Extrinsic Con the Evolution of Hawaiian Ocean Island Volcanoes

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Extrinsic controls on the evolution of Hawaiian ocean island volcanoes

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[1] Extrinsic parameters that affect the evolution of magmatic systems within and beneath ocean island volcanoes include physical variables such as confining pressure, which controls magma degassing, and temperature of the underlying lithosphere and crust, which controls magma crystallization during ascent. Other extrinsic parameters are environmental variables coupled to the hydrosphere and atmosphere such as hydrothermal circulation systems and even rainfall. All these extrinsic factors interact with intrinsic parameters, such as magma supply rates or composition, to modulate the evolution of magma chambers and the petrologic processes that take place within them.

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1. Introduction

[2] It is now widely documented that volcanic eruptions can have short- and long-term effects on the atmosphere, hydrosphere, and biosphere. It is far less widely recognized that the environment around a volcano affects the magmatic and volcanologic characteristics and behavior of the volcanic system. Magmatic systems reflect variations in both intrinsic and extrinsic parameters. Intrinsic parameters include the depth and degree of melting that produce the magmas, the composition and mineralogy of the source rocks, the amount of magma delivered to the surface per unit time, and the transport mechanisms that enable the magmas to rise to the surface, among others. Extrinsic parameters that affect the evolution of volcanic systems and the processes that modify magma compositions include variables as innocuous as rainfall. These parameters and how they interact to modify magma compositions and volcano evolution are examined using Hawaiian volcanoes as a natural laboratory.

2. Hawaiian Eruptive Stages

[3] Hawaiian volcanoes grow in a series of well-documented stages [e.g., *Stearns*, 1940; *Clague and Dalrymple*, 1987]: an alkalic preshield stage,

a main tholeiitic shield stage, an alkalic postshield stage, and a strongly alkalic rejuvenated stage. Lavas that are geochemically similar to rejuvenated stage strongly alkalic lavas also erupt in front of (the South Arch lavas [*Lipman et al.*, 1989]) and to the sides of the island chain (the North Arch lavas [*Clague et al.*, 1990; *Dixon et al.*, 1997; *Frey et al.*, 2000]). These precursory and peripheral magmas migrate through lithosphere unmodified by prior transport of magmas that form the islands and also avoid migrating through the volcanic edifices themselves. The North Arch lavas erupted between 0.5 and 1.6 Ma [*Dixon et al.*, 1997]. The South Arch lavas erupted ∼20 ka (D. A. Clague et al., The distribution of alkalic lava around Hawaii and the geometry of the Hawaiian plume, submitted to *Journal of Petrology*, 2000, hereinafter referred to as Clague et al., submitted manuscript, 2000), concurrently with shield stage volcanism on Kilauea and Mauna Loa Volcanoes and postshield stage volcanism on Hualalai and Mauna Kea Volcanoes.

[4] The shield stage can be further subdivided into three distinct phases: an effusive submarine phase, an explosive phase characterized by both effusive and explosive eruptions, and an effusive subaerial phase. Rift zones and calderas develop during the preshield stage (both have already formed on Loihi Seamount) and play a fundamental role in the growth of the volcano. Some characteristics of the sequential eruptive stages are summarized in Table 1. This framework of petrologic evolution during the growth of Hawaiian volcanoes has been used to construct complex and elegant models of mantle sources and melt generation and transport [e.g., *Ribe and Christensen*, 1994; *Hauri et al.*, 1994; *DePaolo and Stolper*, 1996]. The intrinsic variables that generate Hawaiian magmas have been the subject of decades of study [e.g., *Clague and Frey*, 1982; *Budahn and Schmitt*, 1985; *Clague*, 1987a; *Frey and Rhodes*, 1993; *Yang et al.*, 1996] and will not be further discussed here.

[5] Changes in the development and evolution of the magma storage system are superposed on this well-established eruptive sequence [*Clague*, 1987b, 1996a]. These changes are characterized by the initial creation and eventual solidification of magma chambers or storage zones within and beneath the volcanoes.

3. Formation of Magma Chambers

3.1. Magma Storage Near the Boundary Between the Crust and the Mantle

[6] The formation of magma chambers mainly reflects increases in the magma supply rate over time and is therefore broadly related to magma composition and eruptive stage (see Table 1). *Clague* [1987b] proposed that magma starts to pond at moderate depth near the boundary between the crust and the mantle beneath the growing volcano during the preshield stage. We refer to this magma storage zone near the boundary between the crust and the mantle as the deep magma chamber. It probably begins as an ephemeral plexus of intertwined intrusions and sills. The development of the deep chamber during the preshield stage indicates that it begins to form when magma supply rates are still low. The evidence that supports the location and timing of this important transition in the magmatic system is the fractionated character of lavas from Loihi Seamount and the presence of only crustal cumulate and uppermost mantle lherzolite xenoliths in Loihi alkalic basalts [*Clague*, 1988]. All xenoliths derived from deeper in the mantle have been retained in the deep depth magma chamber at, or just below, the boundary between the crust and the mantle as the magmas pass through it. *Garcia et al.* [1998b], on the basis of the early appearance of augite as a fractionating phase in certain Loihi tholeiitic basalts, also argued for magma storage and crystal fractionation in a deep chamber located within the ocean crust (∼8–9 km below the summit of Loihi). Such deep chambers, first depicted by *Wright* [1971] for Kilauea Volcano (as a "staging area"), probably are only inefficiently cooled by hydrothermal circulation and thus lose heat slowly. These deep chambers persist until late in the evolution of the volcano. Once magmas pond in the deep magma chamber the influence of confining pressure, an extrinsic physical variable, becomes important in that CO_2-rich bubbles in the stored magmas will separate from the melt phase. Hawaiian magmas are saturated with a CO_2-rich vapor at depths greater than the deep magma chamber (between 10-km depth beneath Loihi and 25-km depth beneath Mauna Kea) and therefore lose CO_2 and rare gases during storage [*Dixon*, 1997; *Dixon et al.*, 1997]. Unfortunately, there is no present-day Hawaiian volcano at this developmental stage.

