, p. 72, 1923.
April–May 1924	Phreatic explosions in Halemaumau.	Subsidence of at least 4 m; inward-directed horizontal displacement possibly as much as 1.4 m.	(1) East rift zone near Cape Kumukahi, 40 km from Halemaumau. (2) Koae fault system, about 2 km southwest of Aloi Crater. (3) East rift zone, near Alae and Makaopuhi Craters.	(1) Ground cracks, normal faults, and graben; subsidence up to 3.6 m. (2) and (3) Ground cracks.	Bull Hawaiian Volc. Obs., v. 12, p. 18–19, 66, and 68, 1924; Wilson (1935).
July 1924 July 1927 February 1929 July 1929 November 1930 December 1931 September 1934	Halemaumau	None described	None described	None described	Stearns and Macdonald (1946).
May, June, August 1938	No eruption	Subsidence indicated by measurements of ground tilt; incomplete leveling data for September 1934 to January 1939 indicate subsidence of 42 cm, probably due largely to 1938 events.	Upper east rift zone and eastern Koae fault system.	Ground cracks. New steaming area formed just northwest of Kokoolau Crater.	Volcano Letter, no. 459, p. 2; no. 460, p. 3; no. 462, p. 1–4, 1938; no. 463, p. 4, 1939.
December 1944	No eruption	Abrupt summit subsidence, 3.6 seconds of arc on pendulum seismometer.	None described	None described	Macdonald (1954).
December 1950	No eruption	Subsidence of at least 30 cm calculated from ground tilt data.	Upper east rift zone and eastern Koae fault system.	Ground cracks	Volcano Letter, no. 510, p. 1–3, 1950.
June 1952	Halemaumau	Probably less than 5 sec of inward tilt.	None described	None described	Macdonald (1955).
May 1954	Halemaumau and floor of caldera within 1 km northeast of Halemaumau.	Ground cracks; probably horizontal displacements of 1 m or more accompanied by opening of new fissure.	do	do	Macdonald and Eaton (1957).
February–April 1955	East rift zone, from near Heiheiahulu to Kapoho Cone.	Subsidence of at least 42 cm calculated from ground tilt data.	Same area as eruption	Ground cracks, uplift, normal faults, graben; horizontal movements up to 1.6 m.	Macdonald and Eaton (1964).

November 1959	Kilauea Iki Crater	Inward tilt of 15 μrad, followed soon thereafter by outward tilt of 40 μrad; ground cracks.	None described	None described	Eaton (1962); Richter and others (1970).
January–February 1960	East rift zone, near Kapoho Cone.	Subsidence of about 1.5 m, calculated from ground tilt data.	(1) Same area as eruption. (2) Koae fault system, 7 km south of Halemaumau.	(1) Ground cracks, graben. (2) Ground cracks, normal faults.	Richter and others (1970; Eaton (1962); Hawaiian Volcano Observatory (unpub. data).
February, March, July 1961	Halemaumau	None detected	None described	None described	Richter and others (1964).
September 1961	East rift zone, Napau Crater to near Heiheiahulu Cone.	Subsidence, indicated by ground tilt data.	Same area as eruption	Ground cracks and normal faults, vertical displacements up to 6 m.	Do.
December 1962	East rift zone, Aloi Crater to Kane Nui o Hamo.	do	(1) Same area as eruption. (2) Koae fault system 1 km west of Aloi Crater.	Ground cracks, normal faults, uplift.	Moore and Krivoy (1964).
May 1963	No eruption	do	Central and western part of Koae fault system, 3 to 8 km south and southwest of Halemaumau.	Ground cracks and normal faults	Kinoshita (1967).
July 1963	do	do	Central and eastern part of Koae fault system, 4 to 5 km south and southeast of Halemaumau.	Ground cracks, normal faulting; local buckling.	Hawaiian Volcano Observatory, Summary 31, 1963.
August 1963	East rift zone, in and near Alae Crater.	do	Same area as eruption	Ground cracks	Peck, Wright, and Moore (1966).
October 1963	East rift zone, between Napau Crater and Kalalua Cone.	do	do	do	Moore and Koyanagi (1969).
March 1965	East rift zone, Makaopuhi Crater to near Kalalua Cone.	Subsidence of 25 to 30 cm	do	Ground cracks, graben	Wright, Kinoshita, and Peck (1968).
December 1965	East rift zone, Aloi Crater to Kane Nui o Hamo.	Subsidence, indicated by ground tilt data.	From Kane Nui o Hamo across Koae fault system for 20 km to Kamakaia Hills.	Ground cracks, normal faults, uplift, graben.	Fiske and Koyanagi (1968).
November 1967[3]	Halemaumau	Subsidence of 4.5 cm	None described	None described	Kinoshita and others (1969).
August 1968	East rift zone, Hiiaka Crater to near Kalalua Cone.	Subsidence of 19 cm; inward-directed horizontal displacement of 14 cm.	Same area as eruption	Ground cracks, uplift, graben.	Jackson and others (1975).
October 1968	East rift zone, Kane Nui o Hamo to Puu Kamoamoa.	Subsidence of more than 15 cm; large horizontal contraction.	do	do	Do.
February 1969	East rift zone, Aloi Crater to Puu Kamoamoa.	Subsidence of 12 cm; inward-directed horizontal displacement of 11 cm.	do	do	Swanson and others (1976).
May 1969[4]	East rift zone, from 1.6 km west of Aloi Crater to Alae Crater.	Small subsidence and contraction.	do	Ground cracks and uplift.	Hawaiian Volcano Observatory (unpub. data).
December 1969	East rift zone, Mauna Ulu, between Aloi and Alae Craters.	do	North and northeast of Alae Crater.	do	Do.
April 1970	East rift zone, in and west of Aloi Crater.	do	Same area as eruption	do	Do.
May 1970	No eruption	Uplift of 18 cm and outward-directed horizontal displacement of 29 cm in south part of caldera.	East rift zone adjacent to south part of caldera.	do	Duffield, Jackson, and Swanson (1976).
July 1970	East rift zone, north and northeast of Alae Crater.	Small subsidence and contraction.	Same area as eruption	do	Hawaiian Volcano Observatory (unpub. data).
January 1971	East rift zone, south of Puu Huluhulu.	Small subsidence	do	do	Do.

TABLE 1.—*Eruptions and inferred intrusive events at Kilauea Volcano, 1900–1974*—Continued

Date[1]	Location of eruption	Deformation in summit area[2]	Deformation in areas outside summit area		References
			Location	Nature	
August 1971	Southeast part of Kilauea Caldera, between Keanakakoi Crater and Halemaumau.	Uplift of 22 cm and outward directed horizontal displacement of 34 cm.	None described ----------	None described ----------	Hawaiian Volcano Observatory (unpub. data).
September 1971	Southwest part of Kilauea Caldera and southwest rift zone to 3 km southwest of Mauna Iki.	Uplift of 48 cm; outward directed horizontal displacement of more than 1 m.	Same area as eruption ----	Ground cracks and uplift--	Do.
<u>February 1972</u>[5]	East rift zone, Mauna Ulu--	Very slight subsidence ----	None detected ------------	None detected ------------	Do.
May 1973	East rift zone and adjacent Koae fault system, Hiiaka Crater to 1 km west of crater, and Pauahi Crater.	Inward tilting of more than 15 μrad.	Same area as eruption, but extended several kilometres farther west into Koae fault system.	Ground cracks and uplift--	Duffield (1975); Koyanagi, Unger, and Endo (1973).
November 1973	East rift zone, Pauahi Crater to 1 km downrift of Puu Huluhulu.	Subsidence of more than 4 cm; small horizontal contraction.	Same area as eruption ----	------do-----------------	Hawaiian Volcano Observatory (unpub. data).
July 1974	Southeast part of Kilauea Caldera, in and near Keanakakoi Crater.	Uplift of more than 15 cm; large horizontal expansion.	None described ----------	None described ----------	Do.
September 1974	Halemaumau and southwest part of Kilauea Caldera.	Outward tilting of more than 15 μrad just before eruption, inward tilting of 10 μrad during eruption; horizontal contraction of 16 cm.	------do-----------------	------do-----------------	Do.
December 1974	Southeast edge of upper southwest rift zone, northeast of Puu Koae.	Inward tilting of more than 100 μrad; horizontal contraction of as much as 80 cm.	Same area as eruption ----	Ground cracks and uplift--	Do.

[1]Date for commencement of eruption or intrusive event. Where several dates are listed, each indicates a period of eruption or deformation. Range of dates indicates time span during which most deformation occurred. Underlined date indicates that no inferred forcible intrusions accompanied the eruption; most of these eruptions took place in Halemaumau.

[2]Includes possible deformation, generally minor, shortly before and after main event.

[3]Eruption lasted until July 1968.

[4]Eruption lasted until October 1971.

[5]Eruption lasted until July 1974.

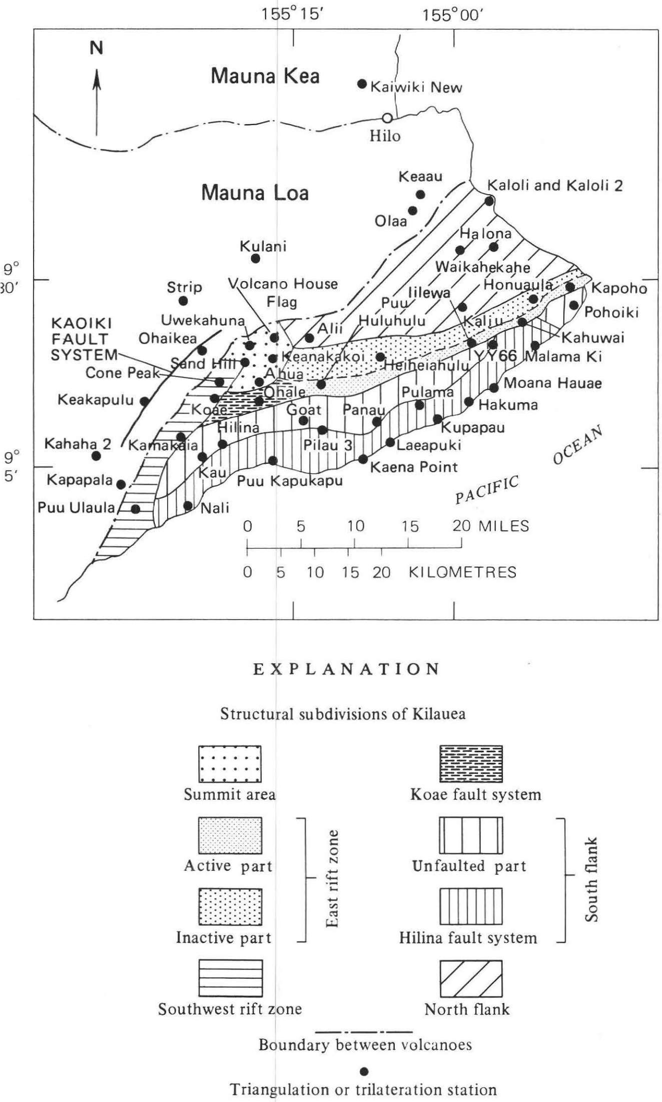


FIGURE 3.—Structural subdivisions of Kilauea Volcano and locations of triangulation stations referred to in this paper and established before 1970. Boundaries of subdivisions are gradational in most places. Compare with principal structures shown in figure 2.

subdivided into upper, middle, and lower parts. The upper part extends from the summit area to about Makaopuhi Crater, the middle part from Makaopuhi to Heiheiahulu, and the lower part from Heiheiahulu to the seacoast at Cape Kumukahi. The rift zone can also be divided into active and inactive parts on the basis of data presented in this paper. The east rift zone forms a prominent submarine ridge that extends far beyond the coastline (fig. 1).

3. The *southwest rift zone*, about 4 km wide, is characterized by structures similar to those of the east rift zone. It is the site of many eruptions, though fewer than the east rift zone. The southwest rift zone extends only a short distance beyond the shoreline (fig. 1; Moore and Fiske, 1969).

4. The *Koae fault system* is an east-northeast trending, 3-km-wide zone of open cracks and normal faults connecting the east and southwest rift zones.

5. The *south flank* is that part of the volcano seaward of and bounded on the north and west by the east and southwest rift zones and the Koae fault system. The northern part of the subaerial section of the south flank is characterized by unbroken, seaward-dipping lava flows. The southern part is disrupted by the Hilina fault system, a set of predominantly south-dipping normal faults.

6. The *north flank* is bounded by the summit area and the east rift zone. This flank is characterized by east- to northeast-dipping lava flows that abut and interfinger with flows from Mauna Loa.

The rift zones are believed to have a structure similar to that outlined by Fiske and Jackson (1972). Chiefly on the basis of seismic evidence, the dikes are considered to be shallow, bladelike bodies, largely if not entirely confined within the volcanic edifice and fed either directly from the summit reservoir system or from secondary high-level reservoirs in the rift zone itself (Swanson and others, 1976). The dips of the dikes are believed to be steep to vertical on the basis of (1) observations in older, deeply eroded Hawaiian shields such as Niihau (Stearns, 1947) and Koolau (Stearns and Vaksvik, 1935), (2) seismic evidence, which shows that earthquakes and tremor directly associated with the intrusion of magma occur in a narrow zone directly below the point where lava reaches the surface, and (3) interpretations of ground deformation data related to the intrusion of magma on the upper east rift zone in August 1968 (Jackson and others, 1975). Contrary interpretations regarding the dip are examined later.

The flanks of the volcano are assumed to consist of lava flows above sea level and pillow lavas below sea level, with a layer of hyaloclastic debris between; data for this interpretation are given by Moore and Fiske (1969).

EARTHQUAKE DISTRIBUTION

The distribution and timing of crustal earthquakes provide probably the most unequivocal evidence for instability of the south flank related to magmatic events in the rift zones. Koyanagi, Swanson, and Endo (1972) summarized this seismic evidence, some of which is shown in figure 4. Numerous earthquakes take place beneath the summit area and are presumably related to stress release in and near the main reservoir system. Almost all other earthquakes occur at shallow or intermediate depths beneath the south flank or its boundary zones, the east and southwest rift zones and Koae fault system. Thousands of earthquakes ranging up to

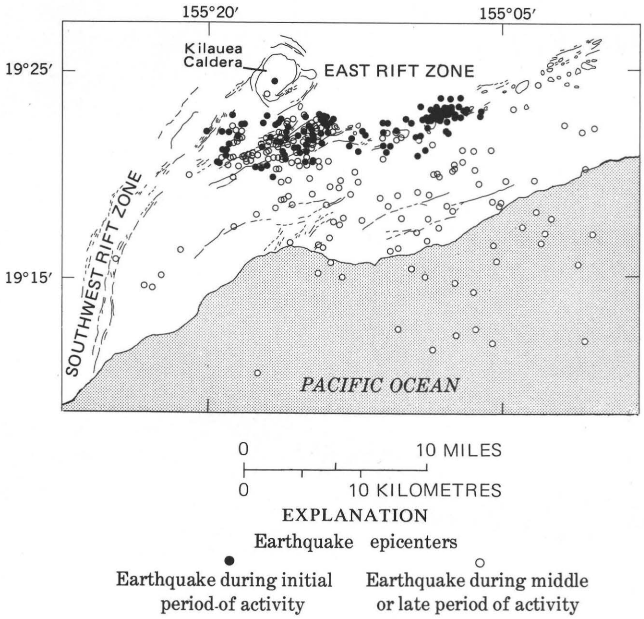


FIGURE 4.—Location of earthquake epicenters from brief seismic swarms related to eruptions and ground-cracking events between 1963 and 1969. Earthquakes plotted are of $M \geqslant 2$ and occurred at depths less than 15 km. Note that earthquake epicenters migrate south-southeastward across south flank following initial seismicity along east rift zone and Koae fault system and that earthquakes are virtually absent from north flank, a distribution also typical of periods between eruptions (Koyanagi and others, 1972, fig. 1). Slightly modified from Koyanagi, Swanson, and Endo (1972, fig. 6).

magnitude (M) 5.0 take place each year in these areas away from the summit, furnishing clear evidence of instability within the volcanic pile. Study of the focal mechanisms of south-flank earthquakes shows that most maximum stress axes are oriented southeast and are horizontal or plunge gently seaward (Koyanagi and others, 1972; Endo, 1971; E. T. Endo, oral commun., 1974).

In contrast, the north flank of Kilauea has few earthquakes, and most of those that do occur emanate from upper mantle depths well beneath the base of Kilauea. Thus, an immediate conclusion is that the north flank is relatively stable seismically compared with the south flank.

The high seismicity of the rift zones is not surprising, for these are areas where intrusion and eruption take place. The reason for the high seismicity of the south flank, on the other hand, is not so readily apparent, but the timing of earthquake activity relative to rift eruptions and intrusive events appears pertinent to an explanation. Figure 4 shows that earthquakes related to magmatic events occur on the south flank only during the middle and late stage of seismic activity, after the intrusive event has largely or completely ended, whereas earlier earthquakes cluster along the east rift zone and adjacent parts of the Koae fault system, near where intrusion is taking place. This sequential relation, discussed in more detail by Koyanagi, Swanson, and Endo (1972), indicates that intrusion is a precursor to major south-flank seismicity, which in turn suggests a cause-effect relation between them. The inference is that stress is built up in the south flank by intrusion along its northern edge, then relieved over a period of a few days during the late-stage earthquake activity.

One effect of forceful intrusion of magma is lateral displacement of wallrock on either side of the dike. This may provide a mechanism to account for the south-flank earthquakes. If the observed south-flank seismicity is caused by such displacement away from the rift zone, it should be possible to detect this displacement by geodetic means. For this reason we undertook a geodetic study of Kilauea, results of which demonstrate that large ground displacements on the south flank and elsewhere have indeed taken place in a remarkably systematic fashion.

The following presentation of the horizontal and vertical deformation data is long and detailed as we attempt to document our case. Many readers may wish to examine figures 5–14 and then proceed directly to the three summary sections on long-term horizontal deformation, south-flank horizontal deformation, and vertical deformation on the flanks. The principal interpretive part of the paper begins with the section, "A Structural Model for the South and North Flanks of Kilauea."

LONG-TERM HORIZONTAL DISPLACEMENTS DURING THE 20th CENTURY

EXISTING DATA AND THEIR QUALITY

Horizontal ground displacements over several periods a decade or more long have been derived from six regional horizontal control surveys. Four of the surveys are triangulations, in 1896 (1914)[1] and 1949 by personnel of the U.S. Coast and Geodetic Survey, and in 1958 and 1961 by personnel of the U.S. Geological Survey. The other two surveys are trilaterations conducted in 1970 and 1971 with a model 8 laser-beam geodimeter by personnel of the Hawaiian Volcano Observatory; the general procedures and measured distances for the 1970 survey, the trilateration most used in this paper, are given by Swanson and Okamura (1975). The regional triangulation surveys by Wilson (1935) and Wingate (1933) are not used in this paper because neither was tied to a baseline that we consider

[1]Most of the 1896 (1914) survey took place in 1896, but stations on the south flank were not triangulated until 1914. See text discussion of figure 5*D* for details.

to be stable. Figure 3 shows the location of all triangulation stations occupied at least once prior to reoccupation in 1970.

Each of these surveys took place over a period of weeks to months, and the 1896 (1914) survey was split into two intervals separated by 18 years. As Kilauea sometimes deforms greatly over much shorter periods of time, the effect of possible contemporaneous ground deformation on the quality of each survey must be considered. This is done in the section "Supplemental Information," with the conclusion that the problem of contemporaneous ground deformation is minimal, with the possible exception of the summit area in 1961.

The general quality of the four triangulation surveys is given in table 2. The 1949 and 1958 surveys are adequate to define reliable long-term displacements of the magnitudes shown by Kilauea. The 1896 (1914) and 1961 data are poor, owing to large closure errors and, in 1961, very poor network geometry on the south flank. The precision of the 1970 trilateration is on the order of 5–8 mm/km, far superior to precisions of all the triangulations.

