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Comparison of hydrothermal systems in layered gabbros and granites, and the origin of low–^{18}O magmas*

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Abstract—The style of hydrothermal alteration in layered gabbros is very different from that in granitic plutons. Non-equilibrium ^{18}O/^{16}O effects are observed in both types of bodies, but whereas the granitic rocks with ^{18}O–exchanged feldspars commonly contain abundant chlorite, sericite, epidote, etc., most such gabbros are mineralogically virtually unaltered (*e.g.*, they contain fresh olivine). These differences imply that the bulk of the externally–derived hydrothermal fluid passes through the gabbros at temperatures of 450°–900°C, much higher than the range of 250°–450°C generally found in the granites. This contrasting behavior is in part explained by the higher solidus temperature, higher latent heat of crystallization, and generally higher melt/phenocryst ratio of the gabbro magmas. However, even more important is: (1) *The geometry of crystallization.* Cumulate gabbros crystallize from the floor upward, and the last part to crystallize is a sub-horizontal sheet of liquid near the roof. This magma sheet provides a thermally insulating lid that is impermeable to the (hydrostatic) hydrothermal convection system in the country rocks, typically producing two decoupled hydrothermal systems, a lower–temperature system at high water/rock ratios above the intrusion, and a much higher–temperature system at low water/rock ratios within the cumulate gabbro below the late–stage magma sheet. (2) *Presence of a magmatic H_2O envelope.* Granitic magmas typically contain much more H_2O than tholeiitic gabbro melts; hence, evolution of a magmatic gas phase will usually occur at the late stages of crystallization of granites. This envelope of magmatic water fills available fractures and is under lithostatic pressure. It is thus impermeable to the outer, convecting meteoric–hydrothermal fluids, which gain access to the pluton only after the magmatic H_2O has dissipated and the body has cooled.

The two major Cenozoic occurrences of low–^{18}O magmas are Iceland and the Yellowstone Plateau, Wyoming. These types of magmas are much less common than heretofore believed, and they typically seem to be developed only in extensional tectonic environments where brittle fracture of the crust allows continued replenishment of the magma reservoir from below, as well as penetration of meteoric ground waters down to great depths. Such magmas are formed by melting or assimilation, or both, of hydrothermally altered volcanic roof rocks by the underlying magma reservoir, *not* by influx of low–^{18}O ground waters directly into the melt.

INTRODUCTION

THE AIM of this paper is two–fold: (1) To point out intrinsic differences in the style of hydrothermal alteration between layered gabbros and granites, specifically trying to explain the observation that ^{18}O–depleted granites are almost invariably *mineralogically* heavily altered to chlorite, sericite, epidote, etc., whereas their counterparts, the low–^{18}O layered gabbros (that also have clearly been altered by meteoric–hydrothermal fluids), are either mineralogically *unaltered* or only weakly altered (*e.g.,* typically containing fresh olivines, for example). (2) To review some of the examples of low–^{18}O magmas formed on Earth, to discuss their possible origin, and to make a case that the mechanism of formation of low–^{18}O magmas is usually operative only in rift–zone environments, which is also where the best examples of hydrothermally exchanged, but mineralogically unaltered, low–^{18}O layered gabbros are found.

First, we briefly review the isotopic systematics displayed by the Skaergaard intrusion (the best described ^{18}O–depleted gabbro in the world), and then we compare the sub–solidus hydrothermal isotope effects observed in such low–^{18}O gabbros with those in low–^{18}O granites. Less discussion is devoted to the granites, because the hydrothermal alteration effects in these bodies are familiar to most petrologists, and the effects are produced at ordinary hydrothermal temperatures of 200°–400°C (and thus are not controversial). In contrast to the above types of effects that are characteristic of low–^{18}O igneous *rocks,* in the second part of the paper we briefly review the problems of low–^{18}O *magmas,* particularly in Iceland and Yellowstone Park, Wyoming, the two largest and best documented areas of low–^{18}O magmas in the world. We conclude with a critical evaluation of some of the pertinent facts that bear on the mechanism of formation of low–^{18}O magmas.

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Some of the discussion that appears in this paper is also included in several recent papers by CRISS and TAYLOR (1986), LARSON and TAYLOR (1986), HILDRETH *et al.* (1984), and TAYLOR (1986).

COMPARISON OF HYDROTHERMAL EFFECTS IN GABBROS AND GRANITES

Permeabilities of rocks

Permeability is the single physical parameter that is of most importance in determining hydrothermal fluid flow through rocks. The permeabilities of common geologic materials vary by a factor of about 10^{12}, from about 10^{-6} cm^2 (100 darcies) to about 10^{-18} cm^2 (10^{-7} millidarcies). A permeability of about 10^{-14} cm^2 is generally considered to be impermeable, as far as hydrothermal systems are concerned (NORTON and KNIGHT, 1977).

The permeabilities of detrital sediments increase markedly with increased grain size and with a decrease in the degree of cementation and compaction. Limestones exhibit widely ranging permeabilities, with low values in argillaceous samples and with extremely high values characterizing cavernous types. Basalt permeabilities are also highly variable, with low values being typical of ancient flows that are strongly altered or metamorphosed; high values are found in highly jointed lavas in areas of recent volcanic activity. Crystalline rocks have relatively low permeabilities except where they are fractured. Because fracture permeability is a large-scale phenomenon that may be dominated by a few, large, widely-spaced fissures, the permeabilities of crystalline rocks measured on a small scale in the laboratory are commonly several orders of magnitude lower than the true *in situ* permeabilities (*e.g.*, as measured in drill holes, BRACE, 1984).

It has become very clear through studies of ^{18}O/^{16}O effects in metamorphic terranes (*e.g.*, GARLICK and EPSTEIN, 1967; RYE *et al.*, 1976; FORESTER and TAYLOR, 1977) that many rocks undergoing ductile or plastic deformation are relatively impermeable to hydrothermal fluids. This is conceptually easy to understand, because such rocks will deform rather than fracture, thus closing up any through-going fractures and greatly reducing permeability. Examples would be salt domes and evaporite beds in sedimentary basins or relatively pure limestones and marbles in areas of strong folding or metamorphism (CRISS and TAYLOR, 1986).

There are some very peculiar notions about permeability in the literature. For example, CATHLES (1983) writes: "We conclude that intrusions hotter than ~350°C must be quite impermeable—otherwise surface venting of solutions 400°C to 500°C or even 600°C or hotter would be commonplace. We conclude that significant fluid circulation does not occur in rock hotter than ~400°C." This statement seems to be contradicted by a great deal of geological and geochemical data, particularly by combined stable isotope and mineralogical data on layered gabbros (see below). However, such statements merely highlight the fact that we have at present very little *direct* knowledge of bulk permeabilities of rocks at high temperatures and pressures.

^{18}O/^{16}O effects in hydrothermally altered intrusions

Figure 1 shows the steep, linear trajectories on plots of δ^{18}O feldspar vs. δ^{18}O quartz or δ^{18}O pyroxene that are a characteristic of hydrothermally altered rocks in the Earth's crust (TAYLOR and FORESTER, 1979; CRISS and TAYLOR, 1983; MAGARITZ and TAYLOR, 1986). The systematics and implications of these trajectories are discussed by CRISS *et al.* (1986) and GREGORY and TAYLOR (1981, 1986). Basically, the trajectories result from the fact that: (1) feldspar exchanges ^{18}O with hydrothermal fluids *much* faster than does the coexisting quartz or pyroxene; and (2) typically, external waters en-

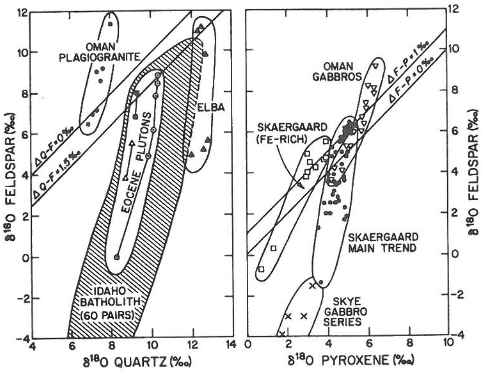


FIG. 1. The left-hand diagram is a plot of δ^{18}O–feldspar versus δ^{18}O quartz for hydrothermally altered granitic rocks from the Idaho batholith (lined pattern), individual Eocene plutons from the Idaho batholith, a Tuscan granodiorite from the island of Elba, Italy, and a plagiogranite from the Oman ophiolite complex (data from CRISS and TAYLOR, 1983; TAYLOR and TURI, 1976; and GREGORY *et al.*, 1980). The right-hand diagram is a plot of δ^{18}O–feldspar versus δ^{18}O pyroxene for hydrothermally altered rocks from the layered gabbro complexes from the Skaergaard intrusion, the island of Skye, Scotland, and the Oman ophiolite (data from TAYLOR and FORESTER, 1979; FORESTER and TAYLOR, 1977; and GREGORY and TAYLOR, 1981). In all cases, marked isotopic disequilibrium is observed in these altered rock suites.

tering a rock system will have $\delta^{18}O$ values that are either lower (meteoric H_2O, high-temperature sea water) or higher (metamorphic waters, sedimentary formation waters, or low-temperature sea water) than the H_2O that would be in equilibrium with a given igneous mineral assemblage at a particular temperature.

Because the in-flowing H_2O is not in isotopic equilibrium with the mineral assemblage, the mineral assemblage will be pulled to either higher or lower $\delta^{18}O$ values; however, because the feldspar reacts faster with the H_2O than does the quartz or pyroxene, the feldspar undergoes ^{18}O exchange at a much faster rate than the coexisting quartz or pyroxene, thereby producing the steep trajectory. The process seldom goes to completion, so the final mineral assemblage is in isotopic disequilibrium, and the disequilibrium $\delta^{18}O$ values provide a "signature" of the hydrothermal event. This isotopic signature has a distinct advantage over the simple mineralogical signature of the hydrothermal episode (*e.g.*, the presence of chlorite, sericite, etc.), because the effects are still observable even at very high temperatures where the original igneous mineral assemblage may be stable (*e.g.*, pyroxene hornfels facies). In addition, the $\delta^{18}O$ effects can provide a much better estimate of the amount of water that moved into the pluton than one can get from the mineralogical data alone. This amount of water is invariably far in excess of the small amounts required simply to make the new OH-bearing minerals. These $\delta^{18}O$ effects are thus the best and least ambiguous way to characterize a fossil hydrothermal system, *if* the external water is isotopically sufficiently different from the pluton.

Isotopic relationships in the Skaergaard intrusion

The best studied example of a fossil meteoric-hydrothermal system associated with a gabbro body is that of the 55 Ma Skaergaard intrusion (TAYLOR and EPSTEIN, 1963; TAYLOR and FORESTER, 1979; NORTON *et al.*, 1984). This 70 km^2 pluton was emplaced during the early stages of opening of the North Atlantic Ocean as a single pulse of magma of tholeiitic basalt composition. The pluton cuts across a subhorizontal unconformity separating Precambrian gneiss from an overlying, 9-km thick sequence of tholeiitic lavas (see Figure 4 below).

