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WATER/ROCK INTERACTIONS
AND THE ORIGIN OF
H₂O IN GRANITIC BATHOLITHS*

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SUMMARY

D/H and $^{18}\text{O}/^{16}\text{O}$ data have now been obtained on a wide variety of granitic batholiths of various ages. The primary δD values of the biotites and hornblendes are remarkably constant at about -50 to -85, identical to the values in regional metamorphic rocks, marine sediments and greenstones, and most weathering products in temperate climates. Therefore the primary H_2O in these igneous rocks is probably not "juvenile", but is ultimately derived by dehydration and/or partial melting of the lower crust or subducted lithosphere. Most granitic rocks have $\delta^{18}\text{O} = +7.0$ to $+10.0$, probably indicating significant involvement of high- ^{18}O metasedimentary or altered volcanic rocks in the melting process; such an origin is demanded for many other granodiorites and tonalites that have $\delta^{18}\text{O} = +10$ to $+13$. Gigantic meteoric-hydrothermal convective circulation systems were established in the epizonal portions of all batholiths, locally producing very low $\delta^{18}\text{O}$ values (particularly in feldspars) during subsolidus exchange. Some granitic plutons in such environments also were emplaced as low- ^{18}O magmas probably formed by melting or assimilation of hydrothermally altered roof rocks. However, the water/rock ratios were typically low enough that over wide areas the only evidence for meteoric H_2O exchange in the batholiths is given by low D/H ratios (δD as low as -180); for example, because of latitudinal isotopic variations in meteoric waters, as one moves north through the Cordilleran batholiths of western North America an increasingly higher proportion of the granitic rocks have δD values lower than -120. The lowering of δD values commonly correlates with re-setting of K-Ar ages, and in the Idaho batholith two broad zones (10,000 km^2) can be defined where δD biotite < -100 and K-Ar "ages" have all been re-set to values less than 60 m.y., suggesting that the Ar loss was caused by the meteoric-hydrothermal circulation systems. In certain Precambrian batholiths, a much different type of very low-temperature, regional alteration by surface-derived waters took place over an extended period long after emplacement, producing "brick-red" feldspars and markedly discordant Rb-Sr isochron "ages".

1. INTRODUCTION

The purpose of this paper is to review some of the recent isotopic evidence for deep circulation and interaction of heated meteoric ground waters with granitic plutons. These phenomena are amenable to study because: (1) Most plutonic granitic rocks throughout the world have relatively uniform and characteristic isotopic compositions ($\delta^{18}\text{O} \approx +7.0$ to $+10.0$ and $\delta\text{D} \approx -50$ to -85 , in per mil relative to SMOW, standard mean ocean water). (2) Simple fractional crystallization of granitic magmas cannot of itself produce any major changes in $\delta^{18}\text{O}$ or δD in the later differentiates, because the equilibrium isotopic fractionations between silicate liquids and minerals are very small at magmatic temperatures. (3) Surface-derived ground waters such as ocean waters, fresh meteoric waters, and connate formation waters, all have relatively unique and characteristic $\delta^{18}\text{O}$ and δD values, typically much different than the values found in deep-seated igneous rocks; thus, major isotopic changes can occur when such waters interact with the rocks.

The above assertions will not be justified in any great detail in the present paper, even though they form an essential foundation for the following discussion. Several recent reviews of these problems are available to the interested reader (Garlick, 1966, 1969; Taylor, 1968, 1974a; Savin and Epstein, 1970a, 1970b; Shieh and Taylor, 1969; Magaritz and Taylor, 1976a; Friedman et al., 1964; Craig, 1963; Anderson et al., 1971).

For purposes of discussion, in the present paper all granites, quartz monzonites, granodiorites, and tonalites may be conveniently classified into 3 groups on the basis of their whole-rock $\delta^{18}\text{O}$ values. These somewhat arbitrary groupings are as follows: Low- ^{18}O Granitic Rocks with $\delta^{18}\text{O} < +6$, Normal- ^{18}O Granitic Rocks with $+6 < \delta^{18}\text{O} < +10$, and High- ^{18}O Granitic Rocks with $\delta^{18}\text{O} > +10$. Another feature of the Low- ^{18}O Group is that such rocks usually exhibit δD values lower than -85 , commonly as low as -150 or lower. Such low δD values can also be found sporadically throughout the other two groups, but by far the majority of these other samples have δD values in the range -50 to -85 , and this is termed the "normal" D/H range.

Very few igneous or metamorphic rocks have δD values outside the interval -50 to -85 , and such rocks tend to be confined to unique geologic environments. For example, even though δD values as low as -200 are known in igneous rocks, values lower than -85 generally occur only in the vicinity of localities where at least some of the rocks also have unusually low $\delta^{18}\text{O}$ values indicative of meteoric-hydrothermal alteration. Igneous biotites and hornblendes with δD higher than -60 are extremely rare, and values higher than -35 are virtually non-existent. Whole-rock samples with $-60 < \delta\text{D} < -35$ are typically confined to muscovite-rich granites and pegmatites, to phlogopites from kimberlites, and to rocks such as the Na-rich trondhjemites associated with ophiolite complexes, which have probably interacted in some manner with heated ocean waters or marine pore fluids (Taylor, 1974a; Sheppard and Epstein, 1970; Magaritz and Taylor, 1974, 1976a).

In addition to having whole-rock $\delta^{18}\text{O}$ values in the range +6 to +10, another important feature of Normal- ^{18}O Granitic Rocks is that the coexisting minerals in a given sample always increase in ^{18}O in the same sequence, namely: magnetite-biotite-hornblende-muscovite-plagioclase-alkali feldspar-quartz (Taylor and Epstein, 1962). This well-defined sequence is a result of an approach to isotopic equilibrium in the mineral assemblage at the time of formation, and it is characteristic of all plutonic granitic rocks that have not been markedly "disturbed" by some later event such as an episode of hydrothermal alteration. The $^{18}\text{O}/^{16}\text{O}$ fractionations between any two minerals in such a rock are reported as $\Delta^{18}\text{O}$ values, which are defined simply as the difference in $\delta^{18}\text{O}$ between the two minerals. Not only do these $\Delta^{18}\text{O}$ values in Normal- ^{18}O rocks always have the same sign, they usually exhibit a very narrow range of values; for example, $\Delta^{18}\text{O}_{\text{quartz-alkali feldspar}}$ typically will be about 1.0 to 1.5, and $\Delta^{18}\text{O}_{\text{quartz-plagioclase}}$ will be about 1.5 to 2.5, being higher the more calcic the plagioclase.

Just as with the $^{18}\text{O}/^{16}\text{O}$ ratios, in an "undisturbed" mineral assemblage there is usually a constant sequence of δD values, as a result of an approach to hydrogen isotopic equilibrium. The sequence of increasing D/H ratios is typically as follows: Fe-rich biotite and hornblende-Mg-rich biotite and hornblende-chlorite-muscovite (Taylor and Epstein, 1966; Suzuoki and Epstein, 1976). If coexisting biotite and hornblende both have essentially the same Mg/Fe ratio, then at equilibrium $\Delta\text{D}_{\text{biotite-hornblende}}$ will be

about zero. Also in most plutonic igneous rocks $\Delta D_{\text{muscovite-biotite}}$ values are about 15 to 30, increasing with increasing Fe/Mg in the biotite.

The wealth of D/H data on rocks and minerals gathered in the last decade has established that the δD values of almost all regional metamorphic rocks, altered marine volcanic rocks, and sedimentary rocks, as well as most clay minerals formed by weathering in temperate to sub-tropical climates, exhibit a relatively narrow range of $\delta D \approx -40$ to -85 (Savin and Epstein, 1970a, 1970b; Lawrence and Taylor, 1971; Taylor and Epstein, 1966; also see the compilation in Taylor, 1974a). A similar range of δD values was also found for the graywackes, shales, basalts, and ophiolite complexes from the eugeosynclinal Franciscan formation in California, as shown in Fig. 1 (Magaritz and Taylor, 1976a). Note that this range of hydrogen isotopic compositions is also practically identical to that established for so-called "normal" igneous rocks, as discussed above.

Is this overlap of D/H values in all these different kinds of rocks a coincidence, or is there some kind of cause-and-effect relationship involved? Before the concepts of plate tectonics became firmly established, it was generally thought that these isotopic similarities were probably coincidental, because the δD values of the sedimentary rocks and weathering products were definitely known to be fixed by D/H exchange at low temperatures with surface waters, whereas the δD values of many deep-seated igneous rocks were thought to be due to incorporation of "juvenile" H_2O in the magmas. However,

if sea-floor spreading has been going on at anything like its present rate during a good portion of geologic time, the reason for the overlapping δD values of all these different kinds of rocks is easily understood.

In a slab of oceanic lithosphere that is being returned to the upper mantle along a subduction zone, the dominant OH-bearing minerals (clays, chlorites, zeolites, etc.) will be those present in the marine sediments or those that have been formed by submarine weathering or hydrothermal alteration at an ocean-ridge spreading center. Ultimately, the fate of all this hydroxyl water must be release by dehydration as the subducted slab heats up, or incorporation in any silicate liquids formed during melting of the slab. Over the course of geologic time, any original "juvenile" H_2O in the upper mantle (but not the deep mantle!) would probably have been completely overwhelmed by the enormous amounts of subducted water that have been cycled through the outermost 400-700 km thick shell of the Earth. Therefore, in terms of its D/H ratio, it would make no difference whether a magma was derived by anatexis of marine sedimentary and volcanic rocks in a deep-seated metamorphic complex, or whether it formed by differentiation of a "primary" magma derived directly from the upper mantle.

Summarizing the concepts outlined above, we cannot at present utilize $^{18}O/^{16}O$ and D/H measurements to conclude anything definitive about the origin of isolated bodies of igneous rocks where $\delta^{18}O$, $\Delta^{18}O$, δD , and ΔD values happen to

lie in the so-called "normal" range, but a special origin is required for any granitic rock in which these parameters have anomalous values.

Granitic magmas with $\delta^{18}\text{O}$ values higher than +10 certainly require either derivation from, or exchange with, some type of high- ^{18}O sedimentary rock or altered volcanic rock (or the metamorphosed equivalents of such rocks). Such High- ^{18}O

Granites are found locally in the Cordilleran batholiths of western North America (see below), but in certain cases they make up entire igneous provinces. Good examples are the Tuscan province in Italy (+12 to +16, Taylor and Turi, 1976) and a belt of granitic plutons in Japan (+10 to +13, Matsuhisa, et al., 1972), as well as practically all muscovite-bearing granites (Taylor, 1968).

Granitic rocks with $\delta^{18}\text{O}$ lower than +6 cannot be derived by any known differentiation process from "normal" basaltic magma, and they must be formed (1) by melting or exchange with preexisting low- ^{18}O rocks, (2) by later exchange with low- ^{18}O hydrothermal fluids either in the magmatic state or later under subsolidus conditions, or (3) by differentiation from a low- ^{18}O magma formed by one of the above processes.

2. INTERACTIONS BETWEEN METEORIC GROUND WATERS AND IGNEOUS INTRUSIONS

Distribution and General Characteristics

When epizonal igneous intrusions are emplaced into highly jointed, permeable country rocks, they act as gigantic "heat engines" that provide the energy necessary to

promote a long-lived convective circulation of large amounts of ground H_2O in the country rocks surrounding the igneous body. This produces a depletion of ^{18}O in the pluton and its country rocks and a corresponding ^{18}O enrichment or " ^{18}O -shift" in the water (Craig, 1963). Inasmuch as the deuterium contents of meteoric ground waters are always lower than ocean water, and in many cases are lower than the primary magmatic waters in igneous rocks, the δD values are also commonly lowered in such rocks.

