INVITED REVIEW

Geology, Geochemistry and Earthquake History of Lō`ihi Seamount, Hawai`i

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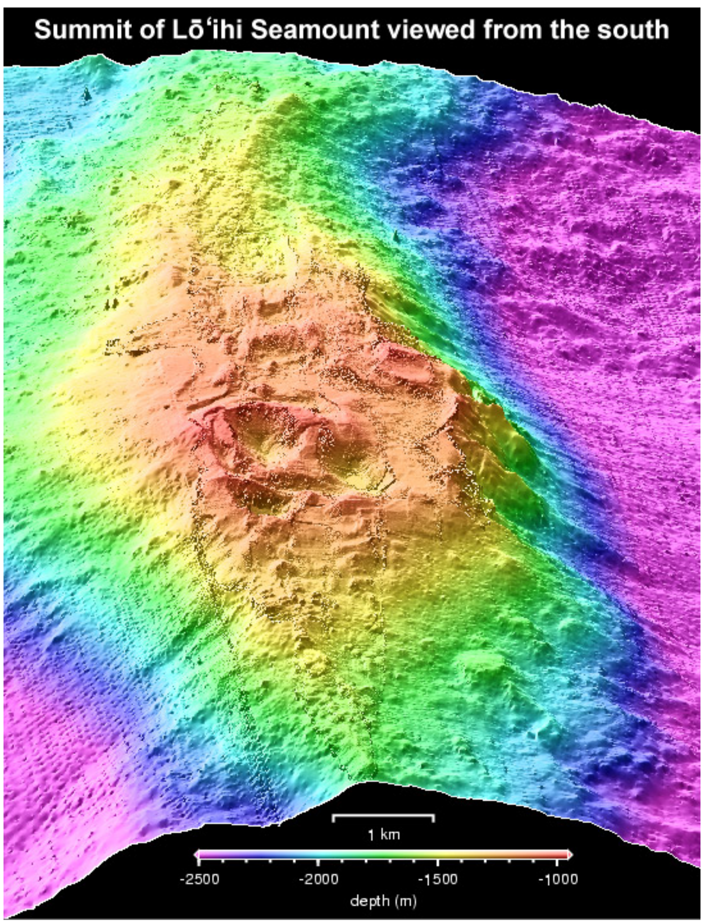

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Abstract

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 1970's led to a dredging expedition in 1978, which recovered young lavas. This led to 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 three months, captured some of the volcano's earthquake swarms. The 1996 swarm, the largest recorded in Hawai`i, was preceded by at least one eruption and accompanied by the formation of a ~300-m deep pit crater, renewing interest in this submarine volcano. 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 led to a refined 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. Seismic and bathymetric data have highlighted the importance of landsliding in the early formation of an ocean island volcano. Lō`ihi's internal structure and eruptive behavior, however, cannot be fully understood without installing monitoring equipment directly 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^{o}C), and on the detection of

water temperature, methane and ^3He anomalies, and clumps of benthic micro-organisms in the water column over the volcano in 1982. Submersible observations in 1987 confirmed a low temperature system (15-30^{o}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 hydrothermal fluids >250^{o}C. Vent temperatures have decreased to ~60oC during the 2004 dive season indicating the current phase of hydrothermal activity may be waning.

Introduction

Lō`ihi Seamount, the youngest and smallest of 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 studies. Its proximity to the island of Hawai`i (Fig. 1) offers an excellent unique opportunity to monitor an active submarine volcano (Klein, 1982; Caplan-Auerbach and Duennebier, 2001b). Here we review existing 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.

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 islands 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$^o$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 chain on Earth, with at least 129 distinct volcanoes (Clague, 1996). The trend of decreasing age to the south was first 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 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.; e.g., 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 obtain by drilling indicate that the hotspot may have drifted rapidly (~4 cm/year) during late Cretaceous to early Tertiary times (81 to 47 Ma) as the Emperor seamounts were formed (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). 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 to 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. 2003), 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 volcano (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. At this stage, as the source once again samples cooler temperatures, magma compositions gradually switch back to alkalic during the post-shield stage (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 to more fractionated lava compositions (hawaiites to trachytes, on some volcanoes such as Kohala; Spengler and Garcia, 1988), as magmas pond at greater depths (~30 km) before eruption (Frey et al., 1991). After about ~1.5 million years of growth (Fig. 2), the volcano dies, having formed one of the largest geologic features on Earth (up to 13 km in height and 80,000 km^3 in volume; Mauna Loa). Many but not all of Hawaiian volcanoes experience a period of renewed volcanism that occurs 0.6 to 2.0 m.y. after the end of post-shield volcanism (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). It should be noted that 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 a more on the geology of Hawaiian volcanoes, the reader is referred to Clague and Dalrymple (1987).

Discovery and Early Work on Lō`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 U.S. 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 Islandss, 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 (e.g., Engel and Engel, 1966).

A large earthquake swarm in 1952 first brought attention to Lō`ihi Seamount. Macdonald (1952) noted that epicenters for this swarm 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), not Emory (1955) as commonly cited, deserves credit for first proposing that Lō`ihi seamount was an active volcano. However, the east-west distribution of the epicenters and the lack of recorded volcanic tremor on seismic stations distant from the volcano led Macdonald (1952) to conclude that the 1952 earthquake swarm was related to a faulting rather than an eruption.

In recognition of its elongated shaped (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 U.S. 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 a 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 1970's (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 revealing 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 deposits 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. This led to the suggestion that Lo`ihi was in the earliest phase of Hawaiian volcanism, reflecting lower degrees of partial melting that produced alkalic magmas (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 (SASS) 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 mbsl, with a summit area containing two prominent pit craters (275 and 256 m deep) and two sub-parallel rift zones extend 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 a 1996 eruption (Loihi Science Team, 1997). Highlights of these expeditions are presented below along with some new observations on the geology, geophysics and geochemistry of Lō`ihi.

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, if Lō`ihi began forming 400,000 years ago, then it is likely to have formed on only a thin veneer of debris from the island of Hawai`i and be considerably thicker (~5 km; Fig. 4). Lō`ihi's summit consists of a small (12 km^2) platform at ~1200 mbsl with several large cones and three, 300 to 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 pit craters walls of thick sections of columnar joints (>20 m), rare in Hawaiian shield volcanoes, suggests that the craters were 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 both craters following an intense earthquake swarm (Loihi Science Team, 1997; Caplan-Auerbach and Duennebier, 2001a). This 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^{o}. In contrast, the deeply dissected east flank of the volcano dips 35-40^{o}, 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 would 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 of the island of Hawai`i and to the trend predicted by models for the formation of rift zones parallel to the rift zones of adjacent volcanoes, which are thought to buttress the younger volcano (Swanson et al., 1976). The rift zones dip more gently that the summit flanks (~10^{o} vs. 14-40^{o}).

Lō`ihi's shorter north rift zone (~11 km long) comprises two subparallel segments (Fig. 3) with numerous 10-30 km high pillow cones. The eastern segment is only ~2.6 km long. 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, observations from submersible dives along both segments indicate little sediment on the cones on both rifts suggesting both have been recently active (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. 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 east 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^{o} 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). 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 (Holcolm and Robinson, 2004). Thus, landsliding is clearly an important process in the evolution of even the youngest Hawaiian volcano.

Ages: Implications for Magmatic Evolution and 1996 Eruption

Radiometric 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 preferred method for dating young lavas, yielded duplicated ages of 5 ± 4 to 102 ± 13 ka for 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 (Guillou et al., 1997). These ages were used to estimate lava accumulation rates of 3.5 mm/yr for the lower part of the section and

7.8 mm/yr for the upper part (Guillou et al., 1997). An increase in lava accumulation rate is consistent with a shift to less silica undersaturated lava composition upsection (~90% to ~20% alkalic) 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 of the overall 3.5 km thick section; 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 ~40% of the volcano has formed in the last 100,000 years. Assuming linear growth, Lō`ihi is possibly 250,000 years old (Guillou et al. 1997). 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 simple exponential growth model, 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.

A shallow magma reservoir has been inferred from the presence of summit pit craters (e.g., Malahoff, 1987), which may have repeatedly formed at Lō`ihi based on the presence of thick sections of columnar jointed lavas in the West Pit crater walls (Garcia et al., 1993). Thus, it is likely that the preshield stage of growth as observed at Lō`ihi is longer than previously assumed (>400 vs. 250 k.y.), which impacts models for the Hawaiian plume melting region. For example, given the northwest drift of the Hawaiian Islands at ~10 cm/yr (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. Thus, Lō`ihi may be taller (~5 km) and more voluminous than previously assumed (1.7 vs. 0.8 x 10^3 km^3).

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 dated using the ^{210}Po-^{210}Pb method which relies upon the fact that ^{210}Po is volatile at magmatic temperatures (Vilenskiy, 1978; Le Guern et al., 1982) and degasses nearly completely (75-100%) from shallow erupted basalts (<2 km deep water; 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 two-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 eruption and represent the Lō`ihi's first documented eruption.

Seismicity

Earthquake swarms

Lō`ihi's 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 HVO seismic network, located 35 to 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 on Lō`ihi's summit near 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 changes in hydrothermal vent temperatures and summit elevation, suggestive of deflation due to magmatic withdrawal (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, hydrophones collected 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 just after 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 and are limited in time to several days. The two exceptions are the 1971-2 activity, spanning many 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 1971-2, earthquakes were seen to migrate 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 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 shallow magma reservoir (Garcia et al., 1998a; Caplan-Auerbach and Duennebier, 2001a).

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 ocean-bottom seismometer (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 5 OBS's was deployed on the summit and flanks of Lō`ihi for approximately one 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 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 velocity 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 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 high-rate hydrophone. HUGO operated between January and April 1998 until a short-circuit in the main cable terminated operations. 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 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. Data from HUGO confirm an observation made based on HVO seismic data, that there was little Lō`ihi seismicity for several years following the 1996 swarm (Figure 6; Caplan-Auerbach and Duennebier, 2001b). In spite of Lō`ihi's quiescence during this time, data from HUGO confirm 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.