3.2. Formation of Shallow Magma Chambers

[7] Shallow subcaldera magma chambers probably form within the submarine volcano somewhat later during the preshield stage, as increasing magma supply increases temperature of the lithosphere and crust. Similar to the deep magma chambers, these may at first be ephemeral and best described as a plexus of intertwined intrusions and sills rather than

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Table 1. Characteristics of Hawaiian Eruptive Stages

Stage	Location	Lava Types	Eruption Frequency	Eruption Style	Eruption Volumes	Duration	Xenolith Types	Magma Chambers
Precursory (submarine)	South Arch	basanite, alkalic basalt	10^4–10^5 years	effusive	10^{7-8} m^3	unknown	none known	none
Preshield (submarine)	Loihi, nearly done	basanite, alkalic basalt, transitional basalt	10–10^3 years?	effusive, Hawaiian, some phreatomagmatic	10^{4-6} m^3?	0.2 Ma	dunite, harzburgite, olivine-gabbro	deep chamber forms
Shield-submarine phase	Loihi, just beginning	tholeiitic basalt	1–10 years?	effusive	10^{4-8} m^3?	0.1 Ma	none known	deep chamber continues, shallow chamber forms
Shield-explosive phase (shallow submarine to low elevation subaerial)	Kilauea	tholeiitic basalt	1–10 years	effusive, Strombolian, Hawaiian, interspersed phreatomagmatic	10^{4-9} m^3	0.3 Ma	olivine-gabbro, rare dunite, dike, and sill fragments	deep and shallow chambers continue
Shield-subaerial effusive phase	Mauna Loa	tholeiitic basalt	1–10 years	Hawaiian to effusive	10^{8-9} m^3	0.3 Ma	gabbro, rare dunite	deep and shallow chambers continue
Postshield (subaerial)	Hualalai, Mauna Kea	alkalic basalt, hawaiite, trachyte	10^2–10^3 years	Hawaiian to effusive	10^{6-8} m^3	0–0.2 Ma	dunite, gabbro, rare pyroxenite, dike, and sill fragments	shallow chamber solidifies early, deep chamber solidifies late
Peripheral (submarine)	North Arch	nephelinite, basanite, alkalic basalt	10^5 years	effusive and Strombolian	10^{8-10} m^3	1.6 Ma	lherzolite? (Fo_{91} olivine xenocrysts only)	none
Rejuvenated (subaerial, some submarine)	e.g., Honolulu volcanics; flows between Oahu and Kauai	melilitite, nephelinite, basanite, alkalic basalt	10^4–10^5 years	Strombolian to effusive, some phreatomagmatic	10^{6-9} m^3	0–3.2 Ma	dunite, wehrlite, lherzolite (±garnet), pyroxenite (±garnet)	none

Rock types, xenoliths, and magma chambers from *Clague* [1987b]. See text for references for precursory and peripheral eruptive stages. Eruption frequency is the average time between eruptions. Eruption volumes are for individual eruptions and are estimated for all submarine stages. Duration of stages estimated from *Moore and Clague* [1992] modified to include longer durations estimated from the Hilo Scientific Drilling Project [*DePaolo and Stolper*, 1996].

a single, homogeneous chamber, but eventually the magma supply is great enough to maintain long-lived shallow magma chambers (here referred to as shallow magma chambers). For example, the summit of Loihi Seamount has three pit craters [*Fornari et al.*, 1988], the most recent formed in 1996 [*Loihi Science Team*, 1997; *Davis and Clague*, 1998]. The three pit craters are nested inside overlapping shallow calderas that form the summit platform of Loihi. These pit craters and calderas formed during collapse of the surface into such shallow magma chambers, although it is unknown if these chambers are presently long-lived or remain ephemeral. In any event, the structure of Loihi Seamount establishes that the shallow magma chambers form during the preshield stage.

[8] Typically, the tops of shallow Hawaiian magma chambers during the tholeiitic shield stage are ~3 km below the volcano summit [e.g., *Decker*, 1987]. The upward migration of the shallow magma chamber continues throughout the growth of the volcano because it appears to remain several kilometers below the surface regardless of the height of the volcano [e.g., *Decker*, 1987]. *Ryan* [1988] argued that the shallow magma chamber resides at this level because of neutral buoyancy between the wall rocks and the magma in the chamber. However, *Hooft and Detrick* [1993] showed that the top of the axial magma lens on mid-ocean ridges is much deeper than the neutral buoyancy horizon. *Gudmundsson* [1990] proposed that the level where magma stopped rising at divergent plate boundaries is a stress barrier where horizontal compressive stresses are greater than vertical stresses causing sill formation. *Rubin* [1993] suggested that on the basis of analysis of Icelandic and Hawaiian rift zones the level was the brittle-ductile boundary caused by tectonically driven stress differences. In yet another view, *Phipps Morgan and Chen* [1993] suggested that magma ascent stops at a freezing horizon owing to viscous stresses. These models are all similar in that the level at which magma ponds is controlled by thermomechanical properties of the crust. We think that the tops of Hawaiian magma chambers, similar to the tops of mid-ocean ridge melt lenses discussed above, are controlled by thermomechanical properties of the crust rather than neutral buoyancy.

