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Hyaloclastites and the slope stability of Hawaiian volcanoes: Insights from the Hawaiian Scientific Drilling Project's 3-km drill core

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Abstract

Core samples recovered during the Hawaiian Scientific Drilling Project (HSDP) drilling project reveal that the upper 1 km of the submarine flank of Mauna Kea is comprised mainly of hyaloclastites. Progressive, very low-temperature alteration of these hyaloclastites has been accompanied by systematic transformations in physical properties of these deposits. Hyaloclastite deposits which directly underlie ca. 1 km of subaerially-emplaced lavas are very poorly consolidated. But over a depth interval of ca. 500 m, compaction and, especially, precipitation of zeolitic, pore-filling cements associated with palagonitization of sideromelane, have eliminated porosity as well as promoted the consolidation of these hyaloclastites. The latter is reflected in unconfined compressive strengths which increase from mean values, respectively, of 2.5 and 4.6 MPa in weakly consolidated, smectite-rich hyaloclastites from the incipient (1080 to 1335 mbsl) and smectitic (1405–1573 mbsl) alteration zones, to a mean value of 10.0 MPa in the more highly consolidated hyaloclastites of the palagonitic zone of alteration (from 1573 mbsl to the bottom of the drill hole). Conversely, overlying, intercalated, and underlying lava flows are generally much less altered, and have mean compressive strengths which are 1 to 2 orders of magnitude greater then hyaloclastites at equivalent depths. The shear strengths of the hyaloclastites also increase with depth and grade of alteration, but are uniformly and substantially lower in the lavas. Those hyaloclastites exhibiting the highest grade of alteration (i.e., palagonitic) also exhibit the highest measured strengths, and thus the alteration of hyaloclastites appears to strengthen as opposed to weaken the flanks of the edifice. However, the contrast in strength between hyaloclastites and lavas may be a primary factor in localizing destabilization, and the zones of weak and poorly consolidated hyaloclastites may facilitate slumping by servings as hosts for shallow detachment surfaces on the unsupported flanks of these volcanoes.

Keywords: Hawaiian; hyaloclastites; alteration; smectite; palagonite; compressive strength; shear strength; edifice stability; Hawaiian Scientific Drilling Project (HSDP)

1. Introduction

In the past few decades, there has been an increased recognition that the submarine flanks of many oceanic volcanoes are largely comprised of large slumps and debris-avalanche deposits (Moore, 1964; Moore and Clague, 1992; Moore et al., 1994; Carracedo, 1996; Lipman et al., 2000, 2002, 2003; Morgan et al., 2000; Garcia and Davis, 2001; Clague and Moore, 2002; Clague et al., 2002). Studies of active volcanoes, e.g., in the Hawaiian Islands, have

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led many workers to agree that the observed slumps and landslides are initiated by the injection of magma along active rift zones and consequent slip along a basal decollement surface at the boundary between the volcanic edifice and the underlying lithospheric plate (Swanson et al., 1976; Dieterich, 1988; Delaney et al., 1990; Got et al., 1994; Denlinger and Okubo, 1995; Iverson, 1995; Walter and Troll, 2003). Others (e.g., Borgia et al., 2000; Morgan, 2006) have hypothesized that spreading on the unsupported flanks of Hawaiian volcanoes may be driven by gravitational forces alone. However, there is much less agreement about how far subaerially-exposed surface faults on the seaward flanks of Hawaiian volcanoes (e.g., the Pali faults of the south flank of Kīlauea) extend at depth (Okubo et al., 1997; Morgan et al., 2000; Cannon et al., 2001; Lipman et al., 2002) and little is known about what factors might affect the localization of faults or slope failures within the flanks of these edifices.

The widespread recognition of giant debris-avalanche deposits on the flanks of subaerial composite volcanoes is also relatively new (prompted by the events at Mt. St. Helens in 1980), although there has been considerably more learned about the processes related to these deposits relative to their submarine counterparts. Although many mechanisms have been proposed for the initiation of large edifice (or sector) collapses on the flanks of continental composite volcanoes, hydrothermal alteration has become an increasingly accepted process for initiating or facilitating these events (Lopez and Williams, 1993; Crowley and Zimbelman, 1997; Watters et al., 2000; Reid et al., 2001; Hedenquist and Stilltoe, 2003; Sisson et al., 2003).

The purpose of this paper is to explore the possibility that alteration may also play a role in destabilizing the flanks of submarine volcanoes. This inquiry arises from the results of studies conducted on the alteration mineralogy of core collected from the Hawaiian Scientific Drilling Project 2, Phase 1 deep drill hole (Walton and Schiffman, 2003). This 3-km-deep hole sampled various lithologies, primarily hyaloclastites and lavas (Fig. 1), which constitute the upper 1.9 km of the submarine flank of Mauna Kea volcano (beneath a cap of 1.1 km of subaerially deposited lavas and tephra deposits). Petrographic study of the HSDP hyaloclastites has revealed a series of progressive alteration zones–termed incipient, smectitic, and palagonitic by Walton and Schiffman (2003)–that appear successively with increasing depth.

The extremely high (>90%) core recovery from the HSDP 3-km hole affords an extraordinary opportunity to make continuous measurements on rock strength through the flanks of a Hawaiian volcano, especially in the context of the proposed alteration zones. Below we present new rock strength data from hyaloclastites and other lithologies from these HSDP cores. These data provide us with the first opportunity to test the effects that variations in alteration and lithology may contribute to the slope stability of Hawaiian volcanoes.