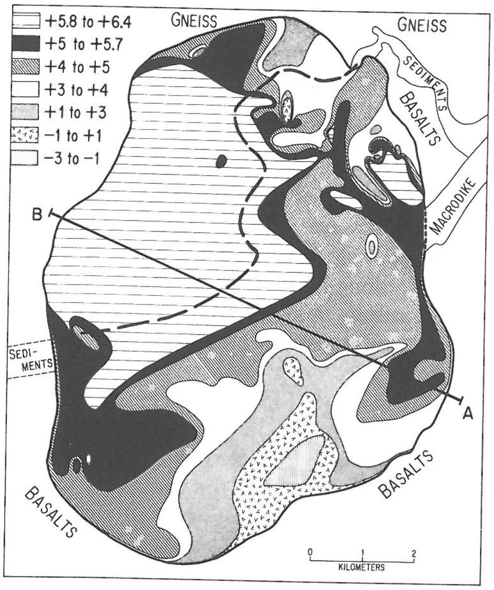


FIG. 2. Generalized distribution of $\delta^{18}O$ values of plagioclase within the Skaergaard intrusion (after TAYLOR and FORESTER, 1979). The 'normal' $\delta^{18}O$ values (+5.8 to +6.4) occur in the NW portion of the pluton, stratigraphically below the heavy dashed curve. This curve represents the projected trace on the outcrop of the unconformity between the highly permeable basalts and the gneiss. The line A–B is the position of the cross-section shown in Figures 4 and 5.

The $\delta^{18}O$ values in the Skaergaard intrusion decrease markedly from northwest to southeast (Figure 2). Essentially pristine $\delta^{18}O$ values ($\delta^{18}O$ plag $= 6.2 \pm 0.2$) occur in the northwest, in the lower parts of the cumulate sequence that lie stratigraphically beneath the projection of the major regional unconformity. The lowest $\delta^{18}O$ values (to -2.4) occur highest in the intrusion in the Upper Border Group; this part of the pluton is surrounded by the very permeable Early Tertiary basaltic lavas. Oxygen isotopic disequilibrium between coexisting plagioclase and pyroxene in this intrusion is clearly indicated by the steep positive slope of the data points in Figure 1.

The contrast in permeability across the regional unconformity clearly played an important role in the development of the hydrothermal $\delta^{18}O$ pattern in the Skaergaard intrusion (see Figure 4 below). Vigorous circulation in the permeable basalts allowed penetration of considerable amounts of H_2O into the solidified gabbro *above* the level of the unconformity, southeast of the heavy dashed line on Figure 2. Below this horizon, much smaller amounts of H_2O penetrated into the gabbro, because such H_2O had to travel through the much less permeable gneiss.

Hydrous minerals such as hornblende, biotite, chlorite, actinolite, and stilpnomelane are found only as rare alteration products within the gabbro matrix, or as replacement and open-space fillings along fractures, in veins, and in miarolitic cavities.

As shown in Figure 3, all of the OH–bearing minerals in the vicinity of the Skaergaard intrusion (in the gabbros themselves, in the basement gneiss, and in the plateau basalts) have very low δD values indicative of meteoric–hydrothermal exchange. The overall water/rock ratios in the deepest part of the system (the basement gneiss) had to be less than about 0.05, however, because there is a general lack of any discernible ^{18}O depletion in the gneiss, except (1) locally along the western contact of the Skaergaard intrusion, and (2) along a north–trending fracture zone containing brick–red, turbid feldspar and thoroughly chloritized mafic minerals (red gneiss $\delta D = -136$, Figure 3). These data are important because they indicate that along major fracture conduits, there *was* significant meteoric water circulation through the basement, approximately 10 km beneath the Eocene surface.

Blocks of basaltic roof rock and gabbros from the early–crystallized Upper Border Group were depleted in ^{18}O by the hydrothermal activity *before* they fell into the magma (TAYLOR and FORESTER, 1979). For example, a 6 m–wide basalt xenolith in the Middle Zone of the layered series has $\delta^{18}O = -4.0$. However, in spite of the fact that the hydrothermal system was operating for the entire 130,000–year crystallization history of the pluton (NORTON and TAYLOR, 1979), there was no measurable depletion of ^{18}O in the liquid magma. This conclusion is proved by the fact that the pyroxenes throughout the layered series have relatively normal $\delta^{18}O$ values, commonly higher than the more easily exchanged plagioclase. Thus, most of the observed ^{18}O depletion in the Skaergaard intrusion was confined to the plagioclase, and this occurred after crystallization under sub–solidus conditions.

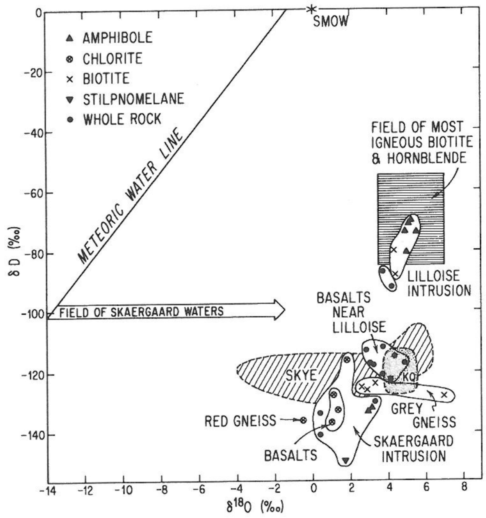


FIG. 3. Plot of δD vs. $\delta^{18}O$, showing the position of the meteoric water line and SMOW (standard mean ocean water, defined as equal to zero for both $\delta^{18}O$ and δD). The positions of the Skaergaard and Lilloise plutons and the East Greenland basalts and gneiss are indicated, as well as the general field of the Skye igneous center in NW Scotland and the field of another pluton in East Greenland (Kangerdlugssuaq, KQ), together with the 'normal' values of most primary igneous biotites and hornblendes (modified after TAYLOR and FORESTER, 1979).

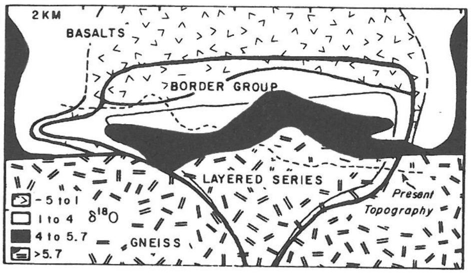
Numerical modeling of the Skaergaard intrusion

NORTON and TAYLOR (1979) carried out a computer simulation of the Skaergaard magma-hydrothermal system, producing detailed maps of the temperature, pressure, fluid velocity, integrated fluid flux, and $\delta^{18}O$ values in rock and fluid as a function of time for a two–dimensional cross–section through the pluton. An excellent match was made between calculated $\delta^{18}O$–values and the measured $\delta^{18}O$ values in the three principal rock units, basalt, gabbro, and gneiss, as well as in xenoliths of roof rocks that are now embedded in the gabbro cumulates (compare Figures 4a and 4b). The best match was realized for a system in which the bulk rock permeabilities were 10^{-13} cm^2 for the intrusion, 10^{-11} cm^2 for basalt, and 10^{-16} cm^2 for gneiss. These represent *average* permeabilities for the lifetime of the hydrothermal system. Because of self–sealing by mineral deposition in fractures, it is likely that during the early, highest–temperature stages of hydrothermal alteration, the pluton had a somewhat higher permeability of about 10^{-12} cm^2 (NORTON *et al.,* 1984).

The thermal history calculated for the Skaergaard system by NORTON and TAYLOR (1979) showed that extensive fluid circulation was largely restricted to the permeable basalts and to regions of the pluton stratigraphically above the basalt-gneiss unconformity. During the initial 130,000–year period of crystallization, fluids circulated in all the rocks surrounding the magma body, but fluid flow paths were deflected around the impermeable magma sheet. After final crystallization of the late–stage sheet of magma, fractures could form in the gabbro, allowing the circulation system to shift toward the center of the intrusion (Figure 5).

During the initial 150,000 years, the average temperature of the intrusion was high (>700°C) and reaction rates were fast; thus, fluids flowing into the intrusion quickly equilibrated with plagioclase. By 500,000 years, the pluton had cooled to approximately ambient temperatures, and the final $\delta^{18}O$ values were 'frozen in'. Reactions between hy-

A



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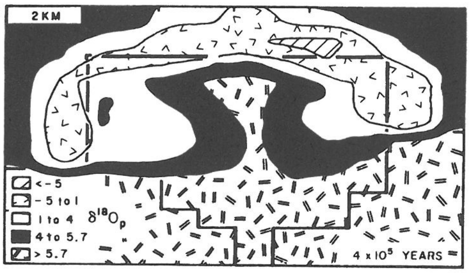

FIG. 4A. Composite geological cross-section of the Skaergaard intrusion, showing interpolated δ^{18}O-values in plagioclase with respect to major rock types and topography along the restored geological section line A–B shown in Figure 2 (after data by TAYLOR and FORESTER, 1979).

FIG. 4B. Calculated δ^{18}O values of plagioclase in an idealized cross-section of the Skaergaard intrusion at 400,000 years (see Figure 5), based on the numerical modeling study of this hydrothermal system by NORTON and TAYLOR (1979). The good match between the above figures indicates that the bulk permeabilities assumed for the various rock types (gneiss = 10^{-16} cm^2, gabbro = 10^{-13} cm^2, basalts = 10^{-11} cm^2) are probably quite accurate (NORTON and TAYLOR, 1979).

drothermal fluid and the intrusion occurred over a broad range in temperature, 1000°–200°C, but 75 percent of the fluid circulated through the intrusion while its average temperature was >480°C (NORTON and TAYLOR, 1979). The relative quantities of water to rock integrated over the entire cooling history were 0.52 for the upper part of intrusion, 0.88 for the basalt, and 0.003 for the gneiss (weight units).

One of the important conclusions of NORTON and TAYLOR (1979) is that most of the sub-solidus hydrothermal exchange in the Skaergaard pluton took place at very high temperatures (400°–800°C); this is compatible with the general absence of hydrous alteration products in the mineral assemblages, and with the presence of clinopyroxenes and high-temperature amphiboles in the veins and fractures; the latter were deposited both before and after the intrusion of cross-cutting granophyre dikes and sills (NORTON *et al.*, 1984). These granitic magma bodies are not chilled at their margins, so emplacement clearly took place at quite high temperatures (WAGER and DEER, 1939). Of course, outside in the country-rock basalts the average temperature of hydrothermal alteration was much lower (200°–400°C), and there was extensive development of hydrous minerals such as chlorite, epidote, and prehnite; locally, along late-stage veins, these minerals also developed in the pluton.

