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INVITED REVIEW

Geology, geochemistry and earthquake history of Lō`ihi Seamount, Hawai`i's youngest volcano

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Abstract

A half-century of investigations are summarized here on the youngest Hawaiian volcano, Lō`ihi Seamount. It was discovered in 1952 following an earthquake swarm. Surveying in 1954 determined it has an elongate shape, which is the meaning of its Hawaiian name. Lō`ihi was mostly forgotten until two earthquake swarms in the 1970s led to a dredging expedition in 1978, which recovered young lavas. The recovery of young lavas motivated numerous expeditions to investigate the geology, geophysics, and geochemistry of this active volcano. Geophysical monitoring, including a real-time submarine observatory that continuously monitored Lō`ihi's seismic activity for 3 months, captured some of the volcano's earthquake swarms. The 1996 swarm, the largest recorded in Hawai`i, was preceded earlier in the year by at least one eruption and accompanied by the formation of a ~300-m deep pit crater, Pele's Pit. Seismic and petrologic data indicate that magma was stored in a ~8–9 km deep reservoir prior to the 1996 eruption.

Studies on Lō`ihi have altered conceptual models for the growth of Hawaiian and other oceanic island volcanoes, and refined our understanding of mantle plumes. Petrologic and geochemical studies of Lō`ihi lavas showed that the volcano taps a relatively primitive part of the Hawaiian plume, producing a wide range of magma compositions. These compositions have become progressively more silica-saturated with time, reflecting higher degrees of partial melting as the volcano drifts toward the center of the hotspot. Helium and neon isotopes in Lō`ihi glasses are among the least radiogenic found at ocean islands, and may indicate a relatively deep and undegassed mantle source for the volcano. The north–south orientation of Lō`ihi rift zones indicates that they may have formed beyond the gravitational influence of the adjacent older volcanoes. A new growth model indicates that Lō`ihi is older, taller and more voluminous than previously thought. Seismic and bathymetric data have clarified the importance of landsliding in the early formation of ocean island volcanoes. However, a fuller understanding of Lō`ihi's internal structure and eruptive behavior awaits installation of monitoring equipment on the volcano.

The presence of hydrothermal activity at Lō`ihi was initially proposed based on nontronite deposits on dredged samples that indicated elevated temperatures (31 °C), water temperature, methane and ^{3}He anomalies, and clumps of benthic micro-organisms in the water column above the volcano in 1982. Submersible observations in 1987 confirmed a low temperature geothermal system (15–30 °C) prior to the 1996 formation of Pele's Pit. The sulfide mineral assemblage (wurtzite, pyrrhotite, and chalcopyrite) deposited after the pit crater collapsed are consistent with

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hydrothermal fluids with temperatures $>250\,^{\circ}C$, although the highest measured temperature was $\sim 200\,^{\circ}C$. Vent temperatures decreased to $\sim 60\,^{\circ}C$ during the 2004 dive season indicating a waning of the current phase of hydrothermal activity.

Keywords: Hawaii; Loihi; Volcano; Seamount; Petrology; Geochemistry; Earthquakes

1. Introduction

Lō`ihi Seamount, the youngest and smallest Hawaiian volcano, has had a remarkable impact on our understanding of oceanic island volcanism and mantle plume processes since its rediscovery in 1978. It is considered the type example of the early phase of growth of plume-related oceanic island volcanoes (Moore et al., 1982) and has been the focus of numerous multi-disciplinary studies. Its proximity to the island of Hawai`i (Fig. 1) offers an excellent opportunity to monitor an active submarine volcano (Klein, 1982; Caplan-Auerbach and Duennebier, 2001b). Here, we review previous results and present new data on the geology, geochemistry and earthquake history of Lō`ihi, and correct the record on the first published report on the volcano.

2. Hawaiian geologic setting

The Hawaiian Island chain is one of the most isolated land masses on the planet, some 3800 km from the nearest continent. This isolation has contributed to the chain's distinct geological and biological character. A famous 19th century American writer considered Hawai`i to be "The loveliest fleet of islands that lies anchored in any ocean" (Twain, 1872). This fleet of islands and seamounts, the Hawaiian-Emperor chain, is anchored in the central Pacific basin at ~19°N. The chain extends ~6100 km from Meiji seamount near Kamchatka in the north to Lō`ihi seamount south of the island of Hawai`i (Fig. 1). It is the longest volcanic chain on Earth, with at least 129 distinct volcanoes (Clague, 1996). The trend of decreasing age to the south was first

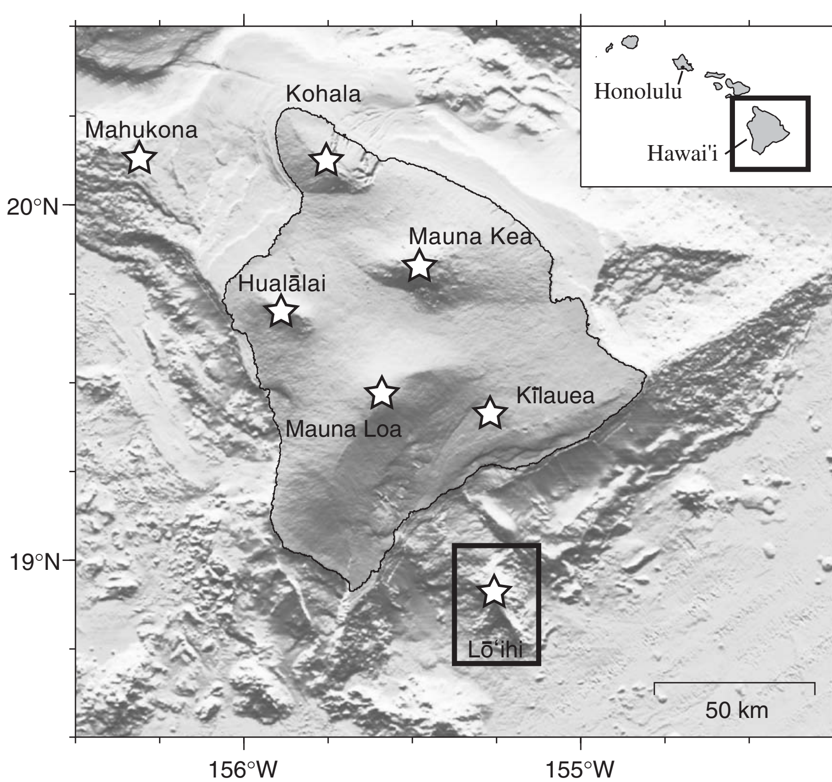

Fig. 1. Shaded relief of merged topography and bathymetry data (NGDC coastal model, available online at http://www.ngdc.noaa.gov/mgg/coastal/grddas10/grddas10.htm for the Island of Hawaii. Artificial illumination is from due north. The seven major volcanoes that form the island and its flanks are labeled, and their summits are indicated by a star symbol. Box around Lō`ihi indicates area shown in Fig. 3. Inset: location of this figure in relation to the major islands of the Hawaiian archipelago and its capital, Honolulu.

recognized by early Hawaiians in their oral tradition of the fire goddess Pele, who moved southward along the island chain with her fire (Westerveldt, 1916) causing successively younger eruptions to the south. Early western explorers to Hawai`i also noted the apparent decreasing age of the islands to the south (e.g., Dana, 1891). The overall age progression of the islands has been confirmed in several studies using radiometric isotopes (e.g., Clague and Dalrymple, 1987; Garcia et al., 1987).

The linear orientation of the Hawaiian-Emperor chain, with its prominent bend, and the age progression of its volcanoes led to the hypothesis that it was formed over a stationary mantle plume (Wilson, 1963; Morgan, 1972). The ages and orientation of the volcanoes were used to infer the rate and direction of motion of the Pacific plate (~10 cm/year towards the northwest for at least the last 15 m.y.; Clague and Dalrymple, 1987; Garcia et al., 1987). However, some aspects of the original hotspot model are controversial (e.g., is it fixed?). New paleomagnetic results on samples obtained by drilling indicate that the hotspot may have drifted ~4 cm/year during late Cretaceous to early Tertiary times (81–47 Ma) as the Emperor seamounts were formed (Kono, 1980; Tarduno et al., 2003). In contrast, other geophysical approaches suggest the hotspot was fixed then and now (Wessel and Kroenke, 1997).

Various evolutionary sequences have been proposed for the growth of Hawaiian volcanoes starting with Stearns (1946). These model sequences have evolved with new discoveries, including the revelation that Lō`ihi is the youngest Hawaiian volcano (Moore et al., 1982). A current popular scheme begins with the preshield stage (Fig. 2), lasting for ~250,000–300,000 years (Guillou et al., 1997) and producing alkalic magmas (Moore et al., 1982; Garcia et al., 1995). Although only observed at Lō`ihi and possibly Kīlauea (Lipman et al., 2002), this stage is thought to be at the core of all Hawaiian volcanoes (e.g., Clague and Dalrymple, 1987). As the volcano moves closer to the center of the hotspot and its source experiences higher temperatures and degrees of partially melting, the magma composition switches to tholeiitic (Garcia et al., 1995). Perhaps 50–100,000 years later, the volcano emerges above sea level, forming a subaerial shield volcano. After another ~100,000 years, the growing volcano reaches the size of Kīlauea (Quane et al., 2000). Vigorous activity persists for another ~700,000 years before the volcano enters the post-shield stage (Fig. 2). It is now ~1.25 m.y. old and has drifted off the center of the hotspot. Melting temperatures and the degrees of partial melting decrease during the post-shield stage causing magma compositions to gradually switch back to alkalic (Feigenson et al., 1983; Frey et al., 1990). A rapid decline in eruption rate occurs over the next 250,000 years, which is accompanied by an abrupt shift

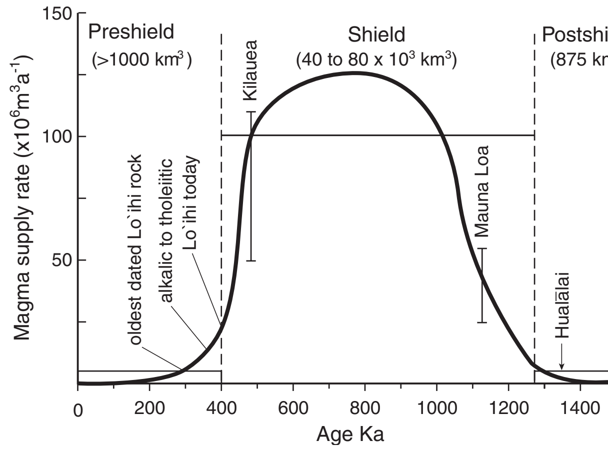

Fig. 2. Growth history model for a Hawaiian shield volcano. This composite model is based on volume estimates for each stage (boxes). Magma supply rate estimates (vertical bars) for Kīlauea (Pietruszka and Garcia, 1999a), Mauna Loa (Wanless et al., in review), and Hualalai (Moore et al., 1987) are shown for comparison. Our re-evaluation of the age of Lō`ihi has resulted in longer duration and large volume estimates for the preshield stage than proposed by Guillou et al. (1997).

to more fractionated lava compositions (hawaiites to trachytes) on some volcanoes such as Kohala (Spengler and Garcia, 1988), as magmas pond and cool at greater depths (~30 km) before eruption (Frey et al., 1991). After ~1.5 m.y. of growth (Fig. 2), the volcano dies, having formed one of the largest topographic features on Earth (up to 13 km in height and 80,000 km^3 in volume for Mauna Loa). Many, but not all, Hawaiian volcanoes experience a period of renewed volcanism that occurs 0.6–2.0 m.y. after the end of post-shield volcanism (e.g., Tagami et al., 2003). The lavas produced during this rejuvenated stage are generally strongly silica-undersaturated and tend to be explosive (Winchell, 1947; Walker, 1990). However, not all Hawaiian volcanoes follow this sequence. Some lack post-shield and or rejuvenated stages (e.g., on the island of O`ahu, Ko`olau volcano is missing post-shield stage lavas and Waianae has no rejuvenated lavas; Macdonald et al., 1983). For more on the geology of Hawaiian volcanoes, see Clague and Dalrymple (1987).

3. Discovery and early work on Lo`ihi

Lō`ihi Seamount is located ~35 km south of the island of Hawai`i (Fig. 1). The first appearance of this bathymetric feature in the literature was on the US Coast and Geodetic Survey chart 4115 in 1940, which was included in a summary of the geology of Hawai`i (Stearns, 1946). However, no specific mention was made of this distinct topographic high. The seamount is one of many that surround the Hawaiian Islands, some of which have been dated by K–Ar methods as Cretaceous in age (e.g., Dymond and Windom, 1968) and probably formed near the East Pacific Rise (EPR) (e.g., Engel and Engel, 1966).

A large earthquake swarm in 1952 first brought attention to Lōʻihi Seamount. Macdonald (1952) noted that epicenters for these earthquakes plotted on and near the seamount, which he suggested was a shield volcano lying along the extension of a trend that includes the two active Hawaiian volcanoes, Mauna Loa and Hualalai. Thus, Macdonald (1952) deserves credit for first proposing that Lōʻihi seamount was an active volcano. However, the east–west distribution of the epicenters and the lack of recorded

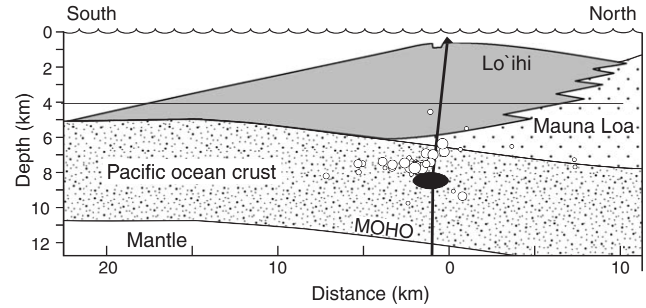

Fig. 4. Cross section of Lōʻihi Volcano drawn without vertical exaggeration showing the inferred basement for the volcano including southern flank of the island of Hawaiʻi, the Cretaceous (~105 Ma; Waggoner, 1993) Pacific oceanic crust and the mantle. The best-located earthquakes related to the 1996 eruption, with circle size proportional to relative magnitude (Caplan-Auerbach and Duennebier, 2001a), and the inferred intermediate depth (8–9 km; Garcia et al., 1998a) magma chamber (black ellipse) are shown. Earthquake locations have horizontal errors averaging ~2 km; vertical errors average ~0.6 km. The location of the new Pele's pit crater is indicated by a notch near the inferred vent location for the 1996 eruption breccia.

