RESEARCH ARTICLE

Susumu Umino · Miyuki Nonaka · Jim Kauahikaua

Emplacement of subaerial pahoehoe lava sheet flows into water: 1990 Kūpaianaha flow of Kilauea volcano at Kaimū Bay, Hawai`i

Abstract Episode 48 of the ongoing eruption of Kilauea, Hawai`i, began in July 1986 and continuously extruded lava for the next 5.5 years from a low shield, Kūpaianaha. The flows in March 1990 headed for Kalapana and inundated the entire town under 15–25 m of lava by the end of August. As the flows advanced eastward, they entered into Kaimū Bay, replacing it with a plain of lava that extends 300 m beyond the original shoreline. The focus of our study is the period from August 1 to October 31, 1990, when the lava buried almost 406,820 m^2 of the 5-m deep bay. When lava encountered the sea, it flowed along the shoreline as a narrow primary lobe up to 400 m long and 100 m wide, which in turn inflated to a thickness of 5–6 m. The flow direction of the primary lobes was controlled by the submerged delta below the lavas and by damming up lavas fed at low extrusion rates. Breakout flows through circumferential and axial inflation cracks on the inflating primary lobes formed smaller secondary lobes, burying the lows between the primary lobes and hiding their original outlines. Inflated flow lobes eventually ruptured at proximal and/or distal ends as well as mid-points between the two ends, feeding new primary lobes which were emplaced along and on the shore side of the previously inflated lobes. The flow lobes mapped with the aid of aerial photographs were correlated with daily observations of the growing flow field, and 30 primary flow lobes were dated. Excluding the two repose periods that intervened while the bay was filled, enlargement of the flow field took place at a rate of 2,440–22,640 square meters per day in the bay. Lobe thickness was estimated to be up to 11 m on the basis of cross sections of selected lobes measured using optical measurement tools, measuring tape and hand level. The total flow-lobe volume added in the bay during August 1–October 31 was approximately 3.95 million m^3, giving an average supply rate of 0.86 m^3/s.

Keywords Pahoehoe sheet flow · Submarine lava · Supply rate · Inflation · Kaimū Bay · Kilauea · Hawai`i

S. Umino (✉)
Shizuoka University, Institute of Geosciences,
Ohya 836, Suruga-ku,
Shizuoka, 422-8529, Japan
e-mail: sesumin@ipc.shizuoka.ac.jp
Tel.: +81-54-2384789
Fax: +81-54-2380491

M. Nonaka
University of Tokyo, Earthquake Research Institute,
Yayoi 1-1-1, Bunkyo-ku,
Tokyo, 113-0032, Japan

J. Kauahikaua
U.S. Geological Survey, Hawaiian Volcano Observatory,
P.O. Box 51 Hawai`i National Park,
HI 96718, USA

Introduction

When lava flows from land into water, it shows a variety of lava morphologies associated with volcaniclastic facies which result from interaction of the hot lava with water. `A`a flows commonly form a fan-shaped lava delta on the shore which extrudes fingers of lava into the water. Examples include the Izu-Oshima 1684 lava (Sakaguchi et al. [1987](#)), Miyakejima 1643 lava (Isshiki [1960](#)), 5,000-year BP Ohmuroyama lava, Higashi-Izu Volcano Group (Kuno [1954](#)) and Fuji 864-866 Aokigahara Lava (Obata and Umino [1999](#)). The Aokigahara Lava flowed down a valley to the Lake Motosuko and spread as an `a`a lava delta on the coastal plain. Although the majority of the lava delta was emplaced on dry land, the flow front entered into the lake and stagnated under water. Numerous flow lobes protruded from the flow front, which then inflated to form subaqueous flow-lobe tumuli (Obata and Umino [1999](#)). The elongate tumuli are 1–20 m wide and typically have a few flow lobes branching from the most inflated front of the fumulus. The lobes have also inflated to form another tumulus downslope. Parent and daughter tumuli are connected by a less-inflated narrow conduit, giving them the appearance of frog's fingertips. Unlike Hawaiian examples, the tumuli do not have any ooze-outs or outflows from

longitudinal or diagonal cracks. In contrast, the 1969 lava from Mauna Ulu, Kilauea, flowed into the sea as a 30-m wide `a`a flow, but no fan-shaped lava delta was formed (Moore et al. 1973). The flow maintained its integrity down to a depth more than 70 m, extruding several cylindrical lava tongues from one side of the flow below 25 m water depth. These lobes are about 1 m in diameter and 10–15 m long. Circular cross sections and highly fractured, bread crust surfaces on the lava lobes indicate that inflation took place after a thick crust had developed.

Fluidal pahoehoe flows commonly produce a large amount of volcaniclastic debris upon entering into water (e.g., Jones and Nelson 1970; Moore et al. 1973; Skilling 2002). Although many examples are reported from Iceland (Jones 1969; Furnes and Fridleifsson 1974), Gran Canaria (Furnes and Sturt 1976) and the Columbia River Basalt (Fuller 1931), there is little direct observation of active flows and the formation of clastic deposits under water (Moore et al. 1973; Tribble 1991). Unlike `a`a lava, some pahoehoe flows do not enter water directly, but are deflected along the shore. The 1971 flow fed by lava tubes from Mauna Ulu produced a delta of pahoehoe lava 1.5 km wide that extends 450 m beyond the old sea cliff (Moore et al. 1973). The surface of this delta is underlain by hummocky, tube-fed pahoehoe consisting of low tumuli and small ooze-outs of lava. At the ocean entry, narrow (<0.5 m wide) streams of lava were immediately quenched and fractured on contact with water, forming angular glassy fine sand and coarse hyaloclastite. However, streams more than 0.5 m wide were seldom totally shattered but instead maintained their coherence to feed lava tongues under water. The lava delta began to develop when small streams of lava emerged from tubes and spread across the newly forming black sand beach. Over the course of several days, the small flows coalesced to form a single pahoehoe flow. As a mound of pahoehoe was produced opposite to the largest feeder tube system, later flows diverted laterally off this mound, spreading along the base of the old sea cliff and widening the delta several hundred meters beyond its primary feeding tubes. This mode of growth also generated several prehistoric pahoehoe deltas along the south coast of Kilauea (Moore et al. 1973). On Surtsey, lava streams were witnessed to have stagnated and turned laterally along the coast when they approached the sea (Bár∂arson 1991; Thorarinsson 1967). Diversion happened when the flow was relatively slow, resulting in a dam of solidified lava along the seashore. Likewise, a branch of Fuji 864 Aokigahara lava formed interconnected flow-lobe tumuli that flowed along the complicated coastline upon entering the Lake Syojiko. On the contrary, lava lobes in Motosuko Lake broke out from the front of the `a`a lava delta and headed directly downslope.

It is of interest how coastal pahoehoe lava lobes are emplaced and why they behave differently from those on land. The majority of previous studies have focused on the formation of hyaloclastite and associated pillow lavas. Few have paid attention to the emplacement and structure of the pahoehoe lobes that entered into water. Different behavior of pahoehoe flows may well be affected by supply rates of lava, topography of the ground, and pre- and syn-eruptive cooling history of the flows (e.g., Cashman et al. 1999; Dragoni and Tallarico 1994; Kilburn 2000; Polacci et al. 1999). However, quantitative data on extrusion rates of lava flows are very limited because of the paucity of detailed records and difficulty in measuring thicknesses of active lava flows during growth of the flow field. Episode 48 of the Pu`u `Ō`ō-Kūpaianaha eruption of Kilauea, Hawai`i, offered an excellent opportunity to examine the behavior of lava that entered the sea. In March 1990, continuous extrusion of lava from the Kūpaianaha vent, 3 km downrift of the presently active Pu`u `Ō`ō vent, reached the coast and eventually buried Kaimū Bay under a plain of lava up to 11 m in thickness (Mattox et al. 1993; Figs. 1, 2). The lava formed a series of elongate, curved lobes which flowed along the embayed coastline or on the seaward side of the parent lobe (Figs. 2c, 3a).

We present a record of the growth of the Kaimū Bay flows based on direct observation combined with recent detailed field mapping and flow lobe measurements, and the calculations of rate of advancement and lava supply calculations, and we discuss the emplacement of the pahoehoe flows from the land into the sea.

