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$\xi = 1 - \exp[-6.7/dxt]$

where ξ is the fractional amount of transformation, *d* is grain size, \hat{x} is the growth rate [see (17)], and *t* is time. The extrapolation assumes that the reaction mechanism and rate observed in the laboratory also occur in nature—and this has not been proven. The extrapolation also applies only to natural situations with trace amounts of volatiles [see (5)].

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 $\dot{x} = b(kT/h) \{\exp(-\Delta G^*/RT)\}$

 $[1 - \exp(-\Delta G_{rxn}/RT)]\}$

where *b* is an empirical constant, *k* is Boltzmann's constant, *h* is Planck's constant, ΔG^* is the activation energy for growth, $\Delta G_{\rm rxn}$ is the volume free energy, *R* is the universal gas constant, and *T* is temperature [D. Turnbull, *Solid State Phys.* **3**, 225 (1956)]. We find that $\Delta G^* \approx 141$ kJ mol⁻¹ and $b \approx 1 \times 10^{-14}$ m; note that *b* is several orders of magnitude smaller than the interplanar spacings in carbonate structures, a value that is not expected on the basis of this simple theory.

- The equilibrium boundary below 600°C is from W. Johannes and D. Puhan [*Contrib. Mineral. Petrol.* 31, 28 (1971)]; that above 500°C is from B. R. Hacker and S. R. Bohlen (unpublished work).
- We thank B. Hankins and J. Pinkston for laboratory assistance; we thank G. Ernst, R. Fournier, J. G. Liou, J. Mosenfelder, G. Nord, D. C. Rubie, and two anonymous reviewers for comments. This work was supported by the Department of Energy and the U.S. Geological Survey.

20 April 1992; accepted 8 July 1992

An Experiment-Based Model for the Petrogenesis of High-Alumina Basalts

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To understand magmatism at convergent margins, one must know the origin of their characteristic, plagioclase-rich, high-alumina basalts (HABs). Wet melting experiments on basalts at 3 kilobars yielded high-alumina liquids and a coexisting mineral assemblage with little or no plagioclase. An isothermal pressure drop to 1 kilobar caused 20 to 30 percent plagioclase crystallization in these melts, while mafic minerals underwent limited crystallization or even resorption. These results suggest that hydrous (\geq 4 percent H₂O) HAB liquids, presumably formed by fractionation of a hydrous basalt at depth, will precipitate voluminous plagioclase as pressure drops during ascent and eruption. Plagioclase accumulation is not necessarily required to explain the petrogenesis of plagioclase-rich HAB.

High-alumina basalt is the most common mafic rock type in most arc systems. HABs are characterized by high Al_2O_3 , abundant modal plagioclase, and relatively evolved basaltic compositions (low Mg/Fe, low Cr, low Ni) (1–5). The controversy over whether HABs represent primary melts of the subducted slab or fractionated, mantlederived basalts (1–3, 6–7) has subsided in favor of the latter view. However, there remain questions as to whether typical HABs represent liquid compositions or are simply fractionated basalts with accumulations of plagioclase. Results of earlier experimental work appear to support both conclusions. Some experiments demonstrate that plagioclase is the sole liquidus phase in HABs at pressures of up to 19 kbar (dry) (8-10), suggesting that they could be plagioclase cumulates. Other work shows that HAB liquids can be generated by fractionation of an ultramafic phase assemblage from basalts under high pressure or high partial pressure of H_2O (P_{H_2O}) (11–13). Our experiments examined the crystallization behavior of water-saturated basalts at 3 kbar and during isothermal decompression to 1 kbar. They showed that HAB liquids can be generated by fractionation of a

SCIENCE • VOL. 258 • 2 OCTOBER 1992

plagioclase-free or -poor phase assemblage from basalt at 3 kbar ($P_{\rm H_2O} = P_{\rm total}$). In addition, our results demonstrated that these same HAB liquids crystallize plagioclase to the near exclusion of other phases when $P_{\rm H_2O}$ is isothermally lowered to 1 kbar. The apparently disparate behaviors observed in previous experiments (crystallization of a plagioclase-free and a plagioclase-rich phase assemblage from HAB liquids) were shown to depend on experimental conditions.