Water/rock ratios in the Skaergaard intrusion

The results of the calculations of NORTON and TAYLOR (1979) for the Skaergaard intrusion illustrate the relationship between material-balance water/rock ratios and the calculated amounts of water that physically flow through the system and thus might actually interact with the rocks. Figure 5 shows the integrated mass flux of fluid at various times as a function of stratigraphic position in the Skaergaard system. Over the 500,000-year lifetime of this hydrothermal system, and above the level of the unconformity, integrated amounts of 100 to 5000 kg of H_2O have flowed through each square centimeter cross-section of rock. This means that *along* the flow path each cm^3 of rock has been exposed to something on the order of 100,000 to 5,000,000 cm^3 of fluid (remember that the permeability is only 10^{-12} to 10^{-13} cm^2!). However, the water/rock (W/R) ratios for various parts of the system, calculated either by material balance or by choosing a sufficiently large-sized mass of rock, are much smaller. For example, the overall, integrated W/R ratio for that part of the intrusion lying above the unconformity is only 0.52 in weight units (NORTON and TAYLOR, 1979), and the instantaneous W/R ratio is probably no more than 0.01 to 0.03 (the latter is essentially completely controlled by the numerical value of the interconnected porosity of the crystalline gabbro at the appropriate pressure and temperature).

The above "discrepancy" in W/R is a simple consequence of the fact that each cm^3 of rock along a given flow path sees the same packet of water that passed through all of the previous cm^3 volumes of rock encountered along this flow path. Because each cm^3 of fluid has chemically and isotopically exchanged with all of these earlier volumes of rock, it will have very little effect on the next cm^3 that it encounters. The overall isotopic shift is produced by adding up the effects of all these individual fluid packets in cumulative fashion as they cross each cm^2 cross-section of rock, whereas in the overall material-balance calculation for the whole body, each cm^3 packet of fluid entering the intrusion is, of course, only counted once.

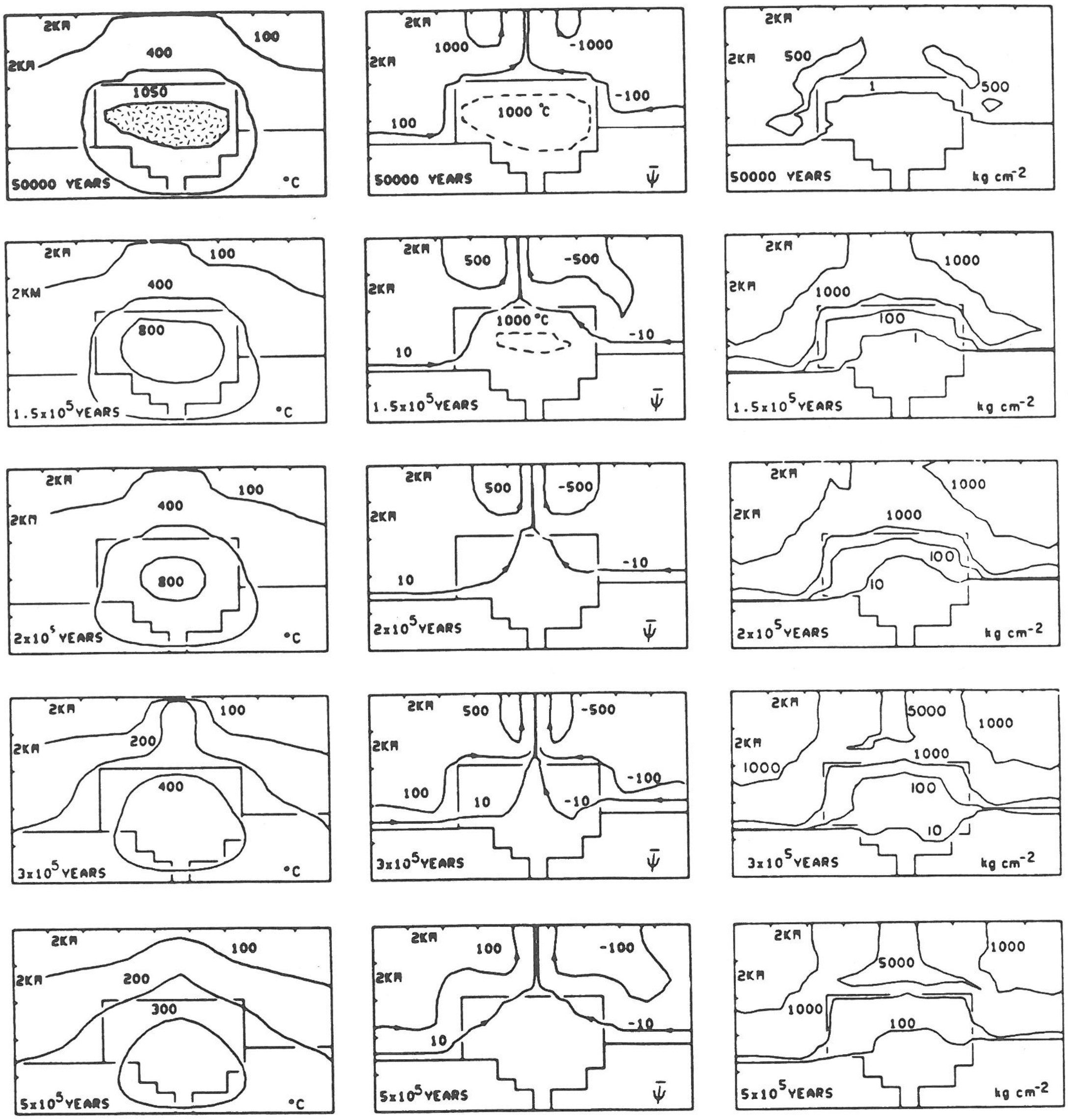

FIG. 5. Temperature (°C), streamlines (ψ), and integrated fluid flux (kg cm^{-2}) determined as a function of time by NORTON and TAYLOR (1979) in their numerical modeling study of the Skaergaard intrusion. In the streamline plots, note that at any instant, the largest amounts of H_2O are flowing through zones where the streamlines are closest together. Stippled pattern represents magma and dashed contour in streamline plot shows region where intrusion is assumed to be impermeable, *e.g.*, where $T > 1000°C$. After 400,000 years, this history of groundwater flow leads to the calculated $\delta^{18}O$ pattern shown in Figure 4b.

It is very important to understand that these seemingly gigantic "actual" water/rock ratios of 10^5 to 10^7 along a flow path are not only real, but also are perfectly compatible with the overall material-balance water/rock ratios of about 0.5 to 1.0 that are required both by the energy balance and by the mass balance of oxygen isotopes (or strontium isotopes, for example, if the initial and final concentrations and isotopic ratios of Sr are known for both the fluids and the rocks, see MCCULLOCH *et al.*, 1981). When one realizes that for a hand-specimen sized volume of rock, the actual, cumulative, fluid/rock ratios in hydrothermal systems (measured over the lifetime of the system) are commonly as high as 1,000,000 to 1, even in such low-permeability systems, it may be easier to understand the thorough "soaking" experienced by these rock systems, as well as the pervasiveness and uniformity of the ^{18}O-depletions that are observed in the feldspars of such hydrothermally altered rocks.

Other layered gabbro bodies

Other layered gabbro plutons that have established meteoric–hydrothermal convective systems are Jabal at Tirf in Saudi Arabia, Stony Mountain in Colorado, the Islands of Skye and Mull in Scotland, and Ardnamurchan in Scotland (TAYLOR and FORESTER, 1971; 1979; TAYLOR, 1980, 1983; FORESTER and TAYLOR, 1976, 1977, 1980). Analogous examples involving ocean water hydrothermal systems are observed in all ophiolite complexes, including Cyprus and the Samail ophiolite, Oman (GREGORY and TAYLOR, 1981), as well as in dredged samples from the Indian Ocean (STAKES *et al.*, 1983). In all these areas, large portions of the layered gabbro complexes display marked $^{18}O/^{16}O$ disequilibrium between coexisting pyroxene and plagioclase (see Figures 1 and 6).

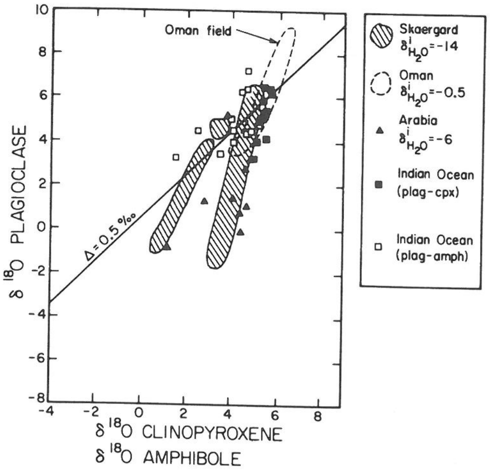

FIG. 6. Comparison of $\delta^{18}O$ data on clinopyroxene-plagioclase pairs from the Skaergaard intrusion (diagonal lined pattern) with analogous data from Indian Ocean dredge hauls (STAKES *et al.*, 1983), the Oman ophiolite (GREGORY and TAYLOR, 1981), and the Jabal at Tirf complex, Arabia. The Cuillin gabbros of Skye (data not shown) display similar relationships indicative of alteration by meteoric–hydrothermal fluids (FORESTER and TAYLOR, 1977). The $\delta^{i}_{H_2O}$ values (including $\delta^{18}O = 0$ for the Indian Ocean) indicate the most probable initial $\delta^{18}O$ values of the surface waters involved in these hydrothermal systems. Seven out of the ten data points at Jabal at Tirf correspond very well with the data from Oman, the main Layered Series rocks at Skaergaard, and the Cuillin Gabbro Series. The other 3 samples from Jabal at Tirf contain unusually low-^{18}O pyroxenes and, like the Eucrite Series gabbros from Skye, and the eastern rift–zone basalts from Iceland, these probably formed from low-^{18}O magmas (see text).

Although in these gabbro bodies there is local development of chlorite, epidote, actinolite, talc, sphene, prehnite, and other low–temperature (greenschist facies) minerals, particularly in areas of diking, heavy fracturing and veining, or multiple intrusion, the bodies in general are astonishingly free of *any* petrographic or mineralogic features indicative of hydrothermal alteration. This statement, of course, does not apply where an older gabbro body has been invaded and hydrothermally metamorphosed by a younger intrusion (*e.g.*, the heavily altered Broadford gabbro body on Skye, FORESTER and TAYLOR, 1977), but it definitely applies to *all* gabbro bodies involved in a single cycle of meteoric hydrothermal activities (*i.e.*, the convection cells set up by that particular gabbro magma chamber itself).

Petrographic evidence for hydrothermal alteration can usually be observed through a careful study of such ^{18}O-depleted gabbros, although the $^{18}O/^{16}O$ "reversals" between plagioclase and pyroxene always represent the most definitive relationship. These petrographic features are typically fairly subtle and include: (1) clouding or turbidity of some of the plagioclase; (2) development of minor talc–magnetite rims on olivine; (3) coarsening of exsolution lamellae around microfractures in clinopyroxene grains; (4) macroscopic veins easily seen on good, glaciated outcrops (on careful examination these veins often prove to contain pyroxene, magnetite, and high–temperature amphiboles such as hornblende); and (5) development of minor amounts of actinolite, biotite, chlorite, and epidote in zones transitional to lower temperature hydrothermal activity (*e.g.*, late–stage veins). All of the above features are described in the references given above, or in NORTON *et al.* (1984), or both. In addition, FERRY (1985b) has demonstrated that secondary magnetite (relatively pure Fe_3O_4) is ubiquitous in the Skye gabbros, and also that the calculated temperatures of formation of the talc–olivine–orthopyroxene assemblages are in the range of 525°–545°C. The early veins throughout practically the entire section of layered gabbro in the Skaergaard intrusion contain hydrothermal clinopyroxenes with *minimum* solvus temperatures of 500° to 750°C (MANNING and BIRD, 1986).