Method

Observations of the growing flow field in Kaimū Bay from April to early November in 1990 were done from both the ground and air by the Hawaiian Volcano Observatory (HVO) of the US Geological Survey and the daily outlines
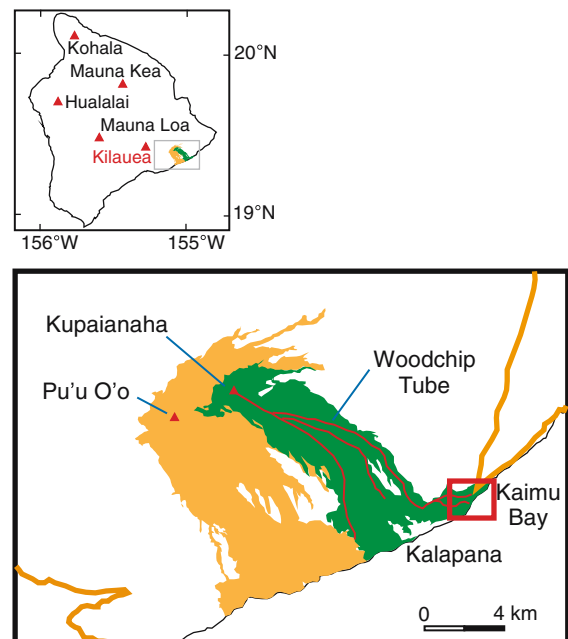

Fig. 1 1983–2001 lava from Pu`u `Ō`ō-Kū;paianaha vents on the East Rift zone of Kilauea, Hawaii. The 1986–1991 Kūpaianaha flows and their master lava tubes are shown as a *green area* and *red lines* (Mattox et al. 1993). The *rectangle* indicates the mapped area shown in Fig. 3
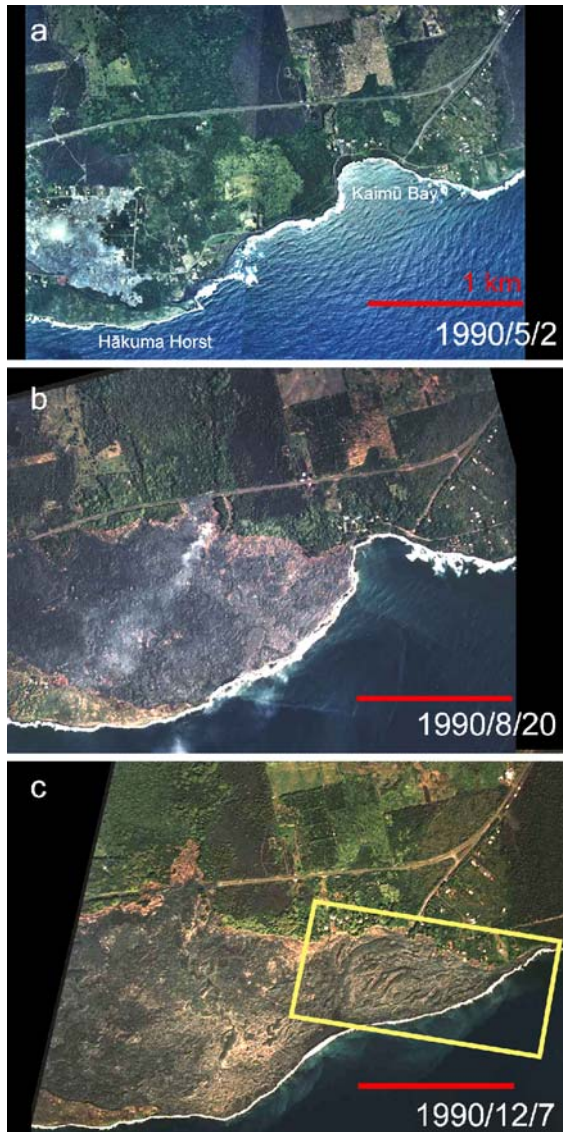
Fig. 2 Aerial photographs of Kalapana taken before (**a** May 2 and **b** August 20, 1990) and after (**c** December 7, 1990) burial of Kaimū Bay by lava. Kaimū Bay is shown in the *center right*. The *rectangle* in **c** is the location of maps in Fig. 5. Aerial photos courtesy of HVO

of the flow field and breakouts from active tubes were mapped on aerial photographs (Mattox et al. 1993). In August 2001 and 2002, we conducted field mapping and measurements of the inflated flow lobes in the former Kaimū Bay, now largely buried by the lava, with the aid of aerial photographs taken in February, May, August and December, 1990 (Fig. 2). We mapped and measured the positions and depths of major inflation cracks, large ropy wrinkles indicative of flow directions and the positions and depths of inflation pits by using a measuring tape. We observed the mutual relationships between adjacent flow lobes. Although the aerial photograph is slightly oblique, the frame on Fig. 2c has an area of 0.9237 km^2 before correction and 0.9300 km^2 after correction, which indicates calculated areas are underestimated by no more than 2%. Selected flow lobes were measured using optical instru-
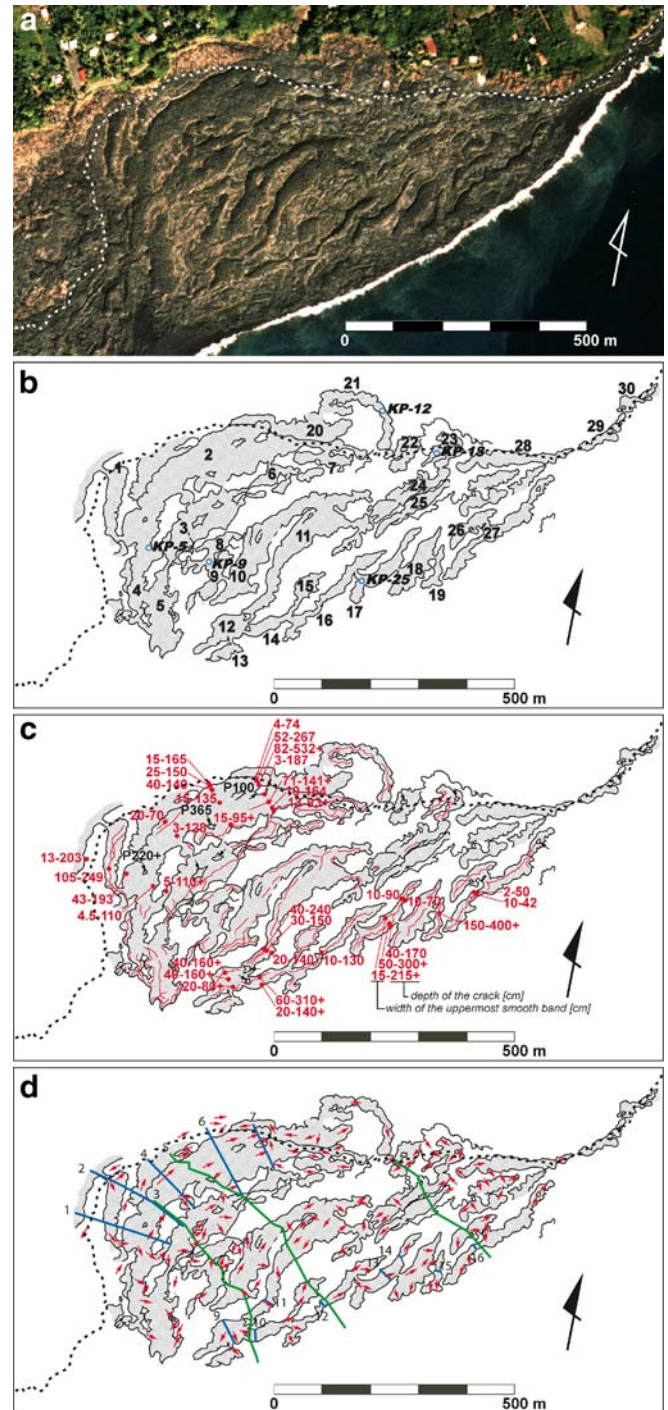

Fig. 3 Enlarged aerial photograph of flows filling Kaimū Bay (**a**) and maps showing outlines and structures of primary flow lobes (**b**–**d**). *Dotted line* represents the pre-eruption coastline. *Gray areas* on **b**, **c**, and **d** denote the primary lobes and *white spaces in between gray areas* are breakouts. The position of the mapped area is shown in Fig. 2. **b** Flow-lobe numbers (1–30) and some localities (*circles* with *KP-5, 9, 12, 13, 25*) are shown. **c** *Arrows* indicate inflation pits, some of which are labelled *P* with depth (in centimeter) of pit. *Thin red lines* are inflation cracks with numbers connected with a hyphen showing width of the uppermost brittle band on the crack wall (*left*) and depth of the crack (*right*) in centimeters. **d** *Arrows* indicate flow directions inferred from ropy wrinkles. *Blue and green lines with numbers* indicate positions of cross sections shown in Fig. 6

ments (Rangefinder, Laser Atlanta Optics, USA; and LaserAce300, Measurement Devices Limited, UK) capable of resolving three-dimensional coordinates of a distant point relative to the instrument, measuring tape and hand level in order to get 3-dimensional shapes and volumes of the lobes. Altitude above sea level was determined by measuring several known reference sites of which altitudes are shown on a 1:24,000-scale topographic map of Kalapana, HI. HVO provided all available daily outlines mapped on pre-eruption aerial photographs. The daily outlines were mapped without GPS reference using only local landmarks and are therefore least accurate for flow outlines that extended offshore. We compared and correlated the growth lines of the flow field shown by the USGS mapping with the present flow lobes and determined the initial dates of emplacement of 30 primary flow lobes. The flow outline of August 20 between Hakuma Horst and

Kaimū Bay by the USGS maps is 40–50 m behind the real coastline shown by the aerial photograph of the same date. We have adjusted these flow outlines to fit the true coastline of August 20. The USGS flow outlines filling Kaimū Bay are almost consistent with the growth patterns shown by the connecting primary flow lobes, except those on October 5–8 which cross the inflated primary lobes. These outlines indicate the reversed order of emplacement of the lobes determined by the field mapping. We have slightly expanded the western part of the outlines seaward to fit the growth pattern of the primary lobes. Daily aerial extent in the mapped area (Fig. 4) was calculated by pixel counting on digitized maps with Adobe Photoshop. Textures of selected rock samples were examined under a conventional optical microscope and JEOL JXA-733 electron probe microanalyzer of the Center for Instrumental Analysis, Shizuoka University.