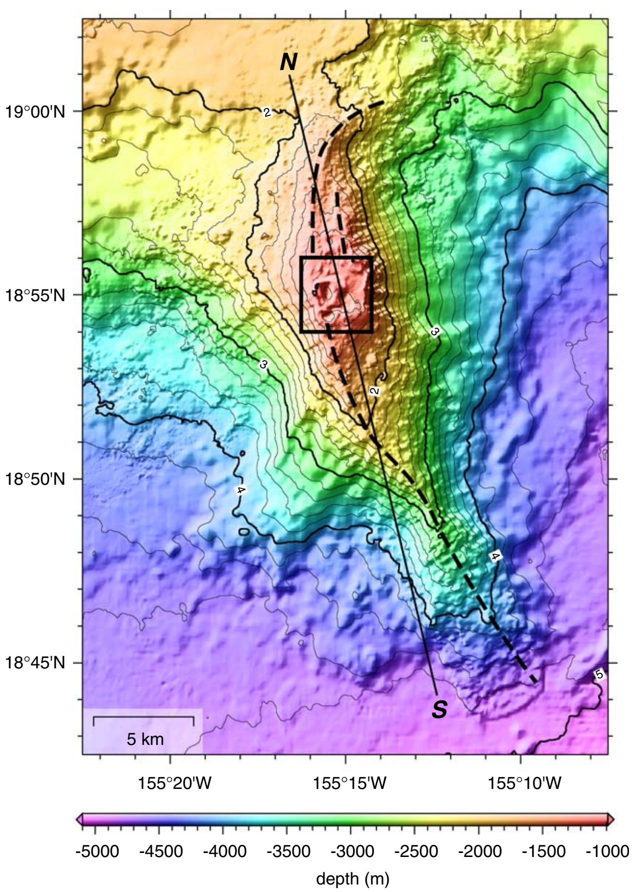

Fig. 3. Bathymetric contour map with shaded relief of Lōʻihi Seamount. Map data are from various surveys compiled and processed by J.R. Smith, F.K. Duennebier and T. Duennebier (see Smith et al., 2002). Contour interval is 200 m with annotations every 1 km. Artificial illumination is from the northwest. Rift axes are marked by dashed lines. Note the well-defined south rift zone curves to the east and the two-pronged north rift. The longer western segment of the north rift curves to the northeast. Box shows the location of Fig. 5, a detailed map of the summit area. The north (N) and south (S) line shows the location of the cross section in Fig. 4.

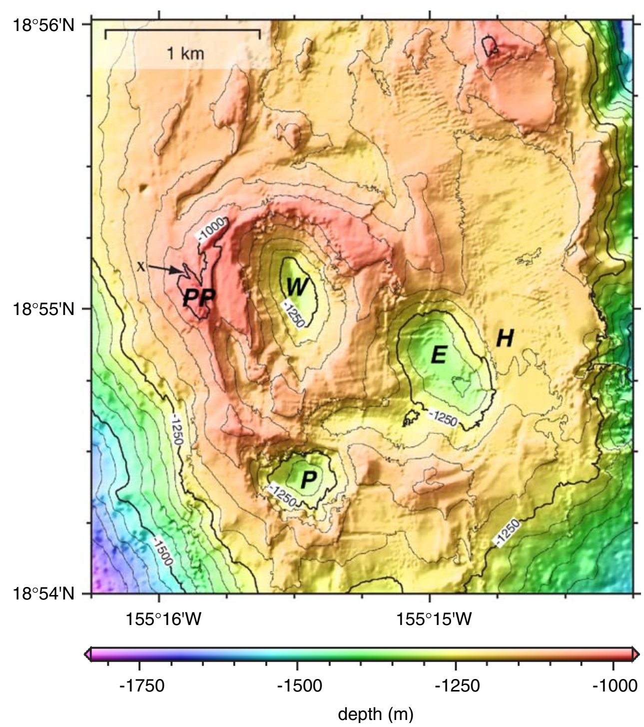

Fig. 5. Bathymetric contour map with shaded relief of the Lōʻihi Seamount summit. Contour interval is 50 m with annotations every 250 m, and artificial illumination from the northwest. The pit craters are labeled, W (West), E (East) and P (Pele's). The location where 1996 eruption breccia samples were collected is shown by the "x" just north of Pisces Peak (PP). The former location of the Hawaii Undersea Geological Observatory (HUGO) is shown by the "H". See Fig. 3 for location of this figure. Data sources as in the Fig. 3 caption.

volcanic tremor on seismic stations distant from the volcano led Macdonald (1952) to conclude that the 1952 earthquake swarm was related to faulting rather than an eruption.

In recognition of its elongated shape (Fig. 3), created by its nearly parallel north and south rift zones, the volcano was named Lo`ihi after the Hawaiian word for "to extend, to be long" by Mary Pukui and Martha Hohuhe of the Bishop Museum in Honolulu, and Gordon Macdonald of the US Geological Survey's Hawaiian Volcano Observatory (Emery, 1955). In addition to reporting a new name, Emery (1955) presented a new bathymetric survey and a suggestion that this seamount was the youngest Hawaiian volcano, a southern extension of the Hawaiian chain.

Lo`ihi was largely ignored and even classified as an 'older volcanic feature' on some subsequent geologic maps of the island of Hawai`i (e.g., Moore and Fiske, 1969) until two earthquake swarms in the 1970s (Klein, 1982) prompted a marine expedition in 1978 to survey (bathymetry, gravity, magnetics, photography and high-resolution reflection profiling) and sample the seamount to determine if it was a young Hawaiian volcano. This was confirmed by photographs from a camera towed across the seamount's summit that revealed fresh coherent pillow lava. Its youth was reaffirmed by a single dredge haul on Lo`ihi's summit that recovered ~300 kg of pillow lavas with fresh glassy crusts and thin red-brown, iron-rich coatings containing nontronite of possible hydrothermal origin (Moore et al., 1979). A 1979 expedition undertook more extensive sampling with 17 dredgehauls on the summit and rift zones (Moore et al., 1982). A wide range of rock types was recovered including alkalic lavas, which had not been found on the neighboring active volcanoes, Kīlauea and Mauna Loa. The presence of alkalic lavas led to the suggestion that Lo`ihi represented the earliest phase of Hawaiian volcanism, reflecting lower degrees of partial melting (Moore et al., 1982). A 1980 expedition found extensive hydrothermal fields associated with recent lava flows in the summit area (Malahoff et al., 1982), further supporting the hypothesis that Lo`ihi is a young, active volcano. The first high-resolution multi-beam bathymetric survey of the volcano showed Lo`ihi to be a significant feature rising at least 3 km from the deep ocean floor to 980 m below sea level (mbsl), with a summit area containing two prominent pit craters (275 and 256 m deep) and two sub-parallel rift zones extending north and south forming a 30 km long volcano (Malahoff et al., 1982). These pioneering studies ushered in two decades of intense exploration of Lō`ihi punctuated by its 1996 eruption (Loihi Science Team, 1997). Highlights of these expeditions are presented below along with new observations on the geology, geophysics and geochemistry of Lō`ihi.

4. Morphology and structure

The morphology of Lō`ihi Seamount has been delineated by numerous bathymetric surveys of the volcano (e.g., Emery, 1955; Malahoff, 1987; Fornari et al., 1988; Eakins et al., 2003). Lō`ihi is built on the flanks of two other active Hawaiian shield volcanoes (Mauna Loa and Kīlauea), which sit on ~105 Ma Pacific ocean crust (Epp, 1984; Waggoner, 1993; Fig. 4). The maximum thickness of Lō`ihi has been estimated using regional bathymetry at ~3.5 km (Garcia et al., 1995), comparable in height to Europe's largest volcano, Etna. However, as discussed below, it is likely that Lō`ihi began forming 400,000 years ago, and therefore, formed on only a thin veneer of debris from the island of Hawai`i, and is considerably thicker (~5 km; Fig. 4) than previously thought. Lō`ihi's summit consists of a small (12 km^2) platform at ~1200 mbsl with several large cones and three, 300–370 m deep pit craters (Fig. 5). These craters are similar in diameter to some Kīlauea subaerial pit craters but are ~100 m deeper. Hawaiian pit craters are thought to form when magma in a shallow reservoir is erupted or intruded laterally within the volcano (e.g., Okubo and Martel, 1998). The presence in the West Pit crater walls of thick sections of columnar jointed lava (>20 m), rare in Hawaiian shield volcanoes, suggests that the crater was repeatedly formed and filled. The Western Pit crater has been interpreted to be older because it contains alkalic basalts and is truncated by the East Pit crater (Fornari et al., 1988; Garcia et al., 1993). Pele's pit formed in 1996 south of the West and East Pit craters following an intense earthquake swarm (Loihi Science Team, 1997; Caplan-Auerbach and Duennebier, 2001a). The volume of lava erupted in 1996 is too small to explain the large volume of Pele's Pit (~0.1 km^3; Malahoff, 1998). The 1996 earthquake data do not show a pattern indicating lateral magma drainage. Thus, the cause of the 1996 collapse of Pele's Pit is unknown. However, the sequence of pit crater formation from older West, intermediate East, and younger Pele's pit may be part of an overall southward shift in the locus of Lō`ihi's volcanism (Fig. 5). The western flank of the summit, where unaffected by mass wasting, dips ~14°. In contrast, the deeply dissected east flank of the volcano dips 35–40°, with some sections of the flank standing nearly vertical.

Two prominent rift zones striking north and south-southeast extend from Lō`ihi's summit platform creating the elongate shape of the volcano (Fig. 3). This shape has been inferred to indicate that the rift zones formed early in the volcano's growth (Fornari et al., 1988). Otherwise, the volcano might have a more conical or starfish shape (see Vogt and Smoot, 1984). The north–south trend of the rift zones is perpendicular to the south coastline and submarine slope of the island of Hawai`i suggesting they did not develop within the gravitational influence of the adjacent volcanoes, which were thought to buttress the younger volcano (Swanson et al., 1976). The rift zone crests dip more gently that the summit flanks (~10° vs. 14–40°).

Lō`ihi's shorter north rift zone (~11 km long) includes two subparallel segments (Fig. 3) with numerous 10–30 m high pillow cones that were observed and sampled during three PISCES V dives (145–147; Garcia, unpubl. data). The double ridge character of the north rift has been attributed to lateral migration of magmatic feeders (Fornari et al., 1988), perhaps related to the collapse of the eastern flank. However, the limited sediment on the cones of the two rifts indicates recent volcanism on both rifts (M. Garcia, unpubl. data). The south rift is ~19 km long (Fig. 3), with several cones along the upper part (<1400 mbsl) of the sharp-crested rift but few along the lower part (Fornari et al., 1988; Garcia et al., 1993). The lavas along this crest range from sheet flows to rubbly breccias, with little or no sediment cover indicating recent eruptive activity (Garcia et al., 1993). The axis of the south rift curves to the southeast adjacent to sections where it collapsed on both sides of the rift (Fig. 3).

A bulge with three, 60–80 m high cones extends west from the northern summit area at 18°56.7′N (Fig. 3). It may represent a third rift zone or a product of isolated flank eruptions (Fornari et al., 1988). No submersible observations have been made of this area and there is no geophysical expression of a rift. The most prominent cone along this bulge has been dredged yielding weakly alkalic basalts typical of other young Lō`ihi cones (Moore et al., 1982).

The morphology of Lō`ihi has been extensively modified by landslides (Fornari et al., 1988; Moore et al., 1989). Landslides have over steepened the eastern flank of the volcano (Garcia et al., 1995) and created large gaps in its southwestern portion. A block on the western side of the south rift separating two, landslide-formed amphitheater valleys (Fig. 3) may have undergone gravitational slumping, although its more resistant nature has been interpreted as an indication that it is underlain by a rift zone (Fornari et al., 1988). However, there is no surface expression of this rift (e.g., cones or other signs of recent volcanism), so it may be an erosional remnant. Whatever the origin and history of this block, it is clear that more than half of Lō`ihi's surface area has been affected by landslides (Fig. 3). The debris from some of these landslides has created an extensive avalanche debris field that extends southeast of the volcano (Holcomb and Robinson, 2004). Thus, landsliding is clearly an important process in the evolution of even the youngest Hawaiian volcano.

5. Ages: implications for magmatic evolution and the 1996 eruption

Determining radiometric ages for Lō`ihi lavas is inherently challenging given their young age and moderate to low $K_2O$ content (0.3–1.6 wt%; Frey and Clague, 1983). However, ages have been determined for two suites of Lō`ihi rocks: a composite stratigraphic section collected from its dissected east flank (Guillou et al., 1997) and juvenile breccia from the 1996 eruption (Garcia et al., 1998a). Unspiked K–Ar analyses, a method for dating young lavas, yielded duplicated ages of 5 ± 4 to 102 ± 13 ka for the east flank section of Lō`ihi (Guillou et al., 1997). An older age, 201 ± 11 ka for the middle part of the section, was considered unreliable, a result of excess argon. These ages were used to estimate lava accumulation rates of 3.5 mm/y for the lower part of the section and 7.8 mm/y for the upper part (Guillou et al., 1997). An increase in lava accumulation rate is consistent with decrease in the percentage of alkalic lavas upsection (~90% to ~20%), indicating higher degrees of partial melting and higher eruption rates (Garcia et al., 1995). These geochronological results were combined with geological constraints for the growth of Hawaiian volcanoes to infer an overall age for Lō`ihi (Fig. 2). The dated east flank section samples represent only the uppermost part of the volcano (~0.5 km; Garcia et al., 1995). The magma budget for Kīlauea suggests that only ~1/3 of the magma intruded into Kīlauea volcano is extruded (Dzurisin et al., 1984), which is consistent with the idea that the deeper parts of Hawaiian volcanoes are dominated by intrusives (e.g., Hill and Zucca, 1987; Moore and Chadwick, 1995). If this ratio is valid for Lō`ihi, then ~30% of the volcano has formed in the last 100,000 years. Assuming linear growth, Lō`ihi is possibly 330,000 years old. However, the limited geochronology results and simple modeling studies (Garcia et al., 1995) suggest that lava accumulation and eruption rates were lower during the earlier part of Lō`ihi's growth. Assuming a progressive increase in accumulation rate, at least 400 k.y. are needed to form the volcano. This model is highly dependent on the assumed extent of endogenous growth (e.g., Francis et al., 1993) for this youthful volcano, which is unknown and dependent on the presence of a persistent shallow magma reservoir. This extrapolated age for the initiation of Lō`ihi volcanism is older than the estimated age for alkalic volcanism at Kīlauea volcano (180–330 ka; Lipman et al., 2002). However, if Lō`ihi's alkalic volcanism started at 400 ka or earlier, and has continued until recently (see below), then the duration of Kīlauea's alkalic volcanism is probably longer. A >400 ka start date for Lō`ihi impacts models for the Hawaiian plume melting region. For example, given the northwest drift of the Hawaiian Islands at ~10 cm/y (e.g., Garcia et al., 1987), Lō`ihi started forming at least 40 km southeast of its current location, out on the sedimentary apron surrounding the island of Hawai`i. This expands the region of plume melting by ~15 km, and makes Lō`ihi taller (~5 km) and more voluminous (1.7×10^3 km^3) than previously assumed (3.5 km and 0.8×10^3 km^3; Garcia et al., 1995).

