SUBMARINE GROWTH OF A HAWAIIAN SHIELD VOLCANO
BASED ON VOLCANICLASTICS IN
THE HAWAIIAN SCIENTIFIC DRILLING PROJECT 2 CORE

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ABSTRACT

Hawaiian volcanoes are predominantly formed subaqueously. The processes of submarine growth on the flanks of these volcanoes are poorly understood due to their difficult accessibility. Hence, little is known about what lithologies comprise the bulk of these large oceanic volcanoes. The Hawaiian Scientific Drilling Project 2 (HSDP2) core provides a thick (2,019 m), nearly continuous stratigraphic section from the eastern submarine flank of Mauna Kea volcano. During deposition of these deposits, the volcano's summit was above sea level. This section is dominated by volcaniclastic rocks (55%), which are used here to assess fragmentation and depositional processes on the submarine flanks of Hawaiian volcanoes during the "subaerial" shield stage. Macroscopic structures (e.g., bedding, fabric, grading and contacts), grain size and sorting, clast lithologies, vesicularities, crystallinities, and shapes and alteration states of volcaniclastics are described. These features are used to interpret transport and depositional processes. Major element and sulfur analyses of glass clasts were made to assess source heterogeneity within volcaniclastics units and to constrain eruption depths of lavas supplying the clasts. Clast shapes are useful indicators of fragmentation processes and extent of reworking. Quantitative clast shape data are presented in order to minimize ambiguous shape descriptions and to assess the usefulness of certain shape parameters including form factor, aspect ratio, solidity and roundness.
Most of the volcaniclastic rocks in the 11 units examined consist of poorly sorted, fine lapilli to block basaltic lava clasts resting in a matrix of sideromelane ash. Silty intervals occur within some of the coarser volcaniclastics, although distinct beds of different grain sizes (fine ash to fine lapilli) and normal grading occurs only in two of the units. Clasts are usually glassy, equant with planar to curviplanar margins and have low vesicularities (<50%; average ~20%). Other clast shapes include blade-like sideromelane ash and fluidal lapilli. Alteration of the deposits includes palagonitization and smectite replacement of variable degrees from thin margins of the clasts to entire ash sized clasts. Incompatible element ratios (TiO₂/K₂O) of glass clasts are within 4σ error within each volcaniclastic unit and variations in major elements are relatively small (MgO ranges within most units are <1 wt %). All the volcaniclastics above 2,215 mbsl are degassed (S <0.04 wt %) and probably erupted subaerially. Two units at ~2,460 mbsl are less degassed (S >0.04 wt %) and probably erupted subaqueously.

Beneath the subaerial-submarine transition in the HSDP2 core is a 0.9 km thick section dominated by volcaniclastic rocks (84%; average ~11 m thick) interbedded with thin massive lava flows (average ~3 m thick). Underlying this section is a 1.1 km thick lava flow-dominated section consisting of interbedded pillow lavas (65%; average 26 m thick) and volcaniclastics (average ~10 m thick). Transport processes of the clasts away from the lava flows vary from pulsed
debris flows to low concentration, vertically stratified density currents, to in-situ deposition. The clast shapes and low vesicularities indicate they were fragmented from lava flows by passive cooling-contraction, collapse of oversteepened lavas flows, spallation, autobrecciation and erosion. These characteristics and the dominance of glassy clasts suggests that passive cooling-contraction is the principal fragmentation process and it occurred during the emplacement of submarine lava flows. Erosion is not a major process in clast formation, probably because of the high accumulation rates on the submarine flanks. All the glass clasts in the volcanioclastics are derived from single lava sources and variations in major elements are due to variable degrees of crystal fractionation equivalent to cooling temperature ranges of less than 43°C.

The types and proportions of lithologies that comprise the submarine flanks of Hawaiian volcanoes are evaluated using the products and processes identified in the HSDP2 core. Hyaloclastites and syn-eruptive pillow breccias dominate the deeper water volcanioclastics (650 to 1450 mbsl emplacement depths), whereas peperitic lapilli tuffs and secondary pillow breccias comprise the bulk of the shallower volcanioclastics (0 to 650 mbsl emplacement depths). Due to subsidence, these deposits accumulate 1.1 and 0.9 km thick sections respectively. The processes that form these deposits should not be affected by increasing water depth. Thus, deep deposits within the Mauna Kea section are probably a continuation of the deep water-type interbedded volcanioclastics and pillow lavas. However, the proportions of pillow lavas may increase and the
volcaniclastics may become thinner as the flank slopes get less steep closer to the sea floor because of the reduction of collapses of oversteepened lava flows.
1. INTRODUCTION

Ocean island volcanoes are universally associated with the eruption of generally basaltic a'a and pahoehoe lavas, which build the characteristic broad forms of shield volcanoes. However, a significant portion of many ocean island shield volcanoes are made up of subaqueously deposited volcaniclastic rocks (~50%, Staudigal and Schmincke, 1984). The emplacement of these fragmental deposits is far less well understood than the associated lavas. Proposed origins include a range of processes such as primary fragmentation via magmatic explosivity or magma-water interactions, secondary fragmentation by erosion, secondary transport (e.g. from mass wasting), and re-deposition (e.g., Schneider et al., 1998).

The Hawaiian Scientific Drilling Project drilled a nearly continuous core that penetrated nearly 2 km into the submarine flank of Mauna Kea volcano in Hawai'i (DePaolo et al., 2001). This core provides an opportunity to better understand the formation and deposition of volcaniclastics on oceanic islands. The stratigraphic section documents the emergence of the flanks of a Hawaiian volcano from a submarine to subaerial environment (Seaman et al., 2000). Submarine volcaniclastics are abundant within this core. Such rocks are normally exposed only on the submarine flanks of Hawaiian volcanoes and,
therefore, are poorly represented in existing sample collections from these oceanic islands. This study characterizes the textures and petrology of these volcaniclastics, revealing how decreasing water depth affects the occurrence, type, and products of fragmentation of lava on the flanks of oceanic volcanoes.

1.1 Geological setting

The Hawaiian archipelago is a chain of oceanic volcanoes that has grown as the Pacific plate moved northwest over a mantle plume that is believed to be located currently beneath the island of Hawai'i (Ribe and Christensen, 1999). The volcanoes grow from the ocean floor (~5,000 meters below sea level, mbsl) and emerge as islands up to 4,205 m above sea level (masl; e.g. Mauna Kea). Hawai'i is the largest island of the Hawaiian archipelago and it consists of five volcanoes ranging in age from Kohala volcano in the north, which last erupted ~60,000 yr ago (Spengler and Garcia, 1988), to Kilauea volcano in the south, which is currently erupting (Heliker et al., 2003; figure 1.1). Mauna Kea, the tallest Hawaiian volcano (~4,205 masl), is thought to have emerged above sea level about 420,000 years ago (DePaolo et al., 2001) and its most recent eruption was ~4,500 years ago (Wolfe et al., 1997). The southern flanks of Mauna Kea are overlain by lavas from the largest Hawaiian volcano, Mauna Loa, which last erupted in 1984.
Figure 1.1 Map of Hawai'i showing the locations of all five volcanoes that make up the island and the HSDP drill site.

Figure 1.2 Schematic of three stages of oceanic volcano growth that correspond to different depositional regimes. a) initial deep water stage is dominated by pillow lavas and their fragmented counterparts; b) processes of passive and explosive fragmentation become prevalent and produce more fragmental deposits during the shallow submarine to emergent stage; c) deposits of the subaerial stage are lava flows with very limited occurrences of fragmental material because the exclusion of water inhibits fragmentation (After DePaolo et al., 2001).
1.2 Models for the growth of shield volcanoes

1.2.1 Magma supply and stages

The magma for Hawaiian shield volcanoes is formed by partial melting in the upper mantle (Yoder, 1976). The buoyancy of the magma and lithostatic pressure of the surrounding rocks force it to ascend through conduits in the lithosphere and volcano edifice (e.g., Decker, 1987). A shallow reservoir is located at 3 to 7 km depth, from which the magma travels either vertically to the summit region, or laterally to flanking rift zones (Tilling and Dvorak, 1993). Eruption durations range from days to decades and are separated by repose periods because long term rates of the magma supply to the shallow reservoir are lower than the discharge rates (Dzurisin et al., 1984). Geochemistry of historical lavas from Kilauea volcano indicate that the residence time in the shallow reservoir is in the range of ~30 to ~110 years since 1790 (Pietruszka and Garcia, 1999).

The growth of typical Hawaiian volcanoes is divided into a preshield stage (duration ~250 ky; alkalic lavas), a shield stage (400-900 ky; tholeiitic lavas), a postshield stage (duration 250 to 300 ky; alkalic lavas) and an erosional stage of variable length (200 to 250 ky; e.g., Feigenson and Spera, 1981; Guillou et al., 1997; Frey et al., 1990). Many volcanoes also have a rejuvenation stage (Macdonald et al., 1983). The chemical evolution of lavas between the stages
has been attributed to variations in the degree of partial melting of the magma source as the volcanoes travel across the hotspot (e.g. Frey et al., 1991).

Products of the shield stage dominate Hawaiian volcano edifices. Only small volumes are preshield and postshield stage products (e.g., Macdonald, 1963; Lipman, 1995; Garcia et al., 1995; Frey et al., 1990). Based on volumes and durations of the stages, lava accumulation rates on the flanks of Hawaiian volcanoes are inferred to increase from about 35 cm per 100 years during the preshield stage (Guillou et al., 1997) to 78-110 cm per 100 years during the shield stage (Sharp et al., 1996) and then decrease to 9 cm per 100 years during the postshield stage (Sharp et al., 1996). The shield stage is subdivided into submarine, emergent, and subaerial phases based on the height of the volcano summit (Walker, 1990; figure 1.2). Submarine and emergent-type (i.e., littoral environment) processes continue to occur on the volcano flanks during the subaerial stage.

1.2.2 Deposits

As water depths decrease throughout the submarine and emergent stages of Hawaiian volcanoes, the incidence of explosive fragmentation of lava is predicted to increase and volcaniclastics should, therefore, become volumetrically more significant (e.g. Fisher and Schmincke, 1984). An increase in the generation and accumulation of non-explosive, epiclastic volcaniclastics may also be expected.
due to flank slopes becoming steeper as the volcano grows (Staudigal and Schmincke, 1984). Volcaniclastic deposits at La Palma in the Canary Islands increase from 20% in deep water to 70% in shallow water (Staudigal and Schmincke, 1984). In the La Palma deep-water succession volcaniclastic deposits are dominated by pillow lava breccias, whereas in the shallow water successions it consists of pillow lava and lapilli breccia in roughly equal proportions along with minor (5%) volcaniclastics, which were called hyaloclastic. Staudigal and Schmincke (1984) used the inferred volatile fragmentation depth (VFD; ~800 mbsl for alkalic lavas) to subdivide the La Palma sequence into shallow submarine and deep submarine environments.

At Hawaiian volcanoes, the deposits of the deep-water stage are generally thought to be pillow lavas and their passively fragmented counterparts (e.g. autobrecciated pillow breccias and quenched hyaloclastites). However, Clague et al., (2000) describe volcaniclastics on Loihi (~1000 mbsl) that are akin to the more explosively formed limu-o-Pele of littoral environments. Observations of Loihi during submersible dives indicate that these and other fragmental deposits are localized (e.g., on some topographic highs) and are volumetrically insignificant compared to the pillow lavas (Garcia et al., 1993; Garcia et al., 1995). Similarly, observations of Mauna Loa's western submarine flank indicate that the submarine section is dominated by pillow lava with interbedded pillow lava breccias and sand layers (Garcia and Davis, 2001). However, submersible
observations of Kilauea volcano indicate that heterogeneous volcanioclastics are the predominant lithology on the lower submarine flanks and coastal-derived glassy sands cover the upper flanks (Lipman et al., 2000).

Kilauea volcano's shallow submarine slopes are steep (25° to 40°) down to 250 mbsl and, consequently, slumps and slides occur frequently (Tribble, 1991). Shallow slopes (<50 mbsl) off Kilauea's south coast where lavas are currently flowing into the ocean are covered with >80% fragmental material that is produced by the combination of littoral explosions, \( \text{H}_2 \) combustion, autobrecciation of flows, and mass wasting of lava (Tribble, 1991). This corresponds with the aforementioned model that emergent stage-type deposits are much more fragmental than those forming in a deeper submarine environment. General causes for the dominance of fragmental deposits may include fragmentation as subaerial lava flows enter the ocean, increased magmatic explosivity and oversteepening of slopes. In very shallow depths (10's m), the higher energy environment may lead to increased erosion and lower deposit accumulation rates (Kokelaar, 1986).

1.2.3 Subsidence and Collapse

Hawaiian islands subside predominantly due to lithospheric flexure under the weight of new volcanoes (e.g. Decker, 1987). Rates decrease away from the hotspot as the lithosphere accommodates the weight of the island and isostatic
equilibrium is approached. Current subsidence rates at Hawai’i are estimated at 2.4 to 2.6 mm/yr (Moore and Thomas, 1988, and Moore and Clague, 1992).

Major flank collapse events have occurred at several Hawaiian volcanoes (e.g. Ko’olau volcano, O’ahu; e.g., Moore and Normark, 1994). Seaward-dipping, poorly-lithified volcaniclastics weakened by high pore-fluid pressures may increase the instability of the volcano edifices (Garcia and Davis, 2001 and Schiffman et al., in review). In addition, magma injection along rift zones may induce slope failures along clay-rich sediment horizons (Schiffman et al., in review), which have extremely low coefficients of friction (Kopf and Brown, 2003). Poorly compacted (i.e., high permeability and high porosity) sediments in combination with clay-rich (i.e., low permeability and low friction) may generate high pore pressures (Schiffman et al., in review) and situations susceptible to hydroplaning. Deposits of large landslides are not expected to be present in the HSDP2 core because the steep gradient of the submarine flanks (10 - 18°, DePaolo and Stolper, 1996) may inhibit deposition. Typically, landslide deposits around the Hawaiian islands have run-out beyond the volcano upper flanks (Moore and Normark, 1994).
1.3 Hawaiian Scientific Drilling Project

1.3.1 Introduction.

The Hawaiian Scientific Drilling Project (HSDP) involves a range of educational and research institutions. Primary institutions are the University of Hawai'i, the University of California at Berkeley, and the California Institute of Technology. Funding for the HSDP came from the National Science Foundation and the International Continental Scientific Drilling Program. HSDP is designed to recover a continuous record of deposits of Mauna Kea volcano, Hawai'i. This record provides a unique opportunity to study the internal stratigraphy of a Hawaiian volcano. The purpose of HSDP was to develop an understanding of the origin of Hawaiian volcanism, volcanic hazards, the Earth's palaeomagnetic field, and groundwater movement in oceanic islands (DePaolo et al., 2001). A pilot hole was drilled in 1993 down to 1,056 mbsl. The second drill hole, HSDP2, reached a depth of 3,098 m and is the deepest continuous core into an oceanic volcano. Plans are to deepen the hole to 4,500 m in 2004.

The HSDP2 drill site is located near the coast (4.2 masl), which was selected to decrease the thickness of subaerial deposits, and to avoid the rift zones of Mauna Loa and Mauna Kea, so influences of intrusives and high-temperature alteration are minimized (DePaolo et al., 2001; figure 1.1). Lavas from the younger Mauna Loa volcano (<130 ka; Lipman and Moore, 1996) cap the Mauna Kea products (>200 ka; Sharp et al., 1996). The boundary between Mauna Loa
and Mauna Kea stratigraphy is at 246 m depth in the HSDP2 core (figure 1.3). This contact is recognized by the lower SiO₂ and total iron, and higher total alkali at a given MgO of the alkalic Mauna Kea lavas (Rhodes, 1996). No interfingerling of lavas from the two volcanoes has been identified (Rhodes, 1996).

Overall core recovery was ~90%, although recovery was <25% in the shallow submarine zone of poorly consolidated volcanics (DePaolo et al., 2001). Poor recovery and other complications forced a change in coring to tri-cone drilling from 1,140 m to 1,222 m and from 1,243 m to 1,260 m. At 1,260 m, the volcanics became more consolidated and recovery improved to almost 90% below this depth (DePaolo et al., 2001).

The HSDP2 core was split longitudinally into an archive portion and a working portion. The working portion is stored at the American Museum of Natural History, New York, and only half of the 6 cm-wide working portion can be sampled. Petrological, geochemical, isotopic, magnetic and geophysical borehole analyses have been made of the HSDP2 core (e.g., Seaman et al., 2004; Stolper et al., 2004; Walton and Schiffman, 2003; Feigenson et al., 2003; Huang and Frey, 2003; Kurz et al., 2004; Kontny et al., 2003).
1.3.2 Basic stratigraphy of the HSDP2 core.

The HSDP2 core contains three basic components (figure 1.3): subaerial Mauna Loa (0 to 246 mbsl), subaerial Mauna Kea (246 to 1,079 mbsl), and submarine Mauna Kea (1,079 to 3,098 mbsl). The submarine-subaerial transition was identified at 1,079 mbsl by the appearance and subsequent dominance of glassy volcaniclastic (>50% glass) and the disappearance of subaerial features such as soils and oxidation of lava flow tops. A single lava flow with massive upper and lower sections (12 and 2.2 m thick) and fragmental middle section (1.7 m thick) spans the transition. Below this flow, glassy volcaniclastics are interbedded with lava flows. Argon-argon dates of the HSDP2 core indicate that Mauna Kea shield deposits are 330 to 635 ka, which yields a vertical growth rate of approximately 8.6 m/ka (Sharp and Renne, in review; figure 1.4).

The upper submarine section (1,079 to 1,984 m) is dominated by glassy basaltic volcaniclastics (~80%) interbedded with massive lava flows with an average thickness of ~3 m (DePaolo et al., 2001; figure 1.3). The lower submarine section (1,984 m to 3,098 m) is characterized by pillow lavas (~65%), interbedded with volcaniclastics (DePaolo et al., 2001). Throughout the submarine section, the volcaniclastic units (<71 m thick) form continuous packages <100 m-thick that are not interrupted by lava flows.
Figure 1.3 a) Basic stratigraphic column outlining the structure of the HSDP core, including the Mauna Loa-Mauna Kea transition, the subaerial-submarine boundary, and the major changes in lithology. Blue box marks the area of study. Note distinct change in volcaniclastic-lava flow proportions in the submarine deposits at ~1,984 mbsl. b) Insert listing units studied and arrows indicate depths of samples examined within the units. Units with asterisks are those described in most detail (after DePaolo et al., 2001).
Figure 1.4 Age-depth relationship of deposits in the HSDP cores based on $^{40}\text{Ar}/^{39}\text{Ar}$ dating (after Sharp and Renne, in review).

- **Age-depth model from DePaolo-Stolper (1996) based on HSDP1 samples**
  - Oldest Alkalic Flow

- **$^{40}\text{Ar}/^{39}\text{Ar}$ Isochron Ages**
  - Samples from HSDP-1
  - Samples from HSDP-2
  - Errors are 95% C.I.

- **Best fit line for samples >380 mbsl** (Sharp and Renne, in review)

- **BASE OF CORE**
  - 550 ka
  - 635 ka
1.3.3 Introduction to the HSDP2 volcaniclastics

Volcaniclastics within the HSDP2 core are summarized as consisting of millimeter to decimeter basaltic clasts in a glassy matrix (DePaolo et al., 2001). The basaltic clasts have a range of vesicularities and olivine contents. The majority of their source magmas were subaerially degassed (based on low sulfur contents; Stolper et al., 2004) and the clasts are thought to have accumulated from the disintegration of lavas originally emplaced near the shoreline (DePaolo et al., 2001). However, the latter interpretation of clast origin is speculative.

Initial logging of the HSDP2 core described bedding, grading, the maximum clast size and the average size of the ten largest clasts, the framework (matrix or clast supported), sorting, lithologies, and lithologic diversity (i.e., monomict versus polymict; Seaman et al., 2000). Basaltic clast types were grouped using vesicularity and mineralogy. Lithologies were named using the diversity of clast types and the mineralogy of the dominant clast type (e.g., basaltic hyaloclastite, polymict, aphyric to moderately olivine-phyric basalt clasts). The relative abundance, modal clast size, size range, angularity, and other comments (e.g., alteration, phenocryst size and shape, and vesicle size and shape) were recorded for every lithology in each core box and internal variations were noted.

HSDP2 unit boundaries were identified using changes in the clast lithology or sedimentary features (e.g. sorting and grain size, etc; Seaman et al., 2000). Unit
boundaries within polymict volcaniclastics were defined wherever the dominant lithology changed (i.e., where the mineralogy indicated a change in the primary lava flow source). Fine grained, well-sorted intervals within an otherwise continuous and consistent volcaniclastic unit were not defined as unit boundaries. Decimeter-thick intervals of coherent crystalline basalt that were indistinguishable between being a clast, a flow or an intrusive were logged as a different unit if they were thicker than ~60 cm.

Modern temperatures in the HSDP2 submarine section are low (varying from 15 to 44°C; DePaolo et al., 2001). Thus, alteration of the core mainly involved hydration of glass and olivine forming clays and zeolites, which are the primary cementing agents for the volcaniclastics. This paragraph summarizes parts of the alteration study by Walton and Schiffman (2003). Sideromelane clasts within HSDP2 volcaniclastics have undergone various stages of weathering involving fracturing, dissolution, smectite replacement, conversion to palagonite, and mineralization of zeolites and Ca-silicates. Dissolution of sideromelane grains is manifested by either empty pore space, residues of fresh olivine, plagioclase and pyroxene microlites, or scattered titaniferous nodules. Smectite constitutes 4 to 32% of the whole rock and there are four types; brown, pore-lining smectite 1, reddened grain-replacing smectite, pale-green, pore-lining smectite 2, and pale-green, grain-replacing smectite. Brown smectite 1 coats all clast types (sideromelane, basalt and crystals) but is absent at contact points between
clasts. Coatings have round corners even where shard margins are angular. Reddened smectite grows in irregularly distributed masses and patches that extend from sideromelane margins inwards. Pale-green smectite contains titaniferous nodules and does not coexist with microbe tubules. Both the pore-lining and replacement types grow from the grain margins and the pore-lining smectite layers occur on the pore-side of brown smectite 1 layers. Palagonite replaces sideromelane from the margins inwards and both the inner and outer contacts are smooth and sharp. Relative to sideromelane, palagonite is depleted in all major elements except TiO₂ and FeO, which are comparatively enriched. Three textures can be present within clasts that have been completely palagonitized; an outer pale-yellow rind, a high relief pitted ring, and an orange-yellow center. Phillipsite occurs in samples deeper than 1403 mbsl, where it constitutes 1 to 6% of the whole rock. Chabazite fills pores and is intergrown with Ca-silicate masses, resulting in an irregular boundary between the two. Chabazite only occurs in samples with palagonite and vice versa. Calcium-silicates partially fill pores and replace early phillipsite.

Microbe tubules appear as thin tracks propagating from vesicle and shard margins into glass interiors. Early stages of development resemble neatly spaced, near-parallel hairs oriented perpendicular to the margin, with transparent 'heads' at the tips. Advanced stages are thicker mats that become less dense towards the clast interior. Longer tracks tend to bend towards olivine.
phenocrysts. Clusters of titaniferous nodules indicate complete dissolution of grains.

Walton and Schiffman, 2003) distinguish three successive zones of alteration within the HSDP2 core. These zones are characterized by various stages of the alteration products described above; Incipient Zone (1080 to 1335 mbsl), Smectitic Zone (1405 to 1573 mbsl) and Palagonitic Zone (1573 mbsl to core base). The incipient zone is characterized by pore-lining brown smectite, dissolution, reddened grain-replacing smectite, titaniferous nodules and microbe tubules. The smectitic zone is distinguished by pale-green pore lining/filling smectite, pale-green grain-replacement smectite, and cementation by phillipsite and Ca-silicate mineralization. The palagonitic zone represents an alternate post-incipient zone trend to the smectitic zone, rather than being consecutive processes (Walton and Schiffman., 2003). The palagonite zone includes palagonite replacing sideromelane clast margins, and pore-filling chabazite. Palagonitization may overprint features formed in the smectitic zone, but smectite is unlikely to replace palagonite.