Fig. 4 Growth of lava flow field in Kaimū Bay area from August 1 to October 31. *Scale bars* on the *right bottom* are 500 m long. Pre-eruption coastline is shown by the *dotted line*. Selected lavas which flowed on the following specific dates are shown in *black*; **a** August 1, **b** August 6, **c** August 8, **d** August 15, **e** September 2–22, **f** September 24–26, **g** September 27, **h** September 28, **i** September 29, **j** September 30, **k** October 6, **l** October 7, **m** October 9, **n** October 11, **o** October 17, **p** October 31. The lava which flowed during intervening periods are shown by *orange areas*. Some minor flows were emplaced during periods not shown in the figures, however, their areal extent can be deduced from comparing the *black and orange areas* of the two adjacent figures such as Fig. 3c,d. Based on maps by HVO personnel on the ground as the flows were emplaced, and modified after further mapping in 2001 and 2002
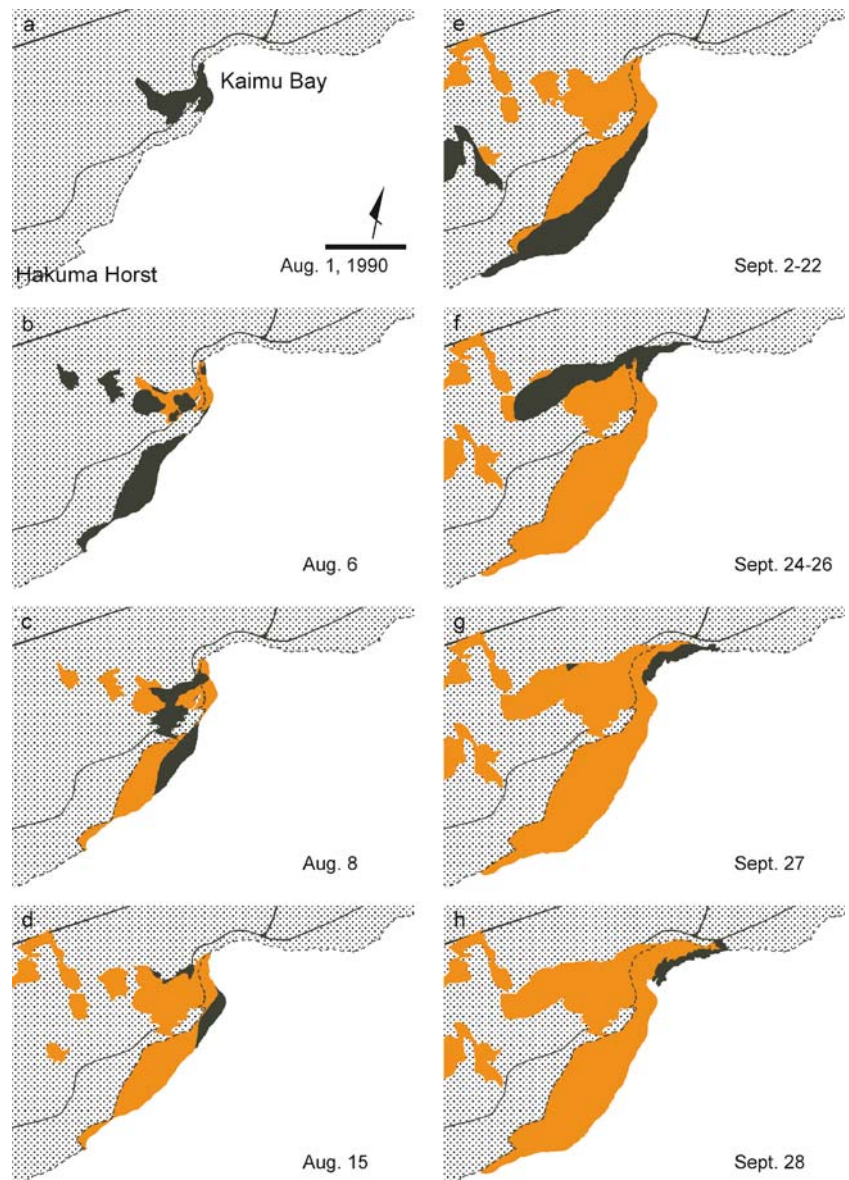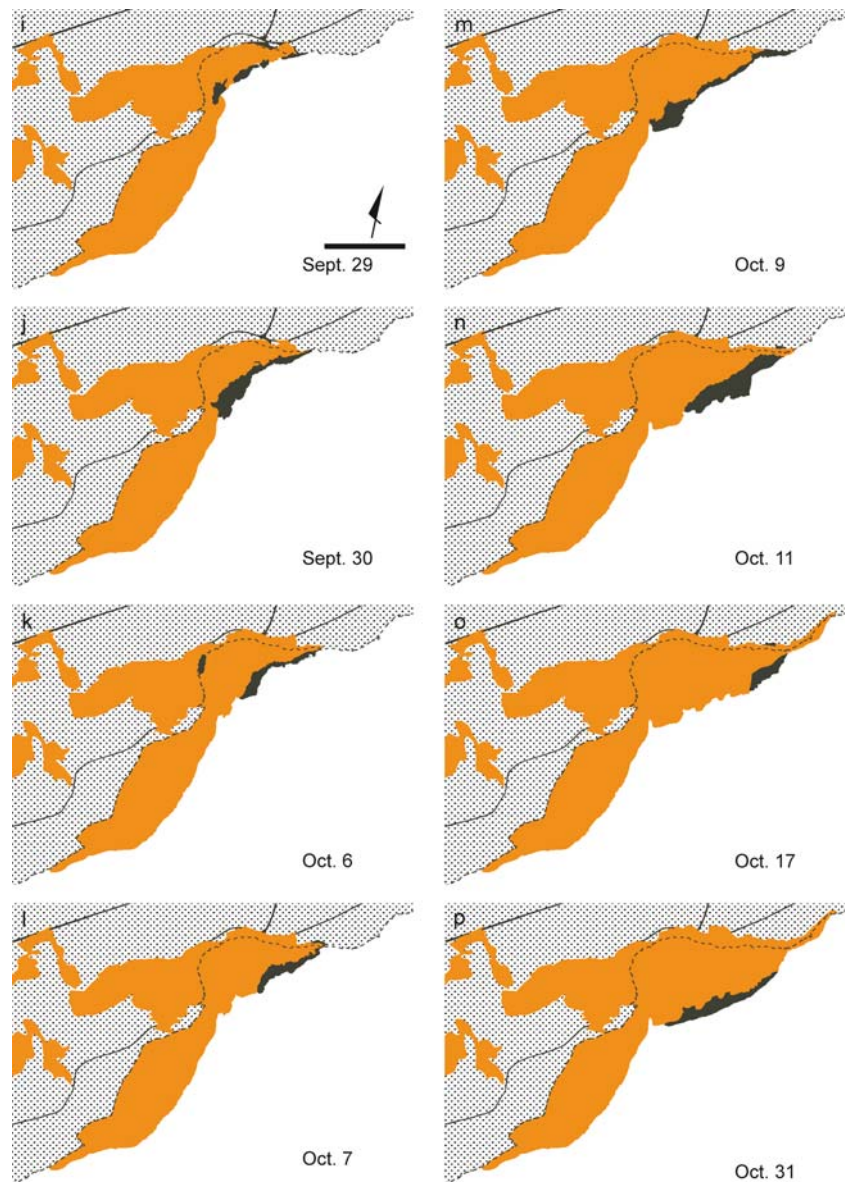
Fig. 4 (continued)

Kūpaianaha eruption and Kaimu Bay flow field

The current eruption of Kilauea volcano has been the most voluminous and longest lava eruption from the east rift zone of Kilauea in the past 600 years, and has produced 2.2 km^3 of lava that covered 105 km^2 by the beginning of 2002 (Heliker and Mattox 2003). The eruption commenced at Pu`u `Ō`ō on January 3, 1983, and for the next 3 years it formed a 255-m high cinder-and-spatter cone by episodes of fountaining that took place every 3–4 weeks. High effusion rates during these eruptive episodes yielded mainly channeled `a`a flows which never reached the coastline due to their short durations of effusion. The eruptive style changed from channelized `a`a flows to quiet extrusion of lava when the vent migrated 3 km downrift to Kūpaianaha in July 1986 (Heliker and Mattox 2003). The Kūpaianaha eruption lasted for the next 5.5 years and formed broad fields of pahoehoe flows which gradually

developed tube systems 12 km long toward the south-eastern coast. A branch of surface flow that broke out from a pre-existing "Kalapana tube" in March 1990 slowly grew to form another tube system (Woodchip tube). The Woodchip tube obtained an increased supply of lava in early July to feed rapidly advancing new flows (Woodchip flow) to Kalapana Gardens. By August 1990, the Woodchip flow began to fill in Kaimū Bay, continuing through the end of October 1990. It is this period that we are primarily interested in describing in detail.