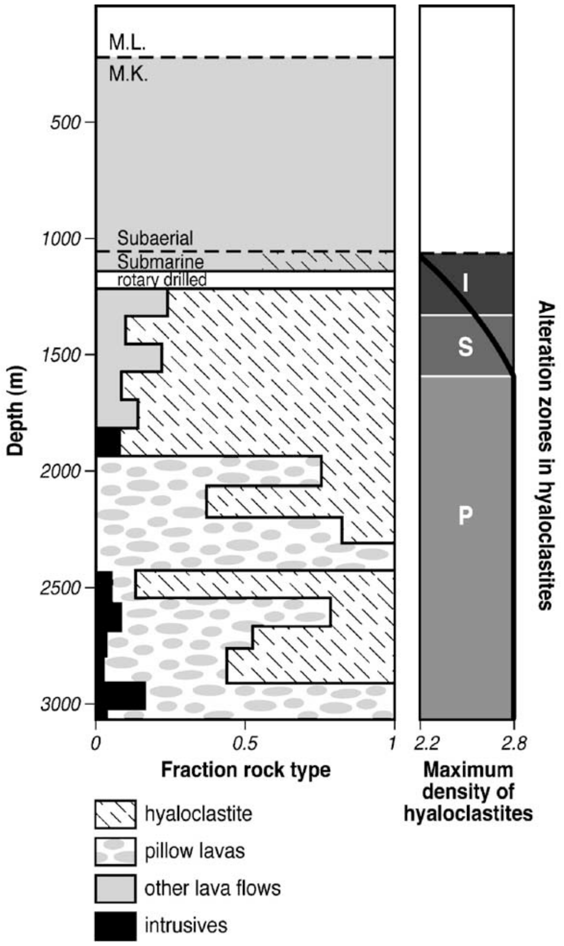

Fig. 1. Lithologic and alteration zones in deposits from Mauna Kea volcano as deduced from the HSDP-2 core. The column on the left (adapted from DePaolo et al., 2000) depicts the major lithologic units in the core hole. The column on the right depicts hyaloclastite alteration zones (after Walton and Schiffman, 2003) and the variation in the maximum density of hyaloclastites with depth (after Moore, 2001). Abbreviations: M.L. = Mauna Loa, M.K. = Mauna Kea, I = incipient alteration zone, S = smectitic alteration zone, P = palagonitic alteration zone.

2. Methods

The samples for this study come mainly from the HSPD 2 Phase 1, 3-km hole that penetrated the southeast flank of Mauna Kea volcano to a total depth of 3.06 km, with a core recovery of almost 95% (DePaolo et al., 2000; Project HSDP, 2001). Hyaloclastite samples from the working portion of the 3.85″ diameter cores were obtained from the former core repository at the California Institute of Technology and the present core repository at the American Museum of Natural History. For comparative purposes, we also examined selected core samples of hyaloclastites from the State of Hawai‘i-1 (SOH-1) Drill Hole, which penetrated the southwest rift zone of Kīlauea volcano (Quane et al., 2000), and we examined volcaniclastic sandstones sampled by submersible during the 2002 Japanese Marine Science and Technology (JAMSTEC) expedition off the south flank of Kīlauea (Sisson et al., 2002; Coombs et al., 2006). Study of petrographic thin sections by light microscopy was augmented with back-scattered electron imaging using a Cameca SX-100 electron microprobe equipped with an energy dispersive spectrometer.

The strength testing conducted on HSDP cores for this study was not designed to thoroughly characterize the rock core behavior but rather to obtain the strength of selected portions of the core that would identify the weakest zone(s) in which failure paths were more likely to develop, thus promoting flank collapse. A comprehensive strength profile of the core would require shear strength and confined (triaxial) compressive testing combined with a quantitative assessment of the effects of the fracturing within the individual zones, and the results could then be used for slope stability modeling studies. This type of detailed strength analysis was outside the scope of our study. However, utilizing unconfined compressive strength testing can show relative strength differences between rock alteration zones and unaltered lavas. Though the unconfined rock strength is not directly applicable to slope stability modeling, in general the lowest unconfined strengths will also have the lowest shear strengths on testing (Goodman, 1989). Consequently the unconfined test results show where shear testing should be performed to obtain the shear characteristics of the zones in terms of cohesion and frictional resistance that can be used in slope stability modeling studies.

The normal unconfined testing method using cylindrical core to obtain the strength was impossible to perform as the 3.85″ core had been split longitudinally for archival and mineralogical purposes. Producing cylindrical core from the existing split core would have required laboratory core drilling to obtain “pencil” cores 19 mm in diameter. This was found to be exceedingly difficult and impossible for the weakest samples as the rock fell apart on drilling. Hence a point load strength testing apparatus (which accommodates irregular rock geometries) was employed, which enabled the split rock core to be tested without further sample preparation and ensured that all the test results were comparable (ISRM, 1985; ASTM D-5731, 2002). The split core was positioned between the testing platens and load applied to the sample. The load and platen separation at failure were recorded, and the results reduced to obtain the unconfined compressive strength. At completion of each test the failure surface was inspected to ensure that failure occurred through intact rock and not along pre-existing major fractures that would yield unrepresentative low strength values relative to the intact rock. A total of 302 strength tests were performed on selected core samples. The number of tests performed on each core sample depended on the length of the core, with the majority of samples providing at least 2 and occasionally 3 strength tests. Sample failure paths followed clast/mineral boundaries or microfractures or passed through the sample matrix. The detailed testing procedures using a point-load apparatus are given in Watters et al. (2000).

Reconnaissance shear strength measurements were also performed on selected samples containing natural rock joints in the HSDP cores. The tests were performed at normal loads equivalent to what the rock joints would experience in situ given the overlying rock load as calculated using average rock density values obtained by Moore (2001). The shear strength procedures for testing rock joints followed the recommendations of ASTM D-5607 (1995). The selected rock core containing a natural rock joint is placed in a mould and both sides of the rock joint encapsulated in a cement material so that both the upper and lower parts of the rock joint are covered by the molding material. The molding cement does not cover the actual joint. The molded rock joint is gently placed into the direct shear machine and the normal load slowly applied. The shear load is gradually and continuously applied, noting the shear displacement with increasing load. At least four different normal loads (stepped loading) are applied which provides four different peak shear loads allowing for calculation of the average friction and cohesion and mechanical behavior of the rock joint for the stress range tested.

