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Pleistocene snowlines and glaciation of the Hawaiian Islands

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Abstract

High volcanoes on two of the southeastern Hawaiian Islands experienced Pleistocene ice-cap glaciations. Drifts of three glaciations on the upper slopes of Mauna Kea (4206 m) are interstratified with lavas. During marine isotope stage (MIS) 2, an ice cap (70.5 km^2) formed when the snowline fell ca. 930 m to an altitude of ca. 3785 m. Drifts of MIS 4 and 6 age, largely buried by hawaiite lavas, record a snowline about 100 m lower. Holocene lavas bury the upper slopes of Mauna Loa (4169 m), the summit of which likely intersected the snowline during the last glaciation. The reconstructed history of snowline variations and glaciation on Maui's Haleakala (3055 m) relies on evidence of the island's growth, isostatic subsidence, and tilting. The snowline may first have intersected Haleakala about 800,000 years ago; ice caps waxed and waned during the next ca. 400,000 years (MIS 22-12) until subsidence and diminishing eruptive activity brought the summit below the snowline. Erosional morphology and thick diamictons indicative of subglacial eruptions are consistent with repeated summit glaciations. Whereas Haleakala's glacial history has ended, ice caps likely will regenerate on Mauna Kea ca. 60,000 and 100,000 years in the future, after which the summit will subside below the glacial-age snowline.

1. The island of Hawaii

The Hawaiian Islands include the only high land areas in the vast central Pacific Ocean basin that retain evidence of former glaciation. None of the volcanoes comprising these islands rise high enough to intersect the modern (interglacial) snowline, but several summits lay above the snowline during the last and earlier glaciations, leading to the formation of mountain ice caps. The island of Hawaii, which is the southernmost and highest of the archipelago, provides a record of Pleistocene glaciation that has been studied since early in the last century (see discussions and references in Porter, 1979a, b). Mauna Kea and Mauna Loa volcanoes are the only two summits that now rise high enough to intersect the full-glacial snowline (Fig. 1). In addition, Haleakala, one of two volcanoes comprising the island of Maui, may contain a lengthy record of earlier glaciations, about which few details are known.

1.1. Mauna Kea volcano

Mauna Kea (19°46.5′ N Lat., 155°29′ E Long; 4206 m.), the highest summit in the Hawaiian Islands, is the only volcano with unequivocal evidence of Late Pleistocene glaciation (Porter, 1979a, b). Exposed rocks span an interval from ca. 250,000–4500 years ago. These rocks have been divided into two lithic units (Wolfe et al., 1997). The Hamakua volcanics (ca. 250,000–65,000 yr old), consisting primarily of alkali and transitional basalt, form the surface over the intermediate and lower slopes of the volcano. They also are exposed to a depth of ca. 75 m in deep gullies on the southwestern slope of the mountain, as well as in sea cliffs along its northeastern landward margin. The Hamakua flows are overlain by the Laupahoehoe volcanics (ca. 65,000–4000 yr old), which are dominated by hawaiite lavas and associated pyroclastic deposits.

1.1.1. Glacial geology

Moraines of the last (Makanaka) glaciation delimit a ca. 10-km-diameter ice cap that mantled the upper

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slopes of the mountain (Fig. 1). They extend farther downslope (to ca. 3420 m altitude) on the southeastern side of the mountain than on the southwestern (3465 m), northeastern (3510 m), or northwestern (3570 m) sides. The moraines locally are overlain by postglacial lava flows and tephra cones, and overlie volcanic rocks and intercalated deposits related to two earlier glaciations.

Moraines of the penultimate (Waihu) glaciation are exposed discontinuously along the southwestern flank of the mountain downslope from the Makanaka moraines (Fig. 2). Their more-subdued morphology points to substantial modification by post-depositional slope processes. The eroded moraines have an average lower limit of ca. 3000 m, or ca. 465 m below the average Makanaka ice limit. In places, the moraine belt is crossed by Laupahoehoe hawaiite flows from upslope; elsewhere on the mountain the moraine system has been completely buried by such flows.

Pokahuloa drift is well exposed in two deep gullies that indent the southern slope of the volcano above the Mauna Loa-Mauna Kea saddle, as well as in a small kipuka on the upper east slope. This deposit consists predominantly of basaltic lithologies and lies near the top of the Hamakua lavas. Because the drift has limited exposure, the configuration of its limit is not known. However, in the two gully exposures the till facies appears to reach approximately the same downslope limit as the overlying Waihu moraines. These relationships suggest that the two ice caps may have been of similar size.

Several Laupahoehoe lava flows lying inside the Makanaka moraine belt have features indicating that they were erupted through glacier ice, e.g., abnormally high and steep margins, pillow structures with radial fractures and hyaloclastite, spiracles, and glassy flow surfaces, all pointing to rapid cooling in contact with ice or meltwater (Porter, 1987).

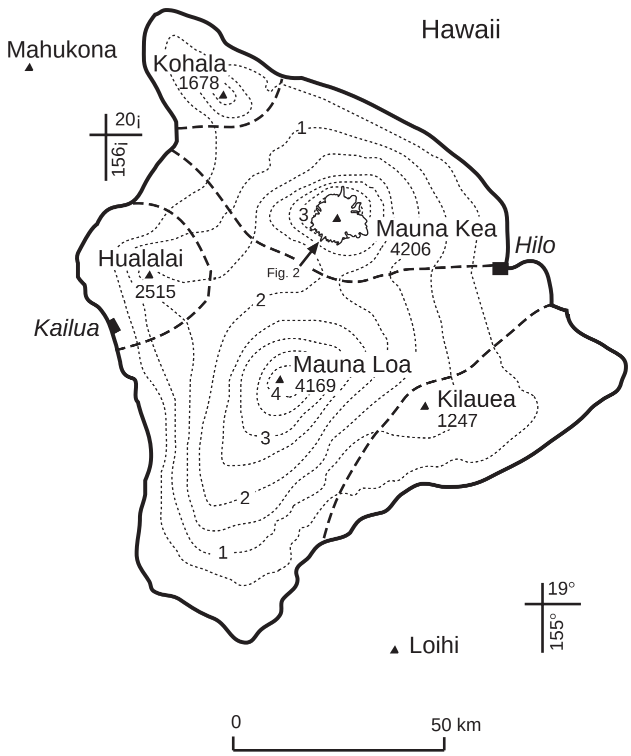

Fig. 1. Map of Hawaii showing location of five volcanoes that comprise the island (contours interval 10^3 m). The summits of Loihi and Mahukona volcanoes lie below sea level. Only Mauna Kea has an exposed stratigraphic record of past glaciations, but Mauna Loa may have had an ice cap during the last glaciation. Extent of the last (Makanaka) ice cap is outlined on Mauna Kea.