Filling of Kaimu bay

We hitherto use "the flow field" for the specific area shown in Fig. 4. This area includes subaerial outflows presumably fed through the same tube system which fed the Kaimū Bay flows after August 1. However, the area is only a part of the

entire flow field of the Pu`u `Ō`ō-Kūpaianaha eruption. The majority of lava entered Kaimū Bay in two periods: from August 1 to 15 and from September 24 to October 17 (Fig. 5). During the first period, lava invaded only 100–150 m offshore along the western coast of the bay at a low rate of areal extension (2,440 m^2/day). Although the total area of the flow field during the first period was 504,530 m^2, only 46,160 m^2 of the bay was covered with new lava.

Rapid filling really started during the second period of activity. On September 24 after a 40-day hiatus in lava supply to Kaimū Bay, a large breakout occurred from the Woodchip tube 700 m west of the original coastline (Fig. 4f). This breakout flowed along the northern coastline of the bay, building the new coastline outwareds 100 m for the next 2 days. Lava rapidly extended seaward at a rate of 20,750 m^2/day until October 1 and filled 189,300 m^2 of the bay. By October 2, the flows from the Woodchip tube had covered 1,050,810 m^2 of the area in Fig. 4 at an average rate of 25,070 m^2/day.

After a short halt during October 2–4, the flows continued farther into the sea until October 17 (Fig. 4k–o), after which seaward expansion temporarily stopped until October 31. The average rate of areal extension was 22,640 m^2/day and 153,170 m^2 of the bay was covered during October 5–11. On October 31, new flow lobes entered the ocean to the east and southeast (Fig. 4p). By October 31, 95% of the flows had been emplaced.

Structures of primary flow lobes

The Kaimū Bay flows are today composed of rows of curved, elongate gentle mounds which run subparallel to the original coastline. These mounds are inflated flow lobes with flat top "inflated sheet flows" (Hon et al. 1994) or "lava rise" (Walker 1991) and large tumuli with a triangular cross-sectional shape(Figs. 2c, 3a, 6). The mounds are bounded by topographic lows filled with uninflated or slightly inflated pahoehoe lobes. A number of small (<1 m in width) flow lobes that leaked out from the inflated lobes cover the majority of the original surface of the inflated lobes and parts of the breakouts. Most of these later small flow lobes are only slightly inflated. The majority of the tumuli and inflated sheet flows are interconnected and appear to have been emplaced lobe by lobe by breakouts from their parental inflated flow lobes, which formed a long-lived tube system lasting for more than 48 days. These inflated flow lobes are termed "primary flow lobes", the emplacement style of which is our prime interest.

Primary lobe sections

Inflated primary flow lobes near the original coast of Kaimū Bay curve along the old shoreline and are concave seaward. Later flow lobes emplaced farther offshore tend to be straight, as the lava that buried the inner part of the bay had straightened the shoreline. Unburied portions of the primary flow lobes range from a few tens to 100 m in width and extend up to 400 m in length. The primary lobes stand out 2–5 m above the surrounding breakouts. Lobe cross sections (Fig. 6) gradually change in shape from an obtuse triangle to a steep sided pancake as they become wider. Lobes with triangular cross sections have a major crack along the highest axis of the lobe (Figs. 6c, 7b). Where the width of unburied portion exceeds 25 m, at least two major cracks are present along margins of the flat-lying upper surface of the lobe. The cracks opened as the upper surface was uplifted from the surrounding outflows (Fig. 7a).

