Characteristics of Hawaiian Volcanoes
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Chapter 4

Instability of Hawaiian Volcanoes

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Abstract

Hawaiian volcanoes build long rift zones and some of the largest volcanic edifices on Earth. For the active volcanoes on the Island of Hawai‘i, the growth of these rift zones is upward and seaward and occurs through a repetitive process of decades-long buildup of a magma-system head along the rift zones, followed by rapid large-scale displacement of the seaward flank in seconds to minutes. This large-scale flank movement, which may be rapid enough to generate a large earthquake and tsunami, always causes subsidence along the coast, opening of the rift zone, and collapse of the magma-system head. If magma continues to flow into the conduit and out into the rift system, then the cycle of growth and collapse begins again. This pattern characterizes currently active Kīlauea Volcano, where periods of upward and seaward growth along rift zones were punctuated by large (>10 m) and rapid flank displacements in 1823, 1868, 1924, and 1975. At the much larger Mauna Loa volcano, rapid flank movements have occurred only twice in the past 200 years, in 1868 and 1951.

All seaward flank movement occurs along a detachment fault, or décollement, that forms within the mixture of pelagic clays and volcaniclastic deposits on the old seafloor and pushes up a bench of debris along the distal margin of the flank. The offshore uplift that builds this bench is generated by décollement slip that terminates upward into the overburden along thrust faults. Finite strain and finite strength models for volcano growth on a low-friction décollement reproduce this bench structure, as well as much of the morphology and patterns of faulting observed on the actively growing volcanoes of Mauna Loa and Kīlauea. These models show how stress is stored within growing volcano flanks, but not how rapid, potentially seismic slip is triggered along their décollements. The imbalance of forces that triggers large, rapid seaward displacement of the flank after decades of creep may result either from driving forces that change rapidly, such as magma pressure gradients; from resisting forces that rapidly diminish with slip, such as those arising from coupling of pore pressure and dilatancy within décollement sediment; or, from some interplay between driving and resisting forces that produces flank motion. Our understanding of the processes of flank motion is limited by available data, though recent studies have increased our ability to quantitatively address flank instability and associated hazards.

Introduction

The southern end of the Hawaiian-Emperor volcanic chain (fig. 1) includes some of the largest volcanoes on Earth. The sheer size of these volcanoes is enabled by the development of long rift zones that extend outward from the summits for 100 km or more (Fiske and Jackson, 1972). These rift zones grow preferentially upward and outward to the seaward side, spreading laterally on a décollement formed along the interface between the volcanic edifice and the seafloor (Swanson and others, 1976). From studies of the active volcanoes Mauna Loa and Kīlauea, we know that integral to the growth of these long rift zones is persistent seismicity within the flanks and along their décollements (Koyanagi and others, 1972), punctuated by large earthquakes and, occasionally, by catastrophic collapse as a large avalanche or landslide (Moore, 1964; Moore and others, 1989; Moore and Clague, 1992).

The evidence for catastrophic collapse is clear in maps of debris on the deep seafloor surrounding the Hawaiian Islands (figs. 1 and 2). The debris forms fanlike aprons connected to landslide and slump structures that form large cliffs (or pali) along the coasts of each island in the Hawaiian chain (Moore and others, 1989). Islands as large as O‘ahu and Moloka‘i are inferred to have been entirely dissected, with huge landslide scars forming the prominent pali for which the islands are justly famous. The debris scattered across thousands of square kilometers of ocean floor are the missing portions of these islands, and some debris consists of subaerially deposited

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lava (Moore and others 1994a,b). Dating of this seafloor debris, as well as dissected lava flows on the islands, provides constraints for studies that show that these landslides occur episodically during growth of Hawaiian volcanoes and rift zones (Moore and others, 1989), with the largest landslides reserved for the latter stages of growth (Moore and Clague, 1992; Clague and Sherrod, this volume, chap. 3).

Indications of flank instability on active Hawaiian volcanoes come from both geologic structures and historical observations. On the Island of Hawai‘i, large, rapid flank movements (often occurring with large earthquakes) were observed four times during the 19th and 20th centuries, each spaced about 50 years apart. The most spectacular examples are the great 1868 Ka‘ū earthquakes, which opened rift zones on Kīlauea and Mauna Loa and moved the southeast flanks of both Kīlauea and Mauna Loa seaward in two great earthquakes spaced days apart (Wyss, 1988). Large flank movements occurred again in 1924 and 1975, preceded by decades of slow growth of summit and rift magma systems (Swanson and others, 1976). Current observations of gradual seaward flank growth on Kīlauea support the interpretation that the East Rift Zone (ERZ) episodically opens as the flank is gradually compressed by deep magma and cumulate intrusion and that this compression also triggers décollement slip, which propagates the south flank seaward (Montgomery-Brown and others, 2011). Similar patterns of summit and rift growth, followed by flank spreading, are displayed by detailed models of progressive volcano growth. These models simulate many of the geologic structures observed on the south flank of Kīlauea, but they do not detail the triggers of rapid flank movement or wholesale flank collapse. Our understanding of the mechanics of rapid flank movements and what drives them is incomplete.

What conditions result in flank instability, or an imbalance between forces driving seaward flank motion, relative to forces resisting flank motion, as islands grow? Is rift-generated flank instability the cause of landslides large enough to dissect entire islands? How can large landslides or avalanches occur catastrophically when subaerial slopes are moderate and the seafloor actually slopes inward towards the center of each island as the weight of the island depresses the oceanic lithosphere? These questions have motivated research into Hawaiian Island stability since discovery of large landslide deposits on the seafloor more than 20 years ago (Moore and others, 1989), but, despite exhaustive studies, remain largely unanswered. These questions are important for hazards, as well as research, because we know that active volcanoes on the Island of Hawai‘i are deforming and potentially could one day produce a landslide large enough to dissect the Island of Hawai‘i. However, we do not know what to look for, or where we are in the evolution of what appears to be, given abundant evidence for flank failure along the Hawaiian chain, a characteristic and ubiquitous pattern of volcano growth and decay (Clague and Sherrod, this volume, chap. 3).

We show, in this chapter, that the observed pattern of decades-long subaerial flank compression and uplift, followed by rapid flank extension and summit subsidence, is part of a systemic pattern of growth of Hawaiian volcanoes. We investigate these growth processes by drawing on onshore and offshore field

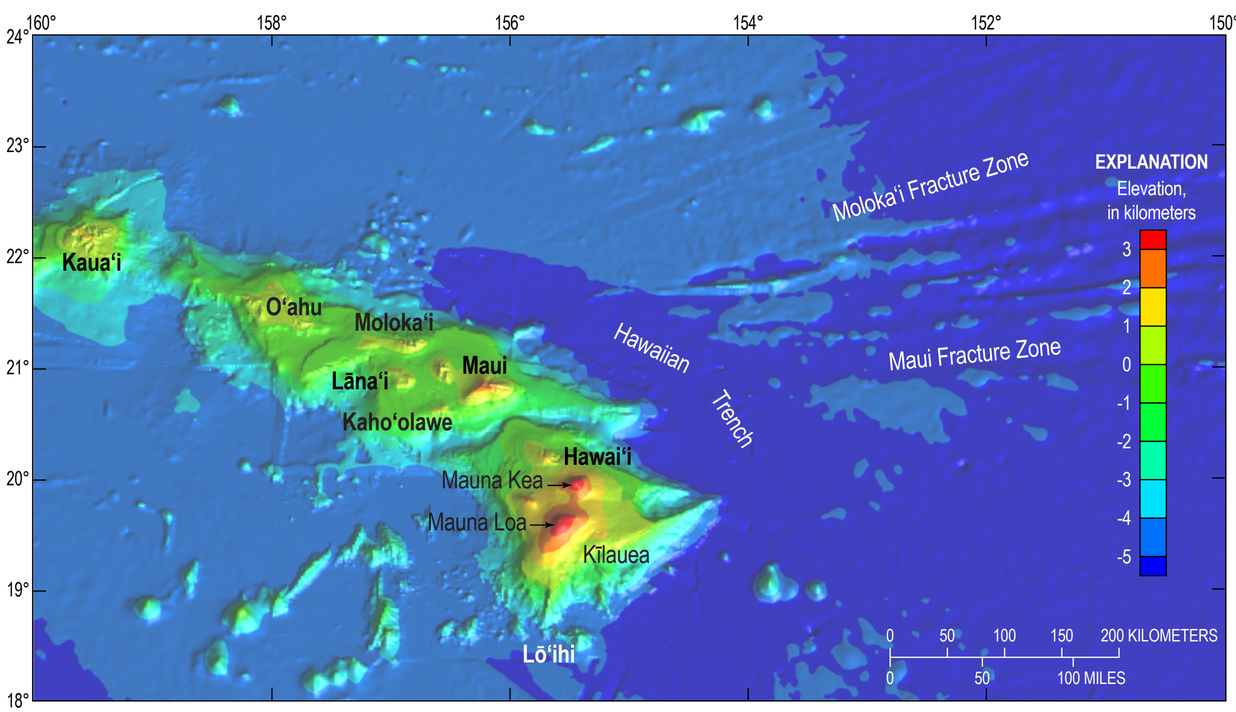


Figure 1. Bathymetric map of the Hawaiian Islands, showing debris scattered around pediments on which islands sit. All islands have significant aprons of landslide debris associated with previously active volcanoes. Significant deposits are also associated with currently active Mauna Loa volcano on the Island of Hawai‘i.

studies of Kīlauea and Mauna Loa, mechanical models that are analogs for flank motion, and sophisticated models for volcano growth, flank failure, and décollement resistance. We find that the processes of volcano growth migrate rift-zone flanks seaward and are associated with a cyclic evolution of flank instability. The mechanisms of seaward flank migration involve all of the basic structural elements of Hawaiian volcanoes and provide clues to promising paths for future research.

Geologic Evidence for Flank Instability

Subaerial Evidence

Although evidence that the flanks of Hawaiian volcanoes could be unstable and collapse catastrophically was present in subaerial morphology and structures (that is, fault-bounded steep cliffs, or pali, along the coasts), the significance of these structures was not fully appreciated until the late 20th century. In their prescient model for motion of the south flank of Kīlauea, Moore and Krivoy (1964) (1) attributed the ERZ eruptive fissures formed during the 1962 eruption to be a continuation of a listric Koaʻe Fault System, (2) identified antithetic faulting in the Koaʻe system, (3) suggested that the ERZ was fed by the same source of magma that fed the summit, and (4) went on to suggest that the Hilina Fault System is also a seaward-dipping listric fault. They interpreted all seaward-facing subaerial flank scarps as listric faults connected to a décollement under the south flank at depths well below sea level. They further suggested that southward, gravity-driven flank motion opened the ERZ and allowed it to be fed from summit magma chambers. Although not all of these assertions have withstood continued study, many aspects of their model are contained in modern concepts of south-flank deformation (fig. 3).

The concept of a magmatically driven rift and flank was developed by Swanson and others (1976) using trilateration, triangulation, and spirit leveling to constrain compression and uplift of the south flank of Kīlauea between the summit and the coast, using geodetic data collected from 1896 to 1975. They then correlated these changes to observed changes in the magmatic system of Kīlauea. Between 1924 and 1968, the lava level in Halemaʻumaʻu Crater, within Kīlauea's summit caldera, gradually increased from 600 to 100 m below the rim (Macdonald and others, 1983). The high lava level was associated with a voluminous eruptive output at Kīlauea in the 1960s and early 1970s, including an active lava lake at the summit during 1967–68, frequent summit eruptions, eruptions combined with flank movement that opened both the Southwest

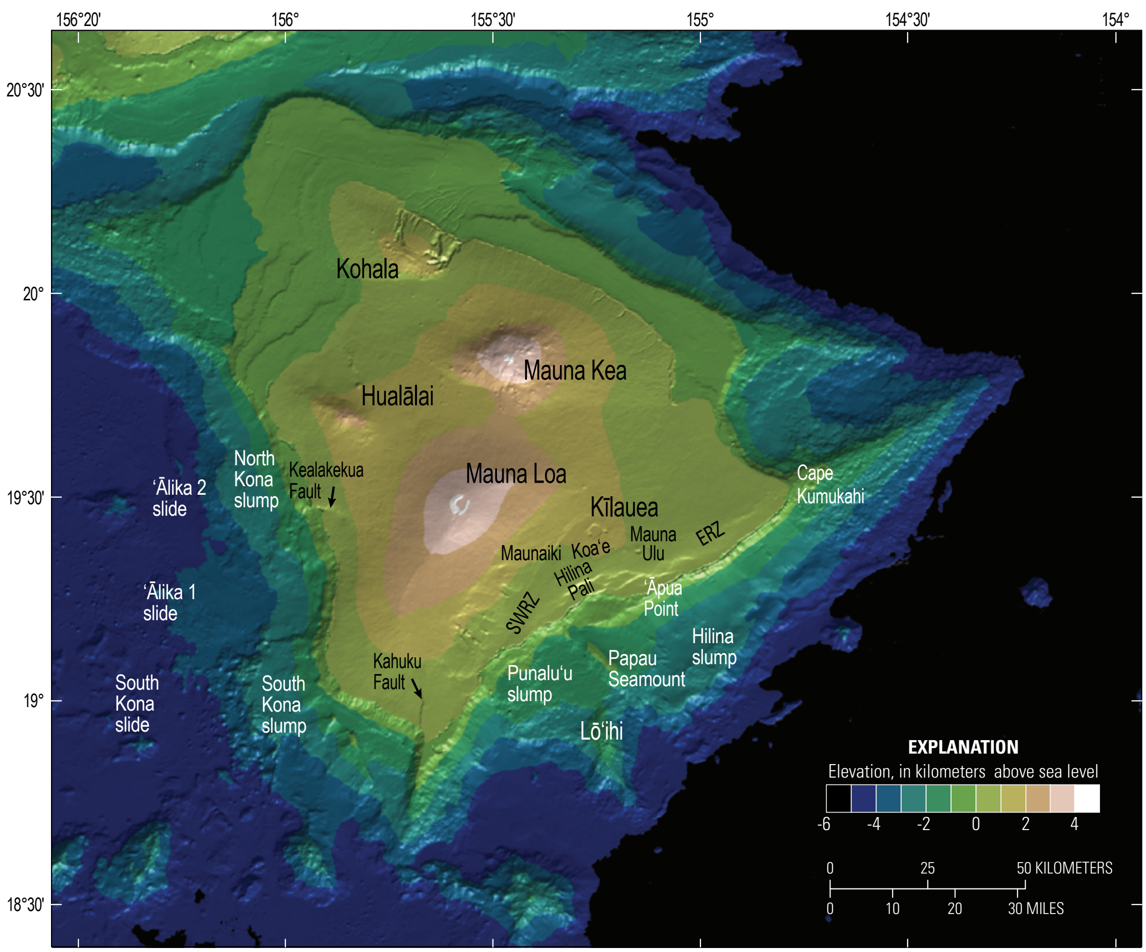


Figure 2. Map showing the Island of Hawaiʻi, surrounding seafloor, structures off the south flank of Kīlauea, and locations of features indicated in the text. ERZ, East Rift Zone; SWRZ, Southwest Rift Zone.

