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Unspiked K-Ar dating of young volcanic rocks from Loihi and Pitcairn hot spot seamounts

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Abstract

A geochronological study of volcanically active seamounts from the Pitcairn and Hawaii hot spots successfully utilized the unspiked K-Ar technique to obtain ages for young volcanic rocks. These ages were used to calculate lava production rates for the youthful stage for the Hawaiian volcanic chain. The lavas from Pitcairn seamounts have ages ranging from 3 ± 1 to 344 ± 32 ka. The reliably dated lavas from Loihi are from a 500 m thick section on the east flank of the volcano and they range in age from 5 ± 4 to 102 ± 13 ka. These lavas are all alkalic and are from the preshield stage of volcanism. These ages yield lava accumulation rates that increase from 3.5 mm/a for the dominantly alkalic lower section to 7.8 mm/a for the predominantly tholeiitic upper part of the volcano. These values are consistent with those obtained for other Hawaiian volcanoes. The duration of the transition from tholeiitic to alkalic lava during the preshield stage is about 17 to 40 ka, which is similar to the tholeiitc to alkalic transition during the postshield stage of Hawaiian volcanism. The total duration of the preshield stage of Hawaiian volcanism is estimated to be at least 250,000 years, which is more than twice previous estimates. The overall length of magmatic activity, for a typical Hawaiian volcano is now estimated to be ~ 1.4 Ma. © 1997 Elsevier Science B.V.

Keywords: hot spot; lava accumulation rates; K-Ar geochronology; Loihi; Hawaii

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1. Introduction

Accurate and precise age determinations for young (< 500 ka) volcanic rocks are essential for detailed magmatic and neotectonic interpretations. K-Ar dating has proved very successful for determining crystallization ages for some volcanic rocks (e.g., Dalrymple and Lanphere, 1969) but it has been a challenge to adapt this method for dating young basaltic rocks, especially submarine basalts. To date such rocks, one must avoid the effects of mantle-derived ^{40}Ar and be able to detect and analyze very small amounts (< 1%) of radiogenic ^{40}Argon (^{40}Ar*). Duncan and Hogan (1994) found that mantle-derived ^{40}Ar could be avoided if the holocrystalline part of the basalt was used (at least 4–5 cm from the glassy margin).

The incremental heating method takes advantage of the observation by Hall and York (1978) that the amount of ^{40}Ar* in a sample could be enhanced by heating it at temperatures as high as 400°C. This step

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removes some of the surficial atmospheric Ar, which resulted in greater relative amounts of $^{40}Ar^*$ at higher temperature intervals. This method has been successfully combined with improvements in the sensitivity and resolution of mass spectrometers and in ultra-clean extraction systems to analyze low K content, young basalts (300–850 ka) from the East Pacific Rise in small quantities (Duncan and Hogan, 1994). It has not yet been used to date very young submarine basalts (< 200 ka).

An alternative method of conventional K-Ar dating has proven successful with dating very young subaerial, K-rich volcanic rocks, some as young as > 10 ka (Gillot and Cornette, 1986). This unspiked