Ages were also determined for two very glassy blocks collected just after the 1996 seismic event from a thin breccia deposit just north of Pisces Peak along the western margin of the summit platform (Fig. 5). The samples were radiometrically dated using the ^{210}Po–^{210}Pb method. ^{210}Po is part of the ^{238}U decay scheme and is volatile at magmatic temperatures (Vilenskiy, 1978; Le Guern et al., 1982). It degasses nearly completely (75–100%) from magmas erupted at water depths <2 km (Rubin et al., 1994). An age is determined by repeatedly analyzing the activity of ^{210}Po in a sample over a few of its 138.4-day half-lives and fitting the resulting data to an exponential ingrowth curve. Lava ages are reported as eruption windows, the most probable time of eruption between the calculated maximum and estimated minimum ages, because of uncertainty in the extent of Po degassing during the eruption (Garcia et al., 1998a). The 2-month eruption windows for these samples are during the first half of 1996 (Fig. 6), prior to the summer earthquake swarm that led to the summit collapse event that produced Pele's Pit (Garcia et al., 1998a) and not during periods of significant seismic activity at Lō`ihi (Fig. 6). Although the eruption 'windows' for the samples do not overlap even when analytical and regression errors are considered, they were collected from the same thin, localized breccia deposit and are identical petrographically. Thus, they were probably formed during the same episode and represent Lō`ihi's first documented eruption.

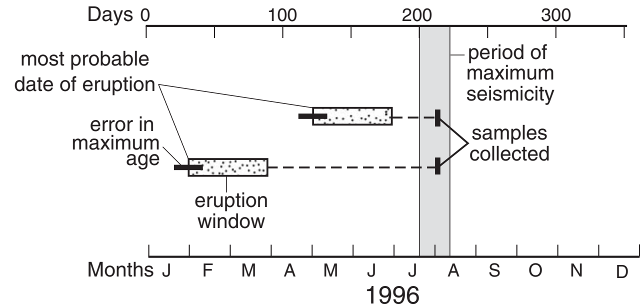

Fig. 6. ^{210}Po–^{210}Pb ages for of the Lō`ihi 1996 eruption based on two very glassy samples from a new breccia deposit. Ages for these lavas are show by the stippled boxes, with the most probable age on the left side of the box. These eruption windows are bounded by the dates of maximum (100%) and assumed minimum (75%) initial extent of Po degassing (left and right sides of the stippled boxes). Also shown is the error in maximum age (black horizontal bars) based upon data regression quality, the date of sample collection (vertical black bars) and the time period of the earthquake swarm (gray vertical band). These results indicate that the two glassy breccia lavas were erupted prior to the swarm. Although these two samples have different eruption windows, the geologic field relations suggest there were part of the same event. Time scales are given in both days (upper scale) and months during 1996 for reference. Modified after Garcia et al. (1998a).

6. Seismicity

6.1. Earthquake swarms

The 1996 earthquake swarm was the largest but not the first at Lō`ihi. Its historical record of seismicity begins with a large swarm of earthquakes in March 1952 (Macdonald, 1952). Since that time, researchers have used Lō`ihi seismicity as evidence of its activity and relationship with the Hawaiian hot spot (Klein, 1982), to investigate the volcano's velocity structure (Bryan and Cooper, 1995; Caplan-Auerbach and Duennebier, 2001a) and to examine the relationship between seismicity and Lō`ihi eruptions (Malahoff, 1993; Caplan-Auerbach and Duennebier, 2001a).

As recorded by the Hawaiian Volcano Observatory's (HVO) seismic network, located 35–120 km from the seamount, Lō`ihi seismicity is relatively low, on the order of a few earthquakes per month. This background activity is punctuated by periodic earthquake swarms, in which tens to hundreds of earthquakes of similar magnitude occur in the course of days to weeks (Fig. 7). Volcanic earthquake swarms are commonly associated with dike intrusion or magma reservoir inflation and as such may point to the location and extent of eruptive or intrusive activity (Klein et al., 1987). Intriguingly, Lō`ihi earthquake swarms do not typically locate beneath the volcano's summit, although the distribution of seismic stations introduces large epicentral uncertainty in the NE-SW direction (Caplan-Auerbach and Duennebier, 2001a). The swarms of 1971, 1986, 1993, 1995 and the initial stages of the 1996 event locate on Lō`ihi's southwest flank (Fig. 8; Klein, 1982; Caplan-Auerbach and Duennebier, 2001a). The earthquakes locate near a feature interpreted by Klein (1982) as an active, mobile flank and by Fornari et al. (1988) a possible failed rift zone (as discussed above; Fig. 3). In contrast, the 1990 and 1991 swarms cluster to the northeast of the volcano, with some activity located beneath the summit and south rift (Fig. 8). Only the 1975 and 2001 swarms and the later phase of the 1996 activity occurred beneath the summit region and south rift zone.

Two of Lō`ihi's earthquake swarms are believed to be associated with eruptions. At the time of the 1991 earthquake swarm, a temporary ocean-bottom observatory (OBO) was deployed. It was deployed for several months during 1991 and 1992 on Lō`ihi's summit near

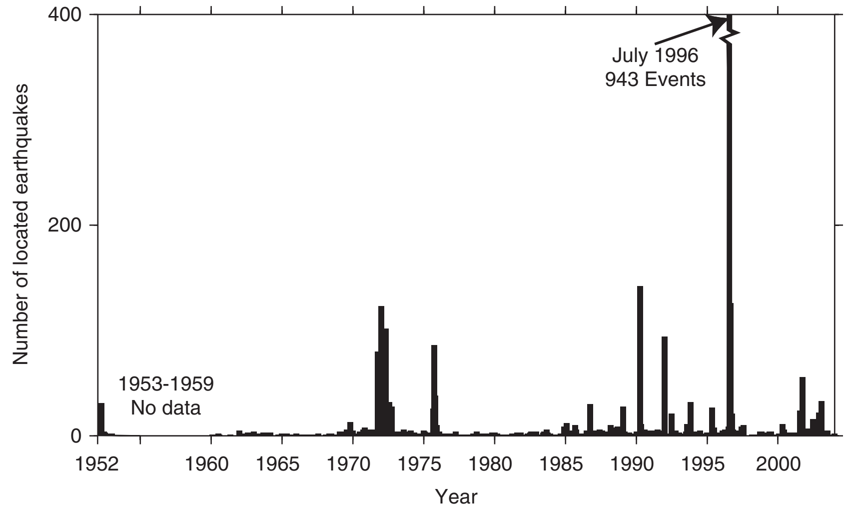

Fig. 7. Histogram of Lō`ihi monthly earthquakes based on detection by the HVO seismic network. No data exist between 1953 and 1959. Lō`ihi's seismicity is characterized by low background rates punctuated by occasional earthquake swarms. The apparent increase in monthly seismicity after 1986 reflects improvements to the HVO seismic network and data acquisition system.

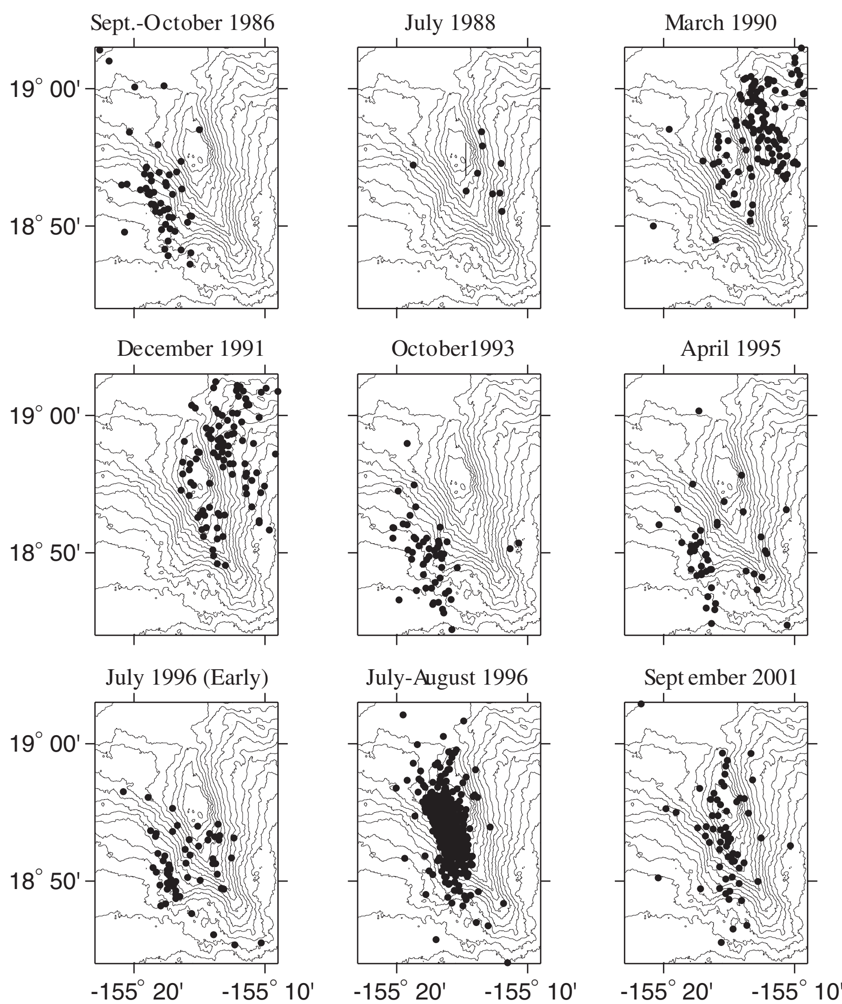

Fig. 8. Epicenters for earthquake swarms on Lō`ihi volcano between 1986 and 2001 overlaid on contoured bathymetry with intervals of 250 m. The 1996 swarm is shown in two phases: the early phase took place between July 16–18 and the second phase began July 20. Most swarms locate beneath the flanks of the volcano with the exceptions of the late phase of the 1996 swarm and the 2001 swarm. The 1991 and 1996 swarms are believed to be associated with eruptions.

the hydrothermal system known as Pele's Vents (Malahoff, 1993). The OBO recorded seismic, pressure and thermal data in 1991, overlapping the time period of the December earthquake swarm. Coincident with the seismic activity, the OBO recorded an increase in hydrothermal vent temperatures (~1 °C) and summit elevation decrease of ~40 cm, suggestive of magma withdrawal from the summit (Malahoff, 1993). The 1996 earthquake swarm was accompanied by the collapse of Pele's vents and the formation of a 300-m deep pit crater dubbed Pele's Pit (Loihi Science Team, 1997; Caplan-Auerbach and Duennebier, 2001a). Although no eruption was observed, numerous hydrophones deployed during two cruises in 1996 recorded explosion signals suggestive of eruptive activity emanating from the northeast section of Lō`ihi's summit. However, as discussed above, Po^{210} dating of rocks collected during this seismic swarm indicated that they erupted just prior to the earthquake swarm (Garcia et al., 1998a).

The majority of Lō`ihi earthquake swarms comprise tens to hundreds of earthquakes over a few days. The two exceptions are the 1971–1972 activity, spanning 5 months, and the 1996 event, which was shorter in duration but included thousands of earthquakes. Both of these swarms also saw a migration of earthquake epicenters between the volcano's flanks and summit (Klein, 1982; Caplan-Auerbach and Duennebier, 2001a). In 1971–1972, earthquakes migrated from a location just east of the summit to a broad region beneath the southwest flank, a sequence interpreted as summit rifting followed by motion on a mobile flank (Klein, 1982). The 1996 swarm began beneath Lō`ihi's south flanks and, following a day of seismic quiescence, migrated to the summit region (Fig. 8).

Because the 1996 activity was coincident with the formation of a pit crater on the summit, Caplan-Auerbach and Duennebier (2001a) interpreted the swarm as resulting from faulting on the south flank, possibly related to an eruption in early 1996. This faulting changed the stress field such that magma withdrew from a summit reservoir, inducing pit crater collapse (Caplan-Auerbach and Duennebier, 2001a). An ocean-bottom seismometer (OBS) deployed during the 1996 event enabled Caplan-Auerbach and Duennebier (2001a) to calculate hypocentral depths for a subset of swarm earthquakes. These events locate at 7–8 km depth beneath Lō`ihi's summit (Fig. 4). This is just above the depth at which rocks erupted in 1996 crystallized and thus is believed to represent the location of a magma reservoir (Garcia et al., 1998a; Caplan-Auerbach and Duennebier, 2001a).

6.2. Earthquake monitoring

Because Lō`ihi is >30 km from any of the seismometers operated by HVO, the magnitude detection threshold for Lō`ihi seismicity is relatively high, at ~M_L1.0. Thus, much of the volcano's seismicity either goes undetected or is detected at too few stations to be robustly located. The presence of a large number of earthquakes below the HVO network's magnitude detection threshold was confirmed with data from an OBS deployed in 1996. These data show that 10 times more earthquakes were detected by the OBS than were visible on stations in the HVO network (Caplan-Auerbach and Duennebier, 2001a).

The largest Lō`ihi earthquake for which a magnitude has been calculated occurred in September 2001 with magnitude $M_L$5.1. Two events with magnitude $M_L$4.9 occurred in 2001 and 1996. No volcanic tremor has been observed at Lō`ihi. However, the distance between the volcano and the HVO seismic network precludes detection of low-level signals, so weak volcanic tremor may not have been detected.

Lō`ihi's proximity to shore and frequent swarm activity has made it an excellent candidate for OBS and other focused seismic studies. Immediately following the 1986 earthquake swarm, a network of five OBSs was deployed on the summit and flanks of Lō`ihi for approximately 1 month (Bryan and Cooper, 1995). Most earthquakes recorded by the 1986 OBS network had magnitudes <1.5 and were not detected by the land-based HVO stations. Events located using only the OBS network locate on the summit and western flank of Lō`ihi (Bryan and Cooper, 1995). The few earthquakes detected by both networks locate to the north beneath the Big Island or between Lō`ihi and the island (Bryan and Cooper, 1995).

In the waning days of the 1996 swarm, a single OBS was deployed on Lō`ihi's summit. Although most of the earthquakes recorded by the OBS were not detected by the land-based network and could not be located, 42 earthquakes were well-recorded by both systems, allowing the production of a new velocity model for Lō`ihi (Caplan-Auerbach and Duennebier, 2001a). This model indicates that shallow (<7 km) velocities beneath Lō`ihi are slower than those used to locate earthquakes beneath Kīlauea. A conclusion of the 1996 OBS study, however, was that while the new velocity model improves our understanding of Lō`ihi and its seismicity, earthquake hypocenters will remain poorly constrained until longer-term instruments are deployed on Lō`ihi itself (Caplan-Auerbach and Duennebier, 2001a).