Rocks

Sampling and Submersible Observations

The summit and rift zones of Lō`ihi have been extensively sampled providing good spatial and temporal coverage of its volcanism. 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 manned submersibles and KAIKO, a remotely operated vehicle. Thus, the volcano has attracted broad international interest with only the French submersible 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 the 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^o (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^{o}W to N10^{o}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) using 1987 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 and was buried in this sediment prior to being rescued in 2002. 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 (Fig. 9). 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^{o}C) from their rubble (Karl et al., 1988). The uppermost cone was the site of the 1996 collapse that produced Pele's Pit (Loihi Science Team, 1987). 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, 1987). Fresh glassy 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 it collapsed in 1996; 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., 2000a).

Petrography and Mineral Chemistry: Magma History Implications

The petrography of Lō`ihi rocks have been describe in numerous studies (e.g., Moore et al., 1982; Frey and Clague, 1983; Hawkins and Melchoir, 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 Melchoir, 1983; Garcia et al., 1989). There is no correlation of olivine abundance with rock type, although in a subset of submersible-collected alkalic lavas (hawaiites), all with rare olivine (≤0.1 vol.%; Garcia et al., 1995). Olivine is generally euhedral and undeformed, with inclusions of chromite and glass, 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. 11). Clinopyroxene is usually euhedral and commonly sector zoned. Plagioclase is less common in Lō`ihi lavas (e.g., none are present 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., 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 to 43 vol.%) with no obvious correlation with rock type (e.g., Moore et al., 1982; Frey and Clague, 1983; Hawkins and Melchoir, 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) was examined and found to contain 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 the glass and mineral fragments suggest this deposit is young. The silt extends into a hydrothermal field with numerous small chimneys, 0.5-2.0 m high (Malahoff, 1987). However, it is not clear whether the silt is related to the hydrothermal field.

Lō`ihi minerals have been analyzed by electron microprobe in several studies (Hawkins and Melchoir, 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 in relatively calcic magmas (e.g., Garcia et al., 2002).

Clinopyroxene crystals vary markedly in composition in Lō`ihi lavas (Hawkins and Melchoir, 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 available data show that anorthite contents range widely (47-71%; Hawkins and Melchoir, 1983; Garcia et al. 1995). 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 the more evolved alkalic lavas contain Ti-magnetite (Hawkins and Melchoir, 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^{o}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. 10) 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, predominantly 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).

Rock Types and Temporal Magmatic Variation

Lō`ihi lavas have been included in numerous studies characterizing submarine basalts (e.g., Yi et al., 2000; Kaneoka et al., 2002; Boyet et al., 2005). This fascination with Lō`ihi started with the discovery of alkalic lavas in 1981 (Moore et al., 1982), which was unanticipated. The early submarine phase of Hawaiian volcanism was thought to consist of tholeiitic basalt (Macdonald et al., 1983), although it had not be sampled prior to work on Lō`ihi. Subaerial Hawaiian volcanoes had been well studied prior to this expedition resulting in a well established evolutionary sequence for their growth (e.g., Stearns, 1946; Macdonald, 1963). These giant shield volcanoes (20 to 80 x 10^3 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 was based only subaerial studies, ignoring the 5+ km of submarine growth.

The dredged alkalic lavas have on average thicker palagonite alteration rinds than the tholeiitic lavas, although thicknesses overlap for the two rocks groups (1-12 vs. 0.5-4 um; Moore et al., 1982). The alkalic lavas are also more vesicular causing a concern that alkalic lavas might alter faster than tholeiitic lavas (Moore et al., 1982). Nonetheless, the alkalic lavas were presumed to be older, which was in agreement with experimental work and fit a simple melting model whereby early magmatism was alkalic reflecting lower degrees of melting. As the volcano drifted towards the hotspot, 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 examined and collected the walls of the summit pit crater and from 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 overwhelming alkalic (14 of 16 flows), whereas the upper section is mostly tholeiitic (8 of 14 flows). If the pit crater sections are including in this analysis, tholeiites are the dominant recent rock type at Lō`ihi (59 of 75 samples), including during the 1996 eruption (Fig. 13). However, alkalic volcanism has continued until recently based on the presence of several young alkalic cones along the uppermost south rift including Pele's cone, where the 1996 eruption occurred (Garcia et al., 1993). Based on these results and the previous work on Hawaiian volcanoes, a composite Hawaiian volcano cross section illustrates the proportions of the three stages and their rock types (Fig. 12) confirming the model of Macdonald (1963) that alkalic volcanism represents a minor component of Hawaiian volcanoes.

Whole-Rock Compositions

Relatively few whole-rock XRF analyses are available for Lō`ihi lavas compared to glass analyses (Frey and Clague, 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). There is an enormous range in MgO content of these lavas (3.4-25.2 wt.%). 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 explain by the observed modal differences in mafic minerals (~3 vol.%; Garcia et al., 1998a). The relatively high Mg# [Mg/(Mg + Fe^{2+}) x 100] of the 1996 lavas (58 to 63) indicates that they probably were not stored in the crust for significant time periods (Garcia et al., 1998a).

Glass Geochemistry: Magma Chamber Processes

Glass is common on Lō`ihi lavas and has been used extensively to its characterize rock types because it represents liquid compositions (e.g., Moore et al., 1982). The glass data show a remarkable variation, spanning the 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). Hawaiian basalt glass MgO composition 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 that magmas for these glasses were stored under similar conditions resulting in a relatively narrow temperature range (~40^{o}C based on the Kīlauea glass geothermometer; 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). This large range probably indicates that the alkalic magmas were stored for variable periods during times of lower magma supply, which led to ephemeral magma chambers. Magmas with high MgO content 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 tholeiites overlap with Kīlauea tholeiites (Fig. 15).

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). This attention resulted from the early and exciting He isotope evidence that Lō`ihi magmas were derived from a relatively primitive source (Kurz et al., 1983). Early work on the dredged samples showed the glasses had relatively high S contents (0.11-24 wt.%) despite the presence of S globules and high vesicularity in some samples (Moore et al., 1982). In general, the glasses show a good correlation of H_2O with K_2O content 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). Thus, extensive but variable CO_2 degassing is thought to have occurred. Cl concentrations are even more erratic and are not correlated with K_2O, with some very high concentrations for Hawaiian basalts (up to 0.17 wt.%; Byers et al., 1985; Kent et al., 1999). The high concentrations were interpreted as evidence for widespread assimilation of a seawater-derived component, probably brines (Kent et al., 1999). These high Cl concentrations were thought to be beyond those in glasses from adjacent Hawaiian volcanoes suggesting that seawater contamination of magma is more likely during the preshield stage of volcanism (Kent et al., 1999). Even higher Cl concentrations were found in some Mauna Loa glass inclusions (up to 0.36 wt.%; Davis et al., 2003). High Cl correlated with high F contents in these inclusions, which led to the suggestion that hydrothermal deposits rather than brines have contaminated the Mauna Loa magmas (Davis et al., 2003). Little F data are available for Lō`ihi glasses to test this possibility.

Stable isotope data for Lō`ihi glasses are limited and provide equivocal evidence for seawater contamination. Only two samples with moderate to high Cl contents have been analyzed for hydrogen isotopes yielding δD values of -69 and -84, which are thought to be representative of the normal mantle (Garcia et al., 1989). Oxygen isotope values for 16 glasses are relatively low ($\delta^{18}O$ of 4.7-5.2) compared to those from mid-ocean ridge basalts (5.3-6.0; Ito et al., 1987). However, these 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).

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). Analyzes of glasses show systematic variations for highly incompatible trace elements regardless of rock type (Fig. 16), suggesting that these rock types shared similar sources (Garcia et al., 1995). Wide variations are noted 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).

These results are consistent with a simple model of variable amounts of partial melting of a common source (see isotope discussion below) for to produce parental magmas (higher degrees produce tholeiitic magmas with lower ratios; e.g., Allègre and Minster, 1978). However, these ratios can be somewhat modified by clinopyroxene fractionation (Fig. 16).

Glasses from the ~350 m thick East Pit stratigraphic tholeiitic 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 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 core, where individual cycles spanned thousands of years of the flanks of Mauna Kea volcano (Yang et al. 1996; Blichert-Toft et al., 2003), and for the historical lavas at the summit of Kīlauea, where recent individual cycles may be hundreds of years long (Pietruszka and Garcia, 1999a). These trace element trends correlate with Pb and Sr isotopic variations indicating that the cyclicity is related to changes in proportions of the source components, which may be controlled by melting processes (Pietruszka and Garcia, 1999b). Another interesting corollary with other Hawaiian volcanoes is the reversal in the La/Yb ratios following the collapse of the volcano'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 need to explain the isotopic variations in Hawaiian basalts (Staudigel et al., 1984). This and subsequent studies (Garcia et al., 1993; 1995; 1998a) found no correlation of these isotopes with major elements (i.e., 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 these alkalic rocks were formed by high pressure clinopyroxene crystallization 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 rock groups (e.g., La/Yb; Fig. 16) could not be explain by high-pressure clinopyroxene fractionation. Modeling of incompatible trace element concentrations in lavas with the similar Sr, Nd, Pb isotopic ratios, using a tholeiite to indicate the source composition, and assuming nonmodal equilibrium melting of a garnet peridotite produced a range of melt compositions reflecting varying degrees of partial melting (Fig. 19). The shape of the trace element patterns from melting the tholeiitic source (dotted lines in Fig. 19) parallel an alkalic lava for ~8% melting. Although the true extent of partial melting for alkalic magmas may be different, the modeling demonstrates that variable degrees of partial melting of a common source could explain the range of rock types erupted at Lō`ihi.