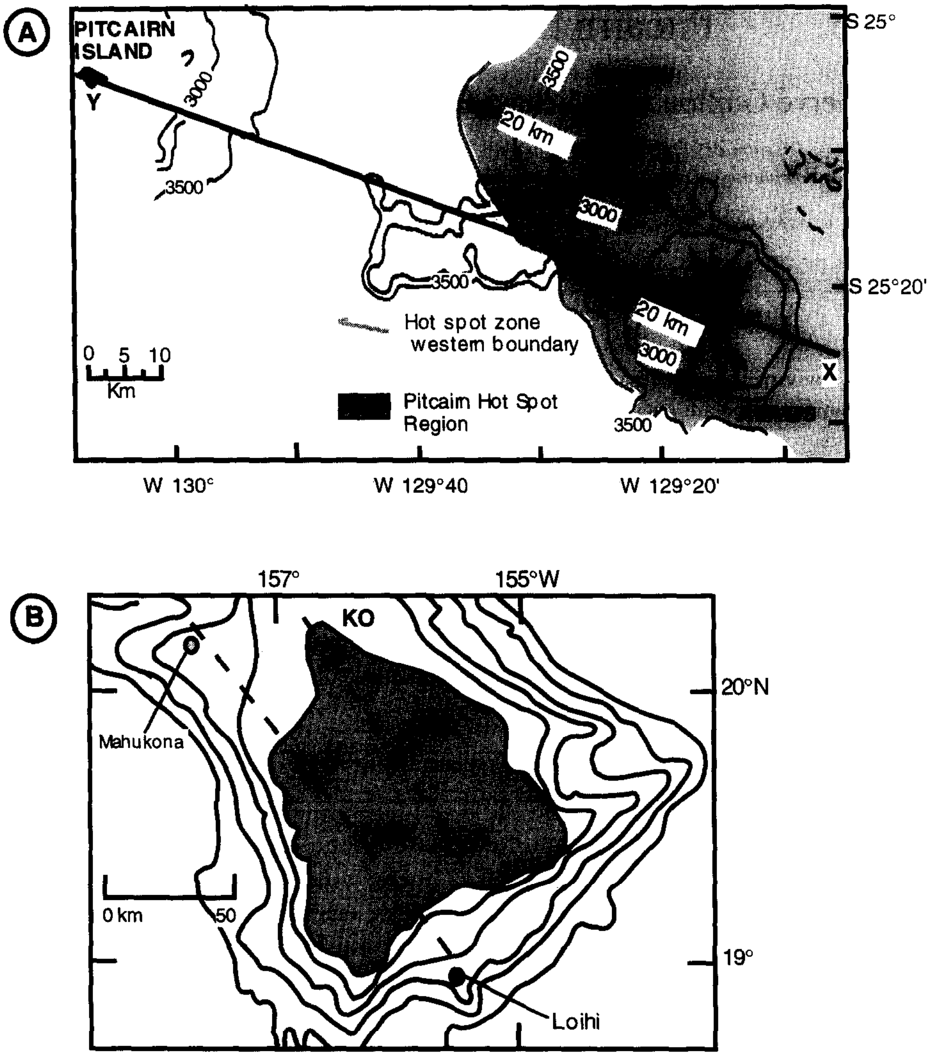

Fig. 1. General bathymetric map of the Pitcairn hot spot region (HSR), (from Binard, 1991) and Loihi (from Garcia et al., 1995a). (A) Pitcairn HSR: The limit of the hot spot magmatic activity is inferred to be shown by the 3500 m contour line. The two arrows indicate the distance between the seamounts and the western boundary of the hot spot zone. The direction of plate motion is shown by the line *X–Y*, which is oriented N110°. The contour interval is 500 m. (B) Loihi Seamount: This map shows the location of this volcano on the south flank of the island of Hawaii and the summit locations of the volcanoes (O) that form the island (*KO* = Kohala; *H* = Hualalai; *MK* = Mauna Kea; *ML* = Mauna Loa; *K* = Kilauea and Mahukona Seamount). The two dashed lines show the orientation of the two parallel chains of volcanoes that form the southern end of the Hawaiian islands. Contour interval is 1000 m.

technique (also known as the Cassignol method) dynamically compares the isotopic composition of an aliquot of pure atmospheric Ar with the sample Ar composition to accurately determine minor variations of the $^{40}Ar/^{36}Ar$ isotopic ratio between the standard and the unknown. This method provides a precise correction for atmospheric argon contamination and it avoids any discrimination effects of the mass spectrometric measurements (Cassignol and Gillot, 1982).

We have utilized the unspiked technique to date some young to very young submarine basalts from the Pitcairn and Hawaiian hot spots. The Loihi samples have been collected from a ~ 500 m thick section on the deeply dissected east flank of the volcano. This well established stratigraphic section allows us to check the geological significance of the results and to estimate lava accumulation rates. These results are the first 'data-constrained' estimates of the duration and growth rate for a Hawaiian volcano during its early preshield stage of development. The duration of Loihi's preshield stage is probably at least 250 ka and the calculated lava accumulation rates near the volcano's summit increase from 3.5 mm/a during the upper part of the alkalic substage to 7.8 mm/a for the overlying predominantly tholeiitic volcanism, which probably marks the end of the preshield stage (Garcia et al., 1995a).

2. Geological background

The active region of the Pitcairn hot spot (Fig. 1) is located ~ 100 km southeast of Pitcairn Island (near 129°30′W, 25°20′S). It is built on 30 Ma Pacific ocean crust and consists of twenty volcanic hills (~ 500 m high) and two major edifices, the Adams and Bounty seamounts, which rise ~ 3500 m in height to within 60 and 450 m of sea level (Stoffers et al., 1990; Binard, 1991). The Pitcairn samples for this study were dredged in 1989 during the 65th cruise of the German research vessel Sonne. The recent volcanic rocks from this region have compositions ranging from basanites and alkali basalts to trachytes. No tholeiitic rocks have been described from this area.