The subaerial-submarine boundary identified in the HSDP2 core is ~5,280 m below the present day summit of Mauna Kea, and the maximum depositional depth in the core is ~1,450 m (see section 4.2 for details). This relationship indicates that the "submarine stage" of growth (i.e., when the summit of the
volcano is submarine) is unlikely to be recorded in the core. Therefore, it is likely that the volcanlastic rocks were deposited on the submarine flanks of Mauna Kea during its "subaerial stage". Two fundamental phases of volcanlastic production, transport and deposition on the flanks of Mauna Kea are hypothesized:

1. *Deep water flank phase.* Although the summit of Mauna Kea is subaerial, the HSDP2 site is deep submarine (i.e., depth of emplacement is >400-500 mbsl; Seaman et al., 2004). Lavas (or fragments of lavas) being deposited at the HSDP2 site may be erupted from subaerial or submarine vents on Mauna Kea. Any eruptions occurring near the HSDP2 site are dominated by non-explosive fragmentation processes, such as hyaloclastic, autoclastic, and epiclastic. Volcanlastic lithologies from these eruptions are dominated by pillow lava breccias and pillow fragment breccias. Other lithologies include hyaloclastites and isolated-pillow lava breccias.

2. *Shallow water flank phase.* The summit of Mauna Kea is subaerial and the HSDP2 site shoals from <400-500 mbsl to sea level. An increase is expected in the occurrence of material from subaerially erupted lavas that fragmented during flowage into the ocean. The predominant lithology at the drill site is dependant upon the activity of the volcano. If subaerial eruptions are long-lived, voluminous, and relatively constant, it is expected that deposits are similar to those forming today at Kilauea's active ocean entry (e.g., Tribble, 1991; Mattox, 1993; Mattox and Mangan, 1997). That is, fragmentation of subaerial lavas
occurs by steam explosions, quenching, autobrecciation and wave action as the flows enter the ocean and create unstable deltas. These deltas commonly collapse and gravity flows redeposit the material downslope on the volcano's submarine flanks (e.g., to the HSDP2 site). Fragmentation during periods of quiescence (10's to 100's of years long) is dominantly erosional and products are primarily beach deposits of eroded lava flows and pre-existing volcaniclastics.

The following questions will be addressed:

- What fragmentation processes (see section 1.5 for details) contributed to the formation of volcaniclastic deposits and did a particular process dominate?

Deciphering the fragmentation processes will help determine the phase of growth recorded by the HSDP2 core and assess the eruptive activity of Mauna Kea: By recognizing the fragmentation processes that produced the clasts within the volcaniclastics it will be possible to determine the environment of clast fragmentation (see section 1.5 for discussion). For example, clasts that are fragmented by magmatic explosivity are most likely to be derived from a subaerial eruption. However, there are exceptions and complications (e.g., magmatic explosivity can occur subaqueously; Clague et al., 2000). Therefore, it is important to distinguish all the fragmentation processes involved and determine which one is dominant. For example, if a deposit includes clasts produced by magmatic fragmentation but is dominated by clasts fragmented by
quench-granulation then the source environment was probably subaqueous rather than subaerial.

- Do the deposits have homogeneous glass chemistry (i.e., products from one eruption of a uniform magma composition) or heterogeneous chemistry (i.e., products from more than one eruption)? Volumetrically dominant homogeneous deposits may indicate that growth at the HSDP2 site was dominated by the products of single eruptions that were deposited too rapidly to be mixed with material from other eruptions. In this case, the fragmentation process identified must have been efficient at fragmenting a high volume flux of magma. Thin homogeneous deposits suggest there were periods of relative quiescence with pulsed or small eruptions. Heterogeneous deposits indicate either a very active stage of growth (multiple eruptions occurring simultaneously within the supply region of the site; primary deposits) or, more likely, extended periods of erosion during a time of infrequent eruptions within this region.

- To what extent is the primary clast population modified by events following initial fragmentation, such as thermal contraction and cracking, grain-grain collisions during transport, and abrasion and/or reworking by wave or current action? Clast modification represents the processes that occurred after initial clast formation (during transport, deposition and, potentially, redeposition). These processes can indicate the environment of transport and deposition (e.g., wave action affects clasts deposited in a littoral environment) and stability of the area (e.g., evidence for extensive wave action suggests the clasts remained
exposed in the environment for a substantial period of time). Clast modification can indicate erosive processes (e.g., collapse) that are important for understanding the stability of the volcano flanks, and hence relative slope angle.

- What depositional processes were involved in the production of the volcaniclastics? These processes may reflect the distance the material traveled from the source and the type, rate and volume of material production. For example, intervals of continued cooling contraction may induce a turbidity current of relatively fine grained and well sorted material, whereas collapse of a lava delta may instigate a debris flow of coarser grains.

### 1.4 Terminology

HSDP2 volcaniclastics were classified as predominantly hyaloclastites (DePaolo et al., 2001). This term was first used to describe the glassy, sand-sized fragmental deposits produced by the non-explosive process of quenching hot lava (Rittmann, 1960). Subsequently, it has been used to describe deposits of many types and origins. For example, "any sand-sized volcaniclastics" (e.g. Schmincke et al., 1979), "any glassy, subaqueous basaltic clastic material" (e.g. Staudigal and Schmincke, 1984), "any clastics formed by explosive fragmentation" (i.e., via magmatic exsolution and steam explosivity; e.g. Batiza et al., 1984), and autoclastic lava flow breccias of subaqueous origin (e.g., Cas and Wright, 1987). Even more confusing are the papers in which the term has been used without clarification of its definition.
This digression from the original definition has allowed investigators to lump all subaqueous volcaniclastic material under the same term, *hyaloclastite*, regardless of their origin. Consequently, the issue of what these rocks actually represent is evaded and progress in the interpretation and understanding of subaqueous fragmental material has been hindered. It is clear that a consensus is needed regarding the classification of submarine fragmental rocks of volcanic origin, however, this re-evaluation is outside the scope of this project.

The following terms are used in this work: *Volcaniclastic* is any clastic material made up of volcanic constituents regardless of the fragmentation, transport and depositional mechanisms (e.g. Fisher, 1961; Schneider et al., 1998). The terms *pyroclastic, autoclastic* and *epiclastic* imply contrasting styles of fragmentation of the clastic material via magmatic explosivity or explosive magma:water interaction, autobrecciation and erosion of pre-existing rocks, respectively. *Hyaloclastite* is a describes fragments produced by the passive interaction of water and magma. These terms are not associated with specific transport or depositional mechanisms. Volcaniclastics that have undergone secondary transportation are *reworked* deposits (Schneider et al., 1998).

Volcaniclastic units are also classified according to their grain size, for example ash (0.2 to 2 mm), lapilli (2 to 64 mm) or breccia (>64 mm). Vesicularity terms follow the index in Houghton and Wilson (1989) where non-vesicular is <5%
vesicles, incipiently vesicular is 5-20%, poorly vesicular is 20-40%, moderately vesicular is 40-60%, highly vesicular is 60-80%, and extremely vesicular is >80%.

1.5 Subaqueous fragmentation of basaltic magma

Four mechanisms of lava fragmentation are recognized in subaqueous environments (e.g. Fisher and Schmincke, 1984): Primary fragmentation via magmatic explosivity, explosive and passive magma-water interaction (figure 1.5), and secondary fragmentation via disintegration of pre-existing rocks by erosion.

1.5.1 Magmatic explosivity

When the vesicularity of magma reaches some critical value (typically 75-85 volume %), fragmentation can occur via breakage of the bubble walls (Head and Wilson, 2003). Volatile exsolution is dependent on volatile solubility and therefore, the critical amount of dissolved volatiles needed in the melt to reach this vesicularity is principally dependent on pressure. Critical weight percents of dissolved CO₂ and H₂O were calculated to be 1.1-6.7 and 1.9-12.1 respectively for magma between 500 and 3500 mbsl (Head and Wilson, 2003). Such high values are rare in mafic magmas and do not occur in the HSDP2 core (Seaman et al., 2004), but if the effects of CO₂ and H₂O are combined, the critical values decrease and magmatic explosive fragmentation of mafic magma at greater depths is theoretically viable (Head and Wilson, 2003).
Theoretically, submarine fountaining can occur at depth even if the volatile content of the magma is below the critical values. Rapid and disequilibrium gas exsolution from a melt may disrupt and fragment magma in the conduit (Head and Wilson, 2003). If this occurs, the magma and bubbles may ascend at roughly equal and rapid speeds (>1 m per second) resulting in a submarine Hawaiian-style fountaining eruption. Strombolian-style eruptions occur when the bubble ascent rate is faster than the magma ascent rate. Eruption dynamics of subaqueous Strombolian eruptions are similar to subaerial cases (Head and Wilson, 2003). Vulcanian eruptions are generated when pressure from gas accumulation in the magma exceeds the confining pressure of the overlying relatively viscous melt. In subaqueous environments, water interacts with newly exposed magma causing contact-surface steam explosivity and bulk interaction steam explosivity (Head and Wilson, 2003).

1.5.2 Steam explosivity

Steam explosivity involves the heating of water by magma followed by either explosive expansion of trapped water (bulk interaction; Kokelaar, 1986) or cyclic expansion & collapse of steam (contact-surface interaction; e.g. Wohletz, 1983; Wohletz, 1986; Sheridan and Wohletz, 1983). Steam explosivity is inhibited by pressure: when water is heated to 500\(^\circ\)C, steam expands by 3566 times at sea level but only by 1000 times at 45 mbsl (Kokelaar and Durant, 1983).
Figure 1.5 Flow chart of passive and explosive fragmentation processes that can arise from magma-water interaction (Modified after Kokelaar, 1986, and Wohletz, 1986).

PASSIVE QUENCHING

Water quenches the surface of the lava and forms a brittle skin around the hot interior

- Cooling-Contraction
- Granulation

Skin fragments as the interior cools and contracts

PASSIVE FRAGMENTATION

MAGMA-WATER INTERACTION

EXPLOSIVE FRAGMENTATION

CONTACT-SURFACE

- Steam bubbles coalesce into a thin steam film at the magma-water contact
- Expansion and collapse of steam layer on microsecond scale causes fine fragmentation of magma and turbulent mixing
- Spontaneous Nucleation (superheated water vaporizes instantaneously)
- Thermal Detonation (rapid vaporization behind a propagating shock)

BULK INTERACTION

- Rapid heating and vaporization of water either enclosed in magma or trapped near magma
- Explosively expanding steam
- Pressure waves
- Clasts form by tearing apart of fluid magma
- Clasts form by shattering of rigid magma
Bulk interaction requires trapping water either near the magma (e.g., in wet sediment) or beneath the magma (e.g., as lava enters the ocean). Fragmentation is inferred to occur via tearing, as the steam explosively expands, and shattering of brittle magma by shock waves (Kokelaar, 1986). Contact-surface steam explosivity occurs at the magma-water interface and is analogous to fuel coolant interactions (FCI). Two triggers for FCI have been modeled: spontaneous nucleation and thermal detonation (Wohletz, 1983; Wohletz, 1986). Both models are self-sustaining and need a sudden and irregular contact between the water and magma in order to have a high surface area for adequate heat transfer.

*Littoral magma:water interaction*

Once Hawaiian volcanoes breach the ocean surface, pahoehoe and a'a lava flows build the subaerial portion of the shield. Lateral growth of the subaerial portion of shields occurs partly via flowage of lava into the ocean. Pahoehoe flows initially reach the shoreline as surface flows and are passively quenched into glassy blocks and lapilli as they cross the surf zone (Mattox and Mangan, 1997). The fragmental material builds debris slopes in the submarine littoral zone that interfinger with pahoehoe flows. When the debris pile reaches sea level it is capped by pahoehoe lavas and a lava delta is formed within which lava tube systems can develop (Moore et al., 1973). Lava can now enter the ocean from tube terminals located at or below sea level.
Lava deltas are prone to collapse into the ocean because they are built on ocean-facing slopes, they often contain ocean-facing failure scarps (pre-existing cliff or new failure surface; Mattox and Mangan, 1997), and there are layers of wet sand, which create slip-surfaces. When deltas collapse, magma in the tubes is abruptly exposed to ocean water, which often results in explosions. Four types of explosions can occur: tephra jets, tephra blasts, bubble bursts, and littoral lava fountains (Mattox and Mangan, 1997).

Complete collapse of lava deltas allows open mixing of water with molten lava, creating tephra jets, and contact of water with hot rocks, creating tephra blasts (Mattox and Mangan, 1997). Wave action and continuous flow of lava through the severed tube allows tephra jets to occur sporadically for hours to days, until the tube crusts over, whereas tephra blasts are singular events (Mattox and Mangan, 1997). Tephra jets include steam, juvenile lapilli/bombs, and lithic ash. Tephra blasts consist of lithic fragments ripped apart from recently solidified magma and surrounding rock. Clasts fragmented by tephra blasts will be more blocky than those produced by tephra jets (Mattox and Mangan, 1997).

Partial collapse of lava deltas allows confined mixing of water and magma within the tubes, producing sporadic lava bubble bursts through holes in the tube roof and overlying crust 10’s to 100’s of meters inland of the ocean, and lava fountains (Hon et al., 1988; Mattox and Mangan, 1997). Lava bubbles are
formed when thin sheets of melt are blown into large (<10 m) dome-shaped bubbles by the expansion of steam from trapped ocean water. Almost immediately, the bubble bursts and limu-o-Pele fragments are thrown outwards and the remaining magma drains back into the lava tube. Unlike the products of explosions following complete delta collapse, the fragments of bubble bursting are entirely juvenile (Mattox and Mangan, 1997). Similar processes to those that form limu-o-Pele may also occur subaqueously (<1200 mbsl; Clague et al., 2000).

Subaqueous magma-water interaction

The ratio of magma to water affects the explosivity of subaqueous magma-water interaction (Sheridan and Wohletz, 1983; Wohletz, 1986). Too much water inhibits contact surface processes that may cause explosive activity. Therefore, the probability of explosive subaqueous magma-water interaction is small. In addition, explosivity is influenced by the rate and amount of heat exchange between the magma and water and therefore, the contact geometry and duration are an important factor (Sheridan and Wohletz, 1983; Wohletz, 1986).

Formation of limu-o-Pele clasts requires stretching of fluidal lava into a very thin film. This stretching and fragmentation can occur via production of 'lava bubbles' by rapid expansion of external water trapped by the lava or of magmatic volatiles, otherwise the lava quenches before the bubble forms. The rate of boiling of
water into steam is inhibited by pressure, so with increasing depths, greater energy is needed to form limu-o-Pele. However, submarine equivalents of littoral limu-o-Pele have been observed at Loihi at depths of 1200 +/- 300 mbsl (Clague et al., 2000).

1.5.3 Passive quenching and autobrecciation of lava.

The surface of lava erupted subaqueously cools so rapidly that it forms a brittle, glassy skin, beneath which the rheology grades inward from viscoelastic to fluid (e.g., Hon et al., 1994). Thermal contraction upon cooling will cause the brittle skin to crack into sand and gravel sized fragments (*cooling-contraction granulation*). This process is often enhanced by inflation (*spallation*). Rapid quenching inhibits vesiculation (unless a high content of volatiles are present), thus clasts are often non-vesicular. Fragments are flake-like to equant, and are cold upon deposition because they rapidly lose heat to the surrounding water (Fisher and Schmincke, 1984). *Autobrecciation* also requires chilling of the lava margin but fragmentation occurs in dynamic fashion, as the still viscous body of the flow continues to move, breaking apart the brittle margin. Steep slopes promote autobrecciation.

1.5.4 Secondary fragmentation of basaltic magma.

Mass wasting and longer-term erosion (e.g., wave action) fragment lava flows that have cooled and solidified. These processes of secondary fragmentation
may be enhanced by steep slopes, unstable strata, or shallow submarine environments. Factors contributing to the failure of submarine slopes include high internal water pressures (from rapid deposition of sediment and trapping of water), oversteepening of slopes (from high sedimentation rates), and ground shaking during earthquakes. Oversteepening and subsequent mass wasting events (e.g. slumps, slides, and debris flows) are common at littoral submarine slopes of Hawaiian volcanoes where lava is flowing into the ocean (Moore and Chadwick, 1995). Mass wasting events disaggregate pre-existing rocks and redeposit them downslope. Turbidite currents associated with debris flows are able to carry material for 100's of kilometers from their source (e.g. Garcia, 1996). Erosive wave and current actions increase as the volcano shoals, intensifying the redistribution of fragmental material.
The HSDP2 core provides us with a nearly continuous lithostratigraphic record of ocean island deposits, far exceeding the thickness of surface outcrops. Thus, not only does the core give us an opportunity to reconstruct the stratigraphic record, but also to study lithological characteristics in great detail (particularly vertical changes in texture, structures, componentry and chemical composition).

2.1 Descriptions of the Deposits
Criteria used to characterize the volcaniclastic deposits in the HSDP2 core include macroscopic structures, grain size and sorting, clast lithologies, clast vesicularities and crystallinities, clast shapes, chemical composition, heterogeneity of glassy clasts and alteration. Techniques based on information in two- or three- dimensions, such as geometric relationships, are of limited value in studying drill cores.

Macroscopic structures of volcaniclastic lithologies incorporate bedding, framework, fabric, grading and contacts. Together with grain size and sorting, these are standard criteria used to interpret transport and depositional processes of volcaniclastic rocks (Compton, 1985). Clast shapes are useful indicators of fragmentation processes and degree of reworking involved in the production of
Table 2.1 Table summarizing key features of sedimentary deposits in a marine environment and their corresponding unique or extreme interpretation. The endless variations on the interpretations and combination of features are not included.

A. BEDDING / GRADING / SEDIMENTARY STRUCTURES (10’s cm to 10 m scale features)

<table>
<thead>
<tr>
<th>Reference</th>
<th>FEATURES</th>
<th>INTERPRETATION</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1.1</td>
<td>Nonbedded</td>
<td>No changes in the pattern of sedimentation</td>
</tr>
<tr>
<td>A1.2</td>
<td>Beds of similar grainsize</td>
<td>Sequence of events with recurring similar sedimentary mechanisms with time</td>
</tr>
<tr>
<td>A1.3</td>
<td>Fluctuations in grainsize</td>
<td>Fluctuations in transport energy</td>
</tr>
<tr>
<td>A1.4</td>
<td>Alternating coarse- and fine- (silt/clay) grained beds</td>
<td>Constant sedimentation of fine material overprinted with periodic sedimentation of coarse debris.</td>
</tr>
<tr>
<td>A2.1</td>
<td>Massive</td>
<td>No traction transport or sedimentation from high concentration suspension, and very rapid deposition (or post-depositional bioturbation or liquefaction).</td>
</tr>
<tr>
<td>A2.2</td>
<td>Cross bedding/lamination and ripples</td>
<td>Turbulent lateral emplacement.</td>
</tr>
<tr>
<td>A2.3</td>
<td>Planar bedding/lamination</td>
<td>Equivocal - multiple possible interpretations.</td>
</tr>
<tr>
<td>A3.1</td>
<td>Mantling beds</td>
<td>Clasts settled and blanketed pre-existing topography</td>
</tr>
<tr>
<td>A3.2</td>
<td>Bed thickens in channels</td>
<td>Lateral transport with enough energy to overtop highs</td>
</tr>
<tr>
<td>A3.3</td>
<td>Bed confined to channels</td>
<td>Lateral transport with insufficient energy to overtop highs.</td>
</tr>
<tr>
<td>A4.1</td>
<td>No grading</td>
<td>Constant source and supply of sediments and/or rapid deposition</td>
</tr>
<tr>
<td>A4.2</td>
<td>Normal grading</td>
<td>Equivocal - multiple possible interpretations.</td>
</tr>
<tr>
<td>A4.3</td>
<td>Reverse grading</td>
<td>Size fractionation during settling of clast population with a range of densities (e.g., including scoria).</td>
</tr>
<tr>
<td>A5.1</td>
<td>Non erosive base</td>
<td>Deposition from transport system with insufficient transport to erode sediment below.</td>
</tr>
<tr>
<td>A5.2</td>
<td>Erosive base</td>
<td>Lateral transport (sub-parallel to pre-existing topography) of higher energy than the transport of the previously deposited bed.</td>
</tr>
<tr>
<td>A6.1</td>
<td>Flat top</td>
<td>Equivocal - multiple possible interpretations.</td>
</tr>
<tr>
<td>A6.2</td>
<td>Irregular top (large grains projecting)</td>
<td>Settling or winnowing of finer sediments within deposit or matrix-supported flow.</td>
</tr>
<tr>
<td>A6.3</td>
<td>Rippled top</td>
<td>Current or wave action.</td>
</tr>
</tbody>
</table>
### B. GRAIN SIZE & SORTING

<table>
<thead>
<tr>
<th>Reference</th>
<th>Feature</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>B1</td>
<td>Fine-grained, well sorted</td>
<td>Source of exclusively fine grained clasts or long transport time.</td>
</tr>
<tr>
<td>B2</td>
<td>Coarse-grained, well sorted</td>
<td>Source of exclusively coarse grained clasts or sorting by transport energy (including winnowing). Close to source.</td>
</tr>
<tr>
<td>B3</td>
<td>Poorly sorted</td>
<td>Rapid deposition and/or minimal size sorting with transport.</td>
</tr>
</tbody>
</table>

### C. CLAST SHAPES

<table>
<thead>
<tr>
<th>Reference</th>
<th>Feature</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>C1</td>
<td>Well-rounded to subrounded ovoid</td>
<td>Secondary reworking</td>
</tr>
<tr>
<td>C2</td>
<td>Glassy, perfect spheres</td>
<td>Lava fountaining</td>
</tr>
<tr>
<td>C3</td>
<td>Curved shard</td>
<td>Spallation of lava or bubble bursts (limu-o-Pele)</td>
</tr>
<tr>
<td>C4</td>
<td>Blocky, wedge-shapes</td>
<td>Cooling-contraction fragmentation of pillow lavas</td>
</tr>
<tr>
<td>C5</td>
<td>Irregular, angular margin (ragged)</td>
<td>Fragmentation of brittle magma with involvement of magmatic volatiles</td>
</tr>
<tr>
<td>C6</td>
<td>Irregular, fluidal margin</td>
<td>Fragmentation of fluidal magma</td>
</tr>
<tr>
<td>C7</td>
<td>Planar/curviplanar, angular margin</td>
<td>Fragmentation of brittle magma with no involvement of magmatic volatiles</td>
</tr>
<tr>
<td>C8</td>
<td>Planar/curviplanar, fluidal margin</td>
<td>Fragmentation of fluidal magma with no involvement of magmatic volatiles</td>
</tr>
<tr>
<td>C9</td>
<td>Fluidal form</td>
<td>Fragmentation and transport of fluidal magma.</td>
</tr>
</tbody>
</table>

### D. SECONDARY FEATURES OF CLASTS

<table>
<thead>
<tr>
<th>Reference</th>
<th>Feature</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>D1</td>
<td>Oxidization</td>
<td>Alteration in subaerial environment</td>
</tr>
<tr>
<td>D2</td>
<td>High temperature alteration</td>
<td>Alteration near to or within the vent</td>
</tr>
<tr>
<td>D3</td>
<td>Low temperature alteration</td>
<td>Alteration outside area of heat from magma (far from vent).</td>
</tr>
</tbody>
</table>
### E. Juvenile Clast Groundmass Textures

<table>
<thead>
<tr>
<th>Reference</th>
<th>Feature</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>E1</td>
<td>Crystalline basalt</td>
<td>Derived from interior of lava flows or dikes</td>
</tr>
<tr>
<td>E2</td>
<td>Tachylite: moderate to high vesicularity</td>
<td>Slow quenching; fragmentation involved magmatic volatiles</td>
</tr>
<tr>
<td>E3</td>
<td>Tachylite: no to low vesicularity</td>
<td>Slow quenching; fragmentation did not involve magmatic volatiles</td>
</tr>
<tr>
<td>E4</td>
<td>Sideromelane: moderate to high vesicularity</td>
<td>Rapid quenching; fragmentation involved magmatic volatiles.</td>
</tr>
<tr>
<td>E5</td>
<td>Sideromelane: no to low vesicularity</td>
<td>Rapid quenching; fragmentation did not involve magmatic volatiles.</td>
</tr>
</tbody>
</table>

### F. Chemical Heterogeneity of Glassy Clasts

<table>
<thead>
<tr>
<th>Reference</th>
<th>Feature</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>F1</td>
<td>Homogeneous</td>
<td>Single magma batch and probably a short eruption time (no long-term evolution of chemical composition) and a short depositional time (no extraneous material included)</td>
</tr>
<tr>
<td>F2</td>
<td>Slight range in composition</td>
<td>Single magma batch erupted after significant repose period so that the eruption taps a heterogeneous magma source.</td>
</tr>
<tr>
<td>F3</td>
<td>Small percentage of chemically distinct clasts</td>
<td>Limited incorporation of previously erupted material during transport or incorporation during deposition</td>
</tr>
<tr>
<td>F4</td>
<td>&gt;1 major population of chemically distinct clasts</td>
<td>Material from &gt;1 eruption and &gt;1 magma batch entering depositional system. Implication of extended duration for depositional processes (e.g., erosion).</td>
</tr>
</tbody>
</table>

### G. Nonjuvenile Clasts

<table>
<thead>
<tr>
<th>Reference</th>
<th>Feature</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>G1</td>
<td>Pelagic microfossils</td>
<td>&lt;1 vol % High sedimentation rate of volcanioclastics</td>
</tr>
<tr>
<td>G2</td>
<td></td>
<td>1-10 vol % Intermediate sedimentation rate of volcanioclastics</td>
</tr>
<tr>
<td>G3</td>
<td></td>
<td>&gt;10 vol % Slow sedimentation rate of volcanioclastics</td>
</tr>
<tr>
<td>G4</td>
<td>Interbedded pelagic muds</td>
<td>Intermittent volcanic deposition with very long repose periods.</td>
</tr>
<tr>
<td>G5</td>
<td>Reef limestone</td>
<td>Low energy, sediment-starved, shallow environment</td>
</tr>
<tr>
<td>G6</td>
<td>Reworked fragments of pre-existing rock</td>
<td>Sudden catastrophic events (tectonic collapse or volcanic explosions) or prolonged energetic erosion</td>
</tr>
</tbody>
</table>
the clasts and a separate section (2.3) is dedicated to describing the quantitative work done on the samples. Secondary mineralization and veins are often specific to certain post-depositional environments and can be used to estimate palaeo-locations (e.g. proximity to heat provided by a vent). Compositional or mineralogical heterogeneity in clast populations are key indicators of the number of sources involved and it is especially useful for distinguishing primary versus reworked deposits.