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. Due to these unusual isotopic compositions and the availability of quenched submarine glasses suitable for volatile measurements, Lō`ihi is among the best studied volcanoes for noble gases.

Helium

The first reported helium measurements from Lō`ihi basalts revealed $^3He/^4He$ of ~32 times the atmospheric value [Ra] (Kurz et al., 1982). This value is far above the average value for mid-ocean ridge basalts (MORB) of ~8 Ra, and created considerable interest in Lō`ihi among noble gas geochemists. More detailed studies on the dredged lavas yielded $^3He/^4He$ ranging from 20 to 32 Ra, with higher values in the tholeiites than the alkali basalts (Kurz et al., 1983; Kaneoka et al., 1983; Rison and Craig, 1983).

The initial interpretation of these high $^3He/^4He$ values was that unradiogenic helium indicated undegassed mantle, i.e. having high He/(Th+U) since the formation of the Earth, and is 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 et al., 1997), coupled with evidence for penetration of subducted slabs into the lower mantle (van der Hilst and Karson, 1999). Alternative models for 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 (rather than undegassed). This residue could then retain unradiogenic helium isotopic signatures (e.g., Anderson, 1998; Meibom et al., 2003). This debate is far from resolved and has considerable geodynamic importance.

Shield tholeiites from neighboring Hawaiian volcanoes commonly have $^3He/^4He$ higher than MORB, but no other Hawaiian lavas are as unradiogenic as Lō`ihi seamount. Existing data from older volcanoes in the Hawaiian Emperor chain (i.e., 45-75 Ma) include $^3He/^4He$ 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, temporal variations within Hawaiian shields, so a small number of samples do not necessarily characterize a volcano. In every case where age or stratigraphic data are available for Hawaiian volcanoes, the oldest shield building tholeiites have the highest $^3He/^4He$ ratios. 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 (Mukhopadyay et al., 2003). These studies vary in their time scales, although higher $^3He/^4He$ values were found in the earliest shield lavas, with later tholeiites and alkali basalts approaching MORB values. A temporal trend for Hawaiian shield volcanoes is evident for He, Sr, and Nd isotopic composition when comparing Lō`ihi and Mauna Loa, the adjacent volcano (Fig. 1). The oldest Mauna Loa lavas (~ 250 ka in age) are closest in isotopic composition to Lō`ihi, whereas younger Mauna Loa lavas have successively lower $^3He/^4He$ with time (Fig. 20). Temporal helium isotopic evolution has not been observed at Lō`ihi, possibly due to the lack of age control for most analyzed samples. However, Garcia et al. (1998a) reported a $^3He/^4He$ of 26 Ra from the 1996 near-summit eruption, which is in the range of the other reported data, and shows that there has been no significant recent change in Lō`ihi helium isotopes. The high $^3He/^4He$ in Loihi 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 represents the present-day center of the Hawaiian hotspot (e.g., Kurz et al., 1983; 2004; Kaneoka, 1987).

Argon, Xenon and Neon

The first heavy noble gas studies of Lō`ihi seamount revealed argon and xenon isotopic compositions close to air (Allègre et al., 1983). 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 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 $^{129}Xe/^{130}Xe$ was 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 difficult to eliminate the possibility that the Lō`ihi Xe isotopic data were affected by late stage atmospheric contamination rather than I/Xe systematics. Ar and Xe 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 contamination. Staudacher et al. (1991) countered that the Lō`ihi glass abundance ratios were not consistent with atmosphere or sea water interaction and defended the original interpretation. It is plausible that Ar and Xe isotopes in the Lō`ihi glasses reflect some atmospheric or seawater contamination effects, given the other evidence for seawater influence (e.g., 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. Similar values have been 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), but this important observation was first made at Lō`ihi seamount. Neon data for Loihi samples from three different laboratories and sample suites form a 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. 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 most of the variations along the linear arrays are caused by atmospheric contamination. This set of assumptions leads to a single neon Lō`ihi neon isotopic composition (Fig. 21), despite the range in helium and other isotopes for these samples. The Lō`ihi mantle source neon isotopic composition is clearly less radiogenic than MORB, which is consistent with the helium data.

A concern was raised about the possible effects of degassing on the Lō`ihi noble studies, which were carried out on samples mostly from shallow depths near the summit. This led to new noble gas measurements on samples from the deeper flanks of Lō`ihi (Valbracht et al., 1997; Kaneoka et al., 2002). Like the earlier studies, they found Ar and Xe isotopic compositions close to atmosphere ($^{40}Ar/^{36}Ar$ ranging from 296 (air) to 2600), and total gas concentrations similar to the earlier studies (Fig. 21). Another issue was the relatively low helium concentrations in Lō`ihi glasses. This is important because undegassed sources (i.e. high He/(Th+U)) should also have higher helium concentrations [Mark, reference for this idea?]. 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.

Trieloff et al (2000) observed non-atmospheric xenon in two Lō`ihi dunite xenoliths. Although it is unclear how the xenolith xenon relates to the mantle source of the basalts (helium isotopes are similar), the $^{129}Xe/^{130}Xe$ ratios are not distinguishable from MORB xenon. This is important because the xenolith data contrast with the near-atmospheric Xe isotopic values obtained for Lō`ihi glasses, and may indicate that the glass xenon data was influenced by atmospheric contamination, or that the xenon in the xenoliths was derived from the MORB lithosphere. 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.

In summary, Lō`ihi is one the world's best studied oceanic island volcanoes for noble gases. He and Ne isotopes show that the source for Lō`ihi is distinct from the MORB source and define an important unradiogenic mantle source. The three isotopes of neon allow a clear elucidation of the mantle and atmospheric components. It seems likely that Ar and Xe isotopes in Lō`ihi glasses have been influenced to some extent by outgassing and atmospheric contamination. Therefore, the original interpretation, that Lō`ihi noble gas isotopic compositions represent the lower mantle, must be viewed with caution. However, the noble gas measurements have fundamental importance to geodynamic models, and there is an ongoing debate on the origin of unradiogenic helium and neon isotopes in the mantle. Some recent papers advocate a layered mantle, with unradiogenic noble gases derived from the lower mantle (e.g., Allègre and Moreira, 2004), while 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). The Lō`ihi noble gas data provide a benchmark for ocean island basalt studies and remain central to debates regarding the inner workings of the planet.

Geochemistry of Hydrothermal Minerals and Sediments

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

The hydrothermal deposits encrusting lava dredged in 1982 from the northeast corner of Lō`ihi's summit yielded smectite formation temperatures of 31-57^{o}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 vent fluids percolate upward through lava talus and mix with oxidizing, 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 Lō`ihi, which would indicate the presence of a high-temperature hydrothermal system (Malahoff et al., 1982; De Carlo et al., 1983).

Following the formation of Pele's Pit in summer of 1996, sulfide minerals were found and temperatures up to 200^{o}C were measured at vents within the pit confirming the presence of an extensive high-temperature hydrothermal system within Lō`ihi. The millimeter- to centimeter-sized sulfide samples included brassy-yellow minerals (pyrite or marcasite), less common shiny black hexagonal zinc sulfide (wurtzite) and clear, euhedral barite crystals (Davis and Clague, 1988). Electron microprobe analysis of brassy minerals 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 are consistent with the existence of high-temperature (>250^{o}C) hydrothermal fluids at Lō`ihi, the first documented evidence of high-temperature fluids at any hotspot volcano (Davis and Clague, 1998). This sulfide mineral assemblage is similar to that found in black smokers at mid-ocean ridge spreading centers (Craig and Scott, 1974). The Lō`ihi 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 the 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^{o}C vented in the talus (Fig. 9C). The barite crystals 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 similar to those found in white smokers at mid-ocean ridges, although the $\delta^{34}S$ of the Lō`ihi sulfides are lower and attributed to lose of magmatic sulfur (Davis et al., 2003a). The dissolution features observed on some sulfides suggest that these minerals are unstable (Davis et al., 2003a), which may explain why they were not seen prior to the 1996 event. The temperature of vent fluids in Pele's Pit appears to be decreasing from a high of ~200 in 1996 to a high of ~60^{o}C during the 2004 dive season suggesting the current phase of hydrothermal activity may be subsiding.

Geochemistry of Hydrothermal Solutions

The first ocean water column geochemical anomalies were observed over Lō`ihi in 1982 (Horibe et al., 1986). The presence of extensive hydrothermal plumes was confirmed by more extensive sampling in 1985 in the summit area, which found water rich in methane (up to 569 ppm), He (91.8 nl/l, a record 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 (Sakai et al., 1987; Gamo et al., 1987). Comparison with other submarine hydrothermal areas showed that the Lō`ihi hotspot has 10-100 times lower $CH_4/^3He$ ratios than the EPR and Galapagos Spreading Center, implying a more 'primitive' source. 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^oC) in 1987 using titanium samplers revealed strong enrichments and depletions in the same species as observed in the ocean water in 1985, and enrichments in dissolved Li, PO_4^{3-}, NH_4^+, Rb and Ba (Karl et al., 1988). The greatest difference between the composition 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 is thought to be responsible for their high dissolved Fe concentrations (~1 mM), two to three orders of magnitude greater than in fluids from low temperature hydrothermal vents on Axial Seamount and the Galapagos Rift (Karl et al., 1988). The high concentrations of dissolved Fe probably is responsible for the precipitation of the abundant low-temperature iron-rich deposits around Lō`ihi vents (e.g., Malahoff et al., 1982).