The problems of imprecision and the generally poor geometry of the survey networks make the triangulation data unrealiable for strictly quantitative studies. Nonetheless, most long-term ground displacements have been large enough to overshadow these problems, so that the triangulation data can be compared with the trilateration data to derive general patterns and semiquantitative amounts of displacements. The selection of stable baselines and the procedure used in determining ground displacements are outlined in the section "Supplemental Information."

RESULTS

Horizontal displacements derived from comparison of the four triangulation surveys with the 1970 trilateration survey are shown in figure 5. The net displacements for most stations, especially those on the south flank, are large and systematic for the survey periods, which range from 9 years to 74 years in duration. In this section we correlate these displacements with magmatic (eruptive or intrusive) events that occurred during the survey intervals and find that the displacements on the south flank are systematically related to the opening of new ground cracks in the east and southwest rift zones and in the Koae fault system. However, the exact timing between specific magmatic and displacement events is unknown because the survey intervals are long and may include several events; later we examine shorter survey intervals in which specific magmatic and displacement events can be correlated. In this section, the results are presented in the order of decreasing reliability instead of chronologically.

TABLE 2.—*Precision of the triangulation surveys*

Triangulation	Closure of triangles used (sec) Average	Closure of triangles used (sec) Maximum	Order[1]	Probable error in length less than[1]
1896 (1914)	3.9	12	3	1 in 5,000 (20 cm/km)
1949	1.19	3.09	2	1 in 20,000 (5 cm/km)
1958 rift	4.0 } 3.6	13.6	3 } 3 (good)	1 in 10,000 (10 cm/km)
summit	2.3 }	5.6	2 }	
1961	4.8	9.6	3	1 in 5,000 (20 cm/km)

[1]Taken from Gossett (1959, p. xiv–xv). Results of the present study suggest that the 1896 (1914) and 1961 surveys have probable errors of 1 in 10,000 (10 cm/km) or less.

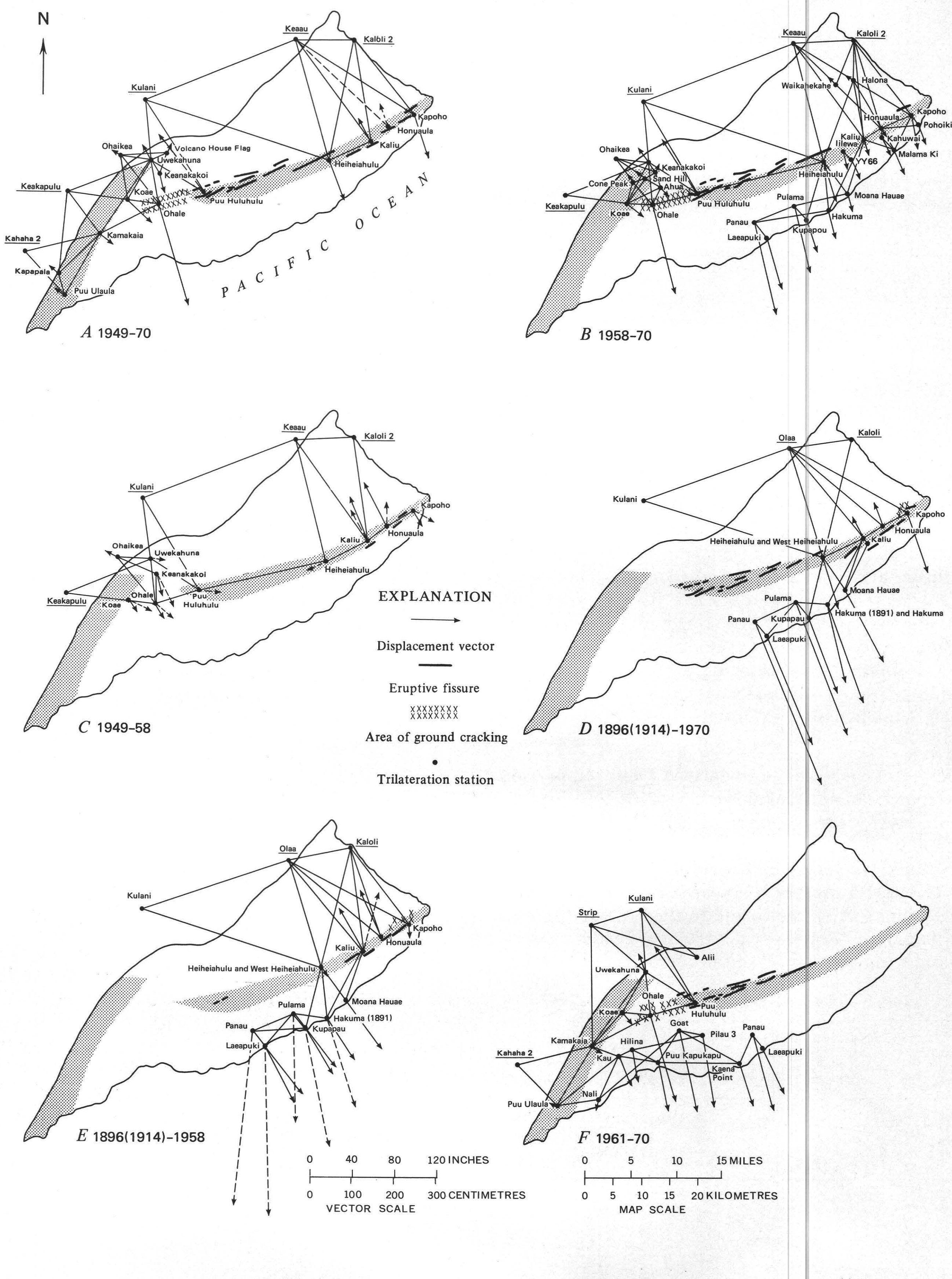
1949–70

Displacement vectors on the lower east rift zone generally point perpendicularly away from nearby fissures that opened in 1955, 1960, and 1961 and are of magnitudes close to the probable aggregate amount of opening of these fissures, judging by observations of the fissures soon after they formed (fig. 5A). Kaliu and Honuaula are both located within the rift zone, less than 400 m northwest of some of the 1955 fissures (Macdonald and Eaton, 1964). Kapoho is southeast of the easternmost 1955 vent and the main 1960 vents (Richter and others, 1970). Heiheiahulu is southeast of the nearest 1961 fissure (Richter and others, 1964).

The radial pattern of displacement defined by the Volcano House Flag, Keanakakoi, Uwekahuna, and Ohaikea vectors indicates tumescence of the summit region centered northwest of Keanakakoi. Such tumescence is reasonable in view of the similar location of swelling centers documented for many later periods of inflation (for example, Jackson and others, 1975). This radial pattern is defined by comparatively small displacements, 30 cm or less, and would probably have been distorted or even completely obscured if the baselines had moved. For this reason we have confidence in the stability of the baselines between 1949 and 1970. The displacement of Keanakakoi is larger than at the other summit stations, perhaps partly the result of the opening in 1954 of a fissure oriented N.75°E. about 1.5 km north of the station (Macdonald and Eaton, 1957).

Puu Huluhulu shows a large northwest displacement of about 1.7 m that is not perpendicular to the trend of nearby eruptive fissures. This displacement is possibly the net result of major north-northwest movement directly away from the rift zone during the nearby eruptions of 1962, 1963, 1965, and 1968–70 combined with lesser westward movement away from an area of uplift centered over a magma reservoir near Makaopuhi Crater (Swanson and others, 1976).

Ohale, in the Koae fault system, was displaced about 2.6 m south-southeastward, perpendicular to cracks that formed in the area during major faulting events in


N
Keaau
Kaloli 2
Kulani
Kapoho
Honuaula
Kaliu
Ohaikea
Volcano House Flag
Uwekahuna
Keanakakoi
Heiheiahulu
Keakapulu
Koae
Puu Huluhulu
Ohale
Kahaha 2
Kamakaia
Kapapala
Puu Ulaula
PACIFIC OCEAN
A 1949-70
Waikahekahe
Halona
Pohoiki
Iilewa
Kahuwai
YY66
Malama Ki
Cone Peak
Sand Hill
Ahua
Pulama
Moana Hauae
Panau
Hakuma
Laeapuki
Kupapou
B 1958-70
C 1949-58
EXPLANATION
Displacement vector
Eruptive fissure
Area of ground cracking
Trilateration station
Olaa
Kaloli
Heiheiahulu and West Heiheiahulu
Kupapau
Hakuma (1891) and Hakuma
D 1896(1914)-1970
Hakuma (1891)
E 1896(1914)-1958
Strip
Alii
Goat
Pilau 3
Hilina
Kau
Puu Kapukapu
Kaena Point
Nali
F 1961-70
0 40 80 120 INCHES
0 100 200 300 CENTIMETRES
VECTOR SCALE
0 5 10 15 MILES
0 5 10 15 20 KILOMETRES
MAP SCALE

1950 (Finch, 1950), 1963 (Kinoshita, 1967), and December 1965 (Fiske and Koyanagi, 1968), and several minor events as recently as May 1970 (Duffield and others, 1976).

The pattern of displacement directly away from the southwest rift zone suggests rift dilation, for Koae and Kamakaia are southeast of the axis of the rift zone and Puu Ulaula and Kapapala northwest. The several episodes of faulting in the nearby Koae fault system could also have contributed to the displacement of Koae and Kamakaia.

1958–70

All stations within the lower east rift zone were displaced southeastward, directly away from the rift zone, in a pattern consistent with rift dilation during the 1960 and 1961 eruptions (fig. 5*B*). The 1960 vents were near Kapoho, and the eastern 1961 vents near Heiheiahulu. The four stations between Kapoho and Heiheiahulu are not located near vents active since 1958, but their displacements may record the effects of intrusion accompanying the two eruptions.

Stations on the south flank (Pohoiki, Malama Ki, YY66, and all stations southwest of YY66) were displaced by large amounts directly away from the east rift zone. These displacements are largest (2.3 m at Panau) at stations opposite the most active part of the rift zone between 1958 and 1970, even though the nearest vent areas are more than 5 km away. The displacements generally decrease away from the rift zone; for example, compare Panau with Laeapuki and Pulama with Kupapau.

FIGURE 5.—Horizontal ground displacements (solid vectors) at Kilauea Volcano derived from comparison of 1896 (1914), 1949, 1958, and 1961 triangulation surveys with 1970 trilateration survey. Procedures used outlined in the section "Supplemental Information." Base stations considered stable are underlined. Southwest rift zone and active part of east rift zone (fig. 3) are shown by shading. Locations of rift eruptions and major episodes of ground cracking that occurred during a given survey interval are indicated. Vectors in *C* were obtained by subtraction of vectors in *B* from *A*; vectors in *E* were obtained by subtraction of vectors in *B* from *D*. Dashed vectors in *C* and *E* were computed by Lloyd (1964) using standard least-squares analysis with no geologic constraints. Dotted vector for Honuaula in *A* indicates station was intersected but not occupied during 1949 survey. Alii in *F* shows no displacement greater than survey error. See text for discussion of the vectors at Panau and Laeapuki in *F*. Parts of network common to both surveys are indicated for each survey interval; further information can be obtained from the U.S. Geological Survey, the National Geodetic Survey, or the senior author.

Waikahekahe and Halona, located a comparable distance from the axis of the rift zone as Laeapuki and Kupapau but on the north flank, registered little if any displacement.

The radial pattern of displacements in the summit area during this period indicates expansion around a center 1–2 km southwest of Keanakakoi. This center, south of that for the 1949–70 interval (fig. 5*A*) corresponds to an area of uplift defined by leveling between 1958 and 1971 (Okamura and Swanson, 1975). The fact that these small horizontal displacements define the center of deformation quite precisely is evidence that the assumption of baseline stability is justified.

Puu Huluhulu shows about the same displacement for the 1958–70 interval as the 1949–70 interval; this indicates that most displacement took place after 1958, consistent with the absence of upper east rift eruptions between 1949 and 1958. The displacement at Ohale during the 1958–70 interval has about the same azimuth as between 1949 and 1970 but is somewhat smaller, probably because it does not reflect the 1950 ground-cracking event in the surrounding Koae fault system. Koae moved away from the southwest rift zone, and its magnitude of displacement suggests that about half of its 1949–70 displacement took place after 1958.

1949–58

Vector subtraction of the 1958–70 displacements from the 1949–70 displacements yields the 1949–58 values. Kaliu, Honuaula, and Kapoho, all within the rift zone, clearly responded to dilation accompanying the 1955 eruption (fig. 5*C*). Heiheiahulu, Puu Huluhulu, and Uwekahuna did not move measurably, Ohale reflects the 1950 ground-cracking episode near it, and Keanakakoi was displaced away from the 1954 vent fissure nearby in the caldera.

The dashed vectors in figure 5*C* are displacements calculated by Lloyd (1964, fig. 3*A*) using data from the 1949 and 1958 triangulation surveys adjusted by least-squares methods with no geologic constraints. The pattern of these displacements is rather similar to the one we derived, except for Puu Huluhulu and Heiheiahulu, where Lloyd's data show large displacements along azimuths nearly parallel to the east rift zone, and ours show none. This discrepancy clearly relates to the methods used in the derivation of the displacements. We believe it important that there is no geologic evidence for deformation that would give rise to displacements at Puu Huluhulu and Heiheiahulu of the azimuths and magnitudes indicated by the least-squares adjustment.

1896(1914)–1970

In 1896, stations Kulani, Olaa, Kaloli, West Heiheiahulu and Heiheiahulu (100 m apart), Kaliu, Honuaula, Kapoho, Moana Hauae, and Hakuma (1891) were triangulated, and our derivation of their displacements up to 1970 is shown in figure *5D*. In 1914, Heiheiahulu-Kaliu was used as the base for triangulating Hakuma (150 m south of Hakuma (1891)) and all stations farther west. Assuming that this base was stable between 1896 and 1914, we derived the shortest vector shown in figure *5D* at Hakuma for the period 1914–70. The Hakuma and Hakuma (1891) vectors are so similar in magnitude and azimuth, with the vector for the longer period somewhat larger, that the assumption of stability of the Heiheiahulu-Kaliu baseline for the period 1896–1914 seems justified. Furthermore, this period was one of almost continuous weak eruption in the famous Halemaumau lava lake at the summit of Kilauea (table 1) and on the basis of recent work at Kilauea, it is unlikely that significant deformation took place along the rift zone during such summit behavior, especially as far as 30–50 km downrift. We therefore assumed baseline stability and derived the 1914–70 displacements for Kupapau, Pulama, Laeapuki, and Panau.

The general relations between displacements and eruptive fissures were the same for the 1896(1914)–1970 period as for the previously discussed examples. Heiheiahulu, Kaliu, Honuaula, and Kapoho were all displaced at right angles away from nearby vent fissures and ground cracks. Moana Hauae and other south flank stations farther west were also displaced away from the nearest active parts of the east rift zone, and the seaward stations (Laeapuki and Kupapau) were displaced less than landward stations (Panau and Pulama) (compare fig. *5D* with *5B*). The largest displacement, nearly 4.4 m at Panau, was opposite the most active part of the rift zone.

1896 (1914)–1958

Displacements for this period (fig. *5E*) were derived by subtracting vectors of figure *5B* from corresponding vectors of figure *5D*. Lloyd (1964, fig. *3A*) also computed vectors (dashed in fig. *5E*) for this period, again by a least-squares analysis. The displacements derived by Lloyd and by us are rather similar, although ours are smaller and more systematic, lacking the clockwise swing suggestive of adjustment or survey error. Both sets of displacements show that the south-flank stations moved seaward by large amounts, most likely along azimuths nearly perpendicular to the trend of the east rift zone. Comparison of figures *5E* and *5D* shows that only about half of the displacement of the south flank from 1914 to 1970 took place in the 44 years before 1958. Thus the rate of displacement has been almost four times greater in the last 12 years, a time during which eruptions along the middle and upper east rift zone have been far more frequent than during the earlier period (table 1).

Displacements of Heiheiahulu, Kaliu, and Honuaula between 1896, 1958, and 1970 are reasonable in light of the eruptions in these areas. However, the displacement of Kapoho between 1896 and 1958 is unexpectedly small, considering the major episode of ground cracking and graben subsidence that took place just north of Kapoho in April 1924 (Finch, 1925; Jaggar, 1924). In fact, the overall displacement at Kapoho between 1896 and 1970 can be largely accounted for by the 1960 eruption (compare figs. *5B* and *5D*). We cannot eliminate the possibility that significant displacement during the 1924 event has gone undetected owing to unrecognized survey errors. However, the mode of ground deformation in 1924 (maximum subsidence along the north edge of the graben and northward tilting of the graben floor: Finch, 1925; Jaggar, 1924) suggests the development of an asymmetric graben similar to that which Cloos (1968, fig. 18) modeled in clay. In this model, the north side of the graben would have moved northward, and little if any ground displacement south of the graben would be expected.

1961–70

Three major episodes of faulting (May and July 1963 and December 1965) in the Koae fault system (fig. 2) and thirteen upper and middle east rift eruptions (December 1962, August and October 1963, March and December 1965, August and October 1968, February 1969, and five events associated with the Mauna Ulu eruption in 1969 and 1970) were the dominant structural events of this interval. Ohale moved southeastward more than 2 m, and Puu Huluhulu moved northwestward more than 1.5 m, perpendicular to new ground cracks and vent fissures (fig. *5F*). Dilation of the southwest rift zone is suggested by displacements at Koae, Kamakaia, and possibly Puu Ulaula. South flank stations within, and inland of, the Hilina fault system (fig. 2) moved seaward. The displacement at Kaena Point is similar to that at nearby Laeapuki for the 1958–70 period (fig. *5B*), not surprising since the two stations are equidistant from the east rift zone, only 4 km apart, and are not separated by intervening ground cracks of faults.

Laeapuki itself was not triangulated in 1961, as shown in figure *5F*, but Panau was triangulated from Kaena Point and Laeapuki. In order to estimate the displacement of Panau between 1961 and 1970, it is necessary to assume a displacement vector at

Laeapuki. If we assume that the displacement at Laeapuki for 1961–70 is the same as that measured at Kaena Point for the same period, the displacement derived for Panau compares closely with that for the 1958–70 period (compare figs. 5*F* and *B*). This close agreement seems reasonable and constitutes evidence that the 1961–70 displacements are indeed reliable, despite the poor quality of the triangulation (table 2) and the weak network geometry.

Just as for earlier periods, south flank stations opposite the most active parts of the east rift zone and Koae fault system were displaced more than other stations (for example, compare displacements west of Goat with those east). The north flank apparently remained stable, as the displacement at Alii is well within probable survey error.

AUGUST–SEPTEMBER 1970 TO OCTOBER 1971

This period is bracketed by two trilateration surveys conducted on the western half of Kilauea. A more closely spaced survey network was used than during the previous periods discussed. Major eruptions and related ground cracking occurred in August 1971 in Kilauea Caldera and September 1971 in the caldera and along the southwest rift zone. In addition, a small eruption took place in late January and early February 1971 from new cracks south of Puu Huluhulu (fig. 6) on the east rift zone; this episode was related to the contemporaneous eruption at nearby Mauna Ulu (Swanson and others, 1971).

The effects of the August and September eruptions dominate the displacement map. HVO 34, about 1.5 km west-southwest of HVO 10, shows the largest displacement, more than 1.25 m, which is best interpreted as the net result of deformation attending the two eruptions combined with lesser inflation of the summit area. Most stations in the summit area show some effect of this inflation, which took place largely before the August eruption. Displacements decrease abruptly down the southwest rift zone beyond the eruption site.