The features listed above are used to describe the deposits and the most straightforward interpretations that can be drawn from each are tabulated from large- to small-scale (see table 2.1). Supplementary detail and explanation is given in the accompanying text. Reference numbers are used to link features in the table with the accompanying text. Only structures likely to be found on the submarine flanks of ocean island volcanoes are considered. For simplicity, only the most probable and/or extreme interpretations are presented in these tables. Alternative interpretations are possible if the environment of deposition is poorly known.

2.1.1 Macroscopic features (centimeter to decimeter scale)

Descriptions made on scales ranging from centimeters to several 10's of meters are divided into six categories based on types of bedding, structure, grading and contacts (Table 2.1A). Some features, such as cross-bedding (A2.2), bedding
geometries (A3), and contact characteristics (A6) may not be obvious or well-preserved in a drill core such as the 6 cm wide HSDP2 core. In addition, evidence of erosion (e.g. nonconformities) may be difficult to detect.

**Bedding**

It is unlikely that deposition on the submarine flank of a Hawaiian volcano is continuous, although sedimentation rates will be high at times. Breaks or changes in deposition are reflected in bedding, whereas absence of bedding planes may indicate continuity in supply and sedimentation. Presence of sharply defined bedding planes indicate episodic deposition or intervals of erosion. Repetitive sequences of beds of similar grainsize indicate recurring deposition by similar sedimentary mechanisms (figure 2.1a). Fluctuations in the transport energy levels can be represented by well-defined beds of contrasting grainsize or, in absence of bedding planes, by gradational changes in grainsize (figure 2.1b). However, sequences of interbedded coarse volcanioclastics and silt- to clay-size material are more likely to develop by constant deposition (settling) of the fine fraction overprinted by sporadic but rapid deposition of coarse particles from a separate source rather than switches of the sedimentation regime (figure 2.1c).
Figure 2.1 Sketch showing variations of types of bedding. a) Episodic accumulation of beds of similar grain size; b) Gradational changes in grain size and hence transport efficiency; c) Coarse-fine couplets.

**Sedimentary Structures**

Sedimentary structures reflect the intensity and transport direction of the energy depositional media. Cross bedding, cross lamination and ripples indicate deposition from lateral transport whereas planar bedding and planar lamination can form both by vertical settling and during lateral transport. Beds with upper contacts that are parallel with the preexisting topography are formed by clasts settling from suspension and blanketing the ocean floor (figure 2.2a). Beds that thicken into, or are confined to, topographic lows (e.g., channels) are transported laterally and the transport energy is reflected by the deposit thickness on topographic highs (figure 2.2b). Absence of deposition on highs indicates that the sediment current was completely confined to topographic depressions (figure 2.2c).
Figure 2.2 Examples showing the possible relationship between bedding planes of newly formed deposits and preexisting topography. a) Mantling beds: upper bedding plane is parallel with preexisting topography. b) Bed thickens in channels. c) Beds confined to channels.

*Grading*

Changes in sediment supply (e.g., rate of sedimentation) and properties of the transport media during deposition may result in size or density grading of clasts. Often no single, unique explanation is available. For example, normal grading could indicate either increasingly efficient fragmentation at the site of clast production or a decline in the energy of the transport medium. Opposite changes in the clast supply and transport medium create reverse grading. Absence of grading reflects steady state conditions of sediment supply and deposition.

*Basal and Upper Contacts*

Erosion at basal contacts is a measure of the transport strength and periods of current activity and/or deposition (i.e., lateral current activity). The shape of the
upper contact is specific either to the type of clast support (matrix versus clast) and transport, or to post-depositional reworking. For example, flows supported mainly by matrix-strength may have enough cohesion between the grains that larger clasts are held at the upper margin of the flow and, therefore, cause an irregular upper margin (figure 2.3, example A6.2; Mulder and Alexander, 2001).

Figure 2.3 Graphic log illustrating examples of three endmember types of upper contact: planar (A6.1), irregular (A6.2), and rippled (A6.3).

Sediments deposited out of suspension always have non-erosive bases. However, deposits with sedimentary structures indicative of current transport that have a non-erosive base must be deposited by a current with equal or lower energy than that of the preceding bed, or the underlying material is resistant to erosion (either due to diagenesis or high surface tension, e.g., clay- and silt-rich beds). An erosive base denotes transport and deposition in a higher energy environment than that of the preceding bed in order to move the material.
A flat upper contact can form by several processes ranging from current planning to deposition from suspension to deposition from a lateral current. Upper contacts with projecting large clasts (figure 2.3) are most likely to indicate a matrix-supported flow with sufficient internal strength that the larger clasts 'float' at the top (i.e., cohesive flows such as debris flows). Similar upper margins can develop by post-depositional winnowing of fines between larger clasts. Rippled upper contacts are formed by laterally moving transport media, whether unidirectional (currents) or bi-directional (waves).

2.1.2 Grain Size & Sorting

Source conditions (e.g. fragmentation efficiency), transport mechanisms and intensity, and depositional processes, determine grain size and sorting. Well-sorted deposits indicate either a uniform fragmentation process or size sorting of material during transport. For example, fine-grained and well-sorted material can be from a single source of fine clasts (e.g., volcanic ash) or deposition from a low-capacity transport medium (e.g., turbidity currents). Coarse grained deposits may be products of inefficient fragmentation mechanisms such as autobrecciation of lava flows and pillow lavas. Well-sorted accumulations of coarse grained deposits may indicate proximity to vent or post-depositional winnowing out of fine sediments from the deposit. Sorting of both fine grained and coarse grained material can result during transport. Processes leading to
poorly sorted beds include rapid deposition after little or no transport and deposition from a high particle concentration medium such as a debris flow.

2.1.3 Lithologic diversity and clast types

The sources of the volcaniclastic material are basaltic lava from upslope Mauna Kea (Rhodes, 1996). Eruption style and locale determine fragment characteristics such as grainsize, groundmass texture, vesicularity and shape. Therefore, it is possible to group clasts into lithologic types based on simple observations and then use these estimates of lithologic diversity to distinguish deposits that are derived from a single sustained versus multiple eruptions. As a first approximation, units containing texturally uniform clasts are taken to represent deposits from the products of a single eruption, whereas texturally diverse clast deposits are taken to indicate products from more than one eruption. However, the products of a single eruption can exhibit significant petrologic and textural variations (e.g., crystallinity can range from sideromelane to holocrystalline in a lava flow; Kawachi and Pringle, 1988). Correspondingly, unrelated eruptions can produce fragments that appear similar. In an attempt to resolve such issues, major element analyses of glass clasts were conducted to determine chemical compositions. Deposits characterized by heterogeneous glass compositions are clearly derived from products of multiple eruptions that are not co-magmatic, whereas homogeneous composition of clasts points towards an origin involving the products of a single eruption (section 2.1.7).
During hand-specimen core inspection, clasts were grouped into types based upon vesicularity, crystallinity and phenocryst content, mineralogy, shape, and degree of secondary mineralization and alteration. Modal analysis of clast types was achieved by 100-point counts on grid sizes ranging from 5 x 5 mm to 11 x 13 mm, depending upon grain size (>4 times the size of the largest clast).

2.1.4 Clast Shapes

Clast shapes are predominantly controlled by source fragmentation mechanisms and subsequent modification via abrasion and fragmentation during transport (for details see section 2.2). Nine key clast types for submarine flanks of oceanic volcanoes are recognized here (Figure 2.4). The original fragmentation process for rounded clasts (C1) may be obscured by abrasion during transport. Glassy spheres (C2) form as individual melt droplets with high surface tension in lava fountains of low-viscosity melts (Heiken and Lofgren, 1971) and in phreatomagmatic deposits (Wohletz, 1986; Morrissey et al., 2000). Curved shards (C3) created by spallation will produce less glassy (including tachylite) and thicker shards than those produced by the bursting of very large bubbles or from vesicular melt (e.g., limu-o-Pele, Mattox and Mangan, 1997; Batiza et al., 1984). Deposits consisting of grains formed by spallation may also contain wedge-shaped clasts with crystallinities ranging from tachylite (i.e. cryptocrystalline) to holocrystalline. Limu-o-Pele shards are thin, delicate
sideromelane that may have elongate vesicles formed by deformation (stretching) of the fluid magma (e.g., Clague et al., 2000).

Blocky, wedge-shaped clasts produced by cooling-contraction fragmentation of pillow lavas should display a range of crystallinites because the outermost selvage of the pillow is cooled rapidly upon contact with water and quenched to glass, whereas the insulated pillow interior cools slower and thus exhibits higher degrees of crystallization (Carlisle, 1963; Kawachi and Pringle, 1988).

Figure 2.4 Illustrations of the endmember types of clast shapes that can be produced by magma:water interaction.

Clast shape types C5 through C8 represent four extreme shapes produced by fragmentation of brittle magma (angular margins; C5 and C7) to molten magma (fluidal margins; C6 and C8) with variable vesicularities from high (irregular margins; C5 and C6) to none (planar margins; C7 and C8). There are many
situations that may result in the production of any one, or any combination of these extremes and a mixture of all the clast shapes is expected within a deposit. Other collaborating evidence is needed to determine more detailed processes of formation. For example, if type C7 (planar, irregular margins) is blocky and glassy with pervasive cracks then it was probably formed by quench fragmentation (cracking and fragmentation following thermal relaxation). However, if type C7 is blocky and has incipient cracks on the surface then it had a hot interior that continued to expand after the margins were brittle (breadcrust texture). Finally, if C7 has a holocrystalline texture then it is most likely to have formed by erosion or autobrecciation of a lava flow or pillow lava. Fluidal clasts (C9) are shaped by resistance of air or water during transport whilst the lava is still molten.

Phreatomagmatic eruptions produce a wide variety of clast shapes and therefore the proportions of C2 to C8 clasts will be important for distinguishing the dominant eruptive processes. For example, a deposit consisting predominantly of angular clasts with planar margins (C7) with lesser amounts of fluidal clasts with planar margins (C8) and spalled curved shards (C3), and rare limu-o-Pele curved shards (C3) probably originated in a littoral environment where lava flows and/or tubes entered the ocean. Spatter bombs may also be produced by littoral eruptions. In this case, magma-quenching and autobrecciation would be the volumetrically dominant fragmentation processes (producing C7 clasts ranging in
texture from sideromelane to crystalline) and other mechanisms would include tephra jets (producing C8 clasts), spallation (C3) and bubble bursts (C3).

2.1.5 Secondary Features of Clasts

High temperature alteration of clasts is most likely to occur near to or within the vent soon after eruption. Alteration that may indicate high temperatures include the replacement of volcanic components (e.g., glass, calcic plagioclase, pyroxene and olivine) by minerals such as quartz (100-230°C), laumontite (100-200°C), adularia (>230°C), albite (150-230°C), and chlorite (150-250°C; Wohletz and Heiken, 1992). In contrast, alteration by low temperature fluids starts at deposition and requires significant time periods to develop (1000's of years). Deposits with mixed altered and unaltered material are likely to be derived from more than one source (separated by age and/or location).

Nomenclature used in the descriptions of alteration within the units is from Walton and Schiffman (2003), which describes volcanioclastic alteration in the HSDP2 core (outlined in section 3.3).

2.1.6 Juvenile Clast Groundmass Textures

Vesicularity reflects the degree of involvement of magmatic volatiles in the fragmentation process. Clasts with vesicularity >75 volume % are fragmented by magmatic exsolution (Head and Wilson, 2003). Fragmentation of clasts with
fewer vesicles may be influenced by magmatic volatiles but there must be an alternate dominant process. The degree to which magma can vesiculate is affected by the rate of cooling (e.g., rapid quenching by external water may inhibit vesiculation). Similarly, the degree of crystallization of a melt is dependent on the rate of cooling. Holocrystalline basalt (E1) is derived from slowly cooled interiors of lava flows, pillow lavas and dikes. Cryptocrystalline tachylite (E2 and E3) and sideromelane glass (E4 and E5) form by rapid quenching of magma. The rate of quenching that produces tachylite is slow enough to permit microlite nucleation and growth. The quenching rate that produces sideromelane is too rapid for microlite nucleation. Clast assemblages that have all three textures may originate from, for examples, disintegration of fresh lava with a range in crystallinity.

2.1.7 Chemical Heterogeneity of Glassy Clasts
The geochemistry of glass clasts can be used to distinguish deposits derived from products of single versus multiple magma sources. Major element oxides (e.g., MgO) and ratios (e.g., TiO2/K2O) are used here to decipher chemically homogeneous versus heterogeneous deposits, and assess possible causes for heterogeneity (see section 2.4 for details). Homogeneous deposits are inferred to be derived from a single magma batch (with uniform composition) over such a brief duration that neither fractionation nor mixing alters the chemistry of the magma between or during the eruptions (F1). Heterogeneous compositions (i.e.,
variable major and minor element concentrations) suggest that the clasts are from either a single eruption of magma with zones of various degrees of fractionation (i.e., cooling temperatures) or mixing, or more than one magma batch and eruption. Incompatible element ratios are used to assess the number of magma batches supplying the volcaniclastic units. Heterogeneous deposits with uniform incompatible-incompatible element ratios may represent a single eruption with various degrees of fractionation (i.e., cooling temperatures) or mixing (F2). Deposits with variable incompatible-incompatible ratios indicate that there must be more than one parental magma batch supplying the clasts (i.e., more than one eruption; F3 and F4).

Figure 2.5 Cartoon of an environment that can lead to deposition of a homogeneous deposit derived from a single source (F1); a deposit of clasts from a single parent magma that exhibit a slight range in composition (F2); a small percentage of chemically distinct clasts from a different source than the bulk of the deposit (F3); and a mixture of two or more major populations of chemically distinct clasts (F4).
A small portion of clasts with different composition can be attributed to limited incorporation of previously erupted material during transport or deposition (F3). There is a range of possibilities that can explain how a deposit on an oceanic island can contain significant subpopulations of chemically distinct clasts. The two extreme interpretations are a major volcanic collapse (e.g. Nu'uanu landslide, O'ahu) and a prolonged erosional period of slowly accumulating lava flows (F4; figure 2.5). On the slopes of the volcano, deposits from a major collapse would be poorly sorted (with a wide range of grain sizes including blocks) with angular clasts (Clague et al., 2002). Deposits accumulated from slow erosion (e.g. wave and current action) of lava flows would be bedded, well sorted (sand to block grade) and the clasts could be significantly rounded depending upon transport distance. Note that a slight range in composition (F2) and a small portion of chemically distinct clasts (F3) could also result from a minor collapse of relatively homogeneous substrate (e.g., an avalanche).

2.1.8 Non-Juvenile Clasts

Non-volcanic material may occur in the deposits. These clasts reflect background sedimentation distinct from volcaniclastic processes and they can be used to constrain the location and rate of sedimentation. The volume percentages of pelagic material within a succession can be expressed either as the proportion of separate beds of pelagic microfossils within a sequence of deposits or as a component mixed within a single deposit. In an environment
where deposition and emplacement of volcaniclastic debris is high, as it was where the deposits of the HSDP2 core accumulated, the abundance of pelagic microfossils (G1-G3) and pelagic muds (G4) is likely to be small. Although the environment and depth of reef limestone development is well constrained, the occurrence of reef limestone clasts in volcaniclastics (G5) can only be used as an indicator of depositional environment under the assumption that the deposition of the volcaniclastics was proximal to that of the reef limestone. The angularity of the reef limestone clasts can be used to access if this is a reasonable assumption, because angular clasts would indicate clasts were deposited close to the reef-forming environment, whereas well-rounded clasts could have been transported far from their environment of production and the depositional environment may be very different. Abrupt large events, such as landslides and volcanic explosions that fragment and eject wall rocks (e.g. Halemaumau eruption, Kilauea, 1924), or prolonged periods of energetic erosion are required to excavate and fragment older, buried deposits (G6).

2.2 Interpretation of transport and depositional processes

The above characteristics of deposits and the associated interpretations can be compiled to determine endmember processes of transport and deposition (Table 2.2). Volcaniclastic deposits are divided into primary, syn-eruptive, and secondary deposits based on the source for the clasts as well as the transport
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<thead>
<tr>
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<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Redeposited pillow breccia</td>
<td>1. Fine grained, well sorted</td>
<td>3. Poorly sorted</td>
<td>11. Planar to curvilinear, fluidal margin</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Redeposited hyaloclastite</td>
<td>3. Poorly sorted</td>
<td>11. Planar to curvilinear, fluidal margin</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Avalanche</td>
<td>11. Planar to curvilinear, fluidal margin</td>
<td></td>
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<td></td>
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<td></td>
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<tr>
<td></td>
<td>Debris flow</td>
<td>11. Planar to curvilinear, fluidal margin</td>
<td></td>
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<td></td>
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<td></td>
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<tr>
<td></td>
<td>Hyperconcentrated flow</td>
<td>11. Planar to curvilinear, fluidal margin</td>
<td></td>
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<td></td>
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<tr>
<td></td>
<td>Concentrated flow</td>
<td>11. Planar to curvilinear, fluidal margin</td>
<td></td>
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<tr>
<td></td>
<td>Turbidity flow - surge type</td>
<td>11. Planar to curvilinear, fluidal margin</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Turbidity flow - steady type</td>
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<td></td>
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<td></td>
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<tr>
<td></td>
<td>Wave-reworke</td>
<td>11. Planar to curvilinear, fluidal margin</td>
<td></td>
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</tbody>
</table>

Table 2.2: Summary chart linking end-member transport and depositional processes with key deposit characteristics.
<table>
<thead>
<tr>
<th>D: Secondary features</th>
<th>E: Juvenile clast groundmass textures</th>
<th>F: Chemical heterogeneity of melt</th>
<th>G: Nonjuvenile clasts</th>
</tr>
</thead>
<tbody>
<tr>
<td>2. High temp alteration</td>
<td>2. Tachylyte: mod to high ves</td>
<td>2. Slight compositional</td>
<td>1. Interbedded pelagic muds</td>
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<td>3. Low temp alteration</td>
<td>3. Tachylyte: none to low ves</td>
<td>range</td>
<td></td>
</tr>
<tr>
<td></td>
<td>4. Sideromelane: mod to high ves</td>
<td>3. Small % of chemically</td>
<td></td>
</tr>
<tr>
<td></td>
<td>5. Sideromelane: none to low ves</td>
<td>distinct clasts</td>
<td></td>
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<tr>
<td>Syn-eruptive deposits</td>
<td>In situ pillow breccia</td>
<td>4. &gt;1 major population of</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Redeposited pillow breccia</td>
<td>chemically distinct clasts</td>
<td></td>
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<td></td>
<td>Redeposited hyaloclastite</td>
<td>5. Reef limestone</td>
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<td>secondary deposits</td>
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<td>6. Reworked fragments of</td>
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<td></td>
<td>Peperite</td>
<td>pre-existing rock</td>
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<td>Subaqueous fallout</td>
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<tr>
<td>Pyroclastic density</td>
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<td>?</td>
<td>?</td>
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<tr>
<td>current - surge</td>
<td>?</td>
<td>?</td>
<td>?</td>
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<tr>
<td>Pyroclastic density</td>
<td>?</td>
<td>?</td>
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<td>Hyperconcentrated flow</td>
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<tr>
<td>Concentrated flow</td>
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<tr>
<td>Turbidity flow - surge</td>
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<tr>
<td>Wave-rewarded</td>
<td>?</td>
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<td>?</td>
</tr>
</tbody>
</table>

51
history of clasts between source and deposition. Subdivisions are structured according to types of transport and depositional processes.

**Primary deposits**

These deposits record only one transport process of clasts directly from the source to the final depositional environment (e.g., subaqueous pyroclastic fallout from explosive activity at the vent or tumbling from disintegration of lava as it enters the sea). They include deposits transported very limited distances, such as in situ pillow lava breccias and peperites.

Material deposited by fallout may show bedding, particularly if the eruption was pulsatory. Various factors, such as eruption intensity, proximity to vent, frequency of pulses and wind strength/direction, could influence whether the beds contain similar grain size, weak fluctuations in grain size, or coarse-fine couplets. In a subaqueous setting, gradational vertical changes in grain size and good sorting are expected due to the more pronounced fractionation of clasts during settling. Bedding should be planar and mantle the pre-existing topography. Reverse size-grading can be produced by density fractionation of clasts (e.g., non-vesicular lithics and pumice). The basal contact of the deposit will be non-erosive and both lower and upper contacts will be parallel with the pre-existing topography. Clasts suffer little attrition during plume and fallout
processes and accordingly clast shapes reflect the original mechanism of fragmentation.

Pyroclastic surges have low solid concentrations, are density-stratified, and are dominantly turbulence-supported. Due to the low solid concentrations, subaerial surges flowing towards the ocean usually propagate across the water surface rather than continuing downslope. Material then settles through the water column and forms well sorted, normally graded deposits that mantle the ocean floor. Therefore, it is difficult to distinguish between subaqueous fallout and surge deposits. Clasts are well rounded to angular depending upon the amount of attrition during the subaerial portion of the transport (e.g., Heiken and Wohletz, 1992).

Pyroclastic flows have higher solid concentrations and, therefore, may be able to continue downslope upon crossing the coastline (although increased lofting is expected). If flows cross the water surface, they may transform into a water-supported density current, such as a debris flow. Flows have a plug-like velocity profile with higher velocities at the base and they can carry a range of clast sizes from ash to blocks. Thus, their deposits are coarse grained, poorly sorted, and bedding may show variations in grain size. The high solid-fluid ratio and rapid deposition accounts for the massive and ungraded nature of deposits. Pyroclastic flows are controlled by gravity to such a degree that deposits are
generally confined to topographic lows and these ground-hugging conditions allow extensive erosion of the pre-existing topography. Upper contacts tend to be flat. Attrition is likely to be more prevalent than in pyroclastic surges because of the higher solid-fluid ratio. Thus, clasts are more likely to be rounded. Both types of pyroclastic density currents are produced by explosive eruptions and are likely to contain wall rock fragments and have a range of juvenile groundmass textures.