The goal of longer-term seismic monitoring of Lō`ihi was achieved in late 1997 with the deployment of the Hawai`i Undersea Geo-Observatory (HUGO) at the volcano's summit. HUGO was designed as a permanent observatory, with real-time power and data connections via a 50-km long electro-optical cable to the Big Island (Duennebier et al., 2002). The initial experiment package included a seismometer, which failed shortly after deployment, and a hydrophone. HUGO operated

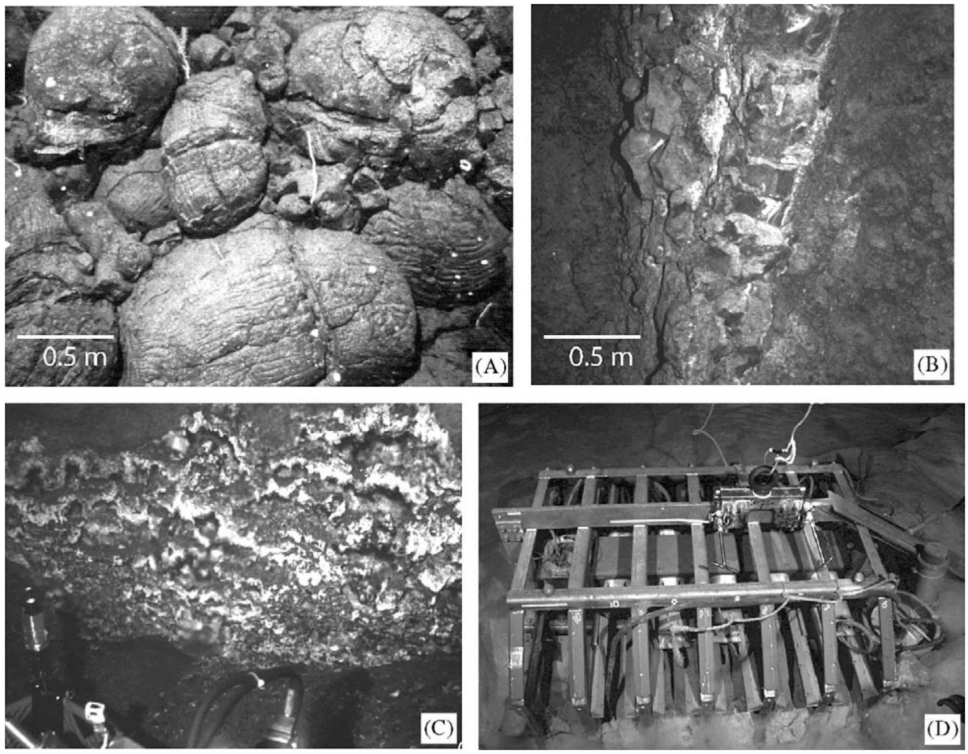

Fig. 9. Photos of Lō`ihi outcrops and HUGO: (A) Young bulbous pillows with deep-sea coral and brittle star from upper north rift zone at ~1200 mbsl. (B) Steeply dipping (~75°) dike intruding pillow lava, east flank at ~1450 mbsl. (C) Hydrothermal venting at Pele's Pit in 1997 creating barite mounds (white areas). Corner of Pisces V sample basket in foreground of photo. Field of view about 70 cm. (D) HUGO stuck in the mud and being recovered in November 2002 by JASON2. Photos A–C taken by PISCES V cameras; photo D taken by JASON2 camera.

between January and April 1998 until a short-circuit in the main cable terminated operations. The unarmored cable was apparently damaged by abrasion over the volcanic terrain. During the four months that HUGO was operational, hydroacoustic data were recorded continuously and delivered to the shore station in real-time (Caplan-Auerbach and Duennebier, 2001b). HUGO was recovered in October 2002 (Fig. 9D) and, following development of a new and sturdier cable, may eventually be redeployed.

Although a hydrophone records pressure fluctuations in the water, it is also able to detect earthquakes, as seismic signals can couple with the water column at the instrument site. HUGO was deployed during a period of seismic quiescence at Lo`ihi. Only 15 Lo`ihi earthquakes were recorded during the 4 months that HUGO was active, and none of these earthquakes was large enough to trigger the HVO seismic network. However, data from HUGO were useful in constraining the locations of earthquakes occurring between Lo`ihi and Kīlauea. The results from the HUGO study demonstrate that the presence of an offshore sensor dramatically improves locations and formal errors associated with seismic activity offshore of Hawai`i Island (Caplan-Auerbach and Duennebier, 2001b).

Data from HUGO and from the OBS studies performed on Lō`ihi confirm that the volcano's seismicity, internal structure and eruptive behavior cannot be fully understood without sensors positioned on the volcano itself. Further seafloor instrumentation is necessary in order to answer fundamental questions related to the growth and behavior of Lō`ihi as well as other submarine volcanoes.

7. Rocks: keys to unlocking Lo`ihi temporal magmatic evolution

7.1. Sampling and submersible observations

The summit and rift zones of Lō`ihi have been extensively sampled providing good spatial and temporal coverage of its volcanic products. Early rock sampling (prior to 1987) was by dredging (Moore et al., 1982; Hawkins and Melchior, 1983). Most of the subsequent sampling utilized a submersible, primarily the PISCES V but also the ALVIN, SEACLIFF, MIR and SHINKAI 6500 manned submersibles and KAIKO, a remotely operated vehicle. Thus, the volcano has attracted broad international interest with only the French submersible, NAUTILE, not having visited Lō`ihi. Submersibles provide invaluable opportunities to directly observe Lō`ihi, and to collect rock and water samples. These vehicles also make it possible to collect stratigraphic rock sections to evaluate Lō`ihi's temporal magmatic evolution. Three sections were collected in the two older pit craters (310–350 m thick; Garcia et al., 1993) and another three were plucked from the walls of the deeply dissected east flank of the volcano (total stratigraphic thickness ~550 m; Garcia et al., 1995). These sections help document the volcano's post ~100 ka magmatic history (Guillou et al., 1997).

Observations from the Pisces V submersible of the east flank of Lō`ihi revealed a dike complex (Garcia et al., 1995). The dikes are steeply dipping, 60–90° (Fig. 9B), creating nearly vertical walls with intervening sections of pillow lavas mantled with talus. The dikes range in thickness from 20 to 150 cm (most are 50–100 cm) and strike generally N10°W to N10°E, subparallel to the cliff face. The east wall dikes are similar in density to subaerial Hawaiian dike complexes (e.g., Ko`olau volcano; Walker, 1986). The abundance of dikes decrease upsection, especially above 1400 mbsl and none were observed in the upper part of the section (<1200 mbsl). Most of the east flank is composed of pillow lavas (Fig. 9A) with no clear stratigraphic breaks. However, the section is locally draped with younger sheet flows forming pronounced angular unconformities (Guillou et al., 1997). One thin (~1 m thick) volcaniclastic deposit was encountered near the top of the section (1160 mbsl). In situ samples were taken from the east flank pillow section at regular intervals (20–40 m) wherever possible (Garcia et al., 1995).

The walls of the two older summit pit craters were examined in three traverses (two in the deeper East Pit) during 1987 using the ALVIN submersible (Garcia et al., 1993). The traverses began on the floor of the craters, which are covered with silt. This silt extends to the south and east of the craters and is especially thick on the southeast corner of the summit platform where HUGO was deployed. HUGO sank into this sediment prior to being rescued (Fig. 9D). About half the West Pit section was draped with talus or contained massive sections, ~10 m tall, of columnar basalt. The East Pit is better exposed, revealing thick sections of pillow lava with no obvious stratigraphic breaks. The sections were densely sampled, every 10–20 m, where possible. The summit platform north of the pit craters is covered with young looking (i.e., lightly sedimented and glassy) bulbous pillows. Similar flows were found on the distal tip of the south rift zone (Umino et al., 2000).

A traverse along the upper south rift zone (<1400 mbsl) in 1987 found young lavas, including sheet flows, and several small cones (Garcia et al., 1993). Two of the cones (Pele and Kapo) contained blocky lava and were venting warm, shimmering fluids (~15–30 °C) from their rubble (Karl et al., 1988). Pele's cone was the site of the 1996 collapse that produced Pele's Pit (Loihi Science Team, 1997). Other areas of weak, warm hydrothermal venting have been reported on the western, eastern and southern flanks of the summit (Malahoff, 1987; Hilton et al., 1998; Wheat et al., 2000).

Following the 1996 earthquake swarm, several Pisces V dives were made to investigate its consequences (Loihi Science Team, 1997). Fresh glassy breccia samples were found on the west flank of the summit platform just north of Pisces Peak (Fig. 5), now the highest point on Lō`ihi (Pele's cone was the highest peak on the volcano before its 1996 collapse; Fornari et al., 1988). Pisces Peak was visited in 1987 with the ALVIN submersible and was found to be covered with pillow lavas with a thin surface coating of iron oxides (Garcia et al., 1993). After the 1996 earthquake swarm, this area was littered with broken weathered rock debris, scattered glassy rocks including delicate glass shards, the freshest that have been recovered from Lō`ihi (Garcia et al., 1998a; Clague et al., 2000).

7.2. Petrography and mineral chemistry: magma history implications

The petrography of Lō`ihi rocks has been described in numerous studies (e.g., Moore et al., 1982; Frey and Clague, 1983; Hawkins and Melchior, 1983; Garcia et al., 1989, 1993, 1995, 1998a). These glassy lavas commonly contain olivine crystals, like most Hawaiian basalts (e.g., Macdonald, 1949). This is especially true in the dredge rock suites, where olivine abundances range widely (1–52 vol%; Frey and Clague, 1983; Hawkins and Melchior, 1983; Garcia et al., 1989). Generally, there is no correlation of olivine abundance with rock type, although the hawaiites all have rare olivine (⩽0.1 vol%; Garcia et al., 1995). Olivine is generally euhedral and undeformed, with inclusions of chromite and glass (Fig. 10B), although some crystals are resorbed and a few are weakly kink banded. Chromite may also occur as small crystals (<0.5 mm) in the matrix.

Clinopyroxene is the second most common mineral in Lō`ihi lavas. It is common in tholeiitic lavas including those from the 1996 eruption (Garcia et al., 1998a). Clinopyroxene generally occurs as the second crystallizing phase. This is dramatically illustrated in the 1996 lavas where small olivine inclusions occur within large clinopyroxene crystals (Fig. 10A). Clinopyroxene is usually euhedral and commonly sector zoned. Plagioclase is less common in Lō`ihi lavas (e.g., none were observed in the 1996 lavas). When present, it is generally small (<0.5 mm). Many Lō`ihi lavas also have FeS globules in matrix glass (Byers et al., 1985; Yi et al.,

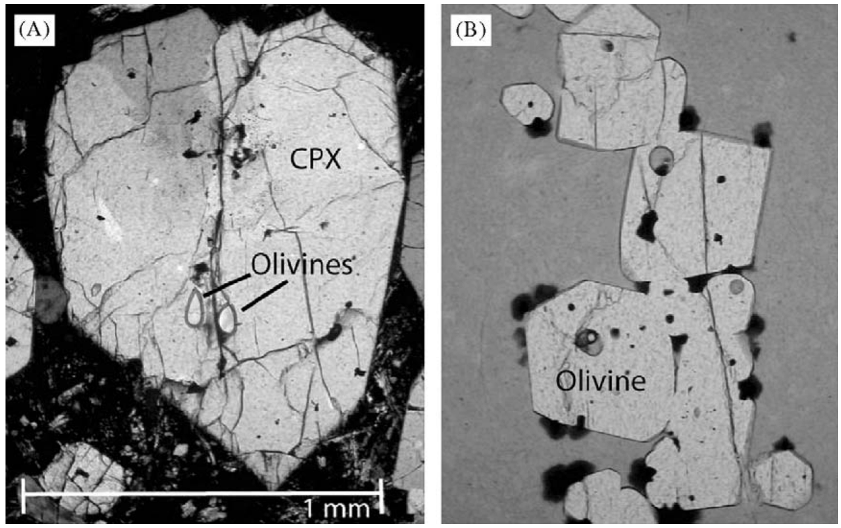

Fig. 10. Photomicrographs of Lō`ihi basalts: (A) Crossed nicols view of a clinopyroxene (CPX) with two small, elongate olivine inclusions in sample 286-2. Olivine was the liquidus phase but it reacted with the melt and was overgrown by clinopyroxene. See text for discussion of the magmatic conditions that created this texture. (B) Plain light view of euhedral olivine crystals with inclusions of chromite and glass set in clear brown glassy matrix of sample 286-1 from the 1996 eruption. The scale is the same for both photos. See Garcia et al. (1998a) for additional information on these samples.

2000), which are rarely observed in lavas from other Hawaiian volcanoes (e.g., Davis et al., 2003b).

Vesicularity in Lō`ihi rocks ranges widely; dikes have low vesicularity (<5 vol%), reflecting solidification under pressure. Lavas show a dramatic range in vesicularity (0.1–43 vol%) with no obvious correlation with rock type (e.g., Moore et al., 1982; Frey and Clague, 1983; Hawkins and Melchior, 1983; Garcia et al., 1993, 1995). For example, some upper south rift alkalic and tholeiitic lavas are strongly vesicular (~40 vol%; Garcia et al., 1993). The vesicularity of the 1996 lavas is moderate (5–20 vol%).

Silt from the southern summit platform adjacent to HUGO (Fig. 9D) contained abundant (25–35 vol%) pristine glass shards with 20–30 vol% unaltered mineral fragments (mostly olivine with some clinopyroxene and plagioclase), and 30–40 vol% rock fragments with varying degrees of alteration. The pristine nature of some glass and mineral fragments suggest this deposit is young. The silt extends into a hydrothermal field with numerous small clay chimneys, 0.5–2.0 m high (Malahoff, 1987). However, it is unknown whether the silt is related to the hydrothermal field.

Lō`ihi minerals have been analyzed by electron microprobe in several studies (Hawkins and Melchior, 1983; Garcia et al., 1995, 1998a) to better understand their crystallization histories. Olivines usually have high forsterite contents (80–90.3%), typical of Hawaiian basalts (e.g., Clague et al., 1995; Garcia et al., 2003), although a few analyses are reported with lower forsterite contents (65–78%; Clague, 1988). The vast majority of olivine grains are normally zoned. CaO contents in the olivines are moderate (0.2–0.4 wt%), indicating crustal depths (<12 km; Fig. 4) of crystallization (e.g., Garcia, 2002).

Clinopyroxene crystals vary markedly in composition in Lō`ihi lavas (Hawkins and Melchior, 1983; Garcia et al., 1995, 1998a). Most have relatively high Mg#s (80–84), moderate to low TiO_2 (0.8–2.0 wt%), Cr_2O_3 (0.3–1.0 wt%) and Na_2O (0.2–65 wt%), but highly variable Al_2O_3 (2.5–12 wt%). The large range in Al_2O_3 might reflect the wide range in Lō`ihi lava compositions (tholeiite to basanite; Moore et al., 1982), with higher values in alkalic lavas as observed for lavas elsewhere (e.g., Dobosi, 1989). However, many Lō`ihi lavas show strong variations within individual crystals regardless of rock compositions. In some tholeiitic lavas, clinopyroxene cores are more Al-rich than the rims (e.g., 5.1 vs. 2.3 wt%; Garcia et al., 1998a), which may reflect polybaric crystallization (e.g., Gasparik and Lindsley, 1980). In other tholeiites, the rims may have much higher Al_2O_3 contents than the cores (e.g., 4–10 wt%; Garcia et al., 1995), which may reflect disequilibrium growth (e.g., Allègre et al., 1981). These variable conditions seem to overprint any compositional variations due to magma composition. A complex history for some clinopyroxene-bearing rocks is also reflected in the presence of reverse zoning in some crystals (e.g., 1996 eruption lavas; Garcia et al., 1998a).