3. Results of strength measurements

Physical measurements of HSDP core samples reveal significant differences in their unconfined compressive strength (Table 1 and Fig. 2). The data in Table 1 are presented as the mean (average) and the median of the unconfined compressive strength testing. The point load data are best presented as the median values, as the median strength value is less sensitive to individual extreme strength results. The median strength values illustrate a progressive increase from the incipient (2.4 MPa) to smectitic (4.2 MPa) to palagonitic (9.4 MPa) alteration zones, in which strengths–although very low–appear to double between alteration zones. The incipient alteration zone samples tested reflect the strongest portions of this alteration zone; the weaker samples of this zone could not be tested using the point load as they are essentially cohesionless with only frictional strength.

The strength differences include both differences among hyaloclastites as a function of their alteration, as well as differences between hyaloclastites and the various forms of lava flows and intrusive bodies. Specifically, the unconfined compressive strengths of hyaloclastites from the incipient alteration zone range from 0.0 to 7.0 MPa, hyaloclastites from the smectitic alteration zone exhibit strengths between 0.7 and 10.5 MPa. Hyaloclastites from the palagonitic alteration zone exhibit the greatest strengths of all the hyaloclastites ranging from 1.7 to 28.9 MPa. The unconfined compressive strengths of extrusive submarine and subaerial lavas as well as of intrusive rocks from the HSDP core samples are much greater than that of any of the hyaloclastites, and range from 82 to 150 MPa. The weakest submarine lavas (53 MPa) have nearly twice the unconfined compressive strength of the strongest hyaloclastites from the palagonitic alteration zone.

Results of shear tests performed on natural rock joints from representative lithologies in the HSDP cores are presented in Table 2. The cohesive and frictional strength data for each lithology in this table are the average value from a minimum of 3 sheared rock joints for each lithology. Both the cohesive and frictional strengths of hyaloclastites are uniformly less than those of subaerial and submarine lavas. The cohesive strengths of hyaloclastites increase serially with depth and type of alteration (from 0.9 MPa for the incipient zone to 3.2 MPa for the palagonite zone). Conversely, the frictional strengths of the hyaloclastites show a decrease with increasing depth and alteration zone (from 17.6° for hyaloclastites from the incipient zone to 13.7° for the palagonitic zone). The decrease in frictional strength with increasing depth and overburden stress is normal for rock and soil material. The Mohr strength envelope changes from a linear relationship at shallow depth to become curvilinear with increasing depth and consequently shows higher cohesion and lower friction.

Statistical significance testing was performed on the unconfined compressive test results. No analysis was done on the test values of shear strength as an inadequate number of shear tests were carried out to enable a statistical comparison to be made. The unconfined test results were analyzed using a statistical software package (Elliott, 2004). *P*-value statistical significance tests were run on all the strength results from the unconfined testing, specifically *t*-statistics for testing the mean

Table 1
Unconfined compressive strengths of HSDP core samples

Rock type (n = number of measurements)	Range in estimated unconfined compressive strength (MPa)	Mean (S.D.) estimated unconfined compressive strength (MPa)	Median estimated unconfined compressive strength (MPa)
Subaerial basalt flow (Aa), n = 55	20.0–324.6	133.7 (73.9)	124.2
Subaerial basalt flow (Pahoehoe), n = 22	37.7–249.1	123.3 (53.6)	116.2
Subaerial basalt flow (Transitional), n = 10	2.7–136.3	70.9 (41.7)	82.2
Incipient altered hyaloclastite, n = 41	0.0–7.0	2.5 (2.0)	2.4
Smectitic altered hyaloclastite, n = 45	0.7–10.5	4.6 (2.5)	4.2
Palagonitic altered hyaloclastite, n = 87	1.7–28.9	10.0 (5.0)	9.4
Submarine basalt flow, n = 5	114.4–187.9	138.4 (29.6)	131.7
Submarine massive intrusive basalt, n = 2	116.1–182.7	149.4 (47.1)	149.4
Submarine pillow basalt flow, n = 35	53.1–435.1	185.8 (81.2)	150.3