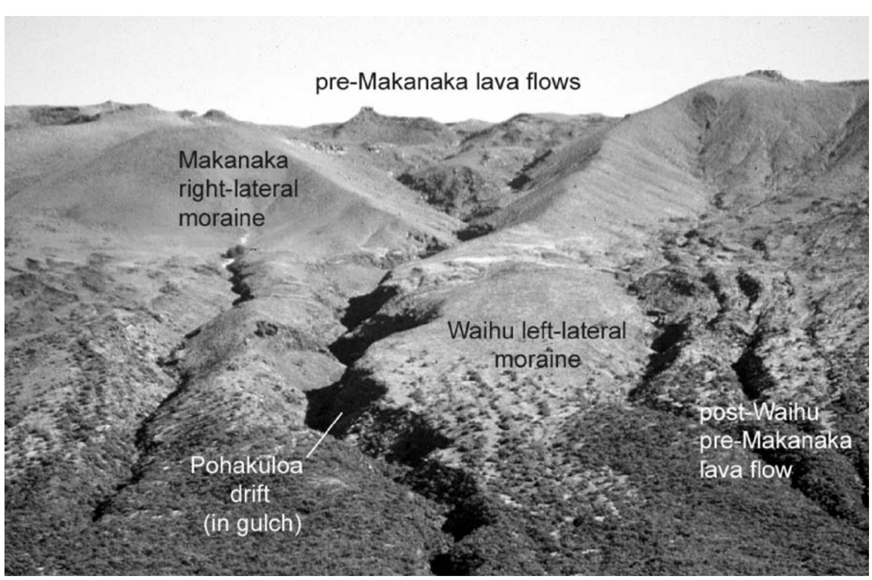

Fig. 2. View looking northeast toward the summit of Mauna Kea from the Mauna Kea-Mauna Loa saddle showing moraines of Makanaka and Waihu glaciations and relationship of drift units to interstratified postshield lavas. Width of the view at the moraine belt is ca. 2 km. Flow of Makanaka and Waihu glaciers was toward the viewer.

1.1.2. Chronology of glacial drifts

The chronology of glaciations on Mauna Kea is based largely on K/Ar dates of lava flows that overlie and underlie successive drift units (Fig. 3), and on several surface-exposure (^{36}Cl) ages of lava flows and moraine boulders. Dates of hawaiite lavas constrain the Makanaka drift between 31,000 ± 9000 and 18,000 ± 10,000 K/Ar years (Wolfe et al., 1997). A boulder on the outer Makanaka moraine system has a reported surface-exposure age of 18,300 ± 200 ^{36}Cl years (Dorn et al., 1991). These provide a minimum age for moraine stabilization and imply an age for the greatest expansion of ice during the last glaciation of ca. 20,000 years, within MIS 2 (Martinson et al., 1987). A ^{36}Cl exposure age (14,700 ± 500 yr) for a boulder upslope from the innermost moraines suggests that the summit was deglaciated by ca. 15,000 yr ago. This age is consistent with ^{14}C dates of 13,000–14,000 ^{14}C yr B.P. (ca. 15,000–16,000 cal yr B.P.) from basal sediments in Lake Waiau (3971 m), located in a crater near the volcano's summit (Wolfe et al., 1997).

Poorly constrained Waihu drift is overlain by lavas with K/Ar ages of 167,000 ± 32,000 to 70,000 ± 3000 years and overlies lavas as young as 121,000 ± 4000 years (Wolfe et al., 1997, Fig. 11). Although some of the dates are stratigraphically inconsistent, Wolfe et al. suggest that the age of the drift likely is in the range of 100,000–150,000 years (i.e., MIS 5 and 6), but could be as young as 70,000 years (MIS 4). Dorn et al. (1991) preferred a more-restricted age range of 60,000–70,000 years based on a minimum limiting ^{36}Cl age of 63,000 ± 2300 yr for a Waihu moraine boulder and on rock-varnish cation-ratio dates of similar age.

The Pohakula drift is less-tightly constrained because of large 1-sigma error ranges of the K/Ar dates (±13–25%) that result from a low K content of the bounding lavas. Available K/Ar dates suggest that this unit is older than ca. 100,000–150,000 years, but younger than 150,000–200,000 years, and therefore, likely correlates with MIS 6 (Dorn et al., 1991; Wolfe et al., 1997).

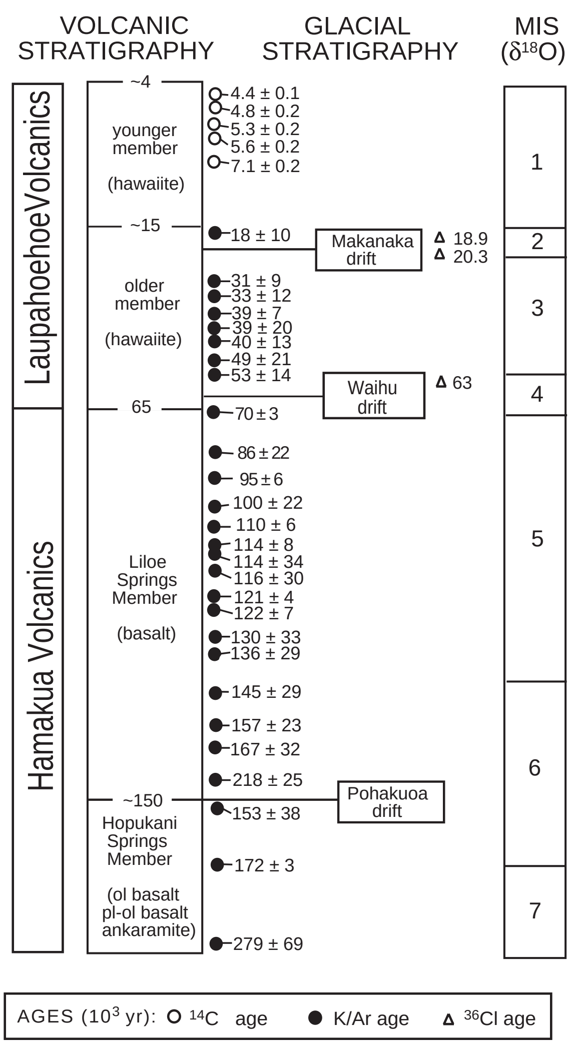

Fig. 3. Volcanic stratigraphy and chronology of postshield volcanics and approximate relationship of the three exposed drifts to dates of lavas. K/Ar and ^{14}C dates are from Wolfe et al. (1997, Tables 4–6 and Fig. 11). The three ^{36}Cl dates are from Dorn et al. (1991). Ages of marine oxygen-isotope stage (MIS) boundaries are from Martinson et al. (1987). Dating uncertainties are $\pm 1\sigma$.