Starting materials were two rock powders, a mafic [(Mg)/(Mg+Fe) = 58] HAB (AT-4) from Atka in the Aleutians and a mafic [(Mg)/(Mg+Fe) = 68], olivine-phyric mid-ocean ridge basalt from the Deep-Sea Drilling Project leg 37 (L-37) (Table 1) (6, 14). The starting assemblage for both rocks includes plagioclase, olivine, clinopyroxene, oxide minerals, and glass. The experiments were conducted in gold capsules in an internally heated pressure vessel (15). All were done water-saturated at 1025°C because (i) water contents proposed for some natural HABs suggest that water saturation might occur between 1 and 3 kbar and (ii) gold capsules could be used to minimize Fe loss. One set of experiments was done at 1 kbar (4 days), one at 3 kbar (4 days), and one at 3 kbar (4 days) and then reduced isothermally to 1 kbar and held there for an additional 4 days. During the quench, solidus temperatures were reached in 30 to 90 s. Oxygen fugacity (f_{O_2}) was not buffered. Past experiments with a similar experimental setup yielded f_{O_2} values approximately 100 times that of the Ni-NiO buffer (12), and this value is consistent with the composition of magnetite in the 1-kbar experiments on AT-4. Other spinels in the experiments were rich in Cr. Melt and mineral phases were identified by backscattered electron imaging and analyzed with a focused, rastered, 5-nA, 15-kV beam on a Cameca Camebax microprobe. Sodium loss from hydrous glasses under these conditions has been shown to be around 30% (12), and glass analyses in Table 1 were corrected accordingly.

At 3 kbar, sample AT-4 was 89% melted at 1025°C. The melt, with roughly the composition of a HAB, coexisted with an assemblage of olivine, clinopyroxene, and chromium-rich magnetite. Abundant dendritic amphibole crystallized when this sample was quenched (Fig. 1). This texture required the use of a broad beam to determine prequench glass composition. The least differentiated glass analysis that we obtained from this sample is also given in Table 1 to show the effects of quench crystallization on glass compositions. Under the same conditions, L-37 produced 42% melt, coexisting with olivine, clinopyroxene, amphibole, plagioclase, and Cr-spinel.

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Plagioclase (93 mole percent anorthite, An_{93}) made up 9% of the charge. The liquid was rich in alumina (>24% Al_2O_3) and was andesitic rather than basaltic. Although the textural evidence was not as obvious, the amphibole in this sample may

also have formed upon quench. Magnesium-rich rims on olivine were perhaps quench features (Fig. 1). Therefore, the pre-quench liquid may have been more basaltic than the glass analysis in Table 1 shows.

Fig. 1. Backscattered electron images of the experimental charges. Scale bars, 100 µm; m, melt; p, plagioclase; o, olivine; c, clinopyroxene; a, quench amphibole. (A) AT-4, water-saturated, at 3 kbar and 1025°C. Note extensive crystallization of quench amphibole. Before quench, charge was mostly liquid. Clinopyroxene and olivine have accumulated at bottom of charge. (B) L-37, water-saturated, at 3 kbar and 1025°C; melting is not as extensive as in (A). Minor plagioclase (An₉₃) is present with olivine, clinopyroxene,



and amphibole (perhaps quench). (C) AT-4, water-saturated, 1025°C, held at 3 kbar for 4 days then dropped isothermally to 1 kbar and held for 4 days. Extensive crystallization has occurred. Of particular interest are the euhedral plagioclase crystals, up to 1 mm in length, that contain inclusions of glass and clinopyroxene, and the skeletal (tubelike) olivine crystal. (D) L-37, with conditions as in (C). Large, lathlike plagioclase crystals with glass inclusions are present. Quench crystallization of amphibole in (C) and (D) is relatively minor.

In sample AT-4, approximately 28% plagioclase (An₈₀) crystallized during and after the isothermal drop from 3 to 1 kbar. The plagioclase formed euhedral crystals up to 1 mm in dimension (in a 3-mm charge) (Fig. 1). Evidence that the plagioclase grew rapidly includes inclusions of glass and Alrich clinopyroxene. The amount of liquid in the sample dropped from nearly 90% to < 50%, and its composition was substantially less aluminous than the 3-kbar liquid (Table 1). Quench amphibole was present but not abundant. Similar effects were noted for L-37. Modal plagioclase increased from 9 to 28%, and the liquid became much less aluminous. Interestingly, the liquid produced in the pressure drop experiment was more basaltic than the liquid produced in the 3-kbar experiment. For the most part this compositional change was due to the resorption of all amphibole and 25% of the clinopyroxene during the pressure-drop experiment. In sample AT-4, 68% of the new crystallization that occurred during pressure drop was plagioclase. In sample L-37, 95% of the new crystallization was plagioclase, and olivine was the only other mineral phase with an increase in abundance (Fig. 2).