Low- ^{18}O igneous rocks produced by interaction with large quantities of meteoric ground waters have now been observed in many different localities (Taylor and Epstein, 1963, 1968; Taylor, 1968, 1971, 1973, 1974a, 1974b, 1974c; Taylor and Forester, 1971, 1973; Forester and Taylor, 1972; Sheppard and Taylor, 1974; Friedman et al., 1974; Hall et al., 1974; Muehlenbachs et al., 1974; O'Neil et al., 1973). These types of hydrothermal systems represent the "fossil" equivalents of the deep portions of modern geothermal water systems such as occur at Wairakei, New Zealand; Steamboat Springs, Nevada; and Yellowstone Park, Wyoming (e.g. Banwell, 1961; White, 1968).

The igneous complexes that exhibit extreme ^{18}O -depletions characteristically display the following geological, petrological, and isotopic features. (1) The intrusions often occur as sub-volcanic ring complexes associated with explosion breccias, surrounded by highly fractured country rocks that are very permeable to groundwater movement. (2) The feldspars are commonly depleted in ^{18}O to a greater degree than the other coexisting minerals and they also usually show a "clouding" or turbidity (particularly the

alkali feldspars). (3) The primary igneous minerals are commonly partially altered to uralitic amphibole, chlorite, Fe-Ti oxides, and/or epidote; locally, this process is complete and only pseudomorphs of the primary igneous minerals remain. (4) Granophyric (micrographic) intergrowths of turbid alkali feldspar and quartz are ubiquitous.

(5) Mirolitic cavities are locally present in the intrusives, and veins filled with quartz, alkali feldspar, epidote, chlorite, or sulfides are very common in both the intrusives and the surrounding country rocks.

In certain cases, even in extremely low- ^{18}O rocks, the characteristic petrographic features outlined above are minor or totally absent, indicating that the hydrothermal alteration either occurred at very high temperatures (above the stability fields of some of the OH-bearing minerals) or that the minerals crystallized directly from low- ^{18}O magmas. These low- ^{18}O magmas are themselves probably produced by some process of exchange with meteoric waters, although as will be discussed below, this does not necessarily involve direct influx of meteoric water into the magmas.

In the following discussion we shall briefly review only one of the many occurrences of extremely low- ^{18}O igneous rocks. This is a typical example, however, and it is important to grasp from the above list of references how widespread these phenomena are, as well as to understand the large scale at which they operate. These meteoric-hydrothermal processes are extremely common in volcanic terranes, and analogous systems also clearly operate at mid-ocean ridge spreading centers, there involving massive convective circulation of ocean waters through the newly formed oceanic crust (Spooner et al., 1974). Such systems involve truly enormous amounts of surface waters and are responsible for the hydrothermal alteration of very large volumes of rocks.

Inner Hebrides, Scotland

Isotopic evidence for interaction between meteoric ground waters and epizonal igneous intrusions was first discovered at the Skaergaard intrusion (Taylor and Epstein, 1963), but its enormous scale and pervasive effects were first recognized in the Tertiary volcanic centers of Skye, Mull, and Ardnamurchan in western Scotland (Taylor, 1968). Later work by Taylor and Forester (1971) and Forester and Taylor (1976a, 1976b) established the extent of the ^{18}O effects around each intrusive complex (Fig. 2). Within the boundaries outlined on Fig. 2, essentially all the rocks have been depleted in ^{18}O , some by as much as 13 per mil. The average ^{18}O depletion is about 6 to 7 per mil at both Skye ($\sim 400 \text{ km}^2$) and Mull ($\sim 400 \text{ km}^2$), but only about 4 per mil at Ardnamurchan ($\sim 80 \text{ km}^2$). The effects are smaller at Ardnamurchan because it is a smaller center, but also probably because the country rocks there are largely Precambrian basement rocks; such rocks are less permeable than the overlying highly jointed lavas.

All analyzed samples of early Tertiary volcanic rocks from Skye that are older than the Red Hills granites are plotted on the regional geologic map shown in Fig. 3, and $\delta^{18}\text{O}$ contours are drawn through the data-points. These rocks are mainly basalts and all are relatively fine-grained and have similar mineralogy; therefore they should all have offered approximately equal resistance to hydrothermal exchange.

The whole-rock $\delta^{18}\text{O}$ values contoured in Fig. 3 are extremely systematic, decreasing inward toward the center of the plutonic complex. Essentially all the samples

that lie outside the +5 contour have "normal" $\delta^{18}\text{O}$ values, but the samples within the $\delta = -5$ contour must have been depleted in ^{18}O by at least 10 per mil.

In Fig. 4, the above $\delta^{18}\text{O}$ data for the Skye basalts are generalized and plotted as a function of distance outward from the central intrusions. Similar generalized curves are also shown for some pre-Tertiary sandstone and shale country rocks from Skye; these various curves for the Skye complex can then be compared with the analytical data obtained for the basaltic country rocks surrounding the central complex at Mull (Fig. 4). Note that all of the country rocks surrounding these centers are systematically depleted in ^{18}O , but the isotopic effects extend outward for a greater distance in the basalts than in the sandstones. This can be attributed to a higher permeability in the basalts, as well as to the greater resistance of quartz to hydrothermal exchange. The radial inward depletion in ^{18}O is caused in part by increasing temperatures toward the intrusive centers and in part by an inward increase in the overall, integrated water/rock (W/R) ratio. The latter increases toward the intrusives, because the flow of water is radially inward and then upward, and the rock volume decreases inward as the square of the distance (see meteoric-hydrothermal convection model of Taylor, 1971, and Taylor and Forester, 1971).

The δD values of the OH-bearing minerals at Skye and Mull are also all much lower than "normal", but in contrast to the $\delta^{18}\text{O}$ data the δD values for a given mineral are extremely uniform throughout the entire complex. Just as is found elsewhere in the

world, the sericites concentrate deuterium relative to the Fe-bearing hydroxyl minerals. The sericites have δD values of -104 to -107, whereas the chlorite δD values vary from -122 to -132, and the amphiboles vary from -119 to -128 (Taylor and Epstein, 1968; Forester and Taylor, 1976a).

Utilizing the equilibrium D/H fractionation factors for muscovite-water and amphibole-water determined by Suzuki and Epstein (1976), we can estimate that the Eocene meteoric-hydrothermal waters at both Skye and Mull had $\delta D \approx -85$. Using the well-established relationship between δD and $\delta^{18}O$ in pristine meteoric waters (Craig, 1961), this gives $\delta^{18}O \approx -12$ for the early Tertiary meteoric surface waters on Skye. Because of the characteristic positive " ^{18}O shift" produced in geothermal waters as a result of water-rock exchange (Craig, 1963), this represents a minimum value for all the Eocene hydrothermal waters at Skye; therefore all of the oxygen isotope effects discussed above must have been produced by waters with $\delta^{18}O > -12$.

Red Hills Granites, Skye

All the rocks within the central intrusive complexes in the Scottish Hebrides are depleted in ^{18}O , including for example the entire Cuillin gabbro mass on Skye (Forester and Taylor, 1976a). However, inasmuch as in this paper we are mainly concerned with granitic rocks, only the $\delta^{18}O$ data obtained on the Red Hills granites are plotted in Fig. 5, as a function of sequence of intrusion. The relative ages of

intrusion of these granites were established solely by geologic field mapping (see Stewart, 1965), and are definitively determined only for the Western Red Hills granites.

The isotopic data in Fig. 5 are grouped according to the major geologic discontinuities in the sequence of granitic intrusions. In general, there was a major geographic shift in the focus of granitic ring-dike intrusion at each of these discontinuities. Although there is appreciable scatter in the isotope data, the age discontinuities seem to correlate with breaks in the $\delta^{18}\text{O}$ data. In particular the most abrupt change in the $\delta^{18}\text{O}$ trend of quartz occurs at the major geological discontinuity, namely between the Marsco epigranite of the Western Red Hills and the Glas Beinn Mhor epigranite of the Eastern Red Hills. Note that the quartz and feldspar samples are all lower in $\delta^{18}\text{O}$ than those from "normal" granites, and the $\Delta^{18}\text{O}$ fractionations between the quartz and coexisting feldspar are invariably much larger than the near-equilibrium values of 1.0 to 1.5 per mil that are typically found in "normal" granitic rocks. These large $\Delta^{18}\text{O}_{\text{quartz-feldspar}}$ values imply that the bulk of the ^{18}O depletion in the feldspars must have occurred after solidification of the plutons. The euhedral quartz and feldspar crystals in the miarolitic cavities (indicated by "m" in Fig. 5) tend to have lower $\delta^{18}\text{O}$ values than the host rock minerals, compatible with their direct crystallization from a low- ^{18}O aqueous gas phase in a somewhat higher meteoric water/rock environment. For example, three different samples of miarolitic quartz from a single hand specimen of

the Loch Ainort epigranite have $\delta^{18}\text{O} = +1.7, +2.6, \text{ and } +3.0$, all lower in ^{18}O than the host-rock quartz ($\delta^{18}\text{O} = +4.1$). Therefore, the miarolitic cavities were probably produced by late-stage influx of meteoric-hydrothermal fluids into the crystallizing granite plutons.

Within each age-grouping shown in Fig. 5, much larger ^{18}O depletions in feldspar are seen in the earlier intrusions such as the Maol na Gainmhich epigranite than in the later intrusions. This is probably a result of the fact that each successive arcuate pluton was commonly intruded along a ring fracture interior to an earlier pluton, and at each stage a new hydrothermal convective system ought to have been established or an older, preexisting system modified. Other things being equal, we would then predict that the older intrusives would experience a greater number of overlapping hydrothermal alteration events (and hence more ^{18}O depletion) than the younger intrusions. As time goes on, the meteoric-hydrothermal systems tend to be self-sealing because the joints and fractures become filled with vein minerals deposited from the circulating solutions. These various effects are cumulative, and they all indicate that the feldspars of the older, outer-most ring intrusions should show the greatest ^{18}O depletions.

Inasmuch as quartz is the most resistant rock-forming mineral to hydrothermal $^{18}\text{O}/^{16}\text{O}$ exchange, of all the minerals present in an altered igneous rock the $\delta^{18}\text{O}$ of the quartz should most closely reflect the $\delta^{18}\text{O}$ value of the original magma. Therefore,

the most logical explanation why the $\delta^{18}\text{O}$ values of quartz in the Early Epigranites tend to decrease with time, while the feldspars in the same rocks show the expected increase with time, is that these granitic intrusions were emplaced as low- ^{18}O magmas, with the $\delta^{18}\text{O}$ of the magmas decreasing progressively with time. A general $\delta^{18}\text{O}$ -lowering of the magmas with time might perhaps be expected if meteoric water continuously exchanged with a shallow magmatic reservoir; the youngest magmas would be the most affected because they were exposed to exchange for the longest time. For example, this is the type of $\delta^{18}\text{O}$ -time trend shown by some rhyolitic ash-flow tuff magmas studied by Friedman et al. (1974). An equally likely possibility is that if the granitic magmas were formed by partial melting of preexisting country rocks, as suggested by Pb and Sr isotope data (see Moorbath and Bell, 1965; Moorbath and Welke, 1969; and Taylor and Forester, 1971), the younger magmas might be derived from rocks that had undergone progressively greater ^{18}O depletion as the meteoric-hydrothermal activity built up to its maximum.