1.1.3. Ice-cap reconstructions and paleo-ELAs

The reconstructed configuration of the Makanaka ice cap at the last glacial maximum (LGM) is based on long profiles of the glacier oriented along flow lines (Porter, 1979b, Fig. 5). The crests of several cinder cones rose above the glacier surface, and the minimum height of the glacier against their flanks is marked by hawaiite erratics. Other cones were overridden by ice, and both their flanks and summits were glacially eroded. The profiles were constructed using moraines that mark the glacial limit, the height of overridden cinder cones, and the upper limit of erratics on the cinder cones. The topography of the ice cap is depicted in Fig. 4 by contours at 100-m intervals (Porter, 1979b). The area of the glacier was ca. 70 km^2, and its average thickness, based on the difference between the reconstructed ice-cap topography and present topography, was ca. 70 m; the thickest ice slightly exceeded 100 m.

The full-glacial equilibrium-line altitude (ELA) was derived using the Accumulation-Area Ratio (AAR) method and an assumed steady-state AAR of 0.6 ± 0.05 (Porter, 1979b) (Fig. 5). To evaluate a possible ELA gradient, the glacier was divided into four sectors (NW, NE, SW, and SE), and the average ELA was calculated for each. The ELA was interpolated between sectors to derive an ELA for the ice cap as a whole (bold dashed line in Fig. 4). Contours defining the ELA surface across the summit region were then obtained by connecting equivalent ELA values on opposite sides of the ice cap. The results indicate that the ELA was about

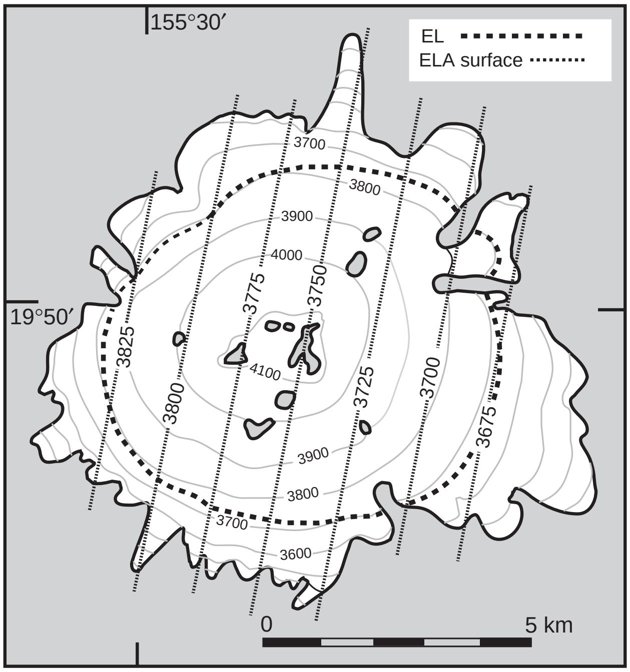

Fig. 4. Map showing extent and surface topography of Makanaka ice cap at the last glacial maximum about 20,000 years ago (Porter, 1979b). Bold dashed line represents reconstructed equilibrium line (EL) of ice cap, and dotted lines show east-southeast-sloping gradient of the equilibrium-line altitude (ELA) surface across the glacier.

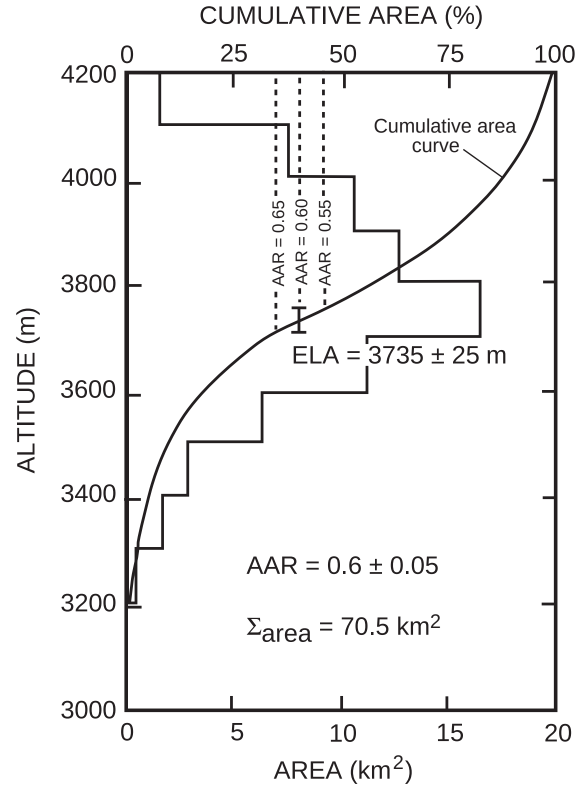

Fig. 5. Area vs. altitude histogram and cumulative curve for the Makanaka ice cap at the last glacial maximum (LGM). The former represents the area of ice cover, at 100-m intervals, between 3200 and 4200 m altitude. The superimposed curve shows the cumulative percent of these values. The ELA of 3735 ± 25 m was calculated using an accumulation-area ratio (AAR) of 0.6 ± 0.05. The ice cap had an area of 70.5 km^2.

125 higher in the NW sector than in the SE sector, and at intermediate levels in the other two sectors. The ELA surface sloped across the ice cap toward the east–southeast with a gradient of about 20 m/km, which is consistent with the greater downslope extent of moraines on the SE side of the mountain compared to that on the NW side. The average reconstructed ELA for the ice cap, without a correction for island subsidence (see below), is 3735 ± 25 m (Fig. 5; all altitudes in this paper are rounded to the nearest 5 m). At this low latitude, insolation varies little across the upper slopes of the volcano. Therefore, the ESE-sloping ELA gradient is interpreted primarily as a response to precipitation, implying a primary ESE moisture source during the glacial maximum.