The limited plagioclase compositional data show that anorthite contents range widely (47–71%; Hawkins and Melchior, 1983; Garcia et al., 1995). The plagioclase in the two tholeiitic lavas that have been analyzed have generally higher anorthite contents than the two alkalic lavas that have been studied (66–71% vs. 47–69%). Spinels in Lō`ihi lavas are typically Cr-rich, although

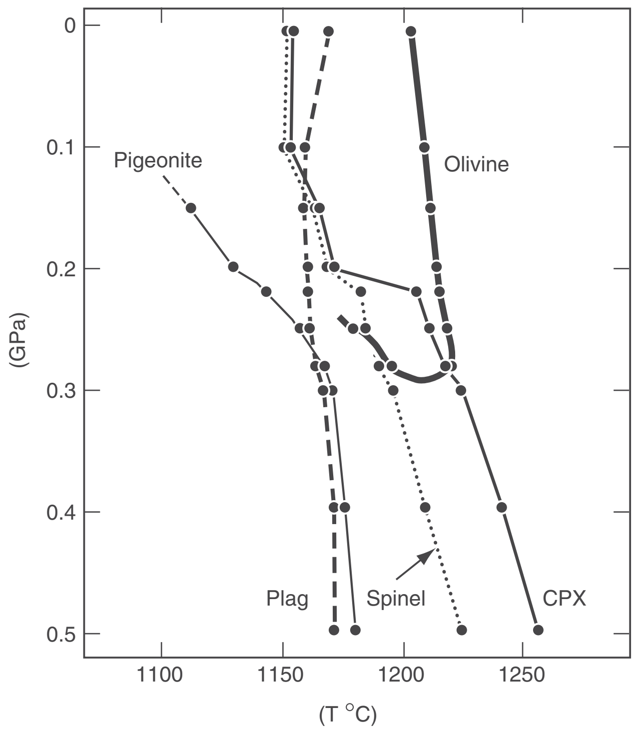

Fig. 11. Calculated phase diagram created using the MELTS program (Ghiorso and Sack, 1995) for lava from the 1996 eruption. For composition of the sample used, see Garcia et al. (1998a). Modeling conditions were 0.5 wt% H_2O (the glass contained 0.61 wt% H_2O), oxygen fugacity of 1 log unit below the FMQ buffer (based on the Fe^{2+}/Fe^{3+} results of Byers et al., 1985) and equilibrium crystallization (i.e., crystals remained with the melt). For each pressure intervals, shown by dots at 1 atm, and 0.1, 1.5, 2.0, 2.2, 2.5, 2.8, 3.0, 0.5–4, 0.5 GPa, the liquidus temperature was determined and then the magma was cooled in 1 °C increments to determine the crystallization sequence at that pressure. Olivine is unstable in a melt of this composition at pressures >0.28 GPa. At pressures of 0.22–0.28 GPa, it dissolves in the melt once clinopyroxene begins to crystallize and completely disappeared after 30 to 40 °C.

the more evolved alkalic lavas contain Ti-magnetite (Hawkins and Melchior, 1983; Garcia et al., 1995). Cr spinels commonly contain 39–48 wt% Cr_2O_3, with moderate Al_2O_3 (12–17 wt%) and TiO_2 (1.5–3.7 wt%). The magnetites are TiO_2-rich (11.5–18.5 wt%), with highly variable Cr_2O_3 contents (0.0–14 wt%).

The mineralogy of Lō`ihi lavas provides insights into the volcano's magmatic processes. For example, the mineralogy of the 1996 lavas records two distinct magmatic processes: moderate pressure crystal fractionation and magma mixing just prior to the eruption (Garcia et al., 1998a). Moderate pressures are indicated by modeling using the MELTS program (Ghiorso and Sack, 1995), which shows that olivine is the liquidus phase only at pressures <0.28 GPa for the 1996 eruption lava composition (Fig. 11). In this modeling, olivine is closely followed by clinopyroxene (<10 °C) at pressures of ~0.22–0.28 GPa and is resorbed by the magma at 2.5–2.8 GPa. The presence of olivine inclusions in clinopyroxene crystals (Fig. 10A) is indicative of this reaction relationship. This texture was used to infer that the 1996 magmas were stored at moderate pressures (0.28–0.25 GPa) prior to eruption (Garcia et al., 1998a). Greater pressures would have prevented early olivine crystallization, whereas lower pressure would have inhibited clinopyroxene formation and resorption of olivine (Fig. 11). These pressures were used to estimate the probable depth of magma storage for the 1996 lavas at 8–9 km (Fig. 4), ~1 km below the main concentration of earthquake hypocenters from the 1996 swarm (Fig. 4). Magmatic earthquakes at Kīlauea volcano commonly occur just above magma bodies (e.g., Klein et al., 1987) and presumably the same is true for Lō`ihi. Thus, the interpretations from the seismic and petrologic modeling are in good agreement. A moderate depth magma chamber may have existed for some time prior to the 1996 eruption based on the common occurrence (33%) of clinopyroxene with olivine inclusions in Lō`ihi tholeiitic lavas (Garcia et al., 1998a). However, such clinopyroxenes are absent in the older alkalic lavas from the east flank section. Thus, the formation of a moderate depth magma chamber may have followed the volcano's transition from alkalic to tholeiitic magmatism. This transition may have started at ~20 ka (Guillou et al., 1997) and may now be essentially complete. The formation of this moderate depth magma chamber may be related to an increase in magma supply rate, which is thought to accompany the alkalic to tholeiitic transition on Hawaiian shield volcanoes (e.g., Frey et al., 1990). The greater depth of Lō`ihi's magma chamber, compared to those at the more active shield volcanoes to the north (i.e., Mauna Loa: 3–4 km depth; Decker et al., 1983; Kīlauea: 3–6 km; Klein et al., 1987), may be a consequence of Lō`ihi's cooler thermal regime. Thus, the depth of Lō`ihi's magma chamber may be controlled by thermal conditions, which are largely governed by magma supply rate, rather than by the volcano's density structure, as was proposed by Ryan (1987).

The 1996 eruption also involved magma-mixing based on the presence of reverse zoning in the clinopyroxene crystals and two compositionally distinct populations of olivine crystals (Fo ~87% vs. 81–82%). The narrow width of the reversely zoned clinopyroxene rims (outer 0.01–0.02 mm) indicates that the mixing event probably occurred shortly before and may have triggered the eruption (Garcia et al., 1998a).

7.3. Rock types and temporal magmatic variation

Numerous studies characterizing submarine basalts have included Lō`ihi lavas (e.g., Yi et al., 2000; Kaneoka et al., 2002; Boyet et al., 2005). This fascination with Lō`ihi rocks started with the unanticipated discovery of alkalic lavas in 1981 (Moore et al., 1982). The early submarine phase of Hawaiian volcanism was thought to consist of tholeiitic basalt (Macdonald et al., 1983). Subaerial Hawaiian volcanoes had been well studied prior to this expedition, resulting in a well established evolutionary sequence for their subaerial growth (e.g., Stearns, 1946; Macdonald, 1963). These giant shield volcanoes ($2–8 \times 10^4\,km^3$) were thought to contain a core of tholeiitic basalts and intrusions with a thin cap of alkalic lava (Fig. 12). However, this interpretation ignored the >5 km of submarine growth.

Among the dredged Lō`ihi samples, the alkalic lavas have on average thicker palagonite alteration rinds than the tholeiitic rocks, although thicknesses overlap for the two rocks groups (1–12 vs. 0.5–4 μm; Moore et al., 1982). The alkalic lavas are also more vesicular causing a concern in assessing the relative ages of Lō`ihi lavas that alkalic lavas might alter faster than tholeiitic lavas. Nonetheless, the alkalic lavas were presumed to be older (Moore et al., 1982), which was in agreement with experimental studies and theoretical models, whereby early alkalic magmatism reflected initial lower degrees of melting. As the volcano drifted toward the hotspot and into a region of higher temperatures, the extent of partial melting increased, producing larger volumes of tholeiitic melts (e.g., Frey et al., 1990).

To test this melting hypothesis, submersible expeditions were undertaken to examine and sample from the walls of the summit pit crater and the deeply dissected east flank Lō`ihi (Garcia et al., 1993, 1995). These studies confirmed that alkalic lavas are generally older than the tholeiitic lavas. For example, alkalic lavas were found in the walls of the older West Pit crater just above its base, although the entire younger East Pit crater section consists of tholeiites. The east flank composite section shows a dramatic variation in rock type (Fig. 13). The lower section (>1450 mbsl) is overwhelmingly alkalic (14 of 16 flows), whereas the upper

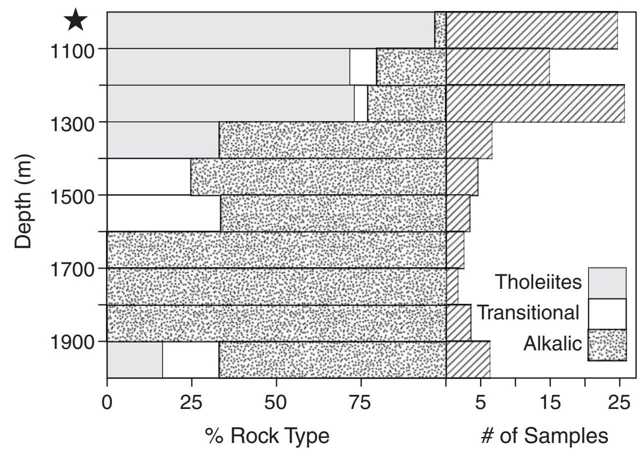

Fig. 13. Lō`ihi rock type variation with depth based on 95 submersible-collected samples from the East Pit crater and the adjacent, deeply dissected east flank. Transitional lavas are those that plot near the Macdonald-Katsura (1964) line on a total alkalis vs. SiO_2 diagram (Fig. 14). The right side of the diagram is a histogram of the number of samples per 100 m depth interval. The lava from the 1996 eruption is shown by the star and counted as a single sample. Note the dramatic increase in the percentage of tholeiitic lavas above 1300 mbsl. Rock type information from Garcia et al. (1993, 1995).

section is mostly tholeiitic (8 of 14 flows). If lavas from the pit crater sections and 1996 eruption are included in this analysis, tholeiites are the dominant recent rock type at Lō`ihi (59 of 75 samples; Fig. 13). However, the presence of several young alkalic cones along the uppermost south rift including Pele's cone, indicates that alkalic volcanism has continued until recently (Garcia et al., 1993). This gradational transitional between rock types is also observed for the preshield stage for Kīlauea (Lipman et al., 2002) and post-shield stage for several Hawaiian volcanoes (e.g., Mauna Kea, Frey et al., 1991; Kohala, Feigenson et al., 1983). The Lō`ihi results and the previous work on other Hawaiian volcanoes allow a composite cross section to be drawn illustrating the proportions rock types for the three primary stages of growth of these volcanoes (Fig. 12). This section supports the model of Macdonald (1963) that alkalic volcanism represents a minor component (<5 vol%) of Hawaiian volcanoes.

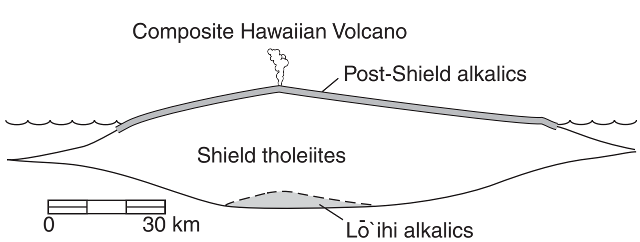

Fig. 12. Schematic cross section of a composite Hawaiian volcano at the end of the postshield stage showing rock type proportions. The preshield stage is represented by Lō`ihi alkalic lavas. The shield stage is based on Mauna Loa and is composed of tholeiites. The postshield stage is based on Mauna Kea volcano (Frey et al., 1990). The section has two times vertical exaggeration.

7.4. Whole-rock compositions

Relatively few whole-rock XRF analyses are available for Lō`ihi lavas compared to glass analyses (Frey and Clague, 1983; Hawkins and Melchior, 1983; Garcia et al., 1995, 1998a). These analyses include rocks that have no glass and those with abundant phenocrysts, and span the rock type range (tholeiites to hawaiites; Fig. 12). These whole-rock data extend the range of

Lō`ihi compositions to both higher and lower MgO contents than reflected in the glass compositions (3–25 wt% vs. 4–10 wt%). The low MgO samples are aphyric hawaiites that lack glass. The high MgO rocks have abundant olivine phenocrysts (>20 vol%), probably resulting from accumulation. For example, the 1996 eruption lavas vary from 8.2 to 10.3 wt% MgO, which can be explained by the observed modal differences in mafic minerals (~3 vol%; Garcia et al., 1998a). At a given MgO content, the concentrations of other major oxides are similar for whole-rock and glass analyses (Garcia et al., 1995). The Al_2O_3/CaO and K_2O/P_2O_5 ratios in Lō`ihi lavas have wide ranges that correlate with rock type (higher in alkalic rocks, lower in tholeiites) reflecting the influence of variable degrees of melting of a clinopyroxene-bearing source (e.g., Frey and Clague, 1983).

7.5. Glass geochemistry: magma chamber processes

Glass is common on Lō`ihi lavas and has been used extensively to characterize its rock types because it represents liquid compositions (e.g., Moore et al., 1982). The glass data show a remarkably wide range, from tholeiitic to strongly alkalic compositions (Fig. 14). This range cannot reflect low pressure fractionation of olivine, which would increase both silica and total alkalis (Fig. 14). Another striking feature of the glass compositional data is the restricted MgO range for tholeiites compared to the alkalic lavas (Fig. 15). The MgO content of Hawaiian basalt glass has been related to temperature of the magma at the time of eruption (e.g., Helz and Thornber, 1987). The small MgO range for tholeiitic glasses (6.2–8.0 wt%) indicates a relatively narrow temperature range (~40 °C based on the Kīlauea glass geothermometer of Helz and Thornber, 1987). This may reflect the presence of a steady-state summit magma reservoir as observed for Kīlauea (Garcia et al., 2003). In contrast, the alkalic glasses span a large MgO range (4–9 wt%; Fig. 15), probably indicating they were stored for variable periods in ephemeral magma chambers during times of lower magma supply. High MgO glasses may have been erupted without mixing with a cooler, lower MgO resident magma, whereas lower MgO glasses reflect storage and crystallization over considerable periods in a magma reservoir that was not being frequently recharged. Lō`ihi glasses are distinct compared to those from other Hawaiian volcanoes in their high CaO contents, although some of the tholeiites overlap with Kīlauea tholeiites (Fig. 15). The high CaO content may reflect melting of a more clinopyroxene-rich source (Garcia et al., 1995).