Fig. 5 shows that the Southern Porphyritic Epigranite has the lowest $\delta^{18}\text{O}$ values found in any of the Skye granites. It is very doubtful that these marked ^{18}O depletions could be the result of subsolidus meteoric-hydrothermal exchange, because the quartz crystals in this rock are typically euhedral, have β -quartz morphology, and are among the coarsest in grain size of any quartz from the Red Hills granites. These data thus suggest that the Southern Porphyritic Epigranite was originally intruded as an extremely

inhomogeneous, markedly ^{18}O -depleted magma (Forester and Taylor, 1976a). If so, this particular intrusion would represent the lowest- ^{18}O magma that has yet been documented, inasmuch as all previous examples have $+2 < \delta^{18}\text{O} < +6$ (see below).

In order to more fully document this hypothesis, Forester and Taylor (1976a) separated a set of single-crystal quartz phenocrysts from the most ^{18}O -depleted sample of Southern Porphyritic Epigranite, and these were then subjected to some fluorine stripping experiments similar to those described by Epstein and Taylor (1971) on the lunar regolith samples. The results of these experiments, in which the external portions of the quartz crystals were successively "peeled" away by reaction with F_2 , are given in Fig. 6. This experiment demonstrates that the quartz crystals are isotopically homogeneous throughout, and that their ^{18}O depletion cannot be attributed to later subsolidus hydrothermal exchange. This in turn implies that the $\delta^{18}\text{O}$ value of -2.6 in this quartz must have been a primary characteristic of the quartz at the time of its crystallization from a magma that had a very low $\delta^{18}\text{O} \approx -2.0$. The argument is strengthened by noting that this particular sample is one of the few granites from Skye with a near normal $\Delta^{18}\text{O}_{\text{quartz-feldspar}}$ fractionation.

Low- ^{18}O Magmas

In light of the data by Forester and Taylor (1976a) indicating that low- ^{18}O granitic magmas existed at Skye, it is worthwhile to discuss this general phenomenon in greater detail.

The first tangible evidence for the existence of true low- ^{18}O magmas was provided by the inner diorite of the Stony Mountain ring complex, Colorado (Forester and Taylor, 1972). Undeniable evidence for other low- ^{18}O magmas has since been furnished by Muehlenbachs et al. (1974) and Friedman et al. (1974) for several young basaltic lava flows on Iceland and some rhyolitic ash-flow tuffs from the western United States, as well as by Taylor (1974c) for the 650 m.y. old Seychelles Islands granites in the Indian Ocean. Mechanisms for generating ^{18}O -depleted magmas have been discussed in these papers and by Taylor (1968, 1974a, 1974b). The most likely mechanisms are: (1) Oxygen isotope exchange between a normal- ^{18}O magma and country rocks that had earlier been hydrothermally altered and thereby depleted in ^{18}O . (2) Partial fusion of the hydrothermally altered, ^{18}O -depleted country rocks in the roof zone above the magma chamber. Direct diffusion of some H_2O into the magmas probably also occurred, but because of the very limited amounts of H_2O that can be dissolved in epizonal magmas, such a process cannot quantitatively account for the large ^{18}O depletions observed in many of these low- ^{18}O magmas.

A question of great interest in this connection is whether or not any considerable amount of meteoric H_2O is able to diffuse directly into the liquid magma, because at first glance this might also appear to be a plausible way to produce low- ^{18}O magmas. However, studies of the Skaergaard layered gabbro intrusion (Taylor and Forester, 1973; also see pp. 861-864 in Taylor, 1974a) have shown that even though a meteoric-hydrothermal convection system was established soon after the magma was emplaced, these hydrothermal

fluids produced little or no ^{18}O depletion of the liquid magma throughout practically the entire period of crystallization, which must have lasted thousands of years. Thus, it is clear that very little meteoric H_2O diffused into the Skaergaard magma, even though the $P_{\text{H}_2\text{O}}$ in the hydrostatic fissure system outside the magma chamber must have been at least 300–500 bars.

The reason why it is difficult for the water to move directly into a magma is that the latter is under a lithostatic pressure a factor of 2.5 to 3 higher than the hydrostatic pressure in the fissure system. The fractures which provide the major conduits for H_2O circulation outside the intrusion obviously cannot be present in the immediate contact zone at the edge of the magma body; otherwise they would be forcibly filled with magma. Therefore, if H_2O is to gain access directly to the magma, it must be by the relatively slow process of grain-boundary diffusion up a thermal gradient through the hot contact zone and into the magma. This will be the rate-controlling step in any such process. Once the H_2O is inside the magma chamber, convective circulation of the silicate melt can, in principle, aid the diffusion process in distributing this H_2O through the interior of the magma body. However, the Skaergaard intrusion is one of the best documented examples of convection in magmas, and such a process did not seem to be very effective there; in any case, any type of convective process will be enormously less efficient in the more viscous, SiO_2 -rich, granitic magmas.

The difficulties outlined above are compounded by the fact that as far as ^{18}O effects are concerned, very large amounts of H_2O are required to produce any significant

$\delta^{18}\text{O}$ lowering of the magma. First of all, any H_2O finally able to diffuse into the magma would probably have already undergone a significant ^{18}O shift to much higher $\delta^{18}\text{O}$ values than those characteristic of the cool ground waters in the surrounding terrane. If $\delta^{18}\text{O}_{\text{H}_2\text{O}} = -5$, for example, then to produce even a modest $\delta^{18}\text{O}$ lowering of a magma from a "normal" $\delta^{18}\text{O} = +7$ to a value of $+5$ would require an amount of H_2O equivalent to about 15 wt. percent H_2O in the magma; this is far more H_2O than can be dissolved in a magma at such shallow depths in the earth's crust, thereby implying that most of this H_2O would have to diffuse or bubble through the magma chamber and then out again! It is therefore doubtful that direct influx of low- ^{18}O meteoric H_2O into a magma chamber can by itself ever account for more than about a one per mil lowering of the $\delta^{18}\text{O}$ of a large magma body.

If the above models for the generation of low- ^{18}O magmas are valid they require that the low- ^{18}O granitic magmas at Skye must have been formed by large-scale melting or assimilation of hydrothermally altered country rocks. In support of the above model, note that the Pb and Sr isotope data of Moorbath and Bell (1965) and Moorbath and Welke (1969) also require major involvement of the country rocks in the magmatic processes that formed the Skye igneous complex. Also, it has previously been suggested that certain granitic rocks on Skye, including the Southern Porphyritic Epigranite, were most likely formed by partial melting of the Torridonian arkosic sandstones (Wager et al., 1953; Brown, 1963; Thompson, 1969). This could help explain the large $\delta^{18}\text{O}$ variations in the quartz from different samples of this epigranite (Fig. 5), as the

source rock itself likely would have suffered varying degrees of meteoric-hydrothermal alteration and would thus have been isotopically non-homogeneous. Any melting would then probably give rise to a magma that was also inhomogeneous in terms of $\delta^{18}\text{O}$. Note that the hydrothermally altered Torridonian sandstones on Skye do in fact show a wide range in $\delta^{18}\text{O}$ (Fig. 4), with some values as low as -4.

Oxygen isotope exchange between the liquid magma and the hydrothermally altered wall rocks that form a shell around the magma chamber would be facilitated by the high H_2O content of the hydrothermally altered country rocks, and by the fact that the ^{18}O -depleted, turbid feldspars in the country rocks are very easily exchanged and thus extremely susceptible to a second alteration event (O'Neil and Taylor, 1967). The efficiency of this exchange process depends on the surface-to-volume ratio of the magma chamber, and would be enhanced by the presence of low- ^{18}O xenolithic debris provided by piecemeal stoping of the shattered roof and wall rocks. Note that this process is not necessarily one of wholesale dissolution and assimilation, but merely communication and selective exchange between oxygen in the silicate melt and oxygen in the country rocks or stoped blocks (probably mainly the feldspars and hydroxyl minerals).

Low-Deuterium Magmas

It is logical to believe that low-deuterium magmas ought to be much more common in nature than low- ^{18}O magmas. This is because the oxygen content of a typical magma

is enormously higher than the amount of hydrogen that can be dissolved in the form of H_2O in such a magma. Therefore, addition of very small amounts of meteoric water can drastically lower the original δD value of a magma. Unfortunately, in contrast to the variety of arguments that are possible utilizing $^{18}O/^{16}O$ data, there are very few criteria for deciding whether the δD values in minerals such as hornblende or biotite represent primary values or are the result of later sub-solidus exchange. Whereas there are order-of-magnitude differences in the rates of $^{18}O/^{16}O$ exchange among minerals (e.g. between quartz and feldspar), such differential kinetic effects are much less apparent for D/H exchange among hydroxyl minerals. However, if the grain sizes are roughly the same, there are some indications that during meteoric-hydrothermal alteration muscovite preserves its D/H ratio slightly better than coexisting biotite (Shieh and Taylor, 1969; Magaritz and Taylor, 1976c).

It is obvious from the above discussion that if the local meteoric waters are sufficiently low in deuterium, then the low- ^{18}O magmas should also be low-D magmas; however, the reverse is not necessarily true. Also, in certain geographical areas the surface meteoric waters may have δD values even higher than the "normal" values in deep-seated igneous rocks (see Friedman et al., 1964); thus, in certain climatic areas (warm, temperate) it is possible to develop low- ^{18}O magmas that by coincidence appear to have "normal" deep-seated δD values. In addition, if sea waters are involved, it is theoretically possible to develop slightly ^{18}O -depleted magmas with abnormally high δD values.

The final δD of the magma is also dependent upon exactly how the meteoric H_2O enters. If it diffuses directly into the magma, which is perhaps plausible for small amounts of H_2O , then the final δD of the contaminated magma could approach the δD of the surrounding meteoric waters. However, if the meteoric H_2O is added by partial melting or by incorporation of stopped blocks of hydrothermally altered roof rock, the final δD value of the magma conceivably could be as much as 30 to 50 per mil lower than that of the local meteoric water. This is because of the fact that under equilibrium conditions at 200° - $500^\circ C$, H_2O strongly concentrates deuterium relative to hydroxyl-bearing minerals. However, to date only a few possible examples of such low- δD magmas have been identified (e.g. the Seychelles Islands, see below), perhaps because later, sub-solidus, meteoric-hydrothermal exchange is very likely in such environments and would tend to produce similar δD values.

In the future, detailed isotopic studies in favorable geographic areas (i.e. where the meteoric waters are very low in deuterium) may allow us to sort out the various effects described above and to prove the existence of low-deuterium magmas. More important, it may even be possible to identify the exact mechanism of formation of such magmas. Also, as discussed in a subsequent section, if true "juvenile" H_2O just happened to have a low δD value, then low- δD magmas conceivably could form in other ways than by interaction with meteoric ground waters. In any case, in the discussion that follows the possibility of formation from a low-deuterium magma cannot be discounted for any rocks which exhibit the combination of low δD values together with "normal" $\delta^{18}O$ and $\Delta^{18}O$ values.