A full reconstruction of the Waihu ice cap is not possible because its limit is known only for its SW sector and no data are available that permit assessment of its thickness upglacier from end moraine remnants. Nevertheless, the general extent and morphology of the SW sector of the ice cap can be inferred based on the assumption that the glacier's steady-state profile was similar to that of the Makanaka ice cap. Using the same AAR, the derived ELA for the Waihu glacial maximum was 3510 ± 35 m, or about 225 m lower than at the last glacial maximum (LGM) for this sector of the glacier; this value also is uncorrected for island subsidence.

The Pohakuloa ice cap cannot be reconstructed. However, based on the downslope extent of the exposed sediments, its limit on the SW slope of the volcano lay close to that of the Waihu ice cap. It, therefore, may have had somewhat similar dimensions. In this case, the average Pohakuloa ELA may have been close to that of the Waihu ice cap.

1.1.4. Effect of island subsidence on ELA estimates

The huge volume of lava erupted during the growth of a Hawaiian shield volcano imposes a great load on the ocean crust, causing it to subside. The altitude of a volcano at any time, therefore, reflects the difference between volcanic accretion (positive) and subsidence (negative). During its glacial history, Mauna Kea was completing its accretionary phase, while adjacent young volcanoes (Mauna Loa and Kilauea) were in their shield-building phases. As a result of this extrusive activity, the island of Hawaii subsided rapidly during the interval encompassing Mauna Kea's successive glaciations.

Reported estimates of the subsidence rate for Hawaii range between 2.3 and 2.7 mm/yr. The rate adopted here, 2.6 ± 0.4 mm/yr, is that proposed by Ludwig et al. (1991) for the submerged northwestern sector of the island, which they infer has subsided at a relatively constant rate during the last 463,000 years. This sector likely experiences the highest rate of subsidence on or around the island because it lies along the axis of the Hawaiian Ridge (Wolfe et al., 1997, Fig. 1). Moore and Clague (1992) suggest that this rate is applicable to the island as a whole.

The rapid subsidence rate would have influenced both the summit altitude and the relative position of the snowline during glaciations. Using this subsidence rate, the summit at the LGM (ca. 20,000 years ago) would have been ca. 52 ± 8 m higher than now (i.e., ca. 4260 ± 8 m). The full-glacial ELA (ca. 3735 ± 25 m), when corrected for average subsidence of 2.6 mm/yr during the past 20,000 yr, would have been at 3785 ± 25 m.

Using the same average subsidence rate for the early Laupahoehoe interval, the summit of the volcano would have stood 170 m higher (i.e., 4375 ± 25) during the Waihu glaciation (ca. 65,000 yr ago). The related ELA would have been at ca. 3680 ± 25 m (i.e., ca. 100 m lower than the Makanaka ELA).

Values for summit altitude and ELA during the less-well dated Pohakuloa glaciation are more speculative. Assuming an age of 133,000 yr (see below) and an average subsidence rate of 2.6 mm/yr, the summit would have been ca. 345 m higher than now (i.e., ca. 4550 ± 25 m).

Not included in these estimates is the addition of lava to the summit region. Post-Pohakuloa lavas cover the summit and locally are tens of meters thick on the upper slopes of the volcano. A reasonable estimate would add an average of up to 100 m of Laupahoehoe lavas and cinder cones to the summit region over the last 65,000 years, requiring an appropriate (positive) adjustment to estimated altitude values.

1.1.5. ELA depression during the LGM

Because the summit of Mauna Kea lies below the modern snowline, ELA depression during the Makanaka glaciation cannot be determined directly. However, it was at least 420 m at the LGM (the difference between mean full-glacial ELA and present summit altitude) (Fig. 6). Nevertheless, in the tropics, where seasonality is minimal, a close relationship exists between the snowline and the mean annual level of freezing isotherm (e.g., Hostetler and Clark, 2000). The mean July freezing isotherm lies close to the snowline, and its altitude (4715 m) above Hilo during 1965–1974 has been used as a proxy for modern snowline altitude (Porter, 1979b). On the assumption that the ELA would lie between the +0.5 and −0.5 °C isotherms (4715 ± 95 m), snowline depression at the LGM would amount to ca. 930 ± 100 m.

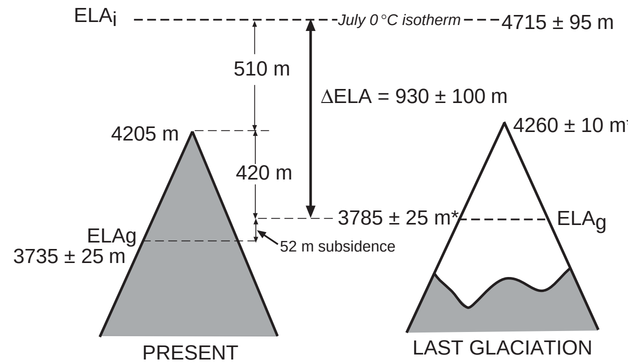

Fig. 6. Diagram showing correction of summit altitude (*) and ELA (*) of Mauna Kea during the LGM based on 60 m of subsequent isostatic subsidence of the volcano (see text). ELA_i = interglacial ELA, ELA_g = glacial ELA, and ΔELA = ELA depression. The average level of the July freezing isotherm is based on rawinsonde measurements at Hilo taken from 1965–1974. Estimated ΔELA at the LGM is 930 ± 100 m, assuming that the ELA_i lies between isotherms of −0.5 and 0.5 °C. All altitude values have been rounded to nearest 5 m.

1.1.6. LGM climate at Mauna Kea

Hostetler and Clark (2000) applied a mass-balance model to evaluate the LGM temperature and precipitation required to generate an ice cap on Mauna Kea. Although the climate changes that control snowline depression are not well constrained, their model results suggest that an equilibrium ice cap could have existed during the last glaciation if precipitation was ca. 63% greater and temperature ca. 3.5 °C lower than now. However, a further cooling of 1 °C would have resulted in a glacier if precipitation was a third less than at present. They suggested that a modest increase in atmospheric lapse rate at the LGM could be related to a change in the mean altitude of the tropical inversion. This might result in colder and somewhat drier conditions at mid-altitude levels (1500–2000 m), and increased precipitation above 2000 m in the zone of glaciation.