Loihi seamount is located 35 km south of the island of Hawaii on the flank of the two other active Hawaiian volcanoes, Mauna Loa and Kilauea (Fig. 1A). It is underlain by 100–105 Ma Pacific ocean crust (Epp, 1984; Waggoner, 1993). It rises ~ 3.5 km from its basement to within ~ 960 m of sea level and is the youngest shield volcano in the Hawaiian chain (Fornari et al., 1988). A large landslide has exposed the deep interior of the east flank of Loihi. Lavas were collected from the landslide scarp during three dives of the Pisces V submersible (Garcia et al., 1995a). The slope of this area varies from the regional slope for areas that have not experienced mass wastage (~ 14°, which is the regional dip of the upper flank lavas) to essentially vertical. The average slope for the composite section (between 1130 and 1960 m) is ~ 30°. Thus, the overall stratigraphic thickness for the samples in this study is ~ 505 m. This section is thicker than the thickest subaerial exposures for the active volcanoes on the island of Hawaii [e.g., Kilauea's Hilina Pali—300 m thick (Easton, 1987); Mauna Loa's Kahuku Pali–150 m (Garcia et al., 1995b)]. Loihi's east flank section yielded a wide range of rock types including tholeiites, alkali basalts and basanitoids (Garcia et al., 1995a). There is a systematic change in the proportion of alkalic to tholeiitic lavas with depth, from predominantly alkalic at the base of the section to almost entirely tholeiitic at the top (Garcia et al., 1993, 1995a).

3. Analytical procedure

Eight unaltered lavas (three from the Pitcairn HSR, five from Loihi) were selected for geochronological study based on their high K content and relative stratigraphical position. Glassy samples were avoided because of possible excess argon caused by rapid quenching which impedes loss of the initial non-atmospheric argon (Dalrymple and Moore, 1968). Major element analyses were made to estimate the degree of alteration, the amounts of potassium and to establish the rock types for these lavas (Table 1). The loss-on-ignition values (L.O.I.) for these samples range from − 0.43 (from oxidation of iron) to 0.35%, except for sample 65 61DS with a loss of 0.76%. Thus, all the samples (except 65 61DS) are essentially unaltered based on the L.O.I. analyses. The lavas from the Pitcairn seamounts

Table 1
Whole-rock analyses of lavas from Pitcairn (Bounty and Adams) and Hawaii howspots (Loihi)

Sample: Rock type:	65 67DS Trachyte	65 61DS Trachyte	65 51DS Tholeiitic basalt	186-14 Hawaiite	187-6 Alkalic basalt	158-6 Alkalic basalt	158-5 Basanitoïd	186-2 Alkalic basalt
SiO_2	59.75	59.70	49.25	46.05	45.76	45.48	42.62	46.86
TiO_2	0.62	0.60	2.75	4.17	2.99	3.28	3.67	3.23
Al_2O_3	17.22	17.27	14.00	14.17	14.95	14.54	15.10	13.53
Fe_2O_3	7.82	7.61	11.92	14.36	13.40	14.67	16.99	12.46
MnO	0.18	0.18	0.14	0.19	0.18	0.19	0.19	0.17
MgO	0.66	0.78	9.43	5.33	6.18	5.44	6.31	7.05
CaO	2.50	2.47	8.04	10.30	12.62	11.87	11.27	12.93
Na_2O	6.35	6.13	3.14	3.66	2.82	3.13	2.71	2.76
K_2O	4.09	4.10	0.77	1.11	0.89	1.11	0.77	0.77
P_2O_5	0.31	0.30	0.40	0.51	0.3	0.38	0.28	0.35
Sum	99.50	99.14	99.84	99.85	100.12	100.09	99.91	100.11
L.O.I	0.35	0.76	−0.05	0.10	−0.43	−0.02	0.27	0.11

The Pitcairn HSR samples were analysed by Induced Coupled Plasma at the University of Brest. Analyst: J. Cotten. Data are from Binard (1991). DS = dredge sample. Loihi samples were analysed by XRF at the University of Hawaii. Analysts: M. Garcia and T. Hulsebosch.

include an alkali basalt (65 51DS) and two trachytes (65 67DS and 65 61DS). The Loihi rocks include a basanitoid (sample 158-5), alkali basalts (187-6, 158-6 and 186-2) and a hawaiite (186-14).

All the samples were crushed and sieved to 0.250–0.125 mm size fractions and ultrasonically washed in HNO3 to remove any secondary mineral phase that might be present in minute amounts. Phenocrysts and xenocrysts (which may carry excess argon) were eliminated using both heavy liquids and magnetic separations to obtain pure groundmass aliquots. The groundmass is assumed to have formed shortly after eruption and should not contain any excess argon.

K and Ar measurements were performed on the groundmass fractions of each sample. K was analysed by atomic and flame emission spectrophotometry with a relative precision of 1%. A description of the unspiked technique and the instrument used for Ar measurements have been presented elsewhere (Cassignol et al., 1978; Cassignol and Gillot, 1982; Gillot and Cornette, 1986). Argon was extracted from 1 to 2 g samples by radio frequency heating induction in high vacuum glass line and purified with titanium sponge and SAES Zr-Al getters. Isotopic analyses were made on Ar quantities ranging from 1.10^{-11} to 2.10^{-10} mol, using a 186°, 6 cm radius mass spectrometer operated at an accelerating potential of 620 V. The spectrometer was operated in a semi-static mode in which data were measured on a double faraday collector in sets of 100 using a 1 s integration time. The sensitivity of the mass spectrometer is about 5.1 10^{-15} mol/mV and its background is 4.10^{-13} mol for ^{40}Ar and 3.10^{-14} mol for ^{36}Ar. The procedural blank is 8.10^{-12} mol and has an atmospheric composition.