Rift Zone (SWRZ) and ERZ, growth of the Mauna Ulu eruptive vent on the ERZ during 1969–74, and rates of summit and rift-zone extension unseen today (Swanson and others, 1976). This led Swanson and others (1976) to predict that a large flank movement would occur, and their intuition was rewarded when their paper had passed review but was still in press. Repeating a pattern established during the previous two centuries, the elevated level of summit eruptive activity and extrusive growth ended abruptly in late 1975 with a great earthquake and flank movement. The 1975 Kalapana earthquake (*M*7.7; Nettles and Ekström, 2004) produced a maximum of 10 m of seaward flank motion (Lipman and others, 1985) and a tsunami (Ma and others, 1999). In the next 7.2 years, until the eruption of Puʻu ʻŌʻō in January 1983, only 3 small eruptive events occurred, whereas 14 intrusive events occurred at the summit and along the ERZ. This ratio of eruptive to intrusive events was the inverse of that established in the decades-long increase of magma-system head in the summit region (Macdonald and others, 1983; Dzurisin and others, 1984), demonstrating that a significant change in behavior had occurred.

On Kīlauea, this sequence of gradual growth and inflation of the summit, followed by rapid flank movement that collapses the summit, opens the rift zones, and drains the magma system, is a pattern characterizing large, rapid flank movements in 1823, 1868, 1924, and 1975 (Swanson and others, 1976; Macdonald and others, 1983) and probably also occurred on Kīlauea sometime between 1750 and 1790 (Macdonald and others, 1983). The great 1868 *M*7.1 and *M*7.9 earthquakes and associated flank movements affected both Kīlauea and Mauna Loa, opened both rift zones on Kīlauea and the SWRZ of Mauna Loa, and produced coastal extension and subsidence from Punaluʻu to Cape Kumukahi, as well as a tsunami that killed 46 people along the island's southeast coast near Punaluʻu (Macdonald and others, 1983). The 1924 flank movement on Kīlauea was associated with abundant seismicity and strong ground motion that progressed rapidly from the summit eastward, along the ERZ to Cape Kumukahi, without a large earthquake (Macdonald and others, 1983) but caused extension across the entire ERZ, as well as coastal subsidence. Subsidence of Cape Kumukahi and encroachment of the sea, a kilometer up the rift, combined with the 60° dip of the faults flanking the opening of a 1-km-wide graben at Cape Kumukahi, suggests that about 4.6 m of seaward (southward) migration of the south flank of Kīlauea occurred at that time. With the opening of this graben and the ERZ, the level of lava at Kīlauea's summit dropped more than 500 m, well below the water table, resulting in violent phreatic eruptions several weeks later in May 1924 (Swanson and others, 2012). It was not until the 1960s that subsequent summit growth regained the magma levels last seen in 1924 (Swanson and others, 1976), and summit eruptions again became commonplace. As in 1924, high magma levels and extrusive activity did not last, as rapid extension of the flank associated with the 1975 *M*7.2 earthquake dropped the magma level again and ushered in a period of intrusive growth.

Submarine Evidence

The realization in 1975 that previously identified landslide-like fault scarps on the flanks of Hawaiian volcanoes could slip suddenly during large flank movements, first proposed by Moore (1964), helped to motivate mapping of the seafloor surrounding the islands. In fall 1988, side-scan GLORIA (Geologic LOng-Range Inclined Asdic) surveys were completed in a zone extending from south of the Island of Hawaiʻi to north of Kauaʻi. In this region, deposits of 17 well-defined large landslides were identified (Moore and others, 1989), with many overlapping adjacent deposits. Of particular note, the surveys near Oʻahu and Molokaʻi revealed deposits from some of the largest landslides on Earth. Sampling of these deposits indicated that landslide debris had moved as far as 200 km from its subaerial source, had crossed a trough in the Hawaiian Trench (the depression of the oceanic plate as it sags under the weight of the islands), and had traveled upslope on the opposite side. In other slide deposits, curved scars in ocean-floor sediments showed the track of slide debris. In a landslide track off the west coast of the Island of Hawaiʻi, the track curves and points

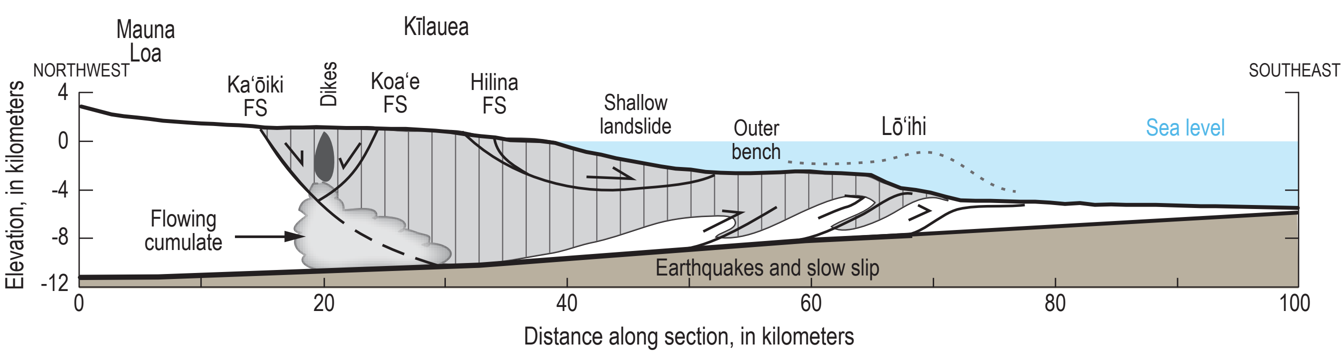


Figure 3. Cross section diagram showing interpreted internal structure for the south flank of Kīlauea Volcano, modified from Morgan (2006). High-angle normal faults underlie summit region and accommodate subsidence and axial extension. Intrusion and subsidence along rift axis and seaward flow of ductile cumulates precipitated from magma chambers help to drive unbuttressed flank seaward. Stress transfer from the magma system to the distal end occurs through flank slip along a basal décollement, producing earthquakes and slow slip events. This basal slip terminates by thrust faulting, building an overthrust structure within the outer bench. Shallow faults on seaward slopes define surficial slump features, such as the Hilina Fault System.

toward Lānaʻi and Kahoʻolawe, where cobbles of coral reef are found stranded hundreds of meters above sea level (Moore and Moore, 1984). This and other evidence are explained by catastrophic failure of the flanks of Hawaiian volcanoes in which the momentum of landslide debris not only carries it far from the volcano, but also generates tsunamis large enough to break up coral reefs and deposit fragments above sea level (Moore and others, 1992). The GLORIA surveys show that both submarine and subaerial evidence are required to interpret the genesis of the morphology and structure of Hawaiian volcanoes.

Beginning in the 1990s, various studies of the submarine flanks of the Hawaiian Islands gradually tested and refined previous interpretations for volcano-flank deformation and evolution. Submersible surveys and dredge hauls of submarine landslide blocks west of the Island of Hawaiʻi confirmed the subaerial origin of these blocks and provided the first age constraints on the deposits (Moore and others, 1995). Seismic surveys reported by Smith and others (1999) revealed the presence of thick, landward-tilted sedimentary deposits on Kīlauea's offshore bench, consistent with thrust faulting at the toe of the flank. High-resolution bathymetric data collected by the University of Hawaiʻi, Mānoa, around the Island of Hawaiʻi, clarified the morphology of the submarine flanks (Moore and Chadwick, 1995), leading to morphotectonic interpretations that helped extend surficial structures offshore (for example, the Punaluʻu slide, Hilina slump, and South Kona landslide complex); however, these interpretations were limited by a lack of subsurface imaging, as well as few direct samples. This situation changed in the late 1990s as a result of a comprehensive multichannel seismic reflection survey offshore the south flanks of Kīlauea and Mauna Loa (Morgan and others, 2000, 2003a,b; Hills and others, 2002), seafloor mapping surveys carried out by the Japan Agency for Marine-Earth Science and Technology (Smith and others, 2002; Eakins and others, 2003), and multiple manned and unmanned submersible dives around all of the islands (Lipman and others, 1988; Lipman and others, 2000; Naka and others, 2000; Takahashi and others, 2002; Coombs and others, 2004; Morgan and others, 2007).

The morphologic features revealed by the new seafloor maps clarified the distribution and geometry of landslides around the islands, leading to reinterpretations of the nature and origin of several landslides previously based primarily on GLORIA imagery (Takahashi and others, 2002). In particular, the large Nuʻuanu and Wailau landslides, which broke away from the north flanks of Oʻahu and Molokaʻi, respectively, and the South Kona landslide complex, west of Mauna Loa, were shown to consist of coherent, dispersed megablocks that could be reconstructed back onto adjacent, broken flanks (Moore and Clague, 2002; Yokose and Lipman, 2004). The more proximal portions of the deformed flanks exhibited broad benches similar to the bench offshore from Kīlauea.

These characteristics reaffirmed, for several different settings, the flank model proposed by Denlinger and Okubo (1995). In their model, the south flank of Kīlauea moves seaward by gradual, long-term creep that builds broad submarine benches braced by imbricate thrusts that build an outer bench along the distal end. They proposed that formation of this structure is accompanied by intermittent seismicity and is occasionally punctuated by rapid catastrophic slip of the seaward flank that generates large earthquakes.

This general picture was further confirmed by seismic-reflection profiles obtained along the south flanks of Kīlauea and Mauna Loa and the north flank of Oʻahu (Moore and others, 1997; Morgan and others, 1998, 2010; Hills and others, 2002). All of these volcano flanks are underlain by a deep reflective surface, interpreted to define a décollement coincident with the top of the oceanic crust (ten Brink, 1987; Morgan and others, 2000; Leslie and others, 2004). The best imaging comes from Kīlauea's south flank, which exhibits seaward-vergent thrust faults rising beneath the prominent midslope bench (Morgan and others, 2000, 2003a). Evidence that the broad benches have undergone progressive uplift and rotation indicates that they are probably large overthrust structures (Smith and others, 1999; Morgan and others, 2000; Hills and others, 2002). These structures have accommodated tens of kilometers of lateral displacement (Morgan and others, 2000, 2003a). Thus, Kīlauea's offshore bench has formed slowly, while accumulating a thick infill of sediment in the closed midslope basin formed by extension on its landward side (fig. 4). This model is more consistent with contraction at the toe of the volcano flank in response to its outward displacement above a continuous décollement (fig. 4), as originally suggested by Denlinger and Okubo (1995), and less compatible with more modest displacements proposed by models for gravitational slumping. A similar model may hold for several benches that compose the Kona landslide complex along the west flank of Mauna Loa (Morgan and Clague, 2003; Morgan and others, 2007), although controversy persists, owing to the lack of high-quality seismic-reflection data and sparse geochemical and stratigraphic data (Lipman and others, 2006).

New observations also confirm that the offshore benches at Kīlauea and elsewhere are constructed largely of fragmental debris shed as lava deltas build and then break off at the shoreline (Moore and Chadwick, 1995; Moore and others, 1995; Clague and Moore, 2002; Clague and others, 2002; Naka and others, 2002; Morgan and others, 2007). Submersible surveys of the outer benches revealed thick accumulations of indurated volcaniclastic materials composed of subaerially derived fragmental lavas, shallow landslide debris, or both. Seismic-reflection profiles over Kīlauea's midslope basin and upper flanks revealed buried normal faults and intensely folded layering—a juxtaposition of normal and reverse faulting that may denote past landslide structures or reveal complex deformation within the currently mobile flank (fig. 4; Hills and others, 2002; Morgan and others, 2003b). Volcaniclastic deposits of mixed compositions and diagenetic states make up the incised bench of Mauna Loa's Kona landslide complex, suggesting a long history of accumulation and burial, followed by exhumation and uplift (Morgan and others, 2007). Mauna Loa's upper submarine slopes, in contrast, are commonly composed of subaerially derived lava flows, as well as volcaniclastic deposits (Garcia and Davis, 2001), forming an