4. Extrinsic Effects of the Hydrosphere and Atmosphere

[9] It is during the submarine preshield stage, after a shallow magma chamber has formed, that the hydrosphere and atmosphere begin to have significant effects on the processes taking place inside the volcano. This is so because small, shallow magma chambers in submarine volcanoes are efficiently cooled by circulating hydrothermal fluids, as suggested by the abundance of hot springs on Loihi Seamount. To maintain the shallow magma chamber requires that magmatic heat input must offset hydrothermal cooling. Such thermal considerations are consistent with formation of the shallow magma chamber somewhat later than the deep magma chamber as magma supply rate increases.

[10] Three important geochemical and petrological changes take place when the shallow magma chamber forms: (1) increased cooling rates of stored magma as hydrothermal circulation convects heat away from the chamber, (2) assimilation of seawater-derived components into the magma, and (3) degassing under hydrostatic pressures sufficient to allow exsolution of most of the initial CO_2 but little of the initial H_2O, S, and Cl.

4.1. Increased Cooling Rates

[11] Higher cooling rates result in increased fractionation of magma, although this effect may be counterbalanced to some degree by increased magma supply rate that results in either shorter residence times in the shallow magma chamber or increased volume of the chamber. At Loihi Seamount most melts are highly fractionated with glass MgO contents between 5.7 and 8.6% for tholeiitic basalt (71 glass rind and 397 sand analyses), 3.6 and 8.9% for transitional basalt (24 glass rind and 87 sand analyses), 4.0 and 9.7% for alkalic basalt (48 glass rind and 180 sand analyses), and 4.3 and 6.2% for basanite (7 glass rind analyses) (data from *Moore et al.* [1982], *Garcia et al.* [1989, 1993, 1995, 1998b], *Honda et al.* [1993], *Clague et al.* [2000], and *Dixon and Clague* [2000]). The vast majority of glasses for each lithology contain less than 7.5% MgO.

4.2. Assimilation of Seawater-Derived Components

[12] Assimilation of seawater-derived components into the magma is an important process within shallow magma chambers on submarine volcanoes. During the early stages of shallow magma chamber formation, the maturing magma chamber has not yet dried out the surrounding crust [*Connor et al.*, 1997], and individual intrusions assimilate various amounts of hydrothermally altered roof rocks as a function of their unique pathway through the

volcanic structure [*Kent et al.*, 1999a]. The assimilation of such altered rocks and either contained brine [*Kent et al.*, 1999a, 1999b] or salts contaminates the stored magmas with seawater-derived components, particularly Cl and heavy rare gases [*Patterson et al.*, 1990; *Honda et al.*, 1993]. Except in isolated cases [*Kent et al.*, 1999a], assimilation does not affect the water contents of the melts [*Dixon and Clague*, 2000].

4.3. Submarine Degassing

[13] In addition to assimilation, magmas within the magma chamber are modified by degassing as CO_2 gas exsolves and bubbles separate from the magma [*Gerlach*, 1986; *Dixon and Stolper*, 1995; *Stevenson and Blake*, 1998]. The discharge of such CO_2-rich fluids is well documented at Loihi Seamount [e.g., *Sakai et al.*, 1987; *Karl et al.*, 1988; *Sedwick et al.*, 1994] and intensified following the recent collapse at the summit [*Loihi Science Team*, 1997]. The CO_2 gas bubbles transport other volatile components, particularly noble gases, out of the melt and into the hydrothermal fluids prior to eruption, but hydrostatic pressure (>100 bar) is sufficient to keep nearly all H_2O, S, and Cl dissolved in melts in the chamber and during eruption in all but the most differentiated or alkalic samples [*Dixon and Clague*, 2000].

5. Breaching Sea Level

[14] The next transition takes place when the summit of the volcano approaches sea level and phreatomagmatic eruptions occur owing to decreased confining pressure and the expansion of seawater, and eventually groundwater, to steam. Large-scale phreatomagmatic eruptions require introducing water into the crustal magma chamber [e.g., *Mastin*, 1995]. This explosive eruptive phase may already have begun at Loihi where thin layers of glass sand, including glass bubble-wall fragments, are common on the summit and rift zones [*Clague et al.*, 2000]. Evidence for more vigorous explosive activity includes bedded ash deposits up to 11 m thick occur on the summit platform, still 1 km below sea level (R. Batiza, D. Clague, and J. Head, unpublished data, 1999). These explosive eruptions, a minor part of Loihi's recent eruptive activity, should intensify as the summit approaches and then breaches sea level. The explosive phase continues until the summit is at least 1.2 km above sea level, as demonstrated by Kilauea, where explosive eruptions have occurred twice in historic times (1924 and 1790) and at least 17 times in the last 50,000 years [*Easton*, 1987; *Clague et al.*, 1995a], although the steam in these eruptions was derived from fresh water confined by faults that bound Kilauea's caldera. The largest of these hydromagmatic eruptions produced ash up to 27 m thick exposed by the Hilina fault southwest of Kilauea's summit [*Clague et al.*, 1995a]. The subaerial explosive eruptions, similar to their submarine counterparts on Loihi, constitute a minor, though hazardous, part of Kilauea's eruptive activity.