In situ primary deposits include peperites, pillow breccias and hyaloclastites. The clasts in these deposits do not undergo any significant transport except minimal rotation of clasts in the pillow breccia and hyaloclastite cases. This lack of transport and almost continuous supply of clasts inhibits size fractionation and bedding, so deposits are nonbedded, massive and ungraded. The basal contact is non-erosive (and reworked fragments of pre-existing rock will be absent), but whether the upper contact is flat or irregular depends upon the shape of the pile of accumulating material, the shape of the pre-existing topography, and the number and juxtaposition of sources. Pillow breccias will be poorly sorted with variable clast shapes and juvenile groundmass textures ranging from fine, shard-like sideromelane from the rinds of pillows, to coarse, blocky cryptocrystalline basalt from pillow lava interiors (e.g., Carlisle, 1963). Grain sizes and clast shapes of in situ hyaloclastites depend upon the fragmentation mechanism, but they are usually fine grained (ash size) curved shards or angular equant shapes.
with irregular or planar margins depending upon vesiculation. These deposits are glassy and are more likely to be dominated by sideromelane textures rather than tachylite.

Peperites are formed by quenched-induced disruption of magma or molten lava upon contact with water-logged sediments. Their characteristic features and origins are much more complicated and varied than in situ pillow breccias and hyaloclastites (see Skilling et al., 2002, for details). This outline therefore simplifies the expected characteristics of peperites. Excluding any pre-existing features, peperites are nonbedded and massive. The grading may be complicated by the shape of the lava/sediment contact. Contacts obviously depend upon the shape of the intrusion and on whether the magma is wholly subsurface or flowing over and/or into pre-existing material. Grain size and sorting depends upon the nature of the host material and on the efficiency of the fragmentation, but peperite is likely to be poorly sorted. Juvenile clast shapes are normally irregular or planar with angular or fluidal margins depending on the rheology of the magma at the time of fragmentation, the type of host rock, and the vesicularity of the magma. In all cases, jigsaw-fit patterns should be present amongst some of the fragments (Skilling et al., 2002).

**Syn-eruptive deposits**

Syn-eruptive deposits consist of juvenile material (no reworking of pre-existing material) that has briefly come to rest before continuing in a different transport
process. For example, clasts from a disrupted pillow that fall to the ocean bed (in situ pillow breccia) before continuing to move downslope in a debris flow (redeposited pillow breccia). These deposits will be difficult to recognize, however, close association with primary deposits derived from the same source may indicate overall proximity to an active source. For example, a hyaloclastite deposited by a turbidity current may be overlain by an in-situ hyaloclastite from the same lava flow.

**Secondary deposits**

These epiclastic deposits have clasts with more heterogeneous chemistry and they often contain more than one clast lithology. They have a higher probability of containing reworked fragments of pre-existing lava/sediment and, in the strict senses, lack juvenile material. Secondary deposits are transported by a range of events from dilute density currents to debris avalanches and rock falls. Debris avalanches are high density, gravity-controlled events that are generally confined to topographic lows and associated deposits are nonbedded, massive and poorly sorted. Depending on the runout distance and velocity (i.e., the opportunity for clasts to form and become size fractionated) they may be clast or matrix-supported. They usually have large blocks and may have an irregular top. Despite high attrition rates, the typical short transport distances mean clasts would not become well rounded.
Unless otherwise referenced, the following classification of density flows is after Mulder and Alexander (2001). Principal divisions are cohesive flows, hyperconcentrated density flows, concentrated density flows and turbidity flows. Cohesive flows have high particle concentration (35 to 95%), are held together by matrix strength from cohesion between the grains, behave pseudoplastically, and are divided into debris flows (poorly sorted) and mudflows (<5 vol % gravel; mud:sand ratio >1). Mudflows are further divided into silty mudflows (<25% clay) and clay-rich mudflows (>40% clay). Hydroplanning allows cohesive flows to travel long distances at high velocities without eroding at their base. Deposition occurs en masse when the shear force drops below gravity, leaving a chaotic deposit with a large range of particle sizes, shapes and compositions. Shear structures may form at the base and margins of the flow due to shear flow, and in the middle due to flow surging (e.g. layers of coarse material and imbrication).

The other divisions of density flows are non-cohesive and they are held together by frictional forces between discrete grains. Hyperconcentrated flows can have a similarly high solid-fluid ratio to cohesive flows but they do not behave plastically. Hyperconcentrated flows have the highest solid-fluid ratio (40 to 70%) of the frictional density flows and are supported primarily by grain-grain interaction, whereas turbidity flows have the lowest solid-fluid ratio (<9%) and are supported dominantly by turbulence. Concentrated flows, which have intermediate solid-fluid ratio (10 to 30%), are dominated by grain-grain support but turbulence
support may prevail at the head and upper parts of the flow. Concentrated flows differ from hyperconcentrated flows because they are Newtonian and their lower solid concentrations allows particle sorting via settling during the passage of flow. Turbidity flows are also Newtonian but clasts are dominantly supported by turbulence and have solid concentrations less than 9% (Bagnolds limit for turbulent suspension).

Frictional freezing (grain-grain interaction) is responsible for deposition from hyperconcentrated flows. Accumulation may occur from accretion of successive small-scale flow pulses. Deposits could include rafts and large clasts, and local reverse grading from traction carpets created by upward velocity gradients and laminar flow, but little or no normal grading due to a lack of suspension fallout. A subset of hyperconcentrated flows are grain flows, which deposit well sorted silt, sand or gravel grade material. Hyperconcentrated flows can transform into cohesive flows by dewatering or bulking-up (e.g., by deceleration), and into concentrated flows by water entrainment and increased turbulence by a slope change. The latter transformation is more likely to occur in subaqueous environments.

Concentrated flows are more internally varied as they have traction loads at the base (high shear; C>9%) and turbulent, with sometimes decoupled, low sediment concentrations (C<9%) at the top and head. Hydroplaning does not occur
because the lower concentrations of solids permit dissipation of pore pressure upwards through the flow. Basal erosion can be pronounced and inclusions entrained into the flow during transport can account for >80% of the clast types. Highly concentrated basal layers can display inverse grading created by traction carpets, sediment supply variations, or pulsing flows. Weakly concentrated basal layers can undergo traction that with time may develop bedforms such as ripples, dunes and antidunes. The body of concentrated flows deposits massive sand and gravel, and material in the more turbulent upper part and head settle out from suspension to form Bouma Tc-e or Td-e facies. As particle concentrations decrease, concentrated flows can change into turbidity flows.

Turbidity flows are subdivided by the duration of constant velocity into surges (very short durations; uncommon), surge-like turbidity flows (short durations; uncommon), and quasi-steady turbidity currents (long durations; common). Deposit characteristics of surge and surge-like turbidity flows are largely controlled by the behavior of the flow head, whereas those of the quasi-steady turbidity flow are controlled by the body. Surges essentially consist of a head and a dilute tail because material from the body feeds into, and is largely deposited by, the head. Similarly, surge-like flows have only a short body. Surges and surge-like flows carry fine material (less than sand size) in suspension and settling along with lateral movement creates typical Bouma Tb-d structures and bedforms. Basal erosion will only occur upon acceleration (e.g.,
increase in slope). Due to the short durations, beds are thin. Quasi-steady turbidity currents are formed by hyperpycnal flows at river mouths and deposits may be recognized by a basal inversely graded unit overlain by a normally graded unit separated by a sharp contact. The most common bedforms are climbing ripples.

Wave or storm deposits may include cross-bedding or ripples (Collinson and Thompson, 1982). Their basal contact is highly erosive. The upper contact is commonly rippled and clasts are well-rounded provided that exposure to wave action has been of a sufficient duration. Non-juvenile material, such as reef fragments, are likely to be present.

2.3 Clast Shape Analyses

2.3.1 Introduction and Previous Work

It has long been recognized that the shapes of clasts reflect the mechanisms of fragmentation that created and modified them (e.g. Wohletz, 1983; Sheridan and Wohletz, 1983). Primary fragmentation can be brittle, ductile or viscous depending on the rheology (viscosity, surface tension and yield strength) of the melt (Wohletz, 1983). Fuel-coolant experiments using thermite (fuel) and water (coolant) by Wohletz (1983) examined the relationship between clast shapes and fragmentation mechanisms (including the physical properties of the melt and heat
energy release rates) in explosive water to melt interactions. Five characteristic clast shapes were attributed to different fragmentation mechanisms (table 2.3).

Field based studies, particularly at Gran Canaria and Iceland as well on fragmental rocks from the seafloor, have described and used particle shapes as an indicator of primary fragmentation mechanisms and influences of secondary processes (e.g., Schmincke and Sumitra, 1998; Batiza et al., 1984). The majority of these studies are qualitative and base interpretations on angularity (planarity) versus rounding (convexity) of clasts combined with vesicularity and crystallinity. For example, Batiza et al., (1984) interpreted small flakes and jigsaw-fit shards as arising from in-situ shattering of larger glass fragments. Glassy globules, which can crack into wedge and shell shaped splinters, and angular glass clasts are interpreted to form via quench fragmentation (Kokelaar, 1986).

Table 2.3 Magma-water fragmentation mechanisms and characteristic clasts shapes that are produced (after Wohletz, 1986).

<table>
<thead>
<tr>
<th>Fragmentation mechanism</th>
<th>Shapes</th>
<th>Process</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stress wave fragmentation</td>
<td>Blocky and equant with curvi-planar surfaces</td>
<td>Brittle fracturing and quenching</td>
</tr>
<tr>
<td></td>
<td>Irregular. Shape controlled by vesicles. Smooth, fused surfaces</td>
<td>Fracturing followed by turbulent mixing of melt and water. Quenching of melt.</td>
</tr>
<tr>
<td>Fluid instability fragmentation</td>
<td>Convoluted and moss-like fluidal fine ash</td>
<td>Turbulent mixing of melt in water at irregular contact (high surface area). Melt deforms viscously.</td>
</tr>
<tr>
<td></td>
<td>Spherical and drop-like fluidal fine ash</td>
<td>Turbulent mixing of melt and water at irregular contact. Surface tension affected the melt.</td>
</tr>
<tr>
<td>Vesculation-stress wave fragmentation</td>
<td>Plate-like ash &lt;100um</td>
<td>Strongly vesicular melt. Almost simultaneous bursting of vesicles and melt/water fragmentation.</td>
</tr>
</tbody>
</table>
Fractal Dimension is found by measuring the clast perimeter with varying measuring units ('step-lengths'). Therefore, finer irregularities will only be detected by smaller step-lengths. When this data is plotted onto a Richardson plot (log step length, L, vs log perimeter, P) the slope yields the tortuosity of the particle. The fractal dimension (D; 1 - slope of Richardson plot) ranges from 1.0, which is planar in shape, to 2.0, which is highly irregular. A clast that produces a single, linear slope and has a low D has smooth margins, whereas a clast with multiple slopes will have irregular margins. Fractal analyses has been used to quantify clast shapes and decipher the fragmentation mechanism of clasts deposited during the submarine stage of Gran Canaria (Carey et al., 1998). Their aim was to distinguish clasts with simple margins (representing deep-water volcanism) from those with complex margins (shallow-water volcanism) based on the principle that vesicularity is a major cause of complex margins. Carey et al., (1998) found a distinction could be made between non-explosive and explosive fragmentation mechanisms, but more specific classification was not possible.

2.3.2 Procedures

Three-dimensional shape analyses was not possible because much of the core is too well indurated to extract individual clasts. Therefore, thin sections were scanned at 1200 dpi and the images were adjusted in Adobe Photoshop to enhance the contrast between clasts and matrix. In Photoshop, clasts were outlined manually, filled black, and saved as a binary tiff image with a white
background. Scion Image (developed at the U.S. National Institute of Health and available on the Internet at http://rsb.info.nih.gov/nih-image/) was used to analyze the area, perimeter, maximum length and minimum length of clasts in the binary tiff images. The maximum and minimum lengths are that of a best-fitting ellipse. Matlab was used to analyze the convex area of clasts, which is the area within an imaginary boundary constructed around the clast as if it were a taut string.

2.3.3 What parameters were used and why

The aim of this part of the study was to survey quantitative procedures for analyzing clast shapes and assess whether resulting data could be useful tools for identifying mechanisms of fragmentation and degree of reworking. Fractal analyses has already been shown by Carey et al. (1998) to yield a first order distinction between rough (vesicular) and smooth (non-vesicular) clasts. This method was not explored in this study because the population of clasts fragmented purely by magmatic volatiles (i.e., the fragmentation mechanism that can be recognized by vesicularity contents) is expected to be minor to none. Therefore, more information of shape and surface texture is needed than just vesicularity. Four parameters were calculated from the measurements taken in this study:

- Aspect Ratio = Maximum length / Minimum length
- Formfactor = (4 * π * Area) / Perimeter²
- Solidity = Area / Convex Area where Convex Area is the area within a string
that is pulled taught around the clast perimeter

Roundness = \((4 \cdot \text{Area}) / (\pi \cdot \text{Maximum length}^2)\)

In addition, deformation factor \((\text{maximum length} - \text{minimum length} / \text{[minimum length / maximum length]})\) was calculated, however, it will not be discussed because it does not appear to yield any further information than the aspect ratio and correlations of these two parameters are positive. Aspect ratios are useful for determining the proportion of shards and elongate, fluidal clasts within a deposit. Pillow breccias will show a range of aspect ratios due to radial fracturing, which creates equant clasts near the margins and wedge shaped clasts closer to the center.

Formfactor is an important indicator of clast margin roughness but it cannot distinguish shape. Clasts with low formfactor have rough margins and those with high formfactor have smooth margins. Usually, roughness is a proxy for vesicularity and therefore the gas-magma ratio at the time of solidification. A high gas-magma ratio would indicate magmatic explosivity played an important role in fragmentation. An additional assessment of roughness is solidity, which is a measure of the total size of embayments. This would be far more informative in combination with complexity (which compares the true and convex clast perimeters) because they can be used together to find the size and number of embayments along clast margins. Roundness defines shape by the deviation of the clast shape from spherical.
2.3.4 End-member samples

Three samples from Kilauea volcano of known origin (Figure 2.6) were analyzed for comparison to the deposits in the HSDP2 core in order to facilitate interpretations by direct comparison. The samples chosen represent three extreme end member processes of clast fragmentation on Hawaiian volcanoes:

*Primary Magmatic Explosive Deposit: Kilauea Iki 1959 Eruption*

In 1959, a series of 17 Hawaiian fire fountaining episodes occurred over a 5 week period at Kilauea Iki, Hawai‘i. Of the ~102 million m$^3$ lava that was erupted about 64 million m$^3$ drained back into the reservoir; the rest remained in a lava lake that had filled the adjacent crater (Richter et al., 1970). Syn-eruption drainback of lava allowed recycling of the fire fountain ejecta that fell onto the lake both during and between episodes. Sample 03_KiL25_10 is taken from the final episodes of the eruption, which had the highest fire fountains and accordingly produced the most vesicular deposits. Therefore, this sample represents a subaerial primary fall deposit produced by an explosive magmatic basalt eruption.

*Reworked Beach Deposit*

The HL98 beach deposit was collected from a black sand beach formed from the 1955 Kilauea lava flow. The 1955 eruption occurred from the East Rift Zone in the eastern Puna district (Macdonald and Eaton, 1964). It lasted for 3 months and consisted of
Figure 2.6 Photographs and corresponding binary images of clasts fragmented or modified by a) primary magmatic explosivity; b) explosive and passive magma-water interaction in a littoral to shallow submarine environment; c) erosive wave and current action at a beach.

a) Subaerial, primary, magmatic explosivity deposit

b) Littoral/shallow submarine, primary, phreatomagmatic deposit

c) Littoral/shallow submarine, secondary, epiclastic deposit
intermittent lava fountains and 'a'a and pahoehoe flows. Lava began entering the ocean after one month and continued until the end of the eruption. Some phreatomagmatic eruptions occurred, however, predominantly the lava continued along the ocean bed as intact lava flows. These lava flows have since been subjected to erosion by waves. The sample consists of well sorted, subrounded fragments that now make up black sand beaches. The HL98 sample is collected from one of these beaches that lie on top of a 1955 lava flow.

**Shallow Submarine Deposit**

Two samples from the ongoing Pu'u 'O'o eruption were collected by Tribble et al (1991) at approximately 25 mbsl on the steep (25° to 40°; angle of repose) littoral slope where the lavas have been entering the ocean routinely since December 1986. During this eruption, lavas have entered the ocean in both a passive and an explosive manner (e.g., Moore et al., 1973 and Mattox and Mangan, 1997). Therefore, this deposit contains a wider variety of clast shapes.

**2.4 Electron Microprobe Analyses**

Major element and sulfur contents of glasses were analyzed using a Cameca SX five spectrometer electron microprobe at the University of Hawai'i. A beam diameter of 10 µm (x15,000) was used with a current of 10 nA and 15 kV accelerating voltage. Peak counting times were 90 seconds for Ca, Ti, Mn and Fe, 60 seconds for P, S and K, 45 seconds for Mg, Al, and Si, and 40 seconds
for Na (counted in the first round of a two part analysis to reduce Na volatilization). Background counting times were half the peak counting times except for Mg, Al and Si, which were counted for 25 seconds. Natural glass and mineral standards (VG2, A99, OR-1 and Apatite) were used for calibration. Internal standards of natural glasses (A99 and VG2; Jarosewich et al., 1979) were used for minor adjustments to the calibration (<1%) for some of the analyses. PAP-ZAF corrections were applied to all analyses. Analytical errors based on 95% confidence are within 0.1 wt% for Na, P, S, K and Mn, 0.2 wt% for Mg, Al, Ca and Ti, and 0.3 wt% for Si and Fe.

Major element plots were used to decipher homogeneous assemblages representing a single magma batch from heterogeneous assemblages that vary in composition due to fractionation or different parental magmas. Glassy clasts with similar MgO contents are inferred to be homogeneous and those with variable MgO are heterogeneous. Incompatible element ratios (e.g., TiO$_2$/K$_2$O) are used to assess the number of magma batches supplying the heterogeneous volcanioclastic units. Whole rock XRF analyses of small individual eruptions at Kilauea volcano since 1820 (~0.1 km$^3$) vary by TiO$_2$/K$_2$O ~0.2. This implies that ranges less than 0.2 probably represent a single magma batch. However, the average total error for glass TiO$_2$/K$_2$O is 0.38. Thus, ranges of TiO$_2$/K$_2$O ratios greater than total errors are assumed to be too significant for a single magma batch to have supplied the clasts. Therefore, glassy clasts with heterogeneous
major element compositions and TiO₂/K₂O values within a range of 0.38 are probably derived from a single magma batch that underwent variable degrees of fractionation. Assemblages of glassy clasts with TiO₂/K₂O values varying greater than 0.38 are inferred to be from more than one separate magma batches.
3. RESULTS

Eleven HSDP2 units were chosen from the submarine core section of Mauna Kea spanning depths from ~1300 to ~2460 mbsl to characterize the volcaniclastic deposits formed during the submarine growth during its subaerial shield stage. Eight units are from the upper submarine section that is dominated by fragmental lithologies (1,079 to 1,984 mbsl) and three are from the underlying pillow lava-dominated deep submarine section (1,984 to 3,098 mbsl; figure 1.3). Volcaniclastic units 297, 286, 261, 260, 238, 202 and 194 were chosen because they are thick (20.9 to 70.9 m) and primarily massive. Thus, they represent volumetrically-dominant eruptive and depositional conditions. In addition, bedded and more structurally heterogeneous units 294, 293, 275, and 274 were selected because they occur at the base and just above of the thickest pillow lava packages in the HSDP2 core. Descriptions of all eleven units are summarized in table 3.1. Due to overlapping characteristic features, only six units (294, 293, 286, 260, 202 and 194) are described in detail. Units 294, 286 and 260 are thought to reflect the dominant processes of formation of the five units that are not described in detail (297, 275, 274, 261 and 238): Units 294, 274, and 275 (6.3 to 8.2 m) are bedded and matrix-supported volcaniclastics with significant grain size fluctuations (maximum clast sizes range from 1 to 16 cm). Units 275 and 294 have alternating coarse (fine sand) and fine (silt) layers. Some silt layers show cross-bedding, pinch-and-swell, and load structures. An
important difference from unit 294 is unit 274 overlain by a lithologically distinct volcaniclastic unit, whereas unit 294 is depositionally overlain by pillow lava. Units 286 and 260, like units 297, 261 and 238, are thick (~21 to 49 m) volcaniclastics consisting mainly of poorly sorted, coarse (~2 to 42 cm) angular clasts. Although these units are principally massive, they contain some moderately sorted sections of medium lapilli and rare bedding.

Units 294 and 293 are the deepest (2,457 to 2,459 mbsl) and most heterogeneous units examined, representing distinctive lithofacies and emplacement modes. These units also document the depositional regime at the base of the thickest package of basalt flows (~220 m) in the submarine section. It is unlikely that these units represent a long-lived stable depositional regime because they are interbedded with thin silt layers and contain some fluctuations in grain size and clast lithology. The base of unit 293 includes isolated pillow lavas in a fine glassy matrix. Unit 294 is a bedded sequence of ash to fine lapilli volcaniclastics. Unit 286 (2,185 to 2,193 mbsl) is representative of the thicker volcaniclastic deposits within the pillow-lava-dominated deeper submarine section. It is a coarse lapilli breccia with intervals of glassy fine lapilli. In the upper part of the HSDP2 submarine section, three units highlight significant changes within the core. Units 260 (1,828 to 1,836 mbsl), 202 (1,428 to 1,437 mbsl) and 194 (1,293 to 1,298 mbsl) were chosen because they represent the lower, middle and upper parts of the volcaniclastic-dominated submarine section.
Table 3.1 Summary of sedimentary, petrographic and geochemical characteristics of representative HSDP2 volcaniclastic units.

Columns are divided into categories after Table 2.1. Bedding planes (A3), non-juvenile (G) columns, and some subcharacteristics (e.g., beds of similar grain size, A1.2) are omitted because they don't occur or can't be assessed in the HSDP2 core.