Water/Rock Ratios

The original $\delta^{18}\text{O}$ values of igneous rocks that have been hydrothermally altered can be estimated from the "normal" values of such rocks, or better yet, by analyzing the same formations outside the hydrothermally altered area. The initial δD value of the H_2O can be estimated from D/H analyses of the alteration mineral assemblages, and then by application of the meteoric water equation we can calculate the initial $\delta^{18}\text{O}$ of this water. Given the above parameters, together with some information about the temperatures of alteration, the total amounts of water involved in any meteoric-hydrothermal convective system can be calculated by material balance:

$$W \delta_{\text{H}_2\text{O}}^i + R \delta_{\text{Rock}}^i = W \delta_{\text{H}_2\text{O}}^f + R \delta_{\text{Rock}}^f$$

where i = initial value; f = final value after exchange; W = atom percent of meteoric water oxygen in the total system; R = atom percent of rock oxygen in the bulk system.

If $\delta_{\text{H}_2\text{O}}^f$ is determined by isotopic equilibration with the rocks, then:

$$W/R = \frac{\delta_{\text{Rock}}^f - \delta_{\text{Rock}}^i}{\delta_{\text{H}_2\text{O}}^i - (\delta_{\text{Rock}}^f - \Delta)}$$

where $\Delta = \delta_{\text{Rock}}^f - \delta_{\text{H}_2\text{O}}^f$. Note that for a given set of initial conditions, if W/R is a constant, then δ_{Rock}^f is determined solely by Δ , which is only a function of temperature.

Conversely, if the temperature is constant, δ_{Rock}^f is controlled only by the W/R ratio.

The W/R values defined above are based on the ratios of water oxygen to rock oxygen. Inasmuch as a typical granitic rock contains only about 45-50 wt. % O, whereas H₂O is 89 wt. % O, the water/rock ratio in weight units (abbreviated w/r) is equal to about 0.5 W/R.

Because of the common lack of attainment of isotopic equilibrium among the various minerals, to be strictly accurate one should calculate a separate Water/Mineral ratio for each mineral in a rock; these Water/Mineral ratios can then be combined with modal abundance data to calculate a W/R ratio. Nevertheless, it is usually more convenient to make the reasonable approximation that in such systems δ_{Rock} at equilibrium is equal to the $\delta^{18}\text{O}$ value of plagioclase (An₃₀). Then we can utilize the feldspar-H₂O geothermometer to calculate Δ at any temperature ($\Delta \approx 2.68 (10^6/T^2) - 3.53$, O' Neil and Taylor, 1967).

The above calculations give W/R ratios integrated over the lifetime of the hydrothermal system, assuming continuous recirculation and cyclic re-equilibration of the H₂O with the pluton. However, some of the heated water will be lost from the system, for example by escape to the surface. In the extreme open-system case in which each increment of water makes only a single pass through the system, it can be shown that the integrated W/R ratio is given by the equation:

$$W/R = \log_e \left[\frac{\delta_{\text{H}_2\text{O}}^i + \Delta - \delta_{\text{Rock}}^i}{\delta_{\text{H}_2\text{O}}^i - (\delta_{\text{Rock}}^f - \Delta)} \right] = \log_e \left[\left(\frac{W}{R} \right)_{\text{Closed System}} + 1 \right]$$

For each of the two extreme cases discussed above, we can calculate the W/R ratios as a function of $\delta^{18}\text{O}_{\text{Rock}}^f$ and temperature; comparisons are made for the specific parameters 500°C, 300°C, and $\delta^{18}\text{O}_{\text{H}_2\text{O}}^i = -14$ in Fig. 7. In actual hydrothermal systems, the true 500°C curve would lie between the two 500°C curves plotted on Fig. 7, probably very close to the right-hand curve, as the single-pass system is a very unrealistic one. Note that both types of calculations are dependent on temperature and in both cases the lower the temperature the higher the required W/R ratio. Also, these models only give minimum values of W/R because appreciable H₂O may move through fractures in the rocks without exchanging (i.e. after the wall rocks next to the fractures have become markedly depleted in ¹⁸O).

Utilizing the above types of calculations, and assuming $\delta^{18}\text{O}_{\text{H}_2\text{O}}^i = -12$ and a reasonable average temperature of 400°-500°C, most of the Skye granitic rocks shown in Fig. 5 must have experienced time-integrated W/R ratios of approximately unity. Locally, however, W/R ratios at Skye clearly reached very large values (>5). The only exception is the Beinn an Dubhaich granite, which typically exhibits W/R lower than 0.3, presumably because this is the only Skye granite emplaced into relatively impermeable Cambrian limestone. Only locally, along fractures or in very fine-grained samples, was the W/R ratio in this granite high enough to reduce the $\delta^{18}\text{O}$ feldspar to less than +4 (Fig. 5). These differences are thought to be due to the relative ease of plastic deformation of the carbonate country rocks surrounding the Beinn an Dubhaich

granite; any fractures that do form in such rocks tend to be quickly sealed, because under hydrothermal conditions carbonates are much more easily recrystallized than silicate rocks (Forester and Taylor, 1976a).

Profound $\delta^{18}\text{O}$ effects clearly will be produced in epizonal intrusions if the meteoric-water/rock ratios are high enough. However, what happens if we are dealing with more deep-seated intrusions or with plutons such as the Beinn an Dubhaich granite that are emplaced into country rocks with relatively low permeabilities? The $\delta^{18}\text{O}$ effects will be smaller, and they may be so small, given the natural isotopic variations in igneous rocks, that no clear-cut $\delta^{18}\text{O}$ changes may be discernible. Nonetheless, even if the $\delta^{18}\text{O}$ effects are very small the δD values of the igneous rocks can still be markedly affected by the influx of relatively minor amounts of external waters.

Such D/H effects will, however, only be definitive in certain geographic areas, such as in submarine environments where $\delta\text{D} \approx 0$, or in high-latitude portions of the continents where the δD values of the meteoric waters are very low. Areas of high elevation can in principle also be used, but generally it is difficult to ascertain the original surface elevation above a pluton presently exposed by erosion. At least for Mesozoic and Tertiary plutons, latitude is a much more definite parameter than original surface elevation.

If the meteoric waters are sufficiently low in deuterium, D/H measurements can be an extremely sensitive indicator of meteoric-hydrothermal alteration. An example of

this effect is given in Fig. 8, where we show what would happen to a hypothetical granodiorite pluton that was recrystallized under conditions of successively increasing water/rock ratio. In this example, the water is assumed to have a $\delta D = -120$, typical of many present-day meteoric waters in the Montana-Idaho-British Columbia region of North America. Initially, the granodiorite is assumed to have "normal" $\delta^{18}O$ and δD values, and during the alteration process the quartz is assumed to be inert. Fig. 8 indicates that as the w/r increases from zero to 0.01 (wt. units) the δD of the biotite in the granodiorite will drop by about 60 per mil without any appreciable ^{18}O depletion in the feldspar. Only after w/r reaches 0.03 is there a significant ^{18}O depletion of the feldspar, and at that stage the δD value of the biotite has already been lowered about as far as it ever will be. From there on, as the w/r increases, the δD biotite remains at about -150 to -160 while the feldspar becomes progressively more depleted in ^{18}O . Thus, if one analyzes a series of granitic rocks altered at varying w/r ratios, on a $\delta D - \delta^{18}O$ diagram one ought to observe a characteristic inverted L-shaped pattern similar to that shown in Fig. 8. If the initial δD_{H_2O} and $\delta D_{biotite}$ are accurately known, it should theoretically be possible to determine meteoric w/r ratios as low as 0.001 in an altered granite.

3. CORDILLERAN BATHOLITHS OF WESTERN NORTH AMERICA

General Features

The previous discussion indicates that if one is interested in discovering possible interactions between small amounts of meteoric groundwaters and deep-seated plutonic

intrusions, it is necessary to study areas where the δD values of such waters were exceptionally low. Initially, this means we should restrict ourselves to plutons of relatively young age, because as one moves backward through geologic history it obviously becomes more and more difficult to determine the latitudinal and topographic variations over the land surface at the time of intrusion. These problems are not so serious, however, for the late Mesozoic and Tertiary rocks of western North America, because we can be reasonably confident that throughout this period the general size, shape, and orientation of the North American continent was similar to the present. Therefore, except for local effects attributable to topographic differences and an overall shift to higher δD values during the warmer climatic periods, during the past 150 m.y. the isotopic variations of meteoric waters across North America should have been grossly similar to the present. In fact, studies of a wide variety of rock types have demonstrated that although the ancient surface waters were somewhat higher in δD , the δD_{H_2O} values did indeed progressively decrease northward from the Caribbean area and Central America through California and Oregon, to British Columbia and Alaska (Wenner and Taylor, 1974; Sheppard et al., 1969, 1971; Taylor, 1974a; Ohmoto and Rye, 1970).

The locations of the great tonalite-granodiorite batholiths of late Mesozoic to Tertiary age are shown on a map of North America in Fig. 9. Each of these large batholiths represents an extremely complex intermingling of individual plutons of various sizes, shapes, ages, and petrographic types. Nonetheless, their overall

similarities far outweigh their differences, and because of their similar geologic settings they probably were all formed by basically the same mechanism. Therefore, in view of their broad latitudinal distribution, even on a reconnaissance basis it ought to be relatively easy to find out if any of the $^{18}\text{O}/^{16}\text{O}$ or D/H variations in these batholiths are the result of meteoric-hydrothermal alteration.

Recently, in cooperation with M. Magaritz, we have undertaken systematic investigations of some of these batholiths, building upon previous studies by Sheppard and Taylor (1974), Godfrey (1962), Taylor and Epstein (1962, 1966), Turi and Taylor (1970), and Levin and Friedman (1962).

Coast Range Batholith, British Columbia

The Coast Range batholith (or Coast Plutonic Complex) is about 1800 km long and 125 km wide, and it is the largest contiguous mass of granitic rocks in western North America. Samples from two sections across this batholith and its country rocks in the Yakutat Bay-Skagway and Prince Rupert-Prince George areas (see Fig. 9) were studied by Magaritz and Taylor (1976b, 1976c). Some of the isotopic data obtained on these samples are presented in Figs. 10 and 11.

The 100-km long Skagway traverse includes 6 separate quartz-diorite plutons (numbered 1 through 6 from SW to NE) and a composite suite of quartz monzonite intrusions confined mainly to the NE half of the batholith (Fig. 10). Effects of meteoric-hydrothermal alteration are very obvious in the data shown in Fig. 10,

as evidenced by the widely varying δD values, the rough correlation between $\delta^{18}O$ and δD , and by the fact that very large ^{18}O -depletions are observed in some of the feldspars. Note that $\Delta^{18}O_{\text{quartz-feldspar}}$ values as large as 10 per mil are observed and all the rocks that have anomalously large $\Delta^{18}O$ values also have $\delta D < -130$.

Except possibly for one sample from pluton 3 and two samples from pluton 1, all of the rocks in the Skagway traverse have apparently interacted with meteoric-hydrothermal fluids. However, the meteoric W/R ratios were low enough in the westernmost third of the batholith so that only the δD values were markedly affected. The most extreme ^{18}O -depletions in the quartz diorite plutons occur where they are juxtaposed against quartz monzonite. Inasmuch as the δD values at these localities are also very low, this suggests that the isotopic effects are the result of convective circulation of large amounts of heated meteoric ground waters set into motion by the younger quartz monzonite intrusions.

Some other interesting aspects of the isotopic data in Fig. 10 are that many of the $\delta^{18}O_{\text{quartz}}$ values are higher than "normal", and the $\delta^{18}O$ values are commonly higher in the quartz diorites than in the quartz monzonites, opposite to the usual relationship. These data clearly require involvement of high- ^{18}O crustal rocks in the formation of the batholith in this area. If the quartz diorites formed by hybridization between quartz monzonite magmas and gabbroic rocks, either many of the quartz monzonite magmas have

been depleted in ^{18}O while still in the liquid stage, or the mafic rocks involved in the hybridization were even higher in ^{18}O than the quartz diorites are now. The only plausible source of such high- ^{18}O rocks are materials analogous to the Franciscan graywackes, ophiolites, and greenstones shown in Fig. 1.