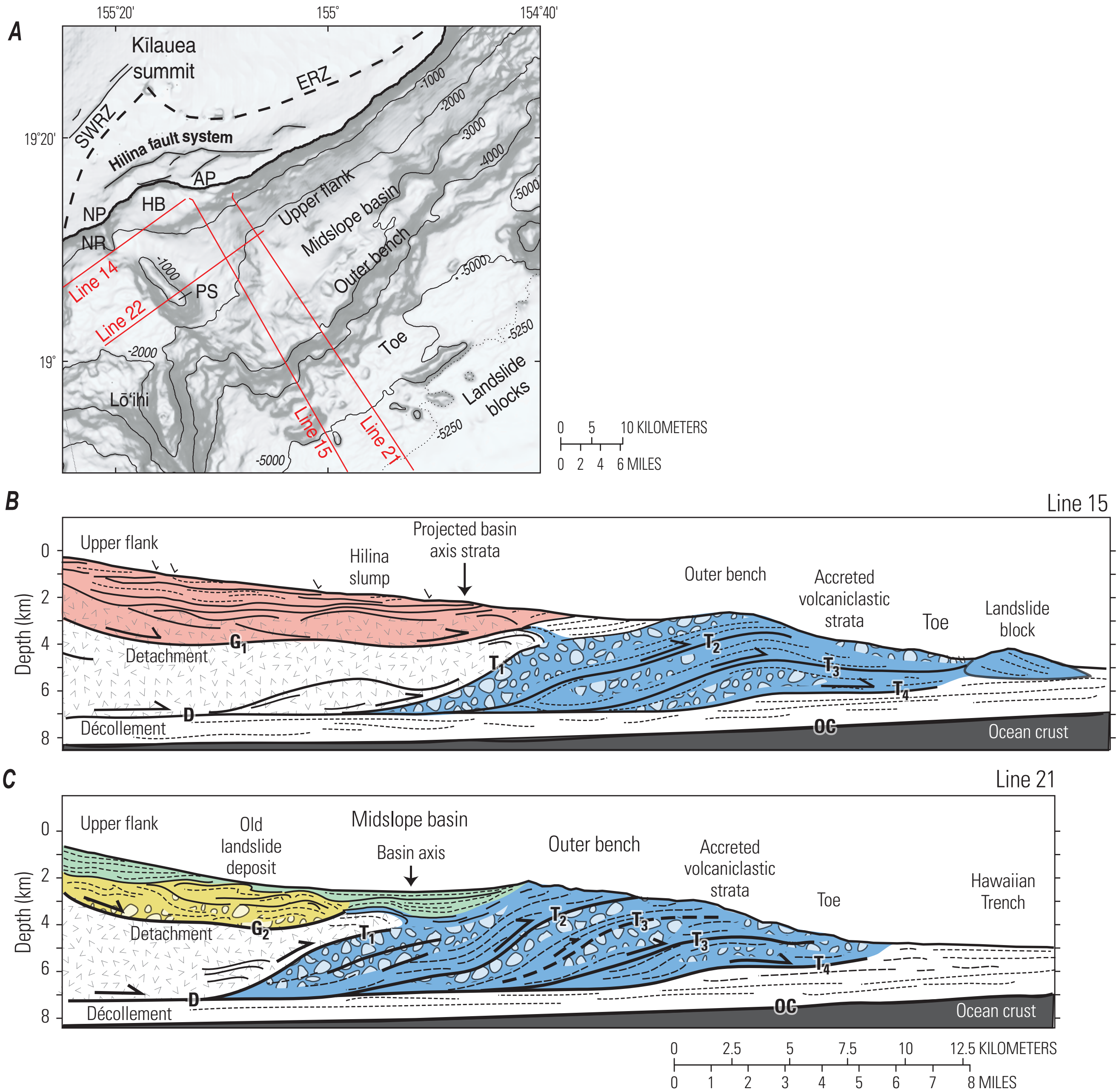
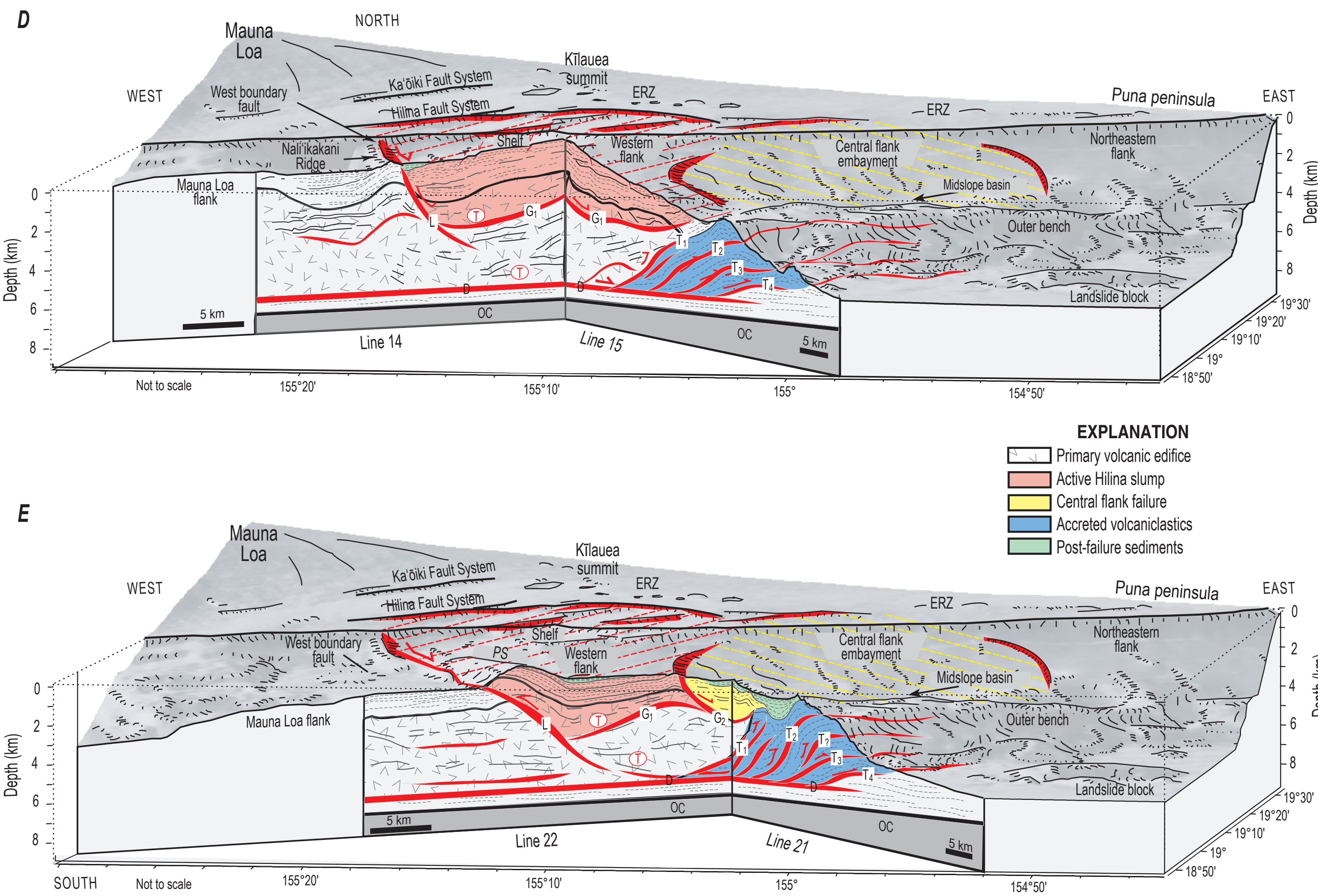


Figure 4. Maps and diagrams showing structural interpretations of Kīlauea's south flank, based on seismic-reflection profiles (compiled from Morgan and others, 2003a). *A*, Shaded slope and bathymetric map of south flank of Kīlauea Volcano, showing locations of seismic-reflection lines 14, 15, 21, and 22 from 1998 survey. Southwest edge of Kīlauea's mobile flank is bounded by an offshore lineament defined by ridges and scarps, including Papau Seamount (PS) and Naliʻikakani Ridge (NR). Upper part of submerged flank is marked by two embayments: Halapē Bay (HB) shoreline reentrant, between Naliʻikakani Point (NP) and ʻĀpua Point (AP); and a central flank embayment above a midslope basin. A broad outer bench fronts the midslope basin, which is partly filled with volcaniclastic sediment. Steep slope of outer bench is incised by several arcuate scarps. Bathymetry gridded at 1,000 m from Smith and others (1994). *B*, *C*, Interpreted depth sections for dip-parallel seismic-reflection lines 15 and line 21. Reflections: D, décollement; G, internal glide plane; OC, top of oceanic crust; T, thrust fault. Transects show contrasting structure: line 15 shows the more coherent Hilina slump (pink), whereas disrupted strata (yellow) underlie bedding-parallel slope and basin sediment (green) on line 21. Blue shows common imbricated stack of accreted volcaniclastic debris. *D*, *E*, Cutaway views through Kīlauea's south flank (looking north) showing subsurface structures compiled from seismic lines mapped in part *A*. Intersection of lines 14 and 15 (part *D*) reveals structure of west flank, detachment G_1, and the Hilina slump (pink). Intersection of lines 22 and 21 (part *E*) shows uplift and westward thrusting of Papau Seamount (PS) due to oblique convergence of the Hilina slump on western boundary fault. Transition to region of central flank failure (yellow) is marked by an arcuate scarp at seafloor and listric G_2 detachment at depth. Imbricate thrust sheets within outer bench (blue) front central flank embayment with ponding sediment (green) within midslope basin. D, décollement; ERZ, East Rift Zone; L, left-hand boundary and western boundary fault; OC, top of oceanic crust; SWRZ, Southwest Rift Zone; T, thrust fault.

interbedded package that appears to drape the distal offshore bench (Morgan and Clague, 2003; Morgan and others, 2007).

An integrated view of the evolution of the mobile flanks of Hawaiian volcanoes, best exemplified by Kīlauea's south flank, implies a complex but repeatable history (Denlinger and Okubo, 1995; Morgan and others, 2003a,b). Debris derived from volcano growth is shed into a moat along the volcanoes' seaward flank, and then pushed seaward by continued flank motion. Flank motion plows the sediment, shortening and uplifting a sediment pile along the distal margin through formation of thrust faults that originate wherever décollement slip terminates. It is this process that builds an outer bench. Flank motion through décollement slip is very efficient and extends the flank more rapidly than sediment can accumulate; thus, a basin forms behind the bench within which subaerially derived volcaniclastic sediment is trapped before it is shoved seaward and deformed, recording a history of deposition and deformation that can be read in seismic-reflection profiles and stratigraphic records (fig. 4).

This general model for submarine flank deformation and evolution may extrapolate well to other volcanoes with less complete datasets than Kīlauea. Mauna Loa's west flank, in particular, has been the focus of intense study (Lipman and others, 1988; Garcia and Davis, 2001; Morgan and Clague, 2003; Yokose and others, 2004) in efforts to constrain the origin and relative timing of multiple interpreted landslide deposits (for example, the South Kona and ʻĀlika landslides, fig. 2). The presence of subaerially derived lava flows deep on the submarine flanks of western Mauna Loa and the radiometric ages of subaerial lavas point to phasing of flank activity similar to that at Kīlauea (Morgan and others, 2007). In particular, it is hypothesized that a large-scale subaerial to submarine flank failure contributed the large blocks that make up the South Kona debris field (fig. 2), preconditioning Mauna Loa's west flank to spread outward and bulldozing the newly deposited debris to build the South Kona bench. This bench was subsequently breached by a smaller landslide, leaving behind a narrow scar and debris track that leads to the lobate ʻĀlika 2 debris field to the northwest of the source region. A similar sequence of events may have resulted in the present benchlike morphologies of Oʻahu and Molokaʻi's north and south flanks (Nuʻuanu and Wailau structures, respectively; Moore and others, 1997). Limited sampling of these features during submersible dives suggests that they are also composed



of heterogeneous volcaniclastic materials (Clague and others, 2002) that may have been reworked during flank spreading and now sit high on submarine benches. Both Nuʻuanu and Wailau may have been subjected to subsequent breakup and local slumping to form the blocky debris-avalanche deposits, similar to, but on a much larger scale than, the late-stage incision of Mauna Loa's western submarine bench.

Geophysical Evidence for Flank Instability

Seismic Studies

The first comprehensive discussion of the seismicity of Kīlauea and Mauna Loa volcanoes by Klein and others (1987) segregated the seismicity of these active volcanoes into three parts, based upon the locations of earthquakes in the edifice. Shallow (<5 km below ground level [bgl]) seismicity is associated with summit and rift-zone eruptions and shallow intrusions (fig. 5*A*), deeper (>5 km and < 10 km bgl) seismicity is associated with magma storage and flank deformation (fig. 5*B*), and very deep (15–40 km bgl) seismicity is associated with magma supply through the mantle and oceanic crust to the volcanic edifice (Wright and Klein, 2006). Seismicity produced by flank deformation in the depth range of 5–10 km bgl within Kīlauea and Mauna Loa volcanoes is offset from their rift zones (Wolfe and others, 2007). As shown in figure 5*B*, this seismicity within the flank of Kīlauea is distributed 2 to 8 km away from the ERZ. This distributed activity results from both internal deformation of the flank (including movement along fault systems forming scarps, such as the Hilina Pali), as well as slip along a décollement defining the base of the mobile south flank of the volcano (Delaney and others, 1998).

Shallow (<5 km bgl) seismicity along the summits and rift zones of Mauna Loa and Kīlauea volcanoes is associated with shallow intrusive or eruptive activity (Klein and others, 1987). In contrast to deeper flank seismicity, interpretation of the causes of this seismicity is usually unequivocal because the occurrence of these events at the summit or along the rift zones is directly associated with eruptive or shallow-intrusive activity that cracks and faults the surface. The effect of the opening of a rift zone at shallow (<5 km bgl) levels on stability or deformation of the adjacent flank, however, is less clear (Wolfe and others, 2007), as studies of long-term deformation and seismicity at Kīlauea demonstrate (Delaney and others, 1998; Owen and others, 2000).

There is a complex interplay between the opening of rift zones and south flank deformation that was first interpreted on the basis of seismic observations (Koyanagi and others, 1972), before deformation could be measured on short time scales (Swanson and others, 1976). Those authors noticed that rift-zone intrusions or eruptions were commonly associated with deeper flank seismicity, though their speculation as to cause and effect was hampered by lack of knowledge of the detailed deformation field that accompanied these earthquakes. More detailed studies of seismicity in the past decade (Brooks and others, 2006; Wolfe and others, 2007) have confirmed earlier suspicions of links between the rift zone and flank seismicity (Dvorak and others, 1986), but the addition of detailed deformation measurements and combined modeling of seismicity and deformation are required to adequately address this coupling.