Fig. 5 Cumulative area of lava flows shown in Fig. 3 (upper) and that of the Kaimū Bay flows (lower) for August 1 (day 1) through October 31 (day 92)
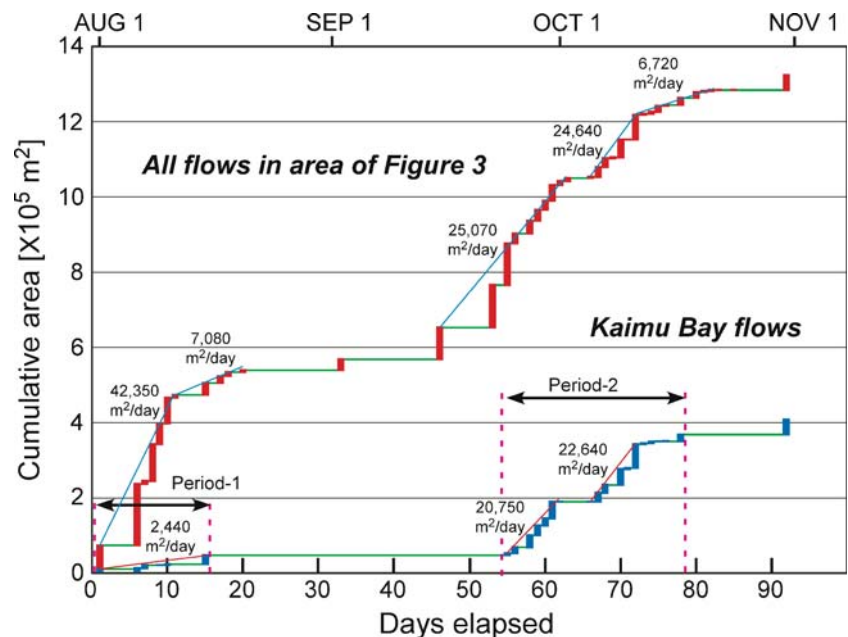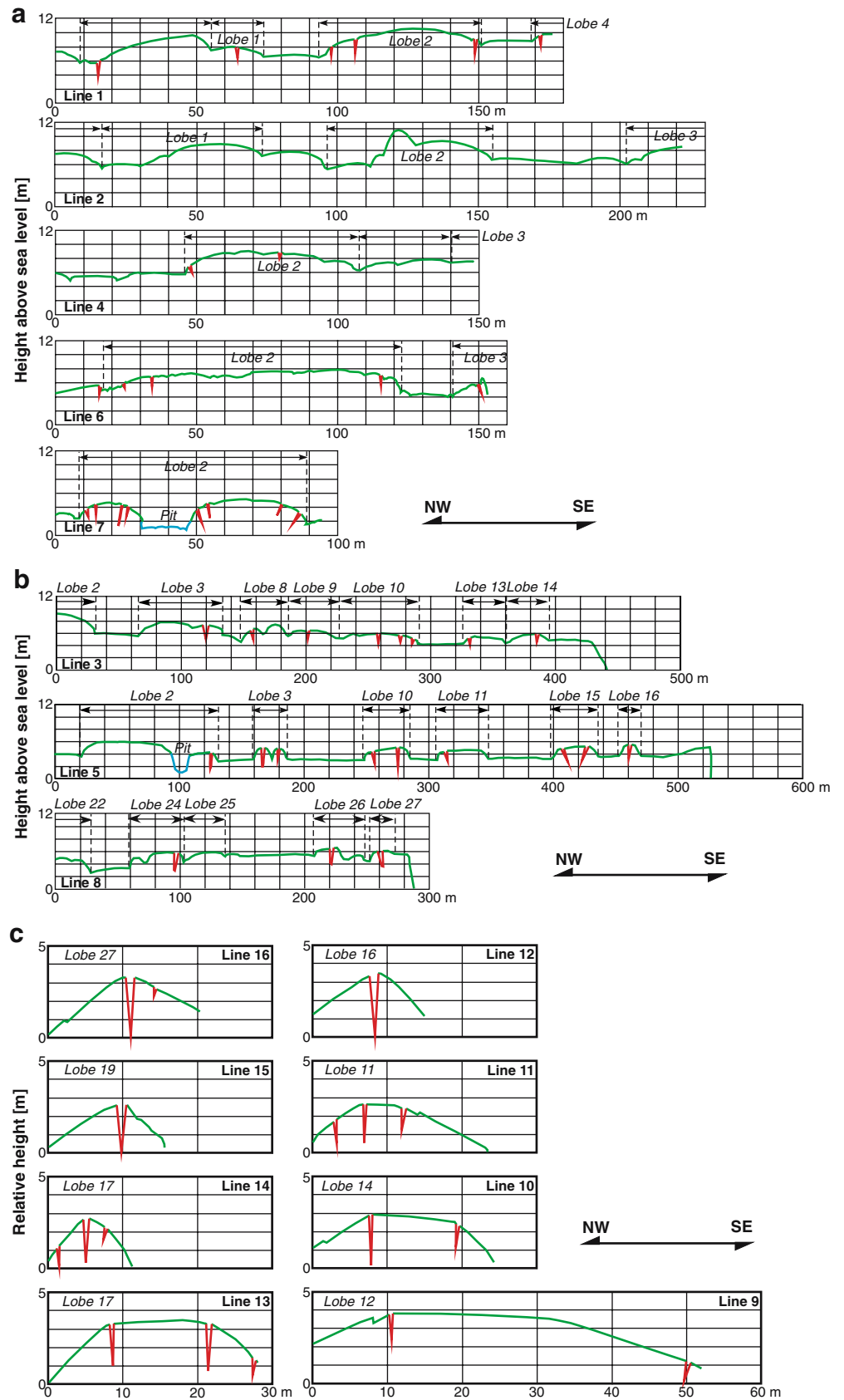
Fig. 6 Transverse sections of flow lobes in Kaimū Bay along lines shown in Fig. 3d. Note that vertical exaggeration differs in each figure. **a** Sections of lobes emplaced near the landward side of the bay. **b** Sections of multiple lobes along long traverses. **c** Sections of lobes on the sea side of the Kaimū Bay flows
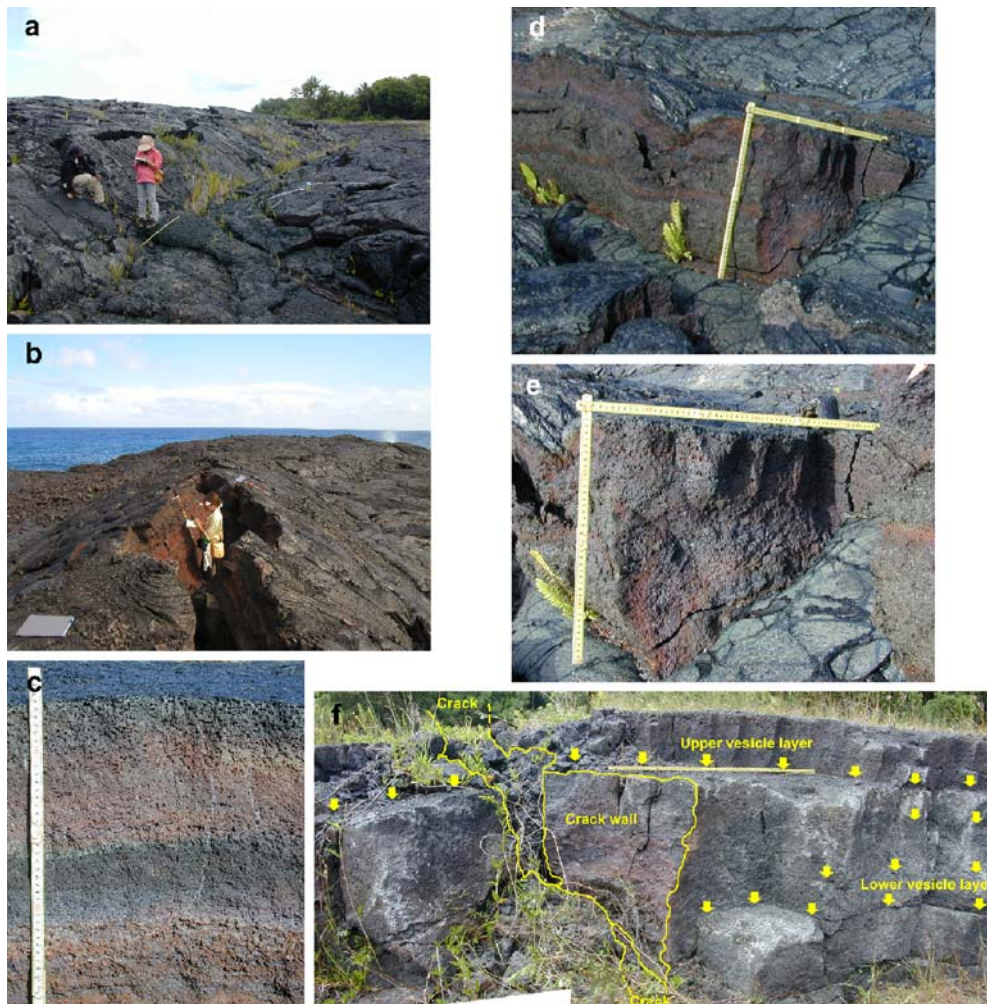
Fig. 7 Structures and textures of flow lobes. **a** Inflated primary flow lobe (*left*) standing out from later breakouts (*right*). A marginal crack developed along the edge of the primary lobe. **b** Axial crack develops on a small flow lobe showing a *triangular cross section*. **c** Bands on an inflation crack wall, showing the upper smooth bands and lower spinose, vesicular bands (*lower red bands* below 32 cm on the scale). Although the upper 32 cm is vesicular, the crack wall surface between the vesicles is flat and smoother than the lower bands. **d** Earlier longitudinal axial crack (*center left*) intersected by a later transect crack (*right of the scale*). The scale is about 40 cm long. Subparallel bands occur on the axial crack, whereas the intersecting crack wall shows distorted bands (**e**). **f** Vesicle-rich layers near the upper surface of an inflated flow lobe of October 1990 exposed in a road cut. The inflation crack is filled with lava rubble. One wall of the crack is exposed. Note the upper vesicle-rich layer is cut by the crack, whereas the lower vesicle-rich layers terminate 15–20 cm from the crack. **g** Ropy wrinkles on buckled-up crust on the side of lobe 3 near the end of a conduit connecting lobes 2 and 3 (at KP-5). (**h**) Initial ropy wrinkles on a primary lobe (lobe 6) at *KP-9*. (**i–l**) Volcaniclastic deposit on a primary lobe (lobe 17) at **KP-25**. **i** Wrinkled and fragmented surfaces of the lobe covered with glassy clasts derived from the shattered vesicular crust beneath the disrupted wrinkled surfaces. **j** Spinous glassy clasts formed by fragmentation of highly vesicular, wrinkled crust. Note the *black vitreous wrinkles* on one side of clasts (right arrow), and smooth convex surfaces on the other side (*left arrow*) which were parts of spherical walls of large bubbles formed beneath the thin surface crust of flow lobes. **k** Partially disrupted, vesicular crust covered with spinose glassy clasts. Note the hollow interior of the crust (*arrows*) formed by large bubbles inside and the oxidized surfaces. **l** Inflation crack wall on lobe 17, showing oxidized bands beneath the glassy surface (*arrows*) covered with hydroclastic deposits

Inflation cracks

Primary flow lobes develop major cracks along the periphery, margins and axis of the lobe and minor cracks may also develop both transverse and subparallel to the axis and margins of the lobe, either crossing or bifurcating from the major marginal and axial cracks (Figs. 3c, 6, 7). Circumferential cracks along the periphery of the lobe are among the first ones to open surrounding the lobe through which the entire upper crust of the lobe is uplifted, but completely buried under later breakouts except those seen in inflation pits. Other major cracks opened when the primary lobes began to inflate after most breakouts fed from the lobes stopped leaking, as they transect small pahoehoe lobes that cover the surface of the main lobes. Most cracks extend subvertically as deep as 3 m, and then incline either toward the center or the margin of the lobes.

Crack walls show bands composed of different colors and textures, as observed in inflation cracks of other sheet flows and tumuli emplaced on land (e.g., Hon et al. 1994; Umino et al. 2002; Walker 1991). These bands are not layers but surface textures which are not tracable to flow

Fig. 7 (continued)
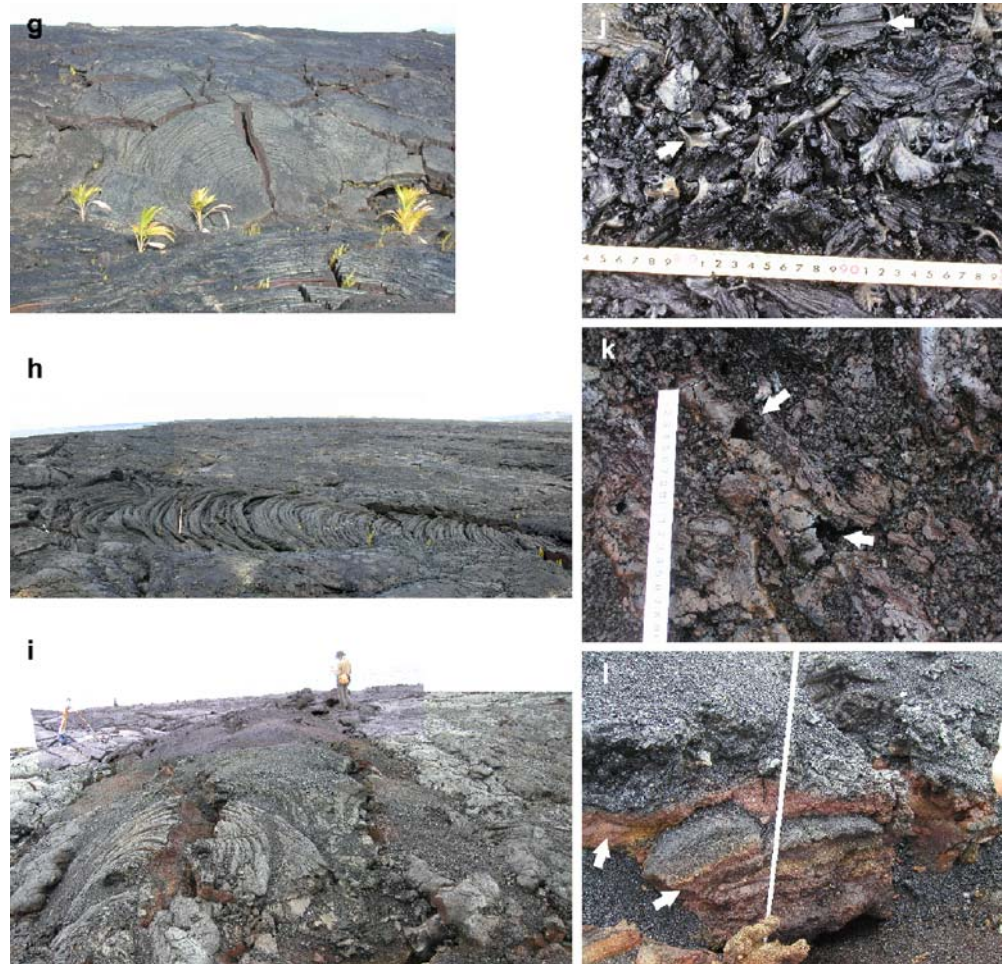
interiors more than a few milimeters from the crack wall. Below the glassy chilled margin on flow tops, crack walls have a smooth, columnar jointed band with spherical vesicles, which grades into rough surfaces to spinose, vesicular bands (Fig. 7c; Walker 1991). Farther downwall, glassy lava shelves may be attached to the crack wall. All the smooth, rough and spinose bands may be repeated downwall, with interspersed glassy lava shelves. Band width varies from 1 cm up to 150 cm, and the uppermost smooth band is typically thicker than the rough and spinous bands downwall. However, the width of the uppermost smooth band may vary laterally even on the same crack wall (Fig. 5c). Spinose bands are less oxidized than smooth bands, indicating rapid quenching. Textures of the bands on a crack wall form due to varying opening rates of the crack and rheological responses of lava to stress when the crack opens (Anderson and Fink 1992; Anderson et al. 1999; Hon et al. 1994; Umino et al. 2002). Hon et al. (1994) concluded that the initial crack formed by brittle fracture of the crust. The fracture then propagated into partially molten, viscoelastic crust that stretched to form the spiny glassy veneers. Continuous crack growth results in thick glass-veneered bands, whereas intermittent growth results in alternating bands of crystalline basalt and spiny glass veneers. Glass shelves attached on crack walls form