The Prince Rupert traverse across the Coast Range batholith lies about 700 km SE of the Skagway area. For purposes of discussion the isotopic data from this traverse (Fig. 11) were grouped by Magaritz and Taylor (1976b) into 4 zones, based mainly on geographic position and geological and petrological criteria. Zone I, the western part of the batholith, is composed mainly of foliated quartz diorite, grading into a gneiss complex (Hutchison, 1967; Hollister, 1975). Zone I is separated from the relatively unfoliated, K-rich granodiorites of Zone II, the eastern part of the batholith, by a boundary analogous to the quartz diorite line defined by Moore (1959). The occurrence of sillimanite-garnet gneiss and kyanite-bearing rocks in Zone I indicates relatively deep-seated conditions in the western part of the batholith.

Zone III is a broad zone that extends from the eastern edge of the batholith for about 200 km to the Pinchi fault zone, where it is separated from the sedimentary and volcanic rocks, blueschists and serpentinites of Zone IV. The country rocks directly east of the batholith include mainly basaltic and andesitic flows, tuffs, and breccias, and graywackes, shales, and conglomerates of both continental and marine origin (Duffell and Souther, 1964). A number of epizonal, K-rich, pink biotite granites and

quartz monzonites are found in Zone III (Carr, 1965). K-Ar ages of these rocks (White et al., 1970) are 138 ± 5 m.y., considerably older than most of the plutonic rocks in the Coast Range batholith, which have U-Pb zircon ages of 84 to 66 m.y. and K-Ar biotite ages of 50 to 43 m.y. (Hutchison, 1970; G. J. Woodsworth, personal communication; Wanless et al., 1974). However, a number of porphyry and andesite dikes also cut the Zone III intrusions, and these are dated by K-Ar at 44 to 50 m.y. (Wanless et al., 1974).

The isotopic data obtained from both of the above traverses across the Coast Range batholith are basically very similar, as summarized in Table 1. A remarkable feature of the Prince Rupert section, however, is that all of the country rocks and satellitic intrusions in Zone III, the broad area to the east of the batholith, also have experienced varying degrees of deuterium depletion as well as local ^{18}O depletion (Fig. 11). Thus, convective circulation of heated meteoric ground waters was extremely pervasive in the less deep-seated eastern half of the batholith and in the country rocks for several tens of kilometers to the east.

Although the NE 2/3 of the Skagway traverse is virtually identical to Zone II of the Prince Rupert traverse, both in geologic setting and in terms of isotopic relationships, the western part of the batholith in the Skagway traverse was somewhat more heavily involved in meteoric- H_2O exchange than was Zone I, the corresponding portion of the Prince Rupert traverse. This is indicated in Table 1 by the higher meteoric water/rock

ratios, lower δD values, and slightly higher $\Delta^{18}O$ values obtained for the SW 1/3 of the Skagway traverse. These differences might perhaps be explained by the fact that Zone I in the Prince Rupert area is composed of extremely deep-seated rocks (Hollister, 1975); at the time of meteoric-hydrothermal activity, the Zone I rocks were apparently at depths too great to be affected by the relatively shallow convective circulation systems. Another difference between the two traverses is that, in contrast to the SW part of the Skagway traverse, only one of the plutonic igneous rocks from the Prince Rupert area could possibly be classified as a high- ^{18}O granite on the basis of its $\delta^{18}O_{\text{quartz}}$ value (+10.8). All the other samples have $\delta^{18}O_{\text{quartz}} < 9.6$.

Some of the quartz samples from these granitic plutons show marked ^{18}O zoning, providing a nice contrast to the quartz data from the Skye epigranite shown in Fig. 6. Quartz monzonite sample 24 from Zone III and granodiorite 67 from Zone II were subjected to a series of step-by-step partial fluorination treatments, with the results shown in Fig. 12. The quartz rims are depleted in ^{18}O relative to the cores by at least 6 to 12 per mil, with the outermost parts of the quartz grains having $\delta^{18}O = +2.5$ and -2.6 . The latter are by far the lowest $\delta^{18}O_{\text{quartz}}$ values found in any of the granitic rocks in British Columbia, approaching the low $\delta^{18}O$ values shown by some of the feldspars.

The $\delta^{18}O$ zoning in these quartz separates from British Columbia is much greater than has heretofore been demonstrated, and because of the known resistance of quartz to hydrothermal ^{18}O exchange these data imply that although alteration of these

particular rocks occurred under subsolidus conditions, this must have been at relatively high temperatures (probably $> 450^{\circ}\text{C}$).

The meteoric-hydrothermal fluids that affected the Coast Range batholith and its country rocks must have been extremely low in deuterium. Based on the lowest- δD rocks in the two traverses, a $\delta\text{D}_{\text{H}_2\text{O}} \approx -120$ can be estimated from the calibration curves of Suzuki and Epstein (1976). Thus the initial $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ was probably about -16 , and the illustrative calculations made for the hypothetical granodiorite in Fig. 8 should be directly applicable. Based on the available K-Ar age dates, the major episode of hydrothermal alteration probably occurred about 43-50 m.y. ago during the period of early Tertiary porphyry intrusion. However, some of the samples from the 140 m.y. old Zone III intrusions suggest more D-rich meteoric waters, perhaps indicating a warmer climate in the Jurassic (Magaritz and Taylor, 1976b).

The relatively high temperature meteoric-hydrothermal effects described above are not found anywhere in Zone IV, compatible with the lack of any granitic intrusions or "heat engines" in Zone IV. The meteoric H_2O interactions in the latter area thus occurred at much lower temperatures than in Zones II and III ($\approx 100^{\circ}\text{C}$ or lower), and only the finest-grained rocks underwent partial D/H exchange with the meteoric waters responsible for local silicification, serpentinization, and vein carbonate deposition (Fig. 11). These data are important, however, because they indicate that even though the Zone IV rocks must have been in contact with low-deuterium meteoric waters for tens of millions of years, many of the rocks have undergone no measureable

D/H exchange with such waters. A corollary of this, of course, is that the extreme deuterium depletions found throughout Zones II and III must have been brought about at elevated temperatures.

Boulder Batholith, Montana

The Boulder batholith (see Fig. 9) is a relatively small batholith with a number of epizonal characteristics, intruded into volcanic rocks of only slightly younger age (the Elkhorn Mtn. volcanics). It is composed principally of a single pluton, the Butte quartz monzonite. Isotopic data by Sheppard and Taylor (1974) showed that although most of the Butte quartz monzonite was essentially "normal" in δD and $\delta^{18}O$, several samples were markedly depleted in deuterium and a few others were depleted in both D and ^{18}O . These low-D and low- ^{18}O rocks are largely confined to two areas near the roof of the batholith on the western side, as shown on Fig. 13. This is the most highly mineralized and hydrothermally altered portion of the batholith, and it includes the Butte Cu-Zn-Mn ore district (Meyer et al., 1968); this important vein deposit was shown by Garlick and Epstein (1966) and Sheppard and Taylor (1974) to have been formed dominantly by meteoric-hydrothermal solutions.

Although most of the samples studied by Sheppard and Taylor (1974) were mineralogically fresh and unaltered, previously published petrographic descriptions by Becraft et al. (1963) and Ruppel (1963) indicated the existence of local variants of the Butte quartz monzonite having all the characteristics commonly found in igneous rocks that have undergone extreme ^{18}O depletion (refer back to the list given in a

previous section). Samples of these rocks were therefore specially collected by Sheppard and Taylor (1974), and, as expected, they proved to be extremely low in both ^{18}O and deuterium (see Fig. 18). In the entire Boulder batholith, only a few samples from the hydrothermal veins at Butte were found to have lower $\delta^{18}\text{O}$ values than these selected samples.

Emplacement of the Boulder batholith occurred about 78 to 68 m.y. ago (Tilling et al., 1968). However, the intense meteoric-hydrothermal alteration associated with emplacement of the Butte vein systems occurred about 5 to 10 m.y. later, and this persisted up to the period of intrusion of post Main Stage dikes at Butte (about 40 to 48 m.y. ago). Thus the meteoric-hydrothermal activity in the Boulder batholith apparently occurred over at least a 15-20 m.y. period extending in time from the latest Cretaceous to the Eocene. The $\delta\text{D}_{\text{H}_2\text{O}}$ values were estimated by Sheppard and Taylor (1974) to have been about -100 to -120, implying that the W/R ratios were typically <0.1 everywhere except for the extreme low- ^{18}O localities and the Butte vein systems.

Idaho Batholith

The Idaho batholith lies immediately west of the Boulder batholith (Fig. 14), but it is much larger and the plutons in it were intruded over a much wider range of ages (Armstrong, 1975). Although much of the batholith was emplaced in the Cretaceous or even earlier, a significant feature is the group of epizonal plutons of Eocene age.

that form a north-trending belt along the east side of the complex (Fig. 14). Smaller intrusions and dikes of early Tertiary age are apparently widespread throughout much of the Idaho batholith (Swanberg and Blackwell, 1973), but only the large, well-documented Eocene plutons are shown separately on Fig. 14. The widespread early Tertiary igneous activity had a profound effect on the K-Ar ages of some of the older portions of the batholith. In two broad zones, termed the Bitterroot Lobe in the north and the Atlanta Lobe in the south, all of the original K-Ar ages were "re-set" during these Eocene events (Armstrong, 1974). K-Ar age contours that illustrate the broad extent of these re-setting events are drawn on the map in Fig. 14.

Taylor and Magaritz (1976) carried out a reconnaissance stable isotope study of the Idaho batholith, concentrating on the Atlanta Lobe. The distribution of sample localities is shown on the map of Fig. 15, and contours are drawn to indicate how the δD values of biotites and/or hornblendes (\pm chlorite) vary throughout the Idaho batholith. Some of the isotopic data obtained on the central and southern parts of the batholith are projected onto 3 section lines in Figs. 16 and 17. The locations of the 3 sections are indicated on Fig. 14 (two E-W sections in the Atlanta Lobe and a bent section that extends from the central part of the batholith into the eastern part of the Atlanta Lobe). The data in Figs. 15, 16, and 17 show some remarkable systematics that correlate almost perfectly with Armstrong's "re-set" K-Ar age data.

The "undisturbed" central part of the Idaho batholith shows normal δD values and $\delta^{18}O_{\text{feldspar}}$ values that are about +10 or slightly higher (Fig. 16). Although not shown on Fig. 16, the $\delta^{18}O_{\text{quartz}}$ values are typically about +11 to +12. Thus, most of these granodiorites and quartz monzonites qualify as High- ^{18}O Granitic Rocks, and sialic crust must have been involved in the formation of the original magmas. This is compatible with the common occurrence of muscovite in these granitic rocks, indicating derivation from high-Al source rocks such as pelitic schists. If the Central Section is followed south into the Atlanta Lobe, significant isotopic changes occur, the first being a lowering of the δD values to about -95. The δD values also become more erratic, even though the $\delta^{18}O$ values remain about constant, just as would be expected to occur on the fringes of a meteoric-hydrothermal convective system. As one approaches the area of Eocene igneous activity and as the K-Ar age contours are crossed, the δD values plunge to a relatively uniform value of about -150 to -160, and many of the $\delta^{18}O_{\text{feldspar}}$ values also drop in a rather erratic fashion.