<table>
<thead>
<tr>
<th>A1</th>
<th>A2</th>
<th>A4</th>
<th>A5</th>
<th>A6</th>
<th>C</th>
<th>D</th>
<th>E</th>
<th>F</th>
</tr>
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<tbody>
<tr>
<td>Bedding</td>
<td>Structures</td>
<td>Grading</td>
<td>Upper contact</td>
<td>Grain size/ sorting &amp; average grain size (cm)*</td>
<td>Clast shapes and margins</td>
<td>Secondary features</td>
<td>Juvenile clast groundmass textures</td>
<td>Chemical heterogeneity of melt</td>
</tr>
<tr>
<td>X = Nonbedded (A1.1);</td>
<td>M = Massive (A2.1);</td>
<td>NE = Non-erosive (A5.1);</td>
<td>F = Flat (A6.1);</td>
<td>FW = Fine grained, well sorted (B.1);</td>
<td>Curved shard (C.3);</td>
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<tr>
<td>FG = Fluctuations in grain size (A1.3);</td>
<td>AB = Cross-bedding/ripples etc (A2.2);</td>
<td>E = Erosive (A5.2);</td>
<td>I = Irregular (A6.2);</td>
<td>p = Poorly sorted (B.3);</td>
<td>Blocky, wedge-shapes (C.4);</td>
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<tr>
<td>GFC = Coarse-fine couplets (A1.4);</td>
<td>PB = Planar bedding/lamination (A2.3);</td>
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<td></td>
<td>Irregular, angular (C.5);</td>
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<td>Planar to curvilinear, angular (C.7);</td>
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<td>Planar to curvilinear, fluidal (C.8);</td>
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<td>Low temperature alteration (D.3);</td>
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<td></td>
<td>Crystalline basalt (E.1);</td>
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<td>Tachylite, mod to high ves (E.2);</td>
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<td>Tachylite, none to low ves (E.3);</td>
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<td>Sideromelane, none to low ves (E.4);</td>
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<td>Sideromelane, none to low ves (E.5);</td>
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<td>194</td>
<td>23.8</td>
<td>M</td>
<td>X</td>
<td>NE</td>
<td>I</td>
<td>P</td>
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<td>R</td>
<td>NE</td>
<td>I</td>
<td>P</td>
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<td>FG</td>
<td>PB</td>
<td>N/R</td>
<td>NE</td>
<td>F</td>
<td>P</td>
<td>5.4</td>
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<tr>
<td>261</td>
<td>31.4</td>
<td>X</td>
<td>PB</td>
<td>N/R</td>
<td>NE</td>
<td>F</td>
<td>P</td>
<td>7.4</td>
</tr>
<tr>
<td>274</td>
<td>7.1</td>
<td>FG</td>
<td>PB</td>
<td>N</td>
<td>NE</td>
<td>F</td>
<td>P</td>
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</tr>
<tr>
<td>275</td>
<td>8.2</td>
<td>FG/CFC</td>
<td>XB</td>
<td>N</td>
<td>NE</td>
<td>I</td>
<td>FW/P</td>
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<tr>
<td>286</td>
<td>42.1</td>
<td>FG</td>
<td>M</td>
<td>N</td>
<td>NE</td>
<td>I</td>
<td>P</td>
<td>5.7</td>
</tr>
<tr>
<td>293</td>
<td>98.4</td>
<td>X</td>
<td>M</td>
<td>N</td>
<td>NE</td>
<td>I</td>
<td>FW</td>
<td>10.6</td>
</tr>
<tr>
<td>294</td>
<td>6.3</td>
<td>FG</td>
<td>XB</td>
<td>N</td>
<td>NE</td>
<td>I</td>
<td>FW/P</td>
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<tr>
<td>297</td>
<td>20.9</td>
<td>FG</td>
<td>PB</td>
<td>X</td>
<td>NE</td>
<td>I</td>
<td>P</td>
<td>6.6</td>
</tr>
</tbody>
</table>

*Average of 10 largest grain sizes except units 189 and 293, which is average of 6 largest grain sizes.

72
Unit 260 is a coarse lapilli breccia with five silty intervals. Unit 202 is the thickest HSDP2 volcaniclastic unit (~71 m) and is remarkably monotonous. It contains coarse lapilli fragments, some of which have fluidal shapes, in a matrix of glassy lapilli. Unit 194 is a massive lapilli breccia consisting of coarse lapilli and blocks of cryptocrystalline to hyaline basalt resting in a matrix of angular sideromelane ash to fine lapilli.

3.1 Unit Descriptions

Macroscopic and microscopic features are described for units 295, 294, 286, 260, 202, and 194 in stratigraphic order (i.e., from the deepest to the shallowest unit). Stratigraphic logs, photographs, and photomicrographs (i.e., thin section scans) are provided for each unit (c/o figures 3.1 to 3.13).

3.1.1 Unit 294

Macroscopic features

Unit 294 is a 6.3 m thick, bedded basaltic volcaniclastic sandwiched between two pillow lava units, although the upper pillow lava grades into an isolated pillow breccia at its base. The basal contact is depositional onto a pillow lava (unit 295) and at the upper contact, unit 294 is normally graded and capped by a pillow lava lobe from the isolated pillow breccia of unit 293. The top 66 cm of unit 294 consists of a massive bed of fine basaltic lapilli tuff (~35 cm thick) overlain by two coarse tuff beds (~5 cm and ~25 cm; figure 3.1). The fine lapilli section is
moderately sorted and contains basalt clasts mostly <4 mm (one clast is ~8 mm) in a fine-grained brown matrix. The top contact of this bed is sharp and depositional. The overlying bed is a thin (~3 cm) coarse tuff of the same material. The uppermost coarse tuff contains fine intervals that are slightly inclined (dip is <4°). These fine intervals are calcium carbonate bearing (Seaman et al., 2000) the beds are subparallel and discontinuous, with millimeter-scale flame structures along some of the upper contacts.

**Microscopic features**

Incipiently vesicular (5 to 20% vesicles) sideromelane is the dominant clast type of the fine lapilli bed. Olivine microphenocrysts (<5 to 10%; <0.2 mm) include glass and spinel inclusions and there are rare olivine microlites (<0.05 mm). Rare clasts are crowded with ~50% plagioclase microlites (<0.01 mm). Vesicles are round and <0.3 mm diameter. The majority (>95%) of clasts are equant and others are irregular due to vesicle boundaries. Vesicular clasts have smooth, scalloped margins and non-vesicular clasts have planar to curviplanar margins.

In-situ cracking of sideromelane has formed planar, angular, equant clasts. Tachylite clasts (<5%) are brown to black in color with irregular margins. The contact between the fine lapilli bed and the overlying coarse ash bed is gradational. Clast sizes grade from ~2.5 mm to ~0.25 mm. The coarse tuff contains higher proportions of tachylite (10 to 20%) and clusters of incipiently to poorly vesicular clasts (10 to 40% vesicles).
Figure 3.1 Stratigraphic log of unit 294 and the base of unit 293 (left) with detailed section of the contact (right). Black boxes show locations of samples taken for thin sections.

- Lava flow with pipe vesicles.
- Top of unit 293 is at 2,359 mbsl.
- Pillow glassy lapilli tuff.
- Thin section photo (Figure 3.3b i).
- Massive, glassy fine lapilli tuff.
- Isolated pillow lava breccia.
- Pillow, fractured glassy margins.
- Thin section photo (Figure 3.3a ii).
- Gradational contacts.
- Coarse tuff.
- Silty intervals.
- Coarse lapilli.
- Lapilli tuff.
- Pillow lava.
- Clast size (cm).
- Graded contacts.
- Core depth (mbsl).
- 2,457.7 mbsl.
- 2,458.8 mbsl.

Top of unit 293 is at 2,359 mbsl.

Lava flow with pipe vesicles.

Figure 3.3a i; major element analyses.

Pillows, fractured glassy margins.

Isolated pillow lava breccia.

Glassy lapilli tuff.

Thin section photo (Figure 3.3a ii).

UNIT 293

Silty intervals

Coarse tuff

Gradational contacts

Thin section photo (Figure 3.3b ii)

Massive, glassy fine lapilli tuff

Figure 3.3b i; major element analysis.
Figure 3.2 Photographs of deposits: a) unit 294 and b) unit 293. Scale bars are 5 cm and are subdivided into cm segments. Core height is in mbsl. Black boxes mark thin section locations.
Figure 3.3 Photomicrographs of thin sections. Scale bars are 10 mm.

a) Unit 294. Thin section 'i' is from the fine lapilli tuff and thin section 'ii' shows the gradational contact to the overlying coarse tuff.

b) Unit 293. Thin section 'i' shows a glassy pillow lava margin and glassy fragments broken off the pillow lava rind. Thin section 'ii' shows a more altered area of the volcaniclastite, which includes a scoriaceous tachylite clast with an altered olivine phenocryst.
Although larger sideromelane clast and vesicle margins have only thin palagonite rims (<0.03 mm), the matrix is extensively palagonitized. Vesicles are lined with pale-green smectite and filled with either chabazite or smectite. Layers of smectite around tachylite clasts have irregular outer margins. Chabazite and phillipsite crystals occur in larger pore spaces, and smectite fills the remaining space between the palagonitized matrix.

3.1.2 Unit 293

*Macroscopic features*

Unit 293 is a ~98 m thick olivine-phyric basalt (Seaman et al., 2000) with a basal 36 cm thick isolated pillow breccia. A gradational and depositional contact at the base of this breccia marks the top of unit 294, which is a bedded volcaniclastite (section 3.1.1). There are 277 internal flow boundaries (i.e., glassy pillow margins) within the lava flow of unit 293. At the top of unit 293, a glassy contact marks the base of a ~30 m thick basalt flow (unit 292).

The isolated pillow lava breccia at the base of unit 293 consists of a few vesicular pillow lava lobes (>10 cm) and rare coarse lapilli sitting in a normally graded (medium lapilli to coarse ash) matrix of glassy clasts. The pillow lava lobes are aphyric, poorly vesicular basalt with glassy margins that are broken into elongate shapes parallel to the edge of the pillow (figure 3.3b-I). The pillow lobes float within the matrix or are linked to the basal pillow lava of the overlying flow.
**Microscopic features**

Vesicles in the isolated pillow lobes are ~25%, round, <1 mm, and are usually filled with zeolites. The overlying flow has unevenly distributed vesicles (~3% in the basal ~2.5 m, and 1 to 16% in the remaining flow; Seaman et al., 2000) with irregular shapes, and some pipe vesicles. The matrix consists primarily of incipiently vesicular, subangular sideromelane clasts with planar to curvilinear margins and some jigsaw-fit patterns. Holocrystalline fragments are absent, and ash-sized cryptocrystalline and tachylite fragments are sparse (~5%). There is one lapilli sized (~1.5 cm) tachylite clast. Sideromelane clasts near the pillow margin are commonly elongate and contain 5 to 20% olivine microlites (~0.01 mm) and microphenocrysts (<0.75 mm) with spinel and glass inclusions, and <5% vesicles (~3 mm). These clasts are larger (~5 mm) and fresher than the majority of sideromelane clasts (~1.5 mm) in the matrix. Smaller sideromelane clasts contain olivine microlites (1 to 10%; <0.08 mm) and rare tiny plagioclase crystals (<1%). Due to the relatively large vesicle sizes (<0.5 mm) compared to clast sizes (<1 mm), estimates of vesicularity may be skewed, but range from 3 to 40%. Consequently, small shards (<0.08 mm) are often blade-like and larger clasts (~1.5; equant) have scalloped margins. Tachylite clasts are moderately vesicular (40 to 50%; <0.7 mm size vesicles) and have smooth scalloped margins. Olivine occurs as microphenocrysts (<10%; <0.8 mm) and as a phenocryst within the lapilli tachylite clast (figure 3.3b-i).
Palagonitization (including fibrous palagonite) of sideromelane clasts is extensive and some clasts are completely replaced by palagonite. However, the olivines in all the sideromelane clasts are fresh, regardless of the degree of palagonitization. Brown smectite layers and patches of pale-green smectite occur along some clast edges and in cracks. Vesicles within both sideromelane and tachylite have geopetal smectite globules and are filled with bladed phillipsite and chabazite or smectite. Microbe tubules propagate from some vesicles and clast margins within the pillow margin. Pores in the matrix are filled with chabazite.

3.1.3 Unit 286

Macroscopic features

Unit 286 is the middle of three pillow breccias (units 287, 286 and 285) that total 97.15 m thickness. The three units are structureless except for eight thin silty intervals (~2 to 20 m apart). Thick packages of pillow lavas lie above (~100 m thick; 2 units) and below (~220 m thick; 5 units) these units. The pillow lavas are not cogenetic. The pillow lava package that underlies the three pillow breccias includes a single, 90 cm thick sandstone layer with silty intervals (Seaman et al., 2000). The basal contact for the three-unit volcanioclastic sequence is flat-lying and depositional. Unit 286 is normally graded near its upper contact, which is gradational into a carbonate silt bed that marks the transition to unit 285 (Seaman et al., 2000).
Unit 286 is ~42 m thick, poorly sorted and contains distinct grain size fluctuations (figures 3.4 and 3.5). Aphyric to low-olivine-phyric basalt clasts are set in a fine ash matrix (58%). Medium lapilli grains to small blocks (>8 cm) are unevenly distributed (average abundance 21%), and are non-vesicular with rare incipiently vesicular clasts. Ash to medium lapilli (average abundance 20%) range in vesicularity from <50%, with decreasing abundance as vesicularity increases. Some largely holocrystalline clasts have glassy margins (<1 cm thick) that are fragmented into smaller angular clasts (figures 3.4 and 3.5). Most clasts (>90%) are equant and have smooth planar to curviplanar, angular to subangular margins, but some clasts have irregular, fluidal margins. Margins become more irregular and scalloped with increasing clast vesicularity.

**Microscopic features**

Basalt textures range from cryptocrystalline to sideromelane. Cryptocrystalline clasts are more abundant in coarser size fractions and are usually very poorly vesicular. Rare poorly vesicular clasts have 30 to 40% vesicles. Sideromelane clasts dominate the ash to medium lapilli fraction (<1 cm). There are two sideromelane clast types, which are distinguished by varying contents of microlites and vesicles (figure 3.6a). The dominant type (>90%) is moderately crystalline (20 to 30% microlites and microphenocrysts) and non-vesicular. Small plagioclase crystals (0.01 x 0.1 mm) dominate the crystal population, followed by
Most clasts have indistinct margins. Others are either very finely crenulated or smooth:

GLASSY LAPILLI TUFF

POORLY SORTED LAVA FLOW BRECCIA

Figure 3.5c

Fluctuations in grain size

Many clasts have dark interiors and light rims.

GLASSY LAPILLI TUFF

Figure 3.4 Stratigraphic log of unit 286 (left) with detailed representative section (right). Small white boxes show locations of point counts and small black boxes show locations of samples taken for thin sections. Arrows indicate larger clasts, which may be on back side of core.
Figure 3.5 Photographs of unit 286 deposits. Scale bars are 5 cm except in (c), which is 3 cm. Scale bars are subdivided into cm segments. a) Working box of unit 286 showing fluctuations in grain size. b) Close up of core showing clast shapes. c) Clast showing gradation from cryptocrystalline basalt with planar, angular margins to hyaline basalt with slightly more irregular and subangular margins. Also shows variation in vesicularity. d) Cross sectional break across core shows angular blocky 3D shape of crystalline clasts and also reveals some fresh black glass.
Figure 3.6 Photomicrographs of thin sections.

a) Unit 286. Note the predominance of sideromelane over tachylite and range of textures.

b) Subunit 260b. Note the variations in vesicularity and crystallinity of the sideromelane clasts, including the tachylite-sideromelane banded clast in the top left corner.

c) Subunit 260a. Note the predominance of tachylite with large olivine phenocrysts.
olivine (<0.4 mm) with glass and spinel inclusions. Locally, plagioclase crystals show flow alignment. Vesicles are small (0.02 to 0.2 mm) and round. The less common type is poorly crystalline (<5% microlites) and incipiently vesicular (5 to 10% vesicles <0.5 mm). Plagioclase dominates (<3%) with smaller olivine (<0.05 mm). Rare clasts display a pervasive sprinkling of tiny plagioclase microlites. Tachylite clasts (<2 cm) are gradational into sideromelane and some individual clasts grade from sideromelane to tachylite. Sideromelane and tachylite clasts are equant with planar to curviplanar margins. There is a single occurrence of a group of ~6 sideromelane clasts (1 mm) forming a jigsaw-fit pattern (Figure 3.6a).

Moderately crystalline clasts have discontinuous palagonite margins (<0.2 mm) with thin brown smectite borders. Ash-sized glass clasts are completely palagonitized. Globular patches of geopetal smectite occur on the outer edge of the brown smectite rim. Vesicle fillings range from none, to thin rinds of palagonite and smectite, to complete filling with pale-green smectite. Poorly crystalline clasts have thinner palagonitic margins (<0.05 mm) and vesicles have brown and pale-green smectite linings with rare palagonite. Pore spaces between clasts are filled with bladed phillipsite and chabazite, which in areas of extensive alteration infills vesicles.
3.1.4 Unit 260

**Macroscopic features**

Unit 260 (~25 m thick) is a poorly sorted breccia with an uneven distribution of coarse lapilli to block size fragments (<28 cm) resting in ash-grade matrix. There are 5 intervals of glassy ash to fine lapilli with an average spacing of ~4 m (Seaman et al., 2000). Towards the base (~60 cm), unit 260 becomes finer grained (fine lapilli to ash) and the abundance of intercalated ash beds increases downwards. The basal contact is sharp and the underlying unit is a basaltic lapilli volcaniclastic deposit of unit 261 (described in Table 5). The upper contact of unit 260 is sharp and depositional. Overlying unit 260 is a ~20 m thick sequence of relatively thin (<5 m) and lithologically diverse volcaniclastics and 'sandstones' (units 259 to 252; Seaman et al., 2000). Deposits categorized as 'sandstones' in the HSDP2 core are clastics that show evidence of reworking, are well sorted and fine grained (Seaman et al., 2000) but may still be volcaniclastic. The bedded sandstone contains silty intervals and a piece of charcoal. Three thick (12 to 30 m), structureless volcaniclastic units underlie unit 260 (units 264, 262 and 261).

The 5.6 m thick section logged in this study consists of a poorly sorted, massive, coarse lapilli tuff overlain by two thin (~1 mm), well sorted, silty beds with planar bedding, separated by ~1 cm thick beds of ash to glassy fine lapilli. The interval of silty and ash/fine lapilli beds is at 1832.3 mbsl. Overlying this interval is a
Figure 3.7 Stratigraphic log of unit 260 (left) with detailed representative section (right). Small white boxes indicate locations of point counts and small black boxes show locations of samples taken for thin sections.
Figure 3.8 Photographs of unit 260. Scale bars are 5 cm long. a) Working box of unit 260 showing a change to finer grain sizes above a large (>10 cm) holocrystalline basalt clast. b & c) Core section showing different types of lapilli clasts ranging from vesicular to non-vesicular, planar to irregular margins, and angular to round corners. d) Close up of variations in clast type. Note the clast with irregular margins has abundant olivine phenocrysts and the clast with planar margins has only rare olivine phenocrysts.

- Moderately sorted, glassy lapilli bed
- Interval of silty and ash to fine lapilli beds
- Massive, poorly sorted, coarse lapilli tuff
- Incipiently to moderately vesicular clasts: margins are usually rougher with increasing vesiularity
- Non-to poorly vesicular with many olivine phenocrysts; irregular margins
- Non-vesicular with few olivine phenocrysts; planar margins
moderately sorted, glassy lapilli tuff (~60 cm), which lacks coarse lapilli and block sized clasts.

The coarse clasts are basaltic with variable vesicularities (none to 25%) and olivine contents (<1 to 25%; microlites to 2 mm). Olivine crystal sizes increase with olivine content and variations in size and abundance occur within individual clasts. Non- to incipiently-vesicular clasts are equant and the margins range from smooth planar/curviplanar to highly irregular, usually as olivine contents increase (figure 3.8d). Poorly to moderately vesicular clasts have irregular and smooth to rough margins respectively (figure 3.8d). The proportions of clast types vary between the poorly sorted coarse lapilli tuff and the glassy moderately sorted lapilli tuff above the silty intervals. Furthermore, two regions are distinguished within the lower coarse lapilli tuff that are referred to as subunits 260a and 260b. The main changes between subunits a and b are the decrease in matrix and non-vesicular clasts, and the increase of clasts with olivine phenocrysts:

<table>
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<tr>
<th>Depth (mbsl)</th>
<th>Point counting data (%)</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>Matrix</td>
</tr>
<tr>
<td>Glassy lapilli tuff</td>
<td>1,832.3-1,831.7</td>
</tr>
<tr>
<td>Silty intervals</td>
<td>1,832.4-1,832.3</td>
</tr>
<tr>
<td>Subunit 260a</td>
<td>1,834.4-1,832.4</td>
</tr>
<tr>
<td>Subunit 260b</td>
<td>1,837.3-1,834.4</td>
</tr>
</tbody>
</table>
**Microscopic features**

The lithologic change between 260a and 260b is also seen in thin section. The deeper sample (260b) is dominated by sideromelane ash and lapilli and the shallower sample (260a) is dominated by lapilli-size tachylite grains. In both thin sections, sideromelane has palagonitization along the margins to an approximately equal thickness (0.1 to 0.3 mm; hence the smaller clasts are completely palagonitized) and contains smaller olivines than the tachylite clasts. Olivine phenocrysts in the sideromelane are unaltered. The olivine phenocrysts within the tachylite have variable alteration from fresh to pervasive palagonitization of glass along cracks. Variation in the alteration is not correlated to the location of the phenocrysts within clasts (e.g., proximity to clast margins) nor with the occurrence of glass inclusions. Both units have poorly crystalline (<10% microlites and microphenocrysts) and highly crystalline (30 to 50% microlites and microphenocrysts) sideromelane clasts with varying vesicularity. Clast margins are smooth and planar except where intersections with vesicles create scallop shapes. Incipiently vesicular glass shapes become more spherical with increasing vesicularity.

**Subunit 260b**

The lower unit is dominated by sideromelane clasts that are non- to incipiently-vesicular and poorly crystalline, incipiently vesicular and poorly crystalline, or non-vesicular and highly crystalline (Figure 3.6b). Non- to incipiently-vesicular
types have <10% vesicles that are small (<0.2 mm) and round whereas incipiently vesicular types (10 to 20% vesicles) have larger (<0.8 mm) vesicles, which are spherical to slightly elongate and aligned. Olivines of poorly crystalline clasts are <0.5 mm, and in the highly crystalline clasts are <1.0 mm. They contain spinels and there are no olivine phenocrysts. Plagioclase crystals are mainly <0.1 mm, but in some of the more crystalline clasts they are tiny (~0.005 mm) and widespread. Some clasts are banded with sideromelane and tachylite zones.

Vesicles and cracks have an outer layer of pale-green smectite and an inner zone of brown smectite. Geopetal smectite grows in globular patches into the vesicles from the brown smectite layer. Sparse groups of evenly spaced and parallel microbe tracks occur around vesicles. These tracks are much less dense than the clumps observed in unit 202. Head-like structures are visible on the tips of these tracks.

*Subunit 260a*

The upper subunit is dominated by tachylite and it has variable degrees of olivine alteration. Sideromelane clasts are small (<1.3 mm) and non-vesicular (<5%; 0.5 mm). Crystals are dominated by olivine microphenocrysts (<0.5 mm), which contain spinel and glass inclusions. Larger clasts (>5 mm) contain 10 to 20% olivine phenocrysts (<3 mm) with spinel and glass inclusions. Plagioclase
microlites also occur (0.02 x 0.2 mm). More crystalline glass clasts are dark brown and grade into opaque tachylite.

Pale-green, pore-lining smectite 2 occurs on the outer edge of some of the palagonitic margins. Vesicles in the center of larger clasts are devoid of palagonite but geopetal smectite and pale-green smectite lines and partially fills them. Sparse microbe tubules extend from vesicles and, less commonly, from clast margins. Clusters of fresh microlites between some clasts indicate complete dissolution of other clasts. In addition, there are some patches of reddened smectite replacement. Highly crystalline sideromelane and tachylite clasts have light yellow, lumpy margins, which is probably pale-green smectite. Pore spaces are filled with chabazite.

3.1.5 Unit 202

Macroscopic features

Unit 202 is a poorly-sorted, structureless pillow breccia (Figure 3.9) that is remarkably monotonous throughout its ~70 m thickness. The unit is overlain by a ~3.8 m thick moderately- to highly-olivine-phyric massive basalt, with a chilled lower margin. Underlying unit 202 is a thin (~1.4 m thick), highly-olivine-phyric massive basalt. Both upper and basal contacts are highly fractured (Seaman et al., 2000), which was probably drilling induced.
Figure 3.9 Stratigraphic log of unit 202 (left) with detailed representative section (right). Small white boxes show locations of point counts and small black boxes show locations of samples taken for thin sections.