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 to 27.0 Ra) are both indicative of a magmatic source contributing to the vent fluids (Sedwick et al., 1994). The fluids collected during this five year period were also enriched in dissolved Si, CO_2, H_2S, alkalinity, K, Li, Rb, Ca, Sr, Ba, Fe, Mn, NH_4^+ but depleted in SO_4^{2-}, O_2, Mg, $^{87}Sr/^{86}Sr$, 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^oC) 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 into Pele's Pit. To better document the fluids from these new vents, OsmoSamplers were deployed at two sites within 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 was 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^{o}C, the boiling point of seawater at vent depth, 1325 mbsl) with magmatic volatiles added 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 0.8 to 58, with low values observed in vents 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 umol/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^{o}C) from Pele's Pit display enhanced alkalinities over bottom seawater, the alkalinity remained much lower that observed in fluids collected prior to the 1996 event (Wheat et al., 2000). This contrasts sharply 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 and a given fluid Si concentration was related to the CO_2 content of fluids, which promotes weathering of basalt and controls fluid acidity. The acidity level is thought to control the Fe/Mn ratio of the fluid (Wheat et al., 2000).

In summary, the 1996 Pele's Pit collapse resulted in a dramatic change in Lō`ihi's hydrothermal system with substantial venting of volatiles and hydrothermal fluids. Evidence of phase separation and segregation suggest temperature in excess of 330°C causing the formation of sulfide minerals. Venting of such effluent was not sustained (i.e., lasted <3 months; Wheat et al. 2000) and there has been a gradual decrease in the flux of a brine phase and a concomitant decrease in thermal, fluid and H_2S fluxes.

Conclusions

Although Lō`ihi Seamount was discovered in 1952 following an earthquake swarm, it was largely ignored for two decades until two earthquake swarms in the 1970's renewed interest in the volcano and instigated a reconnaissance survey in 1978. During this survey, young lavas were photographed and dredged, confirming that Lō`ihi is a Hawaiian volcano rather than a Cretaceous seamount. Lō`ihi's small size, location south of the other active Hawaiian volcanoes, and 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 it geology and hydrothermal vents. The HVO seismic network has continuously recorded Lo`ihi's background seismicity as well as 12 earthquake swarms to date. Geophysical monitoring included a real-time submarine observatory that continuously monitored its seismic activity for three months.

The 1996 earthquake swarm at Lō`ihi, the largest recorded in Hawai`i, 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^{o}C) and the precipitation of a high-temperature sulfide mineral assemblage (>250^{o}C).

Studies on Lō`ihi have altered conceptual models for the growth of Hawaiian and other oceanic island volcanoes, and led to a refined 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. Seismic and bathymetric data have highlighted the importance of landsliding in the early formation of an ocean island volcano. Lō`ihi's internal structure and eruptive behavior, however, cannot be fully understood without installing monitoring equipment directly on the volcano.

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Figure Captions.

COVER: Oblique view of bathymetric data for the summit region of Lō`ihi Seamount, Hawaii, shown without vertical exaggeration. View direction is due north and artificial illumination is from the northwest.

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. The summit of each volcano is indicated by a star symbol. Box around Lō`ihi indicates area shown in Figure 3. Inset: location of this figure in relation to the major islands of the Hawaiian archipelago and its capital, Honolulu.

Fig. 2. Growth history model for a Hawaiian shield volcano. This composite model is based on volume estimates for each stage (large boxes). Magma supply 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).

Fig. 3. Bathymetric contour map with shaded relief of Lō`ihi Seamount. Map data 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. Note the well-defined south rift zone that curves to the east and the two-pronged north rift, with a shorter eastern segment. Rift axes are marked by dashed lines. The longer western segment of the north rift curves to the east. Box shows the location of Fig. 5, a detail of the summit area. North (N) and south (S) arrows show the location of the cross section in Fig. 4.

Fig. 4. Cross section of Lō`ihi Volcano drawn without vertical exaggeration showing the inferred basement for the volcano including southern flank of the island of Hawai`i, the Cretaceous (~105 Ma; Waggoner 1993) Pacific oceanic crust and 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 pod) are shown. The location of the new Pele's pit crater is indicated by a notch near the inferred vent location for the 1996 breccia.

Fig. 5. Bathymetric contour map with shaded relief of the Lō`ihi Seamount summit. Contours interval is 50 m with annotations every 250 m, and artificial illumination is from the northwest. Lō`ihi's summit platform is defined by the 1200 m contour. The pit craters are labeled: W (West), E (East) and P (Pele's). The location where 1996 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 Figure 3 caption.

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%; younger date) initial extent of Po degassing. Also shown is the error in maximum age (black 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 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 for reference. Modified after Garcia et al. (1998a).

Fig. 7. Monthly earthquakes located at Lō`ihi based on detection by the HVO seismic network. No data exist between 1953 and 1959. 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.

Fig. 9. Photos of Lō`ihi outcrops. A. Young bulbous pillows with deep-sea coral and brittle star from upper north rift zone. B. Steeply dipping dike (~75^o) intruding pillow lava, East Flank at ~1450 mbsl. C. Hydrothermal venting at Pele's Pit in 1997 creating barite mounds. 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 Nov. 2002 by JASON2. Photos A-C taken by PISCES V cameras; Photo D taken by JASON2 camera.

Fig. 10. Photomicrographs of Lō`ihi basalts. A. Crossed nicols view of a clinopyroxene (CPX) with two small, elongate olivine inclusions. 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 euhedral olivine crystals with inclusions of chromite and glass set in clear brown glassy matrix. The scale is the same for both photos.

Fig. 11. Synthetic phase diagram created using the MELTS program (Ghiroso and Sack 1995) for lava from the 1996 eruption. 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., 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^{o}C increments to determine the crystallization sequence at that pressure. Olivine shows a limited pressure stability (<0.3 GPa) for this bulk composition. At pressures of 0.22-0.28 GPa, olivine dissolved in the melt once clinopyroxene begins to crystallize and completely dissappeared after 30^o to 40^{o}C.

Fig. 12. Cartoon 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 based on data from Frey et al. (1990). Note the section is two times vertically exaggerated.

Fig. 13. Rock type variation with depth based on stratigraphic sections collected from the East summit pit crater and the 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. 15). 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. Note the dramatic increase in the percentage of tholeiitic lavas above 1300 mbsl. Rock type information from Garcia et al. (1993; 1995).

Fig. 14. SiO_2 versus total alkalis (Na_2O + K_2O) plot for 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). Two glasses from the 1996 breccia are shown by the stars. Trend lines are shown for olivine fractionation from parental alkalic and tholeiitic compositions.

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 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).

Fig. 16. Ta variation diagrams for Lō`ihi summit and east flank glasses. Arrows point in the direction of increase of the process (extent of partial melting, and fractionation of olivine or clinopyroxene). 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 in a small field. The location and size of the field indicate that these glasses were derived at similar amounts of partial melting and experienced limited amounts of crystal fractionation. In contrast, the alkalic glasses (stippled fields) experienced lower degrees of partial melting but greater extents of crystallization of both olivine and clinopyroxene. Two-sigma error bars are given in the lower right corners of each plot. Data 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) and from the Pisces Peak area adjacent to the West Pit. Note the break in section between the East Pit section and uppermost surface samples west and north of the West Pit, including the 1996 eruption (stars). Only samples with >6.7 wt% MgO are

shown to minimize the effects of clinopyroxene fractionation. The East Pit section temporal variation (stippled band) appears to have reversed during of prior to the 1996 eruption. A two-sigma error bar is given in the lower right corner of the plot.

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 summit field. Two-sigma error bars are shown in the upper left 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.

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 high 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. Data sources for Lō`ihi: Kurz et al., 1983; Staudigel et al., 1983; Garcia et al., 1998a; Mauna Loa: Kurz and Kammer, 1991; Kurz et al., 1995.

Fig. 21. Neon isotopes from Lō`ihi seamount glasses determined by step heating and crushing in vacuo in 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. In the

case of step-heating experiments, this is usually the highest temperature step. Neon released at lower temperatures usually has higher uncertainties and is very close to air. Present-day solar wind neon is indicated by the upper left hand rectangle. No difference was found between the results the samples collected on the shallow (Honda et al., 1993) and deeper part of the volcano (Valbracht et al., 1997; Kaneoka et al., 2002).

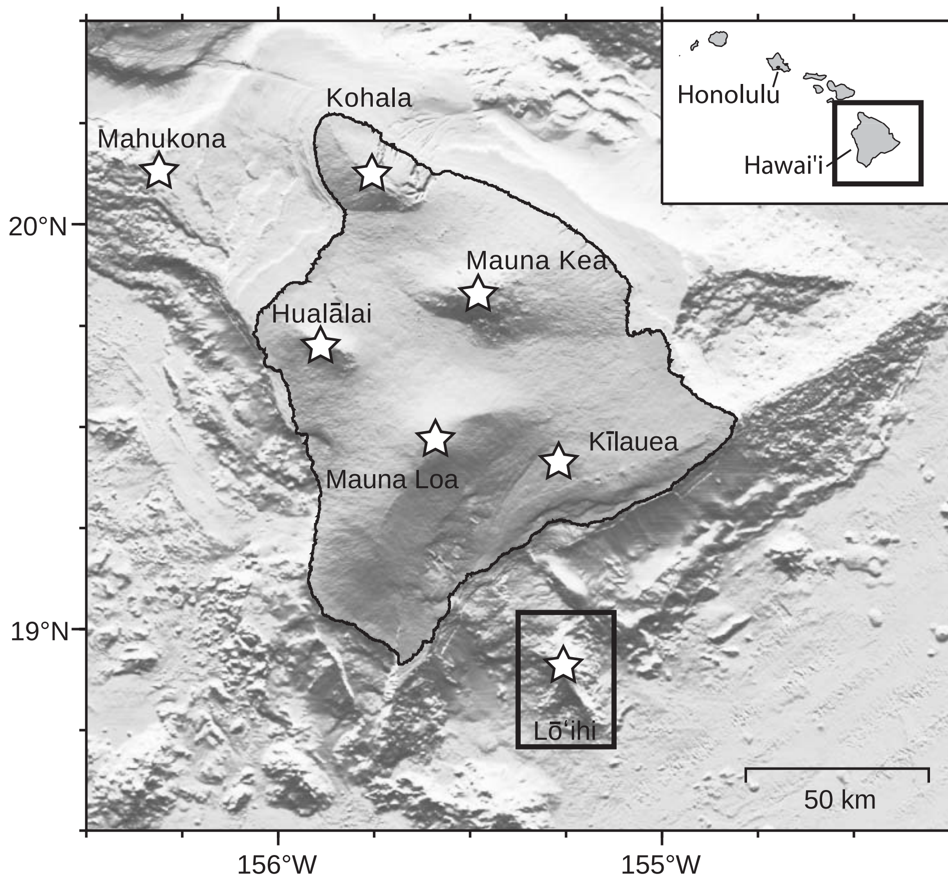
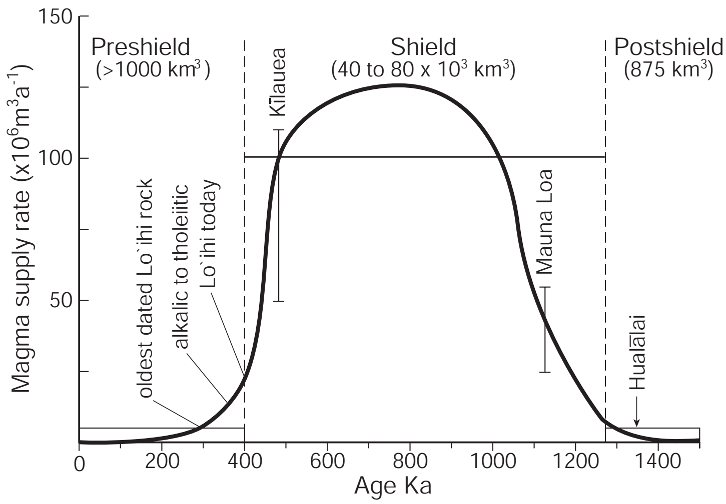