Note that the right-hand sides of Figs. 16 and 17 both include some of the same data points. Particular attention should be called to the closely-spaced set of 8 samples on the extreme right-hand sides of these two diagrams, as those samples were collected specifically to test the correlation between the stable isotope data and the K-Ar "ages". Over an E-W distance of 10 km the K-Ar "ages" show a rapid change from about 40-45 m.y. to about 80 m.y. or older (Armstrong, 1975), and over this same distance the δD values systematically increase from about -160 to a normal value of about -85 on the extreme eastern side (Figs. 16 and 17). Over much of this mini-traverse the $\delta^{18}O$ values are practically uniform, but they do drop abruptly to the west in the vicinity of the Eocene intrusions.

Similar effects are apparent on Fig. 17, which shows that the entire Atlanta Lobe is markedly depleted in deuterium; the only exceptions are the aforementioned easternmost samples and 3 tonalites from the west side of the batholith (the latter 3 samples occur near the boundary separating rocks with initial $^{87}\text{Sr}/^{86}\text{Sr} < 0.704$ from those with $^{87}\text{Sr}/^{86}\text{Sr} > 0.706$, Armstrong et al., 1976). Fig. 17 also shows that many of the $\delta^{18}\text{O}_{\text{feldspar}}$ values in the Atlanta Lobe are extremely erratic, and some are depleted in ^{18}O by as much as 15 per mil! However, the $\delta^{18}\text{O}_{\text{quartz}}$ values are typically about +10 to +12, only slightly lower than $\delta^{18}\text{O}_{\text{quartz}}$ in the "undisturbed" central part of the batholith; $\delta^{18}\text{O}_{\text{quartz}}$ values as low as +7 to +8 are found only in rocks in which the feldspar has been depleted in ^{18}O by several per mil. The isotopic effects shown in Fig. 17 clearly demonstrate that practically the entire Atlanta Lobe has been affected by meteoric-hydrothermal activity in which the W/R ratio was extremely variable (from < 0.1 to > 2.0). The various Eocene plutons and smaller intrusions apparently established a set of overlapping meteoric-hydrothermal convective systems that caused a pervasive isotopic re-equilibration throughout at least 5000 km^2 of the southern part of the batholith. A similar, but less intense, episode also occurred in the Bitterroot Lobe (Fig. 15), and analogous effects probably would also be found if studies were carried out in the vicinity of each of the Eocene plutons shown on Figs. 14 and 15. The paucity of such effects in the central area between the Bitterroot and Atlanta Lobes probably implies that in that region there was very little epizonal Eocene plutonic activity. It also now seems certain that the re-setting of the

K-Ar ages in the Idaho batholith was caused by the meteoric-hydrothermal activity. The stable isotope data thus nicely explain why the K-Ar effects are so widespread and why they extend so far beyond what might be considered the ordinary limits of contact metamorphism. Meteoric-hydrothermal convective circulation systems represent an extremely efficient way to rapidly transfer heat, and they therefore can produce a pervasive high-temperature alteration throughout a large body of rock.

Implications of the Isotope Data

The isotopic effects observed in the more northerly Cordilleran batholiths are best summarized on a $\delta D - \delta^{18}O$ diagram, as shown in Fig. 18. Actual data-points are given for the Idaho batholith and two low- ^{18}O rocks from the Boulder batholith, but only generalized ranges are shown for the two traverses from the Coast Range batholith. Note that most of these batholith samples have more or less normal $\delta^{18}O$ values but very few have normal δD values. Each of the traverses defines an inverted L-shaped pattern very similar to the theoretical prediction illustrated in Fig. 8. The Atlanta Lobe in particular corresponds almost perfectly to the calculations in Fig. 8, except that the $\delta^{18}O_{\text{feldspar}}$ values started out a little higher because most of the rocks from the Idaho batholith were apparently intruded as High- ^{18}O Magmas. If all the data-points from the Boulder batholith were plotted on Fig. 18, they would also show an almost perfect inverted L-shape, although there would be a gap between the two low- ^{18}O rocks and the others. Even this gap could be filled in, however, if

selected samples from the Butte ore deposit were added to the diagram.

Although the exact δD values of the meteoric waters are not known for the various localities shown in Fig. 18, a $\delta D_{H_2O} = -120 \pm 10$ fits most of the data reasonably well. One might well ask why the calculated δD values appear to be so uniform in all of these areas, particularly since the δD_{H_2O} values could easily have varied by at least 20 to 30 per mil or more during the long history of plutonism from the Jurassic to the mid-Tertiary. The answer probably lies in the fact that:

- (1) The Idaho and Boulder batholiths are much farther from the edge of the continent than the Coast Range batholith of B.C.; therefore, just as is true at present, the decrease in δD of the ancient meteoric waters with increasing latitude should have been partially compensated for by the topographic effects.
- (2) Even though meteoric-hydrothermal circulation systems probably operated throughout the long history of intrusive activity in these batholiths, at any given stage these effects would be confined to within 5-10 km of the Earth's surface; thus any evidence of older meteoric-hydrothermal events would tend to be erased, because not only does erosion continuously and selectively remove the epizonal portions of the batholiths, but the later hydrothermal events also cause isotopic re-equilibration and destruction of the previous record.

The above line of reasoning probably accounts for why the hydrothermally altered rocks in these batholiths consistently give Early Tertiary K-Ar ages, because in most

areas this seems to represent the last major plutonic episode. Because of repeated intrusion and re-intrusion, the isotopic evidence of the older hydrothermal events will generally be obscure, and probably can only be obtained by doing very detailed isotopic studies, or by examining outlying plutons intruded well away from the main axis of batholithic emplacement (as for example, the Jurassic plutons in Zone III of the Prince Rupert-Prince George traverse).

Fig. 19 shows a histogram comparing the data given above with all of the available hydrogen isotope data from the Sierra Nevada and Southern California batholiths. There is a clear-cut trend shown in Fig. 19, in that the more northerly batholiths contain a much higher proportion of low-deuterium rocks. This belt must be mainly due to the progressive latitudinal decrease in the δD of meteoric waters as one moves north. For example, we know that extensive areas of low- ^{18}O rocks occur in the extreme NW part of the Southern California batholith in the vicinity of the Corona micropegmatite granite (Larsen, 1948). There has thus been appreciable meteoric-hydrothermal alteration of this epizonal pluton and the surrounding metavolcanic country rocks that form the roof of the batholith. However, the δD values in all these rocks are higher than -115, much higher than are found in the low- ^{18}O rocks of the more northerly batholiths.

The above simple interpretation of the data in Fig. 19 is complicated by some other factors, the chief of which is that the Tertiary history of the Sierra Nevada and Southern California batholiths appears to have been considerably different from that of

the more northerly batholiths.

Whereas Early Tertiary K-Ar age dates are the rule in the northern batholiths, they are virtually non-existent in the two southern batholiths (Kistler, 1974; Krummenacher et al., 1975). In British Columbia and Idaho the plutonic igneous activity apparently continued along a relatively narrow axis from the Mesozoic well into the Tertiary (e.g. see Pitcher, 1975), whereas in the Sierra Nevada and Southern California batholiths this activity largely ceased in the Late Cretaceous. South of latitude 42°N, where the Cordilleran orogenic belt is exceptionally wide, most of the Tertiary intrusive activity occurred well east of the main axis of the Cordilleran batholiths. Tertiary epizonal intrusions are abundant throughout the Basin and Range Province and the southern Rocky Mtns., and were indeed responsible for widespread meteoric-hydrothermal effects (O'Neil and Silberman, 1974; Hall et al., 1974; Taylor, 1973, 1974b; Sheppard et al., 1969, 1971); however, these distant intrusions could not, of course, affect the major batholiths.

The above discussion thus suggests that the isotopic patterns shown in Fig. 19 may in part be due to the absence of strong Tertiary plutonism in the more southerly Cordilleran batholiths. For example, because of climatic shifts, the δD values of meteoric waters might have been much higher in the Cretaceous than in the Tertiary. Also, if the erosion rates have been at all similar, a much greater fraction of the epizonal portions of the batholiths would have been stripped off the two southerly

occurrences. This hypothesis is supported by the general paucity of low- ^{18}O granitic rocks in the Sierra Nevada and Southern California batholiths, although it needs to be qualified because less extensive $^{18}\text{O}/^{16}\text{O}$ data are available for these two batholiths than for the others.

4. PRECAMBRIAN GRANITIC BATHOLITHS

Isotopic Compositions of Ancient Surface Waters

The previous discussion has shown that the epizonal portions of granitic batholiths commonly show widespread D/H and $^{18}\text{O}/^{16}\text{O}$ effects attributable to interactions with meteoric-hydrothermal fluids. Therefore, in addition to our intrinsic interest in such rocks, detailed studies of Precambrian batholiths of various ages should allow us to calculate the isotopic compositions of meteoric waters throughout geologic time. This should also allow us to infer something about the isotopic evolution of ocean water on the Earth, inasmuch as it is by far the major component of the hydrosphere. Under a given set of climatic conditions, the isotopic composition of ocean water will control the δD and $\delta^{18}\text{O}$ values of all waters involved in the meteorological cycle.

In determining the isotopic compositions of ancient surface waters, studies of igneous and contact metamorphic rocks have many advantages over studies of authigenic and biogenic sedimentary rocks such as cherts, carbonates, and iron formations (Perry, 1967; Knauth and Epstein, 1976; Becker and Clayton, 1976). The chief advantage,

particularly in the case of the more easily exchanged D/H ratios of the hydroxyl minerals, is that the above sedimentary rocks are all exceedingly fine-grained and therefore much more susceptible to later exchange and recrystallization. In contrast, once a plutonic-igneous complex has cooled off and the meteoric-hydrothermal convection systems have shut down, these much coarser grained hydrous minerals and silicates have a better chance of preserving their isotopic compositions.

Seychelles Batholith, Indian Ocean (650 - 670 m.y.)

The Seychelles Islands are unique among oceanic areas in that they are almost wholly composed of granite, presumably a remnant of the break-up of the continent of Gondwanaland left isolated between Africa and India in the middle of the Indian Ocean. The main island of Mahé (Fig. 20) is made up of a remarkably homogeneous, leucocratic, hypersolvus, coarse-grained hornblende microperthite granite (Baker, 1963). This granite is latest Precambrian in age, having been dated at 650 to 670 m.y. by K-Ar and Rb-Sr (Miller and Mudie, 1961; Wasserburg et al., 1964; Michot and Deutsch, 1976). No remnants of older country rocks have been found anywhere on the island, except for a few small xenoliths in the granite.

From the point of view of stable isotope studies, the Seychelles batholith is principally of interest because of its age and the fact that a single sample of the Mahé granite just by chance happened to be the first low-¹⁸O granite to be identified (Taylor, 1968). Because of these unique features, a more detailed isotopic study of the

Seychelles was undertaken (Taylor, 1974c); some of these new data are summarized in Fig. 20.