It is thus an inescapable conclusion that layered gabbro bodies typically undergo meteoric–hydrothermal alteration at very high temperatures, in large part in the range 500°–900°C. This conclusion is totally at odds with the statements made by CATHLES (1983) on this matter (see above). Only at such high temperatures could we simultaneously obtain the clear $^{18}O/^{16}O$ evidence for intense hydrothermal alteration and the virtually complete absence of low-temperature hydrous minerals. The temperatures (500°–900°C) and P_{H_2O} values (200–800 bars) are clearly not within the stability fields of chlorite, epidote, serpentine, actinolite, and clay minerals during the bulk of the hydrothermal activity. In fact, much of the mineralogical alteration that *does* occur in these gabbros takes place at 450–550°C, considerably higher than the average temperature of alteration in the granitic plutons described below, but still *lower* than the temperatures at which the bulk of the ^{18}O depletion occurred in these gabbros (TAYLOR and FORESTER, 1979; FERRY, 1985b).

Why then do we mainly see only the effects of 350°–400°C H_2O, for example, at the mid-ocean ridge spreading centers and on land in places like the Salton Sea? As NORTON and KNIGHT (1977) and NORTON (1984) have emphasized, convective systems are strongly controlled by the approximate coincidence of a viscosity minimum, together with a maximum in the isobaric coefficient of thermal expansion in critical to supercritical H_2O ($350° < T < 450°C$; $200 < P < 800$ bars). These are also the conditions where the density of H_2O undergoes the most rapid change as a function of a temperature. Furthermore, the heat capacity of the fluid under these pressure and temperature conditions is quite large, and is maximized at the critical point. Thus, in this general pressure-temperature range, the buoyancy and heat transport properties of the fluid are maximized and the drag forces minimized (NORTON and KNIGHT, 1977). In fact, it is very likely these physical properties of H_2O that control the observed upper limit of 350°–400°C observed during surface venting of modern hydrothermal systems, *not* the "fact" that all rocks hotter than 400°C are impermeable, as proposed by CATHLES (1983).

Granitic plutons

There are no known examples of *mineralogically unaltered* granitic rocks that also exhibit marked disequilibrium $\delta^{18}O$ values in the coexisting quartz and feldspar. The only low-^{18}O granites that are free of such mineralogical alteration products are those that crystallized from low-^{18}O magmas, and these rocks formed as equilibrium assemblages (*e.g.*, the Seychelles granites, see TAYLOR, 1977; also see below). The most conspicuous and quantifiable petrographic change observed in the hydrothermally altered granitic plutons is chloritization of mafic minerals, particularly the biotite, and a major increase in the turbidity of the feldspars. These changes are easily seen with the petrographic microscope, which additionally shows that the conversion to chlorite first proceeds along cleavage planes and grain boundaries of the biotite (CRISS and TAYLOR, 1986). Other common secondary minerals are montmorillonite, sericite, calcite, epidote, prehnite, and zeolites. In most extremely ^{18}O-depleted rocks the alteration is so intense that it can be seen in hand specimen. In such cases the K-feldspar is also strongly clouded, and albite twinning in plagioclase may be absent.

A good correlation exists between the degree of feldspar turbidity and $\delta^{18}O$ in the hydrothermally altered Skye granites (FORESTER and TAYLOR, 1977; FERRY, 1985a). The temperature of much of the alteration in these Skye granites was about 350°–450°C, based on various cation exchange geothermometers and phase equilibria (FERRY, 1985a); locally temperatures as low as 200°C are recorded.

Somewhat higher temperature alteration can occur immediately adjacent to and within the central intrusions of meteoric-hydrothermal systems in granites (CRISS and TAYLOR, 1983). Coarse-grained epidote and secondary amphiboles can occur in this zone, and in many cases intruded country rocks develop hornfelsic textures. Changes in the structural state of feldspar commonly occur, particularly inversion of microcline to orthoclase.

Contrasting effects in gabbros and granites

As indicated above, there are very distinctive mineralogical features observed in ^{18}O-depleted layered gabbros as opposed to those typically observed in ^{18}O-depleted granitic plutons. To recapitulate some of these differences: (1) Gabbros with reversed, nonequilibrium $\Delta^{18}O$ plagioclase–pyroxene values but practically no mineralogic alteration are quite common, whereas (2) Granodiorite, quartz monzonite, and tonalite plutons with large, nonequilibrium $\Delta^{18}O$ quartz–feldspar values are almost invariably strongly altered, containing abundant chlorite, epidote, sericite, turbid feldspars etc. A significant question is, why does most of the *external* (hydrostatic) meteoric–hydrothermal fluid move through layered gabbro bodies at much higher temperatures than in the case of granitic plutons?

There are several aspects to this problem. Seven of the major differences between gabbros and granites that probably contribute to the observed isotopic contrasts are listed in Table 1. It is interesting that each of these seven different igneous rock properties seem to favor a higher temperature of meteoric–

Table 1. Contrasting Properties of Granitic and Gabbroic Plutons

	Granitic plutons	Gabbro plutons
Latent Heat of Fusion	low (40 cal/g)	high (100 cal/g)
Magma Temperature	650°–900°C	1000°–1200°C
Initial H_2O Content of Magma	2–5 weight percent (2nd boiling common)	<1 weight percent (2nd boiling localized in late-stage granophyres)
Initial % Crystallized	20–60% (reduces latent heat)	<10% (little effect on latent heat)
Fracture Network	may be very dense within the pluton because of hydraulic fracturing (e.g., porphyry Cu bodies) and caldera collapse; also very large volume change associated with α-β quartz transition	usually simple contraction cooling and jointing, although occurrence in rift–zone environments is common, suggesting a very deep–seated and pervasive extensional fracture network in the country rocks.
Presence of Magmatic H_2O Envelope at Lithostatic Pressure	very common for H_2O–rich magmas at shallow depths in the crust	non–existent, except perhaps in H_2O–rich alkalic gabbro magmas rich in biotite and hornblende
Geometry of Crystallization	homogeneous solidification of the entire body (only local separation of late–stage melt from crystals); forms a single, integrated meteoric–hydrothermal system, or overlapping systems when complicated by multiple intrusions; these systems cannot migrate into the pluton until the magmatic H_2O envelope is dissipated	strong separation of cumulate crystals from silicate melt, with crystallization upward from the floor of the magma chamber; typically forms two decoupled meteoric–hydrothermal systems, a lower–T system in the country rocks and upper border zone rocks above the late-crystallizing sheet of magma near the roof of the chamber, and a higher–T system in layered cumulates below this magma sheet

hydrothermal alteration in the gabbro as opposed to the granites.

The most obvious reason why the meteoric–hydrothermal systems of gabbros can exist at much higher temperatures than in granites is that gabbros solidify at much higher temperatures, 1000°–1050°C. Granitic materials are still liquid at these temperatures and are thus unable to sustain fractures that would allow penetration by hydrothermal fluids. In addition, the latent heats of crystallization of gabbros are higher and the initial percentage of crystals in the gabbroic magmas is typically lower at the time of intrusion. Both features indicate that there is much more energy available in a gabbro for raising external H_2O to a high temperature than there is in a granite.

Possibly of equal importance to the above is the geometry of crystallization of the magma body. The typical granitic magma probably is intruded with a higher percentage of crystals, and being relatively viscous, both these crystals and newly–formed crystals are probably distributed throughout the mass, with final crystallization taking place in the center and at the lowest levels of the body. This is decidedly not how layered gabbros crystallize. As shown schematically in Figure 5 for the Skaergaard intrusion, these igneous bodies characteristically solidify upward from the bottom, and the last liquid to crystallize is a sheet–like layer of liquid near the roof of the body. This sheet of late liquid is crystallizing very slowly, because crystallization proceeds as the square root of time and the body is at that stage surrounded by a mass of very hot, insulating rock. This means that the crystalline gabbro *underneath* the sheet of late-stage magmatic liquid will remain *very* hot for a long time. When such material fractures, it will be penetrated by very hot aqueous fluids which flow inward underneath the magma sheet, which is itself impermeable to the fracture-controlled hydrothermal system. As indicated schematically for the Oman ophiolite magma chamber (Figure 7), this characteristically seems to produce two decoupled hydrothermal systems in layered gabbros (see NORTON and TAYLOR, 1979; NORTON *et al.*, 1984; and GREGORY and TAYLOR, 1981): (a) A relatively low–temperature, vigorous hydrothermal system above the magma sheet, where water/rock ratios are high and T = 350°–400°C, and (b) A much higher temperature system below the impermeable magma sheet where T = 500°–900°C and water/rock ratios are much lower (0.1–0.3). Only after the final liquid sheet crystallizes

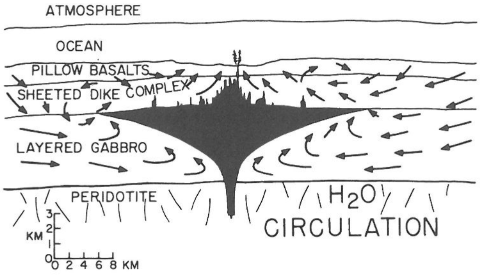

FIG. 7. Cartoon sketch, modified after GREGORY and TAYLOR (1981), showing probable seawater circulation patterns in a cross–section through a fast–spreading mid–ocean ridge. The solid black zone indicates the characteristic, funnel–shaped cross–section of a basaltic magma chamber (*e.g.*, Muskox intrusion, Great Dyke). Note that in the fractured rocks outside the magma chamber there are two essentially isolated circulation regimes, separated by a thin sheet of magma sandwiched between the layered gabbros and the overlying dike complex. The magma is essentially impermeable to the fluids in the two hydrostatic, sea–water convective systems.

can the temperature of the layered gabbro begin to fall rapidly, and at this stage, the two decoupled systems become connected and the temperature of the intrusion sharply declines as strong upward fluid flow is now possible out through the top of the intrusion.

Another major difference is the contrasting effects of magmatic H_2O in the two cases. Granitic magmas typically contain much higher concentrations of H_2O than tholeiitic gabbro magmas. This magmatic water in fact is thought to provide the force that causes abundant fracturing in prophyry copper deposits (BURNHAM, 1979). Any strong fracturing event will increase the permeability enormously, allowing correspondingly larger amounts of fluid to enter the system on a shorter time scale; the pluton would thus be cooled down to 200°–300°C fairly rapidly. These types of effects will not occur in the gabbroic systems. Another contributing aspect is the major volume change that accompanies the α-β quartz transition in granites that is absent in gabbro.