Deformation Studies

Macdonald and Eaton (1957) realized that significant ground deformation accompanied the movement of magma within Kīlauea Volcano, and both Stearns and Clark (1930) and Stearns and Macdonald (1946) recognized that the Hilina Fault System on the flank of Kīlauea resembled the headwall of a large landslide. Other studies have constrained coastal deformation associated with eruptions or with earthquakes. Following the great 1868 Kāʻu earthquakes, Titus Coan (Brigham, 1909) reported that the entire coast from Cape Kumukahi to ʻĀpua Point subsided 1 to 2 m and that the volumetric eruptive output from Mauna Loa was halved. Additionally, aside from a brief summit eruption in 1877 and a persistent lava lake in Halemaʻumaʻu, Kīlauea also stopped erupting until 1919—repeating a pattern established in the 19th century. However, the true significance of the intriguing landslide-like fault structures along Kīlauea's south flank that contributed to this subsidence, and the relation between flank movement and volcanic activity, were unrecognized. Distributed flank deformation had occurred on such a vast scale (tens to hundreds of kilometers of coastline, involving much of the south half of the Island of Hawaiʻi) that it was impossible, with existing measurement capabilities in the 19th and early 20th centuries, to detect any links between coastal warping and nearby volcanic activity. Consequently, the relation between large flank motions and great earthquakes went unrecognized until 1975.

The beginning of geodetic constraints on both volcano and flank deformation at Kīlauea began with regional geodetic-control networks set up in the late 19th and early 20th centuries to connect Hilo to the summit of Kīlauea, to Pāhoa (on Kīlauea's lower ERZ), and to the south coast to survey for road construction and property boundaries (Swanson and others, 1976). These surveys were initially conducted by the U.S. Coast and Geodetic Survey, but beginning in the 1950s, the U.S. Geological Survey began reoccupying and extending these geodetic networks for use in volcano monitoring. Despite the short time scale (50 years) of comprehensive long-term monitoring and the coarse standards by which changes in distances could be measured with triangulation (at best, 1 part in 10^5), the huge displacement rates of the south flank resulted in large signal-to-noise ratios and revealed a wealth of information regarding variations

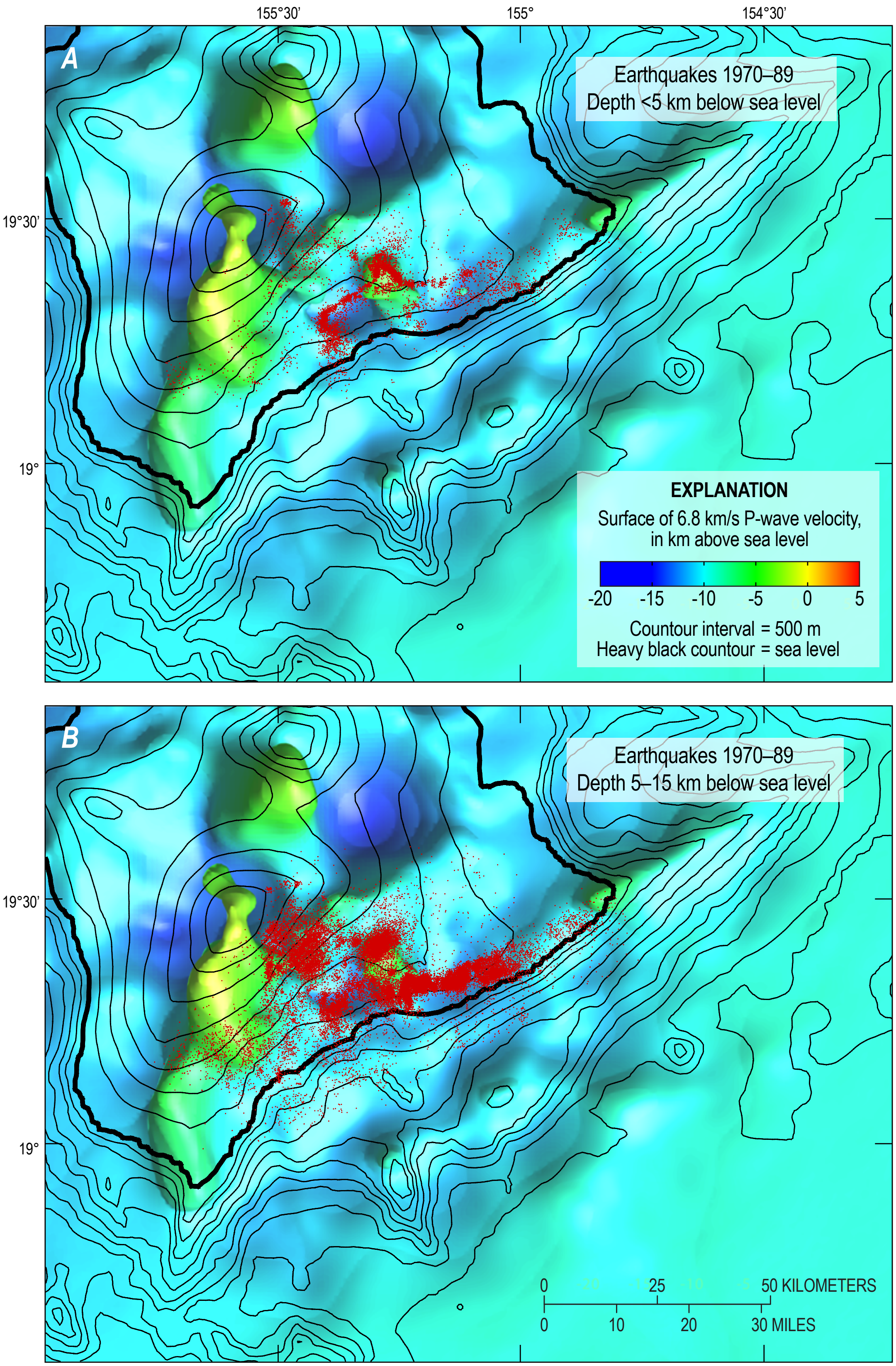


Figure 5. Contour and seismicity maps of the Island of Hawai'i. *A*, Shallow seismicity (<5 km below sea level [bsl]) at Kīlauea Volcano summit aligns with trends of Southwest and East Rift Zones and is associated with extension and eruptions from 1970 to 1989. Earthquakes (red dots) are commonly associated with surface faulting connected with opening of summit and rift zones during extension associated with shallow dike emplacement. Also shown is top surface of internal P-wave velocities >6.8 km/s (from model of Park and others, 2009). Where this surface intrudes volcanic edifice, it is most likely composed of dense dunite accumulated during volcano growth. Areas of high gravity (see discussion below and figure 8) that lack high P-wave velocities are hot. Summit of Kīlauea is also hot, but the volume of accumulated dunite is apparently large enough to create a P-wave-velocity anomaly. *B*, Same P-wave distribution as in figure 5*A* but showing deeper (>5 km bsl) seismicity associated with seaward flank movement. Most of this seismicity is associated with sliding of flank on a décollement between 8- and 10-km depth and is near or within this décollement. Horizontal separations between shallow seismicity in part *A* and deep seismicity here are significant because in gap between them, observations of dike opening and flank movement show that flank moves easily on its décollement (Montgomery-Brown and others, 2011). Stress transmission apparently occurs easily from rift zone to crest of pali overlying décollement seismicity.

in flank movement during the 20th century (Swanson and others, 1976; Owen and Bürgmann, 2006). Application of reasonable geologic constraints to network reduction produces a general pattern of flank motion (Swanson and others, 1976; Denlinger and Okubo, 1995; Owen and Bürgmann, 2006) that is remarkably consistent with that obtained since 1990, using sophisticated satellite geodesy (Owen and others, 2000).

These studies broadly constrain distinct regions of deformation on Kīlauea. The volcano is buttressed to the west-southwest by Mauna Loa. Although the flank north of the ERZ is pushed northward episodically during shallow rift-zone intrusions (for example, Montgomery-Brown and others, 2010, 2011), the flank generally moves southeastward and seaward from the summit and from the ERZ to the coast. The summit and ERZ, though subject to ephemeral eruptive activity, have undergone average long-term subsidence and extension since 1975 (Delaney and others, 1998). This extension is associated with seaward movement of the remainder of the south flank of the volcano, accommodated by slip along a décollement underlying the south flank and its offshore bench (Owen and Bürgmann, 2006).

As the décollement slips, the summit and ERZ extend. The summit occupies a roughly triangular region that has been subsiding and extending for most of the past century (Swanson and others, 1976), with only a few brief periods of uplift associated with rapid summit growth. Extension and subsidence abruptly terminate along an eastern margin between Chain of Craters Road and Escape Road, where the ERZ trends east from the summit to Mauna Ulu. The ERZ at Mauna Ulu, if extended westward to the SWRZ near Maunaiki, would form the south boundary of extension and subsidence within the largely antithetic (north-dipping) Koaʻe Fault System (fig. 2; Duffield, 1975). Extension and subsidence of the summit region thus continue seamlessly southward into the Koaʻe Fault System and southeastward into the ERZ (Duffield, 1975), forming a continuously extending region whose south boundary extends from Maunaiki (SWRZ) through Mauna Ulu (ERZ) to Cape Kumukahi (ERZ) (see Poland and others, this volume, chap. 5).

Despite extension along its crest, growth of the ERZ is upward through accumulation of lava and seaward through flank displacement and rift extension (fig. 6). This one-sided growth pattern, first proposed by Swanson and others (1976), is likely to be the generic pattern for many ocean-flanked rifts on Hawaiian volcanoes (Leslie and others, 2004) undergoing seaward migration of the rift system over time (fig. 6*B*). Here, the "ridge-push" mechanism for seaward motion of the flank (Swanson and others, 1976) is an integral part of Kīlauea deformation: ERZ dikes wedge the flank seaward, compressing it and stimulating slip on a décollement near the base of the edifice at ~8–10-km depth (Dieterich and others, 2003; Syracuse and others, 2010; Montgomery-Brown and others, 2011), as illustrated in figure 6. Using deformation monitoring to estimate strain and then comparing the strain with concurrent seismicity, Montgomery-Brown and others (2009) showed that the link between rift opening and dike injection and flank movement by slip along a décollement is complex and not precisely predictable on the basis of time or

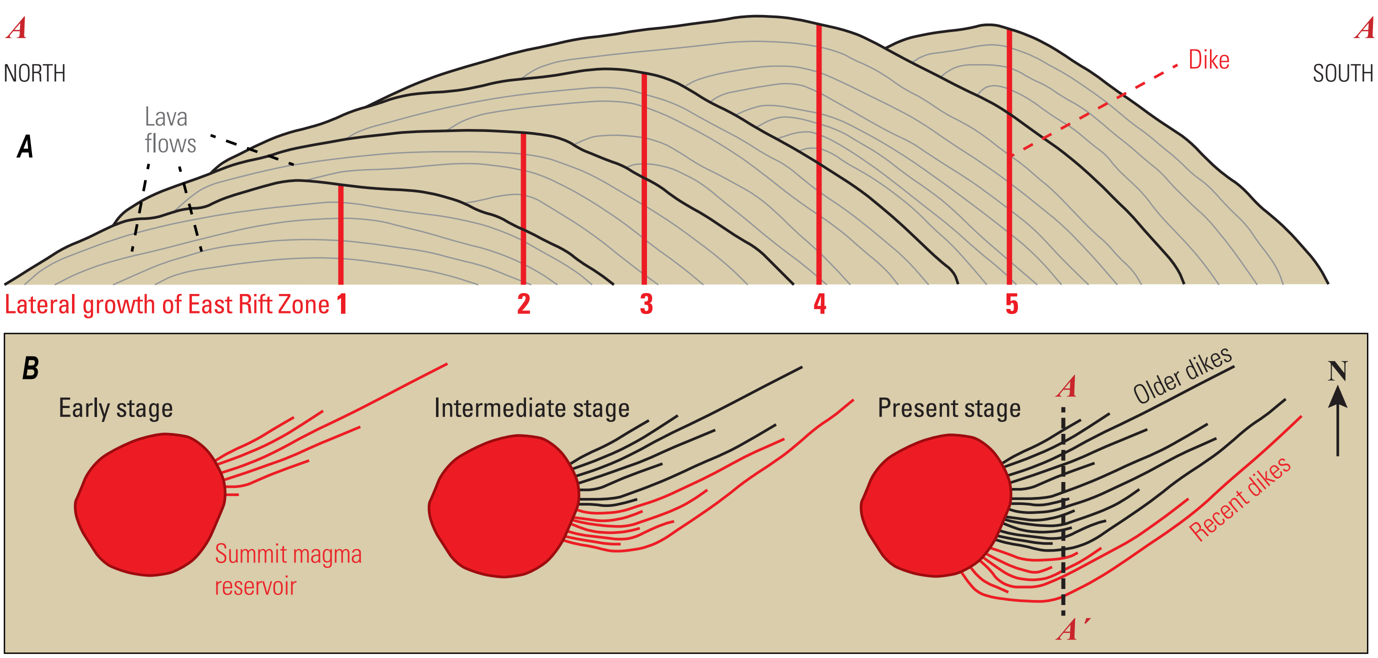


Figure 6. Schematic diagrams of rift zone/flank movement, from Swanson and others (1976). *A*, Cross section showing upward and outward growth of Hawaiian rift zones. This model and repeated measurements of flank deformation led to forecast of 1975 *M*7.7 Kalapana earthquake. *B*, Map view showing how migration of a rift zone seaward, relative to its source at summit, leads to bending of rift zone, as illustrated here and by East Rift Zone of Kīlauea.

average slip, and illustrates the complexity of décollement slip mechanisms.

Deformation of the entire flank south of the Koa‘e Fault System and ERZ undergoes alternating contraction and extension in conjunction with slip triggered on patches of Kīlauea’s south-flank décollement. Décollement slip occurs with large flank earthquakes, such as the 1975 Kalapana event (for example, Tilling, 1976; Lipman and others, 1985), during persistent microseismicity (for example, Got and others, 1994; Wolfe and others, 2004, 2007), and during transient slow slip events (Brooks and others, 2006; Montgomery-Brown and others, 2009, 2010). The association of décollement slip with dike-enhanced rift-zone extension (Delaney and Denlinger, 1999) is illustrated in detail by combined geodetic and seismic studies (Brooks and others, 2006; Montgomery-Brown and others, 2009, 2011) that demonstrate a delicate balance among rift-zone extension, dike injection, and flank deformation coupled to décollement slip.