The water-saturated melting experiment at 1 kbar on AT-4 produced about 24% rhyodacitic liquid coexisting with plagioclase, olivine, clinopyroxene, and magnetite. Although the liquid did not appear to be in equilibrium with the mineral assemblage (the

Table 1. Starting materials and experimental results. All analyses normalized to 100% volatile-free. Glass analyses corrected for 30% Na loss; na, not analyzed.

Component or phase	Starting materials		Experimental results					
			AT-4				L-37	
	L-37	AT-4	1 kbar $(n = 3)$	3 kbar* (<i>n</i> = 1)	3 kbar† (<i>n</i> = 2)	3 to 1 kbar ($n = 8$)	3 kbar (<i>n</i> = 5)	3 to 1 kbar (n = 6)
SiO ₂	48.34	48.62	69.28	52.48	49.67	50.62	56.96	52.96
TiO ₂	0.62	0.91	0.45	0.85	0.80	1.17	0.56	1.21
Al ₂ Ō ₃	15.98	17.38	16.88	21.00	19.40	17.41	24.25	16.89
Fe ₂ O ₃	1.62	6.21						
FeŌ‡	7.80	4.19	0.95	6.42	9.00	10.35	3.70	10.20
MnO	0.13	0.18	0.09	0.34	0.16	0.19	0.14	0.24
MgO	11.27	7.66	0.04	2.43	5.53	4.68	0.70	4.70
CaO	11.84	11.64	2.06	12.51	11.82	10.05	9.26	9.04
Na ₂ O	1.98	2.44	5.10	3.24	2.80	4.62	4.12	4.21
K₂Õ	0.14	0.63	4.38	0.43	0.70	1.01	0.22	0.31
P ₂ O ₅	0.06	0.19	0.79	0.20	na	0.29	0.14	0.20
Total§	95.84	99.31	92.76	93.43	89.90	94.60	91.20	94.66
Plagioclase			37.5	_	_	28.0	9.2	27.5
Clinopyroxene			22.3	7.2	5.9	13.5	24.6	18.0
Olivine			10.8	8.1	4.3	9.0	15.9	16.8
Amphibole			_	9.3	_	_	5.5	_
Spinel			5.9	2.6	0.8	1.7	2.8	1.0
Glass			23.5	72.7	89.0	47.8	41.9	36.7
Sum r²¶			0.50	0.11	0.03	0.56	0.01	0.04

*Glass apparently least modified by quench effects. Other glasses richer in Al, Si, and alkalis and poorer in Fe, Mg, and Ca. tBroad-beam analysis of quench amphibole and glass (areas approximately 300 by 300 μm). This method probably provides the best estimate of prequench glass composition for this sample. Use of a broad-beam largely eliminates Na loss, so no correction was applied to this analysis. Relatively low total resulted from the use of a broad beam. For glasses, all Fe is given as FeO. \$Anhydrous totals for the starting materials and uncorrected microprobe totals for the experimental glasses. #Microprobe analyses of spinel (magnetite) from another charge used in mode calculation. Actual spinels in L-37 charges were Cr-rich. #Sum of the squares of the residuals for the least-squares mixing models. mafic phases are too rich in Mg; zoned minerals are common), the experiment indicated that the equilibrium assemblage at 1 kbar is plagioclase-rich (between 28 and 38% plagioclase by weight). The 1-kbar run of L-37 was also rich in plagioclase.

The force behind plagioclase crystallization in the depressurization experiments was the drop in P_{H_2O} accompanying depressurization (16, 17). This effect occurs in natural systems that become water-saturated during ascent. Current estimates of the water content of natural HAB liquids range from $\leq 2\%$ by weight (8, 9) to 4 to 6% (18, 19). The results of these experiments are most directly applicable to water-rich HAB magmas (4 to 6% H₂O) that achieve water saturation between 1.5 and 3 kbar (17). For these water-rich HABs, crystallization of voluminous plagioclase accompanies ascent through the middle and upper crust, even in magmas that are free of plagioclase at high pressure.