The final feature that may be important is the fact that any magma body that releases significant H_2O under lithostatic pressure can produce a magmatic H_2O envelope that will keep the external, meteoric–hydrothermal system outside the intrusion until this magmatic H_2O envelope is dissipated. This effect seems to occur in porphyry copper systems (SHEPPARD *et al.*, 1971; TAYLOR, 1974). Thus, no low-^{18}O effects would be seen in the pluton until the very late, low–temperature stages. Because of their low magmatic H_2O contents, these effects would not typically be observed in tholeiitic gabbros. However, they might possibly be seen in volatile–rich alkali gabbros, and this may be the reason for the contrasting isotopic behavior of the Lilloise alkalic hornblende gabbro intrusion as compared to the tholeiitic Skaergaard intrusion (Figure 3). Although both intrusions were emplaced into similar country rocks in East Greenland about 55 Ma ago, strong ^{18}O depletions are not observed in the Lilloise body (SHEPPARD *et al.*, 1977).

To conclude this section, we summarize the factors that tend to keep an external (hydrostatic) marine- or meteoric–hydrothermal system from penetrating and causing $\delta^{18}O$ changes in either a gabbroic or a granitic pluton:

(1) Intrusion into relatively impermeable country rocks, for example into either (a) limestones and evaporites, which are susceptible to ductile deformation at pressure-temperature conditions where most other rocks undergo brittle fracture, or into (b) ordinary silicate rocks at sufficient depth in the crust for recrystallization (metamorphism) to occur, particularly in the absence of extensional tectonics (rifting) or strong, brittle deformation.

(2) Presence of the silicate melt itself, which also is essentially impermeable as long as the strain rates are low enough and the percent of melt high enough so that an interconnected fracture network cannot develop. This is very important in layered gabbros because the late–stage magma sheet at the roof of the intrusion provides a barrier between the very hot, lower hydrothermal system in the layered cumulates and a cooler system at 250°–400°C and higher water/rock ratios in the roof rocks. The hydrothermal fluids in both systems are externally derived and are under hydrostatic pressures.

(3) Evolution of magmatic fluids from late–crystallizing portions of a (granitic) pluton, producing a magmatic H_2O envelope under lithostatic pressure that fills all available fractures outward from the crystallization front. This keeps the low–^{18}O meteoric waters outside the pluton until the temperatures have fallen into the range of stability of sericite, chlorite, etc. Thus, two decoupled hydrothermal systems may also commonly occur around H_2O–rich granitic magma chambers, but in such cases the two types of fluids are genetically very different and are under different pressures. Also, the magmatic H_2O system does not produce the kinds of drastic $^{18}O/^{16}O$ changes in the mineral assemblages that are produced by the meteoric–hydrothermal systems.

ORIGIN OF LOW-^{18}O MAGMAS

General statement

The preceding discussion shows that it is relatively easy to demonstrate sub-solidus, meteoric-hydrothermal, ^{18}O/^{16}O exchange in ancient, deeply-eroded igneous centers, and there is definitive evidence that aqueous fluids penetrate in abundance into gabbro intrusions at very high temperatures, well into the temperature range at which granitic melts can be formed. Because of the pervasive hydrothermal effects, it is very difficult to prove the existence of low-^{18}O *magmas* in such environments. Considerable effort is required to "see through" the alteration effects and find out how much of the ^{18}O depletion was an original magmatic phenomenon. On the other hand, if one can analyze fresh, unaltered lava flows that have not been buried and therefore have not suffered any sub-solidus hydrothermal alteration, this is not a problem.

MUEHLENBACHS *et al.* (1974) and FRIEDMAN *et al.* (1974), respectively, made the important discoveries that such low-^{18}O volcanic magmas had in fact been formed in large volumes in two of the major late Cenozoic volcanic fields of the world, namely in Iceland and in the Yellowstone Plateau, Wyoming. This phenomenon is one of the truly unique aspects of the geochemistry of igneous rocks that was developed through the utilization of stable isotope techniques. The existence of such magmas, and the necessity for a special mechanism to explain them, was not even suspected before ^{18}O/^{16}O analyses of silicates became commonplace.

In the past decade, a fairly intensive search has been made for new occurrences of low-^{18}O magmas. Although a few isolated occurrences have been found, two conclusions can be made from this decade-long search: (1) Such magmas are much less abundant than was originally thought; and (2) no new giant occurrences comparable to Yellowstone or Iceland have been found among Late Cenozoic lavas *anywhere* in the world. The following discussion addresses the reasons for the rarity of these types of magmas, which is clearly also related to the problem of their origin.

Iceland

The data of MUEHLENBACHS *et al.* (1974) and CONDOMINES *et al.* (1983) can be briefly summarized as follows (Figure 8): (1) The alkali olivine basalts on Iceland have distinctly higher and more uniform δ^{18}O values than any of the other Icelandic basalts, +5.3 to +5.7; these values are similar to those found in most basalts throughout the world. These alkali olivine basalts are found only on the periphery of the island, at the edges of any deep meteoric-hydrothermal circulation systems. (2) Relative to the alkali olivine basalts and the tholeiites, the transitional alkali basalts occupy an intermediate geographic position in the eastern rift zone (Figure 8) and also have intermediate δ^{18}O values, +3.9 to +4.9. (3) The olivine tholeiites from the western rift zone (Figure 8) also have intermediate δ^{18}O values, +4.0 to +5.7, with the highest values closest to the coast on the SW part of Reykjanes peninsula. (4) In the basalts, the lowest and most heterogeneous δ^{18}O values, +1.8 and +5.4, are all confined to the quartz tholeiites of the eastern rift zone. (5) There are some crude correlations between δ^{18}O and chemical composition; increasing SiO_2 and K_2O tend to be accompanied by decreasing δ^{18}O for each petrologic class of basalts. In particular, the tholeiite with by far the most extreme δ^{18}O value (+1.8) also has by far the highest K_2O content. (6) The rhyolites and obsidians tend to have even lower δ^{18}O values than the basalts (Figure 8), and the rhyolites with the lowest and most variable δ^{18}O values all come from the eastern rift zone. (7) There is a weak correlation between δ^{18}O and ^{3}He/^{4}He, with the intermediate and silicic volcanic glasses typically having ^{3}He/^{4}He ratios close to the atmospheric value.

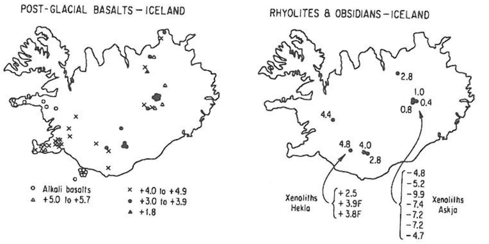

FIG. 8. Maps of Iceland, on the left showing the range of δ^{18}O values determined for post-glacial basalts, and on the right showing the range of δ^{18}O values of rhyolites, obsidians, and siliceous xenoliths (data from MUEHLENBACHS *et al.*, 1974, and MUEHLENBACHS, 1973).

The isotopic relationships described above strongly support the idea that assimilation or partial melting, or both, of hydrothermally altered country rocks in the deeper parts of the rift zone is the most likely mode of formation of the low-^{18}O magmas from Iceland. This explanation is compatible with the Pb isotope data of WELKE *et al.* (1968) and the ^{87}Sr/^{86}Sr and rare-earth data of O'NIONS and GRONVOLD (1973), and it readily explains why the most contaminated magmas are either the rhyolites, or those basalts that have probably most strongly interacted with roof rocks or rhyolite melts, namely, the K–rich and Fe-rich tholeiites. Direct exchange with meteoric waters would not be expected to produce these relationships.

It is probably significant that all of the extremely low–^{18}O basaltic and rhyolitic magmas are confined to the eastern rift zone, which has only been active during the past 3–4 m.y. (Saemundsson, 1974). The magmas in this rift zone are penetrating upward through the lower parts of volcanic and plutonic rocks that presumably were intensively hydrothermally altered in an earlier episode of magmatic activity at the time of their original formation in the western rift zone. Thus, the magmas coming up through the eastern rift zone would be interacting with country rocks that had already suffered heterogeneous ^{18}O depletions, and which, through subsidence, have been brought down into a much higher temperature regime (15–20 km depth?).

Approximately 200 km^3 of low–^{18}O tholeiite has been erupted in the eastern rift zone in the last 12,000 years (Jacobsson, 1972). It was the scale of this process that most bothered Muehlenbachs *et al.* (1974) when they rejected the meteoric–hydrothermal explanation for the origin of these types of magmas. This mechanism for the origin of the Icelandic volcanic rocks has, however, been strongly favored over the past few years by Taylor (1974; 1977; 1979), and recently Hattori and Muehlenbachs (1982) and Condomines *et al.* (1983) provided strong new support for this mechanism.

The model of Condomines *et al.* (1983), based on combined He, O, Sr, and Nd isotopic relationships, is shown in Figure 9. They propose that the primary $^3He/^4He$ and $^{18}O/^{16}O$ ratios of mantle–derived magmas were changed in a deep magma reservoir by exchange or contamination between the magma and the surrounding meteoric–hydrothermally altered basaltic crust. Such processes to not appear to have introduced much water into the magma because Icelandic volcanism, except for subglacial eruptions, is usually not explosive, and hydrous minerals in plutonic ejecta are rare. The rhyolitic magmas are assumed to have been produced at relatively shallow crustal levels, either by melting of hydrothermally altered rocks or as the deeper magmas moved upward and underwent further contamination processes. These processes caused introduction of atmospherically–derived He into the magmas.

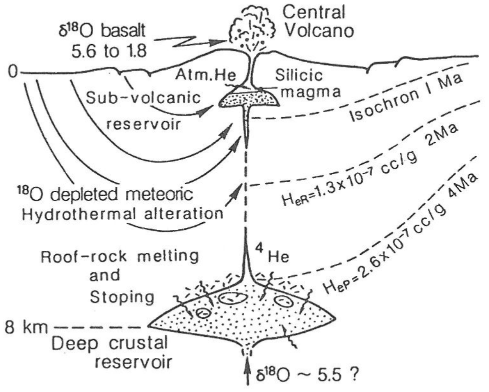

FIG. 9. Schematic section summarizing the model of Condomines *et al.* (1983) for the evolution of the Icelandic magmas. (He_R = radiogenic helium).

Catastrophic isotopic changes in magmas during caldera collapse, Yellowstone volcanic field

The title of this section and much of the following discussion is taken from a very significant paper by Hildreth *et al.* (1984), who followed up the original discovery by Friedman *et al.* (1974) with a detailed chemical and isotopic study of the Quaternary rhyolites of the Yellowstone Plateau, Wyoming.

This 17,000–km^2 volcanic field (Figure 10) consists of three large, overlapping, rhyolitic calderas, apparently a 115 km long extension of the axis of the Snake River Plain. Rhyolitic magmatism has migrated northeastward in this area at a rate of 2–4 cm/year for at least the last 10 Ma (Armstrong *et al.*, 1975; Christiansen and McKee, 1978). Former rhyolitic centers have progressively subsided and been buried by the basalts that now make up the floor of the plain.