- Vesicular clast type: 30-40% coalesced to spherical vesicles <5 mm. Well defined clast margins. Some fresh olivine phenocrysts.
- Clast contains ~20% olivine phenocrysts.
- Dominantly irregular margins & equant shapes. Rod-shaped clasts also common.
- Fluidal shaped basalt clast [10-20% fresh olivine phenocrysts <5 mm; Figure 3.10e]
- Major elements analysis
Figure 3.10: Photographs of unit 202. Scale bars are 5 cm, except in e, which is 3 cm. a & b) Working box of unit 202. c) & d) Irregularly shaped basalt, showing some jigsaw-fit patterns and spheroidal fragmentation in (d). e) Irregular, fluidal shaped clast with smooth margins.
Figure 3.11 Photomicrographs of thin sections.

a) Unit 202. Note the blocky tachylite clast with a sideromelane margin.

b) Unit 194. Note the bimodal vesicularity of the sideromelane clasts and the blocky nature of the poorly vesicular type.
The unit contains only one clast lithology, highly olivine-phyric basalt (28%), set in a fine-grained matrix (<1 mm; 72%). Vesicularity is bimodal: non-vesicular (<10 cm sized clasts) and incipiently to moderately vesicular (<3 cm sized clasts). Clast proportions are 53% non-vesicular clasts, 36% incipiently vesicular clasts, and 11% poorly to moderately vesicular clasts. Non-vesicular clasts are irregular to equant and most are subrounded with finely irregular (sometimes fluidal; figure 3.10e) margins that are rarely partly glassy. Poorly vesicular clasts are rounded and have smooth margins. Some blocks have fractured into medium to coarse lapilli-sized segments and have jigsaw-fit patterns. Fractures are either random or concentric (within glassy margins; figure 3.10d). None are radial as they are in unit 194.

**Microscopic features**

Basalt clasts contain 10 to 20%, evenly distributed, fresh euhedral olivine phenocrysts up to 5 mm with glass inclusions and spinel crystals. Clasts exhibit a range of groundmass textures from cryptocrystalline to hyaline, except poorly to moderately vesicular clasts, which are invariably hyaline. Non-vesicular clasts have <1% round vesicles (<0.2 mm) and sideromelane contains unevenly distributed microlites (1 to 50%; 0.01-0.1 mm). Plagioclase microlites increase in abundance in grains that are flow-banded with tachylite. These clasts are subangular and have planar margins. Incipiently to moderately vesicular clasts have unevenly distributed vesicles (15 to 60%) of various shapes ranging up to 3
Poorly to moderately vesicular sideromelane clasts have 30 to 60% round vesicles (<1 mm) and 5 to 10% microlites. Clasts are subangular with margins that are irregular where they intersect vesicles but are otherwise planar. Tachylite grains are sparse and range from ash to >3 cm in size. They are subangular with planar edges and often have sideromelane margins (Figure 3.11a). All clasts have microvesicles. The above variations in crystallinity and vesicularity can be seen within a single clast. The matrix consists of ash-sized, non-vesicular and moderately vesicular sideromelane basalt with subordinate proportions of tachylite and free olivine crystals (Figure 3.11a).

Microbe tubules are present and are more common in moderately vesicular clasts than non-vesicular clasts. Brown and pale-green smectite coats clast margins and masses of reddened smectite have replaced glass in places. Smaller grains (<2 mm) are often completely altered. Some olivine phenocrysts have been replaced by smectite and vesicles within some tachylite grains are filled with clay.

3.1.6 Unit 194

Macroscopic features

This ~25 m thick, poorly sorted lapilli breccia is structureless and characterized by uneven distribution of coarse lapilli and small blocks (~4 to 8 cm) amongst coarse ash to medium lapilli (<1 mm to 1 cm) and olivine crystals in a matrix of fine-ash to clay (figure 3.12). The unit is sandwiched between two thin (~2 m),
highly olivine-phyric basalt lava flows. The lower contact is sharp and nonerosive, and the top contact is gradational (Seaman et al., 2000). The clast mineralogy (10 to 20% olivine phenocrysts) in volcanioclastic unit 194 is similar to that of the underlying and overlying massive basalts (units 195 and 193; Seaman et al., 2000). There is a bimodal distribution of vesicularity; a poorly vesicular group and a non-vesicular group. The proportion of clasts and matrix varies: core sections with a higher proportion of matrix (>30%) have roughly equal amounts of poorly vesicular and non-vesicular clasts (>30%) and core sections with sparse matrix (<5%) have more poorly vesicular clasts (>70%) than non-vesicular clasts (~20%). Poorly vesicular clasts are subrounded with finely (sub-mm scale) irregular margins. Non-vesicular clasts are predominantly subangular with planar to curviplanar margins. Some clasts have glassy margins, which are irregular and fluidal. Radial cracks commonly divide coarse clasts into 1 to 3 cm (medium lapilli-size) domains that are often slightly separated and rotated (Figure 3.13b to e). These fragments are equant unless they occur where the clast has a glassy margin, in which case they form curved laths (Figure 3.13b). The curved cracks along the inside of glassy edges result in rounded clast margins upon disintegration, which then resemble mature clast shapes. Some coarse clasts have large, irregular central holes and/or zones of poor vesicularity.
Typical large clast: some margins cracking into rod-shaped lapilli; large central voids; (Figure 3.13b)

Most large clasts are fractured into < 1 cm³ sized clasts/domains

Clast with central void; becomes glassy towards margins; in-situ fracturing:

Coarse lapilli and blocks: uneven distribution
Figure 3.13 Photographs of unit 194. Scale bars are 5 cm. a) Archive box 293 shows the monotonous nature of this unit. b) Spherical block of basalt with radial fractures and glassy lower margin breaking into curved, rod shaped lapilli that are parallel with the boundary of the holocrystalline interior. c) An elongate finger of basalt that is glassier towards its thinner end and is fracturing into lapilli sized clasts. d) Crystalline to holocrystalline spherical basalt with radial joints from a large vesicle in center. e) A basalt block with a highly vesicular center and glassy lower margin.
Microscopic features

Coarse clasts are microcrystalline to cryptocrystalline, becoming glassier towards the margins. The matrix (<16 mm) is dominated by brown sideromelane ash (Figure 3.11b), with rare fine ash (<0.2 mm) vesicular tachylite grains and olivine crystals. All the basalt clasts contain evenly distributed olivine phenocrysts that are 0.5 to 5.0 mm in diameter. Sideromelane clasts also contain 5 to 10% olivine microphenocrysts, which are small (<4 mm), euhedral, and have small spinel and glass inclusions. The glassy groundmass of the sideromelane is tan colored and contains 20 to 50% microlites that sometimes change in abundance across grains. Poorly vesicular basalt clasts have 20 to 40% vesicularity, which are <0.6 mm in diameter and are usually round. The finely irregular margins of the poorly vesicular clasts are scalloped by vesicles and have fractured into smaller clasts. Rare, slightly elongate, sideromelane clasts show flow banding and coalescence of vesicles. Incipient vesicular (5 to 10% vesicles) sideromelane have smaller vesicles (<0.1 mm). Non-vesicular clasts contain 0 to 1% vesicles. Incipient and non-vesicular clast shapes are equant and subangular with smooth planar to curvilinear margins. Shapes of clasts dominated by single olivine crystals are controlled by curved fractures, most probably reflecting an elliptical stress field around the olivine. Alteration of this deposit is minor. Brown smectite lines or fills some vesicles within the sideromelane, but the glass is fresh.
3.2 Microprobe Chemical Analyses of Glass Clasts

The HSDP2 submarine deposits contain abundant fresh glass, which document melt composition (Stolper et al., 2004). High-resolution sampling of the HSDP2 core for glass analyses has provided a comprehensive geochemical data set chronicling melt compositions during the growth at Mauna Kea's submarine flank (e.g., Seaman et al., 2004; Stolper et al., 2004). Electron microprobe analyses of glass samples were performed mostly at the University of Hawai'i at Manoa with additional analysis at the California Institute of Technology. Glass clast compositions are useful for interpreting source and depositional processes. Volatile contents are especially useful for determining depth of magma degassing and, therefore, are a proxy for eruption depth. Glasses with S <0.04% degassed at near atmospheric pressures, whereas glasses with higher S (>0.09%) were probably erupted at water depths >500 m and are undegassed (Garcia and Davis, 2001; Moore and Clague, 1992; Seaman et al., 2004). Partially degassed glasses are uncommon in the HSDP2 core, possibly because they form within a narrow depth range (0-500 mbsl; Seaman et al., 2004).

The HSDP2 drill site is located on Mauna Loa tholeiitic lavas and is underlain by Mauna Kea alkalic and tholeiitic lavas (Rhodes, 1996). Undegassed glasses have two groups defined by silica (48-49% and 51-52.5% SiO$_2$), isotopes, and other major and minor element contents (Stolper et al., 2004). Undegassed glasses are typically less fractionated (>7% MgO) than degassed and partially
degassed glasses, which generally have <7% MgO (Seaman et al., 2004). Degassed, more fractionated samples have a range of silica contents and therefore, two distinct primary magma sources are thought to have supplied Mauna Kea's shield stage eruptions with high silica and low silica magmas (Stolper et al., 2004).

Stolper et al., 2004) defined four zones within the submarine deposits based upon the dominant silica content (>50% vs <50%) and alkalinity of glass clasts:

<table>
<thead>
<tr>
<th>Zone</th>
<th>Depth (mbsl)</th>
<th>Silica</th>
<th>Alkalinity</th>
<th>Volatile contents</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1079-1950</td>
<td>High</td>
<td>Tholeiitic</td>
<td>Degassed</td>
</tr>
<tr>
<td>2</td>
<td>1950-2235</td>
<td>Increases with depth</td>
<td>Tholeiitic</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>2235-2480</td>
<td>Low</td>
<td>Transitional</td>
<td>Undegassed</td>
</tr>
<tr>
<td>4</td>
<td>2480-3098</td>
<td>High</td>
<td>Tholeiitic</td>
<td>Undegassed</td>
</tr>
</tbody>
</table>

*Source of data in table is Stolper et al., (2004)*

There are exceptions to the dominant chemical characteristics listed in the table: Zone 1 includes thin intervals of low silica glasses (<50%) between ~1400 and 1800 mbsl. Throughout zone 2, silica contents increase from low to high with depth but all the units are tholeiitic and degassed. The top 45 m of Zone 3 contains alkalic lavas that become less fractionated with depth. Undegassed volcanioclastics are rare in the HSDP2 core and are thin (up to 6 m thick). Most volcanioclastics in the HSDP2 core are degassed, suggesting that the majority of
volcaniclastics (95%) are derived from shallow submarine and subaerial eruptions (DePaolo et al., 2001; Seaman et al., 20044).

Glass clasts from 11 units were analyzed for major elements (SiO₂, TiO₂, Al₂O₃, FeO, MnO, MgO, CaO, Na₂O, K₂O, P₂O₅ and S) in order to assess the chemical heterogeneity within and between units. The number of clasts analyzed within a thin section was usually limited to the number of fresh glass clasts because many were too altered or microlite-rich for analysis. Typically, three points were analyzed on each glass clast. More points (4-6) were analyzed on clasts that displayed gradational textures (e.g., sideromelane to tachylite).

3.2.1 Major element analyses of the HSDP2 units

Introduction

Glass clasts analyzed from the HSDP2 volcaniclastics are all tholeiitic. All the volcaniclastics within zone 1 (units 261, 260, 238, 202 and 194) have high average silica (>50%) and low MgO (<7%; table 3.2; figure 3.14). The upper two units (275 and 274) within zone 2 have lower silica and MgO, but the lower unit (286) within this zone has high silica and low MgO. There are two units (294 and 293) within zone 3, and they have low silica and high MgO. Unit 297, which is at the top of zone 4, has high silica and low MgO. All the volcaniclastics are degassed except those in zone 3, which are all partially degassed or undegassed (S 0.052 to 0.100 wt %). These results are consistent with the overall chemical
stratigraphy of the HSDP2 core (Stolper et al., 2004), including the correlation between degassed, more fractionated magmas and partially degassed or undegassed, less fractionated magmas (figure 3.14).

Cooling temperatures for the HSDP2 volcaniclastics were calculated using the MgO-based equation (temperature (°C) = 20.1 MgO (wt %) + 1014°C) developed by Helz and Thornber (1987). The average cooling temperature calculated from MgO contents of glass clasts in the HSDP2 units studied is 1152°C, although the lowest is 1126°C (unit 188) and the highest is 1208°C (unit 294). Cooling temperatures are generally constant (within 10°C) within individual units, however, unit 294 exhibits a range of 43°C (1165 to 1208°C). Compatible element trends show that most HSDP2 units analyzed in this study have undergone olivine fractionation followed by clinopyroxene and plagioclase crystallization (figure 3.15b and c). However, units 293 and 294 (zone 3; the only undegassed units) only had olivine fractionation.

Among the more fractionated magmas, there are two parallel trends on the MgO-TiO₂ plot (figure 3.15e). All of the units in the lower MgO and TiO₂ trend have high silica and most units in the higher MgO-TiO₂ trend have low silica (only unit 238 has high silica). These trends indicate that there are at least two parental magmas from which these units are derived. Multiple parental magmas are also indicated by the range of Ti/K ratios (based upon 4σ counting errors, ~0.38 Ti/K).
Table 3.2 Average major element analyses of glass clasts in HSDP2 volcaniclastic units (in weight percent). Units focused on in this study are in bold font. In addition, analyses of units surrounding the examined units are included for comparison. Visual estimates of crystallinities are given as modal percent crystals.

<table>
<thead>
<tr>
<th>Unit</th>
<th>SiO$_2$</th>
<th>TiO$_2$</th>
<th>Al$_2$O$_3$</th>
<th>FeO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na$_2$O</th>
<th>K$_2$O</th>
<th>P$_2$O$_5$</th>
<th>S</th>
<th>Total</th>
<th>Ti/K</th>
<th>Modal % crystals</th>
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</thead>
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<td>297</td>
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<td></td>
</tr>
</tbody>
</table>

* = analyses collected from Stolper et al., (2004)
Figure 3.14 Silica, MgO and S of glass clasts in the HSDP2 core versus depth. Data points are average values for single units. Solid points were analysed during this study and open points and trendlines are from Stolper et al. (2004). Nearly all volcaniclastic (squares) have relatively high silica, low MgO, and low S. Pillow lavas (circles) become increasingly silica-poor, MgO-rich, and S-rich with depth. See text for degassing criteria.
Figure 3.15  MgO variation diagrams from microprobe analyses of glass clasts within HSDP2 volcaniclastics.
Assessment of the melt composition of individual units

Incompatible element ratios (Ti/K) are used to assess the number of magma batches supplying the volcaniclastic units. Compositions of small individual eruptions at Kilauea volcano since 1820 (~0.1 km³) vary by Ti/K ~0.2. This implies that ranges less than 0.2 represent a single magma batch. However, the average total error for Ti/K values in these analyses is 0.38. Thus, ranges of Ti/K ratios greater than total errors are assumed to be too significant for a single magma batch to have supplied the clasts. Average Ti/K ratios of HSDP2 units could, therefore, imply that one magma batch supplied clasts to units 194 and 202 (Ti/K 6.34 and 6.31 respectively), another to unit 260 (Ti/K 5.29), and a third to units 286, 293 and 294 (Ti/K 6.76, 6.73 and 6.84 respectively). In addition, average Ti/K ratios do not range significantly within each unit (figures 3.16, 3.17 and 3.18), implying that single magma batches supplied glass clasts per unit (despite most units displaying various clast textures). Furthermore, limited variation in chemical compositions (smaller than analytical errors) of glass clasts in units 293, 286, 202 and 194 (figures 3.16, 3.17 and 3.18) suggests that each unit was derived from a single eruption.

Units 293 and 294 have the highest average MgO and lowest average SiO₂ contents of all units studied. The matrix of unit 293 is so fine grained that good analyses could only be attained from the glass margin of an isolated pillow lava and associated fragments. This is responsible for the limited range in glass
Figure 3.16 Major element analyses of glass clasts within units 294 (crosses) and 293 (diamonds). Unit 294 is derived from a single magma source, but two groups are distinguished by different MgO, representing variable fractionation. Unit 293 has a small range of Ti/K ratios at similar MgO contents and the analyses are tightly constrained for compatible elements, indicating the glass clasts analyzed are from a single magma source.
Figure 3.17 Major element analyses of glass clasts within units 286 (open circles) and 260 (crosses). Values are oxide weight percents. Unit 286 is from a different magma source to unit 260 and it is homogeneous. Two groups are distinguished for unit 260 based upon petrological evidence suggesting that two lithologies are present (see text for details; section 3.2.3). The lower, 260b (+), and upper, 260a (x), subunits are derived from the same magma source.
Figure 3.18 Major element analyses of glass clasts within units 202 (diamonds) and 194 (crosses). Unit 194 is homogeneous in terms of both magma source and fractionation history. Unit 202 is also derived from a single magma source. The two groups of clasts in unit 202 that are distinguished by MgO contents represent slightly different degrees of fractionation, but this small difference can be caused by fractionation of only 1% olivine and it is, therefore, insignificant.
chemistry of this unit. In contrast, unit 294 has large ranges of some elements (e.g., MgO), which is attributed to different fractionation stages (figure 3.16). The more fractionated, lower MgO (<8.0 wt %) glasses correlates with analyses from unit 293 and Ti/K ratios suggest that these units are derived from the pillow lava of unit 293. The lower MgO glasses are non-vesicular with microlites (3-10%; olivine) and the higher MgO clasts have various textures categorized into three types; non-vesicular with no microlites, incipiently to poorly vesicular with no microlites, and highly vesicular with microlites (3-10%; olivine). Although units 294 and 293 units are derived from the same pillow flow, unit 294 is glassier and less fractionated, possibly simply because it cooled faster.

Two thin sections from unit 260 were analyzed; one from subunit 260b at 1841.7 mbsl and one from subunit 260a at 1837.5 mbsl. Glasses from both thin sections have the same Ti/K content and therefore, could have been derived from the parental magma. However, subunit 260b has a wider range of compositions and higher MgO, FeO, CaO, Al₂O₃ and TiO₂ averages (Figure 3.17). Fractionation trends that could account for this variation include crystallization of olivine, plagioclase and clinopyroxene.
3.2.2 Geochemical comparisons to adjacent units

Analyses from the HSDP2 glass composition data set were used to compare the major element contents of units adjacent to those studied with the aim of determining continuous associations versus pronounced shifts in magma supply. Where no data is available for adjacent units, the closest analyzed unit is compared.

Unit 294 is compositionally similar to the underlying lava flow and volcaniclastic units (296 and 295), except it has some clasts with higher MgO, indicating parts of the lava cooled at a higher temperature (table 3.2; figure 3.14). Unit 293 is chemically indistinguishable from the overlying basalt flow, unit 292, which is also undegassed (0.090 and 0.110% S respectively). Units 287, 286, 285 constitute a package of pillow lava breccias that have similar major element compositions to the overlying pillow lava (unit 284). However, the underlying pillow lava (unit 288) has lower silica (48.63 wt % versus 51.66 wt %) and Ti/K ratio (5.7 versus 6.4), indicating a distinctive magma batch. Unit 260 has significantly lower silica than units 264, 262, 261, and 258. Correspondingly it has higher TiO₂, CaO (cf. all units except 262), Na₂O, K₂O, and P₂O₅. Whole rock analyses of unit 202 and the underlying and overlying basalt lava flows are similar in major elements and Ti/K ratios (~8.7), suggesting a single magma batch supplied the material to these units. Major element analyses of unit 194 are similar to volcaniclastic units.
196 and 192, which are separated from unit 194 only by ~2.5 m thick lava flows of units 195 and 193.

### 3.3 Shape Analysis of Clasts

Processes that fragment magma produce clasts with shapes controlled by the fragmentation mechanism and the rheology of the magma at the time of fragmentation. Morphological parameters that aid the recognition of processes of formation include aspect ratio, formfactor (roughness), solidity, and roundness. Quantitative analyses of these features were performed on clasts in thin section from each HSDP2 unit. For comparison, three deposits of clasts formed in known environments and produced by well constrained fragmentation processes were examined: a subaerial fire fountaining deposit where lava was fragmented by magmatic explosivity (Kilauea Iki, 1959); flowage of lava into the ocean where both fluidal and brittle lava was fragmented by phreatomagmatic processes (Kilauea ocean entry 1990); and a beach deposit of lava fragmented epiclastically (1955 Kilauea lava flow; see section 2.2.4 for details).

The end-member samples are valuable examples of Hawaiian fragmentation environments. However, more than one fragmentation process occurs within all three environments: I have identified three clast types in the magmatic deposit that can be attributed to at least two different processes of formation (section 3.3.1). The dynamic littoral environment has multiple processes. The epiclastic
beach deposit is the closest to representing a single fragmentation processes, but even these clasts may have undergone other processes of shape modification (pre-existing shapes). The samples represent endmember environments of clast formation at Hawaiian Islands and are, therefore, useful comparisons for studying paleo-deposits such as those in the HSDP2 core.

3.3.1 Description of clasts within the end-member deposits

Clasts in the primary magmatic explosive deposit exhibit a range of textures that can be divided into two first order categories. The most abundant clast texture is broken acneliths, which have a black, usually shiny (less commonly matt), and smooth surface on some sides of the clasts. The other sides are broken and expose the glassy and moderately vesicular interior. The smooth, fused surfaces and lower vesicularity (than the primary, foamy golden glassy clasts - see below) suggest that these clasts are recycled fire fountain ejecta that outgassed in the lava lake, drained back into the vent and were reincorporated in subsequent explosions involving newly arrived volatile magma below (Bruce Houghton, Pers. comm., 2004). Thus these clasts were hot during transport and have a fluidal texture, yet they were not actively vesiculating. The other clast type is foamy golden glassy clasts. These are primary products of the fire fountaining and their surfaces range from open vesicles to smooth surfaces that were fused by hot gases in flight. Some of the foamy clasts are very irregular in shape because the fused surfaces are pock-marked by underlying vesicles. Striations on some
surfaces are created by tiny but very elongate stretched vesicles. Pele's tears and hairs are abundant in finer size fractions. Vesicles are spherical (except the stretched vesicles in the fused surfaces of the golden foam) and range in size up to ~1mm.

Clasts in the phreatomagmatic deposit are poorly sorted and are divided into new and old material based on luster, oxidization and rounding. Newer clasts are all moderately vesicular and most have some smooth and shiny surfaces similar to the Kilauea Iki acneliths, although there is a wider variation in shape than the Iki deposits. There are three types of old material: (1) oxidized, red to black, sub-angular, poorly vesicular basalt with a dull luster. These clasts have irregular shapes and irregular margins due to vesicles; (2) angular basalt clasts that are incipiently vesicular. Vesicles are convolute and have oxidized walls; (3) poorly vesicular basalt clasts that are glassy and equant.

All the epiclastic beach clasts are black, non-vesicular and sub-rounded. Overall, clast shapes are smooth (due to polishing of surfaces by wave action), although olivine phenocrysts, which are more resistant to mechanical weathering, create a bumpy surface texture.
3.3.2 Parameters

Aspect Ratio

End member samples

All deposits have nearly equant averages of clasts (~1.51; figure 3.19). The magmatically explosive deposit clasts have nearly Gaussian frequency distribution, although a fluidal rod-shaped clast (which is an outlier on the frequency-aspect ratio chart; figure 3.19) forms a long positive tail to the curve. This deposit has the highest minimum aspect ratio (1.10) of all analyzed samples (including the HSDP2 units). The phreatomagmatic deposit has a negatively skewed Gaussian frequency curve. It also has the largest range in aspect ratio (2.91) and it contains clasts with higher aspect ratios than the maximum of the epiclastic and magmatic explosivity deposits. As with the magmatic explosivity deposit, the clasts that constitute the highest aspect ratios in the phreatomagmatic deposit are fluidal and rod-shaped. The epiclastic deposit has a negative trend and the narrowest range of aspect ratios with no high aspect ratio clasts (figure 3.19).

HSDP2 units

The HSDP2 units are predominantly composed of nearly equant clasts (average aspect ratio is 1.6; figure 3.20). Units 293 and 294 have similar aspect ratio frequencies to the negatively skewed Gaussian curve of the phreatomagmatic deposit. Units 188 and 194 have similar distribution curves to the negative trend.
Figure 3.19. Aspect ratio and formfactor analyses of clasts within the endmember deposits. Note the similarity of aspect ratios between all three samples. The phreatomagmatic deposit contains clasts with the highest aspect ratios and the epiclastic deposit has the most equant clasts. Formfactor frequency distributions have more variation between the samples. The phreatomagmatic deposit has the lowest formfactor and the greatest range of clast roughness, and the epiclastic deposit has the highest formfactor.