For systems poorer in H_2O , the results of these experiments are less definitive. Results of earlier experiments indicate that for a given basalt composition, both high pressure and high water content suppress the crystallization of plagioclase in basalts (8, 10–13). They also imply that drier, lower pressure conditions increase the stability of plagioclase (8, 10). Given the maximum crustal pressures attainable in oceanic arcs (7 to 10 kbar) (20) and experimental results that suggest 8 to 19 kbar of pressure is necessary to remove plagioclase from the liquidus of dry HAB (8, 10, 13), some water is probably required for most HAB

Fig. 2. Changes in (A) total mode and (B) mineral mode accompanying the isothermal pressure drop from 3 to 1 kbar. In L-37, clinopyroxene (cpx), amphibole (amph), and oxide minerals (mt) were resorbed the pressure during drop. Plagioclase (plag) was by far the most abundant new crystallizing phase for both starting compositions; ol, olivine. In (B), the percentages of new crystal production for AT-4 were mt, 2%; ol, 11%; cpx, 18%; and plag, 68%. For L-37 in (B), the crystal fractions were mt, partially absorbed; amph, resorbed; ol, 5%; cpx, partially resorbed; and plag, 95%.

liquids to form by fractionation. The key (and presently unresolved) question is: How much water? This and earlier studies indicate that 4 to 6% water is sufficient. The work of Gust and Perfit (13) suggests that substantially less than 4% water (and perhaps none) may be enough under some circumstances. More work is required in systems containing 1 to 3% water to determine whether HAB magmas can form by crystal fractionation in these drier (and hotter) systems.

However, once a HAB liquid has formed and segregated, all experimentally determined phase relations dictate that plagioclase crystallization will eventually occur during ascent, whether or not the magma becomes water-saturated. The initial depth of plagioclase crystallization is a function of magma composition and varies inversely with water content.

Finally, relative superheating of the primitive arc magma with respect to plagioclase may play a subsidiary role in promoting either the initial suppression or later crystallization of plagioclase in HAB (21). Lofgren (22) has shown in 1-atm experiments that elimination of plagioclase nuclei at subliquidus temperatures suppresses plagioclase crystallization in basaltic melts cooled at 2°C per hour (100 hours) to the solidus. The effects were much less pronounced for pyroxene and olivine. In fact, pyroxene crystallized in these experiments even though its appearance temperature is below that of plagioclase. Glass compositions in the experiments in which plagioclase crystallization was suppressed were



extremely aluminous (23). Relative superheating might influence the kinetics of plagioclase nucleation and crystallization in natural HAB. This would lead initially to the formation of highly aluminous liquids supersaturated in plagioclase and, ultimately, to an increase in the force behind plagioclase crystallization.

The chemistry and petrography of HAB can form as a natural consequence of the fractionation, crystallization, ascent, and emplacement or eruption of hydrous basaltic magmas. Passage from high to low pressure (as in ascent) or loss of water (as in emplacement at shallow levels in the crust or during eruption) will result in the onset of plagioclase crystallization. The experiments described here indicate that plagioclase crystallization can be substantial (20 to 30% by weight) and rapid and may proceed to the near exclusion of other phases.

They also show that plagioclase accumulation is not required to explain the chemistry or petrography of HAB. Of course, some plagioclase accumulation may occur in HAB just as in any basalt (24). However, the special pleading for arc magmas that is inherent in the accumulation argument (that is, HAB and, hence, plagioclase accumulation is rare outside of arcs) is not required.

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- 25. We thank B. Marsh for providing a sample of AT-4, A. Lanier and W. Carter for their help with the high pressure-high temperature experimental apparatus, G. McKay and V. Yang for access to and help in using the microprobe facility at Johnson Space Center, and E. Compton-Gooding for drafting the figures. We also thank D. Johnston and an anonymous reviewer for their comments on the manuscript. Supported by National Science Foundation grant EAR-9120035.

26 May 1992; accepted 15 July 1992

Sensitivity of Glaciers and Small Ice Caps to Greenhouse Warming

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Recent field programs on glaciers have supplied information that makes simulation of glacier mass balance with meteorological models meaningful. An estimate of world-wide glacier sensitivity based on a modeling study of 12 selected glaciers situated in widely differing climatic regimes shows that for a uniform 1 K warming the area-weighted glacier mass balance will decrease by 0.40 meter per year. This corresponds to a sea-level rise of 0.58 millimeter per year, a value significantly less than earlier estimates.