The south flank was markedly displaced during this period, although far less so than areas closer to the eruption sites. Displacements are directed away from the general trend of the east rift zone and Koae fault system and are oblique to the southwest rift zone. The magnitude of displacement is greatest in the sector directly south-southeast of the area in which the three eruptions took place and decreases both to the east and west. In contrast to previous periods, the eastern part of the Hilina fault system dilated slightly as indicated by direct measurement of extension between Laeapuki and Panau, and Kupapau and Pulama; this extension is not shown clearly in figure 6 because of the scale. The western part of the Hilina system was displaced

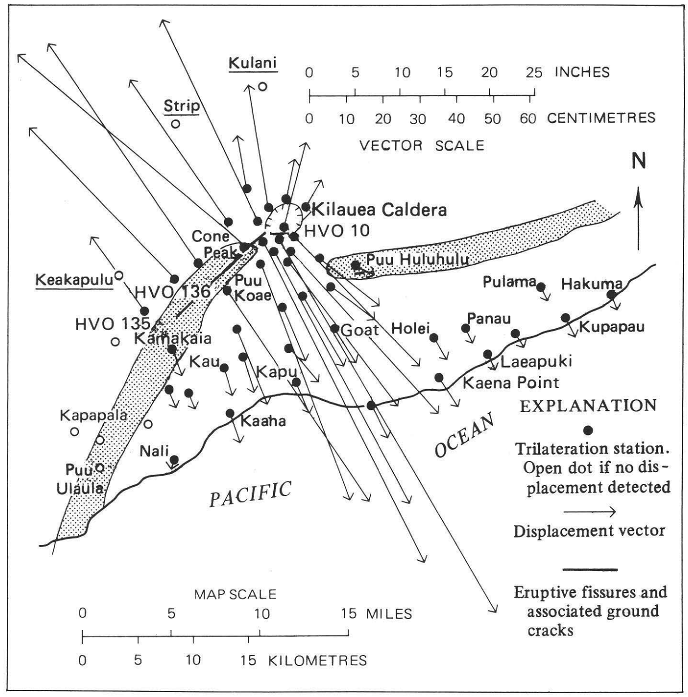


FIGURE 6.—Horizontal ground displacements between August 24–September 23, 1970 and October 3–18, 1971, derived from trilateration surveys. Holei was first occupied on November 18, 1970. Stations held fixed (Kulani, Strip, and Keakapulu) are underlined. Eruptive fissures and associated ground cracks that opened during survey interval are shown. Southwest rift zone and active part of east rift zone (fig. 3) are indicated by pattern. Stations west of Kamakaia and Nali show no displacement greater than survey error. Diagram of network showing lines and stations occupied is given by Swanson and Okamura (1975, fig. 2).

seaward as a unit, with little or no buildup of horizontal strain.

SUMMARY OF LONG-TERM HORIZONTAL DEFORMATION

Long-term horizontal displacements, as measured over periods of years to decades, are of considerable magnitude at Kilauea. The summit and southwest rift zone showed small net expansion for the periods 1949–70 and 1958–70 and considerable dilation related to the 1971 eruptions. Throughout this century, stations within the east rift zone and Koae fault system were displaced along azimuths nearly perpendicular to the rift zone and fault system; stations southeast of active fissures were displaced south-southeastward, those northwest of active fissures, north-northwestward. The entire south flank moved southeastward in a direction almost exactly perpendicular to the east rift zone and Koae fault system; this direction is more nearly perpendicular to the trend of the rift zone as a whole than to individual fissures within the rift zone. The direction of displacement is similar to that of maximum stress axes found from the study of the focal mechanism of south-flank earthquakes. The displacements on the south flank are large, as much as 4.4 m at Panau between 1914 and 1970 and 2.3 m at Panau between 1958 and 1970. Displacement of the south flank is greatest south-southeast of the most active part of the rift zone for a given survey interval, and observations show that the magnitude of this displacement approximately accounts for the probable total width of new fissures and ground cracks in the rift zone. For at least the 1960 eruption, north-flank stations about 10 km from the vents did not definitely move, whereas south-flank stations as far as 10 km from active vents moved tens of centimetres southward. This is consistent with the seismic evidence indicating stability of the north flank relative to the south flank. In contrast to north-flank stations, those close to, but west of, the southwest rift zone were displaced toward Mauna Loa during the 1970–71 period.

Most of this evidence implies a direct causal relation between magmatic events and ground displacement. However, the periods of measurement between these regional surveys were so long that they contain more than one magmatic event; consequently the measured displacements are the net effect of several episodes of deformation. These displacements alone are therefore not adequate to correlate a specific eruptive or intrusive event with a specific episode of ground deformation, and we must turn to shorter survey periods to demonstrate such a correlation.

HORIZONTAL DEFORMATION DURING SPECIFIC INTRUSIVE OR EXTRUSIVE EVENTS

In this section, we present examples of short-term deformation events keyed to specific eruptions or episodes of ground cracking associated with intrusions. It is well known that filling and emptying of the magma reservoir system beneath the summit of Kilauea is generally accompanied by largely reversible ground deformation (Wilson, 1935; Eaton, 1962; Decker and others, 1966; Fiske and Kinoshita, 1969). We focus attention here on other types of events that apparently reflect the forceful intrusion of magma as dikes into the rift zones.

Most of our examples are controlled by geodimeter data of excellent quality, indicated by both repeated measurements and small closure errors. The triangulation data for the 1955 eruption are poor (third order), but the pattern of displacement is compelling.

FEBRUARY–MAY 1955 EAST RIFT ERUPTION

The early part of the February 28–May 26, 1955 eruption took place in an area where a new highway was under construction. Fortuitously, triangulation surveys were carried out just before and just after the eruption; all stations in this local network are located along the east rift zone. The resurvey indicated that substantial ground movement took place during the eruption.

The ground displacements shown in figure 7 were derived using two different assumptions. The preferred assumption is that all displacement of Honuaula, Kapoho, and Kaliu between 1949 and 1958 (fig. 5C) took place during the 1955 eruption. The other assumption, used by Macdonald and Eaton (1964) in computing displacement, is that Kapoho did not move during the eruption. Both interpretations indicate northwest displacement of stations north of the fissure zone, and displacement vectors based on the preferred assumption indicate southeast movement of stations south of fissures. Macdonald and Eaton (1964, p. 107) noted that the pattern of displacements is similar to that produced by "a simple pulling open of the fissure zone," with no evidence of strike-slip movement. Stations opposite the central part of this zone, where eruptive activity was concentrated, moved more than those at either end.

The displacement at Honuaula is less than that at

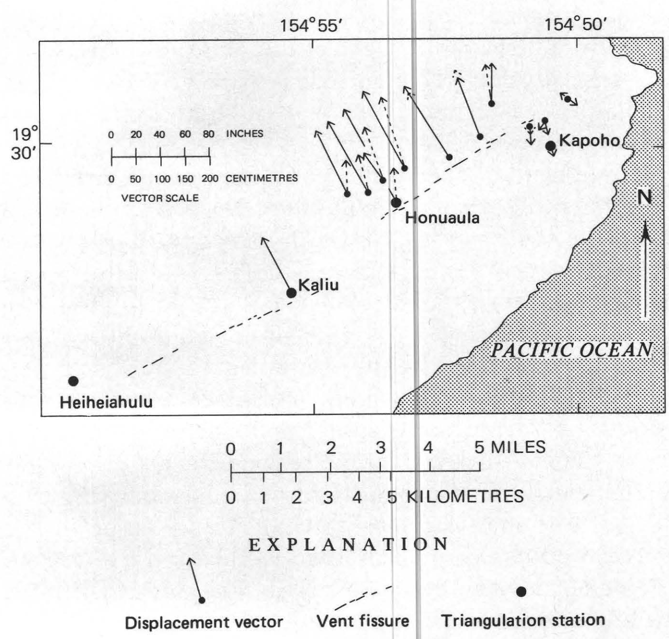


FIGURE 7.—Horizontal ground displacements during February 28–May 26, 1955 east rift eruption. Solid vectors were derived on preferred assumption that all displacement of Honuaula, Kapoho and Kaliu between 1949 and 1958 (fig. 5C) took place during 1955 eruption. Dashed vectors were computed by Macdonald and Eaton (1964) on assumption that Kapoho did not move during eruption. Data for triangulation stations north of Honuaula-Kapoho baseline were obtained in late 1954 (before eruption) and May–June 1955 (after eruption) by the Territory of Hawaii Highway Department.

stations farther northwest. This may result from the combined effect of horizontal dilation and uplift along the fissure zone (Macdonald and Eaton, 1964, p. 105–107), for such uplift should in theory cause the maximum horizontal displacement to be located some distance away from the crest of the uplift (Mogi, 1958; Dieterich and Decker, 1975). Alternatively, the intrusion of northwest-dipping dikes could have caused such a displacement pattern, although first-motion studies of earthquakes during the eruption suggest normal faults dipping 70° to 90° southeast (Macdonald and Eaton, 1964, p. 122).

FEBRUARY–MAY 1969 EAST RIFT ERUPTIONS

Eruptions on February 22–28 and May 24–29, 1969 on the upper east rift zone took place within the newly established geodimeter network (fig. 8*A*). Surveys over the February eruption were conducted on February 11–13 and 24–27; most ground deformation occurred between February 22 and 24 (Swanson and others, 1976). Surveys over the May eruption were conducted on April 21–23 and June 2–4. The two bench marks closest to the eruption sites, Puu Huluhulu (within the rift zone) and HVO 117 (along the northern margin of the south flank), moved away from the vent fissures and associated ground cracks. The displacement vectors are nearly perpendicular to the new fissures and reflect the divergence in trend of the fissures for each eruption.

SEPTEMBER 1971 ERUPTION

An eruption took place on September 24–29, 1971 from a line of fissures between Kilauea Caldera and a point along the southwest rift zone nearly 12 km away (fig. 8*B*). The eruption site lies within a trilateration network that was occupied August 16–19 and October 4–7. Most of the stations are either within the southwest rift zone or near the caldera. Horizontal displacements were directed away from the new fissures and ground cracks and were of roughly equal magnitude on both sides of the eruptive zone. The average extension across this zone, indicated by the displacement vectors, was about 1.4 m, virtually the same as the total amount of opening on new cracks as determined by direct measurement. Stations HVO 111 and HVO 109, on the north edge of the south flank of Kilauea about 6 km from the eruption zone, were displaced an order of magnitude less than stations nearer the zone. This dramatic decrease may largely reflect the partial closing of preexisting open cracks in the intervening Koae fault system. Such closing was documented by detailed measurements of deformation related to the forceful intrusion of magma into the southern part of the caldera in May 1970 (Duffield and others, 1976).

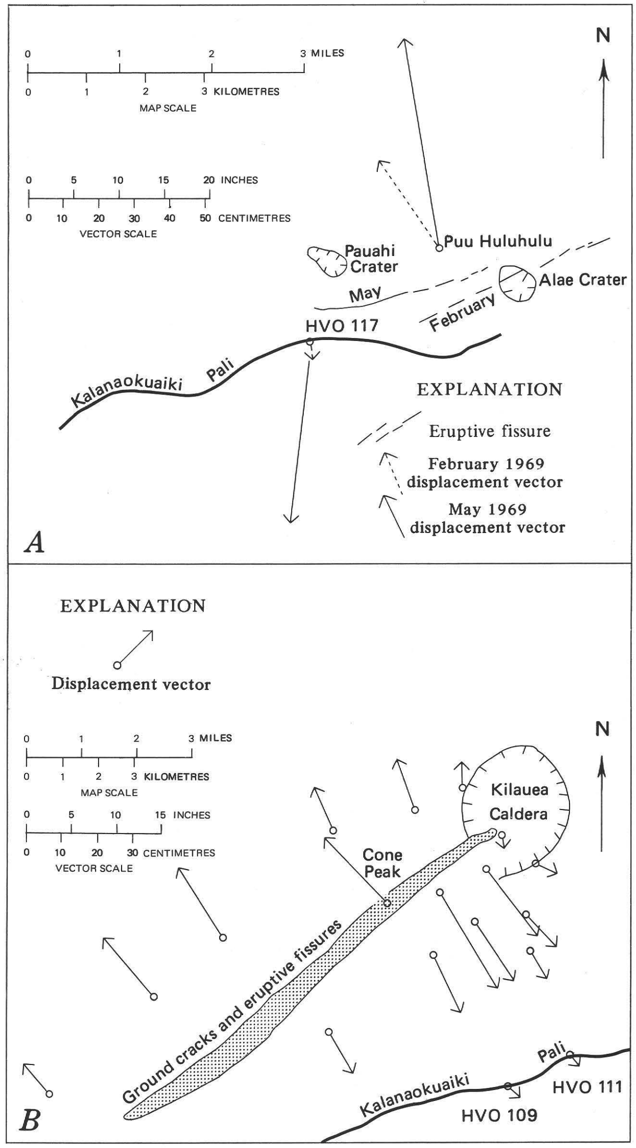


FIGURE 8.—Horizontal ground displacements related to specific recent eruptions at Kilauea. *A*, February and May 1969 upper east rift eruptions; base stations are HVO 135 and HVO 136 (fig. 6). February 1969 data from Swanson and others (1976). *B*, September 1971 summit and southwest rift eruption; base stations are Kulani, Strip, and Keakapulu (fig. 3). Kalanaokuaiki Pali is southernmost fault scarp in Koae fault system.

SPECIFIC DISPLACEMENT EVENTS ON THE SOUTH FLANK

The foregoing examples demonstrate that horizontal deformation accompanying single eruptions is similar in style to that suggested by the long-term geodetic surveys. We feel confident that the long-term displacements are the summation of several smaller episodes of deformation. All of the examples of specific displacement events, however, dealt with the rift zones themselves. The behavior of the south flank of Kilauea must be related in some way to the rift zones, but what is this relation?

The 1970–71 survey period (fig. 6) is the only reasonably short period for which measured ground displacements are available for the south flank. Even this period, however, includes three eruptions, a prolonged episode of summit inflation, and more than a year during which other processes could have affected the south flank. The measured displacements for this period are consistent with the interpretation that they directly reflect the effects of the eruptions, but we now present more specific examples of deformation events on the south flank.

Strains between March 6 and August 27, 1970.—Displacement vectors on the south flank are nearly parallel to one another, so that there is little, if any, extension or contraction along lines perpendicular to them. However, there has been measurable extension or contraction across the Hilina fault system, parallel to the direction of displacement, for all survey periods, so that coastal stations are displaced by different amounts, generally smaller, than inland stations. Thus we can use the observation of strain along a south-southeast direction, with little or no strain perpendicular to that direction, to infer displacement of the type indicated by the more complete survey. This is not the only interpretation, but we believe it reasonable in the context of Kilauea deformation.

As an example, figure 9 shows the results of geodimeter measurements for three triangles on the south flank between March 6 and August 27, 1970. Most lines that cross the Hilina fault system at high angles contracted, whereas lines approximately parallel to the trend of the system did not change length significantly. Contractile strain characterized the area. The principal axis of contraction (E_2) of the calculated strain ellipse is oriented approximately perpendicular to the east rift zone and Koae fault system and parallel to the azimuth of overall south-flank displacement. The minor axis of the ellipse is very small, probably within the limits of survey error. These relations suggest that the south flank underwent seaward displacement during this short survey period.

During this period, two ground-cracking events took place in the upper east rift zone and adjacent part of Kilauea Caldera (fig. 9). On April 9, wide ground cracks opened in and west of Aloi Crater, and the crater was filled by new lava. On May 15, a major earthquake swarm (Endo, 1971) accompanied the inferred forceful intrusion of magma into the southeast part of the caldera; this intrusion resulted in severe uplift, dilation, and some ground cracking (Duffield and others, 1976). A reasonable interpretation is that the measured contraction and inferred displacement of the south flank were caused by one or both of these deformation events.

Linear contraction of a geodimeter line between 1965 and 1970.—We can extend the observation that strain accumulates across the Hilina fault system during displacement events to the interpretation of long-term measurements of a geodimeter line oriented nearly parallel to the azimuth of displacement of the south flank. Figure 10 shows the location of this 5.4–km-long geodimeter line, which was measured 30 times between August 1965 and December 1970; measurements were effectively halted in December 1970, when the line began to skim the top of a newly erupted lava flow, and soon thereafter the inland station was covered by lava. In the 5⅓-year period of measurements, the distance contracted a net amount of 32 cm (an extensional strain of -5.9×10^{-5}). The rate of contraction, at least during 1969–70 when measurements were most frequent, was related in a systematic way to major magmatic and structural events on the upper east rift zone and adjacent areas in the caldera and Koae fault systems. Each new eruption or ground-cracking event in 1969–70 was followed 2 to 4 weeks later by an episode of rapid contraction. Measurements were infrequent before 1969 but are consistent with the later pattern. The period of maximum contraction includes three upper east rift eruptions, all of which were accompanied by substantial ground cracking and large measured horizontal displacements (Jackson and others, 1975; Swanson and others, 1976).

Beginning in early 1970, the geodimeter line began to show periods of extension alternating with periods of contraction. Most of these extensions could be correlated in time with earthquakes of $M \geqslant 3.5$ having epicenters within a few kilometres of the geodimeter line and focal depths between about 5 and 9 km. All such earthquakes between February and November 1970 are noted with respect to changes of line length in figure 10. After the earthquakes in July and September, we predicted extensions later confirmed by measurements. There was no large earthquake on the south flank before the November extension, but a 30-km-deep earthquake with M of 4.0 took place beneath the southern part of the caldera at the end of October.

Summary and interpretation of horizontal deformation on the south flank—The data presented in this section show that the south flank responds to specific rift eruptions and related structural events by undergoing displacement away from the rift zones. The resulting deformation is similar in style to that shown by the long-term survey periods. Almost certainly, the long-term displacements are the cumulative effect of several magmatic and structural events along the rift zones and within the adjacent summit caldera.

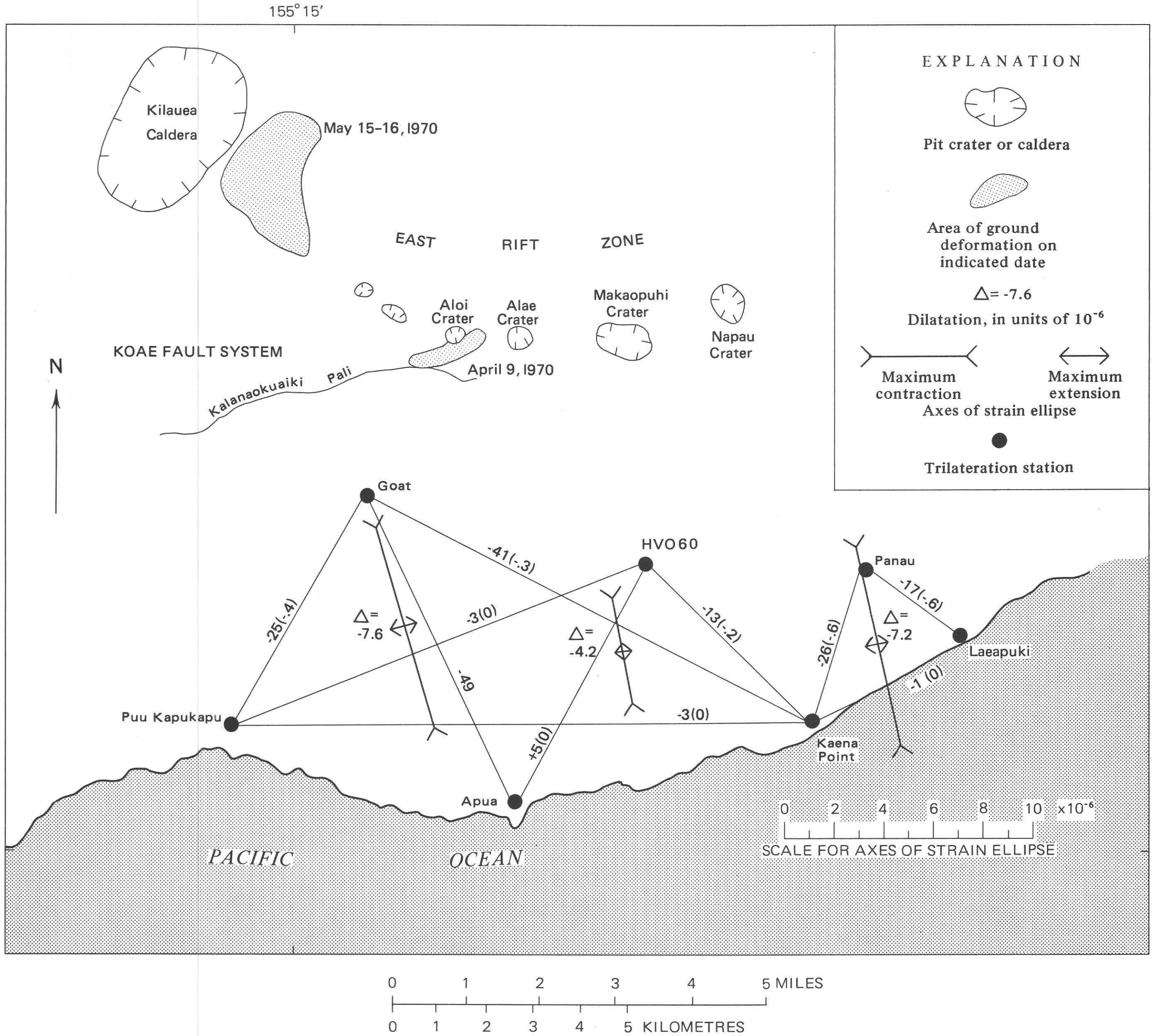


FIGURE 9.—Changes in line length, extensional and dilational strains, and orientation and magnitude of principal axes of strain ellipse for three triangles on south flank between March 6 and August 27, 1970. Changes in length are in millimetres; numbers in parentheses are extensional strains in units of 10^{-5}. Dilation and center of strain ellipse are plotted near center of gravity of each survey triangle, following usual convention. Kalanaokuaiki Pali forms south side of Koae fault system.