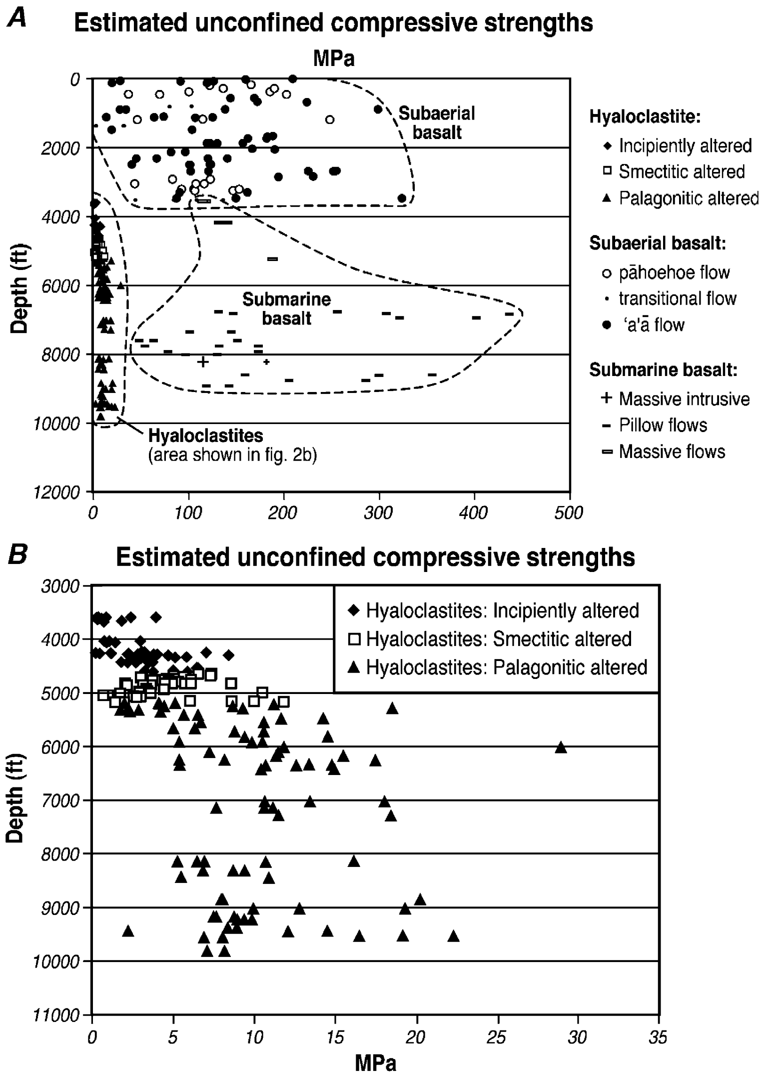

Fig. 2. Unconfined compressive strengths of core samples from the HSDP-2 core hole. (A) All lithologies and (B) only hyaloclastites.

values from the submarine rock assemblages. Two groups of submarine rocks were assessed: the hyaloclastites and basalts. Three types of altered hyaloclastites and three types of basaltic rocks had their mean values compared. Analysis of variance of the means (ANOVA) testing in each group was used to calculate the probability that the differences are not the result of chance but that the results are statistically significant by employing a Neuman–Keuls multiple comparison procedure. Typically a value of $p < 0.05$ is employed to determine if the results are statistically significant.

Table 2
Shear strength data on HSDP core samples

Lithology	Cohesional strength (MPa)	Frictional strength (degrees)
Subaerial/submarine basalts	5.3	20.4
Incipient zone hyaloclastite	0.9	17.6
Smectite zone hyaloclastite	1.5	15.0
Palagonite zone hyaloclastite	3.2	13.7

Comparison of the three groups of the altered hyaloclastites showed that they had p values <0.05, and that the mean strength results from the incipient, smectitic and palagonitic altered hyaloclastites were significantly different from each other. The greatest difference was calculated between the mean value of the palagonitic alteration and the mean values from the smectitic and incipient alteration. As the mean for the incipient alteration was calculated to be significantly less than the means for the smectitic or palagonitic alteration, a further calculation was performed to compare the means only of the smectitic and incipient alteration. This calculation showed that the means had a value of $p > 0.05$ and consequently the means of the incipient and smectitic alteration were not significantly different from each other, though the palagonitic mean remains significantly different from both the incipient and smectitic means. No statistically significant difference was found in comparing the means of the basaltic flows, intrusive and pillow lava values.

4. Discussion

4.1. Alteration and its effect on HSDP rock properties

As outlined by Walton and Schiffman (2003), all hyaloclastites recovered from the HSDP 2, Phase 1 core hole exhibit differing degrees of alteration, and no completely unaltered hyaloclastites have been recognized. Three distinct forms or zones of sequential, very low-temperature alteration (Fig. 1) are characterized by various forms of glass dissolution or replacement textures, as well as by the precipitation of clays and zeolites on the surface of clasts and within primary or secondary pore spaces.

In the incipient zone of alteration, which appears immediately below the contact with overlying, subaerially-deposited lavas (ca. 1080 m depth and extends to 1335 m depth), the external and internal (i.e., vesicle wall) surfaces of mineral and glass shards are ubiquitously coated with a pore-lining form of smectite (Fig. 3A) which may constitute an appreciable volume (i.e., 5–10%) of these hyaloclastites. Although individual shards in hyaloclastites from the incipient zone have experienced some crushing, the extent of compaction is low. The primary porosities in these hyaloclastites are high and may exceed 40% (Walton and Schiffman, 2003).

In the smectitic zone of alteration (from ca. 1335 to 1573 m), smectite replaces portions of sideromelane-rich clasts and also replaces or fills some pores in