Seaward of the summit and ERZ, the relation between slip on the Hilina and Hōlei Pali Fault Systems and décollement slip has been questioned because the 1975 *M*7.7 earthquake produced slip on both (Lipman and others, 1985; Cannon and others, 2001). Kinematic modeling of Global Positioning System (GPS) displacements suggests that the Hilina Fault System has shallow roots, connecting to a low-angle thrust at no more than 4-km depth (Cannon and others, 2001; Cervelli and others, 2002). This same conclusion appears to be supported by consideration of landward surface tilt produced by antithetic faulting on the footwalls (Swanson and others, 1976), as well as by a force balance for the south flank, based on the method of slices (Okubo, 2004) and detailed studies of prehistoric fault slip along the Hilina Pali fault scarp, using elastic-dislocation models (Cannon and Bürgmann, 2001). Finite-strength, particle-based modeling (Morgan and McGovern, 2005a) yields similar results, supporting a shallow faulting interpretation for the Hilina and Hōlei Pali Fault Systems.

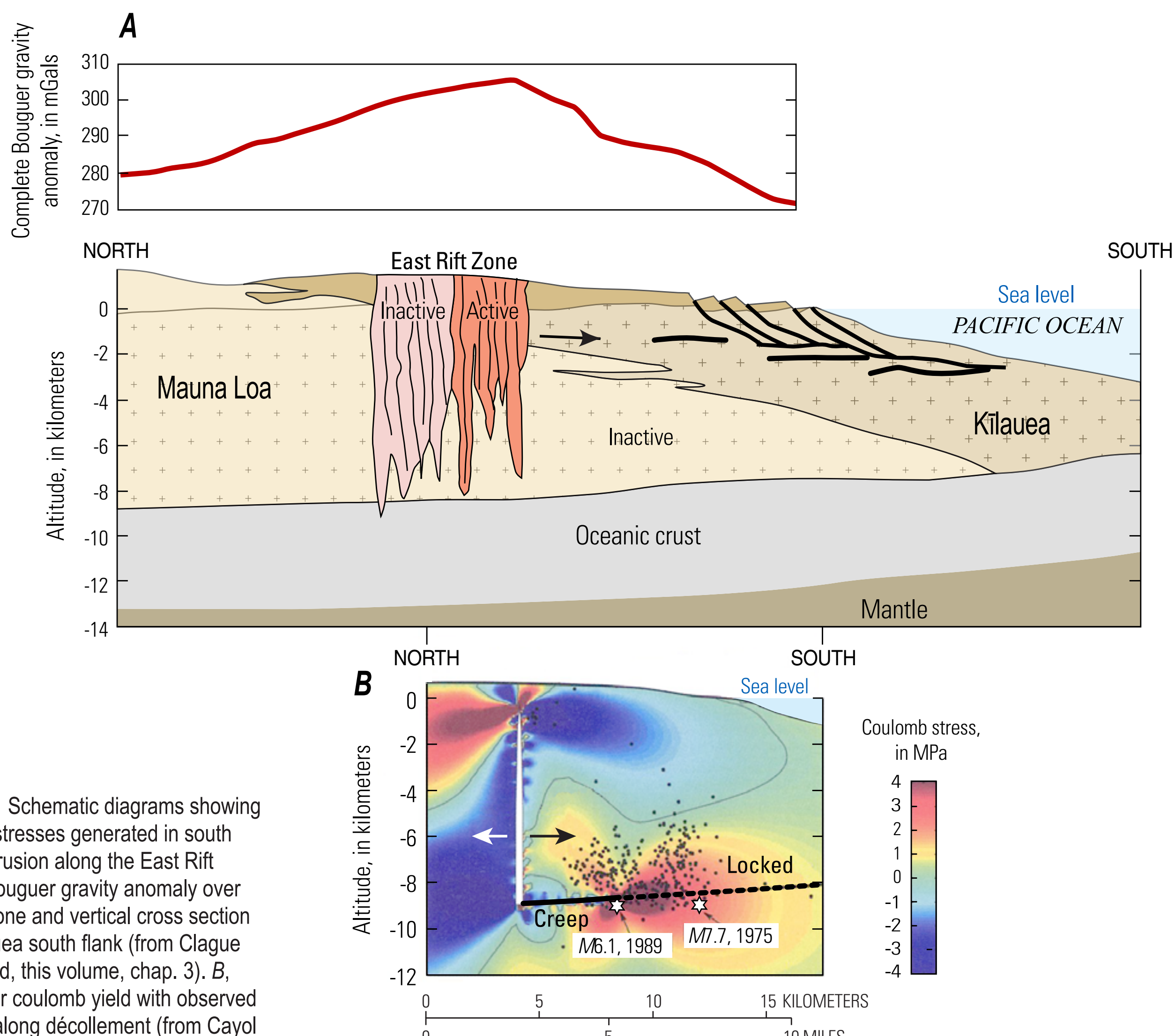


Figure 7. Schematic diagrams showing estimated stresses generated in south flank by intrusion along the East Rift Zone. *A*, Bouguer gravity anomaly over East Rift Zone and vertical cross section of the Kīlauea south flank (from Clague and Sherrod, this volume, chap. 3). *B*, Potential for coulomb yield with observed seismicity along décollement (from Cayol and others, 2000).

Geophysical Evidence for Magma Systems Within Hawaiian Volcanoes and Their Role in Flank Instability

Studies of the deep seismicity below the Island of Hawai‘i suggest that the magma supply to both Kīlauea and Mauna Loa volcanoes may have a common source at great depths (see Okubo and others, 1997a; Poland and others, this volume, chap. 5). From its origin as a mantle hot spot (Klein and others, 1987; Okubo and others, 1997a; Wolfe, 1998; Wright and Klein, 2006), the magma supply diverges beneath the island into separate magmatic plumbing systems (as shown by studies of magma chemistry and deep seismicity), although the relation of deep seismicity to magma transport beneath the Island of Hawai‘i remains controversial (Okubo and others, 1997a; Wolfe and others, 2003, 2004; Pritchard and others, 2006). Within the island edifice, melt storage generates little seismicity (relative to eruptions or flank and rift-zone deformation) and so is hard to image (Ryan, 1988). Thus, the magma plumbing systems for Kīlauea and Mauna Loa have been interpreted mainly from gravity studies (Kauahikaua and others, 2000) and seismic tomography studies (Okubo and others, 1997b; Benz and others, 2002; Baher and others, 2003; Park and others, 2007a, 2009), inferred from event-relocation studies (Okubo and others, 1997b), and more recently, modeled by analyses of surface deformation (see Poland and others, this volume, chap. 5). The segregation of seismicity into that related only to magma transport and that related only to flank deformation is determined by the proximity of earthquake locations to known pathways of magma transport beneath the summit and rift zones and by the coincidence of these paths to sources of gravity and seismic anomalies.

Gravity studies in the Hawaiian Islands (Strange and others, 1965), as well as more detailed studies over the Island of Hawai‘i (fig. 8; Kinoshita and others, 1963; Kauahikaua and others, 2000) indicate that high gravity anomalies are commonly associated with the summits and rift zones of the volcanoes. Where detailed seismic tomography is available, high seismic velocities coincide with dense subsurface regions interpreted as the sources of high gravity anomalies (Benz and others, 2002; Park and others, 2009), as shown in figure 9. On the Island of Hawai‘i, the association of high seismic velocity and high density beneath the summits and along rift zones is hypothesized to result from the continuous precipitation and settling of olivine crystals to form dunite bodies of cumulate within or beneath the tholeiitic magma that resides within each volcano (Clague and Denlinger, 1994). Geologic support for this hypothesis is provided by eruption of picrites, commonly containing numerous dunite nodules, 0.05 to 0.1 m in size, from deep within the magma systems (Wilkenson and Hensel, 1988). A critical attribute of such deep dunite bodies, when maintained at high temperatures and replenished by magmatic intrusions, is their ability to flow viscously under load. As argued by Clague and Denlinger (1994), these ductile dunite bodies may play a key role in mobilizing the south flank of Kīlauea, helping stress transfer to the distal flank (fig. 7) while extending the near-summit regions, such as the Koa‘e Fault System. Evidence for efficient stress transfer through the creeping dunite is that décollement-related flank seismicity begins seaward of these dunite bodies (figs. 5*B* and 7), whose location is interpreted from Bouguer gravity anomalies (figs. 7 and 8).

High gravity anomalies in summit regions and along rift zones, which indicate increasing mass at depth, are also a measure of gradients in gravitational body forces acting within the flank of each volcano. The shape of the gravity field at Kīlauea indicates the presence of material (presumably cumulates) beneath the summit, Koa‘e Fault System, and ERZ that is much denser than in the surrounding flank (Kauahikaua and others, 2000). Gradients in the gravitational field show directions of body forces within the flank, and, for the south flank, indicate whether these forces would help drive flank motion seaward. If flank motion is partly driven by its weight, then the deformation field of the flank and associated gradients in the gravity field should correlate (as shown for south-flank movement only in figure 8*B*). In addition, the pattern of subsidence of Kīlauea’s summit within the past few decades during prolonged extension of the flank has roughly the same shape as the summit gravity anomaly. The steepest gradient, along the southern margin of the gravity high over the summit, coincides with the south boundary of the Koa‘e Fault System and its extension into the ERZ, and the steepest gradient to the east and west corresponds to the upper ERZ and SWRZ, respectively (fig. 8).

Correlations between seismic-velocity inversions and gravity anomalies provide important insights into velocity-density relations within these volcanoes. Different velocity-density relations apply to active and inactive rift zones and summit areas (Park and others, 2007b). For example, a massive seismic-velocity and gravity anomaly detected within Mauna Loa’s south flank (fig. 9*C*) corresponds to a typical high-seismic-velocity, high-density relation. In contrast, the high seismic velocities beneath Kīlauea’s magmatically active summit and ERZ (fig. 9) are associated with lower, but still anomalous, gravity highs. One interpretation of this discrepancy between the two settings is that regions with typical seismic-velocity/density relations denote solidified dunite cumulate, whereas those with lower velocity-density relations indicate the presence of interstitial melt within the cumulate, maintained by higher temperatures and capable of producing magma that can feed eruptions (Park and others, 2007b). If this is true, then the combined analysis of seismic-velocity and gravity anomalies provides a means of distinguishing old, cold magma bodies from young, hot magma bodies, providing a first-order constraint on the mobility of these bodies and their potential contributions to flank instability.

In summary, geophysical evidence for the distributions of high seismic velocity and high density supports the interpretation that seismic velocity and density anomalies are associated with magma pathways within the volcanoes, whether these pathways are active or inactive. The correlation between seismic velocity and density is also useful in defining which areas are still hot (low velocity-density ratio) and potentially active, even without

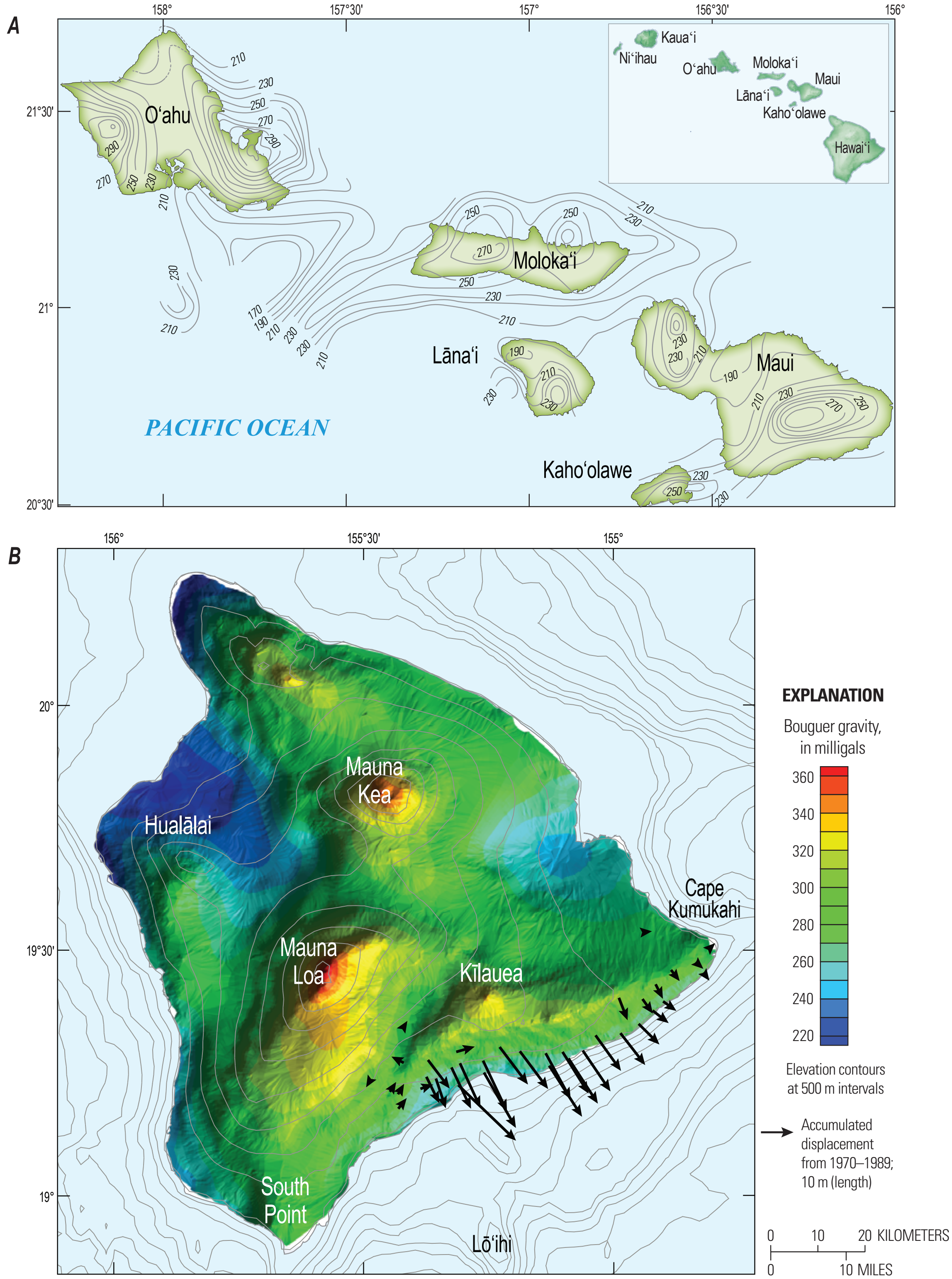
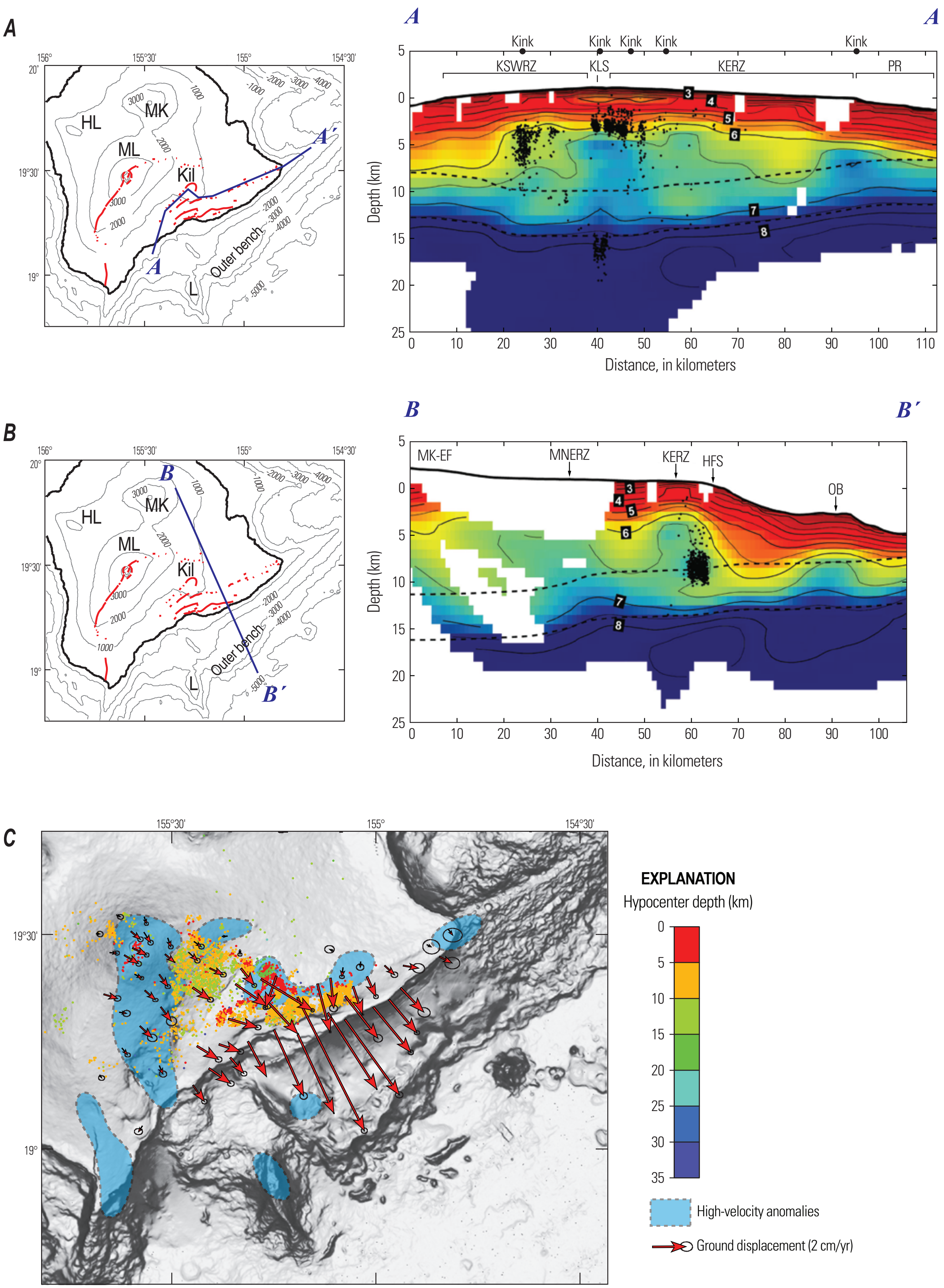