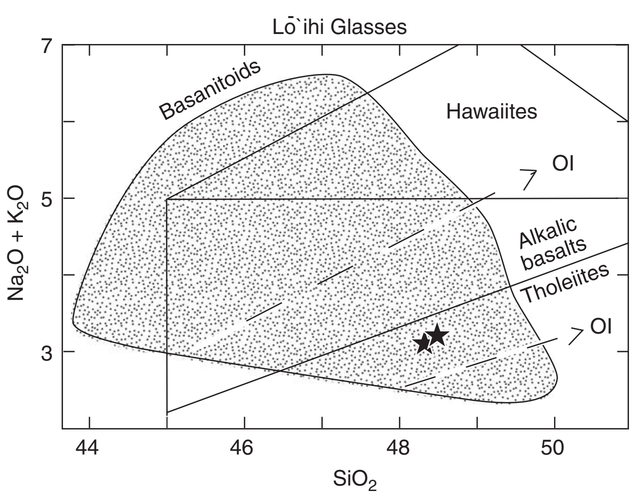

Fig. 14. SiO_2 vs. total alkalis (Na_2O+K_2O) plot for 74 Lō`ihi glasses (all values in wt%). The stippled field shows Lō`ihi glass data from Moore et al. (1982) and Garcia et al. (1989, 1993, 1995, 1998a). The field boundaries for rock compositions are from LeBas et al. (1986) except the dividing line for tholeiites and alkalic basalts, which is from Macdonald and Katsura (1964). Transitional basalts plot along and near this dividing line. Two tholeiitic glasses from the 1996 breccia are shown by the stars. Dashed trend lines are shown for olivine fractionation from parental alkalic and tholeiitic compositions.

Volatile concentrations (H_2O, S, Cl, and CO_2) in Lō`ihi glasses have received considerable attention (e.g., Byers et al., 1985; Kent et al., 1999; Dixon and Clague, 2001) following the exciting He isotope evidence that Lō`ihi magmas have a relatively primitive source (Kurz et al., 1983). Early microprobe analyses of glasses from dredged samples showed high S contents (0.11–24 wt%). These glasses also commonly contain S globules (Yi et al., 2000). In general, the glasses show a good correlation of H_2O with K_2O contents, suggesting that degassing has not affected H_2O concentrations (Byers et al., 1985; Dixon and Clague, 2001). However, CO_2 abundances are more variable and are not indicative of the depths of sample collection (Dixon and Clague, 2001) consistent with variable CO_2 degassing. Cl concentrations are even more erratic than CO_2 with concentrations up to 0.17 wt% and no correlation with K_2O (Byers et al., 1985; Kent et al., 1999). The high Cl concentrations were interpreted as evidence for widespread assimilation of a seawater-derived component, probably brines (Kent et al., 1999). High-Cl glasses were also found among the early submarine rocks from Kīlauea's (Coombs et al., 2004) supporting the idea that seawater contamination of magma is more likely during the early stages of Hawaiian volcanism (Kent et al., 1999). However, even higher Cl concentrations (up to 0.36 wt%) were found in some Mauna Loa glass inclusions (Davis et al., 2003b). The high Cl in these inclusions correlated with high F contents, which led to the suggestion that hydrothermal deposits rather than brines have contaminated the Mauna Loa magmas (Davis et al., 2003b). Little F data are available for

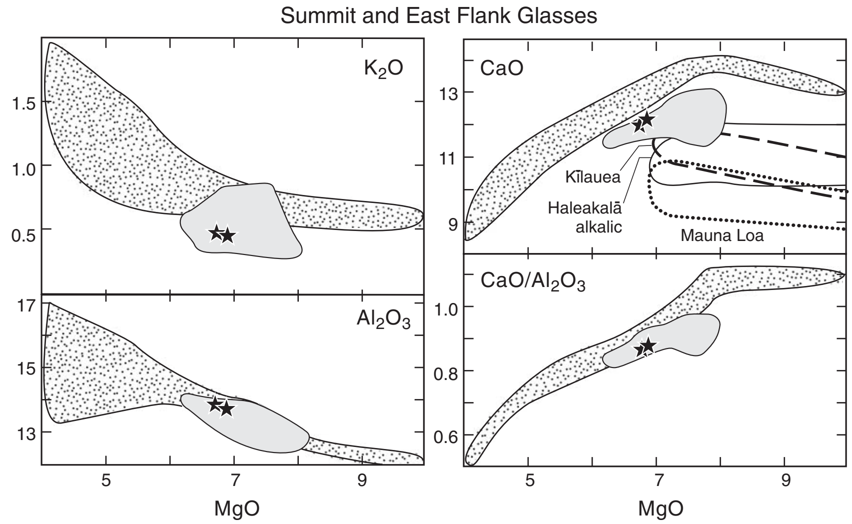

Fig. 15. MgO variation diagrams for Lō`ihi summit and east flank glasses (all values in wt%). Lō`ihi glasses have relatively high CaO contents compared to lavas from other Hawaiian shield volcanoes (fields from Wright, 1971 for Mauna Loa and Kīlauea; Chen et al., 1991 for Haleakala). Two glasses from the 1996 breccia are shown by the stars. Data from Garcia et al. (1993, 1995, 1998a).

Lō`ihi glasses to evaluate if hydrothermal deposits are a potential assimilant for Lō`ihi magmas.

The limited stable isotope data for Lō`ihi glasses provide equivocal evidence for seawater contamination. Two tholeiitic samples with moderate to high Cl contents have δD values of -69 and -84, which are thought to be representative of the normal mantle values (Garcia et al., 1989). Oxygen isotope values for 16 glasses are relatively low (δ^{18}O of 4.7–5.2) compared to those from mid-ocean ridge basalts (MORB) (5.3–6.0; Ito et al., 1987). However, the Lō`ihi values are comparable to other Hawaiian basalts, including those for lavas from the ongoing Kīlauea eruption, which show no other signs of seawater contamination (Garcia et al., 1998b).

The oxygen fugacity of basaltic magma influences the sequence and composition of minerals crystallizing from a cooling magma, and is thought to reflect the oxygen fugacity of their mantle source (Rhodes and Vollinger, 2005). Studies of the redox state of rapidly quenched glasses from Lō`ihi have played an important role in understanding the oxygen fugacity of the Hawaiian mantle plume. Lō`ihi glasses have oxygen fugacities close to, or below magnetite-wüstite (MW) buffer (Byers et al., 1985; Wallace and Carmichael, 1992). Subsequent studies of rapidly quenched lavas from the ongoing eruption of Kīlauea and the 1984 eruption of Mauna Loa have oxygen fugacities also yield values close to MW (Rhodes and Vollinger, 2005). Thus, the plume source for Hawaiian magmas is near MW, which overlaps but is not quite as reduced as the source of MORB (e.g., Christie et al., 1986).

Trace element concentrations have been determined on many of Lō`ihi's lavas using a variety of methods (e.g., XRF and INAA, Frey and Clague, 1983; ICPMS, Garcia et al., 1998a). Analyses of glasses show systematic variations for highly incompatible trace element ratios regardless of rock type (Fig. 16), suggesting that these rock types were derived from sources with similar geochemical characteristics (Garcia et al., 1995). Wide variations are observed for ratios of highly over moderately incompatible elements (e.g., La/Yb), with lower ratios for tholeiitic lavas and higher ratios for alkalic lavas (Fig. 16). Some of this variation in the evolved alkalic lavas is attributable to clinopyroxene fractionation. However, the large range in trace element ratios in the less evolved lavas is probably related to variable amounts of partial melting of a common mantle source (e.g., Allègre and Minster, 1978), which is supported by the Sr, Nd and Pb isotope results presented below.

Glasses from the ~350 m thick tholeiitic East Pit stratigraphic section show a temporal variation in the ratios of highly to moderately incompatible elements (Fig. 17). This trend continues when the young tholeiitic lavas collected north of the West Pit crater are included. Glasses from these young lavas have lower La/Yb ratios than any of the East Pit glasses (5.1–5.3 vs. 5.6–7.4; Fig. 17). This trend reversed with the 1996 eruption near the West Pit, which have slightly higher La/Yb ratios (5.5–5.7; Fig. 17). Cyclic variations were also noted for the Mauna Kea lavas from the two Hawai`i Scientific Drilling Project drill cores from its flanks, where individual cycles (period between crests in trace element ratios) span thousands of years (Yang et al., 1996; Blichert-Toft et al., 2003), and for the historical summit lavas of Kīlauea, where the recent individual cycle may be hundreds of years long (Pietruszka and Garcia,

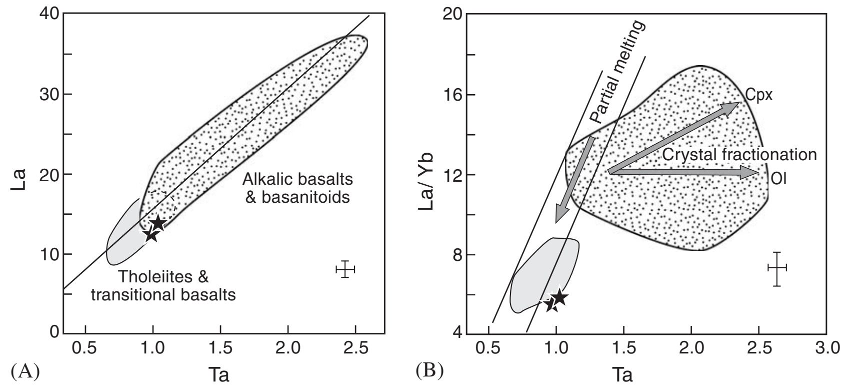

Fig. 16. Ta variation diagrams for Lō`ihi summit and east flank glasses: (A) The co-linear tend for La–Ta indicate that the glasses were derived from sources with similar La/Ta ratios. Similar well-defined trends are observed for plots of other highly incompatible elements. The tholeiitic and transitional glasses data plot together (gray field), whereas the alkalic basalts and basanitoids range widely. (B) La/Yb ratios reflect the extent of partial melting and fractionation of olivine or clinopyroxene. The arrows point in the direction of increasing partial melting and fractionation. The tholeiitic and transitional glasses were derived at similar and higher amounts of partial melting compared to the alkalic glasses (stippled fields). Some of the alkalic glasses experienced greater extents of crystallization of both olivine and clinopyroxene. Two-sigma error bars are given in the lower right corners of each plot. Data are from Garcia et al. (1993, 1995, 1998a).

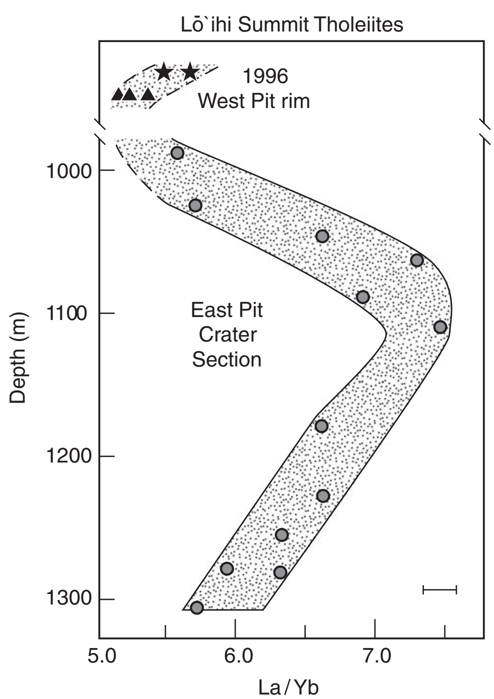

Fig. 17. Variations in the La/Yb ratios with stratigraphic position for tholeiitic glasses from Lō`ihi's East Pit crater (gray circles), rim of the West Pit and 1996 eruption. Note the break in section between the East Pit section, and the other samples. Only samples with >6.7 wt% MgO are shown to minimize the effects of clinopyroxene fractionation. The temporal variation in La/Yb (stippled band) appears to have reversed during or prior to the 1996 eruption. A two-sigma error bar is given in the lower right corner of the plot.

1999a). The fluctuations in trace element trends for Mauna Kea and Kīlauea lavas correlate with Pb and Sr isotope ratio variations indicating that the cyclicity is related to changes in proportions of the mantle source components, which may be controlled by melting processes (Pietruszka and Garcia, 1999b: Blichert-Toft et al., 2003). Another interesting corollary with other Hawaiian volcanoes is the reversal in the La/Yb ratios following the collapse of the Kīlauea's summit in 1924 (Pietruszka and Garcia, 1999a). Although the La/Yb reversal at Lō`ihi predates the 1996 earthquake swarm, it may have been a harbinger of the collapse of Lō`ihi a few months later (Fig. 6).

The first comprehensive study of the Sr, Nd, and Pb isotope ratios in Lō`ihi basalts was by Staudigel et al. (1984). They found unusually large variations in these isotopes for a single volcano, which led to the realization that at least three source components are needed to explain the isotopic variations in Hawaiian basalts. This and subsequent studies (Garcia et al., 1993, 1995, 1998a) found no correlation of these isotopes with rock type (Fig. 18), or ages, in contrast to studies of other Hawaiian volcanoes (e.g., Mauna Loa; Kurz et al., 1995). The significant overlap in isotopes for Lō`ihi alkalic and tholeiitic lavas indicates that they were produced from the same heterogeneous source (Garcia et al., 1995).

The overlap in Sr, Nd and Pb isotopes for Lō`ihi alkalic and tholeiitic lavas led to a modeling study to determine whether the alkalic magmas were formed by high pressure clinopyroxene fractionation of tholeiitic magmas or variable degrees of partial melting (Garcia et al., 1995). It was argued that the systematic variations in highly over moderately incompatible elements for the two magma types (e.g., La/Yb; Fig. 16) could not be explained by high-pressure clinopyroxene fractionation. Modeling of incompatible trace element concentrations in lavas with similar Sr, Nd, Pb isotopic ratios was

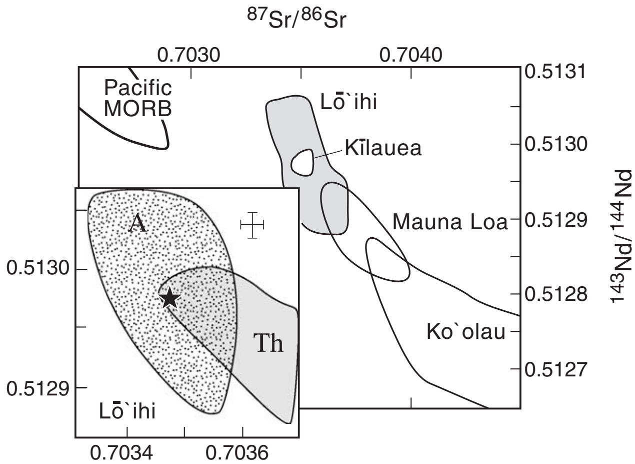

Fig. 18. Plot of Nd vs. Sr isotope ratios for Lō`ihi lavas. The large field encompasses all published Lō`ihi data (~30 analyses). The insert figure shows the Lō`ihi data subdivided by rock type: alkalic—A; tholeiitic—Th. The two new 1996 breccia samples are nearly identical and plot near the center of the alkalic field. Two-sigma error bars are shown in the upper right corner. The MORB field is from King et al. (1993). Hawaiian data from Kurz and Kammer (1991) for Mauna Loa, Pietruszka and Garcia (1999a) for Kīlauea and Roden et al. (1994) for Ko`olau.