The stability of the working conditions required to detect minor amounts of ^{40}Ar* (down to 0.1%) are insured by the purity of the gases introduced in the mass spectrometer as well as by the stability of the ion source parameters, which suppress signal drift. The analysed gases must be as pure as possible because active gases present in the mass spectrometer may react with the source filament and, therefore, change the ionization conditions. After melting a sample, the gas is purified with titanium sponge and SAES Zr-Al getters before its introduction into the mass spectrometer. Zr-Al getter pumps are continuously run during analysis to maintain the purity of Ar in the mass spectrometer and to assure a constant low level of the residual active gases in the instrument. Helium, which can penetrate into the line, especially during the heating of the titanium sponge, is eliminated by pumping just before the introduction of the gases into the mass spectrometer. The measurements are, therefore, performed in a semi-static mode. These procedural improvements almost entirely reduce peak drifting. Memory effects from the mass spectrometer itself can also be responsible for peak drift. These effects are reduced by the use of a double collector, which limits the surface area bombarded by the ion beam, and by using a lower accelerating voltage (620 V). The memory effects are also reduced by continuously running the mass spectrometer at an Ar pressure similar to that of the measured samples.

Owing to the very low radiogenic Ar content of the samples in this study, improvements were made in the error analysis and calibration to improve the accuracy and precision of the age determinations. The volumetric calibration of the spike-free introduction line was refined by cross calibration using the standards GL-O, LP-6 (Odin, 1982), Mmhb-1 (Samson and Alexander, 1987), and HD-B1 (Fuhrmann et al., 1987). This calibration (Charbit et al., 1995) allows Ar content to be determined with a precision of 0.2% ($\pm 2\sigma$). The error on the age determination was then determined as follows: in the absence of error correlation between the two variables ($^{40}\mathrm{Ar}^*$ and $^{40}\mathrm{K}$), the variance of the age is approximate by:

$$\sigma_t^2 = \left(\frac{\partial t}{\partial \mathrm{Ar}^*}\right)^2 \times \sigma_{\mathrm{Ar}^*}^2 + \left(\frac{\partial t}{\partial \mathrm{K}}\right)^2 \times \sigma_{\mathrm{K}}^2$$

where $\sigma_{\mathrm{Ar}^*}^2$ and σ_{K}^2 are the variance of the number of $^{40}\mathrm{Ar}^*$ and $^{40}\mathrm{K}$ atoms, respectively, and $(\partial t/\partial \mathrm{Ar}^*)$ [resp. $(\partial t/\partial \mathrm{K})$] are the partial derivative of the age with respect to $^{40}\mathrm{Ar}^*$ (resp. $^{40}\mathrm{K}$) evaluated at its mean. According to the central limit theorem, the error distribution in the age determination should be gaussian. The main source of error in the determination of $^{40}\mathrm{Ar}^*$ is the noise affecting the $^{36}\mathrm{Ar}$ signal. As the error is propagated symmetrically in the equation above, the distribution of error for $^{40}\mathrm{Ar}/^{36}\mathrm{Ar}$ from each individual measurement was checked and the measurement was rejected when its distribution was not gaussian. Two measurements, one of three for samples 158-5 and 158-6, were rejected, following the procedure aforementioned.

4. Results

The measured isotopic ratios and their errors were calculated based on 100 measurements for each Ar isotope (Table 2). Standard error (1σ) for the isotopic ratios of the samples range between 0.009 and 0.045%. $^{40}\mathrm{Ar}^*$ contents vary from 0.023 to 0.835% with the standard error (SE) of 84.7 and 2.65%. The magnitude of the SE is related to both the stability of the working conditions or the percentage of $^{40}\mathrm{Ar}^*$ detected (e.g., low % Ar gave high SE).