Figure 1. Garcia et al.



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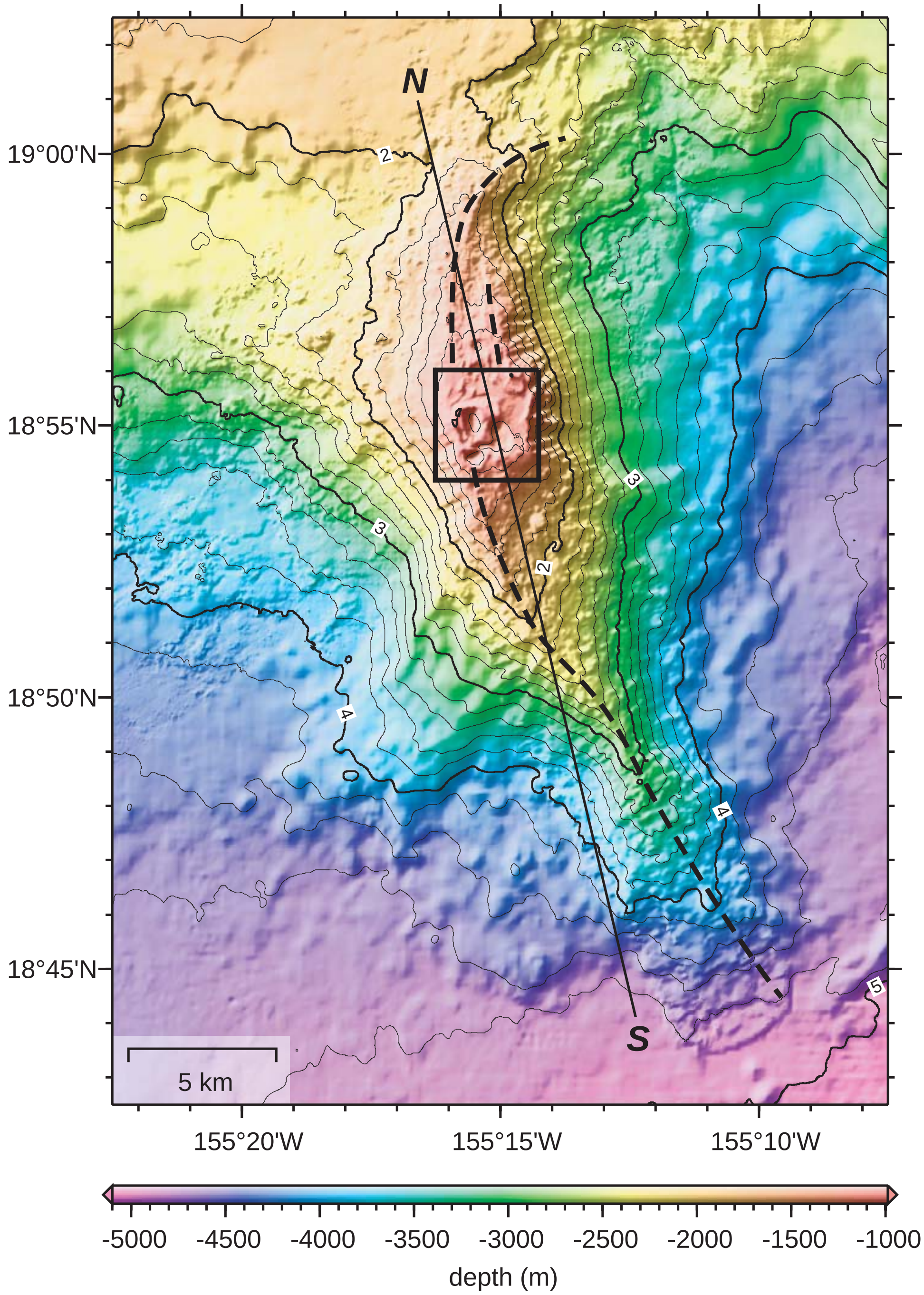
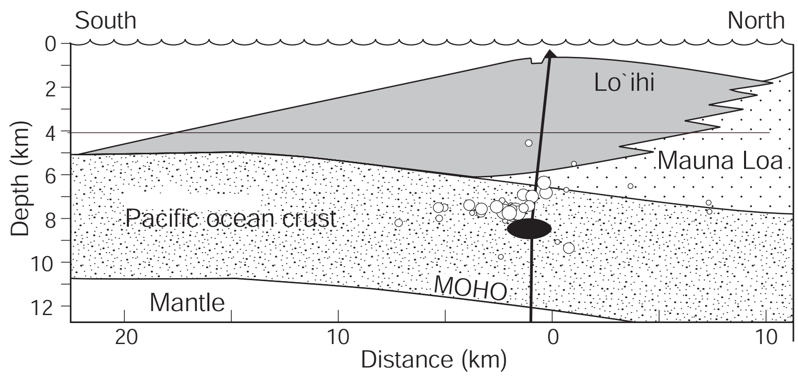


Fig. 3 Garcia et al.



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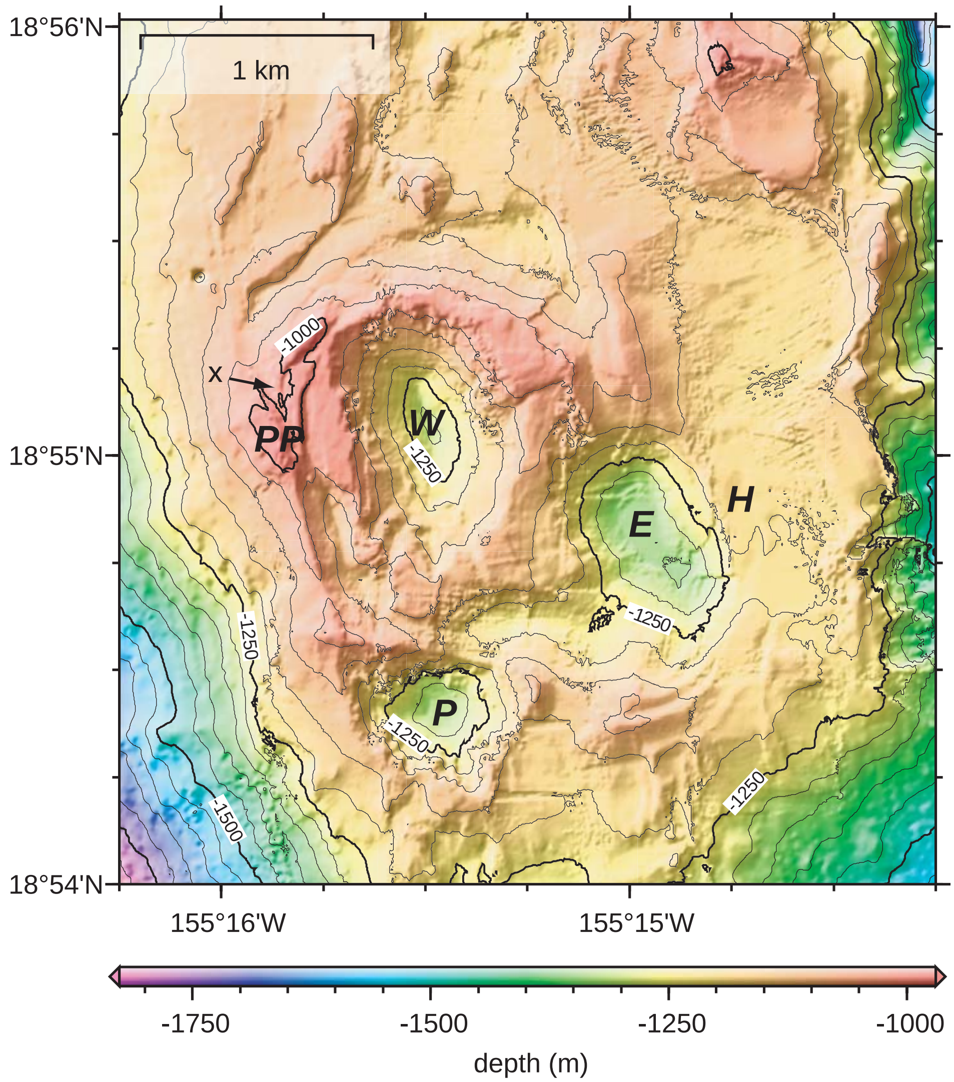
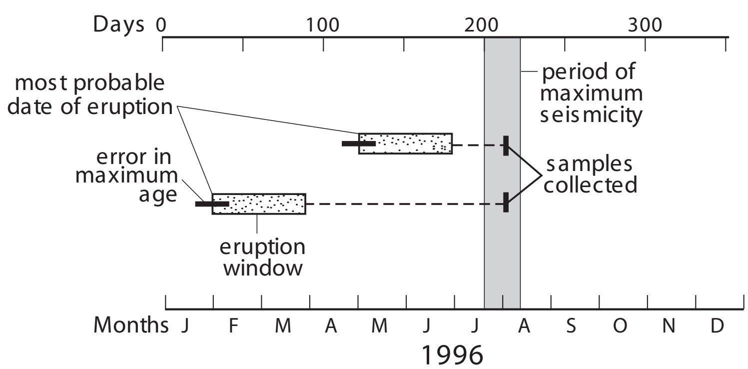


Fig. 5 Garcia et al.