[15] An interesting possibility is that the energy released during these explosive eruptions may also help push the unbuttressed south flank of Kilauea laterally, triggering catastrophic landslides, as have been documented surrounding the Hawaiian islands [*Moore et al.*, 1989]. The push on the flank caused by the combination of dike intrusions in the summit and rift zones [*Swanson et al.*, 1976; *Lipman et al.*, 1985] and flow of cumulate dunite within the volcano [*Clague and Denlinger*, 1994] probably cause large earthquakes, such as the 1975 Kalapana M7.2 earthquake. These large earthquakes, including one estimated at M7.9 in 1868, are associated with slip of the flank but did not result in catastrophic failure. The added lateral force applied to the flank during explosive eruptions may provide the trigger for the catastrophic failures of the flanks of Hawaiian and other ocean island volcanoes. To test this idea requires determination of the timing of explosive eruptions and landslides, probably best accomplished using cores collected away from the islands.

[16] High cooling rates and assimilation that affected magma compositions during the submarine preshield stage continue to affect their compositions after the summit breaches sea level. At Kilauea Volcano, most melts are highly fractionated with glass MgO contents between 5 and 7% for tholeiitic pillow lavas erupted on the Puna Ridge [*Clague et al.*, 1995b], and nearly all subaerial historic eruptions have melt MgO <7.5% [e.g., *Wright*, 1971]. Oxygen isotopes provide evidence that Kilauea magmas assimilate altered crustal rocks in the rift zone [*Garcia et al.*, 1998a], although it is unclear whether these rocks are altered in seawater or fresh groundwater.

5.1. Subaerial Degassing

[17] During the explosive part of the shield stage the summit of the volcano emerges above sea level and low-pressure degassing begins. The loss of magmatic volatiles before eruption is an important

physical process that drives mixing in the magma chamber and therefore affects shallow-level petrogenetic processes [*Dixon et al.*, 1991; *Clague et al.*, 1995b; *Wallace and Anderson*, 1998]. At some poorly defined time, but approximately when the summit breaches sea level, magma stored in the chamber is no longer held under high hydrostatic or lithostatic pressure and can degas significant amounts of water, sulfur, and additional CO_2 [*Dixon et al.*, 1991]. Until such degassing begins, magma stored in the chamber is stratified with the newest, hottest melt at the bottom and the most fractionated melt at the top [*Clague et al.*, 1995b]. The degassing process, however, increases the density of the melt at the top of the chamber, which causes overturn in the magma chamber and initiates large-scale magma mixing. These overturn events may occur episodically (one probably took place at Kilauea Volcano on February 1, 1996 [*Thornber et al.*, 1996]), and introduce fractionated magmas deeper in the chamber where they mix with more primitive melts. *Dixon et al.* [1991] first proposed this process on the basis of the observation that lava erupted along Kilauea's submarine Puna Ridge contained CO_2 contents expected for their eruption depth but had lower than expected H_2O and SO_2 contents. They argued that such characteristics are a consequence of mixing subaerially degassed magma with magma containing its full complement of water and sulfur and enough CO_2 that the hybrid melt contained more CO_2 than the saturation value for its eruption depth. The key feature here is that until the stored magma can degas at, or close to, atmospheric pressure, magma hybridization in Hawaiian volcanoes takes place only rarely and on a small scale.

5.2. Transitions in Hydrothermal Circulation

[18] Hydrothermal systems also evolve as the volcano grows [*Fournier*, 1987]. As the volcano summit grows above sea level, eventually the top of the shallow magma chamber must rise above the boundary between seawater and fresh water, and the circulating hydrothermal fluid changes from being seawater-derived to being fresh water–derived. Kilauea appears to already be at or near this stage, despite the top of the magma chamber being at least 1 km below sea level, since its lavas show little evidence of Cl assimilation [*Clague et al.*, 1995b].

[19] With additional upward growth the summit of the volcano can reach above the main rainfall zone on the island. On Hawaii this probably takes place when the summit reaches ~2-km elevation. With little groundwater to replenish the hydrothermal system it eventually boils away. Mauna Loa Volcano has reached this stage, and no longer has a hydrothermal system near the summit [*Hawaii Institute of Geophysics*, 1983]. The explosive phase of activity ends during the decline of the hydrothermal system. There are also petrologic consequences associated with this transition since it becomes more difficult to cool the stored magma now that heat must be conducted, rather than convected, away. This transition does not necessarily coincide with a decrease in magma supply rates, so slower cooling means less crystallization and less fractionated lava compositions. It also prolongs the life of the shallow magma chamber since less input heat is required to maintain a magma-filled chamber. Only 11 of the 129 volcanoes that make up the Hawaiian-Emperor chain attained heights of ~2 km [*Clague*, 1996b], so most retained their hydrothermal systems until the shallow magma chamber solidified.