Additional factors may play a role in increasing glacial-age precipitation on the Mauna Kea ice cap. For example, a shift to a dominant cool-season weather pattern, marked by a subtropical ridge positioned south of the islands, would allow more fronts and storms to reach the high volcanoes (Schroeder, 1993). Enhanced high-altitude precipitation might result when storms disrupt the trade-wind temperature inversion, allowing clouds to develop on high peaks like Mauna Kea that normally are cloud-free. The probability of a large cyclonic storm reaching the island of Hawaii in any given year is now about 35–40% (Fig. 7) (Lockwood et al., 1990). However, if enhanced cyclogenesis in the eastern Pacific spawned more tropical storms during glacial times, the number reaching the island would rise, thereby leading to greater snowfall on glacierized summits.

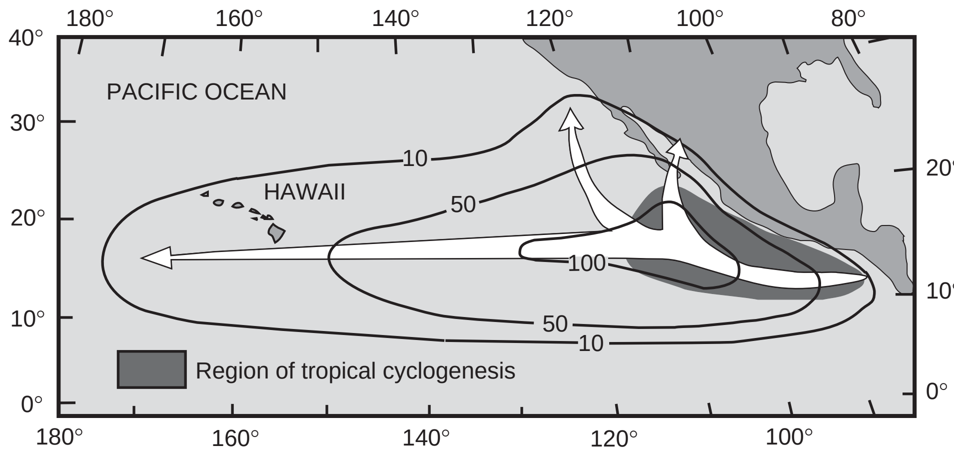

Fig. 7. Average storm tracks for 1965–1985 (white arrows) and isolines of storm probability (% probability of at least 1 storm/5° square in any year) in the eastern tropical Pacific (after Lockwood et al., 1990). A westward shift of the isolines, reflecting an increase in cyclogenesis, would raise the frequency of tropical storms reaching the southernmost Hawaiian Islands.

1.2. Mauna Loa

Like Mauna Kea, Mauna Loa (4169 m) is too low to intersect the modern snowline. However, if the LGM snowline on the two mountains was similar during the LGM, the summit of Mauna Loa likely stood hundreds of meters above the snowline. The crest and adjacent slopes of the volcano are mantled with Holocene basaltic lavas that mask any sedimentary or morphologic evidence of Pleistocene glaciation; the walls of the summit caldera expose only Holocene lavas. Assuming (1) that the ELA on Mauna Loa lay close to that of the Makanaka ice cap on Mauna Kea (ca. 3785 m), and (2) that upward growth of the volcano approximately balanced isostatic subsidence, then the top of the mountain would have stood some 385 m above the LGM snowline. The lowered snowline would have generated an ice cap that may have been similar in size to that on Mauna Kea.

2. The island of Maui: Haleakala volcano

Haleakala (20°42.6′ N Lat., 156°15′ E Lat.; 3055 m) is the higher of the two volcanoes of Maui, but it is too low to intersect the present snowline. The origin of its summit morphology has engendered considerable speculation. A large crater-like depression at the summit, measuring ca. 5×10 km, has variously been attributed to rifting, renting, collapse, and stream erosion (Dutton, 1884; Dana, 1890; Cross, 1915; Daly, 1933; Stearns, 1942). Young lava flows and pyroclastic cones that cover the floor of the depression obscure possible evidence of its origin. Two major valleys, Keanae to the north and Kipahulu to the south, have channeled the largest young flows far downslope.

A decade ago, Moore et al. (1993) suggested that the summit morphology of Haleakala might have been shaped by glacier ice. We based this conclusion on several lines of evidence: (1) DEM images of the volcano point to apparent overfit of the canyons that drain the summit depression, canyons much larger that those that drain comparable areas on other Hawaiian volcanoes; (2) drowned reefs and terraces offshore around Haleakala point to submergence of the volcano by up to 2 km since about 750,000 years ago when its summit likely was as high or higher than that of glaciated Mauna Kea; and (3) diamictons on the lower slopes of the volcano contain clasts with features that are similar to those of presumed subglacially erupted lavas on Mauna Kea.

2.1. Submarine shelf and island subsidence

A submarine shelf that surrounds the islands of Lanai, Kahoolawe, Maui, and Molokai represents portions of their volcanoes that have been submerged due to isostatic subsidence since the end of their respective shield-building stages (Moore, 1987). Beyond the shelf break, which marks the seaward limit of the latest subaerial shield-building lavas that reached the ocean, water depth increases dramatically seaward toward the Hawaiian Deep (ca. 4000–5000 m depth). The edge of the submerged shelf (the H terrace of Moore, 1987) rings the N, E, and S margins of the submarine foundation of Maui (Fig. 8). The terrace has been traced more than 150 km from north of the island (at 400 m depth) to south of Maui and adjacent Kahoolawe (at ca. 2000 m depth) (Moore, 1987, Fig. 2.8; Moore et al., 1990, Fig. 2). The shelf break is presumed to be contemporaneous around the island and to have formed at sea level; therefore, its variable depth

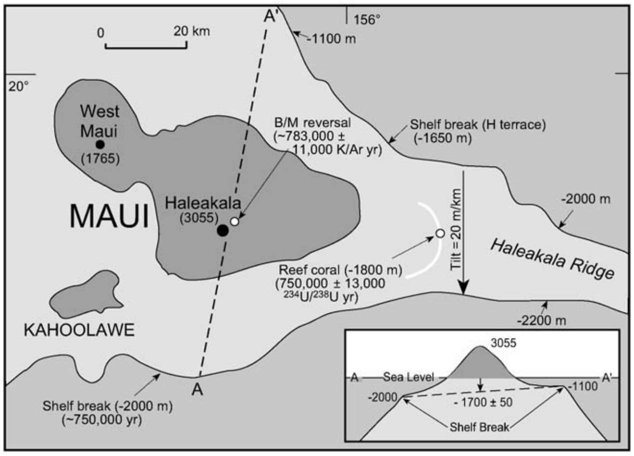

Fig. 8. Map of Maui and vicinity showing location and depth of the shelf break (H terrace, which is tilted 20 m/km southward at Haleakala Ridge. Projected depth of the break under Haleakala is 1700 $\pm$ 50 m (inset, section A–A′). Lavas in the wall of the summit depression record the Brunhes/Matutama reversal (Baksi et al., 1992). A U-series date for a tilted coral reef on Haleakala Ridge at a depth of 1800 m is 750,000 $\pm$ 13,000 years (Moore et al., 1990).

must represent differential subsidence and southward tilting of the island since the shield-building stage ended ca. 750,000 $\pm$ 40,000 years ago (Moore et al., 1990). This submergence and tilting is attributed to crustal loading due to the rapid growth of Maui and the much larger island of Hawaii to the southeast.