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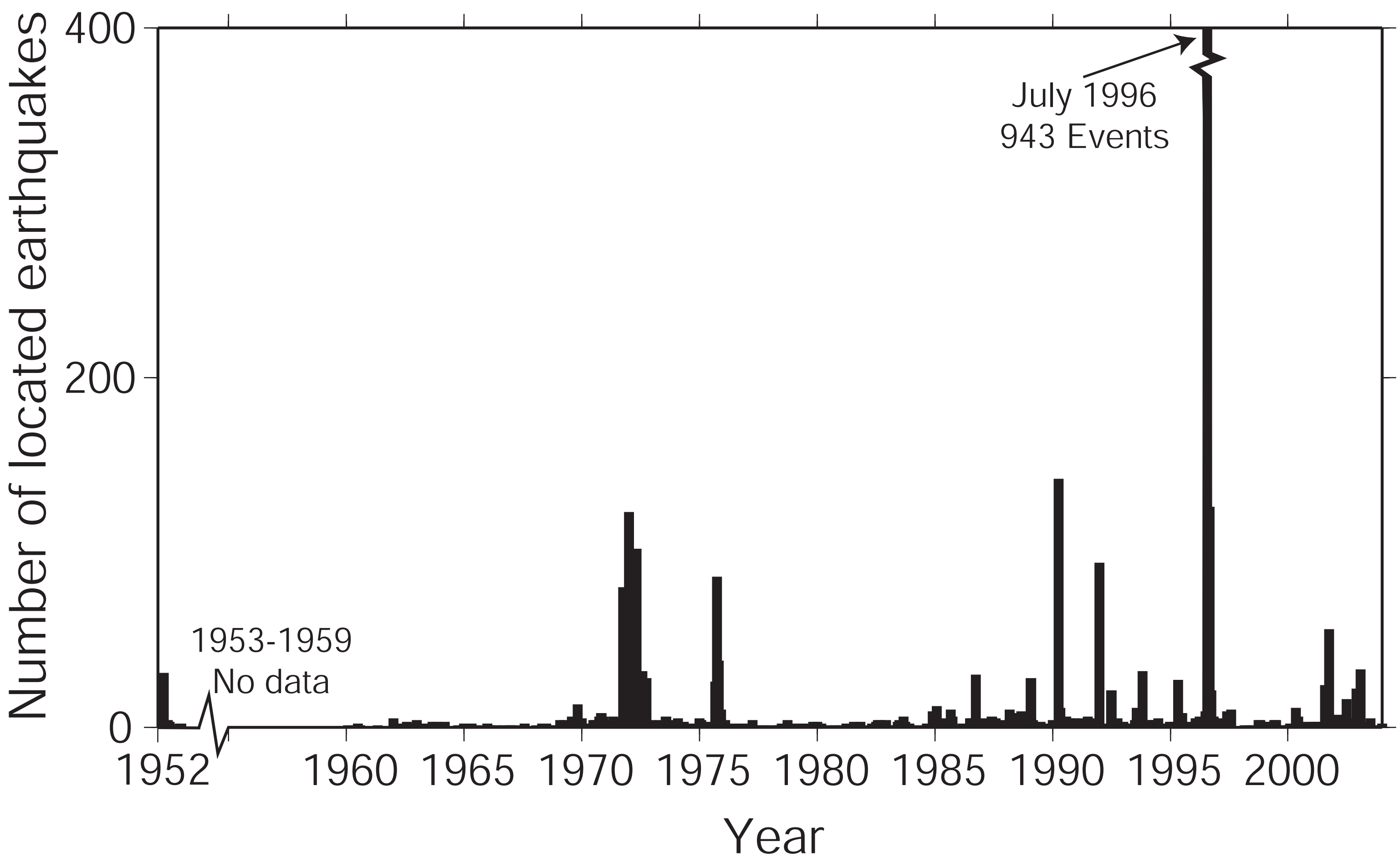

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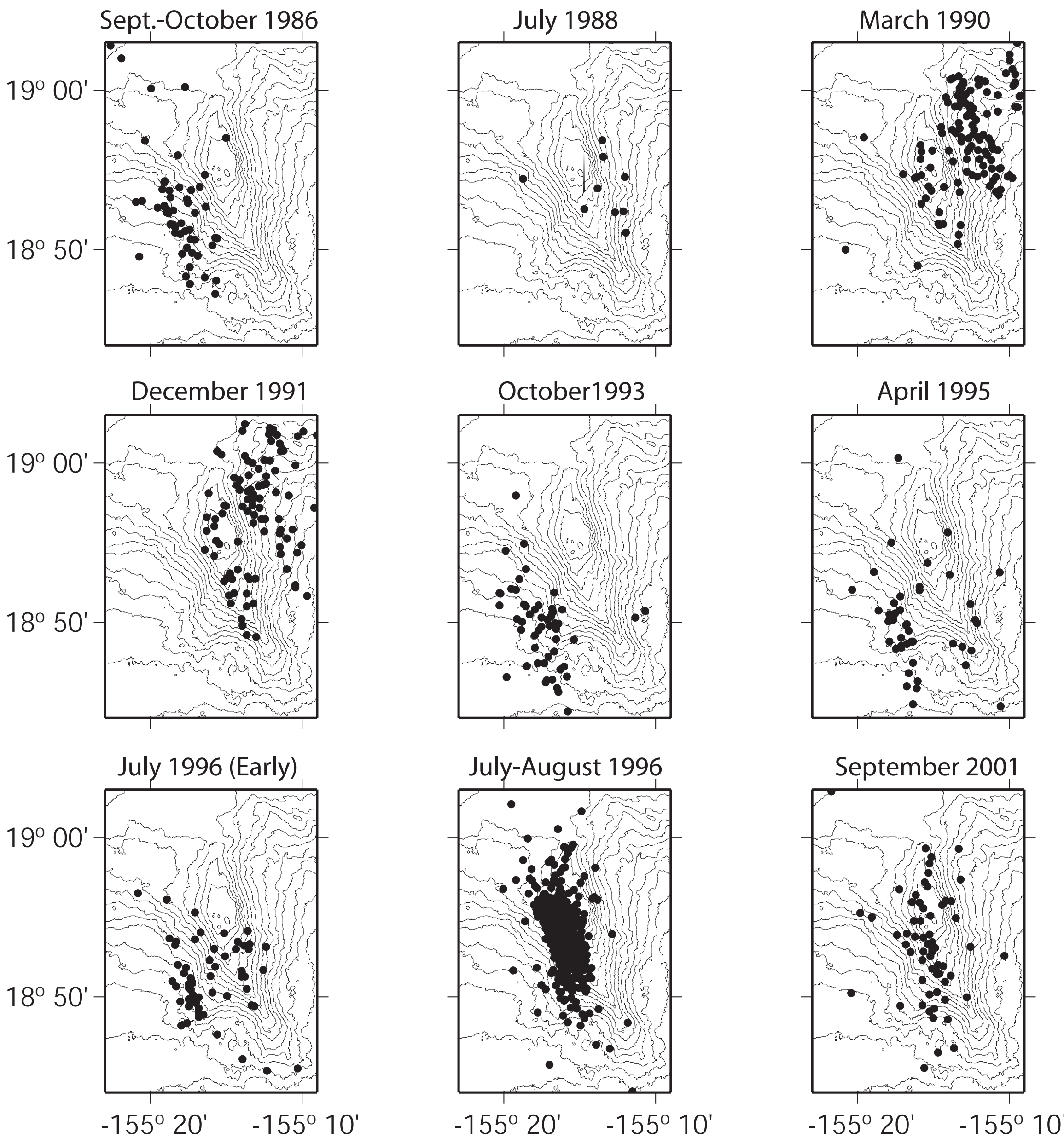