The response of the south flank to rift events is not immediate; instead, there is a lag time of 2 to 4 weeks before the wave of deformation reaches the area 5 to 10 km from the rift zones, an average propagation velocity on the order of 100 m/day. The lag time shows that the displacement is a result, not a precursor, of the magmatic event in the rift zones, a conclusion consistent with the timing of seismicity on the south flank relative to the rift events.

Throughout this century, deformation of the south flank has resulted in net contraction across the Hilina fault system (fig. 5D), and this contraction was closely monitored between 1965 and 1970. Such long-term accumulation of contractile strain may have reached a critical level by early 1970, so that comparatively small but nearby earthquakes could trigger brief episodes of extension. We return to this contention in a later section on the Hilina fault system.

VERTICAL DISPLACEMENT OF SOUTH AND NORTH FLANKS OF KILAUEA

Repeated leveling surveys on Kilauea during this century document substantial vertical displacement of

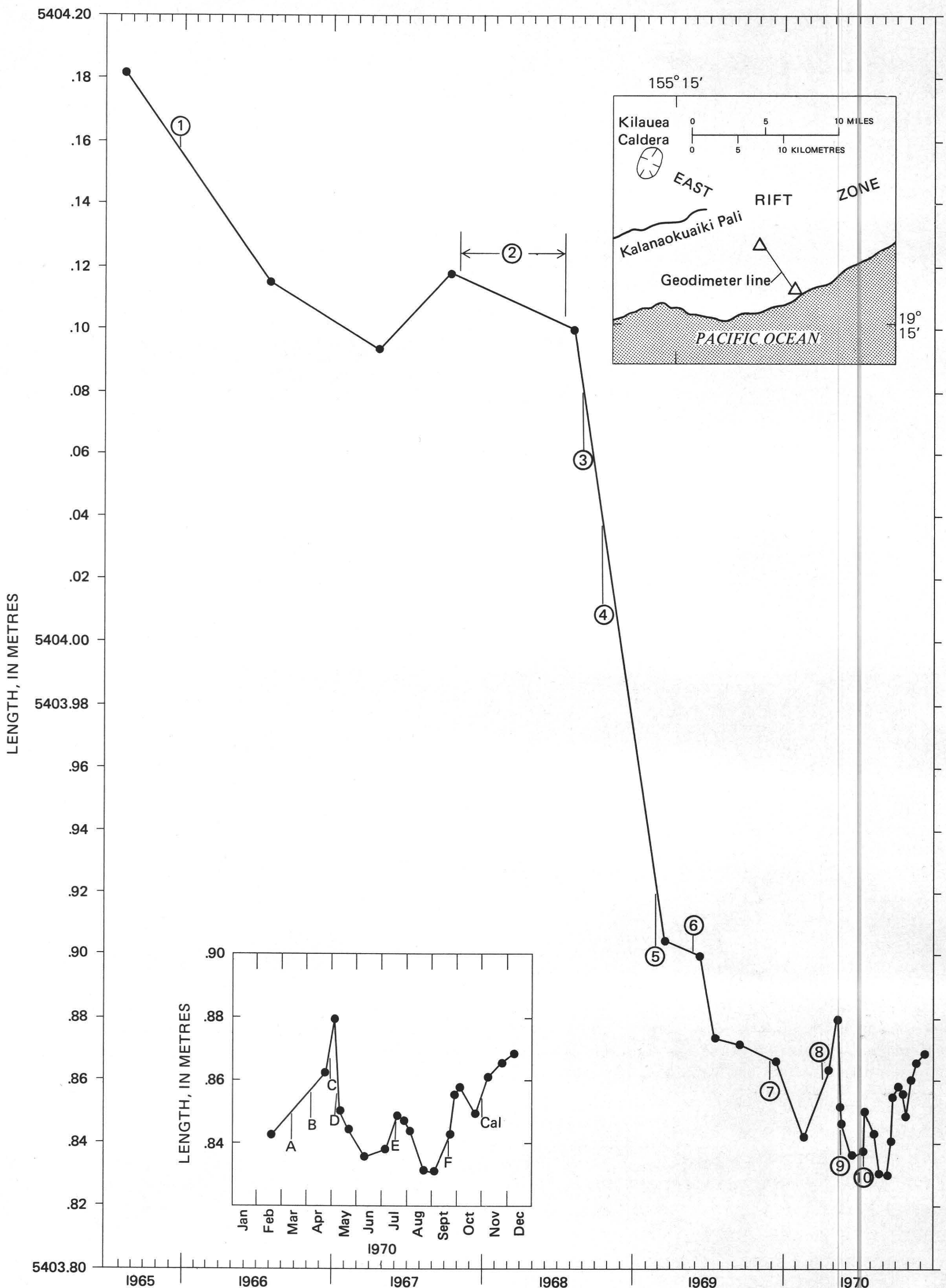


FIGURE 10.—Contraction of 5.4-km-long geodimeter line (upper right inset) on south flank of Kilauea between August 1965 and early December 1970. Major magmatic and structural events are: 1, December 1965 eruption and ground cracking; 2, November 1967–July 1968 summit eruption; 3, August 1968 eruption; 4, October 1968 eruption; 5, February 1969 eruption; 6, beginning of May 1969–October 1971 Mauna Ulu eruption; 7, new fissure north of Alae Crater, December 1969; 8, new fissure and cracks in and west of Aloi Crater, April 1970; 9, intrusion and cracking in southern part of Kilauea Caldera, May 1970; 10, new fissure east of Mauna Ulu, July 1970. Inset in lower left shows changes in line length relative to times of all south-flank earthquakes of magnitude (M) $\geqslant 3.5$ between February and early December 1970 and one caldera (Cal) earthquake at the end of October: A, M=4.1; B, M=3.8; C, M=3.9; D, M=4.1; E, M=3.5; F, M=4.3; Cal, M=4.0.

the south flank and minor, if any, displacement of the north flank. In general, these displacements correlate in time and space with nearby eruptions, a relation also found true for the horizontal deformation. We are not concerned here with deformation in the summit area, although it is well known that this area undergoes uplift between eruptions and, generally, subsidence during eruptions (Wilson, 1936; Fiske and Kinoshita, 1969). We also do not examine vertical deformation within the rift zones themselves, for several workers have previously shown that the area near an eruption site is uplifted, commonly with the development of a keystone graben along the linear crest of the uplift (Macdonald and Eaton, 1964; Fiske and Koyanagi, 1968, fig. 8; Jackson and others, 1975; Swanson and others, 1976).

The leveling data on the flanks were obtained over periods of several days during more extensive surveys. All available information indicates that no deformation occurred during the periods of leveling. The leveling data were evaluated and tabulated by Okamura and Swanson (1975) and Karren (1959). In particular, Okamura and Swanson (1975) recomputed data from a 1921 survey in order to correct for errors in rod length and instrument collimation by somewhat more accurate methods than previously used (Wilson, 1935).

CENTRAL PART OF SOUTH FLANK

Leveling surveys were completed across the central part of the south flank in 1921, 1958, 1965, and 1971 (fig. 11). The estimated qualities of the surveys are: 1921, poor third order; 1958, good second order; and 1965 and 1971, poor second order or good third order (see U.S. Geological Survey, 1966, for definitions of orders).

Bench mark (BM) 10 (inset, fig. 11*A*) is the only station common to all four surveys and is taken as the local datum point for comparison purposes. However, it is located in the area of deformation and may itself have undergone some vertical displacement. In fact, surveys suggest that BM 10 was uplifted about 17 cm relative to a tide gage at Hilo between 1921 and 1958 and 9 cm more between 1958 and 1971 (Okamura and Swanson, 1975, table 3). Although the leveling data, particularly in 1921, are of relatively poor quality, it nonetheless seems safe to conclude that BM 10 has been progressively uplifted at least several centimetres relative to the tide gage since 1921. Consequently, all displacements given in figure 11 are minimum values and would probably be several centimetres larger if computed relative to the tide gage. The Hilo tide gage is sinking at a rate of 4.1 mm/yr relative to sea level, but this is considered as reflecting islandwide isostatic, rather than local, subsidence (Moore, 1970).

Comparison of elevations for the 1921, 1958, and 1971 surveys, which followed the same route but included more bench marks in the later surveys, indicates uplift and seaward tilting of the south flank, increasing in magnitude toward the east rift zone (fig. 11*A*). The amount of uplift near the east rift zone is large, amounting to at least 2 m at BM 2728 since 1921; most of this displacement, more than 1.65 m, took place since 1958. The maximum recorded uplift since 1958 is 2.1 m at BM YY35, near the south edge of the east rift zone; this bench mark was not occupied in 1921. Most bench marks were displaced much more between 1958 and 1971 than between 1921 and 1958, indicating an accelerated rate of uplift in recent years, during which volcanic activity along the east rift zone has been intense (table 1). The rate of horizontal deformation likewise has increased since 1958 (compare figs. 5*E* and 5*B*).

The apparent subsidence of BM 2503 for the 1921–58 and 1921–71 periods is anomalous (fig. 11*A*) but could be removed if the 1921 elevation were about 0.9 m too high. A blunder of this magnitude (about 1 yd) would have been easy to make if the leveling rod were misread, as yard rods were used, but should have been caught in field computations. We rechecked the 1921 field notes but found no error, so that, if our suspicion is correct, an entry error must also have been made. The 1921 elevation of BM 2728 would also have to be lowered by the amount, if any, by which BM 2503 is in error.

An alternative interpretation is that the 1921 elevations are good and the displacements are those indicated in figure 11A. In this regard, the subsidence of BM YY37 (2302) between 1958 and 1971 should be noted. This area may actually experience local subsidence at times, perhaps in some way related to the rather abrupt change in slope 5 km southeast of Makaopuhi Crater (inset, fig. 11*A*), which may reflect an unrecognized fault in the Hilina fault system. Regardless of which interpretation is correct, the overall increasing uplift toward the rift zones is still the dominant pattern of displacement.

The 1965 leveling, and part of the 1971 leveling, followed a route (inset, fig. 11*B*) slightly south and west of the 1921 and 1958 routes, and none of the older bench marks except BM 10 and BM YY35 was occupied. The 1965–71 displacement profile (fig. 11*B*) indicates uplift much like that shown by other profiles (fig. 11*A*) but at an accelerated rate. For example, BM YY35 was uplifted about 1.6 m between 1965 and 1971 but only 0.5 between 1958 and 1965, and BM HVO 53 was uplifted about 0.75 m between 1965 and 1971 but only about 0.4 m between 1958 and 1965 (extrapolating from BM 2503, very close to BM HVO 53). Volcanic activity along the upper east rift zone has been especially intense since 1961 (table 1).

The displacement profiles of figure 11 (except the questionable 1921–58 profile) mimic in a general way

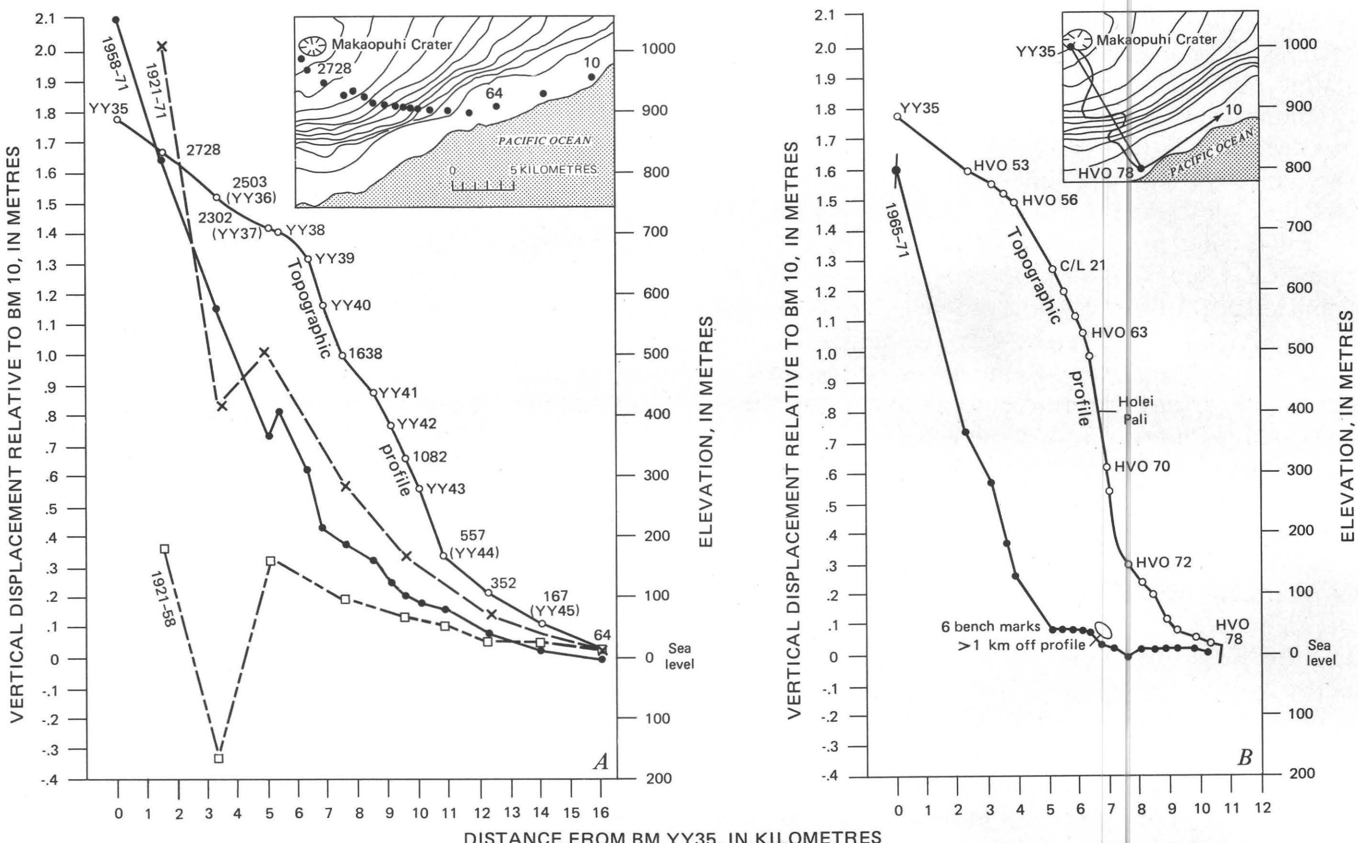


FIGURE 11.—Vertical displacement and topographic profiles across central part of south flank between 1921 and 1971. Datum point is BM 10. *A*, Leveling surveys of 1921, 1958, and 1971, along Kalapana Trail. *B*, Leveling surveys of 1965 and 1971 along Chain of Craters Road. Steep parts of topographic profiles are flow-mantled fault scarps, such as Holei Pali. Bench mark YY35 was destroyed in 1969, but its 1971 elevation can be estimated to within 0.1 m by comparison with a nearby point. All data are from Okamura and Swanson (1975, tables 5 and 7). Inset maps show location of bench marks and leveling routes in *A* and *B* and line of topographic profile in *B*. See figure 2 for location of Makaopuhi Crater. Contour interval in inset maps is 75 metres.

the shape of the topographic profiles. Both leveling routes cross the Hilina fault system, and both the topographic and displacement profiles show abrupt changes in slope across the fault system, suggesting recent fault movement. This effect is especially noticeable for the 1965–71 profile (fig. 11*B*), which is nearly perpendicular to the trend of the fault scarp. The nature of another abrupt change in the 1965–71 profile, between HVO 56 and C/L 21, is not known because lava flows covered the area in 1970–71, prohibiting reoccupation of intermediate bench marks and obscuring evidence of possible surface rupture.

EASTERN PART OF SOUTH FLANK AND LOWER EAST RIFT ZONE

The 1921 and 1958 leveling surveys extended north of BM 10, across the lower east rift zone and the north flank (fig. 12). For comparison purposes, BM 359.3, in Keaau, is used as the datum point. Leveling from the tide gage established in 1926 at Hilo indicates that BM 359.3 has been stable within survey error relative to the gage (Okamura and Swanson, 1975, table 3) since 1926; we cannot document such stability between 1921 and 1926 but think it reasonable. The poor quality of the 1921 leveling must be considered in any quantitative interpretation of the displacement data.

The 1921–58 profile indicates uplift of the south flank and lower east rift zone, with a maximum displacement of nearly 25 cm near the topographic crest of the rift zone. The uplift is asymmetric, the area south of the crest showing substantially more displacement than that to the north. BM 10, the datum point used in figure 11, was uplifted about 17 cm.

Lower east rift eruptive and intrusive events between 1921 and 1958 include the 1955 eruption (Macdonald and Eaton, 1964), the fissure zone of which is crossed by the leveling profile, and the 1924 ground-cracking episode centered 10–15 km east of the leveling route (Finch, 1925).

The section between BM 10 and BM 655 was releveled in 1973 (Hawaiian Volcano Observatory, unpub. data). Assuming BM 655 did not change elevation

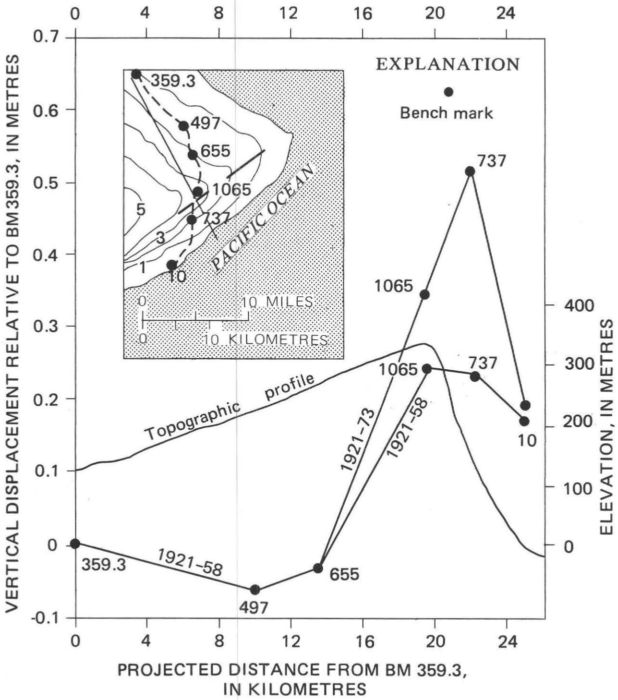


FIGURE 12.—Vertical displacement and topographic profiles across lower east rift zone and eastern part of south flank between 1921, 1958, and 1973. Datum is 1958 elevation of BM 359.3. Inset shows location of bench marks, leveling route (dashed line), line of topographic profile (light solid line), and generalized eruptive fissures for 1955 eruption (heavier solid lines). Displacement data are projected onto line of topographic profile. Contour interval on inset map is 100 m.

between 1958 and 1973, displacements relative to it can be used to indicate the total displacement since 1921. This assumption is dubious but serves to show that a striking change in the pattern of uplift took place (fig. 12). The crest of the uplift shifted southward, with BM 737 being displaced nearly three times as much as BM 1065 between 1958 and 1973. Thus the overall uplift since 1921 is centered 2 to 3 km south of the topographic crest of the rift zone, in an area where no past eruptive activity is known.