Figure 8. Gravity anomaly maps for the Hawaiian Islands and Island of Hawai'i. *A*, Gravity anomalies (in milligals) associated with old rift zones on the Islands of O'ahu, Moloka'i, and Maui, indicating that these rift zones have a dense core, most likely composed of dunite accumulated during their growth. *B*, Current gravity anomalies associated with active and inactive volcanoes on the Island of Hawai'i. Dunite accumulates in deep core of summits and rift zones when magma circulates through these regions. Magnitude of anomaly is related to length of time that magma circulated, and presence of anomalies in other areas (for example, southeast of summit of Mauna Loa) may indicate old rift zones. Vectors on south flank of Kīlauea Volcano denote the accumulated displacement from 1970 to 1989, indicating displacement of flank in response to magma pressure, as well as the gravitational load of dunite. Directions of gradients of gravity anomalies are indicated by gradients in colors used to display gravity.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Kink
KSWRZ
KLS
KERZ
PR
Depth (km)
Distance, in kilometers
B
B
B′
MK-EF
MNERZ
KERZ
HFS
OB
C
EXPLANATION
Hypocenter depth (km)
High-velocity anomalies
Ground displacement (2 cm/yr)

obvious current deformation or seismicity. The correlation of gradients in gravity anomalies with active flank velocities on the south flank of Kīlauea (fig. 8*B*) is evidence that the flow of ductile dunite cumulates contributes to seaward flank motion, because these anomalies record a force imbalance that potentially drives flank motion. In this case, the scaling of the gravitational forces obtained from the flank density distribution is useful for scaling forces driving flank motion and constraining mechanical models for flank instability.

Mechanics of Flank Instability

To explore the mechanical constraints on flank deformation at Kīlauea, we consider several models for flank instability. The simplest model is a rigid wedge sliding on a décollement, in which a force balance is derived for the entire domain. Adding elasticity to the wedge allows for storage and release of elastic strain energy, providing a mechanism for coupling between décollement slip and wedge deformation. Adding finite strength to the wedge localizes deformation within it and limits stress through faulting, which is most notably associated with large earthquakes and slow slip events. Each model scales the forces driving flank instability differently, and we consider each in order of complexity.

The models are constrained by the morphologic, structural, and geophysical evidence for flank instability outlined above. This evidence indicates phenomena that either contribute to, or are a consequence of, the instability of Hawaiian volcano flanks (fig. 3), including (1) magmatic intrusion, which increases mean stress beneath summits and along rift zones, building stress gradients that episodically are transferred seaward by slow slip events; (2) episodic transfer of stress and deformation seaward by slow slip events and small earthquakes, migrating the volcanic pile seaward on a persistent, low friction décollement; (3) termination of décollement slip in thrust faults that build a distal bench from accreted volcanic debris; (4) viscoelastic creep of dunite cumulate, adding to slow progressive loading of flanks beneath summits and along rift zones between episodes of rapid slip; and (5) rapid, large-scale flank movements associated with large earthquakes (for example, the 1975 *M*7.7 earthquake). Flank models show that each of these phenomena plays a different role in altering the forces affecting flank stability.

A rough estimate of the static balance of forces governing flank stability can be obtained from an idealized rigid-wedge model derived from the south-flank geometry (Dieterich, 1988; Iverson, 1995). For the south flank of Kīlauea, we can assume that the back of the wedge approximately corresponds to the south boundary of the Koaʻe Fault System and its extension along the ERZ, as outlined by Duffield (1975) and illustrated in figure 3. The lateral extent of rift extension defines the lateral margins of the wedge, and the distal end of the wedge corresponds to the distal end of the offshore bench. The base of the wedge slides on a décollement, interpreted to coincide with a pelagic sedimentary layer sandwiched between the volcanic edifice and the Cretaceous seafloor (Nakamura, 1982). This layer has been cored and imaged just south of the Island of Hawaiʻi and determined to be as much as 200 m thick (Leslie and others, 2002). The wedge model for the south flank is partly submerged, which influences its specific weight and applied forces (Iverson, 1995; Dieterich, 1988), as shown schematically in figure 3.

The driving forces for a rigid-wedge model, with the geometry illustrated in figure 3, are produced by (1) the weight of magma and cumulates pressurizing the back of the wedge along the rift zone, (2) excess magma pressure, and (3) the component of weight acting tangentially along the décollement (where the surface slope exceeds the décollement slope) (Iverson, 1995). Resisting forces are dominated by (1) the component of weight acting normal to the décollement, resisting wedge motion through fault friction, and (2) excess pore pressures or seepage forces within the décollement that offset fault friction.

Limiting equilibrium provides a framework for assessing the relative importance of each of the driving and resisting forces outlined above. The weight of the flank, combined with a wedge-shaped cross section, produces both a shear stress and a normal stress across the décollement. In the absence of internal deformation, forces driving flank motion are resisted only by friction along the décollement (Dieterich, 1988; Iverson, 1995) formed in the sedimentary layer that is overridden by volcano growth. As the flank overrides and deforms this sediment, effective friction along the décollement is reduced by excess pore pressures generated by burial, shearing, and consolidation. For the south flank of Kīlauea, rigid-wedge models indicate that zero excess magma pressure along the ERZ requires the average effective internal friction

Figure 9. Vertical profiles through P-wave-velocity model derived from first-arrival tomographic inversion around Kīlauea and surrounding volcanoes (Park and others, 2009). Accompanying maps show profile locations, island coastline and 1,000-m bathymetric contours, and main geologic features. Abbreviations are: HL, Hualālai; L, Lōʻihi; MK, Mauna Kea; ML, Mauna Loa; Kil, Kīlauea. *A*, Profile A–A' along Kīlauea's Southwest Rift Zone (KSWRZ), summit (KLS), East Rift Zone (KERZ), and Puna ridge (PR). *B*, Profile B–B' crossing Mauna Kea's east flank (MK-EF) and northeast rift zone (MNERZ), KERZ, Hilina Fracture System (HFS), and offshore outer bench (OB). White areas, unsampled regions of model in vertical profiles. Contour interval, 0.5 km/s, index contours labeled in white. Vertical exaggeration 2x. Earthquakes located within 2 km on either side of each profile are projected onto profiles (black dots). Dashed black lines, interpreted top of oceanic crust and lithospheric Moho. *C*, Maps of southeastern part of the Island of Hawaiʻi (adapted from Park and others, 2007a), showing ground displacements, seismicity, and interpreted extent of high seismic-velocity (~6.5 km/s) anomalies near base of volcanic edifice. Red arrows, average horizontal surface velocities, with 95 percent confidence ellipses, measured by campaign Global Positioning System (GPS) between 1997 and 2002 (Miklius and others, 2005). Colored dots are *M*>2.0 earthquakes with hypocentral depths of 5–13 km from 1995 to 2005.

angle along the décollement to be <16° for any flank motion to occur (Iverson, 1995). Given typical values of internal friction angles for pelagic clays (30°–40°), a zero excess magma pressure condition along the ERZ places stringent limits on the magnitude of excess pore pressure required to produce flank motion.

In contrast to rigid-wedge models, an elastic-wedge model for south-flank deformation deforms internally and can accommodate limited décollement slip adjacent to the rift without requiring either large magma pressures or low friction along the entire décollement. At Kīlauea, even small dike injections move the south flank, as demonstrated by Montgomery-Brown and others (2010). Elastic models (Yin and Kelty, 2000) show that excess magma pressures of ~10 MPa, typical of dike injections along the ERZ (Rubin and Gillard, 1997), can produce limited slip on a ~3-km-wide swath next to the rift zone, comparable to the region of the south flank where rift-zone extension initiates slow slip events (Montgomery-Brown and others, 2011). However, we lack observations that constrain how stress accumulates along the rift zone and propagates seaward through the flank. How do repeated dike injections move a 50-km-wide flank seaward?

Scaling provided by wedge models shows that the primary resistance to flank motion in response to loading along the rift is décollement friction (Iverson, 1995; Yin and Kelty, 2000), and the décollement is where most south-flank earthquakes, including the largest events, occur. We assume that the excess pore pressure, which reduces friction within décollement sediment, builds as the sediment is overridden and compressed by the flank. Pore pressure within pelagic clays will increase through compaction and may be enhanced by fluid circulation within the oceanic crust (Christiansen and Garven, 2004). Although wedge models idealize décollement resistance in terms of some constant average, décollement friction during consolidation is likely to vary broadly in space and time, just as Iverson (1986) describes for landslide motion. In particular, coupling between pore pressure and dilatancy during shear deformation can strongly modulate décollement resistance to flank motion, just as it does along the base of landslides (Iverson, 2005), and the transient nature of this coupling can lead to rapid, unstable landslide motion after decades of creep. Similar mechanisms have been proposed to act along shallow subduction zones (Beeler, 2007).

The time scales for instability obtained by coupling dilatancy and pore pressure depend on the material fabric within the décollement. With reference to Iverson's (2005) model for landslides, the coupling between south-flank motion, shear-induced décollement-zone volume change (dilatancy), and décollement pore-pressure change depends on the rate of volume change with shear strain and shear strain-rate and the intrinsic time scales for pore-pressure generation and dissipation within the décollement. Dilatancy produces a negative pore pressure feedback that can regulate south-flank motion for years. However, if dilatancy decays over time (or with increasing shear strain), then pore pressure can suddenly increase rapidly after years of creep, generating rapid slip. The material properties of the subaerial Minor Creek landslide in northwestern California, reported by Iverson (2005), happen to be consistent with a mixture of volcaniclastic sediment and pelagic clays, giving a wide range of potential time constants for instability. The addition of volcaniclastic sediment to clay-rich sediment produces a fabric in which dilatancy induced by shear will decrease over time. For the values in table 1 of Iverson (2005), but with a décollement thickness of 300 m for Kīlauea's south flank rather than 6 m for Minor Creek, continued south-flank motion will generate rapid pore pressure growth within the décollement after 2 years of steady creep—an intriguing prospect, given the observed 2-year frequency of south-flank slow slip events (Brooks and others, 2006; Montgomery-Brown and others, 2009). And, just as with large landslides, observations of repetitive-slip events support a mechanism that either resets fabric dilatancy along the old slip surface or generates a new slip surface within décollement sediments.