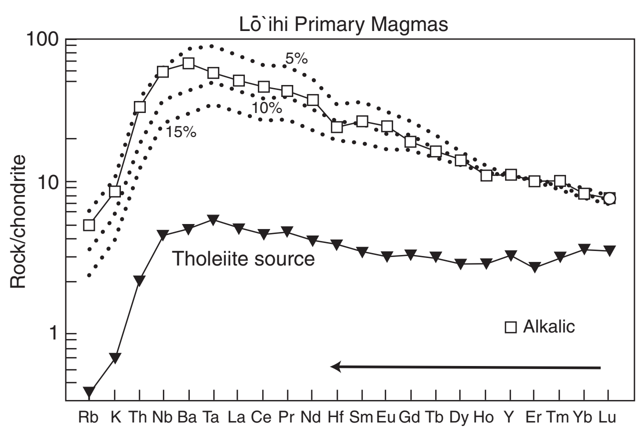

Fig. 19. Partial melting model for Lō`ihi lavas. Source composition was calculated from a Lō`ihi tholeiite assuming 10% nonmodal, equilibrium partial melting of a garnet lherzolite (see Garcia et al., 1995 for partition coefficients). The composition of melts formed by 5%, 10% and 15% melting are shown by dotted lines. A Lō`ihi alkalic primary melt (16 wt% MgO) plots between the 5% and 10% partial melt composition. Arrow shows direction of increasing element incompatibility.

undertaken using a tholeiite to indicate the mantle source trace element composition and assuming non-modal, equilibrium melting of a garnet peridotite (Garcia et al., 1995). Modeling at 5–15% partial melting generated a range of melt compositions (dotted patterns on Fig. 19). The trace element abundances and pattern for ~8% melting were comparable to a Lō`ihi primitive alkalic melt (Fig. 19). Although the true extent of partial melting for alkalic magmas is probably lower, the modeling demonstrates that variable degrees of partial melting of a common source could explain the range of rock types erupted at Lō`ihi.

8. Noble gases: windows into the mantle

Glassy basalts from Lō`ihi have been crucial to mantle noble gas studies because they have helium and neon isotopic compositions that are among the least radiogenic found on Earth. Consequently, Lō`ihi is among the best studied volcanoes for noble gases.

8.1. Helium

The first reported helium measurements from Lō`ihi basalts revealed $^{3}He/^{4}He$ of ~32 times the atmospheric value [Ra] (Kurz et al., 1982). This value is far above the average value for MORB of ~8 Ra, and created considerable interest in Lō`ihi among noble gas geochemists. More detailed studies on the dredged lavas yielded $^{3}He/^{4}He$ ranging from 20 to 32 Ra (Kurz et al., 1983; Kaneoka et al., 1983; Rison and Craig, 1983). The values for alkali basalts and tholeiites overlap, but the highest $^{3}He/^{4}He$ are found in the tholeiites, suggesting some heterogeneity in the Lō`ihi mantle source, as also indicated by radiogenic isotope data (see Fig. 20).

The initial interpretation of these high $^{3}He/^{4}He$ values emphasized ancient undegassed mantle sources that had maintained high He/(Th + U) since the formation of the Earth, most likely derived from the lower mantle (e.g., Kurz et al., 1982; Allègre et al., 1983). This hypothesis has been challenged, based on widespread geochemical evidence for recycling (e.g. Hofmann, 1997), and seismic evidence for penetration of subducted slabs into the lower mantle (van der Hilst and Karson, 1999). The alternative models for the origin of unradiogenic noble gases require that helium be more compatible than Th and U during silicate melting, which could leave behind an ancient residue of depleted mantle with high He/(Th + U) (e.g., Anderson, 1998; Meibom et al., 2003). This debate has not been resolved, but has considerable geodynamic importance.

Shield tholeiites from neighboring Hawaiian volcanoes commonly have $^{3}He/^{4}He$ higher than MORB, but no other Hawaiian lavas are as unradiogenic for helium isotopes as Lō`ihi seamount. Existing data from older volcanoes in the Hawaiian Emperor chain (i.e., 45–75 Ma) include $^{3}He/^{4}He$ values of 10 to 24 Ra (Keller et al., 2004), all of which are lower than the maximum found at Lō`ihi. One problem with making this comparison is that there are large, and sometimes rapid,

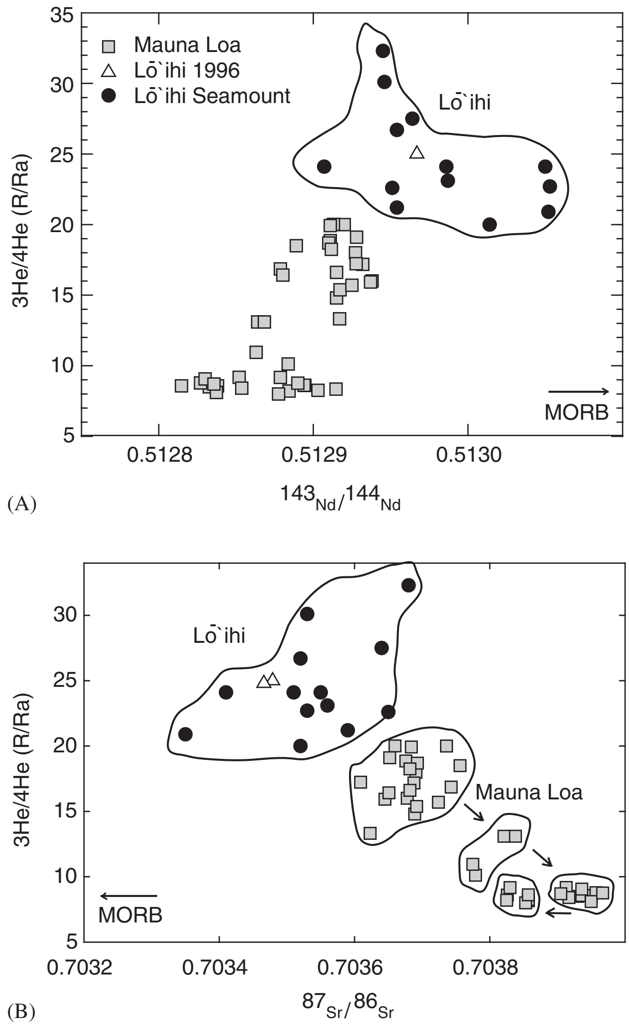

Fig. 20. Sr, Nd, and He isotope data for Lō`ihi seamount and Mauna Loa samples: (A) Lō`ihi samples have distinctly higher He isotope values and somewhat higher Nd isotope ratios than those from Mauna Loa lavas. (B) Temporal evolution for Mauna Loa lava shown by arrows from older (>28 ka in age) to successively younger groups (7–12 ka, 0.6–7 ka to <0.6 ka). The oldest lavas from Mauna Loa are closest to Lō`ihi. The area shows the direction of the MORB field. Data sources for Lō`ihi: Kurz et al., 1983; Staudigel et al., 1984; Garcia et al., 1998a; Mauna Loa and MORB: Kurz and Kammer, 1991; Kurz et al., 1995.

temporal isotopic variations within Hawaiian shields, so a small number of samples do not necessarily characterize a volcano. Where age or stratigraphic data are available for Hawaiian volcanoes, the oldest shield building tholeiites have the highest ^{3}He/^{4}He ratios whereas later tholeiites and alkali basalts approach MORB values. This includes data from Kīlauea (Kurz, 1993), Mauna Loa (Kurz and Kammer, 1991; Kurz et al., 1995), Mauna Kea (Kurz et al., 1996, 2004), Haleakalā (Kurz et al., 1987) and Kaua`i (Mukhopadhyay et al., 2003). These studies vary in their time scales and sampling density, but yield a consistent temporal pattern. The temporal trend for Hawaiian shield volcanoes is evident for He, Sr, and Nd isotopic compositions when comparing Lō`ihi and Mauna Loa (Fig. 20). The oldest analyzed Mauna Loa lavas (~250 ka in age; Kurz et al., 2004) are closest in isotopic composition to Lō`ihi, whereas younger Mauna Loa lavas have generally lower $^3$He/$^4$He (Fig. 20). Temporal helium isotopic evolution has not been established at Lō`ihi, possibly due to the lack of age control for most analyzed samples. However, lavas from the 1996 eruption have ^{3}He/^{4}He values of ~26 Ra (Garcia et al., 1998a), within the range of reported values (20–32 Ra; Kurz et al., 1983; Kaneoka et al., 1983; Rison and Craig, 1983), and shows that there has been no significant recent change in Lō`ihi helium isotopes. The high $^3$He/$^4$He in Lō`ihi lavas and the overall decrease in this ratio for lavas from later stages of volcanism from other volcanoes has led to the hypothesis that Lō`ihi is close to the present-day center of the Hawaiian hotspot (e.g., Kurz et al., 1983, 2004; Kaneoka, 1987). High $^3$He/$^4$He values (17–21 Ra) were also reported for submarine alkali basalts collected on the Hawaiian Arch 190 km south of Lō`ihi, which were interpreted as indicating a strong plume influence (Hanyu et al., 2005).

8.2. Argon, xenon and neon

The first heavy noble gas studies of Lō`ihi revealed argon and xenon isotopic compositions close to air values (Allègre et al., 1983), although distinct from those observed for MORB. Assuming that the helium isotopic compositions reflected a deep mantle origin, this was interpreted to indicate a deep mantle with undegassed but air-like heavy noble gas signatures (Allègre et al., 1987; Staudacher et al., 1986), including neon (Sarda et al., 1988). An additional conclusion was that the difference in ^{129}Xe/^{130}Xe between Lō`ihi and MORB implied isolation of MORB from the lower mantle for at least 4.4 billion years. ^{129}Xe is the stable daughter of extinct ^{129}I ($t_{1/2}$ of 17 Ma), and any intrinsic difference in ^{129}Xe/^{130}Xe must have been produced while ^{129}I was alive, i.e., within Earth's first 100 m.y. This has obvious importance to geodynamic models; for example, these data are consistent with a poorly mixed, layered, mantle (Allègre et al., 1983). However, it is possible that the Lō`ihi Xe isotopic data reflect late stage atmospheric contamination rather than I/Xe systematics and radioactive decay. Argon and xenon are less soluble than helium in silicate melts, more easily lost during outgassing, and are more prone to adsorption and atmospheric contamination. Fisher (1985) and Patterson et al. (1990) suggested that the air-like Lō`ihi heavy noble gas isotopes were produced by atmospheric

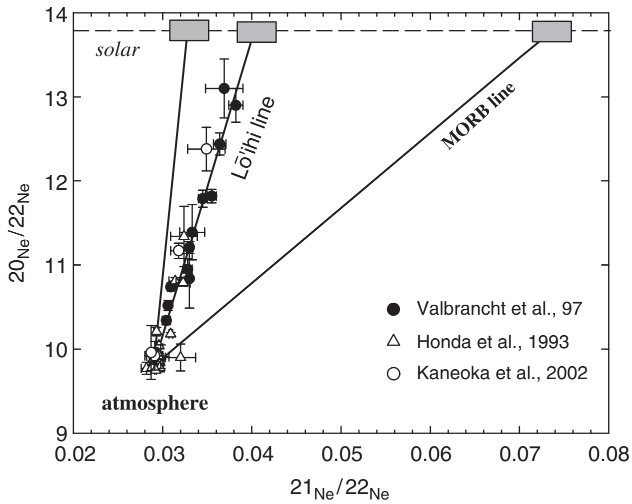

Fig. 21. Neon isotopes from Lō`ihi seamount glasses determined by step heating and crushing in vacuo from three different laboratories (Honda et al., 1993; Valbracht et al., 1997; Kaneoka et al., 2002). For clarity, only the highest values from each reported measurement are plotted. For step-heating experiments, this is usually the highest temperature step. Neon released at lower temperatures usually has higher uncertainties and is close to air. Present-day solar wind neon is indicated by the upper left hand rectangle. No difference was found between the results for the samples collected on the shallow (Honda et al., 1993) and deeper parts of the volcano (Valbracht et al., 1997; Kaneoka et al., 2002). The Lō`ihi trend line is distinct from the trends for mixing atmospheric neon with MORB and solar wind neon. The linear arrays are assumed to be caused by mixing between discrete mantle components (i.e., the Lō`ihi mantle source) and atmosphere (see text for details).

contamination. Staudacher et al. (1991) noted that the Lō`ihi glass abundance ratios were not consistent with atmosphere or sea water interaction. Nevertheless, it is plausible that argon and xenon isotopes in the Lō`ihi glasses reflect some atmospheric or seawater contamination, given the other trace element evidence for seawater influence (e.g., high Cl contents; Kent et al., 1999).

Neon measurements of Lō`ihi glasses led to the first reports of extremely unradiogenic neon isotopes (Honda et al., 1991, 1993; Hiyagon et al., 1992). This has been called "solar" neon because the $^{20}Ne/^{22}Ne$, and particularly $^{21}Ne/^{22}Ne$ ratios, approach those of solar wind, and are presumed to originate from early implantation (i.e., during accretion) of solar noble gases into the earth. Similar values were found for high $^{3}He/^{4}He$ islands such as Iceland, and there is a good general correlation between unradiogenic helium and neon (e.g., Moreira et al., 2001). Neon data for Lō`ihi samples from three different laboratories and sample suites form a single linear array (Fig. 21) suggesting mixing of a high $^{20}Ne/^{22}Ne$ mantle component (close to the present-day solar wind) with atmospheric neon. The linear array of neon isotopes (Fig. 21) demonstrates the importance of atmospheric contamination processes. It is unclear if the atmospheric contamination effects are produced during eruption on the seafloor, during residence in a magma chamber or even in the laboratory. However, the three isotopes of neon allow extrapolation to possible mantle end-members for each linear array (Fig. 21), assuming that the mantle has uniformly high (i.e. solar) $^{20}Ne/^{22}Ne$ ratios and that the variations along the linear arrays are caused by atmospheric contamination. This set of assumptions leads to a single Lō`ihi neon isotopic composition (Fig. 21), despite the range in helium and other isotopes. The Lō`ihi mantle source neon isotopic composition is clearly less radiogenic than MORB, which is consistent with the helium data.

Concern about the possible effects of degassing on the Lō`ihi noble gas results led to the collection and measurement of samples from the deeper flanks of Lō`ihi (Valbracht et al., 1997; Kaneoka et al., 2002). These studies also yielded argon and xenon isotopic compositions close to atmospheric (e.g., $^{40}Ar/^{36}Ar$ ranging from 296, air, to 2600), and total gas concentrations similar to the earlier studies. Another related issue was the relatively low helium concentrations in Lō`ihi glasses. This is important because undegassed sources (i.e., high He/(Th + U)) should have higher helium concentrations. Relatively low helium concentrations (at least 10 times lower than MORB) are found even in the deep glasses, demonstrating that eruption depth is not the primary control on the He content of Lō`ihi glasses.

Non-atmospheric xenon was observed in two Lō`ihi dunite xenoliths, with $^{129}Xe/^{130}Xe$ ratios indistinguishable from MORB xenon (Trieloff et al., 2000). These xenolith data contrast with the near-atmospheric Xe isotopic values obtained for Lō`ihi glasses, and may indicate that the glass xenon data were influenced by atmospheric contamination, or that the xenon in the xenoliths was derived from the MORB lithosphere. It is unclear how the xenolith xenon relates to the source of the basalts, even though helium isotopes are similar. Clague (1988) suggested that the Lō`ihi xenoliths are cumulates formed from Lō`ihi melts, but acknowledged that the noble gases may have a more complex history.