6. Solidification of Magma Chambers

[20] Shallow magma chambers, once formed, persist throughout the shield stage, which is characterized by repetitive caldera collapse events, and then solidify near the end of the shield stage or the beginning of the postshield stage as magma supply rates decrease. There is simply not enough magma delivered to balance conductive heat losses, and the chamber crystallizes. *Clague* [1987b] proposed that the shallow chamber on Hualalai Volcano solidified early in the postshield stage. On Hualalai, solidification of the shallow chamber, indicated by the eruption of trachyte formed by extensive crystal fractionation of alkalic basalt, occurs no more than 20 ka after the end of the tholeiitic shield stage. A deep chamber must still persist today beneath Hualalai, 125 ka after the transition from shield to postshield stages, since all postshield stage alkalic basalts, including those erupted in 1800–1801 A.D., have evolved compositions [*Moore et al.*, 1987]. The xenoliths on Hualalai also provide evidence that the deep magma chamber persisted throughout the tholeiitic shield stage. Cumulate websterite xenoliths of tholeiitic shield magmas that equilibrated at depths of ~20 km were erupted in the 1800–1801 postshield alkalic basalt [*Bohrson and Clague*, 1988].

[21] The petrologic consequences of solidification of the shallow magma chamber are dramatic since magmas delivered from the deep chamber now rise

to the surface and erupt without any shallow repose. Magmas therefore no longer have an opportunity to degas at or near atmospheric pressure, and volatile components, with the exception of some of the least soluble volatiles, such as CO_2 and He, are retained in the magmas. These magmas apparently rise quickly from their storage chamber at the boundary between the crust and the mantle and carry abundant cumulate gabbro xenoliths derived from the ocean crust [*Clague*, 1987b] and from the vertical stack of solidified magma chambers within the volcano [*Bohrson and Clague*, 1988]. They fractionate in the deep magma chamber, rather than in the shallow one, and therefore tend to have high gas contents that drive high fountains and production of tephra and scoria that forms large cones concentrated near the summit of the volcano.

[22] *Frey et al.* [1990] proposed a somewhat different model in which the shallow magma chamber on Mauna Kea Volcano persists until late in the postshield stage when it crystallizes and the deep chamber forms. They showed that late-postshield Laupahoehoe lavas fractionated at ~20 km depth with augite an important crystallizing phase, but that early-postshield Hamakua lavas fractionated at shallow depth, with plagioclase an important crystallizing phase. An alternative interpretation is that the deep chamber existed from early in the growth of the volcano, as proposed above, but that during earlier periods of high magma supply, magma in the deep chamber does not cool sufficiently such that augite joins olivine on the liquidus. It is only when the magma supply decreases sharply during the later part of the postshield stage that the alkalic magmas reside in the deep chamber long enough that augite crystallizes, thereby imparting the chemical characteristics that indicate crystallization at elevated pressure. The magma supply rate declines further as the alkalic postshield stage draws to an end, and the heat input to even the deep magma chamber cannot sustain it, and it too solidifies. This change probably marks the end of the postshield stage.

7. The Earliest and Latest Volcanism in Hawaii

[23] Rejuvenated stage strongly alkalic basalt erupted on the islands is not stored in any shallow or deep magma chamber [e.g., *Clague and Frey*, 1982; *Clague and Dalrymple*, 1988; *Wright and Clague*, 1989; *Maaloe et al.*, 1992]. The submarine precursory South Arch [*Lipman et al.*, 1989; Clague et al., submitted manuscript, 2000] and the peripheral North Arch [*Clague et al.*, 1990; *Dixon et al.*, 1997; *Frey et al.*, 2000] lavas are petrologically similar but not identical to rejuvenated stage lavas. Despite their rapid delivery from the mantle to their eruptive vents, one extrinsic physical variable, the temperature structure of the mantle and crust, leads to distinctions in the compositions of these magmas. The precursory and peripheral magmas travel through relatively cool mantle and crust. Significant amounts of olivine crystallize during their ascent, resulting in eruption of moderately fractionated compositions (whole rock MgO as low as 6.4% [*Lipman et al.*, 1989; *Clague et al.*, 1990]). In contrast, rejuvenated stage lavas erupted on the islands travel through mantle preheated by the prior migration of magmas that built the island. Many of these therefore have near-primary compositions with most having MgO contents between 12 and 16% [*Clague and Frey*, 1982; *Clague and Dalrymple*, 1988; *Maaloe et al.*, 1992]. Initial concentrations of volatiles in these lavas are high because of low degrees of melting of source rocks and degassing of CO_2 and H_2O occurs as a single-stage, closed-system process during ascent [*Dixon et al.*, 1997]. The exsolution of these volatiles may drive rapid ascent of these magmas, but the gas phase does not separate from the melt until eruption.

8. Conclusions

[24] Extrinsic physical and environmental factors affect the formation of shallow magma chambers and the compositions of lava on ocean island volcanoes such as Hawaii. The first transition in the life of a growing volcano takes place when a shallow subcaldera magma chamber first forms during the preshield stage. Three other transitions occur during the shield stage: when the summit approaches and then breaches sea level, when the top of the shallow magma chamber rises above the boundary between seawater and fresh water, and when the top of the shallow magma chamber rises above the zone of heavy rainfall. The final transition takes place during the postshield stage when the shallow magma chamber solidifies. Many of these transitions are related to changes, not in the intrinsic parameters that generate melt and transport it to the surface, but in extrinsic physical and environmental factors that affect the evolution of the shallow magmatic system in profound ways. Changes in each of these parameters affect processes that occur in the shallow magma chamber

and the chemical characteristics of the lavas. These concepts may also improve our understanding of petrogenetic processes on other ocean island volcanoes after consideration of critical intrinsic parameters, particularly magma supply rate, and extrinsic parameters, most notably rainfall.