Minimum, average, and maximum formfactor values of deposits (see figure 3.20 for aspect ratio values). Figures in parentheses are values after removal of outliers.

<table>
<thead>
<tr>
<th>Deposit Type</th>
<th>Minimum</th>
<th>Average</th>
<th>Maximum</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magmatic deposit</td>
<td>0.39</td>
<td>0.55</td>
<td>0.77</td>
</tr>
<tr>
<td>Phreatomagmatic deposit</td>
<td>0.36</td>
<td>0.62</td>
<td>0.83</td>
</tr>
<tr>
<td>Epiclastic deposit</td>
<td>0.61</td>
<td>0.76</td>
<td>0.87</td>
</tr>
<tr>
<td>Unit 194</td>
<td>0.21</td>
<td>0.67</td>
<td>0.85</td>
</tr>
<tr>
<td>Unit 202</td>
<td>0.19(0.56)</td>
<td>0.62</td>
<td>0.81</td>
</tr>
<tr>
<td>Unit 260</td>
<td>0.34</td>
<td>0.58</td>
<td>0.83</td>
</tr>
<tr>
<td>Unit 286</td>
<td>0.21(0.44)</td>
<td>0.63</td>
<td>0.84</td>
</tr>
<tr>
<td>Unit 293</td>
<td>0.24</td>
<td>0.59</td>
<td>0.89</td>
</tr>
<tr>
<td>Unit 294</td>
<td>0.24</td>
<td>0.60</td>
<td>0.96</td>
</tr>
</tbody>
</table>
Figure 3.20 Aspect ratio analyses of clasts within the HSDP volcaniclastic units. Units with asterix are those focused on in the study. Arrows indicate outliers.

Minimum, average, and maximum aspect ratio values of deposits
Figures in parentheses are values after removal of outliers

<table>
<thead>
<tr>
<th></th>
<th>Minimum</th>
<th>Average</th>
<th>Maximum</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magmatic deposit</td>
<td>1.10</td>
<td>1.55</td>
<td>3.32 (2.19)</td>
</tr>
<tr>
<td>Phreatomagmatic deposit</td>
<td>1.02</td>
<td>1.60</td>
<td>3.93</td>
</tr>
<tr>
<td>Epiclastic deposit</td>
<td>1.05</td>
<td>1.39</td>
<td>2.19</td>
</tr>
<tr>
<td>Unit 194</td>
<td>1.06</td>
<td>1.47</td>
<td>3.27 (2.61)</td>
</tr>
<tr>
<td>Unit 202</td>
<td>1.08</td>
<td>1.42</td>
<td>2.08</td>
</tr>
<tr>
<td>Unit 260</td>
<td>1.07</td>
<td>1.60</td>
<td>4.40 (2.66)</td>
</tr>
<tr>
<td>Unit 286</td>
<td>1.05</td>
<td>1.52</td>
<td>2.28</td>
</tr>
<tr>
<td>Unit 293</td>
<td>1.00</td>
<td>2.00</td>
<td>5.67</td>
</tr>
<tr>
<td>Unit 294</td>
<td>1.04</td>
<td>1.61</td>
<td>3.88</td>
</tr>
</tbody>
</table>
of the epiclastic deposits. The other units do not categorically approximate those of the endmember deposits. Minimum and average aspect ratios of the HSDP2 samples are similar to the phreatomagmatic and epiclastic deposits, but all are lower than the magmatic explosivity deposit. Maximum aspect ratios show the most variation but these are usually created by outliers. The only deposit with significantly higher aspect ratio is unit 293. Of the units focused on in this study, units 293 and 294 have the largest range of aspect ratios (4.67 and 2.84 respectively). These are most similar to the range of aspect ratios in the phreatomagmatic endmember deposit (2.91).

**Formfactor**

**End-member samples**

Formfactor yields more varied frequency distribution shapes and ranges between the samples (i.e., there is more similarity between the aspect ratio plots; figures 3.19, 3.20 and 3.21) and the endmembers display Gaussian frequency curves. The epiclastic deposit is clearly distinguished by its smoother clasts, resulting in a narrower range of formfactors (figure 3.19), and there is a complete absence of rough clasts in the beach deposit. The magmatic deposit has the lowest average formfactor.
HSDP2 units

Formfactor ranges for the HSDP2 units are larger than for the endmember samples (figures 3.19 and 3.21). The average formfactors of the HSDP2 units (0.58 to 0.67) are not as extreme as the average values of the magmatic and epiclastic deposits (0.55 and 0.76 respectively). All the HSDP2 units have lower minimum formfactor values (0.19 to 0.34) than any of the endmembers (0.36 to 0.61). These rough clasts are easily distinguished visually and they constitute a minor component of the clast morphologies. Units 293 and 294 are the only HSDP2 deposits with higher formfactor values than the smoothest clast of the epiclastic deposit. Low formfactor clasts are always more vesicular than the smoother clasts and unit 202, which contains the most vesicular clasts (<60%), has the lowest formfactor value.

Solidity and Roundness

There is not much variation in solidity or roundness values of clasts between any of the samples: averages and maximum values are extremely similar between the endmembers and units, and the range of minimum values is a consequence of outliers (figure 3.22). Solidity reveals similar results to formfactor, with epiclastic clasts displaying a narrow range of high solidity clasts. The phreatomagmatic clasts have the greatest solidity range, including lower values than the magmatic deposit. All HSDP2 units have average solidities closest to
Figure 3.21 Formfactor analyses of clasts within the HSDP volcaniclastic units. Units with asterix are those focused on in the study.
Figure 3.22 Solidity analyses of clasts within the end-member samples and some HSDP volcanioclastic units. Arrows indicate outliers occur.
that of the phreatomagmatic deposit and most have similar or greater ranges of values than all endemember deposits.
4. INTERPRETATIONS AND CONCLUSIONS

4.1 Limitations of the drill core
The drill core width (6 cm) precludes descriptions of lateral depositional structures, such as bedding geometries. Hence, interpreting certain aspects of the processes involved in the formation and deposition of the deposits is impractical. For example, the lateral extent and volume of deposits are unknown, which prevents unambiguous characterization of the source, such as the longevity and/or volume flux of eruptive activity. In addition, the catchment area for the deposits and the overall relative abundance of lithologies on the entire submarine flank are unknown. The relationships between the volcaniclastics and surrounding deposits are also poorly constrained, which limits inferences regarding material source (e.g., if a volcaniclastic is derived from the underlying flow, overlying lava flows, or neither).

4.2 Emplacement depths of core deposits
The current estimate of the average subsidence rate of Hawai‘i Island (2.5 mm/yr) is consistent with rates estimated for HSDP1 lavas (DePaolo et al., 2001). The submarine section of the core has an approximate age span of 230,000 years (400 to 630 ka; Seaman et al., 2004). Thus, during the period of deposition of the submarine section in HSDP2, ~575 m of subsidence would have occurred assuming a 2.5 mm/yr subsidence rate. Therefore, the base of
the HSDP2 core (3,098 mbsl) represents an emplacement depth of 1,444 m (which is the total submarine section thickness, 2019 m, minus 575 m of subsidence). The ratio between actual thickness (2,019 m) and the emplacement depth range of the submarine section (1,444 m) is ~0.72. This depth adjustment factor can be used to estimate the emplacement depth \( E \) of deposits in the HSDP2 core by

\[
E = (d - 1079) 0.72
\]

where \( d \) is the depth (mbsl) of the deposit within the submarine section of the HSDP2 core.

4.3 Unit Interpretations

4.3.1 Common features of HSDP2 units

There are three common characteristics of all the volcaniclastic units studied:

1) Sideromelane dominates the volcaniclastic clasts, even though Hawaiian pillow lavas and pahoehoe flows typically have only a thin glassy rind (10-20 mm and 1-5 mm respectively) that constitutes <10-20% of the flow (Naka et al., 2002; Hon et al., 1994). This discrepancy suggests that fragmentation occurred during flow emplacement (rather than after solidification), allowing fluidal magma to be repeatedly exposed and quenched rapidly, resulting in a clast population that has a high glass:holocrystalline ratio.

2) Vesicularity of clasts (<3 to 60%) is always below the volume required for fragmentation by the exsolution of magmatic volatiles alone (~75 volume %; Head and Wilson, 2003).
3) Most of the volcaniclastics are poorly sorted (only units 294 and 293 contain well sorted beds) and matrix-supported. The poorly sorted nature indicates that the source (i.e., the fragmentation mechanism) supplied a wide range of clast sizes, transport was inefficient at size fractionation, and deposition was rapid. The transport medium had enough strength to carry large lapilli to block-size clasts. Transport distance was probably short and non-turbulent (suppressing vertical stratification of clasts by size and density).

4.3.2 Units 294 and 293 (Core depths: 2,465 to 2,459; 2,459 to 2,359 mbsl)

**Emplacement depths and relationship to enclosing units**

Emplacement depths of units 294 and 293 are estimated at ~990-985 and ~985-915 mbsl respectively based on the depth adjustment factor. The moderate to high sulfur contents of the material in both units (0.052 - 0.100 wt % S; figure 3.15) indicates that their source lava(s) erupted in deep water. Unit 294 is a 5.6 m thick, bedded volcaniclastic deposit (figure 3.1). Overlying unit 294 is a predominantly pillow lava unit but the basal 36 cm is volcaniclastic (figure 3.1). Units 295 (pillow lava), 294, 293, and 292 (pillow lava) all have similar chemistry (figure 3.14). This may indicate that these units are part of a single, pulsed eruption. However, the volcaniclastic material in units 293 and 294 was probably derived from the pillow lava of unit 293 because the contacts within unit 293 and between 293 and 294 are depositional. Furthermore, the volcaniclastic in unit 293 has indistinguishable chemistry from the pillow lava that makes up the bulk
of unit 293, and TiO$_2$/K$_2$O ratios indicate that units 294 and 293 are comagmatic (figure 3.16).

**Fragmentation processes of the volcanic sediment source**

Clasts of units 294 and 293 are predominantly equant sideromelane with planar to curviplanar margins, suggesting breakage of the pillow lava occurred within via quench-fragmentation during flow emplacement (Figure 3.3). Furthermore, the dominance of microlite-free sideromelane indicates that this quenching occurred on eruption. Breakage of glassy pillow lava margins into elongate, jigsaw-fit clasts also indicates fragmentation upon extrusion. Subordinate clasts of sideromelane with abundant plagioclase microlites and tachylite represent slightly slower quenching and are probably derived from slightly deeper parts of the lava flow.

Clasts with scalloped margins occur in both units (figure 3.3) and indicate that exsolution of magmatic volatiles could have aided fragmentation. Curved shards in unit 293 may be an effect of large vesicle sizes relative to the sideromelane shards, or they may indicate fragmentation by spallation (e.g., Schmincke et al., 1979). The latter is favored because the isolated pillow lavas are more vesicular than the main overlying flow, indicating late stage vesiculation and expansion.
Depositional/transport process

Unit 294 contains volcaniclastic beds distinguished by different grain sizes (ash to lapilli; figure 3.1). Most beds in unit 294 are normally graded and many are capped by finer (ash-size) intervals. The contact between the massive fine lapilli tuff and the overlying coarse tuff is gradational on a microscopic scale (figure 3.3a-ii), which indicates accumulation of the coarse and fine beds was overlapping and probably from a single depositional medium. These features indicate that deposition was pulsed and each transport medium became weaker with either time or height within the flow. It is probable that recurring vertically stratified, non-cohesive density flows deposited unit 294 (e.g., Mulder and Alexander, 2001). In this scenario, the coarser bed would represent the main body of the flow, and the finer overlying bed would be deposited from the tail of the flow. The uppermost bed in unit 294 is coarse ash with carbonate intervals and a fine lapilli layer (figure 3.1). The thin intervals of carbonate are thought to represent breaks in the supply of volcanic clasts.

In the volcaniclastic portion of unit 293, the absence of bedding indicates that deposition was continuous. The incorporation of isolated lava pillows suggests that the pillow lava was being emplaced simultaneously with the deposition of the sediment. In addition, they indicate that the pillow lava was flowing down a slope that was steep enough for lobes of lava to separate from the main flow and roll or sink down into the glassy sediment. As the flow overran the sediment, lobes
budded into the sediment and some remained attached to the main flow. The sediment was probably, therefore, accumulating at the active flow front in a depression or on a flatter part of the slope, and there was little secondary transport.

**Summary**

Unit 294 is a “syn-eruptive hyaloclastite” produced by the non-explosive fragmentation of an active pillow lava (unit 293; figure 4.1a). It was transported downslope and deposited on a relatively flat area of the submarine flank by vertically stratified density flows. As the pillow lava flowed onto the flatter area, fragmentation decreased and hiatuses occurred in the volcaniclastic accumulation (represented by carbonate intervals). As the pillow lava neared the site of the HSDP2 core, deposition of unit 294 was succeeded by the in-situ deposition of a primary, flow-front hyaloclastite (the “in-situ, isolated pillow breccia” at the base of unit 293; figure 4.1b). This hyaloclastite was derived from glassy rinds spalled from the pillow lava and includes isolated pillow lavas. Finally, the pillow lava overran the volcaniclastics and capped the succession (figure 4.1c).
Figure 4.1a Schematic of processes involved in the formation of unit 294. Not to scale. Average volcano flank slope is 14°.

**Figure Caption:**

- **WATER STAGE mbsl** (represents emplacement depths estimated from the depth adjustment factor)

- **DEEP WATER STAGE**
  - Submarine eruption of an undegassed pillow lava flow
  - Passive quench-fragmentation of pillow lava rinds produces sideromelane and tachylite clasts
  - Clasts become normally graded within the density current
  - Equant sideromelane with planar to curviplanar margins
  - A syn-eruptive density current is triggered; possibly by a local increase in slope
  - Deposits from ~1000 mbsl to ~1,450 mbsl are interbedded pillow lavas and volcaniclastics
**Figure 4.1b Schematic of processes involved in the formation of unit 293**

**DEEP WATER STAGE**

- Isolated pillow lavas fall into the hyaloclastite
- Vesicularity suggests expansion of the isolated pillow lavas occurred after separation
- Jigsaw-fit patterns of curved sideromelane clasts around the pillow lava margins suggest spallation was involved in the production of the hyaloclastite material
- Continued eruption of pillow lava flow
- Continued quench-fragmentation produces sideromelane and tachylite clasts
- Fine grained (ash size), dominantly sideromelane clasts from the pillow lava rinds are deposited near the front of the encroaching lava flow
- Deposit from density current
- Carbonate layer accumulates after density current wanes. During this deposition, the lava flow advances downslope towards the HSDP2 site.
Figure 4.1c Schematic of processes involved in the formation of unit 293

DEEP WATER STAGE

- Pillow lava flow overruns and caps the volcaniclastic deposits.
- Some lava slightly intrudes the hyaloclastite.
- In-situ isolated pillow lava breccia.
- Syn-eruptive hyaloclastite.
- Carbonate interval.
- Secondary hyaloclastite.
- Continued eruption of pillow lava flow.

Continued quench-fragmentation produces sideromelane and tachylite clasts.

Pillow lava flow overruns and caps the volcaniclastic deposits.
4.3.3 Unit 286 (Core depth: 2,215 - 2,173 mbsl)

Emplacement depths and relationship to enclosing units

Emplacement depths of unit 286 are estimated at ~810-780 mbsl based on the depth adjustment factor, although very low sulfur contents in the glass (0.021 - 0.007 wt %) imply that the magma erupted subaerially (e.g., Moore and Fabbi, 1971). Therefore, the source lava and/or the volcaniclastic sediment may have traveled down the volcano's submarine flank for at least 3 km (assuming an average slope of 14°; DePaolo et al., 2001).

Unit 286 is in the middle of a package of three volcaniclastic units (~100 m thick) that have chemically similar glass clast assemblages (figure 3.14) and may be comagmatic. The pillow lava of unit 284 that overlies this package of volcaniclastics also has similar major element compositions (figure 3.14). Therefore, this pillow lava may have supplied the material to unit 286. In contrast, the underlying pillow lava has distinctly lower silica (48.6 wt % versus 51.7 wt %) and Ti/K ratio (5.7 versus 6.4), indicating a different parent magma composition.

All three volcaniclastic units in this package are textually similar basaltic volcaniclastics (e.g., poorly sorted, matrix-supported, and similar grain size), indicating no major shift in the depositional regime. Clast types comprise diverse lithologies such as aphyric and low-olivine-phyric basalt. The lower contact with
unit 287 is gradational, indicating continuous deposition, whereas the upper contact with unit 285 is sharp and marked by a silt layer, which represents a lull in volcaniclastic deposition. Silty intervals (eight unevenly distributed sets of 1 to 10 silt layers <2 cm thick each) are more common in unit 286 that in 287 and 285, reflecting more frequent interruptions in clast supply.

**Fragmentation processes of the volcanic sediment source**

The homogeneous chemistry of glass grains (figure 3.17) and occurrence of glassy margins on some clasts indicate that quenched lava flow(s) from a single eruption supplied the material for unit 286. The equant shapes and planar to curviplanar margins of sideromelane suggest that the lava rind was quench-fragmented by cooling-contraction (e.g., Schmincke et al., 1979) rather than spallation. Fragmentation processes also disrupted the interior of the flow, supplying clasts that cooled more slowly (i.e., cryptocrystalline clasts and clasts that grade in crystallinity). Fragmentation processes may include autobrecciation and erosion. The former is more likely based on the high proportion of sideromelane in the matrix. Another fragmentation process is represented by jigsaw-fit patterns of glassy angular clasts associated with the margins of larger cryptocrystalline clasts. These broke in-situ in brittle fashion on deposition or compaction, probably along fractures produced during quenching.
Depositional/transport process

The range of angularities (angular to subangular) may reflect the combined effects of attrition and continued breakages (angular corners) during transport for many kilometers down the flanks of Mauna Kea. The absence of bedding between the silty intervals and the coarse grained nature of unit 286 (figure 3.9) suggests that deposition was by a series of debris flows (e.g., Mulder and Alexander, 2001).

Summary

The source for unit 286 may have been an active lava flow that was undergoing continuous quench fragmentation and autobrecciation (figure 4.1d). Periodic oversteepening of the lava flow initiated collapse events that produced debris flows, which formed deposits 2 to 30 m thick. These deposits are comprised of finer (ash to fine lapilli) sideromelane and tachylite clasts and coarser (coarse lapilli to block-sized) holocrystalline clasts. Silty intervals indicate that the debris flows were recurring. Unit 286 is a 'syn-eruptive lava flow breccia'.

4.3.4 Unit 260 (Core depth: 1,839 - 1,814 mbsl)

Emplacement depths and relationship to enclosing units

The glass in this unit is strongly degassed (0.002 - 0.016 wt % S), implying a subaerial eruption (e.g., Moore and Fabbi, 1971). However, the depth of emplacement is estimated to be ~545-525 mbsl based on the depth adjustment
Deposits from debris flows contain coarse lapilli to block-sized holocrystalline clasts and ash to fine lapilli sideromelane and tachylite clasts.

An ~18 m thick volcaniclastic underlies unit 286.

Package of undegassed pillow lavas from deep submarine eruptions (~225 m thick)

Deep Water Stage

Active degassed lava flow erupting from a vent above 400 to 500 mbsl

Periodic collapse of oversteepened lava flow initiates debris flows

Figure 4.1d Schematic of processes involved in the formation of unit 286
factor. Therefore, the source lava flow or the volcaniclastic sediment traveled ~2.5 km downslope (assuming an average slope of 14°; DePaolo et al., 2001).

Unit 260 is the uppermost unit of three thick (each ~15 to 30 m thick) volcaniclastics units (260-263). These units are generally similar in thickness, and texture (matrix-supported, poorly-sorted, basaltic lapilli volcaniclastics; Seaman et al., 2000), implying a relatively steady depositional regime. Clast types comprise diverse lithologies. Unit 260 is chemically distinct from the underlying and overlying units (figure 3.14), indicating that the lava source changed. However, contacts between these units are gradational, suggesting that supply was overlapping. The eight units overlying unit 260 are relatively thin (<5 m) sandstones, finer grained (dominantly fine lapilli- and ash-sized) volcaniclastics with seven silty intervals. The abundance of the epiclastic units marks a change in the depositional regime. The finer grain size implies that the supply of clasts declined. Furthermore, there is a reversely graded interval (~35 cm thick) at the base and a normally graded section (~20 cm) at the top of unit 260, reflecting an increase and then decline of the source clast size and/or transport strength.

**Fragmentation processes of the volcanic sediment source**

Most clasts in unit 260 are subrounded (figures 3.7 and 3.8), indicating that the source lava flow was distal, which is consistent with the long transport distance
interpreted from the sulfur contents and emplacement depth. Angular corners of some clasts may reflect breakage during transport. The clasts have a range of shapes (figure 3.8), suggesting fragmentation by erosion rather than autobrecciation or quench fragmentation, which would produce dominantly wedge-shaped lapilli and/or curved sideromelane. In addition, glassy margins are present on only a few clasts, suggesting that the clasts are fragmented from a solidified lava flow.

The sideromelane clasts contain fresh olivines and have uniform thicknesses of palagonitization on their margins (0.1 to 0.3 mm thick), whereas there is a range of olivine alteration within individual tachylite clasts. This suggests that two stages of alteration have affected the clasts (figure 3.6b and c). Alteration of the tachylite probably occurred prior to deposition, whereas the even thickness of alteration of the sideromelane clast margins was post-depositional. The olivine crystals in the tachylite clasts are larger than those in the sideromelane clasts (figure 3.6b). The above differences indicate that more than one lava flow was supplying clasts to this unit. The shallower subunit (260a) contains more altered clasts, so it was probably supplied by an older lava flow. Subunit 260b has more sideromelane and less tachylite than subunit 260a, and a slightly larger range of MgO contents (figure 3.17), indicating that some of the younger pillow lava was quenched at a slightly higher (and larger range of) temperature (1125 to 1156°C) than the older lava (1142 to 1151°C).
Depositional/transport process

There is no discernable macroscopic contact between the two subunits (figure 3.7), suggesting that deposition was continuous. Uneven distribution of coarse lapilli to blocks reaching >13 cm (figure 3.6) and intervals of glassy ash to fine lapilli (Seaman et al., 2000) suggest that transport and deposition was pulsed. The coarse, massive nature of the unit suggests deposition by either a debris flow or a hyperconcentrated flow (e.g., Mulder and Alexander, 2001).

A lull in volcanism/deposition is represented by ~3 thin (<5 mm thick) silty intervals (1832.3 mbsl). This lull was interrupted by a brief production of glassy ash and lapilli (accumulating to ~1 cm thick). The lack of coarse clasts above the fine intervals reflects a weaker supply and/or transport of clasts.

Summary

Unit 260 is a "secondary pillow breccia" composed of epiclastic clasts deposited by pulsating debris or hyperconcentrated flows (figure 4.1e). The source lavas of the volcaniclastics changed with time as erosion fragmented progressively more altered lava (figure 4.1f). This is reflected in changes in the clast types with depth in the volcaniclastic units. This depositional regime continued during the accumulation of ~70 m of pillow lava breccia. Silt deposition was continuous and synchronous with the volcaniclastic deposition, but only formed distinct beds during brief hiatuses of either the erosion or transport of the fragmental material.
Figure 4.1e Schematic of processes involved in the formation of unit 260

- **Continuous settling out of silt from suspension**
- **Degassed volcaniclastic package (~150 m thick) with no interbedded lava flows**
- **Clasts become subrounded**
- **Erosion of solidified, degassed lava flows**
- **Pulsed transport (possibly via debris or hyperconcentrated flows)**

**SHALLOW WATER STAGE**
SUBUNITS ARE DEFINED BY CHANGES IN SOURCE LAVAS (RECOGNIZED BY CHANGES IN CLASTS TYPES)

EROSION EXPOSES AND FRAGMENTS OLDER LAVA FLOWS THAT ARE MORE ALTERED THAN THE YOUNGER FLOWS
4.3.5 Unit 202 (Core depth: 1,485 - 1,414 mbsl)

**Emplacement depths and relationship to enclosing units**

The eruption of the source lava for unit 202 was probably subaerial because the clasts are strongly degassed (0.005 - 0.024 wt % S). Emplacement of unit 202 was probably shallow (~290-240 mbsl) based on the depth adjustment factor. Unit 202 occurs within a ~300 m thick, dominantly volcaniclastic section with thin (<10 m) massive lava flows. Both the underlying and overlying basalt lava flows have similar mineralogy and chemistry to unit 202 (figure 3.14), suggesting a single magma batch supplied the material for all three units. Near the base of the unit 201 lava flow (~75 cm above the base) is an irregular, vertical, glassy contact to a volcaniclastic, suggesting that units 201 and 202 have an irregular or interfingering contact.