where lava squeezes up at the crack tip as it penetrates into molten lava.

Inflation pits

Some primary lobes have pits ranging from 5 to 40 m in diameter and 2–5 m in depth. They are found either near the lobe margins where the flow front bifurcated into two branches which met together ahead, or in the central depression of a large lobe formed where two large flow lobes coalesced (Fig. 3c). The pit floors consist of hummocky pahoehoe lobes partly covered by blocks of broken lava crust which formed parts of the upper crust of the surrounding lobes. Some pit walls are inwardly inclined slabs on the edges of the surrounding inflated lobes, whereas other walls are overhanging and truncated layers of superposed pahoehoe lobes, which are densely coalesced together and show bands as seen in inflation cracks. The positions of pits on a lobe, the presence of the truncated banded pit walls, and the lack of connection to any drained lava tubes are consistent with the interpretation that the pits are spaces which were not inflated whereas the surrounding flow lobes were (Walker 1991; Hon et al. 1994). These features were called "lava-rise pits" by

Walker (1991) and "inflation pits" by Hon et al. (1994). The largest lobe (lobe 2) emplaced along the original shoreline inflated more than 5 m, judging from the depths to the bottom of inflation pits (Fig. 6a,b).

Flow directions

Adjacent primary lobes are connected at one or more points by narrow conduits through which lava was supplied. Many lava conduits, less inflated than the primary lobes, preserve large ropy wrinkles spanning almost the entire conduit width. Active inflating sheet flows form ropy wrinkles by folding of hot, flexible crust in response to locally increased supply rates of lava to the flow front or sudden topographic drops (Hon et al. 1994). The common presence of large ropy wrinkles on the lava conduits feeding primary lobes suggests that discharge of lava through the narrow conduits was rapid. Flow directions inferred from these sets of ropy wrinkles are consistent with the overall flow directions indicated by mapping (Fig. 3d). Where there is no narrow conduit exposed between primary lobes, large ropy wrinkles may be present on the tilted, landward side of a primary lobe indicating flow towards the lobe axis or into the lobe (Fig. 7g). At these locations, a buried lava conduit can be inferred.

Some primary flow lobes also preserve large ropy wrinkles up to several meters wide, spanning almost the entire width of the lobe (Fig. 7h). These wrinkles are presumed to be original surfaces of the primary lobes, as they imply flow directions generally consistent with the observed growth patterns of the primary flow lobes.

Flow directions of lava as deduced from the ropy wrinkle sets on the primary lobes show that the lava flowed into Kaimū Bay from the northwest and extended to the north along the coast, and then progressed offshore to the south (Fig. 3d). Most flow lobes on the southeastern margin of the filled bay were fed from the lobe at the southwestern end and progressed to the northeast subparallel to the present coast. This is precisely the sequence of events described by witnesses.

Quench texture of flow lobes

Most flows that entered water were rapidly cooled and could not fully inflate, and hence were covered by later outflows. Few primary flow lobes with original large ropy wrinkles on the surface have quench textures due to contact with seawater. Lobe 17 (KP-25; Fig. 3b) near the southern end of the flow field is one example. The lobe is 200 m long and 30 m wide, about two thirds of which has a ropy wrinkled original surface partially covered by hydrovolcanic debris (Fig. 7i). The lobe has a poorly vesicular, glassy surface, whereas the vesicularity is 40–70 vol% just a millimeter beneath the surface and vesicles are 1–10 mm in diameter. Large bubbles 1–10 cm in diameter may be present 0.5–1 cm below the ropy wrinkled surface. The vesicular glassy crust is 15–20 cm thick, much thicker than

the glassy crust of subaerial flow lobes, reflecting efficient cooling by water. The remaining one third of the lobe surface is covered by fragile, fragmental lithic and glassy clasts, some of which grade laterally from sheared, wrinkled and vesicular crust into spinose, twisted glassy clasts a few millimeters to tens of centimeters in diameter (Fig. 7j). Fine glassy clasts are surrounded by concavely curved, smooth bubble walls. The curvature of the bubble walls indicates the original bubble diameter of 1–10 mm. A minor amount of brown transparent limu-o-Pele is also present. Unlike common hyaloclastite (Fisher and Schmincke 1984; McPhie et al. 1993), smaller clasts do not have angular, polygonal shapes but instead, have bubble-wall shapes with spines at intersections of coalesced bubbles (Fig. 7j). The presence of clasts of disrupted wrinkled crust suggests an explosive origin, but unlike the typical products of phreatomagmatic explosions, they lack fine ash (Fisher and Schmincke 1984; Wohletz 1983; Mattox and Mangan 1997). These observations suggest that mild explosions took place when wrinkled crust was sheared and torn apart under water, allowing seawater access into the hot lava interior. Seawater in contact with hot lava formed large bubbles that burst, disrupting semimolten lava crust into vesicular, spinose glassy clasts.

The absence of any visible volcaniclastic mounds suggesting the former presence of a littoral cone indicates that lobe 17 was far from the littoral hydrovolcanic explosions and that any cone was probably eroded before the emplacement of the adjacent lobe 18. Although most ropy wrinkled clasts are glassy, some clasts are reddish to orange oxidized. When lava surface solidifies to form crust, the level of molten lava beneath the crust sometimes drops to form a new lava surface that is often wrinkled and oxidized (Fig. 7k). The presence of oxidized clasts among the hydroclastic deposits shows that lava crust formed on land was disrupted and transported later into water. We suggest that the disruption of the crust of lobe 17 took place in the surf zone, where the lava surface were alternately in and out of water.

Discussion

Supply rate of lava

The rate of areal extension in the flow field shown in Fig. 4 during August 1-October 31, 1990, ranges from 6,720–42,350 m^2/day (Fig. 5) and averages 19,200 m^2/day (net average rate=area divided by the emplacement period of flows). This rate is comparable to but smaller than that for all subaerial breakouts at Kalapana during May 31-September 5, 1990 (Mattox et al. 1993), and reflects the fact that there was other activity in the Kalapana area and on the Kūpaianaha flow field during the filling of Kaimū Bay. Finally, 406,820 m^2 of the bay was filled by the lava at a rate of 2,440–22,640 m^2/day (Fig. 5), and the area in Fig. 3 covered by the lava is 1,323,960 m^2. According to several long-time Kalapana residents, the water depth of Kaimū Bay was 10–20 feet (3–6 m) before burial by the

lava. The bottom of the inflation pits is slightly above the sea level, suggesting that the bases of the inflated lobes are approximately at the sea level. Hence, a significant volume of deposits exists between the original seafloor and the bottom of the exposed inflated flow lobe. The average height above sea level of the flow lobes in the bay is 5 m. The depth of the bay before burial by the lava was 10–20 ft, and is assumed to be 5 m on average. Then, the total volume of the lava (including the submerged deposit) is calculated to be 3,953,600 m^3. Because lava supply into the bay was cut off during the 40-day-long repose period, this volume of lava was emplaced in 52 days, provided the flow lobes ceased inflation and attained the final volume by October 31, 1990. Then, the average supply rate of lava to the flow field is estimated as 0.86 m^3/s, which is comparable to that measured during the emplacement of a single Kalapana pahoehoe sheet (Hon et al. 1994).