Collectively, the Lō`ihi studies represent significant advances in understanding mantle noble gases. However, important controversies remain, particularly relating to atmospheric contamination of the heavy noble gases, and the site-of-origin for unradiogenic He and Ne within the mantle. Some investigators advocate a layered mantle, with unradiogenic He and Ne derived from the lower mantle (e.g., Allègre and Moreira, 2004), whereas others assume that the mantle is convecting from top to bottom, and that unradiogenic noble gases must be derived from the core or the core-mantle boundary (e.g., Ballentine et al., 2002). An additional possibility is that helium is more compatible in silicates (relative to Th and U) during mantle melting, so ancient melting events could have left *residual* mantle with elevated He/(Th + U), thus preserving ancient high $^3He/^4He$ ratios. The noble gas silicate/melt partition coefficients are not well known, although they are fundamental to interpreting the unradiogenic He and Ne found at Lōʻihi (e.g., Brooker et al., 2003). Resolving the noble gas story will require further studies.

9. Geochemistry of hydrothermal systems

Hydrothermal activity at Lōʻihi volcano has only been studied since the late 1970s discovery of iron oxides and nontronite coatings on dredged lavas (Moore et al., 1979). An elevated formation temperature for the nontronite (31 °C), water temperature, methane and ^3He anomalies, and clumps of benthic micro-organisms in the water column led to the suggestion of extensive hydrothermal activity at Lōʻihi (Malahoff et al., 1982; Horibe et al., 1986). This was confirmed in 1987 during ALVIN submersible dives by the observation of 'shimmering' water and elevated water temperatures (15–30 °C) at the Pele and Kapo vents along the upper part of the south rift zone (Karl et al., 1988).

9.1. Vent deposits

The hydrothermal deposits encrusting lava dredged in 1982 from the northeast corner of Lōʻihi's summit yielded smectite formation temperatures of 31–57 °C (De Carlo et al., 1983). The green to yellow to red color pattern and variable Fe^{2+}/Fe^{3+} ratios of these deposits are thought to reflect initial deposition of nontronite followed by precipitation of amorphous iron oxide and silica as oxygen-poor vent fluids percolate upward through lava talus and mix with oxygenated, cool seawater (De Carlo et al., 1983). Trace element abundances and X-ray diffraction analysis of these deposits suggested the existence of polymetallic sulfides at Lōʻihi, which would indicate that the volcano had a high-temperature hydrothermal system (Malahoff et al., 1982; De Carlo et al., 1983).

Following the formation of Pele's Pit in the summer of 1996, sulfide minerals were found and temperatures up to 200 °C were measured at vents within the pit, confirming the presence of an extensive high-temperature hydrothermal system within Lōʻihi. The millimeter- to centimeter-sized sulfide samples included "brassy" pyrite and/or marcasite, less common lustrous, black, hexagonal wurtzite and clear, euhedral barite crystals (Davis and Clague, 1998). EDS analysis of the pyrite/marcasite revealed they are relatively homogeneous FeS with only traces of Mn, Co and Cu, with pyrrhotite inclusions containing minute chalcopyrite inclusions (Davis and Clague, 1998). The presence of wurtzite, pyrrhotite, and chalcopyrite, is consistent with the existence of high-temperature (>250 °C) hydrothermal fluids at Lōʻihi, and represents the first documented evidence of high-temperature fluids at a mid-plate hotspot submarine volcano (Davis and Clague, 1998). This sulfide mineral assemblage is typical of that found in black smokers at mid-ocean ridge spreading centers (Craig and Scott, 1974). The Lōʻihi high-temperature sulfides are thought to have formed when a megaplume of hot hydrothermal fluids was ejected through talus and mixed with ambient seawater following the formation of Pele's Pit (Davis et al., 2003a). The relatively homogenous composition of the sulfides suggests that these minerals precipitated through continuous discharge of hydrothermal fluids whose temperature and composition changed little (Davis et al., 2003a).

Barite-rich mounds up to 1 m in diameter and several tens of cm high were constructed in the Pit as fluids up to ~200 °C vented in the talus (Fig. 9C). The barite crystals from the mounds show strong compositional zoning reflecting fluctuations in vent fluid temperature and composition (Davis et al., 2003a). The mounds also contain anhydrite, pyrite and rare zinc sulfide. This assemblage is mineralogically similar to that found in white smokers at mid-ocean ridges, although the $\delta^{34}S$ of the Lōʻihi sulfides are lower and attributed to loss of magmatic sulfur (Davis et al., 2003a). The dissolution features observed on some sulfides indicate that these minerals are unstable (Davis et al., 2003a), which may explain why they were not observed prior to the 1996 event. The temperature of vent fluids in Pele's Pit has decreased from a high of ~200 °C in 1996 to slightly above 60 °C during the 2004 dive season, suggesting the current phase of hydrothermal activity may be subsiding (C.G. Wheat, F. Sansone and E. De Carlo, unpublished results). Recovery of basalt fragments coated with unweathered sulfide mineral assemblages from talus in Pele's Pit during these same dives, however, suggests continued, if sporadic, activity at Lōʻihi.

9.2. Vent fluids

Geochemical anomalies in the water column above Lōʻihi were first observed in 1982 (Horibe et al., 1986). More extensive sampling in 1985 found water rich in methane (up to 569 ppm), He (91.8 nl/l, a record high for open-ocean water), CO_2, Fe, and Mn (Sakai et al., 1987; Gamo et al., 1987). Two plumes with different methane concentrations were detected, indicating at least two summit hydrothermal vent fields. Comparison of the Lōʻihi data with those other submarine hydrothermal areas showed that the Lōʻihi hydrothermal system has 10–100 times smaller $CH_4/^3He$ ratios than the EPR and

Galapagos Spreading Center, implying a higher component of magmatic fluid. Lō`ihi plumes also displayed low pH values (as low as 7.2 vs. 7.8–8.2 for normal seawater), which were attributed to high CO_2 (Sakai et al., 1987; Gamo et al., 1987).

Direct sampling of Lō`ihi's low temperature vent fluids (15–30 °C) in 1987 using titanium samplers revealed strong enrichments (e.g., dissolved Li, PO_4^{3-}, NH_4^+, Rb and Ba) and depletions (SO_4^{2-}, Mg, O_2, and NO_3^-), as observed in the ocean water in 1985 (Karl et al., 1988). The greatest difference between the compositions of these vent fluids vs. low-temperature vent fluids elsewhere was their markedly higher CO_2 ($c_T = 300$ mM) content and lower pH (5.3–5.5; Karl et al., 1988). The low pH in these fluids can also account for their high dissolved Fe concentrations (~1 mM), 2–3 orders of magnitude greater than in low temperature hydrothermal vent fluids on Axial Seamount and the Galapagos Rift. The high concentrations of dissolved Fe probably are responsible for the precipitation of the abundant low-temperature iron-rich deposits around Lō`ihi vents upon mixing with seawater (e.g., Malahoff et al., 1982; De Carlo et al., 1983).

There are several possible sources of hydrothermal methane at Lō`ihi: abiogenic sources, (i.e., mantle volatiles); organic matter that underwent high temperature interactions with basalt and seawater; and biogenic sources associated with thermophilic and methanogenic bacteria. The absence of $C_2^+$ hydrocarbons in the Lō`ihi vent fluids led Karl et al. (1989) to concluded that the methane was extracted from basalt by circulating seawater, as proposed by Welhan and Craig (1983) for other hydrothermal systems.

Between 1987 and 1992, Pele's Vents fluids were sampled repeatedly. Relative to fluid temperature, the volatile content of these fluids showed remarkable variations, with dissolved CO_2 decreasing by ~30%, He decreasing ~20-fold, and the $CO_2/^3He$ ratios increasing by an order of magnitude (Sedwick et al., 1994). The $\delta^{13}C$ values of the CO_2 in the fluids (–5.5 to −1.7 vs. PDB) and corrected $^3He/^4He$ ratios (21.7–27.0 Ra) are both indicative of a magmatic source contributing to the vent fluids (Sedwick et al., 1994). The fluids collected during this period were also enriched in dissolved Si, CO_2, H_2S, alkalinity, K, Li, Rb, Ca, Sr, Ba, Fe, Mn, and NH_4^+ but depleted in SO_4^{2-}, O_2, Mg, $^{87}Sr/^{86}Sr$, and NO_3^- relative to ambient seawater (Karl et al., 1988; Sedwick et al., 1992; Wheat et al., 2000). A strong correlation was noted between dissolved Si and vent fluid temperature, suggesting that dissolved Si concentrations could be utilized to trace mixing of vent fluids with ambient seawater (Sedwick et al., 1992).

Remarkable similarities and differences were noted between the chemistry of Lō`ihi vent waters and warm springs on the Galapagos Rift and at Axial Seamount (Wheat et al., 2000). The higher $CO_2$ and $SO_4^{2-}$ in Lō`ihi fluids were interpreted to reflect the mixing of a high temperature (>200 °C) seawater-derived fluid with juvenile CO_2 and SO_2, and cold, unaltered seawater (Sedwick et al., 1992). The suggestion that high-temperature fluids existed at depth in Lō`ihi was considered highly speculative prior to the summer of 1996.

The surface manifestations of Lō`ihi's hydrothermal system changed dramatically following the 1996 collapse of Pele's vents to form a new pit crater. Major changes in hydrothermal venting associated with eruptive events are well documented at mid-ocean ridges (e.g., Butterfield et al., 1997; Baker et al., 1998, 1999, 2004). To better document the fluids from the new Lō`ihi vents, OsmoSamplers were deployed at two sites within the Pele's Pit and at Naha vents (upper south rift zone) in October 1996 and recovered in September 1997 (Wheat et al., 2000). These devices collected fluids regardless of their flow rate and provided time-series data that were unobtainable using conventional samplers (e.g., 3-L Niskin bottles and 750 mL titanium Walden-Weiss). The sampler in Pele's Pit recorded a decrease in thermal and fluid fluxes involving a high-temperature source (>330 °C, the boiling point of seawater at vent depth, 1325 mbsl) with magmatic volatiles mixing with bottom seawater (Wheat et al., 2000). At the Naha Vents, chlorinity increased and K concentration decreased in the fluids, consistent with two or more distinct fluid sources, including a low-temperature component (Wheat et al., 2000).

The Fe/Mn ratio in the new vent fluids ranged between 58 and 0.8, with low values observed in vent waters collected within Pele's Pit and high values found in fluids from Naha vents. The maximum Fe/Mn fluid value is similar to the value observed for Lō`ihi rocks (e.g., Frey and Clague, 1983). Low Fe/Mn in fluids from Pele's Pit are associated with high-temperature and rock-dominated reactions similar to those observed experimentally (Seyfried and Mottl, 1982) and in high-temperature vent fluids recovered from mid-ocean ridges (e.g., Butterfield and Massoth, 1994, and references therein). Low values of Fe/Mn are typically attributed to removal of Fe from the fluids by precipitation of sulfide minerals, which is consistent with the low (<11 μmol/kg) H_2S concentrations observed in Lō`ihi vent fluids. Low Fe/Mn ratios tend to correlate with alkalinity. Although the fluids with the highest measured temperatures (~200 °C) from Pele's Pit display enhanced alkalinities over bottom seawater, the alkalinity remained much lower than observed in fluids collected prior to the 1996 event (Wheat et al., 2000). This contrasts with other high-temperature hydrothermal systems in which alkalinities are typically extremely low. The decreasing trend in alkalinities from 1993 to 1997 is interpreted to correspond to a decreasing magmatic CO_2. The correlation between the Fe/Mn ratio and alkalinity at a given fluid Si concentration was related by Wheat et al. (2000) to the CO_2 content of fluids, which promotes weathering of basalt and produces fluid acidity. Hence, acidity is thought to control the Fe/Mn ratio of the fluids at Lō`ihi rather than the water:rock ratio or complexation with chloride.

10. Conclusions

Lō`ihi Seamount was discovered in 1952 following an earthquake swarm. It was largely ignored for two decades until two earthquake swarms in the 1970s renewed interest in the volcano and motivated a reconnaissance survey in 1978. During this survey, young lavas were photographed and dredged, confirming that Lō`ihi is an active Hawaiian volcano rather than a Cretaceous seamount. Lō`ihi's relatively small size, location south of the other active Hawaiian volcanoes, and the alkalic composition of many of its lavas led to it being considered the youngest volcano in the chain. The next two decades were a period of numerous marine expeditions to map, instrument the volcano with various sensors, and explore its geology and hydrothermal vents. The HVO seismic network has continuously recorded Lo`ihi's background seismicity over 40 years documenting 12 earthquake swarms. Geophysical monitoring included a real-time submarine observatory that continuously monitored the volcano's seismic activity for three months. A new Lō`ihi observatory is planned for the future, which will allow its internal structure and eruptive behavior to be more fully understood.

Studies on Lō`ihi have altered conceptual models for the growth of Hawaiian and other oceanic island volcanoes, and led to an improved understanding of mantle plumes. Petrologic and geochemical studies of Lō`ihi lavas showed that the volcano taps a relatively primitive part of the Hawaiian plume, producing a wide range of magma compositions. These compositions have become progressively more silica-saturated with time reflecting higher degrees of partial melting as the volcano drifts towards the center of the hotspot. He and Ne isotopes show that the source for Lō`ihi magmas define a unique and important unradiogenic mantle source, not related to a MORB source. The three isotopes of neon provide a clear distinction between mantle and atmospheric components. Ar and Xe isotopes in Lō`ihi glasses appear to be influenced to some extent by outgassing and atmospheric contamination. Therefore, the original interpretation that Lō`ihi heavy noble gas isotopic compositions represent the isotopic composition of the lower mantle, must be viewed with caution. Nonetheless, noble gas measurements of Lō`ihi basalts provide a benchmark for noble gas studies and remain central to debates regarding the inner workings of the planet.

Re-examination of radiometric ages has led to a revised growth model for Lō`ihi indicating that it starting forming on the deep ocean floor at the base of the island of Hawai`i. This expands the area affected by plume melting, and volcano's height and volume over previous models. This interpretation helps explain the north–south orientation of Lō`ihi's rift zones, which would have formed outside the gravitational influence of the neighboring volcanoes. Landsliding has substantially modified the morphology Lō`ihi, affecting $>50\%$ of it surface. Thus, mass wasting plays an important role even during the early stages of growth for ocean island volcanoes.

The 1996 earthquake swarm at Lō`ihi has the largest number of recorded seismic events in Hawai`i. It was preceded by at least one eruption and accompanied by the formation of a ~300 m deep pit crater. Seismic and petrologic data indicate magmas were stored in a ~8–9 km deep reservoir prior to the 1996 eruption. The 1996 events led to the venting of high-temperature fluids (possibly $>330\,^{\circ}C$) and the precipitation of a high-temperature sulfide mineral assemblage ($>250\,^{\circ}C$), although maximum measured vent temperatures were only about 200 °C. Venting of the high-temperature effluent was short-lived, lasting less than 3 months. Since 1997, there has been a gradual decrease in the flux of a brine phase and a concomitant decrease in thermal, fluid and H_2S fluxes.

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