Brief, caldera-forming, ash–flow eruptions occurred at 2.0, 1.3, and 0.6 Ma (Christiansen, 1983), with minimum volumes of 2500, 280, and 1000 km^3, respectively. The $\delta^{18}O$ values of the first-cycle Huckleberry Ridge Tuff (HRT) are among the highest in the Yellowstone area ($\delta^{18}O$ quartz = +7.1 to +7.6, giving a calculated coexisting magma $\delta^{18}O \approx +6.4$ to +6.9). The subsequent caldera-forming eruptions (Figure 11) have slightly lower calculated $\delta^{18}O$ magma values of +4.8 to +5.2 (Mesa Falls Tuff) and +5.3 to +6.0 (Lava Creek Tuff).

Almost all of the Yellowstone rhyolites are somewhat depleted in ^{18}O relative to the 'normal' $\delta^{18}O$ values of +7 to +10 usually observed in silicic volcanic rocks throughout the world (Taylor, 1968, 1974). However, the postcaldera rhyolites of the first and third caldera cycles include some *extraordinarily low–*^{18}O eruptive units (Figure 11).

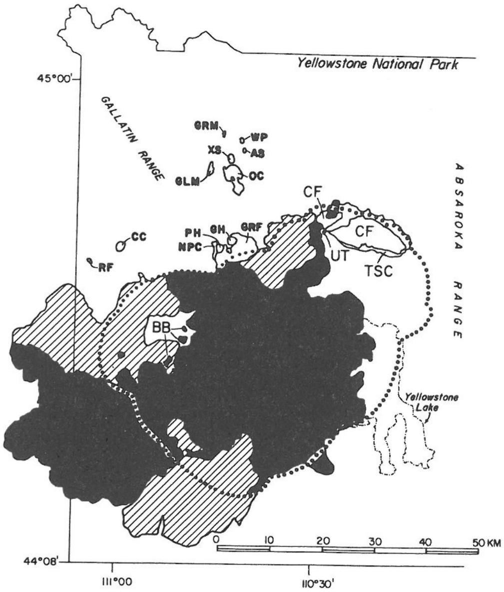

FIG. 10. Generalized map of Yellowstone National Park, showing the outline of the caldera rim (outer fault scarp) of Yellowstone caldera (dotted), and the locations of third-cycle post-caldera and extracaldera rhyolites. The extreme low-^{18}O post-caldera rhyolites (δ^{18}O quartz $< +2.0$) are indicated by symbol: CF, TSC, and UT in the northeast part of the caldera, and BB in the southwest. The moderately low-^{18}O post-caldera rhyolites ($+4.0 < \delta^{18}$O quartz $< +5.0$) are shown in solid black, and the slightly ^{18}O-depleted post-caldera rhyolites (δ^{18}O quartz $\geqslant +5.0$) are indicated by the diagonal-lined pattern. The normal-^{18}O extracaldera rhyolites north of the caldera are the small outcroppings indicated by various symbols. The first-cycle caldera lies about 15 km SW of the western park boundary. Abbreviations: GRM—Gardner River mixed lavas, WP—Willow Park dome, AS—Apollinaris Spring dome, XS—Crystal Spring flow, OC—Obsidian Cliff flow, GLM—Grizzly Lake mixed lavas, CC—Cougar Creek flow, RF—Riverside flow, PH—Paintpot Hill dome, GH—Gibbon Hill dome, GRF—Gibbon River flow, NPC—Nez Perce Creek flow. Modified after HILDRETH *et al.* (1984).

The combined areal extent of the two first-cycle low-^{18}O flows (δ^{18}O = +2.9 to +3.6) is $\approx$66 km^2; their original volume was at least 10 km^3, and they were erupted 30,000 to 350,000 years after the >2500 km^3 Huckleberry Ridge Tuff. These low-^{18}O lavas are similar to most other Yellowstone rhyolites, and contain $\approx$76% SiO$_2$ and 15–20% phenocrysts of quartz, sanidine, plagioclase, clinopyroxene, fayalite, and Fe–Ti oxides.

The second cycle was unique, in that there was no major change in the δ^{18}O of the magma body after eruption of the Mesa Falls Tuff (MFT) and collapse of its 380-km^2 caldera. However, during the third cycle, enormous ^{18}O depletions were observed (Figure 11) in lavas that vented in two separate areas $\approx$45 km apart, soon after eruption of the >1000 km^3 Lava Creek Tuff (LCT), along the compound ring fracture zone of the Yellowstone caldera. Some of these units are petrographically distinct in containing plagioclase in excess of sanidine. In the NE part of the caldera, there are three major low-^{18}O units, which together cover $\approx$140 km^2 and represent >40 km^3 of magma, perhaps as much as 70 km^3, all of it with δ^{18}O between about +0.6 and +1.2 (calculated from δ^{18}O of the quartz). The second area of very low ^{18}O third-cycle rhyolites is in the southwestern ring fracture zone, where scattered exposures in a 25-km^2 area have calculated δ^{18}O values between −0.1 and +0.6, and a *minimum* volume of 2.5 km^3. Chemically, the low-^{18}O third-cycle magma completely overlaps the compositional range of the Lava Creek Tuff, but it also extends to less evolved compositions (HILDRETH *et al.*, 1984).

Pb and Sr isotope data by DOE *et al.* (1982) and HILDRETH *et al.* (1984) are compared in Figure 12, scaled to reveal the correlations between δ^{18}O, ^{206}Pb/^{204}Pb, and ^{87}Sr/^{86}Sr, specifically between lower δ^{18}O and more radiogenic Pb and Sr. The Pb and Sr isotope data define a zigzag pattern that reflects abrupt caldera-forming events and longer intervals of partial recovery toward precaldera isotopic ratios very much like the δ^{18}O patterns. The low-^{18}O early postcaldera rhyolites of the third cycle are particularly enriched in radiogenic Pb and Sr. The rhyolitic units with the most radiogenic ^{206}Pb/^{204}Pb ratios also have the highest ratios of ^{207}Pb/^{204}Pb, suggestive of collapse-related mixing of the rhyolitic magmas with pre-Cenozoic upper crustal components, most

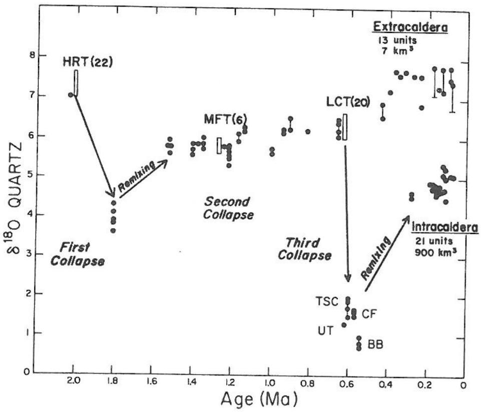

FIG. 11. Plot of δ^{18}O quartz vs. K–Ar age for the rhyolites of the Yellowstone Plateau volcanic field. HRT, MFT, and LCT are (with numbers of analyzed samples) the major ash-flow sheets referred to in the text. Modified after HILDRETH *et al.* (1984).

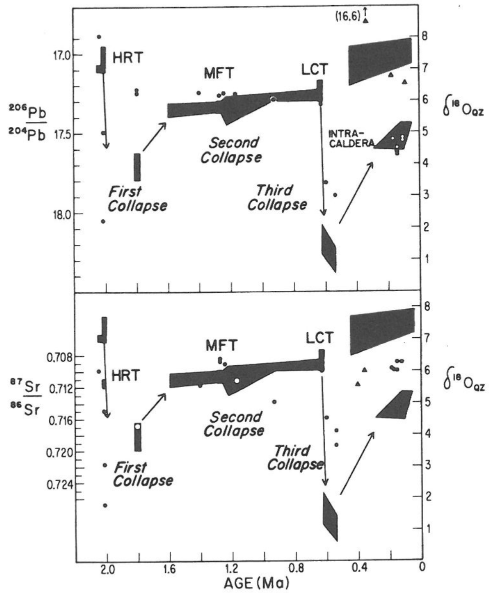

FIG. 12. Initial Sr and Pb isotopic values for rhyolites of the Yellowstone volcanic field (dots, left axis), plotted against ages of the units and superimposed on shaded fields (right axis) representing the $\delta^{18}O$ quartz data of Figure 11. Modified after HILDRETH *et al.* (1984).

of which probably contain more radiogenic Pb and Sr than the precaldera or extracaldera rhyolites (DOE *et al.*, 1982).

Origin of the low-^{18}O magmas at Yellowstone

Following is a summary of the critical features pertinent to the development of low–^{18}O magmas in the Yellowstone caldera complex, modified after HILDRETH *et al.* (1984):

(1) The narrow range of $\delta^{18}O$ in the three major ash–flow sheets (+5 to +7) contrasts sharply with the wide variation in the postcaldera rhyolitic lavas. The earliest postcaldera lavas are depleted in ^{18}O by as much as 3 to 6 per mil relative to the immediately preceding ash flow tuffs. The biggest ^{18}O effects follow immediately after the two biggest ash–flow eruptions.

(2) The ^{18}O depletions were geologically short-lived events (<300,000–500,000 years) that followed caldera subsidence in some cases by less than 50,000–100,000 years. The earliest post–collapse lavas are the most ^{18}O–depleted units, and successively younger postcaldera lavas show partial recovery of the magma toward precaldera $\delta^{18}O$ values, presumably by exchange and mixing with deeper levels of the magma reservoir. However, no intracaldera rhyolite is ever again as ^{18}O–rich as the ash–flow tuff that preceded it.

(3) The pattern of stepwise ^{18}O depletions is reflected in lavas erupted as far apart as 115 km, indicating that these are not isolated events. Particularly for the third cycle, the ^{18}O–depleted rhyolites cover most of the caldera complex (Figure 10), and they are part of an integrated magmatic system that evolved for more than 2.0 Ma.

(4) Depletion of ^{18}O occurred *only* in the subcaldera reservoir; contemporaneous rhyolites that vented just outside the caldera have the highest $\delta^{18}O$ values in the volcanic field (Figure 11).

(5) Voluminous postcaldera rhyolites, all exhibiting the drastic 5 per mil ^{18}O depletion, erupted at opposite ends (45 km apart) of the newly–formed third–cycle caldera, showing that much more than 100 km^3 of magma was affected. More than 1000 km^3 of magma was depleted by 1–2 per mil (Figure 10).

(6) Sr and Pb isotopic ratios of the rhyolites jump to more radiogenic values immediately subsequent to caldera formation. The long-term pattern for Pb and Sr isotopic ratios of rhyolites erupted from the subcaldera magma reservoir is similar to the zigzag pattern displayed by $\delta^{18}O$. Depletion in ^{18}O is clearly related to caldera collapse, and this structural disruption was also important for introducing radiogenic Pb and Sr into the magma.

To explain all the above effects, HILDRETH *et al.* (1984) strongly favored a mechanism of ^{18}O depletion involving influx and solution of low–^{18}O meteoric *water* in enormous amounts directly into the rhyolitic magma. They realized that unless such waters included deep brines exceptionally rich in Pb and Sr (and concentrated brines are *not* low in ^{18}O because they invariably have undergone large ^{18}O shifts), the isotopic shifts of Pb and Sr would require additional exchange with large quantities of foundering roof rocks. Nevertheless, HILDRETH *et al.* (1984) believe that water itself is by far the predominant contaminant.