Given these considerations and recent observations of triggered seismicity associated with slow slip events (Segall and others, 2006; Wolfe and others, 2007), such events may be an endemic process indicative of flank instability at Kīlauea (Cervelli and others, 2002; Brooks and others, 2008; Montgomery-Brown and others, 2009, 2011). Slow slip events occur updip of documented south-flank earthquakes and are presumed to correlate with the transition between seismic and aseismic slip, similar to subduction zones (Liu and Rice; 2007), where slow slip events may occur both downdip and updip of the seismogenic zone (Dragert and others, 2001; Rogers and Dragert, 2003; Norabuena and others, 2004; Obara and others, 2004). At Kīlauea Volcano, slow slip events help translate the south flank seaward, that is, in the same direction as overall flank motion, and are a record of stress transfer.

Elastic modeling of the locations of slow slip along Kīlauea's décollement (Brooks and others, 2006; Montgomery-Brown and others, 2011) shows termination of slip landward of the uplifted outer bench. The bench may restrict further transfer of stress seaward, at least intermittently. Indeed, mechanical models show that décollement slip will terminate by transforming into a thrust fault that curves up into the overburden (Mandl, 1988), similar to the thrust faults recognized within the outer bench at Kīlauea (fig. 4; Morgan and others, 2000, 2003a). The associated bench uplift thickens the overburden, increasing resistance to slip along the décollement (Mandl, 1988) and further enhancing the tendency for thrust faulting and bench growth. New submarine geodetic studies (Phillips and others, 2008) support uplift by thrust faults in the area where geodetic studies show that décollement slip terminates (Montgomery-Brown and others, 2011). Slow slip events may thus be delimited by the outer bench along Hawaiian volcanoes and may be the manifestation of long-term flank-creep-driven upper-flank loading.

The mechanical models discussed above are focused on coherent mobile flanks driven by loading along the rift zones; however, the structure of the volcano flanks, particularly Kīlauea's south flank, reveals both extensional and contractional faults, confirming the importance of internal deformation during flank instability. The relations between flank structure, décollement strength, and internal strength can be explored during volcano growth by using particle-based numerical models. These models, in which volcanoes are constructed by the addition of granular material near the summits above a frictional décollement, result in outward spreading of the volcano flanks, accompanied by faulting (fig. 10). The balance of stresses within the flank, governed by the relative strengths of the flank and the underlying décollement, defines the final geometry and internal structure of the mobile flank (Morgan and McGovern, 2005b). In the examples shown here (Zivney, 2010), low décollement friction (fig. 10*A*) results in deep-seated faulting centered beneath the summit, causing predominantly lateral displacement of the volcano flanks, whereas increasing basal friction (figs. 10*B*–10*D*) results in progressively shallower flank deformation, culminating in shallow landslides that cut the flanks. The addition of a dense, ductile core (shown in purple in figs. 10*E*–10*H*), analogous to the dunite cores interpreted within Hawaiian volcanoes (Clague and Denlinger, 1994), concentrates extensional deformation beneath the summit and is independent of décollement strength. Slip becomes localized along the décollement fault and is enhanced by summit subsidence. The volcano flanks undergo modest extension as they are displaced. An important element of these simple simulations is the incremental nature of décollement slip and flank deformation as stress builds and is released during volcano growth, resulting in discrete slip and deformation events (Morgan, 2006). These examples offer great potential for constructing more elaborate, mechanical-based models to better understand the unique processes currently underway at Hawaiian volcanoes.

Discussion

For the past 200 years, observations of the active volcanoes Mauna Loa and Kīlauea have revealed repeated episodes in which steady growth of the volcano at the summit and along the rift zones was facilitated by gradual seaward spreading of the flanks and punctuated by episodes of rapid seaward slip. For Kīlauea, such large-scale rapid slip occurred on the south flank in the second half of the 18th century and in 1823, 1868, 1924, and 1975, whereas for Mauna Loa it has occurred only twice, in 1868 and 1951 (Macdonald and Wentworth, 1954). Although these flank movements were sometimes associated with a large earthquake and tsunami, they were always associated with extension and subsidence of the coast during seaward movement, opening of the rift zones, and, with extension, a subsequent drop in magma-system head by hundreds of meters.

During the 1975 *M*7.7 earthquake on Kīlauea, downward and seaward movement of the south flank reversed the coastal compression and uplift that had been recorded during growth of the volcano since the last large flank movement in 1924, when the lava level had dropped by hundreds of meters at the summit. As it had in 1924, the lava level dropped again during the 1975 earthquake. Large flank movements and sudden drops in magma-system head, followed by subsequent refilling of the magma system, form a pattern of growth and collapse observed at Kīlauea after each large flank movement since 1823. During each decades-long increase in magma-system head, many smaller episodes of flank deformation occur in which stress is transmitted from intrusions along the ERZ to the south flank. Before each episode of slip, the summit and rift zones are gradually wedged open by magma pressure, presumably facilitated by dunite creep at depth, and the adjacent flank is compressed by shallow magma intrusion. As the rift is wedged open, compression on the seaward flank progressively increases until seismic or aseismic slip is triggered on the underlying décollement. At Kīlauea, seismicity associated with décollement slip begins 2 to 3 km seaward of the rift zone (compare figs. 5*A* and 5*B*; see Montgomery-Brown and others, 2011) and is manifested in the deep seismicity mapped in figure 5*B*.

The offset between the shallow rift-zone seismicity and deeper décollement seismicity indicates separate responses of the flank to push along the rift zone. This response is intriguing, because analogous mechanical models of thrust sheets in which stress is measured (Blay and others, 1977; Mandl, 1988) can explain this offset. Experiments show that if a thrust sheet rides on a frictional base but is underlain by a short section of viscous material at the proximal end, as shown in figure 11, then a push on the proximal end rapidly transmits stress to the distal end of the viscous underlayment. Stress accumulates at the distal end of the viscous underlayment (which separates it from where the push is applied) until frictional slip begins there. As slip begins, stick-slip events then migrate stress along the frictional base toward the distal end. Relating this experiment to the south flank of Kīlauea, the viscous underlayment is an analog for hot dunite, and the frictional base is an analog for the décollement; thus, the offset between shallow rift seismicity and deeper décollement seismicity appears to be consistent with the mechanical indication of a viscous cumulate body adjacent to the rift zone and more distal slip along a décollement. A comparison between the gravity data and seismicity mapped in figure 8 and the velocity data and seismicity mapped in figure 5 also supports this conclusion. The shallow rift seismicity overlies one side of the summit gravity and velocity anomalies, whereas the deeper décollement seismicity forms the seaward boundary of the other side of the gravity and velocity anomalies. Slip events initiate at the distal end of the hot dunite body and migrate the flank seaward, opening the rift zone (Montgomery-Brown and others, 2010). If the intervening aseismic zone is composed of hot dunite, then its viscous response will not only separate the epicenters of shallow rift seismicity from deeper décollement seismicity, but also delay the response of the rift zone to slow slip events or the triggering of slow slip events by rift-zone loading.

At the distal end of décollement slip, each modeled slow slip event (Montgomery-Brown and others, 2010) is likely to terminate beneath the offshore portion of the flank undergoing uplift (Phillips and others, 2008), consistent with mechanical analysis showing that décollement slip on model thrust sheets terminates where thrust faults penetrate into the overburden (Mandl, 1988). A corollary of this model is that if the

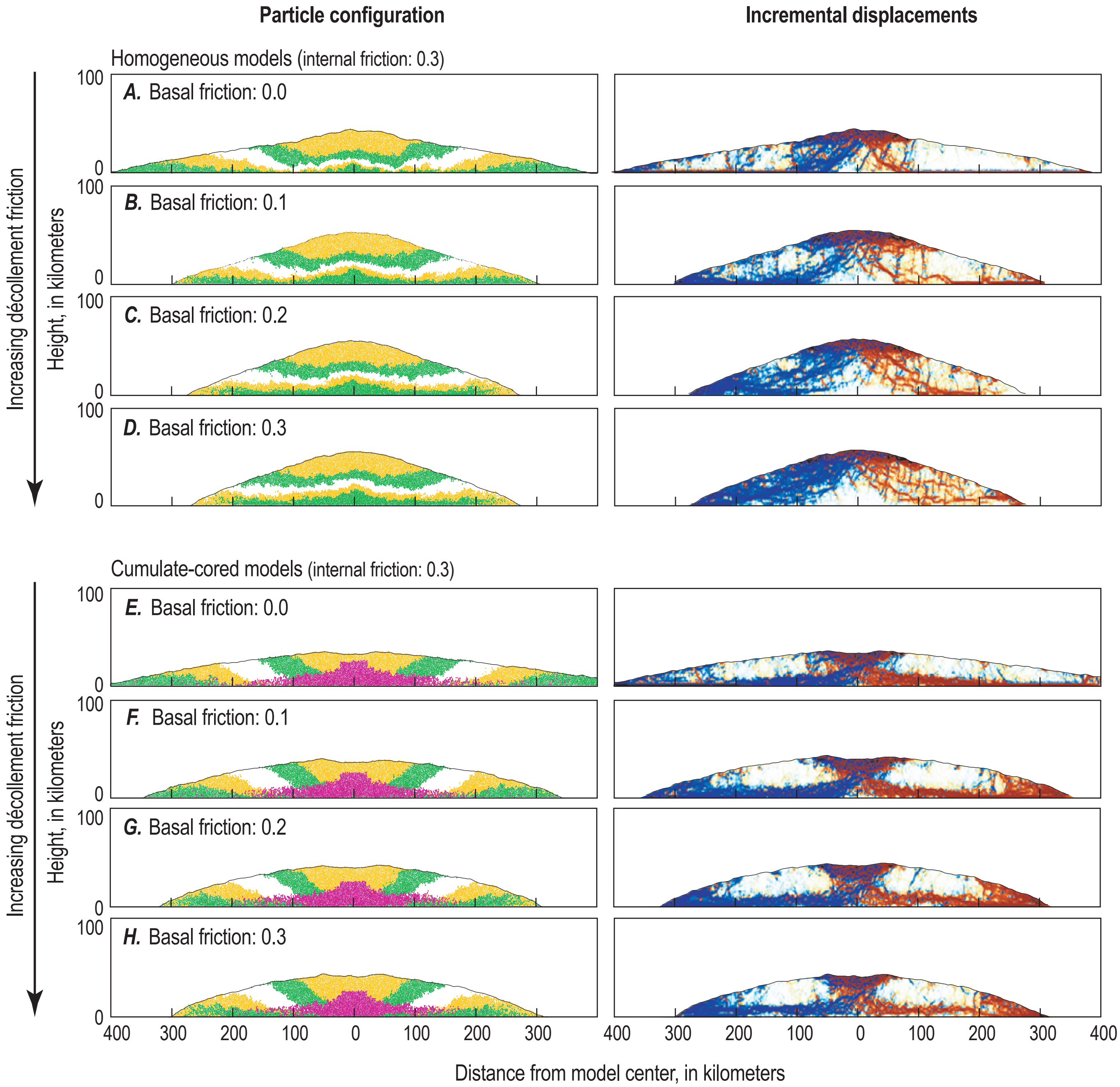

Figure 10. Cross-sectional diagrams showing particle-dynamics simulations of symmetrical volcanic spreading with and without weak cumulate cores (adapted from Zivney, 2010, after Morgan and McGovern, 2005a,b). Interparticle friction noted here differs from bulk friction, which is quantified by Morgan and McGovern (2005a). *A–D*, Effects of basal-friction variation with no cumulate (that is, homogeneous models), showing final particle configuration (left) and incremental distortional strain (right). Red, right-lateral shear; blue, left-lateral shear. Basal friction increases from *A* to *D*, resulting in decreased edifice spreading and steeper flanks. Deformation transitions from slip along base for low basal friction, through enhanced normal faulting, to surficial avalanching at high basal friction. *E–H*, Effects of basal friction variation for cumulate-cored (purple) edifices, otherwise similar to homogeneous volcanoes. Basal friction increases from figure *E* to figure *H*, causing decreased edifice spreading; weak cumulate material spreads laterally, pushing flanks outward, relative to homogeneous edifices. Deformation is characterized by enhanced slip along base of edifice and within weak cumulate; normal faulting and summit subsidence occur above spreading cumulate, and modest surficial landslides still occur. Cumulate-cored volcanoes exhibit dome-shaped morphology with inward-dipping strata.

thickness of the overburden increases seaward (as in the toe of the mobile flank), then the locus of imbricate thrusting from the décollement will move landward. If, over decades, many repeated slow slip events occur (Brooks and others, 2006) and each event builds and uplifts the offshore bench, then, as slip events terminate closer to shore, the distal bench will grow progressively and inexorably landward over time. Evidence for this process exists in the marine seismic data over Kīlauea's south flank but shows that thrusts may also form on the seaward side of the outer bench (Morgan and others, 2000, 2003a), presumably when décollement slip undercuts the bench and terminates there.

The transition of basal slip to thrusts at the distal end of the flank is just one component of flank growth over time that is represented in particle-based models in which stress is limited by a finite-strength criterion and finite strain is allowed (Morgan and McGovern, 2005b). These models show that many of the structures and fault patterns we observe on Kīlauea and Mauna Loa volcanoes evolve over long time periods during volcano growth as the flanks spread over a weak décollement (Morgan, 2006). The correspondence of these models to volcano shape, formation of pali on the subaerial flank, shallow landslide activity, and formation of a distal bench contributes to our understanding of volcano growth.