Except for two granite porphyry intrusions on the west side of the island, practically all of Mahé is made up of a single granite pluton that is just as homogeneous in $^{18}\text{O}/^{16}\text{O}$ and D/H as it is in terms of texture and mineralogy. Only along the northwestern and eastern coasts were any isotopically different granite samples found (Fig. 20), and these samples are also texturally and mineralogically distinct as well. Whereas the main body of granite is absolutely fresh and contains clear, homogeneous microperthite, the anomalous coastal samples typically exhibit somewhat turbid, patchy perthite that has undergone very uneven exsolution. They also show sporadic development of biotite, and hydrothermal zones containing epidote and other alteration features are also locally present; in addition, whereas amphibolite xenoliths are exceedingly rare everywhere on Mahé, they are most common in the northern coastal outcrops, suggesting proximity to the contact and possibly indicating that the present outline of the island may closely approximate the original dimensions of the Mahé pluton.

The data in Fig. 20 leave no doubt that the main Mahé pluton crystallized from a very homogeneous, low- ^{18}O granitic magma having a $\delta^{18}\text{O} = +3.3 \pm 0.3$. This is proven by comparing the Seychelles data with, for example, the data in Fig. 5 for the Skye granites, because: (1) the striking isotopic homogeneity in the Mahé pluton could not possibly have been produced by later sub-solidus hydrothermal exchange, and (2) the

$\Delta^{18}\text{O}_{\text{quartz-feldspar}}$ values are also extremely uniform at 1.25 to 1.50, identical to the equilibrium values in "normal" granites. The $^{18}\text{O}/^{16}\text{O}$ data, together with the total lack of any mineralogical or textural alteration features in the main body of granite, thus also suggest that the δD values of the hornblendes (-102 to -109) are primary as well. This would imply that the granitic magma may have had a $\delta\text{D} \approx -105$, although this must be qualified somewhat because if there is any escape of magmatic H_2O during the latest stages of crystallization it could have a more dramatic effect on the δD of the crystalline granite than on its $\delta^{18}\text{O}$ value.

The Mahé granite pluton therefore must have crystallized from a very homogeneous, low- ^{18}O , low-deuterium magma, but the next question to be considered is how was this magma formed? Inasmuch as both the $\delta^{18}\text{O}$ and the D are lower than "normal", the most logical explanation is that the parent magma underwent some kind of indirect interaction with Precambrian meteoric ground waters. The extreme ^{18}O depletions and the very large $\Delta^{18}\text{O}_{\text{quartz-feldspar}}$ values (up to 3.5) in the northern coastal samples (see Fig. 20) leave no doubt that these marginal portions of the granitic complex underwent sub-solidus meteoric-hydrothermal alteration of some type. An earlier meteoric-hydrothermal event in the country rocks is also confirmed by the occurrence of amphibolite xenoliths in some of these coastal granite samples that are even lower in ^{18}O than the low- ^{18}O granite host rock. In addition, one of the granite samples from Praslin, the second largest island of the Seychelles Group (50 km NE of Mahé), contains very turbid alkali feldspar with a $\delta^{18}\text{O} = -1.6$; this value is markedly lower than in the rest of the granites from Praslin, which have a rather homogeneous $\delta^{18}\text{O}_{\text{feldspar}} \approx +6.9$, distinctly higher than in

the Mahé granite. This altered sample also has a very large $\Delta^{18}\text{O}_{\text{quartz-feldspar}}$ (+8.6). All these data clearly document a hydrothermal event involving waters with an initial $\delta^{18}\text{O}$ probably at least as low as -5.

Referring back to the previous discussion of low-deuterium magmas, it is significant that the δD values of the main Mahé pluton are slightly lower than in the hydrothermally altered marginal granites. This means that the δD of the Mahé granite magma could not have been the result of simple inward diffusion of meteoric water directly into the magma, because the meteoric-hydrothermal fluids in equilibrium with the altered marginal granites should have had a δD at least as high as -50 (Suzuoki and Epstein, 1976; also see Fig. 8 above). Thus the low δD and $\delta^{18}\text{O}$ values of the Seychelles granites are best interpreted as having arisen through exchange, assimilation, or partial melting of previously hydrothermally altered country rocks. Some type of thorough mixing process must then have operated on the magma prior to its intrusion. Applying the meteoric water equation, the ground waters that caused these effects about 650 m.y. ago apparently had $\delta\text{D} \approx -40$ to -50 and $\delta^{18}\text{O} \approx -6$ to -7 , very typical of values in warm temperate climates at the present time (e.g. the southern United States, see Friedman et al., 1964). These data imply that the surface waters on the Earth 650 m.y. ago could not have been much different in δD or $\delta^{18}\text{O}$ than they are today.

St. Francois Mtns. Batholith, Missouri (1400-1500 m.y.)

The 1400-1500 m.y. old Precambrian rocks of the St. Francois Mtns. (Bickford and Mose, 1975) are mainly rhyolitic ash-flow tuffs (Anderson, 1970) intruded by two groups

of plutonic granites (see Fig. 21). This area has a number of features analogous to the extreme low- ^{18}O Tertiary volcanic-plutonic complexes discussed previously. However, in contrast to the Tertiary examples, the St. Francois rocks exhibit a pervasive brick-red alteration of the feldspars that increases in intensity from the NE to the SW in Fig. 21. Wenner and Taylor (1976) concluded that this terrane underwent a meteoric-hydrothermal alteration event at the time of granitic intrusion, but that later on it also suffered a long period of exposure to low-temperature hydrothermal fluids that produced the brick-red alteration.

The most clear-cut isotopic evidence that a low-temperature hydrothermal alteration event has affected an igneous rock is: (1) a reversed (i.e. negative) $\Delta^{18}\text{O}_{\text{quartz-feldspar}}$ value, or (2) a lower $\delta^{18}\text{O}$ value in the quartz phenocrysts than in the coexisting glassy or aphanitic groundmass. Both features are present throughout the St. Francois Mtns. terrane, as well as in the Duluth-Keweenaw granophyre-rhyolite complex of Minnesota and Michigan, the Bushveld red granites and felsites in South Africa, and the granophyre associated with the Muskox intrusion of northern Canada (Taylor, 1968, 1974b; Wenner and Taylor, 1976). In all these areas, the igneous rocks also commonly contain brick-red, turbid feldspars that are full of finely disseminated hematite dust.

Some of the isotopic data obtained by Wenner and Taylor (1976) are given in Fig. 21. The $\Delta^{18}\text{O}_{\text{quartz-feldspar}}$ values are contoured and show a systematic variation ranging

from near-equilibrium values in the northeast, to negative values that are clearly disequilibrium in the southwest. The whole-rock $\delta^{18}\text{O}$ values are not shown on Fig. 21, but they correlate very well with the $\Delta^{18}\text{O}_{\text{quartz-feldspar}}$ values, systematically increasing from about +7 in the NE to about +14 in the SW. This correlation is mainly due to the fact that most of the ^{18}O variations are brought about by variations in the feldspars. The quartz is essentially identical to quartz in "normal" granites ($\delta^{18}\text{O} = 8.8$ to 10.6). In general, the coarsest-grained minerals have the most "normal" $\delta^{18}\text{O}$ values, presumably because of their greater resistance to exchange.

The regional $\delta^{18}\text{O}$ effects in the St. Francois Mtns. are not only in the opposite direction to those found in most Tertiary epizonal intrusions, but they also show no resemblance to the typically concentric $\delta^{18}\text{O}$ contours observed around the Tertiary complexes (see Fig. 3). Instead, the $\delta^{18}\text{O}$ and $\Delta^{18}\text{O}$ contours in the St. Francois Mtns. are essentially independent of the position of the plutons, except in the general sense that the lowest $\delta^{18}\text{O}$ values occur to the northeast where the greatest density of granitic intrusions occurs. The most likely interpretation of the regional increase in $\delta^{18}\text{O}$ to the SW and of the $\Delta^{18}\text{O}$ contours in Fig. 21 is that the feldspar $\delta^{18}\text{O}$ values simply increase upward in the volcanic section, inasmuch as the entire area was apparently tilted subsequent to the alteration event. This upward increase in the $\delta^{18}\text{O}$ of the alkali feldspar could be attributed to a regional downward increase in the temperature of hydrothermal alteration (perhaps along a slightly enhanced geothermal gradient), or

an upward increase in the water/rock ratio or in the extent of ^{18}O exchange.

Wenner and Taylor (1976) calculated $\delta^{18}\text{O}$ values (0 to -6) and δD values (0 to -25) of the low-temperature hydrothermal fluids. These values are higher than in most present-day meteoric waters, but they are similar to modern meteoric waters found in warm, tropical climates at low latitudes and low elevations. They are also similar to many of the saline formation waters in the Paleozoic rocks of the mid-continent region of the U.S. (Clayton et al., 1966). Thus, even going back almost a billion years earlier than the emplacement of the Seychelles granites, the isotopic compositions of surface waters on Earth still seem to have been basically similar to those of present-day waters.

Although the St. Francois terrane is geologically similar to many low- ^{18}O Tertiary volcanic-plutonic complexes that have interacted with meteoric-hydrothermal fluids at high temperatures, only at a single locality (indicated by the large square near the top of Fig. 21) is any low- ^{18}O quartz or feldspar found ($\delta^{18}\text{O}_{\text{quartz}} = 4.6$ to 5.4). Even at this locality, however, the feldspars were later enriched in ^{18}O by the same alteration event that affected the rest of the region. A possible reason why the St. Francois feldspars might have been so susceptible to the later, low-temperature episode is provided by their prior alteration history. The local occurrence of the low- ^{18}O rocks noted above is probably a relict of a much more widespread hydrothermal event, and O'Neil and Taylor (1967) showed experimentally that alkali feldspars that

had previously been subjected to ^{18}O and cation exchange under hydrothermal conditions are transformed to a much more reactive condition. Such pre-treated feldspars are turbid and full of defects, dislocations, and fluid inclusions. If they are later exchanged during a second hydrothermal episode, the exchange should occur at enormously enhanced rates.

The exchange process which produced the brick-red K feldspars in many Precambrian terranes thus probably required long periods of exposure to circulating waters at low temperatures, and it should be most pronounced in those rocks that have previously been subjected to a high-temperature hydrothermal event. The low- ^{18}O Tertiary igneous complexes such as Skye and Mull typically do not contain these brick-red feldspars, probably because these Tertiary rocks have not yet had a chance to be buried by younger rocks and exposed to circulating waters at low temperatures for a sufficient length of time.

A study by Bickford and Mose (1975) of U-Pb in zircons from the St. Francois Mtns. indicated that virtually all of the major igneous events occurred during a very short interval of time, about 1500 m.y. ago; the only exception is a single granite porphyry body in the SW part of Fig. 21, dated at 1400 m.y. However, Bickford and Mose found that in contrast to the systematic picture presented by the U-Pb ages, the whole rock Rb-Sr "ages" are much more variable, ranging from 1408 to 1273 m.y. The inconsistencies represented by these U-Pb and Rb-Sr whole rock ages are best

interpreted as being caused by the low-temperature, pervasive hydrothermal alteration event that produced the regional $\delta^{18}\text{O}$ effects (Taylor, 1974b; Wenner and Taylor, 1976).

From the relationship between Rb-Sr ages and the chemistry of the volcanic rock samples, Bickford and Mose (1975) noted a general positive correlation between the model ages (based on an assumed initial $^{87}\text{Sr}/^{86}\text{Sr} = 0.705$) and Sr concentrations; they interpreted this feature as being due to widespread Sr loss. If similar amounts of Sr were removed from all samples, this would most strongly affect the Rb-Sr ages of those samples that contain low total strontium concentrations. Samples with high Sr concentrations would obviously be less affected by Sr loss and hence exhibit a less "disturbed" age.