Although congratulating HILDRETH *et al.* (1984) for a truly magnificent study, this writer is in serious disagreement with their ultimate conclusion concerning the mechanism of ^{18}O depletion of the magmas. A much more likely mechanism(s) would seem to be (1) partial melting of hydrothermally altered rhyolitic country rocks, with subsequent uprise and mixing of this re-melted material back into the roof zone of the magma chamber and/or (2) foundering of such ^{18}O–depleted roof rocks into the magma chamber, accompanied by melting, assimilation, or exchange of this foreign material with the shallow magma reservoir. Based on constraints imposed by the physics and chemistry of H_2O transport through ductile rocks and silicate melts, which are generally thought to be virtually impermeable to convecting aqueous fluids (see above), these mechanisms seem much more plausible than catastrophic influx of water directly into the magma chamber. These concerns are elaborated on in several papers by TAYLOR (1974; 1977; 1979; 1983), and in fact, this writer believes that the data of HILDRETH *et al.* (1984) actually provide strong new support for those conclusions.

The most abundant country rocks above and along the margins of the Yellowstone magma chamber prior to each major ash–flow eruption were certainly hydrothermally altered, ^{18}O–depleted earlier–cycle rhyolites. In the Yellowstone Plateau, pre-Cenozoic rocks are very subordinate to the volcanic rocks. Such young volcanic rocks cannot represent the contaminant responsible for the Pb and Sr isotope signatures that abruptly appear in the Yellowstone rhyolites; therefore, it is an absolute certainty that *if* foundering of roof rocks or melting of older country rocks is necessary to explain the Sr and Pb isotope effects, then this *must* have been accompanied by prodigious amounts of assimilation or melting of the much more abundant rhyolitic country rocks. The *only* observable chemical effects of this latter process on the magma reservoir would be: (1) lower $\delta^{18}O$ values, (2) lower δD values, and possibly (3) higher H_2O concentrations. This is because these rhyolitic country rocks are chemically and isotopically almost *identical* to their original parent magmas, except for the subsolidus hydrothermal changes they have undergone, which to a first approximation only involve hydration and depletion in ^{18}O and D. Note also that such hydrated rhyolites will be much more easily melted than any other type of country rock.

The other important feature of the HILDRETH *et al.* (1984) study is the catastrophic and abrupt interval over which the isotopic changes occur. The effects are clearly related to caldera collapse. There is no better way of changing the composition of a magma on a short time–scale than catastrophic failure and massive collapse of the roof material into the magma chamber, or rapid subsidence of such rocks into the deep melting zone along the edges of the magma chamber. Not only does the low–^{18}O, hydrothermally altered rhyolite undergo catastrophic subsidence, but new, hotter magma also almost immediately moves upward into the resurgent domes and ring intrusions. Similar, short–time scale, catastrophic processes had in fact been invoked previously by TAYLOR (1974; 1977), but up until the time of the HILDRETH *et al.* (1984) study there was little direct support for the idea that such extreme changes could take place on such a rapid time scale. Even though diffusive transport of enormous amounts of H_2O directly into the magma *a priori* seemed unlikely, there was always the possibility that given enough time it perhaps could occur. The data of HILDRETH *et al.* (1984) show clearly that $\delta^{18}O$ changes in the magma chamber *do not occur* by such a long-term process! In fact, the long-term changes are in the opposite direction, toward recovery of the original magmatic $\delta^{18}O$ value, *after* the intense ^{18}O depletion event that accompanied the foundering of the roof. On such a short geological time–scale, there is no physical way of separating pore waters from their rock matrix, particularly considering the slow rates of diffusion of H_2O through silicate melts or through ductilely deforming rocks. To this writer, it is obvious that water is indeed involved in the process, but this water is only the pore water that permeates the rock matrix and which is essentially carried along in any foundering or re–melting process that affects the rocks themselves.

HILDRETH *et al.* (1984) raised several objections against the partial melting or bulk assimilation process, among which are the fact that there is no evidence of the type of massive cooling or increase in phenocryst content that might be expected in the contaminated magmas, nor is there any obvious change in major or trace element composition. The first objection loses its validity when it is realized that mixing or exchange with enormous volumes of liquid H_2O would have an even more profound cooling effect on the magma than would melting or assimilation of hot, altered country rock. The specific heat of water is three to five times higher than that of the rock. Also, the temperatures in these types of systems are probably controlled by the large reservoir of more mafic, higher–temperature magma underneath the rhyolite sheet. The second argument is refuted by the fact that the rocks being assimilated or melted are predominantly older–cycle rhyolites with essentially identical chemical compositions to the younger magmas.

To this writer, the *only* observations by HILDRETH *et al.* (1984) that may be in disagreement with the roof-rock melting or assimilation mechanism are: (1) The material–balance argument that the effects are too large to be explained by incorporation of altered rocks, which are unlikely to have an average $\delta^{18}O$ any lower than about −5 to −8; and therefore that it is much easier to produce the observed effects with pure H_2O, which could have $\delta^{18}O = -19$ (today's water) or perhaps even −25

(Pleistocene H_2O?). (2) The apparent absence of xenocrysts in the lavas.

The material-balance problem is a difficult one, no matter what mechanism one chooses, but it is really not much more of a problem for one process than for the other, as shown in Figure 13. Even at shallow levels of less than 200 m depth, hydrothermally altered rocks with $\delta^{18}O = -8$ to -9 are not uncommon at Yellowstone (HILDRETH *et al.*, 1984). At depths of several km, these values could be expected to be as low as -12 or lower, judging by the data from the deep Krafla drill hole on Iceland (HATTORI and MUEHLENBACHS, 1982). Note that young volcanic rocks are also very porous, so the foundered roof rocks that fall into the magma chamber or which subside into a deep melting zone during such a catastrophic process are not just rock, but H_2O-saturated rock with perhaps 20–35% pore space. Therefore, the water-saturated rocks could easily have an average $\delta^{18}O = -13$ to -15, or perhaps even lower during the Pleistocene glacial period.

The coupled assimilation-fractional crystallization (AFC) process modeled in Figure 13 is based on R values (ratio of cumulates to assimilated rocks) of only 1.5 to 5, which might seem to be unusually low for such a shallow assimilation process (TAYLOR, 1980). However, these calculations are concerned with the *chemical effects* on the rhyolite magma layer, whereas the *heat effects* in these types of magma systems are dominated by the underlying, higher-temperature, more mafic magma reservoir. Also, the latent heats of fusion of the assimilated glassy rhyolites are probably very low (<30 cal/g).

The explosive shattering that would accompany the engulfment of such water-saturated rocks into the magma chamber would enhance complete assimilation. The lack of xenocrysts may be related to this shattering event or possibly to a super-heated condition in the rhyolite melt, which is being heated from below by the underlying, more mafic magma. In any case, virtually no xenocrysts would be expected to be derived from such very fine-grained to glassy rhyolites in the first place. It is thus conceivable that there were very few xenocrysts, and those that were present were either largely dissolved, or they were so finely shattered and dispersed that they are not recognizable. In any case, this single observation is in no way sufficient to tip the scales against the re-melting/assimilation mechanism. Actually, probably the biggest objection to this mechanism is the fact that low-^{18}O magmas are *not* observed in most rhyolitic caldera complexes (LARSON and TAYLOR, 1986). This problem is discussed in more detail below.

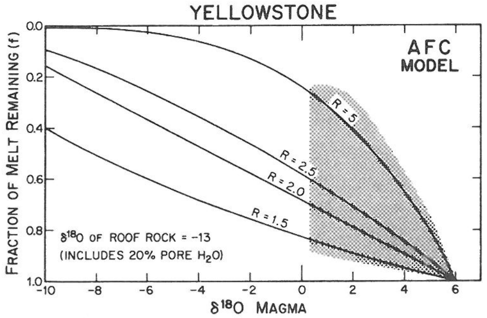

FIG. 13. The curves show AFC model calculations (TAYLOR, 1980) for the Yellowstone caldera system for different assumed values of R, the ratio of cumulates to assimilated rock. The stippled pattern shows the range of $\delta^{18}O$ of the post-caldera rhyolites.

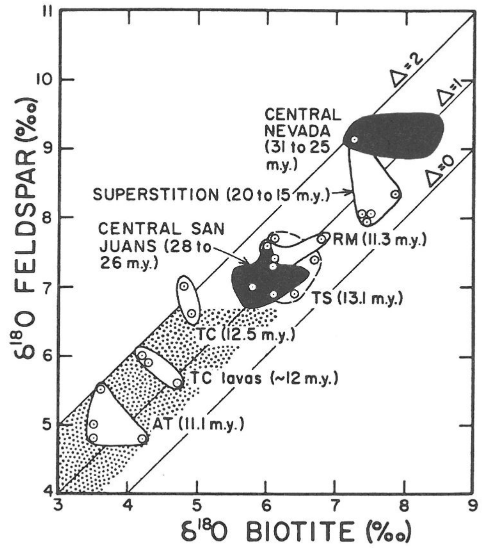

FIG. 14. Plot of $\delta^{18}O$ feldspar vs. $\delta^{18}O$ biotite, modified after LARSON and TAYLOR (1986), for various ash-flow tuff complexes in the western U.S.A. AT = Ammonia Tank, TC = Tiva Canyon, TS = Topopah Springs, and RM = Rainier Mesa; these are different ash—flow units of the southwestern Nevada caldera complex studied by LIPMAN and FRIEDMAN (1975). The stippled pattern indicates the field of Yellowstone rhyolites (HILDRETH *et al.*, 1984). The data on the Superstition volcanic field are from STUCKLESS and O'NEIL (1973).

Other low-^{18}O rhyolite magmas

The only other area in the world where low-^{18}O magmas have been *proven* to have erupted in large quantities is in the Oasis Valley-Timber Mountain caldera complex (Figure 14) in southwestern Nevada (FRIEDMAN *et al.*, 1974; LIPMAN and FRIED-

MAN, 1975). A search for low-^{18}O magmas was specifically made by LARSON and TAYLOR (1986) in two of the largest and most complex caldera complexes in the United States, in Central Nevada and in the Central San Juan Mountains of Colorado, as well as in the Lake City caldera in the Western San Juan Mountains. No evidence of any significant magmatic ^{18}O depletion was found in any of these areas (Figure 14). In fact, in each of these areas, LARSON and TAYLOR (1986) observed a remarkably uniform set of δ^{18}O magma values that persisted from 31 to 25 Ma in Central Nevada and from 28 to 26 Ma in the San Juans. The Central Nevada magmas had $\delta^{18}O = +9.5$ to $+10.0$, whereas the Central San Juan complex is much lower in ^{18}O but just as uniform ($\delta^{18}O = +7.0$ to $+7.5$). These are all primary magmatic values; the two complexes differ because the source-rock materials of these magmas at depth were more than 2 per mil different from one another.