A submerged coral reef crosses Haleakala Ridge, the eastward extension of Haleakala's east rift zone (Fig. 8) (Moore et al., 1990). It has a $^{234}U/^{238}U$ age of 750,000 $\pm$ 13,000 years, consistent with the estimated age of the end of shield construction and of the H terrace. Like the H terrace, the reef has a southward tilt, estimated by Moore et al. (1990) to be 20 m/km.

A point below the summit that intersects a line connecting the shelf break north and south of Haleakala lies 1700 $\pm$ 50 m below sea level (Fig. 8, inset). Thus, toward the end of the shield stage the summit of the volcano would have reached an altitude of at least 4700 m. This is ca. 440 m higher than the altitude of Mauna Kea during the Makanaka glaciation, ca. 915 m above the Makanaka ELA, and ca. 100 m higher than the altitude of Mauna Kea during the Pohakuloa Glaciation. This summit altitude can be considered a minimum value if the subsidence rate beneath the crest of Haleakala was greater than along its submerged margins.

2.2. *Geomorphic evidence of glaciation*

Although Haleakala is inferred to have stood higher than the LGM snowline of Mauna Kea, unambiguous morphologic evidence of glaciation has not been found. Neither moraines nor striated surfaces have been reported, for if they exist, they must lie beneath the lavas that fill the summit depression and the major valleys leading from it. The strongest morphologic argument for glaciation is the shape and size of these valleys.

Keanae Valley and Kaupo Valley head in the summit region and are bounded by steep slopes. Macdonald (1978) estimated that they were as deep as 2100 m prior to eruption of the basaltic lavas of the Hana Formation that mantle their floors. A line of cinder and spatter cones related to these flows crosses the summit depression and now forms the divide between the two valleys. Although these giant canyons are generally thought to have been carved by headward-eroding streams (Stearns, 1942; Stearns and Macdonald, 1942), they are anomalously large for their drainage area, head at a high altitude where annual precipitation is now less than 100 cm, and contrast markedly with much smaller adjacent valleys.

An alternative hypothesis for the summit morphology is that the crest region of Haleakala and the headward sectors of major valleys draining it have been shaped by glacier ice (Moore et al., 1993). If a caldera existed at the summit of the volcano toward the close of the shield-building stage, as is the case at Mauna Loa, it would have been above the snowline altitude of the Mauna Kea glaciations and been an obvious site for ice accumulation. A glacier that formed in such a summit depression may locally have overtopped the caldera rim and excavated the broad upper reaches of the two major valleys that indent the north and south flanks of the mountain. In addition to glacier erosion, meltwater-enhanced streamflow would have increased rates of stream erosion in these valleys. Furthermore, the summit region may have been subjected to periodic eruptions during the waning stage of volcano growth, as on Mauna Loa and Mauna Kea. In this event, there may be distinctive lithologic evidence of subglacial volcanism (see below) that would support this hypothesis.

2.3. *Lithologic evidence of glaciation*

A gravelly diamicton is exposed at the land surface and in seacliffs more than 100 m high at the lower end of Kaupo Valley on the southern slope of Haleakala (Fig. 9). Clasts of a similar diamicton have been found in alluvium at the mouth of Keanae Valley to the north (Stearns, 1942; Stearns and Macdonald, 1942). Stearns and Macdonald inferred these to be mudflow deposits that resulted from erosion and mobilization of taluses along the base of steep valley walls during major rainstorms. However, the Kaupo deposit is unique in terms of its thickness (> 100 m) and exposed breadth (at least 4 km). Somewhat similar diamictons exposed on the glaciated upper slopes of Mauna Kea have been identified and mapped as till (Porter, 1979a, b; Wolfe et al., 1997). However, the deposits on Haleakala lie far

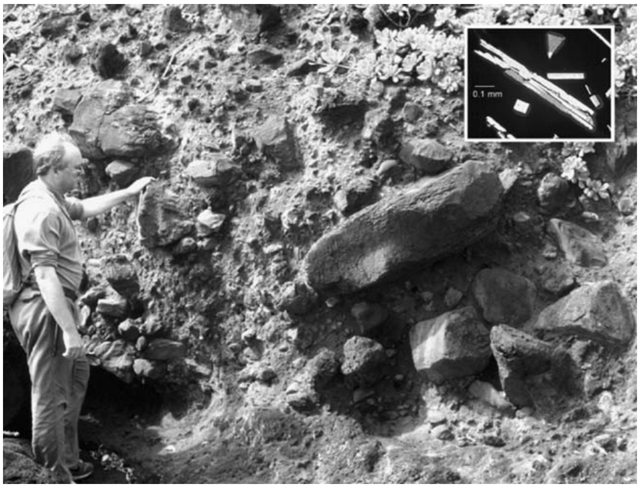

Fig. 9. Seacliff exposure on south coast of Haleakala showing nonsorted mudflow deposit that contains clasts with glassy margins and with ‘swallowtail’’ morphology of plagioclase crystals (inset) typical of rapidly quenched lavas.

too low on the volcano’s slopes (to and below sea level) to be interpreted reasonably as till.

The coarse non-sorted Kaupo deposit contains clasts with glassy, chilled margins like those seen along the margins of subglacially erupted lava flows on Mauna Kea (Porter, 1987). Furthermore, thin sections of the glassy rims of the lava clasts display plagioclase crystals with distinctive ‘‘swallowtail’’ structure, a feature identified in lavas that have been rapidly quenched during subaqueous eruptions (Fig. 9) (Stewart et al., 1973; Winter, 2001).