The amount of ice stored in glaciers and small ice caps is only a fraction of the total amount of land ice on earth. Nevertheless, because glaciers have a shorter response time than the large ice sheets of Greenland and Antarctica, they are thought to be important for sea-level variations on a decadal and centurial time scale. The total volume of water contained in glaciers outside Greenland and Antarctica is not well known—it is probably equivalent to a sea-level rise between 0.3 and 0.7 m. Data on glacier area are more easy to obtain, and a current global estimate is 549 × 10^3 km^2 (1).

Global sea level has risen by 10 to 20 cm over the last hundred years (2). Explanation of this rise is a matter of debate. As it is known that mountain glaciers have retreated in most parts of the world since the late 19th century, it is likely that they made a contribution to the observed sea-level rise. An admirable attempt to quantify this contribution, using available mass-balance data of glaciers, was made by Meier (3), who arrived at a contribution on the order of several centimeters. Critical in his approach was the assumption that the climate sensitivity of glaciers is proportional to the annual balance amplitude (4), a measure of the mass turnover in a glacier. The assumption was needed for extrapolation of the scarce mass-balance measurements to large areas. In spite of the uncertainties, Meier's work made clear that, when considering sealevel fluctuations on a 100-year time scale, glaciers are a factor to deal with.

In this feport, we follow a process-oriented approach and use an energy-balance model for the ice or snow surface to generate annual mass-balance for a selection of 12 glaciers, for which reliable mass-balance observations extending over a number of years exists. After having tuned the model to give a sufficiently accurate simulation of the observed mass-balance fields,

Fig. 1. Mass balance as a function of altitude for the 12 glaciers selected for this study (*5*). Note the wide variety in equilibrium line altitude and balance gradients (change of mass balance with altitude). Small balance gradients are found on glaciers in the drier polar and subpolar regions, like Devon ice cap and White glacier in the Canadian Arctic. Engabreen (Enga) and Nigardsbreen (Nigards), in western Norway, have a tremendous mass turnover and large balance gradients. The same applies to the Rhone glacier in the Swiss Alps. The mass turnover of Hintereisferner (Hintereis) (Austrian Alps) is moderate. Hellstugubreen (Hellstugu) is located in the drier part of southern Norway, we perform a sensitivity analysis.

The specific (or surface) mass balance of a glacier is the net amount of ice mass gained through a year at the surface of a glacier per unit area. We express the mass balance in meters of water equivalent per year. Precipitation, radiation, and air temperature are the most important factors determining glacier mass balance. For a specific region, the altitudinal gradient is most evident and is a direct consequence of the atmospheric temperature lapse rate. We refer to a mass-balance profile B(h) as the balance that depends on altitude h. As elevation increases, B(h) changes sign (from negative to positive) at the equilibrium line. The equilibrium line altitude E varies widely over the globe from greater than 4000 m in the subtropics, to 2900 m in the Alps, and to less than 1000 m in the subpolar and polar regions.

Observed mass-balance profiles for the set of 12 glaciers show a wide variety in character (Fig. 1). These are only a small part of all the glaciers and ice caps for which mass-balance measurements have been performed. The present selection represents well the wide range of climatological settings in which glaciers and small ice caps are found, perhaps with the exception of the tropics. Glaciers in the tropics, however, contribute a negligible amount to the total ice volume.

The mean specific balance of a glacier, denoted by B_m , is the mass balance averaged over the entire glacier. To obtain B_m for a specific glacier, the area-elevation distribution S(h) (hypsometry) must be known. So

$$B_m = \frac{1}{S_T} \sum B(h_i) S(h_i)$$

Here the sum is taken over the entire glacier, where the index *i* refers to the elevation interval centered around h_i . S_T is the total area of the glacier. When a noncalving glacier is in perfect balance, $B_m = 0$.

The mass-balance model we use is based on the following equation:



and Peyto glacier is in the central Rocky Mountains in Canada. Abramov glacier and Tuyuksu glacier are located in central Asia and are more representative of glaciers in the subtropics.

SCIENCE • VOL. 258 • 2 OCTOBER 1992

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