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References

Bohrson, W., and D. A. Clague (1988), Origin of ultramafic xenoliths containing exsolved pyroxenes from Hualalai Volcano, Hawaii, *Contrib. Mineral. Petrol.*, *100*, 139–155.

Budahn, J. R., and R. A. Schmitt (1985), Petrogenetic modelling of Hawaiian tholeiitic basalts: A geochemical approach, *Geochim. Cosmochim. Acta*, *49*, 67–87.

Clague, D. A. (1987a), Hawaiian alkaline volcanism, in *Alkaline Igneous Rocks*, edited by J. G. Fitton and B. G. J. Upton, *Geol. Soc. Spec. Publ.*, *30*, 227–252.

Clague, D. A. (1987b), Hawaiian xenolith populations, magma supply rates and development of magma chambers, *Bull. Volcanol.*, *49*, 577–587.

Clague, D. A. (1988), Petrology of ultramafic xenoliths from Loihi Seamount, Hawaii, *J. Petrol.*, *29*, 1161–1186.

Clague, D. A. (1996a), Crustal processes that modify hot-spot magma compositions, Kilauea Volcano, Hawaii, paper presented at Chapman Conference on Shallow Level Processes in Ocean Island Magmatism: Distinguishing Mantle and Crustal Signatures, AGU, Tenerife, Canary Islands, Nov. 10–16.

Clague, D. A. (1996b), The growth and subsidence of the Hawaiian-Emperor volcanic chain, in *The Origin and Evolution of Pacific Island Biotas, New Guinea to Eastern Polynesia: Patterns and Processes*, edited by A. Keast and S. E. Miller, pp. 35–50, SPB Acad., Amsterdam, Netherlands.

Clague, D. A., and G. B. Dalrymple (1987), The Hawaiian-Emperor volcanic chain, Part 1, Geologic evolution, *U.S. Geol. Surv. Prof. Pap.*, *1350*, 5–54.

Clague, D. A., and G. B. Dalrymple (1988), Age and petrology of alkalic postshield and rejuvenated stage lava from Kauai, Hawaii, *Contrib. Mineral. Petrol.*, *99*, 202–218.

Clague, D. A., and R. P. Denlinger (1994), Role of olivine cumulates in destabilizing the flanks of Hawaiian volcanoes, *Bull. Volcanol.*, *56*, 425–434.

Clague, D. A., and F. A. Frey (1982), Petrology and trace-element geochemistry of the Honolulu Volcanics, Oahu: Implications for the oceanic mantle beneath Hawaii, *J. Petrol.*, *23*, 447–504.

Clague, D. A., R. T. Holcomb, J. M. Sinton, R. S. Detrick, and M. E. Torresan (1990), Pliocene and Pleistocene alkalic flood basalts on the seafloor north of the Hawaiian Islands, *Earth Planet. Sci. Lett.*, *98*, 175–191.

Clague, D. A., M. H. Beeson, R. P. Denlinger, and L. G. Mastin (1995a), Ancient ash deposits and calderas at Kilauea Volcano, *Eos Trans. AGU*, *76*(46), Fall Meet. Suppl., F666.

Clague, D. A., J. G. Moore, J. E. Dixon, and W. B. Friesen (1995b), Petrology of submarine lavas from Kilauea's Puna Ridge, Hawaii, *J. Petrol.*, *36*, 299–349.

Clague, D. A., A. S. Davis, J. L. Bischoff, J. E. Dixon, and R. Geyer (2000), Lava bubble-wall fragments formed by submarine hydrovolcanic explosions on Loihi Seamount and Kilauea Volcano, *Bull. Volcanol.*, *61*, 437–449.

Connor, C. B., P. C. Lichtner, F. M. Conway, B. E. Hill, A. A. Ovsyannikov, I. Federchenko, Y. Doubik, V. N. Shapar, and Y. A. Taran (1997), Cooling of an igneous dike 20 years after intrusion, *Geology*, *25*, 711–714.

Davis, A. S., and D. A. Clague (1998), Changes in the hydrothermal system at Loihi Seamount after the formation of Pele's pit in 1996, *Geology*, *26*, 399–402.

Decker, R. W. (1987), The dynamics of Hawaiian volcanoes: An overview, *U.S. Geol. Surv. Prof. Pap.*, *1350*, 997–1018.

DePaolo, D. J., and E. M. Stolper (1996), Models of Hawaiian volcano growth and plume structure: Implications of results from the Hawaii Scientific Drilling Project, *J. Geophys. Res.*, *101*, 11,643–11,654.

Dixon, J. E. (1997), Degassing of alkalic basalts, *Am. Mineral.*, *82*, 368–378.

Dixon, J. E., and D. A. Clague (2000), Volatiles in basaltic glasses from Loihi Seamount, Hawaii: Evidence for a relatively dry plume component, *J. Petrol.*, in press.

Dixon, J. E., and E. M. Stolper (1995), An experimental study of water and carbon dioxide solubilities in mid-ocean ridge basaltic liquids, Part II, Applications to degassing, *J. Petrol.*, *36*, 1633–1646.