The 1958–73 uplift can be examined more closely by using data from surveys conducted along the same route in 1964 and 1969 by R. W. Decker (1965, 1969; Dieterich and Decker, 1975) (fig. 13*A*). The uplift grew sporadically between 1958 and 1973. The rate of uplift was greatest between 1964 and 1969 (maximum of 3.2 cm/yr) and least between 1969 and 1973 (maximum of 1 cm/yr). The crest of the uplift did not shift laterally during this time. A grabenlike zone developed on the north limb of the uplift, progressively deepening with time; new cracks cutting a paved road apparently define the south side of the graben (R. W. Decker, oral commun., 1969). Both the crest of the uplift and the graben are located south of all known eruptive fissures and vents. BM YY71, only 100 m from a 1955 eruptive fissure (vent R of Macdonald and Eaton, 1964), subsided, possibly owing to thermal contraction during solidification of the 1955 dike and cooling of its wallrock.

In contrast to the profile in figure 13*A*, displacement profiles about 8 km farther east show progressive subsidence relative to BM YY80 since 1958 (fig. 13*B*). The subsidence is of much smaller magnitude than the uplift, amounting to a maximum of about 12 cm at BM YY195. The rate of subsidence was most rapid between 1958 and 1964 and subsequently slowed. Maximum subsidence is centered along the 1955 fissures and may reflect thermal contraction. Removal of this effect by smoothing the profiles leaves a broad basin with a small but sharp uplift centered at BM YY198; this uplift persists on all three profiles and therefore must be real although of small amplitude.

WESTERN PART OF SOUTH FLANK AND KOAE FAULT SYSTEM

Leveling surveys in 1958 and 1971 permit construction of a displacement profile across the Koae fault system and western part of the south flank (fig. 14). The profile is complicated by displacements caused by several faulting events in the Koae fault system, principally in 1965 (Fiske and Koyanagi, 1968). The south flank shows uplift and southward tilting, increasing in magnitude toward the Koae fault system.

NORTH FLANK

The north flank of Kilauea and adjacent part of Mauna Loa between Hilo and the summit were leveled three times between 1926 and 1971. Okamura and Swanson (1975, table 3) tabulated the measured elevations and showed that changes are generally unsystematic and close to or within survey error. Changes are significant only near the summit, presumably as a result of the documented summit tumescence (Okamura and Swanson, 1975, table 3). Displacement on the eastern part of the north flank between 1921 and 1958 was similarly small (fig. 12).

SUMMARY AND INTERPRETATION OF VERTICAL DEFORMATION ON THE NORTH AND SOUTH FLANKS

The following statements characterize vertical deformation of the flanks of Kilauea during the 20th century:

(1) The north flank is virtually stable, as also indicated by both seismic and horizontal displacement data.

(2) In most places, the south flank has been increasingly uplifted toward the rift zones and Koae fault system, with the maximum displacement within or just south of the east rift zone.

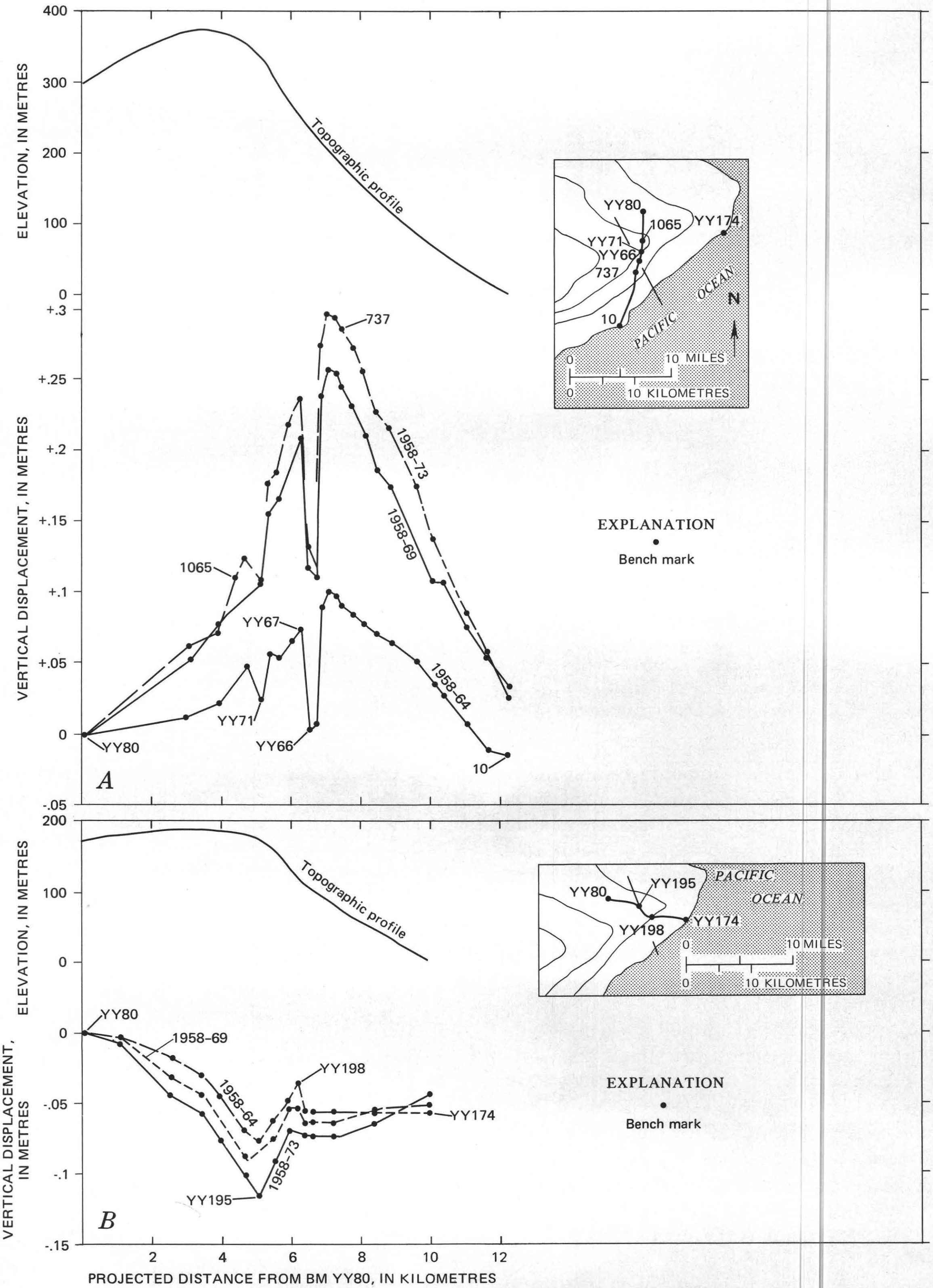


FIGURE 13.—Vertical displacement and topographic profiles across lower east rift zone of Kilauea between 1958 and 1973. *A*, Profiles between BM YY80 (Pahoa) and BM 10 (Kalapana); *B*, Profiles between BM YY80 and BM YY174 (Pohoiki). Datum is 1958 elevation of BM YY80. Inset maps show locations of leveling route, key bench marks, and line of topographic profile. Displacement data are projected onto line of topographic profile. Contour interval of inset maps is 150 m.

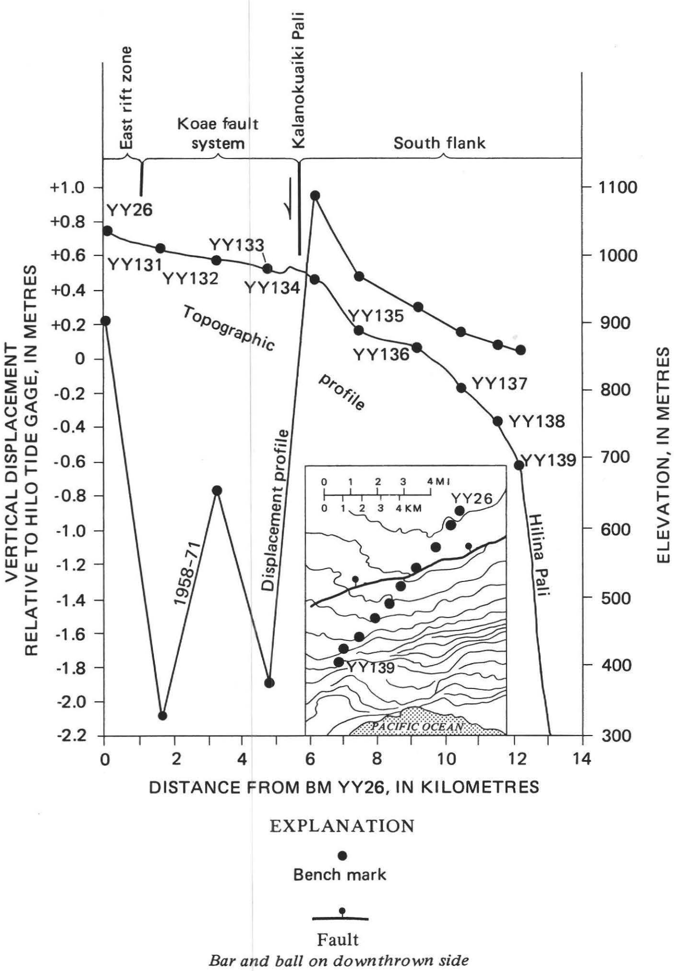


FIGURE 14.—Vertical displacement and topographic profiles across Koae fault system and western part of south flank between 1958 and 1971. Datum point is Hilo tide gage. Inset map shows location of bench marks along Hilina Pali Road and Kalanaokuaiki Pali, a normal fault that forms south boundary of Koae fault system. Contour interval is 60 m. Hilina Pali is a prominent fault-line scarp.

(3) The rate and amount of displacement in areas of large uplift correlate directly with the number and proximity of rift eruptions or related faulting events for any given time period.

The uplift of the south flank can be ascribed to the effects of forceful intrusion of magma into the rift zones. Such uplift was suggested by Macdonald (1956, p. 286) on the basis of geologic arguments. Along the upper and middle east rift zone, intrusion has apparently been followed quickly by eruption, so that a direct relation between the timing of uplift and eruption is evident. Along the lower east rift zone, intrusion appears to have proceeded more slowly or over longer periods of time, or both; consequently, uplift during noneruptive periods may occur (fig. 13*A*). Subsidence, such as that shown in figure 13*B*, may indicate migration of magma to another site.

The displacement profiles that completely cross the lower east rift zone (figs. 12 and 13) are particularly instructive because of their asymmetry. The profiles suggest that either the south flank is virtually decoupled from the north flank along the rift zone or that the uplift is the result of intrusion of magma as south-dipping dikes into the rift zone (Dieterich and Decker, 1975). South-dipping dikes would explain the location of maximum uplift 1–2 km south of the lower rift zone (fig. 13) but would not account for the location of maximum uplift within the central and western parts of the rift zone, where vertical or nearly vertical dikes seem most likely. We examine the possibility that dikes dip southward in the section. "A Structural Model for the South and North Flanks of Kilauea," concluding that some dikes may dip southward at comparatively low angles where gravity faults of the Hilina fault system impinge on the rift zone, but that most dikes along the length of the rift zones are steeply dipping to vertical.

The profile across the Koae fault system and adjacent south flank (fig. 14) is more complicated than the others. Fiske and Koyanagi (1968, p. 19) suggested that the uplift immediately south of the Koae system during the December 1965 eruptive and faulting episode could be attributed to "* * * elastic rebound due to the sudden release of accumulated stresses along Kalanaokuaiki Pali," the fault that forms the southern margin of the fault system. Another possibility is that magma was laterally intruded as a dike into the Koae from the site of concurrent eruption near the intersection of the fault system and east rift zone, causing uplift similar to that elsewhere along the northern edge of the south flank. In May 1973, such a dike apparently was intruded several kilometres into the Koae system from its source along the upper east rift zone (Duffield, 1975; Koyanagi and others, 1973).

RELATIVE MAGNITUDE OF HORIZONTAL AND VERTICAL DISPLACEMENTS

The relative magnitude of horizontal and vertical deformation can be estimated by comparing measured horizontal displacements with measured or extrapolated vertical displacements for approximately the same time intervals at six south-flank triangulation stations (table 3). The horizontal component of displacement is at least several times larger than the vertical component. This generalization appears valid across most of the south flank, judging from comparisons of figures 11–14 with figures 5–8. The relative importance of vertical displacement probably increases near the rift zone, but no triangulation stations are located in the critical area to test this suggestion. Within the rift zones themselves, vertical displacements (both uplift and subsidence) are commonly large

TABLE 3.—*Comparison of horizontal and vertical displacements at six triangulation stations on the south flank of Kilauea Volcano*

[Vertical data have been extrapolated from nearest bench mark if the triangulation station was not included in the leveling survey.]

Station	Time interval		Displacement, m		Ratio of horizontal to vertical
	Horizontal	Vertical	Horizontal	Vertical[1]	
Hilina	1961–70	1958–71	0.8	+0.02	[2]40
Panau	1914–58	1921–58	2.1	+ .30	[3]7
	1958–70	1958–71	2.3	+ .30	8
Kupapau	1914–58	1921–58	1.2	+ .18	[3]7
	1958–70	1958–71	0.8	+ .07	11
Hakuma	1914–58	1921–58	1.4	+ .17	[3]8
	1958–70	1958–71	0.7	+ .09	8
Laeapuki	1961–70	1965–71	1.3	+ .08	[3]16
Kaena Point	1961–70	1965–71	1.3	+ .05	[3]26

[1]Vertical displacement relative to Hilo tide gage or, for the 1921 survey, BM 359.3. The 1965 survey was based on BM 10 and has been adjusted by assigning half the uplift of BM 10 between 1958 and 1971 to the 1965–71 interval.
[2]Probably minimum ratio, because time between horizontal surveys is shorter than that between vertical surveys.
[3]Probably maximum ratio, because time between horizontal surveys is longer than that between vertical surveys.

but smaller than horizontal ones at a given location (Macdonald and Eaton, 1964, fig. 34; Jackson and others, 1975; Swanson and others, 1976), although exceptions associated with the formation of major grabens are known (Jaggar, 1924; Swanson and others, 1972, p. 114).

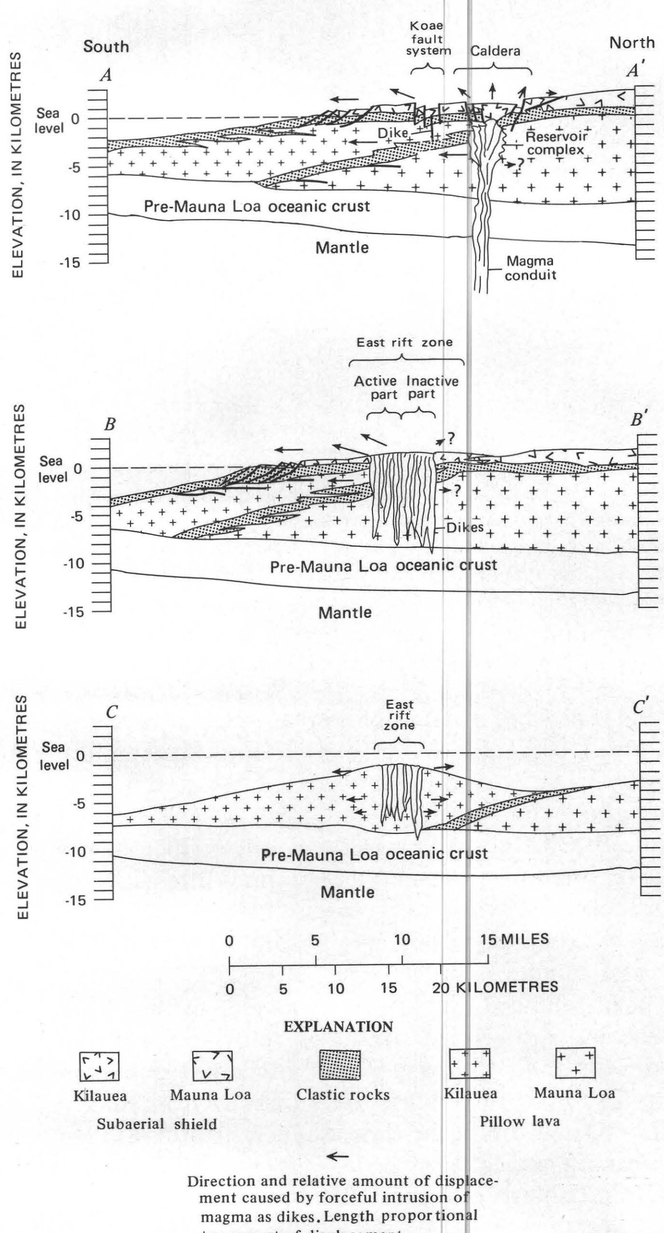
The predominance of horizontal displacement on the south flank probably results from dikes that are forcefully wedged into the rift zones, causing horizontally-directed dilation that manifests itself by the opening of gaping fissures and the corresponding displacement of the south flank away from the rift zones.

A STRUCTURAL MODEL FOR THE SOUTH AND NORTH FLANKS OF KILAUEA

The nature and timing of seismic events and ground displacements prompt us to propose a model in which the south flank of Kilauea moves in response to forceful intrusion of magma into the rift zones (fig. 15). Intrusions, in the form of dikes, literally shoulder the south flank upward and southward and are effectively tearing the volcano apart along the two rift zones. From the timing of seismic and deformation events, we conclude that intrusion is the immediate cause of south flank displacement, not the reverse.

The data further suggest that the north flank is relatively stable and does not respond notably to intrusive events. This lack of response may be explained by the following two interrelated reasons: (1) the north flank is buttressed by the huge mass of Mauna Loa and the other volcanoes on the island, whereas the south flank is free to move southward; and (2) the development of the rift zones of Kilauea is dominated by the gravitational stress field imposed by Mauna Loa, on which Kilauea is constructed, and this gravitational field favors seaward displacement away from the center of the volcanic pile (Fiske and Jackson, 1972).

Figure 16 illustrates the present great difference in size between Kilauea and the rest of the island, particularly Mauna Loa. It is reasonable that this large mass of older rock might buttress Kilauea, especially since intrusion in the rift zones is a shallow process, probably taking place within 8 km, chiefly 5 km, of the ground surface (fig. 15). The buttress effect would be important even if the contact of Mauna Loa and Kilauea were vertical and extends to the old sea floor on which the volcanoes are constructed. In all probability, however, the contact interfingers as shown schematically in figures 15 and 16. In fact, we consider it probable that Mauna Loa was already a large volcano when Kilauea first began forming, because of the present large difference in size.



Not so easily visualized, but perhaps of greater importance, is the control exerted by Mauna Loa on the stress field in which intrusions at Kilauea take place. Fiske and Jackson (1972) demonstrated that gravitational stresses govern the direction of dike propagation in gelatin models free of other stresses, whether the dikes are injected directly beneath the top of the model or at some place on its flanks. If this modeling can be generalized to naturally occurring volcanic edifices, then a younger shield built on the sloping flank of an older shield should come under the influence of the latter's gravitational stress field, which should in turn control dike orientation and overall shape.