There is a positive correlation between the supply rates and the volumes of pahoehoe and sheet flow lobes, because the lobe volume is limited by cooling (Umino et al. 2002). Four separate primary lobes in Kaimū Bay were emplaced in a day: lobe 2 on September 27, lobe 23 on October 5, lobe 10 on October 6, and lobe 29 on October 12. Initial volumes of primary flow lobes before inflation when the flow lobes were emplaced by outbursts are poorly known due to the lack of records of outlines of single flow lobes and the lack of precise thickness data for the lobes. Primary lobes were emplaced in shallow water as deep as, or slightly deeper than, the initial flow-lobe thicknesses, which are approximated to be <1 m. Because primary lobes would have inflated as they spread, and the actual emplacement period for the flow lobes are less than a day, the estimated volumes and supply rates of lava are minimal: 31,710 m^3 and 0.37 m^3/s for lobe 2, 22,420 m^3 and 0.26 m^3/s for lobe 10, 680 m^3 and 0.01 m^3/s for lobe 23, and 2,100 m^3 and 0.02 m^3/s for lobe 29. These values are comparable to those of subaerial pahoehoe sheet flows at Kalapana (Hon et al. 1994) and are consistent with estimates for other subaerial and submarine flow-lobe tumuli from Hawaiian volcanoes (Umino et al. 2002).

Alongshore emplacement of primary flow lobes in Kaimū Bay

When lava encountered the sea, it did not go straight into water but flowed subparallel to the coastline and formed an elongate flow lobe curved along the coast. A burst of new lava eventually came out from one or more points of an inflated lobe, either distal, midpoint or proximal to its source point. The new flow lobe again flowed along the coast or on the seaward side of the parent lobe. Repetition of this process extended the flow field to cover the entire Kaimū Bay. Such a style of emplacement of flow lobes at ocean entry is not uncommon (e.g., Moore et al. 1973; Bár∂arson 1991; Thorarinsson 1967), but is not readily explained. We propose that the subparallel growth of the primary lobes was caused by topographic control of the

submerged delta below the lavas and damming up lavas fed at low extrusion rates.

By analogy with many examples of pahoehoe lava deltas (e.g., Jones and Nelson 1970; Moore et al. 1973; Skilling 2002), the submarine portion of the delta would have largely buried topography in the bay before emplacement of the primary lobes. Therefore, pre-emplacement topography could not affect the lava flow directions. However, the submarine delta itself limited the maximum extent and controlled the subhorizontal shape of a primary flow lobe which was subsequently emplaced and inflated on the delta. It is suggested that the growth of a new primary lobe along and on the seaward side of its parent inherited the formation of the elongate delta along the parent lobe. Low supply rates of lava may enhance quenching and shattering of lava upon entering the ocean, as was observed at the Banana flow ocean entry on June 2, 2004 (Hawaii Volcano Observatory 2004). This process will form a barrier of solidified lava and debris which diverts later flows laterally along the coast, forming a shallow or dry bench along the shoreline for a later breakout to spread over. During the 1971 Mauna Ulu flow ocean entry (Moore et al. 1973), small streams of tube-fed pahoehoe spread across the newly forming beach, coalescing into a single pahoehoe flow that slowly advanced in amoeboid fashion. This pahoehoe became a barrier that diverted later flows laterally, spreading along the base of the old sea cliff. On the contrary, high supply rates at the ocean entry west of Kalapana in 1989 delivered lava under water as a channeled stream on a 25°–30° slope of loose basaltic debris (Tribble 1991). The estimated volumetric rate of lava for the stream was 0.7 m^3/, which is double that estimated above for the large primary lobe in Kaimū Bay. It is likely that a low extrusion rate (<0.4 m^3/s) for the primary lobes resulted in lava dams which diverted the later flows laterally along the coast.

Solidification of the upper crust of inflating flow lobes

Electrical conductivity measurements of inflating flow lobes demonstrated that the upper crust of the flows thickens at a rate 40% higher than the lower crust (Kauahikaua et al. 1998). However, the growth rate of the upper crust is not always uniform over an inflating flow lobe. As it inflates, a lobe develops a downward-propagating axial crack. Because the flow interior is overpressured by up to 0.2 MPa, opening of the crack tip unloads the semi-molten lava exposed at the tip and causes degassing and vesiculation of the lava (Cashman and Kauahikaua 1997; Hon et al. 1994). Degassing raises the liquidus of the lava, resulting in a large degree of undercooling of the semi-molten lava beneath the crack tip. In addition, the opening crack provides new surfaces through which cooling of the lava core proceeds. Such cooling should be efficient in widely open cracks on tumuli such as those of the 1960 Kapoho lava (Walker 1991), and on those formed under water (Obata and Umino 1999). Both degassing and cooling will promote crystallization of the lava core

surrounding the crack tip, locally deepening the solidification front of the upper crust. An example of distorted banding appears in a crack on lobe 23 at KP-13 near the original northeastern coast of the bay. The lobe has an axial crack intersected by a later crack, both of which are partially buried by later outflows (Fig. 7d,e). Most bands run subparallel to the lobe surfaces. However, bands that occur on younger cracks intersecting older deeper bands may be curved and turn subparallel to the older cracks. Samples P1-P4 were taken from the crack wall with the distorted bands along a line subparallel to the flow-lobe surface and traversing the distorted bands to see how the groundmass texture varies away from the axial crack (Fig. 8). Sample P1 adjacent to the axial crack has a groundmass mainly consisting of feathery plagioclase and clinopyroxene, tiny magnetite grains, glass and sporadic microlites of plagioclase and clinopyroxene. The groundmass of sample P2, 3 cm from P1, has more plagioclase and clinopyroxene microlites and the interstices are filled by intergrown platy plagioclase and clinopyroxene, and dendritic magnetite. Groundmasses of samples P3 and P4, 6 and 9 cm farther away from P1, respectively, have well-developed thick microlites of plagioclase and clinopyroxene and only minimal amount of dendritic magnetite between the microlites. The increasing number density (3,200, 4,600, 12,700, 80,900/mm^2) and the decreasing length (11.2±9.4, 10.7± 8.9, 7.2±3.1, 3.1±5.0 μm) of the groundmass plagioclase from P4 to P1 can be ascribed to an increasing degree of undercooling from the lobe interior toward the axial crack (e.g., Lofgren 1980). The distortion of banding on the younger crack wall near the older crack, and the increase in undercooling toward the latter, can be explained if the thermal structure within a flow lobe was modified adjacent to an opening crack and cooled more efficiently the semi-molten lobe interior (Obata and Umino 1999). Another interpretation is that, when the older crack propagated into the flow lobe, the overpressured interior lava exposed at the crack tip degassed (Hon et al. 1994), and in turn enhanced degassing and undercooling of the lava surrounding the crack tip. Both mechanisms promote solidification of lava adjacent to the opening crack tip ahead of the solidification front beneath other portions of the lobe crust away from the crack.

Enhanced cooling around the inflation cracks is also inferred from flow lobes of October 1990 which were emplaced along the northern coastline of the bay. A road cut through a gentle dome-like lobe shows a subvertical inflation crack >1 m deep and <0.2 m wide. Subhorizontal vesicle-rich layers and blisters occur 20–60 cm below the flow top. The vesicles are 1–3 cm in diameter and concentrated in layers up to 5 cm thick (at KP-12; Fig. 7f). Such vesicle-rich layers and blisters are interpreted as being formed by concentration and coalescence of vesicles trapped in the solidification front beneath a growing upper