Onshore and offshore studies of the structure of Hawaiian volcanoes of different ages, from young (Lōʻihi) through middle-aged (Kīlauea and Mauna Loa) to old (Molokaʻi and Oʻahu), and observations of the behavior of Kīlauea and Mauna Loa volcanoes, can be incorporated into an inclusive model representing the volcanotectonic stages of volcano growth and degradation (fig. 12). During the earliest submerged stages of shield building (stage 1), the rift zones grow upward and outward on their seaward side(s). A modest cumulate core lies within the edifice, and the low gravity anomaly over small edifices like Lōʻihi indicates that the push from the cumulate at this stage of growth is too small to drive outward spreading of the volcano flanks. Steep slopes produce shallow landslides along the volcano flanks, building an apron of debris at the distal edge of the flank. In stage 2, an enlarged cumulate core can push the flank outward, in concert with dike intrusions along the rift axis. A décollement fault develops near the base of the edifice, allowing the mobile flank to be pushed into the apron of debris, which deforms by imbricate thrusting at the distal edge. In stage 3, the upper parts of the volcano are subaerial, and the seaward flank is large and thick enough to sustain shallow landslide activity that does not penetrate to the décollement, similar to that interpreted for Kīlauea's south flank (Morgan and others, 2003a). Décollement slip continues, further building the outer

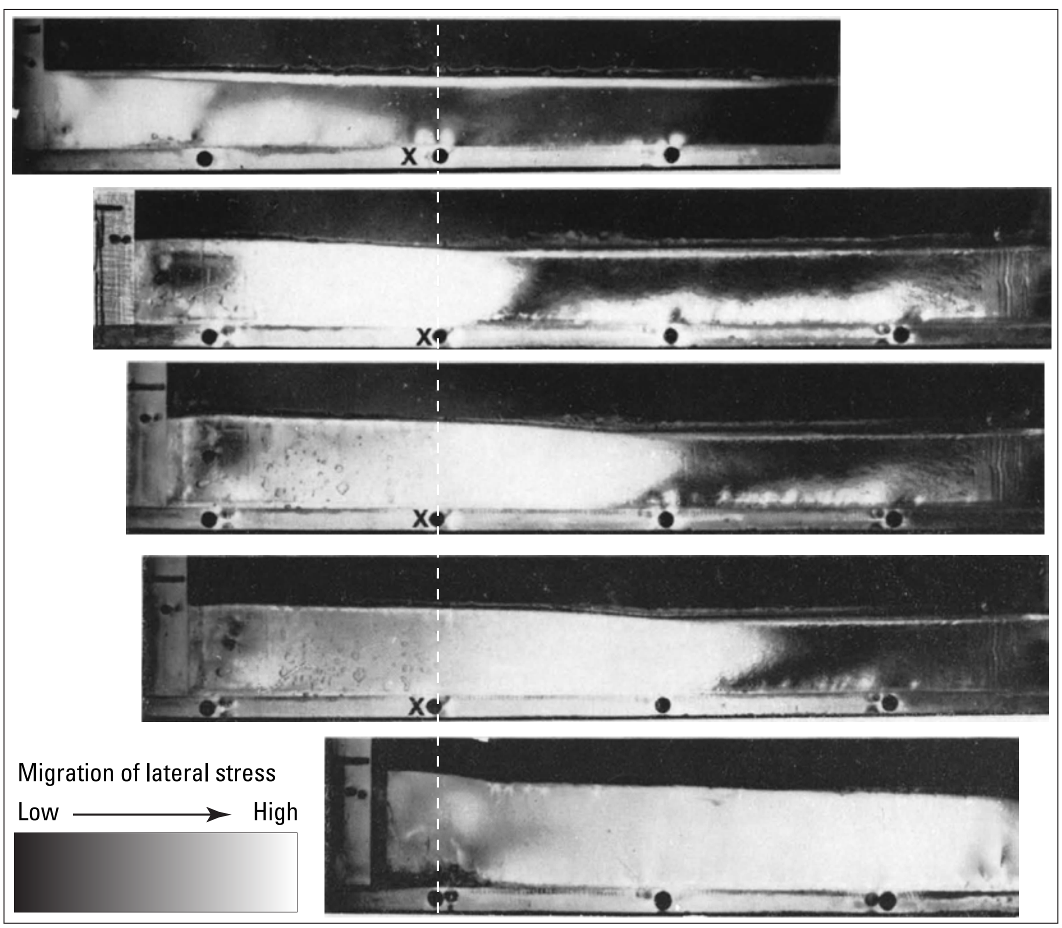

Figure 11. Series of photographs showing stresses in gelatin, capturing migration of lateral stress (higher stress=lighter color) within laboratory model of a thrust sheet pushed from behind (left edge). Base is lubricated from left edge to point X. Applied load on left edge increases in each photograph from top to bottom, which are all lined up, relative to point X. Stress migrates quickly over lubricated zone between this edge and point X and is stored at its distal end, where it accumulates, until sheet can decouple from its base at point X, transferring stress farther toward distal end in a series of small stick-slip events along décollement.

bench. Décollement slip is accomplished both by intermittent large earthquakes and by slow slip events (Brooks and others, 2006). Finally, in stage 4, catastrophic failure of a sector of the flank may occur, possibly including parts of the outer bench, dispersing landslide debris far across the seafloor and leaving telltale scars along the volcano flanks. If growth continues at the summit and along the rift zones, the cycles begin again.

Many of the processes outlined in the evolutionary stages we present here are captured in the particle-based models introduced above, although such models are not yet refined enough to reproduce specific events or structures. Further work using cohesive models with geometries designed to match Hawaiian volcanoes could provide additional insights into these behaviors. Importantly, however, the specific triggers for rapid flank movement are poorly represented in such models because these triggers occur on different spatial and temporal scales from those represented in the models. Despite this deficiency, the particle-based models do show how stress is stored and transmitted within the volcano flanks and the critical importance of décollement resistance to stabilizing the forces that drive flank motion. We therefore argue that flank instability is directly related to décollement processes and properties that are capable of the rapid changes that will influence the resistance to flank motion. Preexisting landslide detachments within the edifice will respond to similar changes. One process that is expected to occur as saturated sediment is sheared is the tight coupling between dilatancy and pore pressure. Using the analysis of Iverson (2005) as a model for the décollement beneath the south flank of Kīlauea, and using his parameters and a thickness (200 m to 300 m) appropriate for a mixture of turbidites (sandy silt) and clay for the pelagic layer overridden by the south flank, we obtain time constants for slow creep, terminated by rupture and rapid slip. For this mixture and thickness, the rapid slip occurs at intervals of one to three years, consistent with the observed range of recurrence of slow slip events at Kīlauea (Brooks and others, 2006). As a consequence, the coupling of induced pore pressure and dilatancy within the décollement represents a promising and intriguing prospect for future research.

Finally, the near-periodic occurrence of large, rapid flank displacements on Kīlauea and Mauna Loa raises the question of whether these events are driven by the same mechanisms that produce sector collapse, such as those documented along the northeast shore of Oʻahu and the north shore of Molokaʻi (fig. 1), as well as along Mauna Loa's west flank (Lipman and others, 1988; Morgan and others, 2007). Our understanding of the scaling driving flank motion, though limited, suggests that each large rapid flank movement is just a smaller version of a much larger sector collapse. The processes affecting resistance to flank motion, such as pore-pressure/dilatancy coupling along the décollement, can drastically and rapidly reduce the resistance to sliding on time scales comparable to those of the slow slip events observed on Kīlauea, but it is unknown how this scales up to areas comparable to a large flank earthquake or scales by orders of magnitude in size to form a sector collapse. In addition, special conditions and circumstances are needed to rebuild pore pressure/dilatancy coupling along the décollement after each slow slip event, as well as after each large catastrophic flank movement. Alternatively, or perhaps in association with pore-pressure/dilatancy coupling, other processes may contribute to instability through rate- and state-dependent variations in friction along the décollement (Dieterich and others, 2003). Finally, explosive volcanic eruptions, which are known to have occurred in all of these settings, provide a potential and unconstrained trigger for large-scale sector collapses (McMurtry and others, 1999; Morgan and others, 2003a). Further study is needed to determine how to constrain these potential mechanisms of flank instability.

Conclusions

Hawaiian volcanoes of different ages exhibit a wealth of evidence for past mobility and instability, including large debris-avalanche deposits scattered across the seafloor, uplifted outer benches at the distal edges of growing volcano flanks, proximal fault scarps, and seafloor scars related to rapid landslide motion. In recent decades, we have gained tremendous insights about the processes responsible for these features through improved seafloor mapping, submersible surveys of the incised and exposed submarine flanks, marine seismic surveys, and seismic-velocity and density modeling. The knowledge gained about the internal structures of Hawaiian volcanoes has revealed the cumulative effects of volcano deformation throughout the growth of these volcanoes, suggesting an evolution of processes during the growth and collapse of Hawaiian volcanoes. Some of these processes promote flank instability, producing episodic and rapid flank movements that ultimately may be the cause of the huge landslides that dissect entire islands as they grow.

The active volcanoes Kīlauea and Mauna Loa on the Island of Hawaiʻi have shown separate volcanotectonic stages of growth, but both volcanoes exhibit cyclic variations in rapid flank movement during growth. Decades-long growth across their summits and along their rift zones is punctuated by brief periods during which the seaward flank of the growing rift zone rapidly moves seaward in seconds to minutes. In this fashion, the growth of rift zones is one-sided, always upward through lava accretion and seaward through flank motion, and it is to the seaward side that rapid movement and associated large-scale gravitational collapse occurs. Sometimes this rapid motion is coseismic, generating a great earthquake and tsunami, but it always results in opening of the rift zone, subsidence of the coast, and collapse of the summit magma-system head by hundreds of meters. Whereas idealized rigid and elastic models of rift flanks shed light on the relative importance of driving and resisting forces that move the flank seaward, successful interpretation of flank evolution requires continual balancing of forces in a flank with finite strength as the volcano grows. Incorporating finite strength and finite strain into models for volcano growth provides considerable insight into the growth of

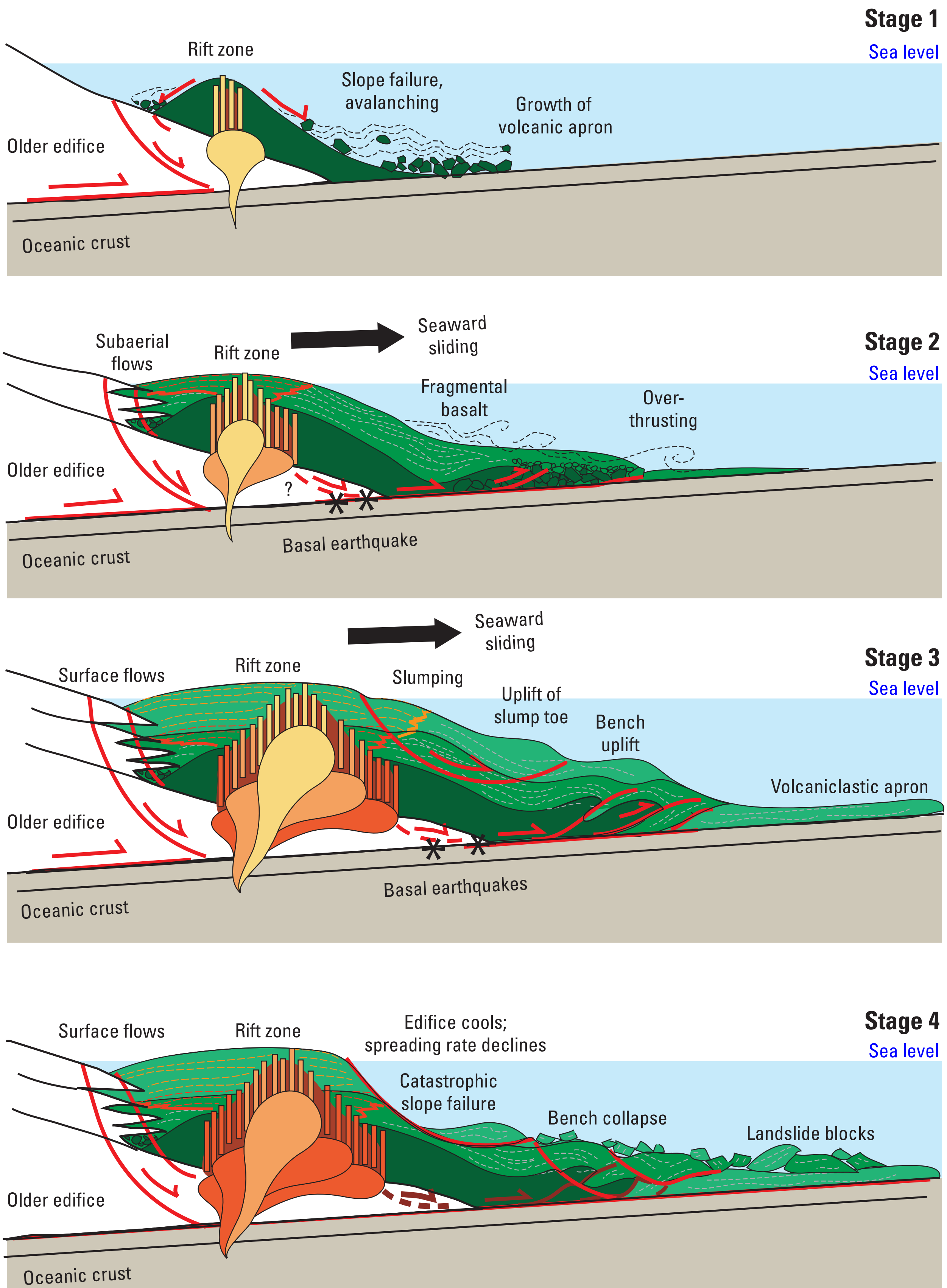

Figure 12. Cross-sectional diagrams showing proposed stages of island growth and degradation, based on integration of morphology, structural geology, and geophysical observations of Hawaiian volcanoes of different ages.

Hawaiian volcanoes and into the mechanisms by which stress may be stored within volcano flanks as they grow. Despite their sophistication, however, these broad-scale models provide few constraints on the mechanisms that suddenly trigger rapid slip of the flank seaward along its décollement and are believed to promote flank collapse.

Instability, as evidenced by the occurrences of rapid slip and the ubiquitous geologic deposits indicating sector collapse, results from rapid changes in an evolving force imbalance driving and resisting flank motion. These forces develop in response to processes on temporal and spatial scales that are challenging to measure and constrain, particularly within and along a deeply buried décollement. Processes that likely occur within the décollement layer, most notably coupling of induced pore-pressure and dilatancy, are capable of inducing rapid and catastrophic changes in the resistance to flank motion after decades of nearly steady or episodic creep. The mechanics of pore-pressure/dilatancy coupling provide an explanation for slow slip events, as well, since décollement models using known properties for saturated mixtures of turbidites and clays 200 to 300 m thick give timescales for slip events tantalizingly close to observed timing of sequences of events on the south flank of Kīlauea. Yet our understanding of these processes is woefully incomplete. Given our inability to precisely scale the forces affecting flank instability, the knowledge we have gained through recent studies only allows us to qualitatively make an educated guess as to the frequency and magnitude of flank instability during the growth of Hawaiian volcanoes.

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