Dixon, J. E., D. A. Clague, and E. M. Stolper (1991), Degassing history of water, sulfur, and carbon in submarine lavas from Kilauea Volcano, Hawaii, *J. Geol.*, *99*, 371–394.

Dixon, J. E., D. A. Clague, P. Wallace, and R. Poreda (1997), Volatiles in alkalic basalts from the North Arch Volcanic Field, Hawaii: Extensive degassing of deep submarine-erupted alkalic series lavas, *J. Petrol.*, *38*, 911–939.

Easton, R. M. (1987), Stratigraphy of Kilauea Volcano, *U.S. Geol. Surv. Prof. Pap.*, *1350*, 243–260.

Fornari, D. J., M. O. Garcia, R. C. Tyce, and D. G. Gallo (1988), Morphology and structure of Loihi seamount based on seabeam sonar mapping, *J. Geophys. Res.*, *93*, 15,227–15,238.

Fournier, R. O. (1987), Conceptual models of brine evolution in magmatic-hydrothermal systems, *U.S. Geol. Surv. Prof. Pap.*, *1350*, 1487–1506.

Frey, F. A., and J. M. Rhodes (1993), Intershield geochemical differences among Hawaiian volcanoes: Implications for source compositions, melting process, and magma ascent paths, *Philos. Trans. R. Soc. London, Ser. A*, *342*, 121–136.

Frey, F. A., D. A. Clague, J. Mahoney, and J. Sinton (2000), Volcanism at the edge of the Hawaiian plume: Petrogenesis of submarine alkalic lavas from the North Arch Volcanic Field, *J. Petrol.*, in press.

Frey, F. A., M. O. Garcia, W. S. Wise, A. Kennedy, P. Gurriet, and F. Albarede (1990), Evolution of Mauna Kea volcano: Petrologic and geochemical constraints on postshield volcanism, *J. Geophys. Res.*, *95*, 1271–1300.

Garcia, M. O., D. Muenow, K. Aggrey, and J. O'Neil (1989), Major element, volatile and stable isotope geochemistry of Hawaiian submarine tholeiitic glasses, *J. Geophys. Res.*, *94*, 525–538.

Garcia, M. O., B. Jorgenson, J. Mahoney, E. Ito, and A. Irving (1993), An evaluation of the temporal geochemical evolution of Loihi summit lavas: Results from Alvin submersible dives, *J. Geophys. Res.*, *98*, 537–550.

Garcia, M. O., D. J. P. Foss, W. B. West, and J. J. Mahoney (1995), Geochemical and isotopic evolution of Loihi Volcano, Hawaii, *J. Petrol.*, *26*, 1647–1674.

Garcia, M. O., E. Ito, J. M. Eiler, and A. J. Pietruszka (1998a), Crustal contamination of Kilauea Volcano magma revealed by oxygen isotope analyses of glass and olivine from Puu Oo eruption lavas, *J. Petrol.*, *39*, 803–817.

Garcia, M. O., K. H. Rubin, M. D. Norman, J. M. Rhodes, D. W. Graham, D. W. Muenow, and K. Spencer (1998b), Petrology and geochemistry of basalt breccia from the 1996 earthquake swarm of Loihi Seamount, Hawaii: Magmatic history of its 1996 eruption, *Bull. Volcanol.*, *59*, 577–592.

Gerlach, T. M. (1986), Exsolution of H_2O, CO_2, and S during eruptive episodes at Kilauea Volcano, Hawaii, *J. Geophys. Res.*, *91*, 12,177–12,185.

Gudmundsson, A. (1990), Emplacement of dikes, sills, and crustal magma chambers at divergent plate boundaries, *Tectonophysics*, *176*, 257–275.

Hauri, E. H, J. A. Whitehead, and S. R. Hart (1994), Fluid dynamics and geochemical aspects of entrainment in mantle plumes, *J. Geophys. Res.*, *99*, 24,275–24,300.

Hawaii Institute of Geophysics (1983), *Geothermal Resources of Hawaii*, 1 map sheet, U.S. Gov. Print. Off., Washington, D. C.

Honda, M., I. McDougall, D. B. Patterson, A. Doulgeris, and D. A. Clague (1993), Noble gases in submarine pillow basalt glasses from Loihi and Kilauea, Hawaii: A solar component in the Earth, *Geochim. Cosmochim. Acta*, *57*, 859–874.

Hooft, E. E., and R. S. Detrick (1993), The role of density in the accumulation of basaltic melts at mid-ocean ridges, *Geophys. Res. Lett.*, *20*, 423–426.

Karl, D. M., G. M. McMurtry, A. Malahoff, and M. O. Garcia (1988), Loihi Seamount: A midplate volcano with a distinctive hydrothermal system, *Nature*, *335*, 532–535.

Kent, A. J. R., M. D. Norman, I. D. Hutcheon, and E. M. Stolper (1999a), Sea water assimilation in an oceanic volcano: Evidence from matrix glass and glass inclusions from Loihi Seamount, Hawaii, *Chem. Geol.*, *156*, 299–319.

Kent, A. J. R., D. A. Clague, M. Honda, E. M. Stolper, I. Hutcheon, and M. D. Norman (1999b), Widespread assimilation of a seawater-derived component at Loihi Seamount, Hawaii, *Geochim. Cosmochim. Acta*, *63*, 2749–2762.