Because of the rarity of low–^{18}O magmas, the one other caldera complex in addition to Yellowstone that does show these low–^{18}O magmatic effects is of considerable interst (LIPMAN and FRIEDMAN, 1975). The major δ^{18}O changes in the southwestern Nevada complex (Figure 14) are as follows. In the first cycle, the Topopah Spring eruption (>300 km^3, 13.1 Ma) had δ^{18}O feldspar $= +6.8$ to $+8.1$, and was compositionally zoned from rhyolite (77 weight percent SiO_2, 1% phenocrysts) to quartz latite (69 weight percent SiO_2, 21% phenocrysts). This eruption was followed by the chemically similar, but much bigger Tiva Canyon eruption (>1000 km^3, 12.5 Ma), which had distinctly lower δ^{18}O feldspar $= +6.4$ to $+7.0$. Late rhyolitic lavas associated with the margins of the Tiva Canyon caldera have even lower δ^{18}O feldspar $= +5.1$ to $+6.1$.

After a hiatus of about 1,000,000 years, a second eruptive cycle began starting with the Rainier Mesa eruption (>1200 km^3, 11.3 Ma), which is zoned from rhyolite (76–77 weight percent SiO_2, 20–25% phenocrysts) to quartz latite (66–68 weight percent SiO_2, 35–50% phenocrysts), and has δ^{18}O feldspar $= +6.6$ to $+7.7$ (*i.e.,* the magma system had essentially recovered from the Tiva Canyon ^{18}O depletion event). This eruption was, however, quickly followed only about 200,000 years later by a very ^{18}O–depleted magma, the giant Ammonia Tank eruption (>2000 km^3, 11.1 Ma), which has δ^{18}O feldspar $= +4.7$ to $+5.5$. The Ammonia Tank unit is also very phenocryst-rich, but is chemically and petrographically almost identical to the higher–^{18}O Rainier Mesa unit.

In each of the two cycles, a normal–^{18}O eruption was followed by a chemically similar, but ^{18}O–depleted eruption. The low–^{18}O magma in each case exhibits some *very slight* differences from the immediately preceding normal–^{18}O eruption, namely: (1) the low–^{18}O magmas lack plagioclase; (2) they contain sphene; (3) they contain more sodic alkali feldspar; and (4) they have very low ilmenite–magnetite ratios. These differences were interpreted by LIPMAN (1971) and LIPMAN and FRIEDMAN (1975) as indicating that the low–^{18}O units crystallized at higher P_{H_2O} and higher f_{O_2} than the immediately preceding units. These features, of course, would be a natural consequence of the low–^{18}O contamination process by either hydrous roof rocks or direct influx of meteoric H_2O. Evidence of an open system involving older, ^{87}Sr–rich rocks is seen in the Sr isotope measurements of NOBLE and HEDGE (1969). Thus, all of the arguments made above favoring a partial melting or AFC model for the ^{18}O depletions at Yellowstone are also valid for these southwestern Nevada occurrences, and the material-balance 'problem' is not as serious because the ^{18}O depletions are much less (counterbalanced, of course, by the fact that the Ammonia Tank eruption was much larger than any of the low–^{18}O eruptions at Yellowstone).

The low–^{18}O magma problem

There are many as yet unanswered questions about the origin of low–^{18}O magmas. This problem should in fact be generalized further to include a variety of other types of magmas generated in sub–volcanic terranes. It cannot be too strongly stressed that our knowledge of the existence of this phenomenon is entirely attributable to our ability to make precise ^{18}O/^{16}O measurements, because at the present time there is really *no other way of recognizing* the existence of such magmas. Thus, exactly the same mechanism may be operating in areas of higher–^{18}O hydrothermal waters (*e.g.,* sea waters, metamorphic waters, formation waters, etc.), but then we would not be able to discern that a unique process was involved, because the δ^{18}O signature would be obscured.

Clearly, a major problem that needs to be solved is why these low–^{18}O magmas are formed in some volcanic fields but not in others. For example, why should some of the eruptive units in the southwest Nevada and Yellowstone volcanic fields show an order of magnitude greater ^{18}O depletion than any of those from the central Nevada, San Juan, and Superstition volcanic fields? This is the most striking problem raised by the work of LARSON and TAYLOR (1986). These dramatic ^{18}O/^{16}O differences cannot be attributed to variations in the δ^{18}O of the associated meteoric ground waters, which certainly were quite low, particularly in Nevada and Colorado (<-10). The differences must be due to some differences in the way the magma chambers at Yellowstone and the southwest Nevada complex interacted with their environment. For some reason, those particular sub-volcanic magma chambers interacted much more strongly with hydrothermally altered roof rocks or with the meteoric ground waters themselves. Based on the new data obtained by LARSON and TAYLOR (1986), this process (whatever it is) must be much less common than heretofore believed (*e.g.,* LIPMAN and FRIEDMAN, 1975, p. 701).

LARSON and TAYLOR (1986) concluded that the ^{18}O depletions in caldera–related rhyolite magmas are *not* directly related to: (1) the size of the eruption; (2) the duration and intensity of magmatic activity; (3) the complexity of the eruption cycle; or (4) overlapping collapse of a number of different calderas. However, (3) and (4) may be a *necessary* condition for development of low–^{18}O ash-flow tuff magmas (see below). LARSON and TAYLOR (1986) were able to discern only two factors that may be significantly different in separating the low–^{18}O and normal–^{18}O caldera complexes, namely emplacement into an extensional tectonic setting, and the existence of a wide range of chemical compositions in the silicic differentiates. It is also possible, however, that the *depth* of the magma chamber may also be important. Obviously, because these low–^{18}O effects involve hydrothermal systems connected to the surface, such low–^{18}O magmas cannot be produced in extremely deep magma chambers where the country rocks are relatively impermeable.

Emplacement into an extensional tectonic setting. This is the most important of the two factors. The two caldera complexes in the United States that exhibit the greatest $\delta^{18}O$ shifts (Yellowstone and Southwest Nevada) are both younger than 15 Ma. The period between 15 and 20 Ma corresponds to the time of transition into the brittle, fracture-dominated extensional tectonic regime of the Basin–Range province (for example, see STEWART, 1978; ZOBACK *et al.,* 1981). The Yellowstone caldera, in fact, lies on the eastern end of the currently–active Snake River Plain rift system, and the Southwest Nevada caldera complex lies in the midst of abundant Basin–Range extensional features. Such region-wide extension must produce fractures that penetrate deeply into the crust. These fractures could allow meteoric water to circulate very deeply, as they clearly have in Iceland (HATTORI and MUEHLENBACHS, 1982). Therefore, the regions around the Yellowstone and Southwest Nevada caldera complexes conceivably could have been subjected to much greater rifting and regional extension than the central Nevada or central San Juan caldera complexes, allowing for much deeper penetration of low–^{18}O meteoric ground waters into the crust. This hypothesis could be tested by specifically carrying out more comparative $^{18}O/^{16}O$ studies of caldera complexes developed in rift and non–rift environments. Nevertheless, at least with the presently available data–set, it is important to realize that the only localities on Earth where there has been clear-cut development of low–^{18}O magmas is in major rift–zone environments. Low–^{18}O magmas are identified with much less certainty in plutonic environments, but virtually all of the examples that have been found to date are also all emplaced into rift–zone settings, as in the Scottish Hebrides (FORESTER and TAYLOR, 1977) and on the margin of the Red Sea (TAYLOR, 1979; 1983).

Chemical composition. Examination of Figure 1 in HILDRETH (1981) shows that many of the known low–^{18}O rhyolites belong to his Groups I and II (high–SiO_2 rhyolites and high–SiO_2 rhyolites zoned to intermediate compositions). These extremely ^{18}O–depleted ash–flow tuffs commonly exhibit SiO_2 contents up to and above 77 weight percent, indicative of some type of very strong chemical differentiation (or fractional melting?). The normal–^{18}O ash–flow tuffs from the central Nevada and central San Juan complexes are either monotonous intermediates (*e.g.,* Fish Canyon and Monotony Tuffs), or, if they do include rhyolites, the rhyolites are zoned to SiO_2 contents no higher than 74–76 weight percent (LARSON and TAYLOR, 1986).

CONCLUSIONS

We have shown that layered gabbros typically undergo hydrothermal interaction with externally–derived aqueous fluids (meteoric water, sea water, etc.) at *much* higher temperatures than in the case of granitic plutons (typically 500°–900°C vs. 200°–450°C). This is manifested in the absence or rarity of hydrous alteration minerals (amphibole, chlorite, epidote, etc.) in gabbros that show clear-cut $\delta^{18}O$ signatures of having interacted with very large quantities of external H_2O. In such rocks, the $\delta^{18}O$ effects may be the *only* obvious indications of intense hydrothermal exchange.

Although a number of characteristics contribute to the higher alteration temperatures of the gabbros, such as higher solidus temperature, greater latent heat of crystallization, higher melt/crystal ratio, and characteristically lower fracture density, the two most important factors appear to be:

(1) *The geometry of crystallization,* wherein the layered gabbro cumulates crystallize upward from the floor of their magma chamber, and thus are overlain by a sheet of magma at $T > 1000°C$ for a protracted period of time. The cumulates are therefore held at a very high temperature long after they are rigid enough to be fractured and made susceptible to penetration by external waters. This goes on for a long period because crystallization proceeds approximately proportional to the square root of the time (NORTON and TAYLOR, 1979), and thus during the late stages, the final sheet of silicate melt takes a very long time to crystallize. This leads to two essentially decoupled hydrothermal systems in layered gabbros: A relatively low-temperature system in the roof rocks operates at high water/rock ratios, and is separated by a thin sheet of impermeable magma from the very high temperature system that operates at much lower water/rock ratios in the cumulates.

(2) *The higher magmatic H_2O contents of granitic plutons,* which is important because as long as a magmatic H_2O envelope under lithostatic pressure is present in and around the pluton, filling all the fractures, the system is totally impermeable to the external, hydrostatic system. The latter can only gain entry at very late stages after the magmatic water system has dissipated.

We also conclude that low-^{18}O magmas are formed mainly by re-melting of hydrothermally altered country rocks (and their pore waters) or by large-scale assimilation of such materials. In conjunction with the new data by HILDRETH *et al.* (1984) and LARSON and TAYLOR (1986), this suggests that low-^{18}O magmas can be formed: (1) as a result of catastrophic caldera collapse; (2) on very short time-scales (<100,000 years); (3) in enormous amounts (>1000 km^3 in several localities); (4) however, these giant occurrences are found only in rift-zone tectonic settings; and (5) such magmas are only rarely formed in other types of plutonic-volcanic environments.

In retrospect, the connection between a rift-zone tectonic setting and low-^{18}O magmas may seem to be a fairly obvious one. After all, where on Earth is there a better chance of bringing into close juxtaposition the two geological materials that are essential to make very high-temperature hydrothermal systems and low-^{18}O magmas? Only in rift zones and spreading centers do we find the large-scale extensions and brittle fracturing that are necessary to allow massive amounts of magma to come upward into the crust, as well as providing the greatly increased fracture permeability that allows surface waters to penetrate to depths of at least 10 or 15 km. Such environments represent the best way to attain the required combination of very high temperatures together with large quantities of low-^{18}O, hydrothermally altered rocks and meteoric pore waters. This is also the environment of essentially all of the layered gabbro bodies that have so far been shown to have interacted with surface waters at very high temperatures.

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