Fiske and Jackson (1972, p. 314–317) showed how this concept could explain the relation of Kilauea to Mauna Loa. Assuming that Kilauea inherited the gravitational stress field of Mauna Loa, intrusion of magma into the rift zones as dikes should produce dilation in a direction consistent with this stress field, that is away from the central part of the edifice and toward the free slope. Thus, the Fiske-Jackson model predicts that the south flank of Kilauea would be far more mobile than its north flank, a relation indicated by the seismic and geodetic evidence.

Much of the east rift zone of Kilauea probably extends beyond the effects of Mauna Loa, however. The submarine part of the rift zone extends at least 110 km beyond the east tip of the island (Macdonald and Abbott, 1970, p. 313). The submarine rift zone forms a prominent ridge standing high above its base (Moore, 1971; fig. 1), and extends well beyond the bulk of Mauna Loa; consequently, it may have its own high-level gravitational stress field. If so, dike-induced dilation should be directed equally northward and southward as *C–C′* in figure 15 portrays. The most recent intrusion into the submarine part of the rift zone may have taken place in 1924 (Finch, 1925), and earthquake activity in this area has been minor since then (Koyanagi and others, 1972, and references therein). Thus it is unlikely that this part of the rift zone is presently undergoing much deformation. The north side of the submarine part of the rift zone is steeper than the south side (Moore, 1971), possibly because of gravity-induced slides, although why slides are concentrated on only one side is not evident.

FIGURE 15.—Diagrammatic cross sections, with no vertical exaggeration, through Kilauea depicting our interpretation of relation between east rift zone and rest of volcano. Locations of sections shown in figure 1. *A–A′*, Summit region. Magma is intruded, primarily as dikes, into east rift zone from upper part of magma conduit or reservoir complex. Magma only rarely enters Koae fault system. *B–B′*, Middle east rift zone. Magma intruded as dikes from beneath summit region wedges south flank seaward and upward. North flank moves little if at all. Dikes in inactive part of rift zone are older, extend to greater depths, and terminate upward at slightly lower elevations than dikes in active part. *C–C′*, Submarine part of east rift zone. Displacements caused by intrusion of magma are assumed to be nearly symmetric across rift zone. In *A–A′* and *B–B′*, gravity sliding is interpreted to take place chiefly in layer of hyaloclastic rocks above and interbedded with pillow lava, displacing seaward part of south flank downward. Hyaloclastic material is generated at shallow depth but sloughed downward by slides and current action (Moore and Fiske, 1969), possibly forming a deposit much thicker than indicated in *A–A′* and *B–B′*. Depths to base of volcanic pile and to mantle from Hill (1969), and bathymetry and rock type from Moore and Fiske (1969). Depths to contact of Kilauea and Mauna Loa are speculative; rocks from the two volcanoes are shown as interfingering only at relatively shallow depths, consistent with our belief that Kilauea did not begin to form until Mauna Loa was already a large edifice. Flows of pillow lava from Mauna Kea may underlie Mauna Loa pile (fig. 16).

The south flank is assumed to be mobile everywhere south and southeast of the rift zones, but the depth to which it remains mobile is not easily determined. The flank is probably mobile to a depth of at least 5 km because seismic evidence during eruptions suggests that some intrusion takes place at such a depth. Almost all earthquakes on the south flank have focal depths of less than 15 km, and most have depths of 8 km or less, calculated using a crustal model devised by J. P. Eaton and E. T. Endo, based on Hill's (1969) refraction study (Koyanagi and others, 1975). The mobility may therefore extend to a depth of at least 8 km, which is almost certainly well within the edifice of Mauna Loa and near the level of the old sea floor on which Mauna Loa is built (fig. 15; Hill, 1969). Seismic evidence suggests that the lithologic break at shallow depths between subaerial flows capping the volcano and pillow lavas forming the submarine bulk of the volcano (fig. 15; Moore and Fiske, 1969) does not influence the overall mobility of the south flank, although the Hilina fault system may largely bottom out at this level, as discussed in a later section.

It seems unlikely that the base of the mobile flank would be defined by a single plane of dislocation. Instead, we favor a model in which the displacement gradually decreases below the depth at which intrusion of magma is concentrated. Thus the base of the mobile south flank may actually be a zone several kilometres thick, and much of the displacement may be taken up by many local adjustments within the pillow complex.

Moore and Krivoy (1964) believed the east rift zone to be a southward-dipping structure forming the sole of a large landslide block (the south flank). They attributed dilation of the rift zone to "* * *southward movement of the flank of the volcano south of the rift zone under the influence of gravity and inflation of the dipping rift zone by magma from the summit reservoir" (1964, p. 2043). However, the documented uplift of the south flank associated with rift dilation and eruption during recent years argues against landslide-type faulting along the rift zone, for subsidence would be

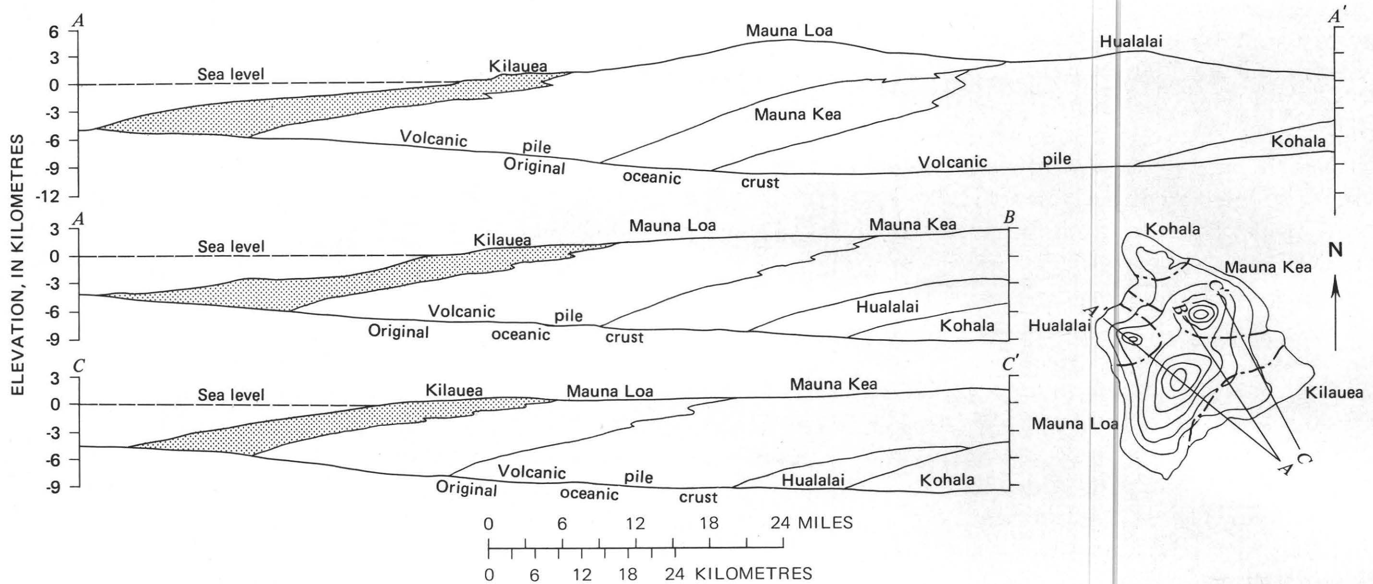


FIGURE 16.—Diagrammatic cross sections, with no vertical exaggeration, across Kilauea, Mauna Loa, Hualalai, and part of Mauna Kea. Depth to base of volcanic pile is from Hill (1969); submarine topography off south flank of Kilauea from Moore and Fiske (1969). We assume that each volcano grew to substantial size before its next younger neighbor began erupting, as indicated by interfingering only at relatively shallow depth. Other volcanoes that never reached above sea level may contribute to volcanic pile. Contour inverval in index map 600 m.

expected, at least on the south side of the rift zone, from gravity faulting during eruption. In addition, seismic studies show that earthquakes associated with intrusion emanate from directly below the surface trace of the rift zone, not south of the zone (Koyanagi and others, 1972). Finally, intrusion and eruption along the east rift zone precede seismic activity and displacement on the south flank, suggesting that forceful intrusion is the immediate cause, not the effect, of the rift dilation and displacement.

We agree with Moore and Krivoy (1964) that gravity faults characterize the Hilina fault system (see later section) but believe that these are only relatively superficial, secondary structures resulting from instabilities generated by injection of steeply dipping dikes into the rift zones. A broad unfaulted area lies between the rift zone and the Hilina system, except locally along the lower east rift zone, and this alone suggests different origins for the rift zones and Hilina fault system.

Gravity plays an important role in our model, just as it does in the model of Moore and Krivoy. Were regional tectonic stresses dominant over local gravitational stresses, the rift and flank structures would doubtless be much different, as Fiske and Jackson (1972) found them to be in isolated Hawaiian shields. However, from the observed timing of events we contend that intrusion, not gravity-inducing sliding, opens the rift zones.

Ryall and Bennett (1968) interpreted the east rift zone to be a north-dipping structure related to a large normal fault that displaces the crust and upper mantle. This interpretation was largely based on gravity data in the east rift area, which, as we show in a later section, can better be interpreted as indicating southward growth of the east rift zone with time. Their model presents a host of other difficulties not easily rationalized with seismic, geodetic, and geologic data largely gathered subsequent to their interpretation.

Dieterich and Decker (1975) suggested that dikes in the lower east rift zone dip 45° S., on the basis of comparison of the 1958–64 leveling profile in figure 13*A* with finite-element models of dike emplacement into an elastic medium. The critical feature of this and other leveling profiles along the same route that suggests such a low dip to these workers is its marked asymmetry, the south flank showing substantial uplift and the north flank almost no uplift. In their interpretation, the graben just north of the crest of the uplift formed at the locus of maximum stress, as located by the model.

Dieterich and Decker's (1975) interpretation conflicts with the magnitude of horizontal displacements in that area. For example, the horizontal displacements at Iilewa and BM YY66 between 1958 and 1970 are each about 50 cm (fig. 5*B*), whereas the maximum displacement predicted from the finite-element model for about the same time period (1958–69) is only 5 cm, an order of magnitude less. The large horizontal displacements do not fit any of the several models given by Dieterich and Decker (1975). Instead, the large extension may reflect the intrusion of a family of dikes, not just one dike. In his helpful review of this paper, Decker (oral commun., 1973) suggested that the horizontal displacements resulted from ground rupture, a process not covered by the finite-element model. This may be true, but no new ground cracks were observed in the critical area along the rift zone during this time, and seismic activity in this area, especially since 1961, has been so slight as to suggest no surface rupturing.

Another problem with the interpretation by Dieterich and Decker is that the most definitive data, those north of the rift zone, lie 3 to 4 km off the profile used in their model comparison, whereas most of the other data are much closer to the profile. Whether the data can be projected so far and still be subject to quantitative interpretation is open to question, particularly in view of the observed rapid change from uplift to subsidence within a short distance along the rift zone (fig. 13). In the absence of data for the north flank closer to the profile, the displacement profile can be interpreted to indicate almost any dip desired.

As a result of these difficulties, we regard as unproven the existence of dikes in the lower east rift zone that dip much less than 90°. Nonetheless, we believe it is possible that some dikes in the area crossed by the leveling line in figure 13*A* do dip southward at angles of 50–70° because of the structural setting of this part of the lower east rift zone. In this area, the Hilina fault system intersects the rift zone (figs. 2 and 3), and magma supplied from steeply dipping dikes farther uprift could conceivably be intruded along some of the preexisting south-dipping fault planes. If such intrusion has taken place, we would view it as a perturbation from the otherwise steep dips of the dikes indicated elsewhere along the rift zone by the patterns of seismicity and uplift, and found on eroded Hawaiian shields (Stearns, 1947; Stearns and Vaksvik, 1935).

NUMERICAL COMPARISON OF VOLUMES OF INTRUSION AND DEFORMATION

If south flank deformation is caused by injection of dikes, then there should be a specific relation between the volume of deformation (defined principally by the amount of uplift) and the volume of intrusion. This relation will not be one-to-one, as the deformed rocks should decrease in internal volume to some degree. Still, a rough balance between the volume of intrusion and the volume of deformation should be maintained.

For the period 1958–71, a cross section of the south flank southeast of Makaopuhi Crater shows an increase in volume of about $10^7 m^3/km$ owing to uplift, computed from data in figure 11*A* assuming negligible deformation of the north flank. A dike 0.5 m wide, a width consistent with many observed Hawaiian dikes, has a volume of $5 \times 10^5 m^3/km$ if it extends from the surface to a depth of 1 km, $1.5 \times 10^6 m^3/km$ to a 3-km depth, and $2.5 \times 10^6 m^3/km$ to a 5-km depth. The calculated volume of deformation could thus be balanced by 4 to 20 such dikes. Table 1 indicates that 8 to 13 intrusive events occurred near Makaopuhi during this time; the uncertainty in number reflects the unknown degree to which events near Alae Crater and Mauna Ulu could have influenced the deformation. At least one of these intrusive events, in August 1968, involved very shallow magma injection—possibly only to a depth of 500 m (Jackson and others, 1975)—and seismic evidence for others suggests that the dikes bottom at depths of 5 km or less. Thus, a reasonable agreement between the volumes of deformation and intrusion is evident.

Similar calculations for the 1965–71 period (fig. 11*B*) show a volume of deformation of $8 \times 10^6 m^3/km$, which could be balanced by 3 to 16 dikes. During this time there were five to eight intrusive events. Again, the two volumes are in reasonable agreement.

Quite obviously these calculations are crude, for the existing data and theoretical models are inadequate for detailed analysis. Nonetheless, the approximate balance between the volumes of intrusion and deformation is of some value, as our model predicts that the dikes make room for themselves by displacing the wallrock up and to the side.

THE RIFT ZONES AND KOAE FAULT SYSTEM AS ZONES OF DILATION

If the south flank of Kilauea is being displaced southeastward because of the forceful intrusion of magma as dikes into the rift zones, its contact with the less mobile part of the volcano should be marked by a zone of dilational opening. We postulate that such a zone of dilation is defined by the two rift zones and the Koae fault system.

The east and southwest rift zones have long been recognized as containing normal faults, grabens, and gaping cracks with matching opposite walls, and in recent years such features have been observed and measured in the process of formation. Fissures that form during any given rift eruption tend to be arranged in an en echelon pattern, with either a right- or left-hand sense of offset depending on their location along the length of the rift zone (Macdonald and Eaton, 1964; Moore and Krivoy, 1964; Duffield and Nakamura, 1973; Jackson and others, 1975). Analysis of these en echelon fissures by the method of Nakamura (1970) and Duffield and Nakamura (1973) suggests dilational opening perpendicular to the rift zones. However, new lava erupted along the rift zones repeatedly covers the fissures, so that the net amount of dilation over long periods of time cannot be estimated adequately. Thus we must look elsewhere for pertinent quantitative information regarding the amount of cumulative dilation.

The Koae fault system (fig. 3) is not the site of eruptions, except for small, infrequent outbreaks near its intersections with the east and southwest rift zones. Further, it has, in the recent geologic past, seldom been covered by lava erupted from vents farther upslope, as the geologic maps of Peterson (1967) and Walker (1969) show. As a result, the Koae fault system serves as a window through which the net effect of displacements produced by many episodes of deformation are well displayed, and it is the best place to find direct structural evidence that displacements such as those measured in this century have characterized the recent geologic past.

Duffield (1975) found that the Koae fault system is characterized by gaping cracks arranged in en echelon fashion, some of which coalesce to form long, sinuous zones of normal faults, generally with north-facing scarps, that define the south margins of asymmetric grabens (fig. 17). Symmetric grabens are uncommon. The fault scarps are as high as 20 m, although most are less than 5 m, and single cracks are as much as 2 m wide. Most cracks and faults dip vertically, strike N. 75° E., and open perpendicular to their strike, as determined by numerous measurements of matching crack walls. A similar direction of opening was determined by Duffield and Nakamura (1973) from a study of the pattern of en echelon cracks, and this direction virtually parallels that of horizontal displacement vectors on the south flank (fig. 5). This evidence suggests rather definitely that similar intrusive events cause both the south-flank displacement and the Koae (and rift zone) dilation.

Knowledge of the amount of dilation recorded by the Koae fault system and the time during which this dilation occurred is helpful in evaluating the history of past displacement events. The total amount of dilation along two profiles across the fault system was measured to be 18.69 m (fig. 17, *A–A′*) and 32.55 m (fig. 17, *B–B′*) (Duffield, 1975). Our field observations suggest that the total dilation decreases westward from profile *A–A′*. We interpret the greater amount of opening in the eastern part of the fault system to reflect proximity to the principal source of dilation—intrusion within the east rift zone.

The measured dilation is that accumulated since the

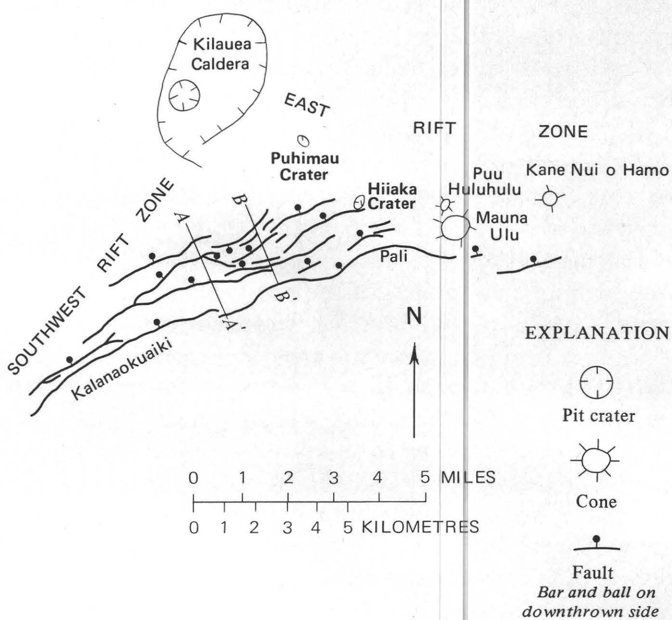


FIGURE 17.—Principal zones of normal faults in Koae fault system. Note that most fault scarps face north. *A–A′* and *B–B′*, traverses along which total dilation was measured. Modified from Duffield (1975). Kalanaokuaiki Pali is southernmost fault scarp in Koae fault system.

last lava flows covered the area. Duffield (1975) broadly estimated this time to have been 500–2,500 years ago from evidence involving, respectively, extrapolation of 20th century rates of displacement on the south flank and the interpreted age of the summit caldera (Powers, 1948; Rubin and Suess, 1956), in which lava erupted at the summit tends to pool rather than flooding areas farther downslope. Hawaiian legends concerning the latest flow in the Koae area suggest that the 500-year age may be more nearly correct (H. A. Powers, written commun., 1974). An ancestral Koae fault system apparently existed before the latest flows because there is field evidence that these flows poured into cracks along the trace of present faults, particularly Kalanaokuaiki Pali. Thus, the available data suggest that displacements similar to those during this century have occurred over at least the past few hundreds to thousands of years.

Fault-bounded blocks in the Koae system appear to have been tilted south-southeastward during deformation, important evidence bearing on the origin of the faults. Such tilting was directly measured over the December 1965 deformation event, when one tiltimeter on a fault block showed more then 200μrad of seaward tilt (Fiske and Koyanagi, 1968, fig. 3). Unusually steep southward slopes on some fault blocks suggest a substantial cumulative effect of such tilting, as least locally.

The tilting and general configuration of the fault locks are remarkably similar to structures obtained n modeling clay by Cloos (1968, fig. 18) during experiments that produced asymmetric grabens. Such a graen forms when one side of a clay cake resting on two verlapping metal sheets is pulled laterally while the ther side remains stationary. The major fault scarps ace toward the stationary block, and the part of the nodel near the zone of separation sags and tilts in the lirection of movement. It is interesting that the direcion of tilt is opposite to that which generally accomanies landsliding, when slopes are rotated toward the tationary block, not away from it. The observed strucures in the Koae fault system are just those expected rom Cloos' experiment, if movement of the south flank way from the less mobile part of the volcano is analoous to pulling apart the clay cake. Duffield (1975, fig.) discussed in more detail the relation of Cloos' model o structural problems at Kilauea.