It is difficult to conceive of rain-generated mudflows occurring on a scale required to produce such a voluminous and extensive mudflow deposit in lower Kaupo Valley that spread beyond the present shoreline. Such deposits are not found in other major valleys on other high Hawaiian volcanoes. A more likely source of the large volumes of water to mobilize such mudflows, especially on the arid southern flank of Haleakala, is subglacial eruptions during times when a glacier occupied the summit depression and adjacent valley heads. Glacier outburst floods generated by such eruptions would likely incorporate pieces of the quenched margins of newly erupted and highly fragmented lavas, together with glacial and stream deposits, and transport the debris down canyons as mudflows. The known deposits are associated only with the two great valleys that originate in the summit depression of Haleakala, which lends credence to this hypothesis.

3. Inferred history of glaciation

Whereas a morphologic and sedimentary record of glaciation exists on Mauna Kea, the case for glaciation on Mauna Loa and Haleakala is circumstantial and based largely on limited morphologic and sedimentary evidence, as well as on the relationship between former snowlines and summit altitudes. Carrying the analysis a step further, volcano growth and subsidence data can be used to place limits on likely times of past glaciation and to derive a tentative comprehensive history of Pleistocene glaciation in Hawaii.

3.1. Volcano growth and subsidence

It is generally inferred that the typical eruptive life of a Hawaiian volcano encompasses as much as 600,000 (Moore and Clague, 1992) to a million years (Wolfe et al., 1997). After an initial preshield phase of relatively slow growth, the volcano enters a long shield phase of rapid growth that is followed by a postshield stage of reduced growth. During the late shield and early postshield stages the volcano likely reaches its maximum altitude.

Most exposed lavas of Mauna Loa are Holocene in age. Mauna Kea’s dated lavas are postshield lithologies that represent the waning phases of volcano growth. On Haleakala, exposed rocks also primarily represent the final interval of volcano construction and few have been dated. A longer record from which to infer the late-eruptive history exists for Kohala volcano, the oldest rocks of which are exposed in northeast-trending valleys that head deep within the core of the volcano. Like Mauna Kea, Kohala’s shield stage was followed by a basaltic substage and a subsequent hawaiitic substage, represented by the Pololu and Hawi volcanics, respectively (Wolfe and Morris, 1996). The oldest exposed lavas (Pololu), 720 m below the constructional surface of the volcano and far up Waipio valley, were erupted within the Brunhes Chron (<780,000 yr ago). The Pololu–Hawi transition, near the end of the shield stage, dates to ca. 245,000 years ago (Moore and Clague, 1992).

The rapid upward growth of a Hawaiian shield is accompanied by isostatic subsidence, a result of the great load imposed on the underlying ocean crust. Subsidence slows the increase in summit altitude due to eruptive activity. As extrusion rates decline at the end of the shield stage, subsidence overtakes upward growth and dominates subsequent volcano history.

3.2. Glacial and snowline history of Mauna Kea and Mauna Loa

The dated stratigraphic record of Mauna Kea shows us that the snowline lay below the summit at least three times during the last ca. 150,000 years, but provides no information about possible earlier glaciations. To assess the possibility of a longer glacial history, data on volcano growth and isostatic subsidence can be used to estimate the change in summit altitude during earlier times.

A shelf break at 375 ± 15 m depth off the northeastern coast of Mauna Kea marks the end of shield construction ($133{,}000 \pm 10{,}000$ K/Ar yr ago) (Ludwig et al., 1991; Moore and Clague, 1992). Using Ludwig et al.'s average subsidence rate of 2.6 mm/yr, the summit altitude at the end of the shield stage would have been ca. 4550 m $[4206 + (0.0026 \times 133{,}000)]$ (Fig. 10). Dashed lines in Fig. 10 placed at altitudes of 3785 and 4715 m represent the full-glacial (MIS 2) and estimated interglacial (present: MIS 1) snowlines. The standard marine isotope curve for the past 300,000 years is placed between these two lines, with MIS 2 and MIS 1 values adjusted to it. This curve is a proxy for global ice volume and is used here as a proxy for snowline variations under the assumption that the timing and relative amplitude of the isotope peaks approximate the gross fluctuations in ELA at Hawaii.

The rising curve of summit altitude first intersects the curves of ELA variations during MIS 6. It crests ca. 130,000 yr ago, where it lies well above the depressed snowline of MIS 6. It also lies above the snowline curve during MIS 4 and 2, each of which is represented by exposed glacial deposits on Mauna Kea. It appears likely that a small ice cap also existed, at least intermittently, during MIS 3. Supporting this inference are K/Ar dates of ice-contact lava flows near the summit that range in age from $33{,}000 \pm 12{,}000$ to $41{,}000 \pm 8000$ years (Wolfe et al., 1997, Pl. 3). A change in glacial and interglacial ELA values and/or summit altitude values by as much as $\pm 20\%$ does not alter the basic conclusion that an ice cap would developed repeatedly on Mauna Kea once the summit rose above the full-glacial snowline. The stratigraphic evidence of glaciation is proof that it did.

Mauna Loa is in its shield-building stage, and its altitude probably is increasing despite ongoing subsidence. It may have been high enough to intersect the Makanaka-age snowline, but was too low to rise above earlier full-glacial snowlines (Fig. 11). If an ice cap formed during the last glaciation, its deposits are now obscured by overlying postglacial lavas.

3.3. Glacial and snowline history of Haleakala

A provisional glacial history of Haleakala can be determined using the same ELA proxy curve (extended

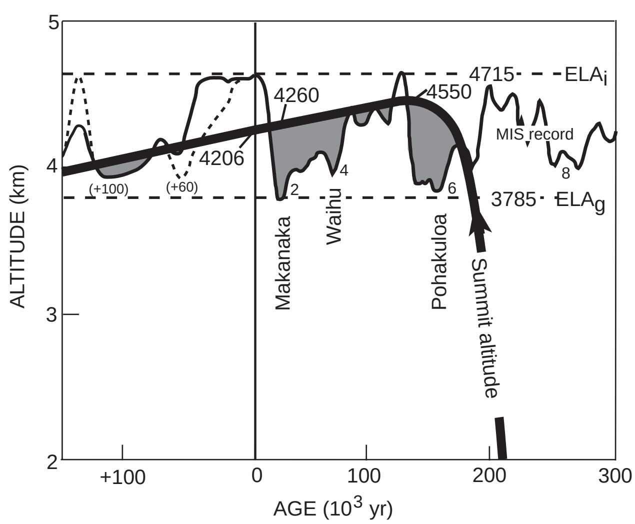

Fig. 10. Summit-altitude curve for Mauna Kea showing time of maximum altitude and subsequent decline as a result of waning volcanic activity and ongoing subsidence. Standard marine oxygen-isotope (MIS) curve (Williams et al., 1988) is used as a proxy for snowline variations. This curve is adjusted so that snowline altitude of the present interglaciation (ELA$_i$) equals ca. 4715 m and snowline for the last glacial maximum (ELA$_g$) equals 3785 m (see text). The summit first intersected the snowline curve near the beginning of MIS 6. Shaded areas represent times when the summit rose above the snowline and a glacier formed. The three mapped glacial drifts coincide with intervals of lowered snowline during MIS 6, 4, and 2. Projection of the ice-volume curve forward in time [from modeling studies of Loutre and Berger (2000) and Berger and Loutre (2002); solid line based on last-interglaciation values, dotted line on constant CO_2 value of 220 ppmv] suggests that small ice caps may form on Mauna Kea ca. 60,000 and 100,000 yr in the future, after which continuing subsidence will bring the summit below the glacial-age snowline.