Age determinations for the seamount lavas are reported in Table 3. Good analytical reproducibility was observed for all samples including the very young lavas (< 10 ka; samples 65 67DS, 65 61DS and 186-14). The duplicate ages are within the two sigma error. The ages for the three Pitcairn submarine lava flows range from 3 ± 1 ka (sample 65 67DS, mean value) to 344 ± 32 ka (sample 65 51DS, mean value). Both of the trachytes from Adams seamount gave Holocene ages indicating that this seamount is still geologically active. The weakly alkalic basalt from Bounty seamount is clearly much older (344 ± 32 ka). The Pitcairn samples were

Table 2
Raw data from mass spectrometric measurements of samples

Sample	$\mathrm{IR_S}$	SE (%)	$\mathrm{IR_A}$	SE (%)	$^{40}\mathrm{Ar}^*$ %	SE (%)
Pitcairn— Adams						
65 67 DS	280.204	0.036	279.316	0.039	0.317	17.066
65 67 DS	278.458	0.042	277.771	0.037	0.247	22.426
65 67 DS	276.613	0.039	275.703	0.041	0.327	18.530
65 61 DS	279.106	0.035	278.396	0.030	0.255	16.466
65 61 DS	280.842	0.045	278.496	0.052	0.835	7.170
Pitcairn— Bounty						
65 51 DS	280.111	0.011	279.362	0.008	0.267	4.749
65 51 DS	278.800	0.008	278.067	0.008	0.263	4.099
Loihi seamount						
186-14	275.934	0.014	275.911	0.013	0.023	84.687
186-14	273.710	0.019	273.475	0.019	0.086	33.072
187-6	275.172	0.010	273.309	0.016	0.677	2.649
187-6	276.517	0.011	274.620	0.012	0.686	2.268
158-6	277.187	0.014	276.072	0.013	0.402	4.494
158-6	274.385	0.012	273.563	0.012	0.300	5.538
158-5	274.296	0.024	274.064	0.021	0.085	36.954
158-5	274.129	0.014	273.858	0.014	0.099	19.493
186-2	278.150	0.011	277.592	0.007	0.201	6.157
186-2	278.291	0.009	277.729	0.008	0.202	5.984

IR = $^{40}\mathrm{Ar}/^{36}\mathrm{Ar}$ isotopic ratio. Subscript S refers to sample, A to atmospheric; SE is standard error.

Table 3
K-Ar ages of samples from the Pitcarin hot spot zone and Loihi. Age calculations are based on the decay and abundance constants from Steiger and Jäger (1977)

Sample	Depth intervals (m)	K (wt.%)	Weight molten (g)	^{40}Ar (%)	^{40}Ar* (10^{-13} mol/g)	Age $\pm 2\sigma$ (ka)
Pitcairn—Adams						
65 67 DS	291-366	3.497 ± 0.035	2.01013	0.317	0.183	3 ± 1
65 67 DS	291-366	3.497 ± 0.035	2.07500	0.247	0.151	2 ± 1
65 67 DS	291-366	3.497 ± 0.035	2.04278	0.327	0.193	3 ± 1
65 61 DS	410-516	3.486 ± 0.035	1.01187	0.255	0.350	6 ± 2
65 61 DS	410-516	3.486 ± 0.035	2.00056	0.835	0.450	7 ± 1
Pitcairn—Bounty						
65 51 DS	1859-1609	0.770 ± 0.008	1.00133	0.267	4.673	350 ± 34
65 51 DS	1859-1609	0.770 ± 0.008	1.02503	0.263	4.514	338 ± 29

Sample	Depth (m b.s.l)	Stratigraphic height (m)	K (wt.%)	Weight molten (g)	^{40}Ar (%)	^{40}Ar* (10^{-13} mol/g)	Age $\pm 2\sigma$ (ka)
Loihi							
186-14	1130	0	1.028 ± 0.010	2.00481	0.023	0.036	2 ± 3
186-14	1130	0	1.028 ± 0.010	1.50268	0.086	0.131	7 ± 5
187-6	1475	245	0.889 ± 0.009	1.05773	0.677	3.0243	196 ± 11
187-6	1475	245	0.889 ± 0.009	1.00889	0.686	3.157	205 ± 10
158-6	1650	310	0.796 ± 0.008	2.00088	0.402	0.638	46 ± 4
158-6	1650	310	0.796 ± 0.008	2.00079	0.300	0.583	42 ± 5
158-5	1830	415	0.962 ± 0.010	1.50028	0.085	0.152	9 ± 7
158-5	1830	415	0.962 ± 0.010	1.58964	0.099	0.175	11 ± 5
186-2	1960	505	0.619 ± 0.006	1.51326	0.201	1.055	98 ± 12
186-2	1960	505	0.619 ± 0.006	1.00630	0.202	1.125	105 ± 13

dredged, so we have no independent test of there relative ages.

The Loihi ages range from 5 ± 4 ka (sample 186-14, mean value) to 201 ± 11 ka (sample 187-6 mean value). We can evaluate whether the Loihi ages are a good indicator of the crystallization age by comparing them to the stratigraphic order in the east flank section. If the sample had any excess argon or was an open system for either gain or loss of K or Ar, the ages would not increase with depth in the section. The five Loihi ages do not increase with depth (Table 3). In the simplest analysis, the ages for samples 187-6 and 158-5 do not fit with their stratigraphical position. The age of sample 187-6 appears to be too old and the age for sample 158-5 appears to be too young. The relative standard error for $^{40}\mathrm{Ar}^*$ is high for sample 158-5 ($\sim 30\%$) and the % $^{40}\mathrm{Ar}^*$ is low but the opposite is true for sample 187-6.