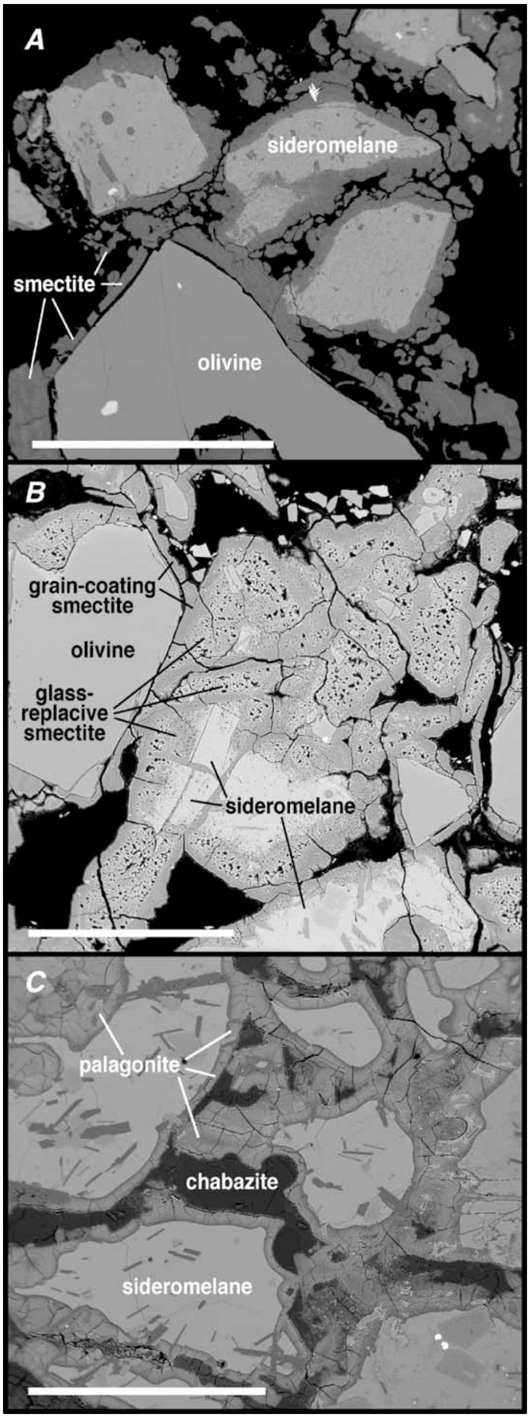

Fig. 3. Back-scattered electron (BSE) images depicting the petrographic characteristics of alteration zones within hyaloclastites in the HSDP-2 core hole. (A) Incipient alteration: highly porous, poorly compacted hyaloclastite in which grains of sideromelane and olivine are uniformly coated with smectite; black regions are open pores (sample R510-7.7; scale bar = 1 mm). (B) Smectitic alteration: porous, moderately compacted hyaloclastite in which sideromelane grains have undergone extensive dissolution and partial replacement by green smectite; the boundary between fresh sideromelane and the glass-replacive smectite is characterized by the presence of high porosity due to dissolution of glass; black regions are open pores (sample R567-1.6; scale bar = 0.5 mm). (C) Palagonitic alteration: non-porous, highly compacted hyaloclastite in which the outer margins of sideromelane grains have been replaced by a rind of yellow, isotropic palagonite; all the primary pore spaces between grains have been filled with grain-coating smectite and a pore-filling zeolitic cement, mainly chabazite (sample R838-15; scale bar = 1 mm).

clasts that have undergone extensive glass dissolution that has produced significant secondary porosity (Fig. 3B). The boundary between fresh glass and smectite-replaced glass is often characterized by extensive arrays of microtubules that contain extant genetic material (Walton and Schiffman, 2003; Fisk et al., 2003) suggesting that glass dissolution may, at least in part, be a microbially mediated process. The porosity of hyaloclastites in the smectite alteration zone is lower than that of the incipient alteration zone, due to increased compaction and in-filling of pores by phillipsite as well as smectite.

In the palagonitic zone of alteration (from ca. 1573 m to the bottom of the core hole at 3109 m), the outer surfaces of glassy clasts have been replaced by rinds of yellow-orange, isotropic palagonite (generally less than 300 μm in thickness). Virtually all pore space remaining after formation of smectite, phillipsite and Ca-silicate has been filled with zeolitic cements, primarily composed of chabazite (Fig. 3C). In these samples, porosity is typically less than 5%.

Alteration of the submarine lavas and minor intrusive bodies that are intercalated with the hyaloclastites of the HSDP core hole has not been studied systematically. Our reconnaissance petrographic examination of these lavas indicates that their glassy components generally exhibit a style of alteration consistent with that observed in adjacent or intercalated hyaloclastites. Otherwise, with the notable exception of various zeolites (mainly chabazite and phillipsite) and other calc-silicates (e.g., gyrolite), which partially fill fractures, and of smectite that partially replaces olivine, HSDP submarine lava flows and intrusive bodies are quite fresh. Therefore, even at depths (i.e., below 1600 m) at which the porosity of hyaloclastites has become negligible through processes of palagonitization and zeolite cementation, pillow lavas have largely retained their primary porosity because their inter-pillow pore spaces have not been extensively sealed with secondary minerals and they are not compacted.

The secondary mineralogical changes observed in the HSDP hyaloclastites are accompanied by systematic transformations of their physical properties. Between the top of the incipient alteration zone and the top of the palagonitic alteration zone, the maximum density of hyaloclastites increases from 2200 to 2800 kg/m^3 (Fig. 1 and Moore, 2001) reflecting both sediment compaction and the progressive infilling of pore space by clays and zeolites. The relative degree of induration and consolidation of the hyaloclastites also increases markedly over this interval.

The styles of alteration described above are apparently not unique to hyaloclastites from Mauna Kea volcano, and may be common to the submarine flanks of actively spreading Hawaiian volcanoes. For example, glassy grains in volcaniclastic sandstones from the active south flank of Kīlauea volcano (Coombs et al., 2006) exhibit alteration textures (Fig. 4) similar to those from both the smectitic and palagonitic alteration zones described from the HSDP 2 core samples. In