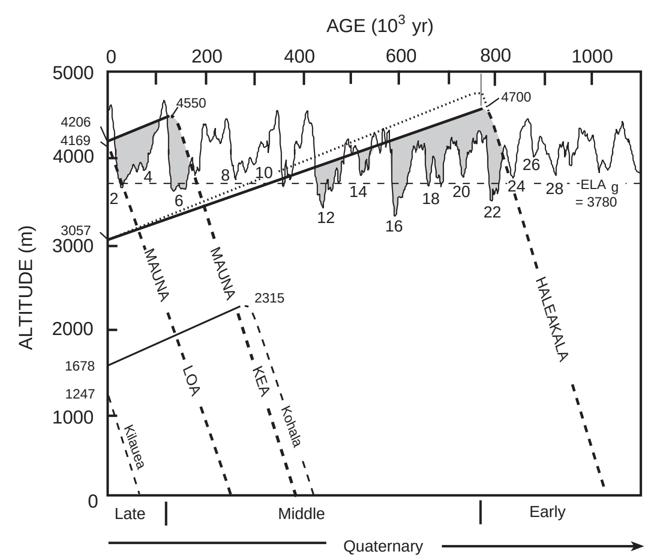

Fig. 11. Composite history of glaciation in the southeastern Hawaiian Islands illustrated by summit-altitude curves and their intersection with the inferred curve of snowline variations for the past million years. ELA$_g$ is the inferred full-glacial ELA based on the value for the Makanaka glaciation on Mauna Kea. The summit of Haleakala first intersected the snowline about 800,000 years ago during MIS 22 and remained above the snowline through MIS 16. The last glaciation of Haleakala may have coincided with MIS 12 (or possibly MIS 10). Dotted line is alternative summit-altitude curve based on an assumption of greater subsidence under the summit than on seaward margins, and using an average subsidence rate of 2.6 mm/yr since the end of the shield stage. Kohala volcano never reached the altitude of the snowline and therefore lacks evidence of glaciation. Mauna Loa reached snowline altitude only during the last glaciation, and Kilauea is likely to remain below the snowline for another 150,000–200,000 years. If a Hawaiian volcano grows to an altitude of at least 4500 m, it may experience a succession of ice-cap glaciations for nearly half a million years.

to 1,100,000 years in Fig. 11). The summit altitude curve is drawn with the maximum altitude (4700 m) at ca. 750,000 years ago, the approximate age of the offshore shelf break that marks the end of the shield stage (Moore and Clague, 1992). Rapid growth of the volcano brought the summit above the snowline near the end of the shield stage, close to MIS 22. Based on the reconstructed summit-snowline relationship, the summit may have stood continuously above the snowline until the end of MIS 16, about 600,000 years ago (i.e., for about 200,000 years). The final glaciation on Haleakala would have occurred during MIS 12 (or possibly MIS 10), following which the summit subsided below the snowline (about 400,000 years ago). The assumption is made that the full-glacial snowline during this long interval was equivalent to the MIS 2 snowline (3785 m). However, even if snowline depression was only 75% of that amount (i.e., to ca. 4020 m), a succession of ice caps would have formed at the summit between 800,000 and 600,000 years ago.

The average calculated rate at which summit altitude decreased over the last 750,000 years is 2.0 mm/yr. However, if the rate of subsidence was greatest at the thickest part of the lava pile (i.e., if the projected shelf break shown in Fig. 8 (inset) was deeper than 1700 m), then the summit may have reached a higher altitude (dotted line in Fig. 11). In this case, the interval of glaciation would have been extended and likely would have included a glaciation during MIS 10. Quite likely the subsidence rate initially was greater (e.g. $\geqslant$2.6 mm/yr), but declined steadily after the shield stage ended and the locus of major eruptive activity shifted southeastward to the island of Hawaii. The southward tilt of the shelf break around Haleakala (Fig. 8) points to continuing differential subsidence in postshield time as Hawaii continued to increase in volume.

3.4. Past and future glaciation of the Hawaiian Islands

Kohala volcano also reached its maximum altitude within the last half million years, but its summit was too low to intersect the glacial-age snowline (Fig. 11). Kilauea, now in its shield-building stage of rapid growth, will not reach an altitude high enough to support a glacier for another ca. 150,000 years. By that time, Mauna Kea likely will have had several additional ice-cap glaciations, at ca. 60,000 and 100,000 years in the future, as can be inferred from modeling studies of future glaciation by Loutre and Berger (2000) and Berger and Loutre (2002) (Fig. 10).

The dated glacial deposits of Mauna Kea, the inferred history of glaciation on Haleakala, and the apparent absence of glacial deposits on lower volcanoes of Hawaii and Maui imply that glaciation is restricted to mid-ocean Hawaiian volcanoes that reach altitudes of at least 3800 m, the level of the glacial-age snowline. If a Hawaiian volcano rises as high or higher than ca. 4600 m, it may experience a succession of ice-cap glaciations lasting as much as half a million years. The high volcanoes of the southeastern Hawaiian Islands may collectively have experienced ice-cap glaciations during most of the Middle and Late Pleistocene (Fig. 11).

The glaciated Hawaiian summits are unique in the vast expanse of the Pacific Ocean basin. Although glaciated mountains rim the basin on all sides, the high Pleistocene volcanoes of Hawaii apparently are the only ones on the Pacific Plate to have grown high enough to become glaciated during the Quaternary Period. Accordingly, they gain unusual importance for assessing past terrestrial climates and environments of a region that encompasses a substantial part of the planet. That the record of these changes may extend back through nearly half the Pleistocene Epoch makes these islands of even greater significance.

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