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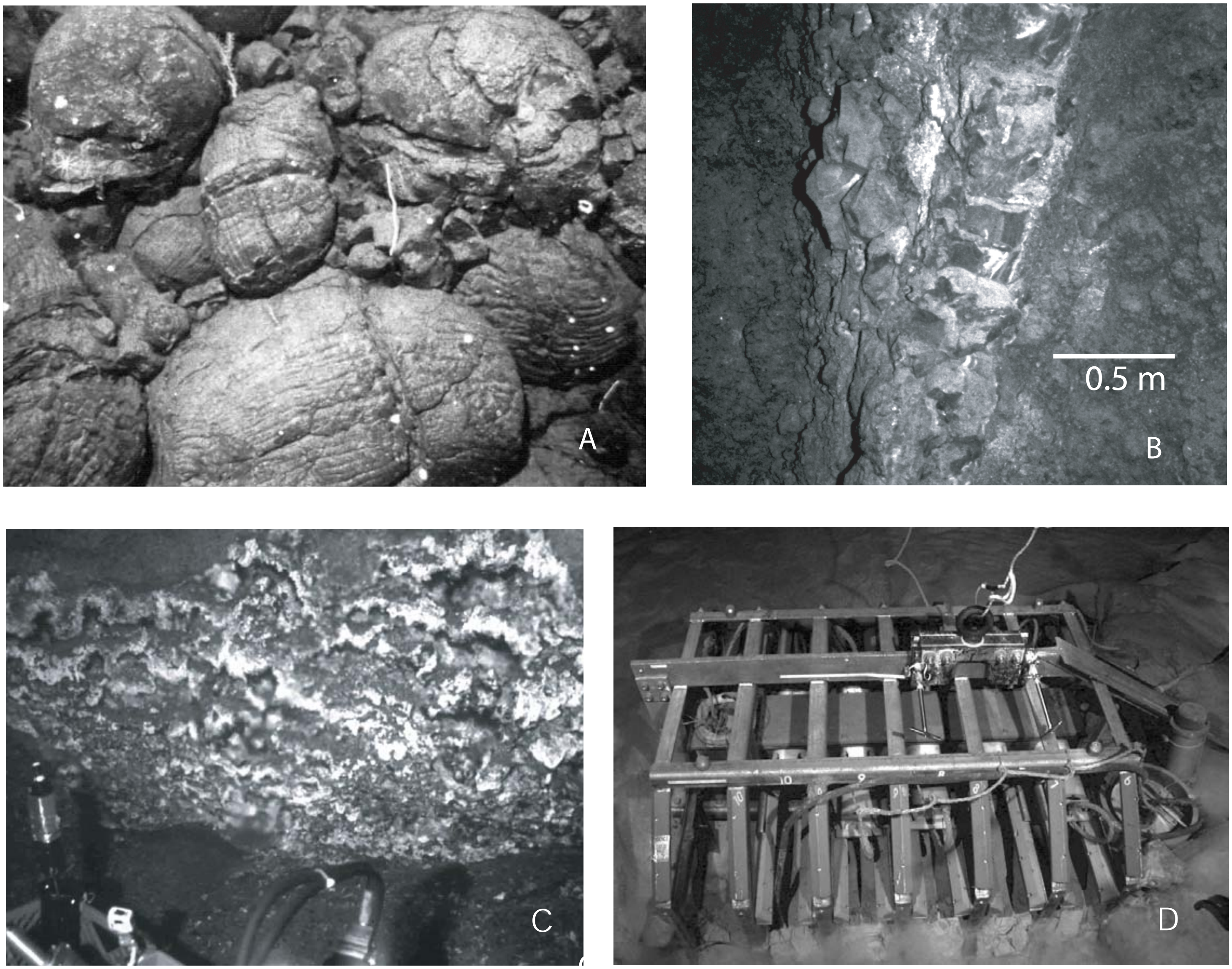
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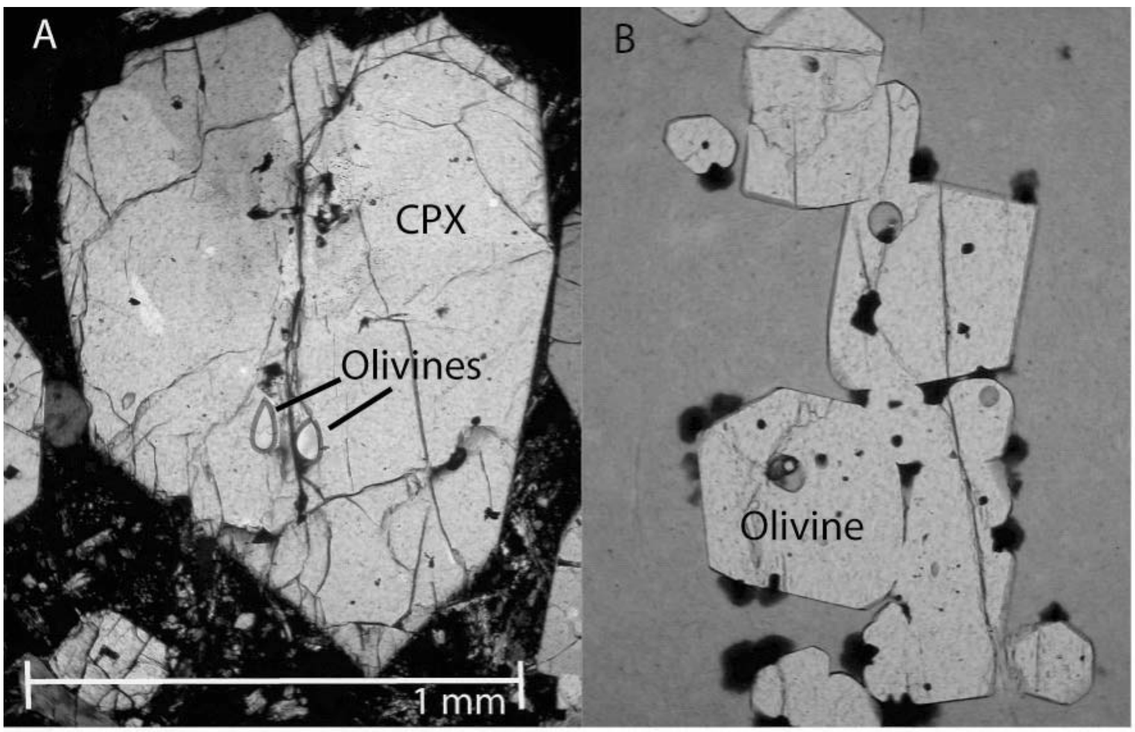

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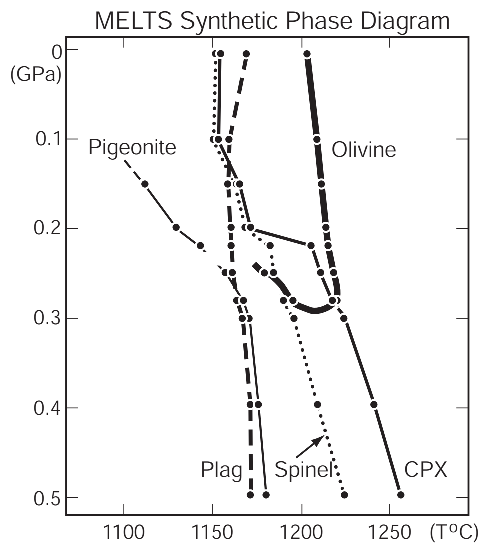

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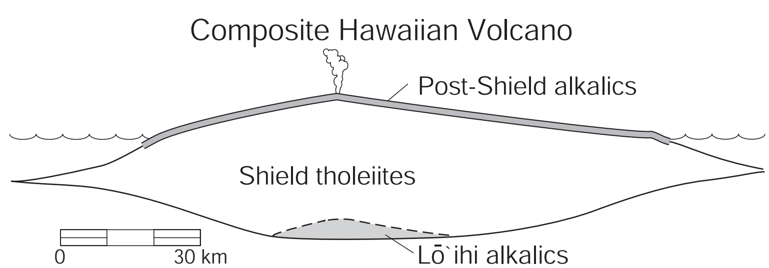

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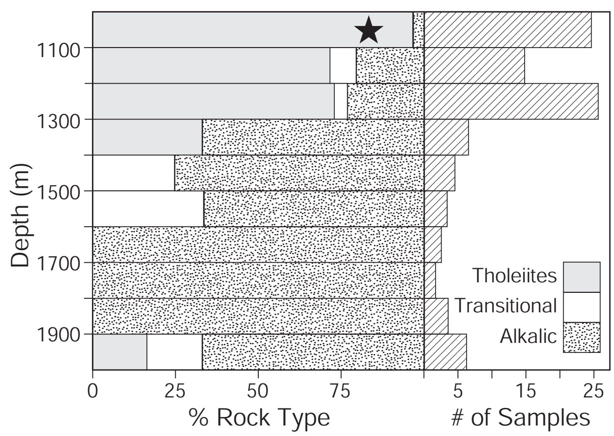

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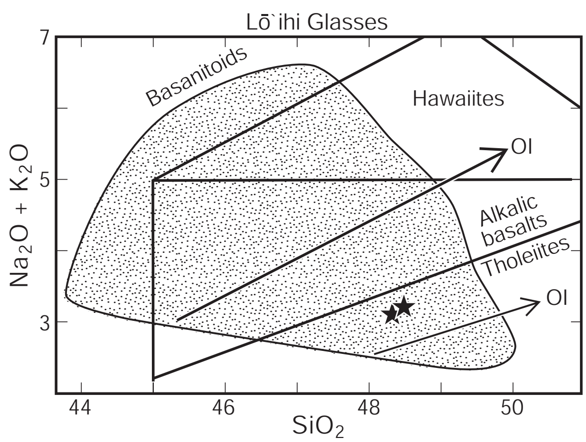

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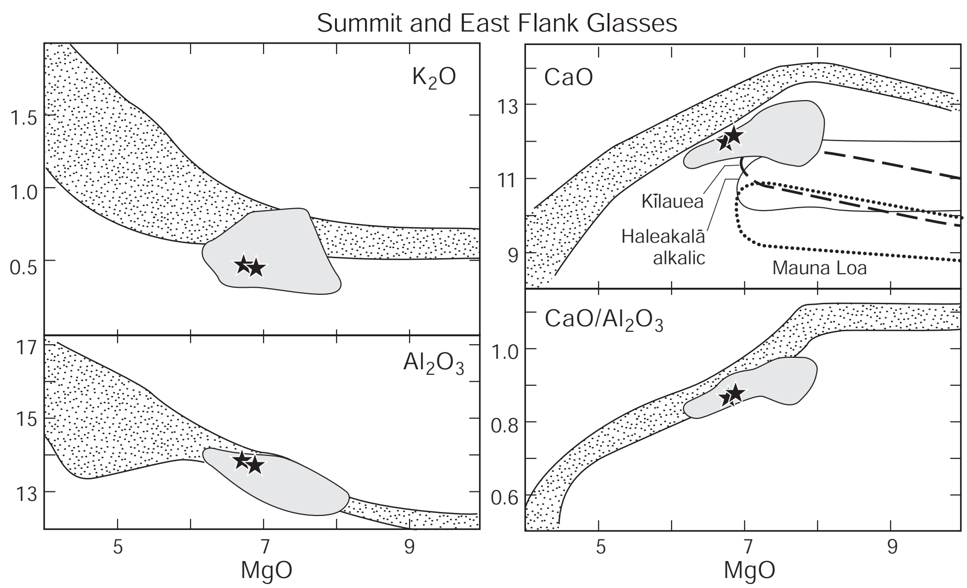