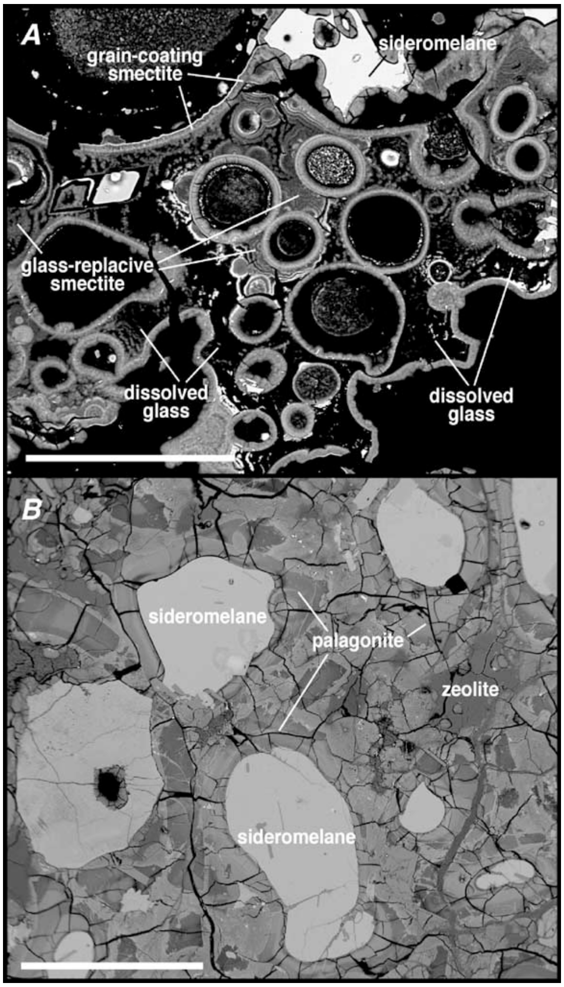

Fig. 4. Back-scattered electron images depicting alteration textures in volcanic siltstones from the south flank of the volcano. (A) Highly porous, poorly-compacted siltstone (JAMSTEC S709-8); large prominent clast has undergone extensive glass dissolution similar to that found within the smectitic alteration zone of the HSDP core samples; some of the glass has been replaced by smectite (e.g., at the middle top of grain), but most of the glass dissolution has resulted in extensive secondary porosity (i.e., the black regions within the interior of the clast); note that – as in the correlative HSDP samples – smectite also coats the exterior surface and vesicle walls of this clast (scale bar = 0.2 mm). (B) Non-porous, highly cemented siltstone (S708-9) whose textures and alteration mineralogy are similar to the palagonitic

sample S709-8, a sandstone that is interbedded with mildly alkalic pillow lavas from the upper slope of Kīlauea's south flank, the grains exhibit extensive glass dissolution and only minor replacement of sideromelane by smectite resulting in development of extremely high secondary porosity. Conversely, in S708-9, a sandstone sample that is from the lower scarp of Kīlauea's midslope bench, palagonitization of sideromelane, with concomitant development of zeolitic pore-filling cements, has resulted in substantial reduction of primary porosity. Similarly, hyaloclastites from the low-temperature (i.e.,<100 °C) portions of the SOH-1 core hole on the east rift zone of Kīlauea exhibit alteration textures similar to HSDP samples: at 800 m in SOH-1, high-porosity hyaloclastites contain fractured, glassy clasts with thin smectitic coatings; at 956m in SOH-1, glassy clasts in very-low-porosity hyaloclastites display 0.1–0.2-mm-thick palagonitic rinds and zeolitic cements.

Alteration of HSDP hyaloclastites undoubtedly occurred at very low temperatures: downhole logs (Thomas, 2000) indicate that temperatures at the bottom of the HSDP 2 core hole do not exceed 45 °C (and have not exceeded this greatly in the past), consistent with the presence of chabazite as the dominant zeolite in pillow lavas at this depth. Temperatures closer to the top of the submarine section, in the zones of incipient and smectitic alteration, have probably not greatly exceeded modern temperatures that range from about 12 °C at the top of the incipient alteration zone (ca. 1100 m) to about 15 °C at the boundary between the smectitic and palagonitic alteration zones (ca. 1600 m). Nonetheless, over this relatively restricted depth and temperature interval, the HSDP hyaloclastites have undergone significant changes in mineralogy, texture, and physical properties.

The most fundamental changes are those in porosity, bulk density, consolidation, and strength. In the incipient alteration zone, primary porosity is high because grain compaction is limited and there is virtually no cementation aside from pore-lining smectite that coats grain surfaces; additionally, glass dissolution has produced some secondary porosity within grains. The measured unconfined compressive strengths of hyaloclastites in the incipient alteration zone are probably maximum strengths because much of this interval could not be cored during the drilling project (DePaolo et al., 2000), and many cores that were collected were just rubble reflecting the lowest intact strength. The samples that were studied came from the most intact parts of the core. In the smectitic alteration zone, compaction is greater, phillipsite and smectite partially fill primary pores, and although glass dissolution can be profound, secondary pores are partially filled with phillipsite and smectite. In the palagonitic alteration zone, primary pore space between highly compacted grains not previously filled by smectite, phillipsite, or Ca-silicate has been cemented by chabazite. Grain volume of hyaloclastites was about 55% when they formed; grain volume, not including volume of intergranular cements, of palagonitized hyaloclastites has risen to about 75% through compaction (Walton and Schiffman, 2003). These changes in compaction and porosity are reflected in the increased maximum measured density of hyaloclastites (Moore, 2001) and in the increased compressive strengths.

The marked increase in the degree of consolidation of the HSDP hyaloclastites from the incipient through the palagonitic alteration zones is reflected in the data from the measurements of unconfined compressive strength. In the incipient alteration zone, alteration is restricted entirely to a thin (generally <0.02 mm) smectite coating on grains, which imparts a very limited amount of compressive strength (i.e., <2.5 MPa). In the smectitic alteration zone, initial grain cementation is augmented by precipitation of phillipsite and smectite in primary and secondary pores, which marginally increases cementation and compressive strength (to a maximum of 4.6 MPa). The mean unconfined compressive strength of HSDP hyaloclastites more than doubles between the smectitic and palagonitic alteration zones. In the latter zone, cementation is mainly the result of the filling of remaining pore space with chabazite rather than smectite. In addition, cementation appears to have been effected where the palagonitized rinds of adjacent glassy grains have "fused" together (e.g., in the right-hand side of Fig. 3C).