5. Discussion

5.1. *Depth vs. age for Loihi lavas*

The lack of consistency of the Loihi ages with their stratigraphic position could be caused by many factors. One factor we can eliminate is that the samples are from displaced blocks because all of the samples were collected in situ. The relatively old age for sample 187-6 (201 ka) from the middle part of the Loihi east flank section appears to be too old compared to samples dated from above and below it. This age also appears to be too old based on lava accumulation rate estimates for Loihi's sister, Kilauea, which were used to estimate the age of the east flank section at between 100 and 150 ka (Garcia et al., 1995a). If we were to consider that the age of the sample 187-6 to be correct, then the 3.5 km of

Loihi would require more than 1 million years to form, even if we allow for half of its thickness to be caused by coeval intrusives. This is much longer than recent estimates for the time required for the formation of an entire Hawaiian volcano [e.g., 650 ka (Moore and Clague, 1992); 750 ka (Frey et al., 1990); 1000 ka (Lipman, 1995)]. The apparent old age for sample 187-6 thus must be attributed to excess $^{40}Ar^*$. This excess $^{40}Ar^*$ can be considered as the result of incomplete degassing of magmatic Ar. Previous studies of Loihi lavas have shown that for some of them, the magmatic $^{40}Ar/^{36}Ar$ ratios range from 300 to 450 (Allegre et al., 1983; Kaneoka et al., 1983; Staudacher et al., 1991).

The age for sample 158-5 is also a problem; it is too young relative to ages above it. If the ages at the top and bottom of the section and this age were all correct, then the lava accumulation rate between 1960 and 1830 m b.s.l. was 1.4 mm/a and the rate between 1830 and 1130 m b.s.l. was 86 mm/a. Such a high lava accumulation rate is unrealistic. If we consider a typical duration of 650 ka of the hot spot related activity for the Hawaiian volcanoes (Moore and Clague, 1992), such a rate will produce edifices as thick as 56 km. Therefore, the young apparent age of sample 158-5 could indicate that the rock has not remained a closed system since eruption (i.e., it lost Ar) or that the sample is out of stratigraphic order. In re-examining notes taken when the sample was collected, we found that the rock was collected from a wall of pillow lavas that were coated with a bacterial mat. Such mats are found on Loihi only in areas of active or recently active hydrothermal venting (Karl et al., 1988). Thus, it is possible that this pillow lava issued from a vent on the wall at this depth and that its depth of collection is not related to its age. The second possibility, Ar loss resulting from alteration or devitrification of glass, is also plausible. Very small amounts of K-rich glass or its alteration products can exert a significant influence on a rock's potassium budget. This explanation has been proposed by Sharp et al. (1996) to interpret some young apparent ages for flows from Mauna Kea volcano in Hawaii. Although both explanations are plausible, the presence of the bacterial mat on the outcrop for sample 158-5 supports the young age for this sample. Vents have been found elsewhere on the flanks of Loihi, including at the top of this section and numerous dikes are present on the east flank (Garcia et al., 1995a).

5.2. Lava accumulation rates and temporal evolution of Hawaiian volcanoes

The lava accumulation rates for the east flank section of Loihi, based on the three stratigraphically reliable ages, are 3.5 mm/a for the lower part of the section and 7.8 mm/a for the upper part of the section (Fig. 2). This change in accumulation rate is consistent with the dramatic but systematic change in the rock types within the section. The lower part of the section (below 1450 m b.s.l.) consists of 88% alkalic lavas. From this depth to 1100 m b.s.l., 57% of the lavas are tholeiitic basalts and the percentage of tholeiitic lavas increases to 95% at the summit of the volcano (Garcia et al., 1995a). This transition is analogous (but in an opposite sense) to that observed during the postshield stage of Hawaiian volcanism, which forms as the volcano drifts off the hot spot (e.g., Frey et al., 1990).