Fig. 8 SEM images showing textural variations of samples P1, P2, P3 and P4 from the crack wall with distorted oxidation bands at KP-13 (Fig. 7d,e). The samples were taken along a line subparallel to the flow lobe surface and perpendicular to the axial crack. Width of the SEM images is 0.34 mm. In P1, *dark gray rectanglar crystals* on the *upper right* are plagioclase microphenocrysts. *Light gray crystals* with brighter Fe-rich rims are clinopyroxene. Bundles of very thin plagioclase laths are mantled by feathery clinopyroxene with interstices filled by tiny magnetite. In P2, spaces between plagioclase and clinopyroxene microlites are filled by thin plagioclase laths and small stubby clinopyroxene <0.015 mm. In P3 and P4, the plagioclase and clinopyroxene crystals are more equant and larger than those in P1 and P2
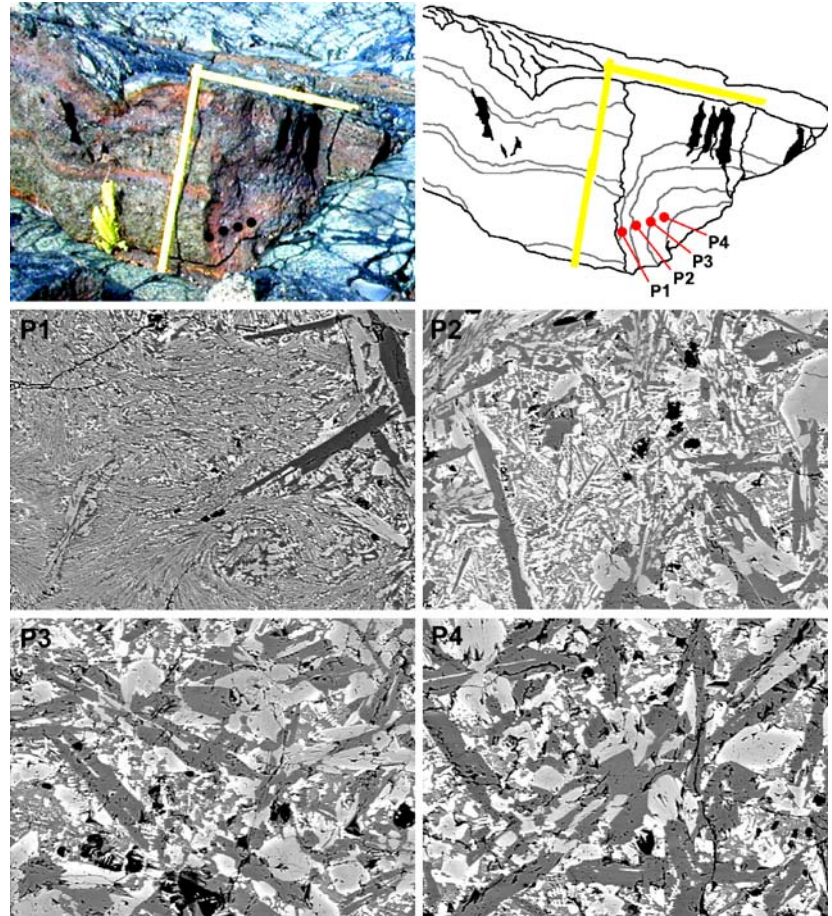
crust (Self et al. 1996, 1997; Thordarson and Self 1998). The vesicle-rich layers extend continuously 5–15 m on both sides of the crack, and the uppermost, largest vesicle-rich layer 20–25 cm below the lobe surface is truncated by the crack. Below the uppermost vesicle-rich layer, the lava within 50 cm of the crack wall is more vesicular, whereas lower vesicle-rich layers become obscure and taper out within 15–20 cm of the crack. This difference indicates that the crack opened after the uppermost 20–25 cm of the lobe had solidified, and propagated deeper ahead of the solidification front before vesicles accumulated to form the lower vesicle-rich layers beneath the solidification front. Either heat loss through the crack wall cooled the lobe interior more efficiently adjacent to the wall than away from it, or decompression at the crack tip caused degassing of still molten lava inside, resulting in a larger under-cooling. Disturbance of the solidification front surrounding an inflation crack is more pronounced as the crack opens wider and exposes more of the lobe interior as has been observed for some dome-like tumuli and lava domes

(Anderson and Fink 1992), and for those formed under subaqueous conditions (Obata and Umino 1999).

Emplacement model of the Kaimu bay flows

We interpret the emplacement and growth of the Kaimū Bay lava lobes as follows: (1) rapid extrusion of lava which broke out from an inflated flow lobe entered shallow water and spread over the nearly horizontal shallow seafloor of an already established delta (Fig. 9a); (2) solidified lava tongues and lobes became barriers on the brink of the delta, diverting later lava along the edge of the delta and the parental flow lobe (Fig. 9b); (3) as the lava lobe spread, the cooling lobe developed a surface crust that retarded advance of the flow front, and inflation of the flow lobe began (Fig. 9c). Many small breakouts leaked from circumferential cracks near the base of the lobe; (4) as the circumferential cracks were healed by the solidified breakouts, the lava supplied into the flow lobe inflated the lobe

Fig. 9 Model for growth of the Kaimū Bay lava. **a** Lava which broke out from an inflated flow lobe entered shallow water and spread over a nearly horizontal delta topset. Solidified lava dammed up on the edge of the delta diverted subsequent lava laterally, resulting in elongate flow lobes running subparallel to the coast. **b** As the lava lobe expanded over the topset, cooling of the lobe developed surface crust and retarded advancement of the flow front, initiating inflation of the flow lobe from behind the front. **c** Most lava flows that leaked out from stagnated, inflating flow lobes were quenched and fragmented to form the subdeltaic deposits. **d** As the circumferential cracks were healed by the solidified breakouts, the lobes inflated to form a domal uplift with deep axial cracks
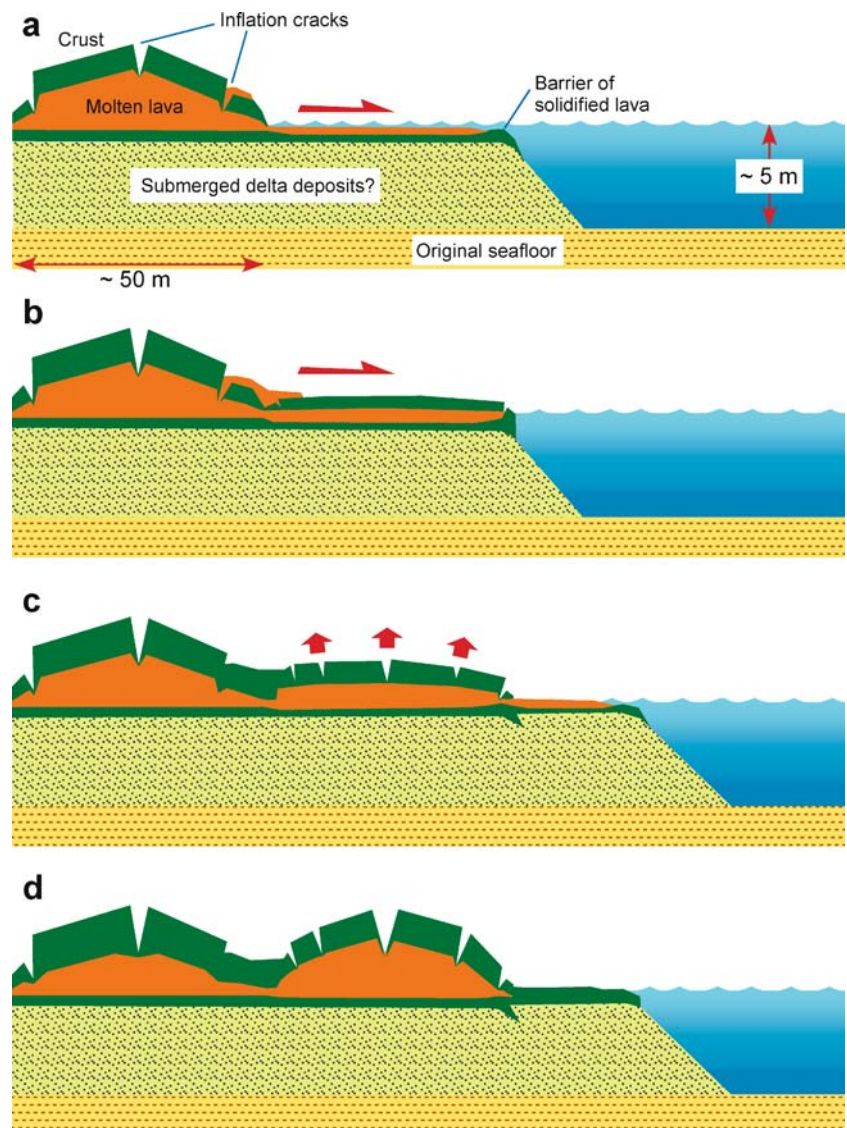
until another burst of lava came out from the inflated lobe, extending a new primary lobe over the shallow surface of the delta (Fig. 9d). Breakouts could have resulted in a short pause in flow lobe inflation and caused slight deflation while the lobe expelled lava via the breakout, but the flow lobe would have resumed inflating until the lava supply to the lobe was finally cut off (e.g., Cashman and Kauahikaua 1997; Kauahikaua et al. 1998). Because all primary flow lobes in Kaimū Bay were fed through interconnected tubes, most flow lobes would have been active and continued to inflate until the end of October.

Summary

The Woodchip flow from Kūpaianaha vent first entered to Kaimū Bay on August 1, 1990, and buried 406,820 m^2 of the bay under a 10-m thick layer by October 31. Growth of the Kaimū Bay flows took place in two periods: August 1–15 and September 24-October 17. The average rate of aerial extension of the Kaimū Bay flows was 2,440–22,640 m^2/day and the total flow-lobe volume is estimated to be 3.95 million m^3. The average supply rate was 0.86 m^3/s for a period of 52 days, which is comparable to that of Hawaiian subaerial sheet flows (Hon et al. 1994) and submarine flow-lobe tumuli (Umino et al. 2002).

Most lava flows that leaked out from stagnated, inflating flow lobes were quenched and fragmented to form the subdeltaic deposits. Solidified lava dammed up on the edge of a shallowly submerged delta diverted subsequent lava laterally, resulting in elongate flow lobes running subparallel to the coast. Inflation structures such as hummocky lobe surfaces, and inflation cracks and pits are ubiquitous on the primary flow lobes. Major breakouts from inflated lobes formed new flow lobes which were connected by narrow conduits to their parental lobes, resulting in a long, meandering tube system. Topographic lows between the inflated flow lobes were buried by later breakouts which scarcely inflated. Unburied portions of the primary flow lobes range from a few tens to 100 m wide and up to 400 m long.

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