The alteration mineralogy, porosity, degree of consolidation, and strength of the HSDP hyaloclastites vary systematically within the submarine flanks of Mauna Kea volcano. The poorly consolidated, highly porous and weak hyaloclastites of the approximately 500 m-thick, combined incipient and smectitic alteration zones lie immediately below subaerially deposited lavas whose compressive strengths are 1 to 2 orders of magnitude higher. Within the underlying palagonitic alteration zone, the hyaloclastites are much more highly consolidated and less porous, and their relative strength contrast with intercalated pillow lava flows and intrusive bodies is less marked. However, the intercalated submarine lavas are more porous than the palagonitized and zeolite-cemented hyaloclastites, and they apparently serve as deep aquifers for down-welling fresh water lenses beneath Hawai'i (Thomas, 2000). Measured fluid pressures in fresh water aquifers at approximately

2926 m depth exceed local hydrostatic pressures by 1.2 MPa (Don Thomas, personal communication, 2004). Therefore, the high-porosity pillow zones represent overpressured zones intercalated with low-porosity, palagonitized hyaloclastites.

4.2. Strength of HSDP core materials relative to lithologies from other volcanic edifices

To our knowledge, no other data exist on either the compressive or shear strengths of the fresh or altered materials underlying oceanic volcanoes. However, such data have recently been obtained for volcanic rocks from composite volcanoes (e.g., in the Cascade Range, by Watters et al., 2000) that also exhibit evidence of edifice collapse. No data sets comparable to the one presented here (i.e., which provides a complete transect through the flanks of an individual edifice) are available. Rather, the available data come from samples collected during traverses across the surface of the edifices. Measurements on lavas and tuffs from Mount Rainier, Washington, and Mount Hood, Oregon, indicate that the strength of these materials is highly dependant on their extent of alteration (Watters et al., 2000). On Mount Rainier, least altered andesitic lavas have unconfined compressive strengths averaging 146 MPa, very similar to mean values of both subaerial and submarine lavas from the HSDP cores (Table 1). Moderately altered tuffs from Mt. Rainier have unconfined compressive strengths averaging 32 MPa, roughly 3 times greater than palagonitized hyaloclastites from the HSDP core. Highly altered, clay-rich matrix from block-and-ash-flow deposits from Mt. Rainier has unconfined compressive strengths (averaging 1.4 MPa) that are comparable to the weakest, incipient alteration zone hyaloclastites from the HSDP core.

The strength measurements from the HSDP core are more representative of vertical strength variations within the volcano than those obtained from the Cascade volcanoes where internal lithology and structure are essentially unknown. The Cascade volcanoes have been shown to contain appreciable clay-rich rocks from inspection of landslide debris and in situ surface observations and geophysical measurements (Finn et al., 2001), though the lateral and vertical extent of the argillic alteration in the edifices is poorly known and therefore it is problematic in utilizing the strength data for stability studies. In contrast, the locations of weak zones within oceanic volcanoes are better known and strength testing results from these zones should lead to superior slope stability appraisals.

4.3. Lithology, alteration and implications for slope stability on Hawaiian volcanoes

The alteration mineralogies and physical properties of hyaloclastites and lavas described from the HSDP 3-km-core hole may provide some insights into the initiation and probable localization of slope failures within the flanks of actively spreading, Hawaiian shield volcanoes. Because shield volcanoes have such gently sloping flanks (<20°), gravitational slope failures are likely to occur principally in poorly compacted sediments weakened by high pore-fluid pressures or alteration, and failures due to magma injection along active rift zones will only occur if the frictional resistance to sliding is anomalously low (Iverson, 1995). HSDP hyaloclastites, especially those in the incipient and smectitic alteration zones, are poorly compacted and contain grain-replacing and/or grain-coating smectite that may account for up to 10% of the rock volume (Fig. 3A and B). Recent shear testing has demonstrated that smectite-rich sediments possess extremely low coefficients of friction (i.e., 0.11) at normal stresses between 10 and 30 MPa (Kopf and Brown, 2003). Thus the incipient and smectitic alteration zones in the HSDP hyaloclastites may be the sites of potentially unstable conditions where excess pore pressures can exist. Pore pressures can temporarily increase as a result of seismic activity or additional sediment load and assist in slope destabilization. Although hyaloclastites of the palagonitic zone of alteration have higher strength contents than overlying hyaloclastites, the pillow lavas intercalated within them apparently possess the permeability to allow circulation of freshwater – at pressures exceeding hydrostatic – deep within the volcano. Garcia and Davis (2001) have suggested that the interlamination of pillow lavas and poorly lithified hyaloclastites should make the flanks of oceanic islands more stable than if they were underlain solely by poorly lithified hyaloclastites. They further suggested that the presence of pillow lavas within the hyaloclastites material should limit the size of landslides or other failures of the island flanks. However, if fluids circulating through pillow lava units are overpressured, the potential for slope failure may be enhanced rather than reduced.

4.4. Hyaloclastites, faulting, and slumping on the flanks of Hawaiian volcanoes

Although no evidence for the existence of faulting (e.g., slicken structures and gouge material) has been described from any lithologic units within the HSDP core, it is not unreasonable to speculate on the role that hyaloclastites might serve in accommodating strain on the flanks of Hawaiian volcanoes. Fig. 5 is a model cross-section (after Moore and Fiske, 1969, and informed by the results of the HSDP 3-km drill hole) through the edifice of an actively spreading Hawaiian shield volcano. Steep, listric normal faults delineate the pali or cliff structures along the unsupported, seaward flanks of these volcano, although the depth to which these faults penetrate is open to debate. Existing data on earthquake hypocenters (e.g., from: Got et al., 1994; Denlinger and Okubo, 1995) indicate that much of the strain produced by rift zone intrusion and gravitational spreading is accommodated on the 7–10 km-deep, subhorizontal basal detachment which separates the base of Hawaiian volcanic edifices from the underlying lithosphere. Nonetheless, seismic reflection data (Morgan et al., 2000) as well as continuously recorded GPS data (Cervelli et al., 2002) indicate that slip events have also occurred at shallower depths, most likely along listric, normal faults which may roll over into shallow-dipping thrust faults.