The lava accumulation rate for Loihi's tholeiitic volcanism is identical to the one calculated for tholeiitic volcanism on the distal flanks of Mauna

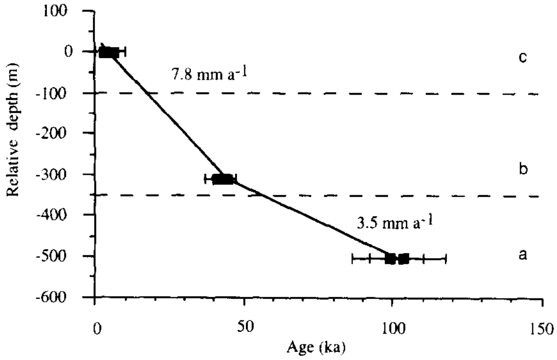

Fig. 2. Age vs. stratigraphic depth relationships for three dated Loihi samples. The depth of the lavas in the section has been calculated using a regional dip of 14° following the procedure discussed by Garcia et al. (1995a). The section has been subdivided into 3 parts; the basal part (below 350 m) is strongly alkalic (~ 88%), the middle section is mixed tholeiitic and alkalic lavas, and the uppermost part (< 100 m) is mostly tholeiitic (> 80%; see Garcia et al., 1995a, for details on rock type vs. depth variation). Both age determinations are plotted for each sample. The lava accumulation rate for the alkalic part of the section is 3.5 $mm\,a^{-1}$; the rate of lava accumulation during the transition from alkalic to tholeiitic volcanism was 7.8 $mm\,a^{-1}$.

Kea (Sharp et al., 1996) but lower than the lava accumulation rate for tholeiites along the axis of Kilauea's lower east rift zone between 42 ka and the present (11 mm/a; Garnier et al., 1996) and an area 10 to 15 km south of Kilauea's summit caldera (~ 10 mm/a; Easton, 1987). Both of Loihi's lava accumulation rates are greater than the rate determined for the lower flanks of Mauna Kea, which has interbedded alkalic and tholeiitic lavas (Frey et al., 1990), during its post-shield stage (0.9 mm/a; Sharp et al., 1996). A similar large variation in lavas accumulation rate was observed by Guillou et al. (1993) for the Fangataufa atoll in Tuamotu archipelago (7 mm/a for the submarine stage which is mainly tholeiitic and 0.7 mm/a for the later alkalic subaerial stage.

The duration of Loihi's transition from alkalic to tholeiitic volcanism can be estimated using an average lava accumulation rate based on our new ages (5–6 mm/a) and the thickness of the transition (100–200 m; Garcia et al., 1995a). This gives the transition a duration of 17 to 40 ka which is similar to the one estimated for the transition during the postshield stage of Haleakala volcano from tholeiitic to alkalic volcanism (< 25 ka; Chen et al., 1991). Thus, the alkalic/tholeiitic transition for both the preshield and postshield stages of Hawaiian volcanism are short compared to the overall duration of shield volcanism for Hawaiian volcanoes (750–1000 ka; Frey et al., 1990; Lipman, 1995). These results are consistent with the experimental results of Hirose and Kushiro (1993) which indicate that only a 25°C temperature increase is needed to change from weakly nepheline-normative melts (2%) to moderately hy-