The interpretation of the Koae fault system as a omplex asymmetric graben produced by seawardlirected dilation in the rift zones accords with our ostulate that the system is part of the tear-away zone eparating the south flank from the rest of the volcano. ntrusion in the rift zones causes dilation, which then ropagates into the Koae system as the south flank moves seaward, gradually dying out away from the site f intrusion. In this sense, the Koae is a relatively passive area, reacting to events in the rift zones without enerating activity itself. Intrusion within the Koae eldom takes place, presumably because it lacks a dirrect connection to the summit reservoir complex. Erupions that do occur are thought to be fed by magma that nters the fault system during rift eruptions near the ntersection with the Koae. Such an eruption occurred n 1972, when the migration of earthquake foci into the Koae indicated that a dike was being intruded from the ite of initial eruption on the east rift zone (Koyanagi nd others, 1973). Such a rift eruption apparently auses dilation, which propagates into the Koae, creatng stress conditions favorable for magma to leave the ift zone and intrude the fault system.

In summary, the Koae fault system and rift zones ogether define the zone of dilation that, according to our interpretation, necessarily bounds the north edge of the mobile south flank.

RELATIVE IMPORTANCE OF THE TWO RIFT ZONES IN GOVERNING MOBILITY OF THE SOUTH FLANK

The horizontal displacements from the September 1971 eruption (fig. 8*B*) indicate that the south flank esponds to dike intrusion in the southwest rift zone, just as in the east rift zone. However, the southwest rift zone has apparently been of subordinate importance in governing south-flank structures. It is much shorter, lower, and narrower than the east rift zone (fig. 1, 2) presumably because it contains fewer dikes and thus has fewer eruptions. Furthermore, Duffield (1975) found that both width and total dilation of the Koae fault system decrease westward away from the east rift zone, a pattern suggesting dominance of the east rift zone in the development of the Koae. Duffield (1975) and Duffield and Nakamura (1973) also showed that the direction of dilation across the Koae parallels that across the east rift zone and the south flank but is in most places oblique to the direction of dilation along the southwest rift zone. Additional evidence suggesting dominance of the east rift zone is the east-west extent of the Hilina fault system; this system extends south of much of the length of the east rift zone and Koae fault system but apparently ends, or becomes much less conspicuous, near the southwest rift zone (fig. 2; Stearns and Macdonald, 1946, pl. 1). From this evidence we conclude that events in the east rift zone largely govern the mode of deformation of the south flank.

SOUTHWARD MIGRATION OF THE RIFT ZONES AND SUMMIT RESERVOIR

EAST RIFT ZONE

Several lines of evidence suggest that the active part of the east rift zone has migrated southward together with the south flank. The topographic crest of the rift zone is as much as 4 km north of the recently active upper and middle sections of the rift zone (fig. 18*A*). This crest is probably a constructional ridge built by eruptions along it, by flows from a parasitic shield centered near the east end of Kilauea Iki (fig. 2; see also the topographic map of the Volcano quadrangle), or both. Relics of possible vent areas along the crest are hidden beneath dense jungle cover or younger flows erupted from the parasite shield. No eruptions have taken place along the crest in late prehistoric or historic time, suggesting that, if the ridge was built by rift eruptions, the active part of the middle and upper rift zone shifted southward from the crest before this time.

The relation of the complete Bouguer gravity anomaly to the recently active part of the east rift zone likewise suggests southward migration (fig. 18*B*). Kinoshita and others (1963; W. T. Kinoshita, oral commun., 1973) interpreted this positive anomaly as caused by a swarm of dikes of greater bulk density than the vesicular flows or loosely packed pillows that make up most of Kilauea and underlying Mauna Loa. These dikes must be the feeders for lava erupted on the

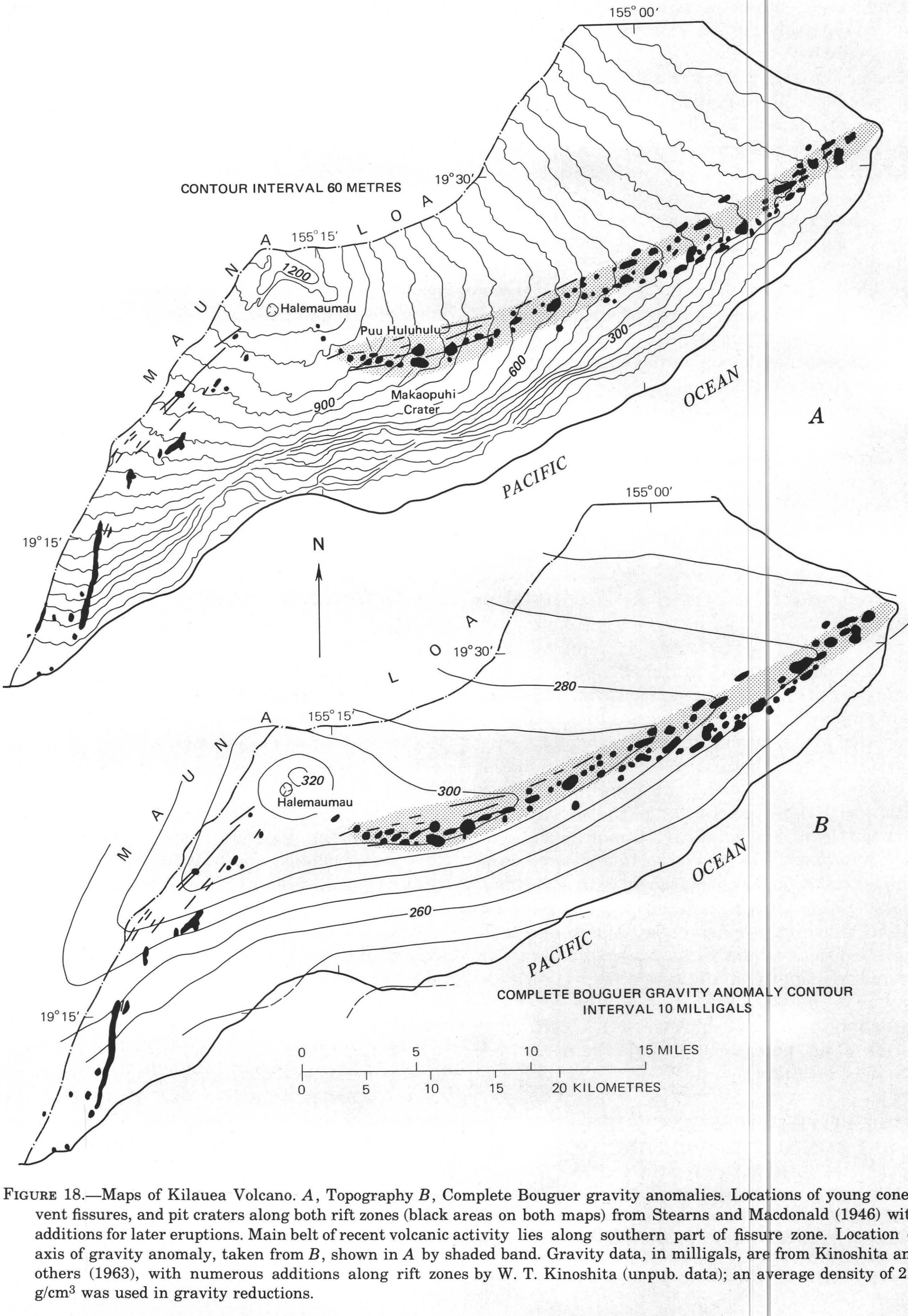


FIGURE 18.—Maps of Kilauea Volcano. *A*, Topography *B*, Complete Bouguer gravity anomalies. Locations of young cones, vent fissures, and pit craters along both rift zones (black areas on both maps) from Stearns and Macdonald (1946) with additions for later eruptions. Main belt of recent volcanic activity lies along southern part of fissure zone. Location of axis of gravity anomaly, taken from *B*, shown in *A* by shaded band. Gravity data, in milligals, are from Kinoshita and others (1963), with numerous additions along rift zones by W. T. Kinoshita (unpub. data); an average density of 2.3 g/cm^3 was used in gravity reductions.

rift zone. However, almost all of the recently active vents are along the south side of the gravity anomaly (fig. 18*B*), not along its crest. This suggests that young dikes have been added preferentially to the south side of the swarm.

The gravity anomaly on the east rift zone is strongly asymmetric, with steeper gradients south of the rift zone. Thus the dike swarm may be thought of as defining a wedge-shaped prism, with a gentle north face and a steep south face (W. T. Kinoshita, oral commun., 1973). Such a prism is the expected product of southward-directed growth of the rift zone with time, because successively younger dikes would on the average be intruded to higher elevations than older dikes to the north as the rift zone grew southward and upward (fig. 19). This interpretation differs from that of Ryall and Bennett (1968), who used the gravity data of Kinoshita and others (1963) to infer a north-dipping rift zone.

The gravity axis is located between the topographic crest and the active part of the upper east rift zone (fig. 18*A*). This is consistent with the hypothesis of southward migration if the rate of migration has recently been sufficiently rapid to prevent construction of a newer and higher constructional ridge. The 1969–74 eruptions of Mauna Ulu created a new, low ridge along the active part of the rift zone.

If the active part of the east rift zone is moving southward, that part of the rift zone nearest the summit should show more displacement because of the likelihood of more frequent dike injection near the summit magma reservoir. This may account for the greater divergence between the topographic and gravity axes, and between the location of young vents along the western and eastern parts of the rift zone. The gravity data (fig. 18*B*) support the concept of greater dike frequency near the magma reservoir, for the Bouguer values are higher along the upper east rift zone than along the lower east rift zone.

The conspicuous bend in the active part of the east rift zone, defined by the location of young cones, pit craters, and fissures (figs. 1 and 18), may reflect the interplay between the greater degree of intrusion near the source reservoir, the southward migration of the rift zone, and two stress systems, one radial to and confined to the reservoir area, the other the gravitational system favoring south-southeastward displacement. Evidence of an approximately radial stress field at the summit comes from studies of ground deformation accompanying episodes of inflation and deflation (for example, Fiske and Kinoshita, 1969; Swanson and others, 1976). Near the summit area, dikes tend to be radial because of the dominance of the radial stress field. Away from the summit, however, the gravitational field controls the orientation of dikes. Figure 20 shows schematically the stages in the possible development of the bend in the east rift zone. At an early stage in the history of the rift zone before much displacement of the south flank (fig. 20*A*), dikes near the summit area are radial to the source reservoir and tend to cluster in a zone perpendicular to the direction of south-flank displacement. As the rift zone migrates southward because of dike accretion (fig. 20*B*), dikes remain radial to the reservoir near the summit, but curve farther downrift to accommodate themselves to the gravitational stress system. This accommodation, combined with the presence of more dikes near the reservoir than far from it, causes the bend in the rift zone. Figure 20*C* shows the present stage, in which the bend has been accentuated as more radial dikes were intruded near the summit area. Possible southward migration of the magma reservoir itself at a slower rate than that of the south flank, as discussed later, could also augment the development of such a bend. During the 1960's and early 1970's, a fairly unobstructed conduit apparently transported magma from the summit reservoir down to and beyond the bend in the rift zone (Swanson and others, 1976). This conduit became periodically plugged, leading to eruptions near the bend and in the summit area. Such a conduit is thought to have developed along the trace of the rift zone at some unknown, but very recent, time.

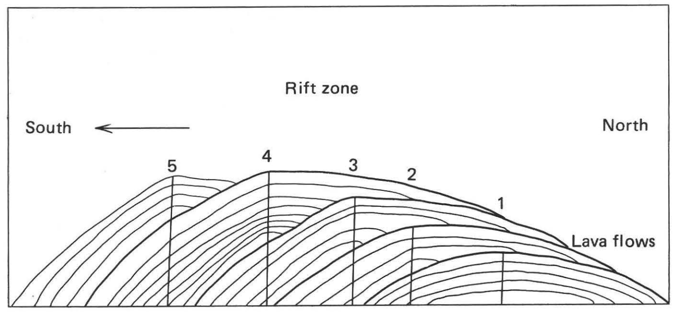


FIGURE 19.—Schematic cross section showing how continued lateral growth of a rift zone generally results in successively younger dikes intruding to higher elevations. Dike 1 is oldest; dike 5 is still erupting lava that may eventually cover flows from dike 4. Arrow, direction of lateral growth. In a less schematic diagram, dikes would be so closely spaced that they would define a wedge-shaped prism of higher density than wallrock, leading to an asymmetric gravity anomaly.

Jackson, Swanson, Koyanagi, and Wright (1975) pointed out the infrequency of eruption along the southeast-trending part of the east rift zone, between the caldera and the conspicuous bend. The southeast trend of this part of the rift zone nearly parallels the direction of horizontal displacement in the zone of dilation and in the south flank, so that intrusion may be more easily accommodated without surface rupturing and resultant eruption than elsewhere along the rift zone.

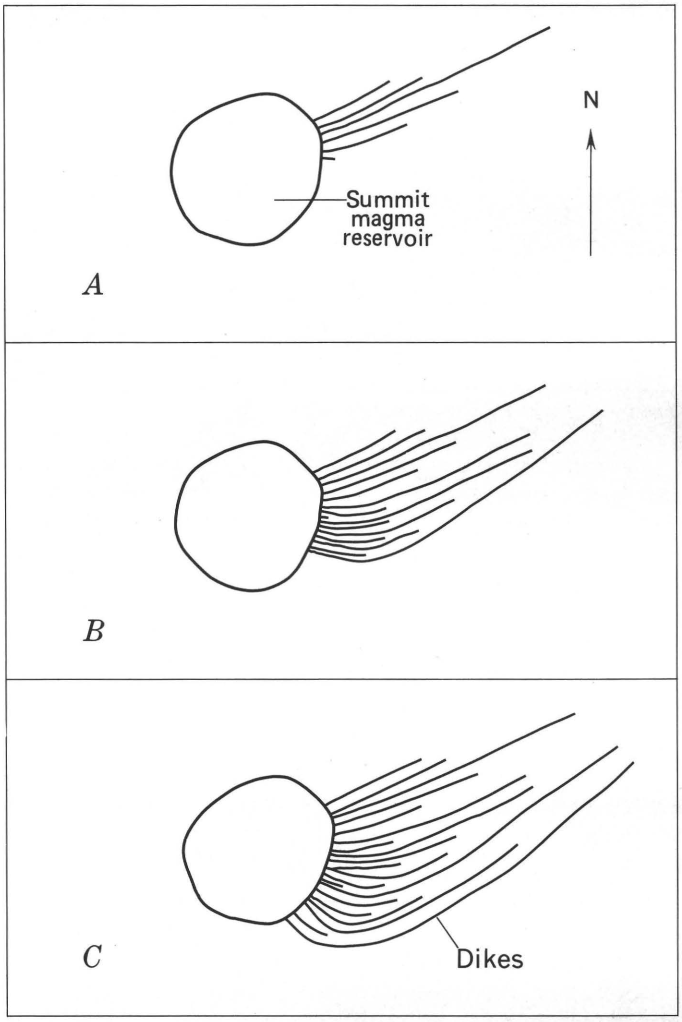


FIGURE 20.—Sketches in plan view showing possible development of bend in east rift zone of Kilauea Volcano owing to accretion of dikes along south side of east rift zone near summit magma reservoir. *A*, Early stage. Dikes near summit area are radial to source reservoir and tend to cluster in a zone perpendicular to direction of south-flank displacement. *B*, Intermediate stage. As rift zone migrates southward because of dike accretion, dikes remain radial to reservoir near summit but curve farther downrift to accommodate themselves to gravitational stress system. *C*, Present stage. Bend accentuated as more radial dikes intruded near summit area. Size of reservoir, from which dikes are fed, is exaggerated for clarity, and shape is greatly simplified from its probable amoeboid form. Model assumes a constant stress field favoring dilation in a south-southeast direction of rift zone, and a radial stress field immediately surrounding reservoir. Trend of dikes near summit reservoir is largely governed by radial stress field, on the east rift zone by south- southeast dilational field.

A slight change in the strike of the fissures within the east rift zone may reflect the southward migration of the rift zone. Recognizable prehistoric and historic fissures north of the main belt of activity trend more westerly by 5°–10° than do those in the main belt (fig. 18*A*; Stearns and Macdonald, 1946, pl. 1; Moore and Krivoy, 1964; Swanson and others, 1976, pl. 1); in other words, northerly cracks are more nearly radial to the summit than those farther south. Such a pattern may be accounted for by the southward growth of the rift zone (fig. 20*B* and *C*). Even fissures of historic eruptions, such as those in August 1968 and February 1969 (Swanson and others, 1976, pl. 1), have somewhat different trends depending on their location, suggesting that older dikes may, to some extent, act as guides for later intrusions, at least beyond a distance of several kilometres from the source reservoir.

Thus, several lines of evidence suggest that the east rift zone is migrating south-southeastward, parallel to the direction of south flank displacement and away from the rest of the island, including the north flank of Kilauea itself. This migration is consistent with our model suggesting that rift intrusion takes place in a predominantly gravitational stress system and that Kilauea is buttressed by the rest of the island and can accommodate intrusion chiefly by moving away from this buttress. In this model, each dike injected into the rift zone effectively shores up and stablilizes the zone, replacing loose pillow lava and highly jointed subaerial flows with relatively dense and coherent dike rock. Successively younger dikes then tend to seek out the contact between dike and flow rocks, accreting along the south margin of the rift zone where south-southeastward dilation is easiest. In this way the active part of the rift zone migrates seaward. This model is derived to explain the net long-term result. Actually the active part of the rift zone is 2–3 km wide, not razor thin, but the net result of intrusion into this zone is that predicted by the model.

SOUTHWEST RIFT ZONE

The southwest rift zone occupies a more complicated structural setting than the east rift zone. The southwest rift zone is bounded on the west by scarps in the Kaoiki fault system (fig. 2), a system of normal faults cutting the southeast flank of Mauna Loa. This fault system is still active, generating many hundreds or thousands of earthquakes yearly (Koyanagi and others, 1972, fig. 2; Koyanagi and others, 1966). One interpretation is that the Kaoiki is a system of landslide-type faults, analogous to the Hilina fault system on the south flank of Kilauea (p. 33), that developed during an earlier stage in the growth of Mauna Loa before Kilauea was very large. The southwest rift zone of Kilauea may overlie part of the Kaoiki system, and it is conceivable that at least some of the southwest rift fissures merge at depth with Kaoiki faults. Further discussion of these possibilities must await detailed mapping of the southwest rift zone and the Kaoiki fault system.

Although little is known about the southwest rift zone, available evidence is consistent with its seaward migration. The axis of the southwest rift gravity anomaly is offset toward Mauna Loa from the active part of the rift zone, as it is on the east rift zone (fig. 18*B*). The geologic maps of Stearns and Macdonald (1946, pl. 1) and Walker (1969) show that young vents tend to be located along the seaward side of the southwest rift zone, although this is complicated by vents within the western part of the Koae fault system (for example, Puu Koae and Cone Crater on the map by Walker, 1969). The vents of the September 1971 eruption also lie near the southeast edge of the rift zone. Thus we believe that the evidence is consistent with seaward migration, possibly less pronounced than along the east rift zone because of less frequent episodes of magma intrusion.

Figure 6 shows that several stations on the Mauna Loa side of the southwest rift zone were displaced away from the zone between 1970 and 1971. We suggest that this displacement was largely absorbed within the Kaoiki fault system by adjustment of fault blocks, in a manner similar to the way in which the east rift zone north of sites of intrusion responds to displacement events (figs. 5*A*, 5*C*, 7, and 8*A*).