If weak, altered hyaloclastites play a role in destabilizing the flanks of these volcanoes, then they would most likely be destabilizing the normal fault zones. Analysis of the 1975 M7.2 Kalapana earthquake by Cannon et al. (2001) suggests that observed ground deformation is best explained by slip on the Hilina (listric, normal) fault between 2 and 3 km of depth. They speculated that at this depth, the Hilina fault might shallow into a decollement surface within hyaloclastites. This depth corresponds well with the depth where hyaloclastites are encountered beneath subaerially deposited lavas in the HSDP core hole. Inversion of GPS data that recorded a sudden aseismic slip on Kīlauea's south flank in November, 2000 (Cervelli et al., 2002) suggests that similar shallow angle thrusts may exist at 4–5 km depth.

We suggest that shallow rotational slumps that occur within the upper few kilometers of spreading Hawaiian volcanoes may be initiated within low strength, poorly consolidated, smectite-rich hyaloclastites similar to those we have described from the incipient and smectitic alteration zones of the HSDP cores. Conversely, deeper slumps might be directed through overpressured pillow lava units as a result of the stronger pillow lava units permitting deeper failure surfaces to develop. Overthrusts, which have been recognized on the edge of the outward flanks of actively spreading volcanoes such as Kīlauea (Morgan et al., 2000) and Mauna Loa (Morgan and Clague, 2003), might also be directed up through hyaloclastites of the incipient and smectitic alteration zones (Fig. 5).

Hyaloclastites having the alteration characteristics of the HSDP palagonitic zone have been recovered from outcrops on the south flanks of Kīlauea (e.g., Coombs et al., 2006) and Hualālai (Hammer et al., 2006). If the depth-dependent distribution of alteration zones described from the HSDP core hole is a feature common to the flanks of actively spreading Hawaiian shield volcanoes–and the recognition of these alteration zones in the SOH-1, Kīlauea core samples implies that it may be–then the presence of palagonitized hyaloclastites on the seafloor requires exhumation of at least 500 m of the submarine flank of the volcano. Although deeper slumps may bring these hyaloclastites to the surface, thrust faults might also be responsible. If the frontal ridges of the seaward flanks of volcanoes are underlain by a series of imbricated nappes (e.g., as has been proposed for the west flank of Mauna Loa by

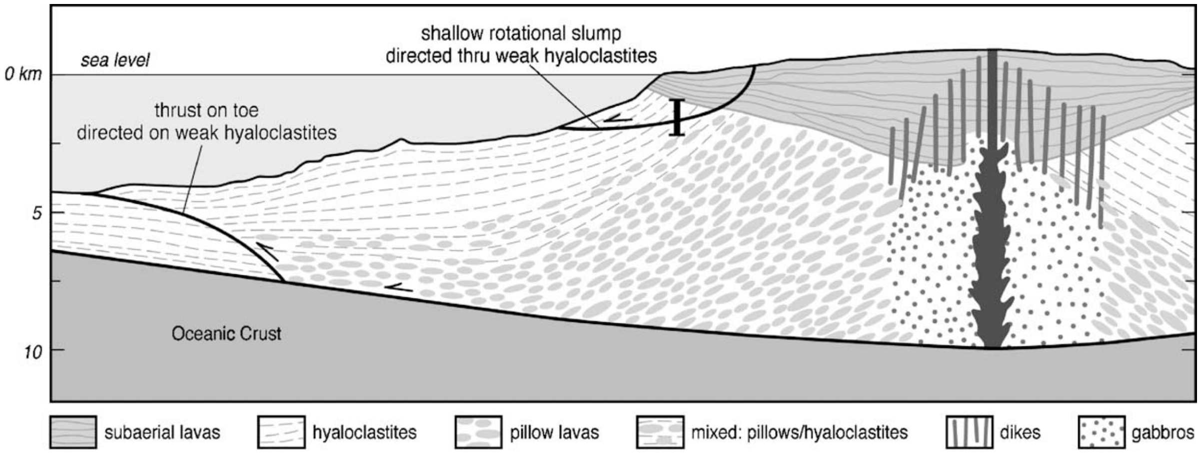

Fig. 5. Model (after Moore and Fiske, 1969 and partially constrained by results from the HSDP 3-km-drill hole) depicting potential effects of rock alteration and lithology on slope stability/failure within the seaward flank of an actively spreading Hawaiian shield volcano. The bracketed vertical bar depicts the approximate depth where hyaloclastites were the main lithology encountered in the HSDP core hole.

Morgan and Clague, 2003) then hyaloclastites within each of these nappes should exhibit an alteration zonation that was imprinted prior to thrusting. As such, alteration zones in hyaloclastites may be useful markers for recognizing thrust sheets and testing models of slope failure on the flanks of Hawaiian volcanoes.

5. Concluding remarks

Strength measurements on hyaloclastites from the HSDP 3-km-core hole indicate that these are very weak materials. Petrographically, these Mauna Kea hyaloclastites appear similar to those from actively spreading Hawaiian shield volcanoes, based on reconnaissance level examination of samples from the East Rift Zone of Kīlauea (i.e., the SOH core samples) and the seafloor on the south flank of Kīlauea (i.e., JAMSTEC dive samples). Alteration processes apparently affect the strength of these hyaloclastites. In the shallower zones of incipient and smectitic alteration, hyaloclastites generally retain their high primary porosities, can develop additional secondary porosity (through glass dissolution) and have low degrees of consolidation. Ubiquitous smectite grain coatings found in these hyaloclastites might further reduce the frictional angle of these deposits, although in the absence of strength data on unaltered hyaloclastites, this has not yet be demonstrated. In the deeper, palagonitic zone of alteration, the hyaloclastites gain both compressive and shear strength, primarily through consolidation and zeolitic cementation. The strength contrast between hyaloclastites and the lavas that overlie and underlie them is dramatic, and may be a primary factor in localizing the destabilization of the flanks of Hawaiian volcanoes.

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