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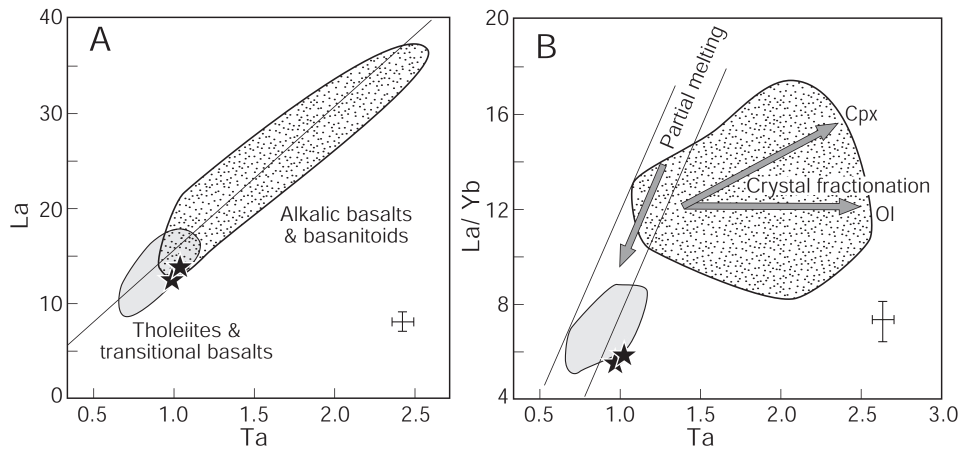

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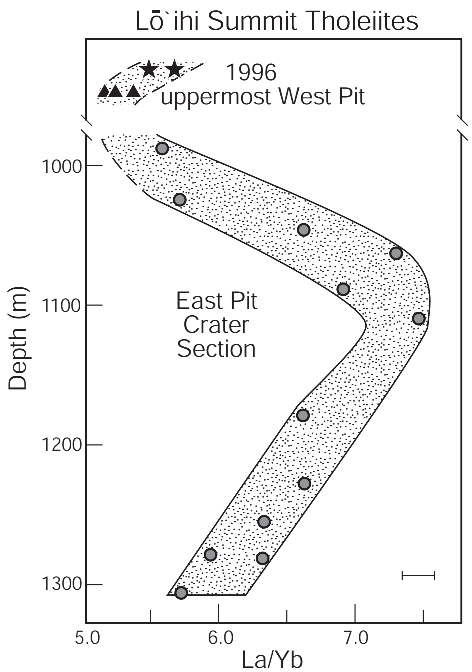

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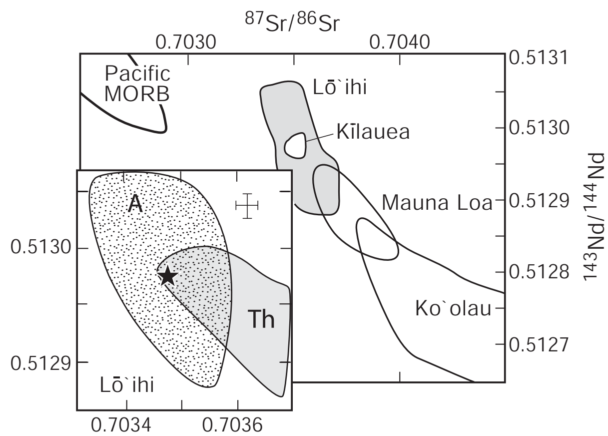

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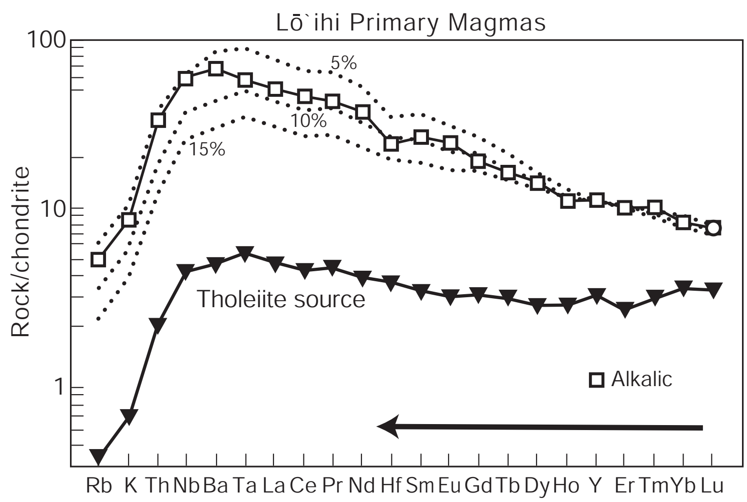

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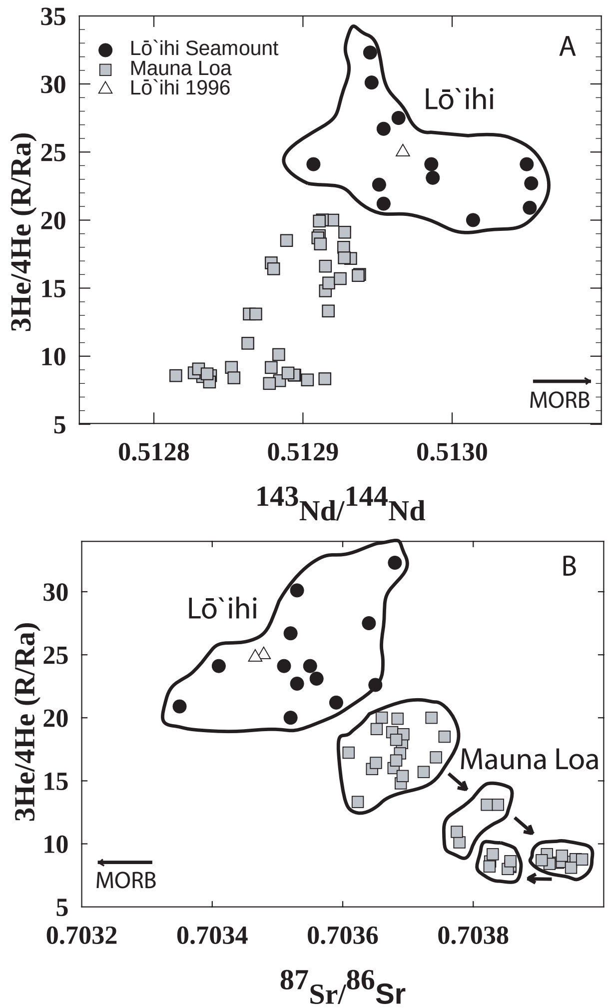

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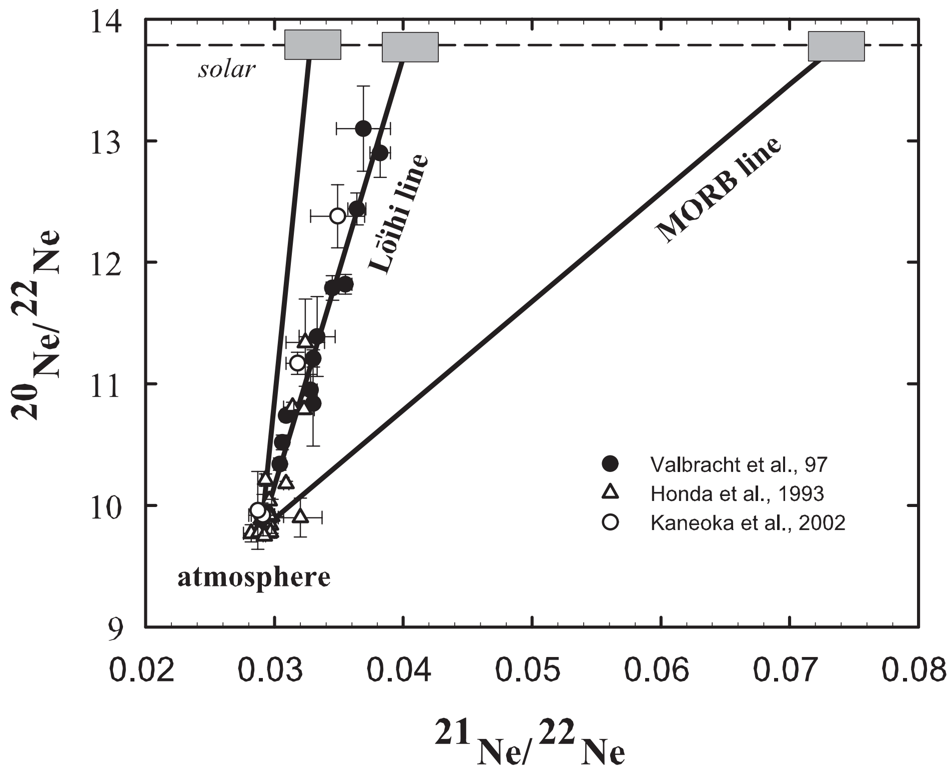

Figure 21. Garcia et al.