SUMMIT RESERVOIR

The summit reservoir system may also be migrating seaward, though at a slower rate than the east rift zone. Such migration is suggested by: (1) the northeast-southeast elongate shape of the caldera (fig. 2); (2) the location of the present Halemaumau pit crater in the south part of the caldera; (3) the general southerly offset of the present caldera from the high area along its north margin; (4) the location of the present-day reservoir system beneath the south part of the caldera (Fiske and Kinoshita, 1969); and (5) the general southerly migration of the centers of ground deformation during periods of inflation and deflation related to reservoir filling and emptying (Fiske and Kinoshita, 1969; Jackson and others, 1975). Moreover, the positive gravity anomaly is centered somewhat north of the area of maximum ground deformation, a relation explained in a manner analogous to our rift-zone argument.

This migration presumably is closely related to the same processes affecting the rift zones, but may be more complex. The summit system is fed by conduits rooted in the mantle and hence is susceptible to lateral movements of the Pacific plate and the asthenosphere. Nonetheless, the postulated seaward migration is an expectable consequence of intrusion in a gravitational stress system favoring southeastward-directed dilation.

THE HILINA FAULT SYSTEM

Leveling surveys since 1921 document uplift in the Hilina fault system (figs. 2 and 3), yet abundant evidence indicates net subsidence of this area over the past few thousand years. Many geologists (for example, Stearns and Macdonald, 1946; Moore and Krivoy, 1964), have commented on the seaward-facing fault-line scarps, which in places are more than 500 m high and bounded by gaping cracks, as evidence of normal faults. Walker (1969) measured a seaward dip of about 50 degrees on one exposed fault plane in the Hilina system and noted that the sense of displacement on most of the faults is down toward the sea. The faults have been recently active, for they locally offset the youngest prehistoric flows in the area (Macdonald, 1956).

Two large subsidence events were recorded along the south coast during the 19th century. In 1823, a series of strong earthquakes accompanied ground cracking and subsidence of at least 46 cm at the village of Kaimu (Ellis, 1827, p. 195–196), a short distance from BM 10 (inset, fig. 13). In 1868, much of the south coastal area subsided 1–2 m or more during severe earthquakes accompanied by a local tsunami (Brigham, 1909, p. 103–113). These two events occurred far from sites of eruption and thus differ from the well-known development of a graben along the rift zone in 1924 near Kapoho, where the east rift zone enters the ocean (Finch, 1925; Jaggar, 1924).

The weight of geologic evidence and the two major recorded subsidence events compel us to believe that subsidence is the dominant sense of displacement in the Hilina fault system. If so, then the measured uplift since 1921 within the fault system is acting against the long-term trend and consequently may be decreasing the stability of the fault system. We suggest that this uplift, probably related to tumescence along the rift zone, combines with large seaward-directed displacements to produce instability, which ultimately triggers faulting along the unbuttressed south coast, such as took place in 1823 and 1868.

We interpret many of these gravity faults to bottom out at shallow depth (fig. 15 *A–A′*, *B–B′*), merging into bedding planes within the hyaloclastic deposits and underlying loosely packed pillow lava that form the basement to the subaerial part of Kilauea (Moore and Fiske, 1969). The displaced blocks are therefore similar to large landslide blocks, as suggested by Stearns and Macdonald (1946) and Moore and Krivoy (1964). Macdonald later objected to the landslide hypothesis, primarily because he believed that "* * *piles of volcanic rocks with average slopes of 4° to 7° should be very stable" (Macdonald, 1956, p. 286). More recent work by Moore and Fiske (1969, fig. 3), however, suggests that a significant thickness of clastic deposits dipping more than 10° seaward underlies the south flank, and we think that these deposits form an adequate medium in which lateral and downward slippage could occur.

Macdonald (1956) offered a challenging alternative to the landslide hypothesis, suggesting that tumescence in the summit region, and by implication the rift zones, causes the formation of nearly vertical fault zones paralleling the rift zones with the inland side upthrown. If this were true, rather straight, linear faults might be expected. Instead, however, faults in the Hilina system curve and branch (fig. 2), and some are arranged in en echelon sets on the limbs of large, arclike master faults (fig. 21). Macdonald's suggestion also does not explain the lack of such faults on the north flank of Kilauea, whereas landslide-type faults would only be expected on the south flank. The seaward dip of 50° that Walker (1969) measured on one fault plane is also inconsistent with Macdonald's suggestion. Finally, broad areas at the base of the Hilina, Holei, and the Poliokeawe fault-line scarps are nearly flat or even slope gently inland (fig. 22), consistent with rotation accompanying landsliding but not with faulting related to rift tumescence.

On the other hand, evidence from the 1965–71 leveling profile (fig. 11B) suggests that the inland side of Hilina faults can at times be displaced upward relative to sea level. That profile shows a sharp offset across Holei Pali, with uplift of 9 cm relative to BM 10. This reference point itself probably moved upward, if at all, during the survey period, so the true amount of uplift relative to sea level may be more than 9 cm. If the fault forming Holei Pali dips seaward, as we believe, then the indication of reverse movement is clear. We conclude that Macdonald's suggestion of uplift on the inland side of Hilina faults is indeed correct for certain periods of strong tumescence of the rift zones, such as between 1965 and 1971, but that this uplift is small and contrary to the dominant long-term sense of displacement within the fault system.

We estimate that the depth to the base of the hyaloclastic layer, within which we interpret many of the Hilina faults to die out, is between 1 and 3 km below sea level. The depth increases with distance from Kilauea. This estimate is based on the results of Moore and Fiske (1969, fig. 3) and considerations of the effects of isostatic subsidence (Moore, 1970), which may have lowered the layer several hundred metres since it formed. It is also possible that some of the Hilina faults bottom out in a deeper hyaloclastic layer formed during growth of Mauna Loa (fig. 15*A* and *B*).

Walker (1969) mapped several thin dikes in the western part of the Hilina fault system; others occur elsewhere in the system. On the basis of field relations, Walker interpreted these dikes as crack fillings fed from overriding flows. The numerous open cracks in the fault system make his interpretation seem likely to us, and we add that neither gravity data (fig. 18*B*; Kinoshita and others, 1963) nor seismic refraction data (Hill, 1969) suggest a dike swarm in this area. Thus we interpret the dikes as bearing no genetic relation to the Hilina fault system.

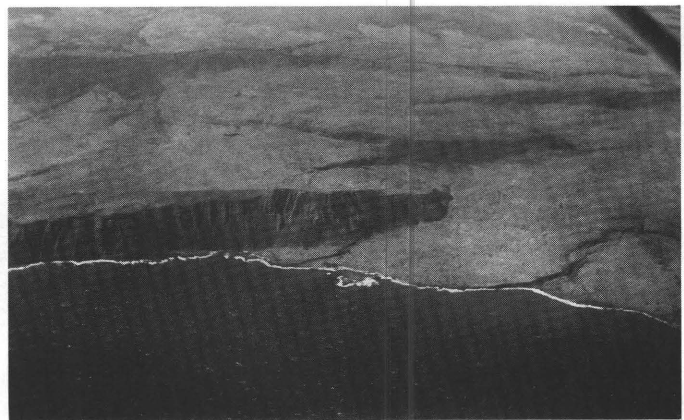


FIGURE 21.—Aerial view of Hilina fault system near Puu Kapukapu (lower left), highest scarp at coastline (fig. 2). Note en echelon arrangement of Puu Kapukapu and two scarps to northeast (middle and upper right). These en echelon fault-line scarps define west limb of a large fault block. En echelon faults also occur around small fault block (lower right corner) southeast of Puu Kapukapu. Puu Kapukapu is 320 m high. Scarp (upper left) is Hilina Pali. View to north-northwest.

FIGURE 22.—Nearly flat ground surface, which may reflect landward rotation during faulting, at base of Holei Pali (background) and unnamed fault-line scarp in Hilina fault system. Part of the area at immediate base of unnamed scarp is shown as virtually flat on Makaopuhi Crater 7½-minute quadrangle; lava ponded in this area in 1970–73 because of the low gradient, which in fact seemed slightly reversed in places when visited in 1970. Lava flows in 1969–74 have greatly modified topography from that shown on quadrangle map. View to northwest from near Kaena Point (fig. 2).

We interpret the shape of the coastline of the south flank to be controlled principally by where and how often lava flows erupted from the rift zones reach the sea, not by faults of the Hilina system. Only locally, as at the horsts near Kaimu (fig. 2) and Puu Kapukapu (figs. 2 and 21), do faults define the trend of the coastline. Most of the coastline is vulnerable to flows from the rift zones, but the area centered near Puu Kapukapu lies between the two rift zones and is less frequently inundated with lava; consequently the rest of the coastline has grown seaward on either side of this sheltered area, leaving a broad embayment. The area inland of this embayment is higher than most other parts of the fault system and includes Puu Kapukapu, the largest and most prominent horst in the Hilina system (fig. 21). This area may stand high because it is pinched between two segments of the south flank moving at slightly convergent azimuths away from the southwest and east rift zones, and so may be kept from subsiding as drastically as elsewhere by this relative compression.

Some of the data suggest that the Hilina fault system is poised for another episode of subsidence, with only minor additional ground displacement needed to trigger it. Figure 10 (lower inset) shows that, in 1970, every south-flank earthquake of $M \geqslant 3.5$ except one (fig. 10*D*) was immediately followed by a brief period of extension of a geodimeter line that otherwise showed strong net contraction in response to rift intrusion. No such correlation could be made in 1969. Possibly the amount of contraction across the fault system was reaching a critical level by 1970, so that earthquakes of even moderate size could trigger brief periods of strain release. Measurements in 1971 (fig. 6) and 1974 (Hawaiian Volcano Observatory, unpub. data) showed small extensions across the Hilina system, continuing the trend shown in 1970 and suggesting a further amount of strain release. Most of the strain acquired since 1965 (fig. 10) remains, however, and we anticipate a subsidence (strain-release) event of unknown magnitude in the not too distant future.

(A major subsidence event took place in the Hilina fault system on November 29, 1975, 15 months after this paper was submitted for publication. The subsidence, accompanied by an earthquake of $M=7.2$, affected much of the south coast of Kilauea and was as much as 3.4 m in the Puu Kapukapu area. Large extensions were measured across the Hilina system, and many fault scarps showed renewed displacement, with seaward side down. Two campers were killed by a tsunami generated by the earthquake and subsidence. These deaths emphasize the fact that the unstable nature of the south flank of Kilauea has broad social as well as geologic implications.)

APPLICATION TO OTHER OCEANIC VOLCANOES

We doubt that Kilauea is a unique volcano, and many of the principles gained from this study should be applicable to other oceanic shields. In Hawaii, for example, large landslide blocks are probably present along the west coast of Mauna Loa, bounded by the Kealakakua fault and unnamed faults between Hookena and Milolii (Stearns and Macdonald, 1946, pl. 1). We consider that the Kaoiki fault system on Mauna Loa (fig. 2) may also be a gravity-controlled landslide area, now partly shored up by Kilauea but still influencing the way in which the southwest rift zone of Kilauea is deformed. The landslide-type faults of Mauna Loa are downslope from a large rift zone and may well have developed in response to intrusion in that zone. Moore (1964) described possible giant landslide masses from Koolau and East Molokai Volcanoes that may owe their origin to dike intrusion in rift zones parallel to the coastline. The flanks of many older Hawaiian shields, such as Haleakala, West Maui, and East Molokai Volcanoes, dip 11 to 18 degrees seaward, much steeper than the average shield slope of several degrees. It is possible that these steep slopes partly reflect the net effect of doming caused by the intrusion of numerous dikes, although all of these volcanoes are at a more advanced stage of evolution than Kilauea and consequently may be subject to processes not observed during the young, tholeiitic stage.

Any oceanic shield volcano that grows on the flank of an earlier shield in the absence of a dominant tectonic stress system should behave similarly to Kilauea. Consequently, there are probably numerous volcanoes in the ocean basins that are characterized by asymmetric dilation of rift zones, a mobile seaward-facing flank, and secondary landslide blocks on the mobile flank. Recognition of these features in deeply eroded shields may be difficult, but there are possible clues to look for. Dikes in rift zones of buttressed volcanoes should on the average become younger from the landward side toward the seaward side; this might be detected in favorable circumstances by paleomagnetic or radiometric age data, temporally related differences in chemistry, or simple stratigraphic evidence. Examples in the Hawaiian chain where rift zones could be examined in this way are East Molokai and Koolau Volcanoes; both of these grew adjacent to an older shield (Fiske and Jackson, 1972) and are dissected sufficiently to expose dikes. On the other hand, unbuttressed shields such as Niihau show no such tendency and when eroded reveal symmetric rift zones in the central part of the edifice. We hope that examination of eroded shields with the model of Kilauea in mind can shed more light on their internal structure, as well as feed back important clues for a better understanding of Kilauea itself.

SUPPLEMENTAL INFORMATION

PROBLEM OF GROUND DEFORMATION OCCURRING DURING THE HORIZONTAL SURVEYS

The 1971 trilateration was carried out between October 3 and 18; the 1970 trilateration took place chiefly between late August and late September, with a few supplementary distances measured in November and early December. The volcano was quite stable during both surveys, as indicated by measurements of tilt and strain at the summit and the absence of major earthquake activity elsewhere on the volcano.

The 1961 triangulation was conducted between March 22 and July 25, spanning the last three days of the March summit eruption, the entire July summit eruption, and the intervening period of substantial summit inflation (about 90 microradians at one tiltmeter at the summit) (Richter and others, 1964, fig. 3). These events, especially the inflation, almost certainly caused some of the poor triangle closures reported in the summit region (table 2) but probably did not greatly affect the rest of the volcano.

The 1958 triangulation lasted from February to June. Only slight summit deformation took place during this time, as judged from records of tilt (Eaton, 1962, fig. 13) and seismicity (Eaton and Fraser, 1958a, b), and probably little if any deformation occurred on the flanks of the volcano.

The 1949 triangulation spanned the March-July interval. A horizontal pendulum seismometer used as a tiltmeter at the summit recorded little deformation during this time (Volcano Letter, 1949).

Virtually nothing is known about ground deformation between 1896 and 1914, for no measurements were made then. Mild summit eruption in Halemaumau Crater continued throughout most of this time. Similar, but more recent, Halemaumau eruptions have been accompanied by very little ground deformation anywhere on the volcano, particularly on the flanks, so we infer that little took place between 1896 and 1914.

BASELINE SELECTION

The selection of baselines and stations considered as stable is one of the most critical factors affecting the study of horizontal displacement. In general, we tried to select stations as far from eruptive activity as feasible. Table 4 shows the stations and baselines that we assumed to be stable. Keaau, Olaa, Kaloli, and Kaloli 2 are far removed from areas of known instability (fig. 3),

TABLE 4.—*Baselines used for deriving horizontal displacements in figure 5*

Survey interval	Stable stations	Baselines (lengths)
1896(1914)–1970	Olaa, Kaloli[1]	Olaa–Kaloli[1] (11,745.71 m).
1949–70	Kulani, Keaau, Kaloli 2, Keakapulu, Kahaha 2.	Kulani–Keaau (27,984.81 m); Keaau–Kaloli 2 (10,929.93 m).
1958–70	Kulani, Keaau, Kaloli 2 Keakapulu.	Kulani–Keaau (27,948.81 m) Keaau–Kaloli 2 (10,929.93 m).
1961–70	Kulani, Strip, Kahaha 2	Kulani–Strip (9,559.14 m).
1970–71	Kulani, Strip, Keakapulu	Kulani–Strip (9,559.14 m).

[1]Kaloli has been destroyed by wave erosion. Distances were measured to nearby Kaloli 2 and reduced to Kaloli using offset data determined by the U.S. Coast and Geodetic Survey in 1949 before the destruction of Kaloli.

and the distance between Keaau and Kaiwiki New, a station on the lower east slope of Mauna Kea, did not measurably change between 1949 and 1970. Thus the assumption of their stability is reasonable.

Kulani, on the southeast flank of Mauna Loa, is closer to the summit of Kilauea and could conceivably have been affected by the great summit subsidence in 1924 (Wilson, 1935). For this reason, Kulani was not used as a fixed point for any interval spanning 1924. In fact, however, Kulani probably was not measurably affected by the great subsidence, for its apparent displacement over the event is negligible (fig. 5*E*). Kulani did not move measurably between 1949 and 1970 because the distance to Kaiwiki New and the astronomic azimuth to Keaau are unchanged and because displacements derived using Kulani-Keaau as a baseline agree within probable error with those using Keaau-Kaloli 2.

Strip, also on the slope of Mauna Loa, was established in 1961. No unusually large structural events affected the summit of Kilauea in the following decade, and Mauna Loa was quiet, on the basis of the monitored seismicity of the volcano and leveling and geodimeter data from the summit (Hawaiian Volcano Observatory, unpub. data, 1961–73; Decker and Wright, 1968).

Kahaha 2 is on Mauna Loa and is approximately along strike of, though southwest of and possibly higher than, the known limits of the Kaoiki fault system (fig. 3), a seismically active system on the southeast flank of Mauna Loa. Displacements for nearby stations are small and geologically reasonable, assuming stability of Kahaha 2.

Keakapulu is located on the northwest edge of the Kaoiki fault system, and most distances from it had to be shortened by amounts approaching triangulation error to achieve geologically reasonable results. These adjustments suggest possible southeastward displacement of the station, although within probable survey error. In any case, the stability of Keakapulu is not critical if Kahaha 2 has remained stable.

In summary, all available information, including the consistency of our results, indicates that the base stations have remained stable within the probable limits of survey error during the 20th century.

PROCEDURE FOR TREATING TRIANGULATION DATA

We used originally observed angles, adjusted only for triangle closure to 180°, in this study. No use was made of published lengths or angles resulting from previous network adjustments, as such adjustments were made in various ways for each triangulation survey and involve different baseline lengths, numbers of stations, and other constraints. The original angles were obtained through the courtesy of B. K. Meade (U.S. Coast and Geodetic Survey data) and J. P. Church, C. R. Lloyd, and R. F. Thurston (U.S. Geological Survey data).

Each triangulation survey was compared with the 1970 trilateration survey by reference to one or more baselines (table 4). The 1970 geodimeter length of each baseline, reduced to sea level by standard techniques, is assumed to have been the length during each preceding triangulation. Using this length and originally observed angles, the triangulation was then recomputed using the law of sines. In this way, distances derived from the trilateration and triangulation surveys are directly comparable. In general, corresponding sides of triangles differ in length for the two surveys, indicating apparent displacement of one or more stations.

The ground-displacement vectors were derived by a graphical method described by Kinoshita, Swanson, and Jackson (1974). This method was selected in preference to the standard least-squares approach normally used for triangulation data because it allows visual inspection and evaluation of the displacements as they are derived. In this way geologic input could be made as to what displacements seem reasonable, and what distances should be given more weight than others on the basis of the displacements they produce. This approach seems justified, given the quality of the data. After geologic evaluation, distances were then changed by trial and error within permissible limits (table 2) to give geologically reasonable displacements. The graphical derivation was then resumed along the chain of triangles until a displacement had been determined at each station. When a geologically preferred, graphically derived length differed significantly from the original computed one, the new length was used for subsequent law of sines computations dependent on it. Finally an acceptable degree of internal consistency and geologic reasonableness was obtained for all triangles in each survey interval.

This time-consuming approach is partly subjective, but the basic data were not violated. All graphical adjustments were made within stated error limits (table 2), and most were far less than the maximum permissible amount. In essence, then, this approach is analogous to a standard least-squares adjustment with an overriding geologic monitor on each operation.

Experience has shown that the basic pattern of displacement could not be changed greatly when adjustments were kept within the limit of survey error, but that, with minor adjustments, the displacements could be better fitted to known geology. An example is shown in figure 5*E*, in which displacements derived by us are compared with displacements calculated by Lloyd (1964) using the standard least-squares approach. A similar basic pattern is shown by both displacement sets, but the progressive clockwise swing of the least-squares vectors (a pattern typically indicative of survey errors) may be eliminated by small adjustments in favor of displacement directly away from open ground cracks that formed during the survey interval, a reasonable geologic constraint. In short, the basic pattern is there regardless of how the data are manipulated, and it is with this basic pattern that we are most concerned in this paper.

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