Lipman, P. W., D. A. Clague, R. T. Holcomb, and J. G. Moore (1989), Newly identified young lava flows on the sea floor south of the Hawaiian Ridge, *Geology*, *7*, 611–614.

Lipman, P. W., J. P. Lockwood, R. T. Okamura, D. A. Swanson, and K. M. Yamashita (1985), Ground deformation associated with the 1975 magnitude-7.2 earthquake and resulting changes in activity of Kilauea Volcano, Hawaii, *U.S. Geol. Surv. Prof. Pap.*, *1276*, 45 pp.

Loihi Science Team (1997), Researchers rapidly respond to submarine activity at Loihi Volcano, Hawaii, *Eos Trans. AGU*, *78*, 229–233.

Maaloe, S., D. James, P. Smedley, S. Petersen, and L. B. Garmann (1992), The Koloa Volcanics suite of Kauai, Hawaii, *J. Petrol.*, *33*, 761–784.

Mastin, L. (1995), Thermodynamics of gas and steam-blast eruptions, *Bull. Volcanol.*, *57*, 85–98.

Moore, J. G., and D. A. Clague (1992), Volcano growth and evolution of the island of Hawaii, *Geol. Soc. Am. Bull.*, *104*, 1471–1484.

Moore, J. G., D. A. Clague, and W. R. Normark (1982), Diverse basalt types from Loihi Seamount, Hawaii, *Geology*, *10*, 88–92.

Moore, R. B., D. A. Clague, M. Rubin, and W. A. Bohrson (1987), Hualalai Volcano: A preliminary summary of geologic, petrologic, and geophysical data, *U.S. Geol. Surv. Prof. Pap.*, *1350*, 571–586.

Moore, J. G., D. A. Clague, R. T. Holcomb, P. W. Lipman, W. R. Normark, and M. E. Torresan (1989), Prodigious submarine landslides on the Hawaiian ridge, *J. Geophys. Res.*, *94*, 17,465–17,484.

Patterson, D. B., M. Honda, and I. McDougall (1990), Atmospheric contamination: A possible source of heavy noble gases in basalts from Loihi Seamount, Hawaii, *Geophys. Res. Lett.*, *17*, 705–708.

Phipps Morgan, J., and Y. J. Chen (1993), The genesis of ocean crust: Magma injection, hydrothermal circulation, and crustal flow, *J. Geophys. Res.*, *98*, 6283–6297.

Ribe, N., and U. Christensen (1994), Three-dimensional modeling of plume-lithosphere interaction, *J. Geophys. Res.*, *99*, 660–682.

Rubin, A. M. (1993), Comparison of rift-zone tectonics in Iceland and Hawaii, *Bull Volcanol.*, *52*, 302–319.

Ryan, M. P. (1988), The mechanics and three-dimensional internal structure of active magmatic systems: Kilauea Volcano, Hawaii, *J. Geophys. Res.*, *93*, 4213–4248.

Sakai, H., et al. (1987), Hydrothermal activity on the summit of Loihi Seamount, Hawaii, *Geochem. J.*, *21*, 11–21.

Sedwick, P. N., G. M. McMurtry, D. Hilton, and F. Goff (1994), Carbon dioxide and helium in hydrothermal fluids from Loihi Seamount Hawaii USA: Temporal variability and implications for the release of mantle volatiles, *Geochim. Cosmochim. Acta*, *58*, 1219–1227.

Stearns, H. T. (1940), Four-phase volcanism in Hawaii, *Geol. Soc. Am. Bull.*, *60*, 1947–1948.

Stevenson, D. S., and S. Blake (1998), Modelling the dynamics and thermodynamics of volcanic degassing, *Bull. Volcanol.*, *60*, 307–317.

Swanson, D. A., W. A. Duffield, and R. S. Fiske (1976), Displacement of the south flank of Kilauea Volcano: The result of forceful intrusion of magma into the rift zones, *U.S. Geol. Surv. Prof. Pap.*, *963*, 30 pp.

Thornber, C. T., C. C. Heliker, J. R. Reynolds, J. Kauahikaua, P. Okubo, M. Lisowski, A. J. Sutton, and D. Clague (1996), The eruptive surge of February 1, 1996: A highlight of Kilauea's ongoing East Rift Zone eruption, *Eos Trans. AGU*, *77*(46), Fall Meet. Suppl., F798.

Wallace, P. J., and A. T. Anderson Jr. (1998), Effects of eruption and lava drainback on the H_2O contents of basaltic magmas at Kilauea Volcano, *Bull. Volcanol.*, *59*, 327–344.

Wright, T. L. (1971), Chemistry of Kilauea and Mauna Loa lava in space and time, *U.S. Geol. Surv. Prof. Pap.*, *735*, 40 pp.

Wright, T. L., and D. A. Clague (1989), Petrology of Hawaiian lavas, in *The Geology of North America*, vol. N, *The Eastern Pacific Ocean and Hawaii, Decade of North American Geology*, edited by E. L. Winterer, D. M. Hussong, and R. W. Decker, pp. 218–237, Geol. Soc. of Am., Boulder, Colo.

Yang, H.-J., F. A. Frey, J. M. Rhodes, and M. O. Garcia (1996), Evolution of the Mauna Kea Volcano: Inferences from lava compositions recovered in the Hawaii Scientific Drilling Project, *J. Geophys. Res.*, *101*, 11,747–11,767.