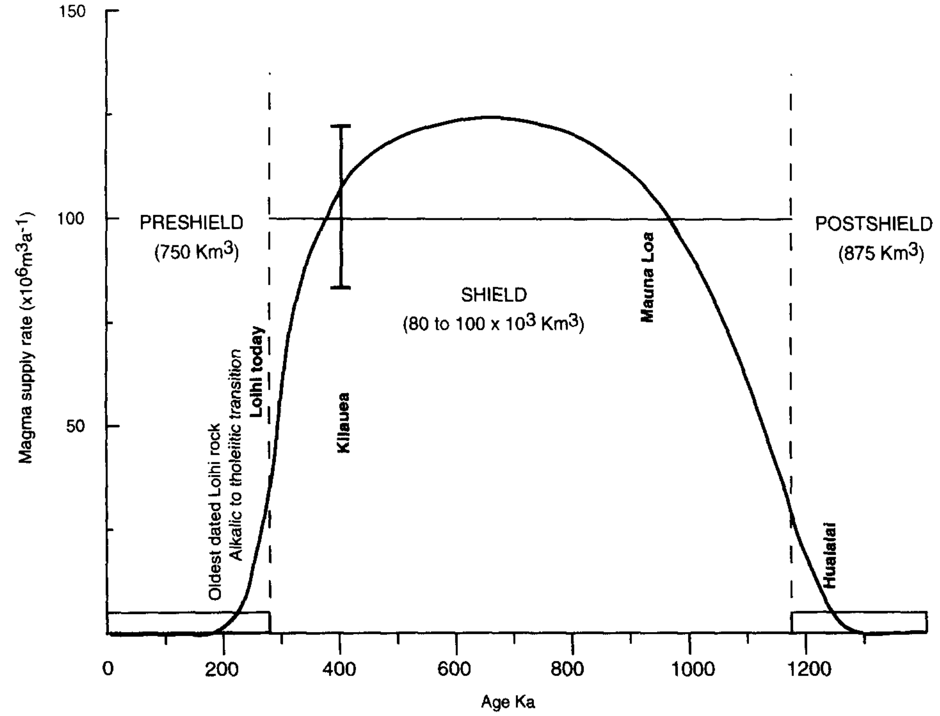

Fig. 3. Growth history model for a Hawaiian shield volcano after Frey et al. (1990) and Lipman (1995). This model is a composite using the volume estimates (rectangles) and growth rates for three different volcanoes for the three stages of growth of a Hawaiian volcano: Loihi for the preshield (Garcia et al., 1995a; this paper); Mauna Loa for the shield stage (Lipman, 1995; Garcia et al., 1995b); and Mauna Kea for the postshield (Frey et al., 1990). The magma supply rate for each stage has been inferred based on its duration and volume and a simple melting model (following the procedure of Frey et al., 1990). The magma supply rate range for Kilauea from 1955 to 1983 is given by the vertical bar for comparison (Dzurisin et al., 1984) and it shows that there are substantial short terms variations in magma supply rates for Hawaiian shield volcanoes. The duration of the preshield stage in this model is more than twice previous estimates.

perstene-normative melts (13%). During our estimated duration of the transition, Loihi would have drifted ~ 2 to 4 km closer to the locus of melting for the Hawaiian hot spot, assuming a 10 cm/a velocity for movement of the Hawaiian islands (Garcia et al., 1987).

Finally, our geochronological results for Loihi allow us to speculate on the duration of preshield volcanism for Hawaiian volcanoes. The east flank section represents only the uppermost part of Loihi (top ~ 0.5 km of the overall 3.5 km thickness of the volcano; Garcia et al., 1995a). The deeper interior of Loihi is probably dominated by intrusives, so the volcanic section may comprise a larger fraction of the volcano's total history than would be indicated by its thickness. On Kilauea, only about 1/3 of the magma that is intruded into the volcano is erupted (Dzurisin et al., 1984). If this ratio for intrusives to extrusives is valid for Loihi, then about 40 to 45% of Loihi's total thickness was formed while the east flank section was deposited (1.5/3.5). The line of reasoning gives us a minimum for the duration of the preshield stage of Hawaiian volcanism, which Loihi typifies. Thus, the preshield stage duration was at least 250,000 years. Although we have been conservative in this estimate, it is much longer than previous estimates (e.g., ~ 100,000 years; Frey et al., 1990; Lipman, 1995), which were based on simplistic melting models. Thus, the preshield stage does not have the rapid increase in melting (compared to the postshield stage) predicted by some models, and the duration of growth of Hawaiian volcanoes is somewhat longer than previously estimated (at least 150,000 years; Fig. 3).

Utilizing our new estimate for the duration of the preshield stage and new increased volume estimates for the size of Hawaiian shield volcanoes (e.g., Garcia et al., 1995b), the length of magmatic activity of a typical Hawaiian is postulated as ~ 1.4 Ma. Although this estimate is considerably longer than some recent estimates, it remains true that Hawaiian shield volcanoes are short-lived compared to many other hot spot volcanoes (e.g., 3.5 Ma for Fangataufa in the Tuamotus; Guillou et al., 1993). Furthermore given their great size (e.g., Mauna Loa is ~ 100 km^3; Garcia et al., 1995b), Hawaiian volcanoes are extremely productive volcanoes, which is consistent with their geoid expression (Sleep, 1990).

6. Conclusions

The unspiked K-Ar technique has been used to successfully date young submarine lava flows with ^{40}Ar* contents less than 1% from the Pitcairn and Hawaiian hot spots. The rocks from two Pitcairn hot spot seamounts give very different ages: Adams—3 to 7 ka; and Bounty—344 ka.

The Hawaiian samples are from the deeply dissected east flank of Loihi seamount. Their ages range from 4.5 to 200 ka, although the older age is probably unreliable based its stratigraphic position relative to other dated samples. Thus, the age range for this ~ 505 m thick section of lavas is from 4.5 to 101 ka. The lower alkalic portion of the section has a lava accumulation rate of 3.5 mm/a and the upper tholeiitic section has a rate of ~ 7.8 mm/a. The duration of the transition from tholeiites to alkalic lava during the preshield stage is about 17 to 40 ka and the total duration of the preshield stage of Hawaiian volcanism as represented by Loihi is estimated to be at least 250,000 years. Both estimates are much longer than previous models for Hawaiian volcanoes (e.g., the total duration of the preshield stage is more than twice previous estimates). We postulate that the average lifetime of a Hawaiian volcano is ~ 1.4 Ma based on our new ages and new volume estimates for these volcanoes. Although this estimate is considerably longer than previous estimates (e.g., 0.65 to 1.0 Ma; Moore and Clague, 1992; Lipman, 1995), it remains true that Hawaiian shield volcanoes have short but productive histories.

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