**Fragmentation processes of the volcanic sediment source**

Clast shapes are irregular and margins are smooth, indicating that the magma was fluidal upon fragmentation (figure 3.10). The occurrence of sideromelane rims on blocky tachylite clasts suggest that fragmentation occurred during quenching (figure 3.11a).

**Depositional/transport process**

Continuous deposition throughout the accumulation of unit 202 (~70 m) is inferred from its massive nature. The preservation of irregular clast shapes
suggests that transport was short. Furthermore, the occurrence of highly fractured fluidal clasts that have not been disaggregated indicate in-situ fragmentation.

**Summary**

The proximity to the source lava flow, occurrence of in-situ fragmentation of fluidal magma, and predominance of sideromelane in the matrix suggests that unit 202 is an in-situ, “syn-eruptive glassy lapilli breccia” with interspersed intrusions of co-genetic fluidal magma (figure 4.1g). That is, the sideromelane matrix was being produced by the quench-fragmentation of a lava flow surface and the clasts were being deposited at the lava flow front or side. The irregularly shaped, smooth-edged coarser clasts were formed by the intrusion of fluidal magma into the hyaloclastite. The thickness of the unit (~70 m) and lack of more disaggregated blocks of lava suggest that the lava flow was moving very slowly on a gentle slope and most of its progression was in the form of the hyaloclastite. This is not a typical flow front breccia, which is much more dynamic (e.g., faster lava flow, additional fragmentation of flow interior, occurrence of lava tubes).
Figure 4.1g Schematic of processes involved in the formation of unit 202

- **Strongly degassed, subaerially erupted lava flow**
- **Continuous deposition of clasts inhibits formation of bedding**
- **~330 m thick package of volcaniclastics with thin (<10 m thick) interbedded lava flows**
- **Lava is fragmented as it flows. A thick deposit (~70 m) of breccia accumulates.**
- **Irregular shapes with smooth margins indicate fluidal fragmentation**
- **Blocky clast shapes and sideromelane along tachylite clast margins**
- **Highly fractured yet coherent clasts**
4.3.6 Unit 194 (Core depth: 1,311 - 1,287 mbsl)

**Emplacement depths and relationship to enclosing units**

The degassed nature of the glass clasts (0.012 - 0.043 wt % S) in unit 194 suggests that the source lava was erupted within a range of depths from shallow submarine to subaerial. Unit 194 is also estimated to have shallow emplacement depths (~165-150 mbsl based on the depth adjustment factor). The unit is bounded by sparsely vesicular massive basalt flows with similar olivine abundances. There are no major element glass analyses of the lava flows. However, they are thin (<2 m) and material within volcaniclastics that overlie and underlie them (units 192 and 196) have similar chemical composition to unit 194. This evidence and the similar mineralogy between unit 194 and the lava flows suggest that there was a single magma batch supplying the material to all the units from 192 to 196. Although the basal contact of unit 194 is sharp and depositional, its upper contact is transitional into the lava flow, suggesting that unit 194 is associated with the overlying basalt rather than the underlying flow.

**Fragmentation processes of the volcanic sediment source**

The limited range of glass compositions for unit 194 (figure 3.18) indicates that a single flow supplied the clasts. Bimodal vesicularity of the clasts may, therefore, be a result of non-uniform vesicle distribution within the lava flow rather than representing two magma sources with different vesicularities. The overwhelming predominance of sideromelane and occurrence of skeletal clinopyroxenes
indicate that the source lava was quenched extremely rapidly (e.g., Lofgren, 1980), probably subaqueously.

Brittle fragmentation of the lava via cooling-contraction is suggested by the dominance of blocky, equant sideromelane with planar and curvilinear margins (figure 3.11b), and pervasive radial cracking of holocrystalline clasts and glassy margins (figure 3.13). Secondary and/or in-situ breakage during deposition is suggested by jigsaw-fit patterns and the high proportion of cryptocrystalline clasts that are fractured into medium lapilli sized domains but are not disaggregated (figure 3.13). Cracks do not cut the central vesicular zones within the clasts (figure 3.13e) and glass rinds are absent or discontinuous and thin (figure 3.13), indicating that the cracks may have formed by continued expansion of the interior after the margins had solidified. This suggests that the clasts were deposited hot and then cracked due to thermal tension between the cooling and contracting center and the already brittle margins (e.g., Carlisle, 1963). Therefore, this volcaniclastic unit was deposited near an active lava flow.

**Depositional/transport process**

Unit 194 has an uneven distribution of matrix and clasts (figure 3.12) but is otherwise structureless, suggesting that supply and/or deposition was continuous but variable in intensity. The lack of bedding and the inferred proximity of the
deposit to the lava flow suggest transport occurred via tumbling of individual clasts.

**Summary**

Unit 194 is a “syn-eruptive glassy lapilli breccia”. It was derived by quench-fragmentation of the overlying lava flow, which may have flowed from a subaerial vent (figure 4.1h). Although this deposit is only ~150 mbsl, there is no evidence that the clasts were fragmented in a littoral environment. Quench fragmentation of the lava flow was efficient. Substantial autobrecciation and erosion was absent (shown by the lack of blocky and wedge-shaped cryptocrystalline coarse lapilli and blocks). Some clasts did not fully outgas prior to quenching, which allowed late stage fragmentation of brittle rinds around vesiculating and expanding fluidal lava. Post-depositional fragmentation also occurred from the contraction of cooling lava with cold, brittle margins. Downslope transport was short with deposition adjacent to the lava flow.

4.4 Growth of Mauna Kea inferred from HSDP2 volcanioclastics

4.4.1 'Deep' versus 'shallow' submarine

Submarine volcanic deposits are generally subdivided into 'deep' and 'shallow' deposits. Deep deposits may be identified based upon volatile fragmentation depths (e.g., Staudigal and Schmincke, 1984), volatile degassing depths (e.g., sulfur contents >0.09 wt%; e.g., Stolper et al., 2004), or significant changes in
Figure 4.1h Schematic of processes involved in the formation of unit 194

Lava flows into the ocean both passively and explosively (phreatomagmatically)

Slow progression of lava flow allows thick (~24 m) volcaniclastic accumulation

Brittle fragmentation via quenching; secondary and/or in-situ cracking

Some clasts were deposited hot and cracked upon deposition, indicating clasts were deposited adjacent to the source lava flow
lithology (e.g., pillow lava-dominated versus volcaniclastic-dominated deposits; DePaolo et al., 2001).

In the submarine section of the HSDP2 core, there is a sudden change in lithology from pillow-dominant (>35% volcaniclastics) to volcaniclastic-dominant (>80%; DePaolo et al., 2001). This boundary is marked at the top of the shallowest unit identified as pillow lava. Accordingly, the deep to shallow submarine boundary in the HSDP2 core is defined at 1,984 mbsl, which correlates to an emplacement depth of ~650 mbsl based upon the depth adjustment factor.

4.4.2 Deep water facies associations: 3,098 to 1,984 mbsl

The pillow lavas below 2,235 mbsl (830 mbsl emplacement depth) are all undegassed, implying that they were all erupted in deep water. Pillow lavas above 2,235 mbsl are degassed, suggesting they erupted at shallower depths and flowed downslope. Volcaniclastics in the deep water section consist mainly of fragments from degassed lava with undegassed clasts in five units, implying that most material was transported downslope either as a lava flow or in a density current.

Pillow lava flows deposited in the HSDP2 deep submarine stage accumulated thick multi-unit packages (up to 175 m). These packages contain interbedded
layers of thin volcaniclastics (up to 10 m; figure 1.3) such as volcaniclastic units 294 and 293. These formed in-situ and were syn-eruptive (e.g., isolated pillow breccias and hyaloclastites). Therefore, multi-unit packages of lava flows in this stage were emplaced in such quick succession that accumulation of secondary volcaniclastics and other sediments probably between the flows was prevented. However, secondary volcaniclastic deposits (e.g., lava flow breccias; unit 286) occur between the multi-unit pillow lava packages. These volcaniclastics form packages up to 97 m thick. Therefore, eruption and emplacement of lava was periodically vigorous (e.g., either voluminous with relatively high eruption rates, or long-lived with continuous supply of lava at relatively low eruption rates) with intermittent breaks when volcaniclastics accumulated.

Four lithological facies are interpreted from the volcaniclastics examined in the deep water section of HSDP2 the core:

1) IN-SITU HYALOCLASTITES. Thin (<30 cm), structureless, fine grained deposits of equant to shard-like glassy clasts with planar margins. These are derived from the passive cooling-contraction fragmentation of glassy pillow lava rinds (e.g., unit 293 base). Deposition occurs at the flow front of the source lava flow and they are rapidly overrun by the lava.
2) **SYN-ERUPTIVE HYALOCLASTITES.** Thick (<10 m), fine-grained, bedded packages of subrounded glassy clasts transported and deposited by non-cohesive density flows (e.g., unit 294). These deposits can be overlain by an in-situ hyaloclastite derived from the same flow or eruption.

3) **IN-SITU, ISOLATED PILLOW BRECCIA.** Isolated, round lobes of lava deposited in a matrix of in-situ hyaloclastites derived from the same lava (e.g., unit 293). These units are relatively rare and thin, probably because pillow lobes don't travel far before breaking into pillow breccia and the source lava flow rapidly over-runs the depositional site.

4) **SYN-ERUPTIVE LAVA FLOW BRECCIA.** Thick (~40 m), coarse grained beds of cryptocrystalline blocks in a glassy ash matrix (hyaloclastite) deposited by debris flows (e.g., unit 286). Fragments are derived from active lava flows that periodically collapsed, perhaps due to oversteepening.

4.4.3 **Shallow water facies associations: 1,984 to 1,079 mbsl**
This section consists entirely of strongly degassed material, indicating that all the source lavas were erupted in shallow water or subaerially. No pillow lavas were identified in the shallow water stage deposits. This stage is characterized by thin, massive lava flows (<10 m) and thick volcaniclastic units (up to 71 m) and multi-unit packages (up to ~93 m).
The absence of pillow lavas in this section suggests either the coarse, holocrystalline volcaniclastic fragments were derived from the interiors of other types of lava flows, or any pillow lavas that formed were brecciated. Volcaniclastics are formed by a range of processes from erosion to quenching.

Two lithological facies are interpreted from the volcaniclastics examined in the shallow water section of the HSDP2 core:

1) **SECONDARY LAVA BRECCIA.** Thick (~25 m), coarse grained beds of lithologically diverse cryptocrystalline blocks in a glassy matrix. Clasts are derived from the erosion of solidified lava flows. Principal transport processes are debris flows and density flows supported by grain-grain interaction (e.g., unit 260).

2) **SYN-ERUPTIVE GLASSY LAPILLI BRECCIA.** Primary, in-situ hyaloclastite formed by quench fragmentation of active lava flows within relatively shallow water depths. Deposition is adjacent to the active flow with late-stage, in-situ fragmentation of clasts (e.g., unit 194). Small (<10 cm) fluidal intrusions of magma are interspersed with the hyaloclastite (e.g., unit 202).
4.5 Revised model for the submarine growth of Hawaiian volcanoes during the shield stage

The lower ~2 km of the HSDP2 core documents the submarine growth on the flanks of Mauna Kea volcano during its subaerial shield stage. The deposits provide insights into the processes involved in the fragmentation, transport and deposition of volcaniclastic that dominate the submarine section of the core. Assessment of the physical limitations of these processes and how they are affected by shallowing water depths as the flank shoals, permits an evaluation of their occurrence during the entire shield stage of growth for all Hawaiian volcanoes. This assessment is used to estimate the types and proportions of deposits constituting Hawaiian submarine flanks. Therefore, examination of the HSDP2 core has provided a new dimension for assessment of the types, proportions and origins of deposits that comprise the bulk of a Hawaiian volcano that was not available from examining surface exposures (e.g., Stearns, 1946; Macdonald, 1963; Peterson and Moore, 1987).

The dominance of pillow lavas in the lower part of the HSDP2 core and switch to mostly volcaniclastics in the upper submarine section generally agrees with the model of Moore and Fiske (1969) for Hawaiian volcanoes and with field work at La Palma in the Canary Islands (Staudigal and Schmincke, 1984). The prevalence of volcaniclastics is not surprising, considering the tendency for lava
flows to oversteepen on ocean island flanks and the fragile nature of water-quenched lavas.

The deep water volcaniclastics (>650 mbsl emplacement depth) include in-situ and syn-eruptive hyaloclastites, in-situ isolated pillow breccias, and syn-eruptive lava flow breccias (figure 4.2). Volcaniclastics deposited in shallow water (0 to 650 mbsl emplacement depth) include secondary lava breccias and syn-eruptive lapilli breccias. Peperites have not been incorporated into any previous models, but a ~70 m thick deposit of hyaloclastite with small magmatic intrusions (Syn-eruptive glassy lapilli breccia; unit 202) was identified in the shallow water deposits of the HSDP2 core.

All the volcaniclastics within the HSDP2 core are dominated by glassy debris. Previous studies that document glass-dominated volcaniclastics on ocean island volcano flanks attribute their formation to quenching and fragmentation of lava at 'ocean entries' (i.e., where lava flows into the ocean; e.g., Moore and Fiske, 1969; Moore and Chadwick, 1995). However, this study shows that undegassed, in-situ sideromelane remains predominant at emplacement depths of up to 1000 mbsl, indicating that quench fragmentation is also widespread in deeper submarine settings. Furthermore, there is little evidence within the HSDP2 core that any volcaniclastics were supplied from littoral zones, and phreatomagmatic material was not identified. It is possible that ocean-entry deposits were common.
Figure 4.2. Schematic of the internal stratigraphy of a submarine flank of a subaerial Hawaiian volcano. The shallow water deposits (~1000 m thick with emplacement depths <650 mbsl) are dominated by volcaniclastics (~80%) interbedded with thin (<24 m thick) massive lava flows. The deep water deposits are dominated by pillow lavas interbedded with volcaniclastics (~35%).
in the upper ~140 m of the submarine section (i.e., mostly in the section of no-recovery) because emplacement depths were very shallow (<100 mbsl) and any proximal lava flows had probably crossed the shoreline into the ocean. However, explosive, submarine deposits were not identified in drill cores from the rift zones of Kilauea either (e.g., Scientific Observation Hole 1, Quane et al., 2000). This also indicates that the lack of phreatomagmatic material within the core is not related to the location of the site on the flank of Mauna Kea, far from the summit and rift zones where most eruptions occur on Hawaiian volcanoes. Active ocean-entry environments do produce fragmental material, but this must be confined to a relatively narrow zone compared to the vast size of the submarine flanks.

Previous studies speculated that the occurrences of pillow lavas and pillow lava breccias are minor compared to volcaniclastics produced in the littoral zone once the summit of the volcano is subaerial (e.g., Moore and Fiske, 1969). However, the HSDP2 core, which was produced while the summit of Mauna Kea was subaerial, indicates that pillow lavas and pillow lava breccias remain the principal lithologies on the flanks throughout most of the subaerial stage of growth. This study suggests that the volcaniclastics are dominanted by subaqueous, passive magma-water fragmentation (e.g., cooling-contraction) of lava flow rinds at a range of water depths. These fragmentation processes are probably not significantly affected by water depths (i.e., they are pressure-independent).
Therefore, similar processes may occur in submarine environments throughout the submarine and subaerial stages of growth. Proportions of volcaniclastics in the HSDP2 core indicate that shallow water processes of volcaniclastic formation are more vigorous (i.e., there is a higher proportion of fragmented material) than deeper water processes. However, the high abundance of sideromelane and low clast vesicularities in the volcaniclastics indicates that fragmentation is subaqueous and controlled by passive quenching processes. Thus, the increase in volcaniclastic material is not due to an increased supply from littoral zone processes. Furthermore, a change in the eruption dynamics of the volcano is not favored as a cause for the onset of increased volcaniclastic deposition because a similar abrupt change in lithological importance is observed at La Palma (Staudigal and Schmincke, 1984). More likely, an increase of slope gradient on the shallow flanks that may have enhanced fragmentation. Shallow submarine flanks of Hawaiian volcanoes may have steeper gradients due to thrust faults and slumps (Mitchell et al., 2002).

4.6 Conclusions
The HSDP2 core provides the deepest and thickest section of deposits within a Hawaiian volcano currently available (DePaolo et al., 2001). The deposits in the HSDP2 core exemplify the continued importance of submarine processes during the subaerial shield stage of growth of Hawaiian and other oceanic volcanoes. These processes may be similar to those that occur during the submarine shield
stage, although the lavas are erupted at both submarine and subaerial vents during the subaerial stage. Volcaniclastic rocks are prevalent within the submarine section of the HSDP2 core, constituting ~35% of the deeper water deposits (below ~650 mbsl emplacement depth) and ~82% of the shallow water deposits. Deep water volcaniclastics include in-situ and syn-eruptive hyaloclastites (e.g., units 294 and 293) and in-situ isolated pillow breccias (e.g., unit 293) that are gradational into their source pillow lava. Syn-eruptive lava flow breccias (e.g., unit 286) comprise the thicker (~40 m) volcaniclastic units of the deep water section and they may be derived from periodic collapse of flow fronts following oversteepening. Shallow water deposits comprise secondary lava breccias (e.g., unit 260) and syn-eruptive glassy lapilli breccias that include small (<10 cm), fluidal intrusions of magma (e.g., units 202 and 194). This change of lithofacies associations from deep- to shallow-water is abrupt and accompanied by a switch in lava flow type from thicker, pillow lavas to thin massive flows.

Both shallow and deep water volcaniclastics are dominated by clasts produced by subaqueous, passive fragmentation of active lava flows. Other fragmentation processes include autobrecciation, spallation, collapse of oversteepened lava flows, and erosion. The observed wide ranges of formfactor values for HSDP2 clasts may reflect the multiple processes (e.g., quenching versus spallation) involved in fragmentation of the HSDP2 core. However, most units lack clasts with low values typical of the beach deposit, indicating that they were not
exposed for significant periods to high-energy reworking. In addition, major element compositions of glass clasts in some volcaniclastic rocks indicate that each unit was derived from a homogeneous source. This is consistent with the dominance of primary volcaniclastics over secondary origin. The scarcity of epiclastic deposits in the HSDP2 core is probably related to rapid accumulation on the flanks of the volcano during the shield stage of growth.

Deposition of most HSDP2 volcaniclastics appears to have been dominantly proximal to the site of clast production. Therefore, fragmentation of lava at a range of water depths (not just at littoral zones) contributes significant amounts of volcaniclastic material to the volcano flanks. There is no evidence in the HSDP2 volcaniclastics that any significant amount of material was produced in littoral zones. Therefore, fragmental material produced at ocean-entry environments is probably confined to a narrow area proximal to the littoral zone and is not a significant contributor to the bulk of volcaniclastic deposits on the submarine flanks of Hawaiian volcanoes.
6. REFERENCES


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APPENDIX A

Assessment of clast shape analysis

Parameters

Aspect Ratio

A problem with utilizing the aspect ratio is that the maximum and minimum lengths are those of a best fitting ellipse of the clast. Therefore, any clast that is irregularly shaped could yield a low aspect ratio, even if it is actually elongate. For example, a piece of string has a high aspect ratio, but if it is laid down in a non-linear pattern (such as a spiral), the aspect ratio calculated from the best-fitting ellipse would be low. In volcaniclastics, one can relate this example to an elongate fluidal clast (such as Pele's tears or hairs) that becomes twisted during transport and settles on the ground in a deformed form (e.g., Duffield et al., 1977).

Formfactor

Formfactor analysis yields much more varied averages and frequency distribution curves than other parameters (figures 3.19, 3.20, 3.21 and 3.22), and as such is the most successful at distinguishing clasts with different morphologies. However, formfactor values correlate with vesicularity; the unit with the roughest clasts (unit 202) also has the highest vesicularities (<60%), whereas the unit with the smoothest clasts (unit 294) is dominated by non-vesicular clasts. This indicates that the roughness of clasts is dominated by vesicularity rather than
fragmentation processes. Accordingly, vesicularity and mean formfactors increase from clasts fragmented by magmatic exsolution (high involvement of vesicles), through those fragmented by phreatomagmatic activity (range of vesicular involvement), to well reworked clasts (inconsequential vesicularities; figure 3.19). However, the Kilauea phreatomagmatic deposit and HSDP2 units have lower formfactor clasts than the Kilauea 1959 magmatic deposit, even though the magmatic deposit contains clasts with rough, scalloped margins. This may be because the clast shape analysis procedure cannot detect fine-scale roughness. Large-scale (cm-sized) irregularities were not detected either, even though fluidal-shaped clasts with large-scale irregularities were found in the HSDP2 core (e.g., unit 202). This is because the overall shape of the clasts is equant, so the large area overprints most irregularities along the clast margins and, hence, formfactor is not significantly affected. Other clasts with high formfactor have high aspect ratios but smooth margins (i.e., shards), which conflicts with the fundamental principle that formfactor is a measure of margin roughness. Clasts in this category need to be interpreted separately from rough clasts, adding an extra step in the data interpretation process.

The range of formfactors for a deposit derived from magmatic exsolution fragmentation or produced by attrition is narrower than deposits of phreatomagmatic activity (figure 3.19). This reflects the more diverse variations of fragmentation processes involved in phreatomagmatic activity. Similarly, all
the HSDP2 units have wider ranges of formfactor than the end-member samples, which was expected because of the relatively pure fragmentation processes that produced each end-member deposit. However, most of the units do not have clasts that are as smooth as those in the beach deposit, implying that none were reworked in a high-energy environment for any significant period of time (i.e., no unit represents a beach environment). Instead, the ranges extend beyond the lowest formfactor clasts of the magmatic deposit. This is unexpected because none of the HSDP2 samples were predicted to be fragmented purely by volatile exsolution.

**Improvements**

Clast shape analyses parameters may be more diagnostic when used in combination. For example, some of the complications of vesicularity affecting the shape and roughness of clasts could be evaded if solidity and complexity are used together to assess the size and number of embayments along a clast edge.

**Procedures**

The procedures for quantitative clast shape analyses (including obtaining digital images, tracing clast margins, creating binary images, running analyses, performing calculations on the data, and sorting out erroneous results) are laborious and time consuming. Such laboratory procedures are worthwhile if the outcome yields definitive data that can be used to assess a hypothesis or to
highlight features not otherwise detected. However, the type of information gained from quantitative shape analyses (e.g., roughness, roundness and aspect ratio) could be estimated by visual inspection in seconds rather than the many hours needed for the shape analysis procedures. Furthermore, quantitative shape analyses, no matter how advanced, will always need to be compared to visual observations to assess the cause of the shapes. For example: (1) Is clast margin roughness a result of vesicularity, phenocrysts, or fluidity? (2) Are low formfactors a side effect of high aspect ratio? (3) Are round, smooth clasts truly mature or has a glassy margin been spalling off along curved fractures around a clast? 4) Is the clast shape influenced by an internal property such as stress around a phenocryst?

The problem of time efficiency may be accepted if the results are quantitative and accurate enough to provide databases for comparison between different working groups. That is, visual inspection is inherently somewhat arbitrary and thus can create misunderstandings, but quantitative procedures can eliminate these kinds of confusions. In addition, a database of analyses of many eruption styles and locations could prove to be a useful guideline for interpreting fragmentation processes involved in the production of clasts. Once fragmentation processes are identified, eruption environments can be more easily and accurately deciphered. This would be especially useful for ancient deposits and eruptions that occur in remote locations (e.g., deep submarine seamounts). Unfortunately,
utilization of this database would be complicated because volcanic eruptions are so varied and clasts usually undergo more than one process of shape modification (either primary or secondary).