Note that the concentration of Sr as well as the $^{87}\text{Sr}/^{86}\text{Sr}$ in the rocks would almost certainly be disrupted during exchange with hydrothermal solutions, because the Sr is largely concentrated in the feldspars, and the oxygen in the feldspars has been totally reconstituted. Such waters may have had a low Sr concentration, thus allowing Sr loss to the aqueous phase, or they could have had a different $^{87}\text{Sr}/^{86}\text{Sr}$, thus causing a change in this parameter. The latter process could also produce a correlation between Rb-Sr "age" and Sr concentration, in the manner suggested by Taylor and Forester (1971) for the early Tertiary intrusive complex at Skye.

The "anomalous" type of Rb/Sr behavior found in the rocks of the St. Francois Mtns. should also be characteristic of all similar rocks in other areas. Thus, any rocks that

contain turbid, brick-red feldspars and show evidence of ^{18}O exchange with low-temperature hydrothermal solutions should be suspect; it is doubtful that Rb/Sr techniques can be used to obtain the original emplacement ages of such rocks, regardless of how well preserved their original structures and textures appear to be.

Swaziland and Barberton Areas, South Africa (>3.3 b.y. to 2.6 b.y.)

Moving considerably further back in time, Taylor and Magaritz (1975) made an isotopic study of some of the oldest, best-preserved granitic batholiths on Earth, the 2.6 to >3.3 b.y. old basement complex (Fig. 22) exposed in the Transvaal craton of southern Africa (Engel, 1968; Viljoen and Viljoen, 1969a; Hunter, 1971). The ancient Gneiss Complex of Swaziland (Hunter, 1970) is probably as old as 3.6 b.y., by analogy with the Mashaba tonalitic gneisses of Rhodesia (Hawkesworth *et al.*, 1975), and because a Rb-Sr mineral isochron age of 3500 ± 200 m.y. was determined for one of the lower stratigraphic horizons in the probably younger Swaziland Sequence of the Barberton greenstone belt (Jahn and Shih, 1974).

There are several important features indicated in Fig. 23: (1) Most of these ancient granitic rocks plot within the "normal" range of $\delta^{18}\text{O}$ and δD established for much younger granites; however, almost all of them plot near the low- ^{18}O end of the "normal" range and none has a $\delta^{18}\text{O} > +9.0$, except for a couple of apparent metasedimentary gneisses from the Ancient Gneiss Complex. (2) Many of the samples are partially chloritized, and the most chlorite-rich samples have a relatively uniform $\delta\text{D} \approx -40$ to -50 ; most of the

D/H data in Fig. 23 can be explained in terms of a simple two-component mixture between such a chlorite and a less deuterium-rich, primary biotite or hornblende.

(3) The very fresh, unaltered biotites and hornblendes of the Ancient Gneiss Complex of Swaziland are distinctly depleted in deuterium relative to the "normal" range. (4) Just as is the case for the much younger batholiths described previously, low- ^{18}O granitic rocks can be found locally within this ancient craton; carrying the analogy even farther, these ancient low- ^{18}O rocks also exhibit large $\Delta^{18}\text{O}_{\text{quartz-feldspar}}$ values and are typically associated with the youngest plutons in the batholithic complex, namely the 2.8 b.y. old Usushwana gabbro-granophyre complex and the 2.6 - 2.7 b.y. old "young" granites. The low- ^{18}O effects are particularly prominent in the vicinity of localities 9, 11, and 19 shown on Fig. 22. (5) Although the data are not presented in the present paper, similar isotopic relationships to those described above were also found in the Rhodesian craton (Taylor and Magaritz, 1975).

Much of the chloritization referred to above was probably caused by meteoric-hydrothermal fluids (e.g. the altered Hood granites, Usushwana granophyre, and the lowest- ^{18}O "young" granite), but the most pervasive effects are best attributed to a regional low-temperature alteration event somewhat analogous to that described above for the St. Francois Mtns., but not intense enough to significantly shift the $\delta^{18}\text{O}$ of the feldspar or to produce a deep, brick-red alteration. These regional alteration effects apparently become less intense with depth, because the mafic minerals in the

deeply eroded Swaziland gneisses are extremely fresh and unaltered, in sharp contrast to the tonalite domes, Nelspruit migmatites, and large parts of the Hood granite. In any case, the uniformity of the δD values in all these different chlorites suggests that regardless of temperature, they were all formed from, or exchanged with, surface-derived waters with a similar $\delta D \approx -10$ to -20 . This range of δD_{H_2O} values can be found today in meteoric waters in equatorial areas, and thus these data are perfectly compatible with an early Precambrian ocean having $\delta D \approx 0$, if the climate was warmer than the present-day climate.

The absence of anything approaching a High- ^{18}O Granite in this ancient craton is compatible with these various granitic rocks having been mainly formed either from primordial low- ^{18}O sialic crust or by differentiation directly from the upper mantle. The $\delta^{18}O$ values of most of these ancient granites thus contrast strongly with those of the Cordilleran granitic batholiths of Mesozoic age. These old granites definitely did not form by anatexis of high- ^{18}O metasedimentary rocks, probably because 3 b.y. ago there was very little high- ^{18}O metasedimentary crust available to be melted.

The Ancient Gneiss Complex does contain a few clear-cut metasedimentary units, but it is dominantly made up of trondhjemites, tonalites and granodiorites of probable igneous origin (Hunter, 1970). A dominantly igneous origin is certainly supported by the generally low $\delta^{18}O$ values, but what are we to make of the anomalously low δD values shown in Fig. 23? The latter clearly cannot be explained away by meteoric-hydrothermal

effects, as can be done for almost all such examples of younger, low-D igneous rocks. One possibility suggested by Taylor and Magaritz (1975) is that the H_2O in these ancient gneisses was in fact derived from "juvenile" water; the presence of such a H_2O could be accounted for if these ancient granitic magmas formed from an upper mantle that had not yet been heavily contaminated with hydroxyl minerals from subducted oceanic lithosphere. This is compatible with most models of the development of the early crust, which do not envision the operation of sea-floor spreading and subduction in anything like its present-day form (e.g. Anhaeusser, 1973; Engel, 1968).

Other models than those invoking a low-D "juvenile" water are also possible, but they are too complex to be taken up in this short discussion. However, we should emphasize that there is no a priori way to predict what the δD value of "juvenile" water ought to be, and if the upper mantle has over the course of geologic time been heavily contaminated with subducted hydroxyl minerals, there is also no possibility of ever directly observing "juvenile" water, except perhaps in (1) the oldest igneous rocks on Earth, which may have formed before the D/H ratio of any original H_2O in the upper mantle was changed by addition of subducted hydroxyl minerals, or (2) the magmas or igneous rocks derived from plumes or hot-spots that possibly originate in the deep mantle, below the zone of heavy contamination by subducted H_2O .

There are great difficulties connected with observing "juvenile" water in either of the above environments. First, the ancient rocks may not have preserved their original δD values; however, at least for the Swaziland gneisses, this does not seem to be a problem, because the hornblendes and biotites in these ancient rocks seem to have preserved their low δD values even though nearby there has been widespread later hydrothermal chloritization by relatively deuterium-rich fluids. Second, any such deep-mantle magma plume must preserve its distinctive δD value during passage through the upper mantle and lower crust, both of which are major potential sources of contamination by "normal" magmatic or metamorphic H_2O .

Many more extensive D/H studies must be undertaken before the above questions can be sorted out, if ever. Nonetheless, besides the data from the ancient gneisses, there is some possible supporting evidence for a relatively low δD value for "juvenile" water. Paradoxically, this supporting evidence comes from D/H data on some younger igneous rocks whose δD values fall below the "normal" range (i.e. below $\delta D = -85$), but which exhibit no evidence that meteoric fluids were involved in their emplacement or subsequent cooling history. As might be expected from the previous discussion, at the present time examples of such rocks are extremely rare. However, they do exist, and there is at present no obvious way to explain their low δD values other than by appealing to an unknown deep-seated source. Examples are the Precambrian Pikes Peak granite-syenite complex of Colorado (Barker and O'Neil, 1975), as well as the Bosmankop syenite and two relatively unaltered samples

of the 2.6 b.y. old Swaziland granites shown in Fig. 23. Considering the paucity of data, there is, of course, some probability that all of the D/H variations discussed above are essentially random, and that the so-called "normal" range of D/H in granitic rocks should be extended all the way from -50 to perhaps -110 or even lower. However, arguing against this is the fact that the δD range of subducted marine hydroxyl minerals is now very well established (see Fig. 1), and if this is what basically controls the "normal" δD values of most igneous rocks, then the H_2O in any rock with an appreciably lower δD value must come from a basically different source. One final qualification must be added to the above statement, namely that ocean water has remained essentially constant in δD throughout most of geologic time. Obviously, if the δD of ocean water was different at any period in the Earth's history, then the δD of any hydroxyl minerals subducted during that period will also show a similar per mil shift.

Isotopic Evolution of Ocean Water

This subject is too involved to be treated in any detail in this discussion. However, summing up the evidence presented above, it now seems clear that no drastic isotopic changes could have occurred in the surface waters on Earth, at least over the past 2,500 m.y. Meteoric-hydrothermal effects involving surface-derived waters have been identified at about 650 m.y., at about 1300-1500 m.y., and at about 2500 m.y.; no unusual isotopic differences were observed in any of these time-periods, except

perhaps that the early Precambrian climates were generally warmer than the present. This implies that ocean water must also have been relatively uniform in δD and $\delta^{18}O$ during the past 2500 m.y. There is certainly no evidence for a much lower $\delta^{18}O$ value in the ancient oceans, as proposed by Perry (1967) and Perry and Tan (1972). The relatively low $\delta^{18}O$ values of extremely ancient cherts are thus better interpreted as due to later metamorphic exchange or to a higher temperature in the Precambrian oceans (Oskvarek and Perry, 1976; Knauth and Epstein, 1976). In fact, the latter concept fits in nicely with the interpretation of warmer Precambrian climates indicated above, as well as with the fact that the surface heat flow and radioactive heat production also must have been much higher in the early Precambrian.

This line of reasoning has a myriad of important consequences, among which are:

(1) If the δD of juvenile water was relatively low, perhaps about -100 or lower, as suggested above, why do the oceans now have a $\delta D = 0$? There are two obvious

possibilities that need to be considered in subsequent studies. Most likely, there may have been a major kinetic isotopic fractionation involving loss of mass-1 hydrogen relative to deuterium from the Earth's atmosphere during the postulated early high-temperature outgassing. This would be favored by high temperatures and intense radiation, particularly if the atmosphere was strongly reducing and contained appreciable CH_4 and NH_3 that could undergo dissociation, or H_2 gas which at equilibrium would be extremely depleted in deuterium. Contrary to the suggestion of Kokubu et al. (1961), any such preferential loss of mass-1 hydrogen from the

upper atmosphere would have had to cease very early in the Earth's history, because there is no evidence for a steady increase in the δD of ocean water with time during the Precambrian. In fact, if anything the data presented above suggest that the early Precambrian surface waters may have been slightly richer in deuterium than they are now. A less likely possibility is that there has actually been a massive net return of H_2O in the form of relatively low- δD hydroxyl minerals to the Earth's mantle by subduction processes; because of the 30 to 50 per mil fractionation between water and OH-bearing minerals, by material balance the H_2O remaining on the surface would have to become enriched in deuterium. (2) If magmatic or juvenile H_2O was outgassed at high temperatures early in the Earth's history, it should have been in equilibrium with mantle-derived magmas and thus should have had a $\delta^{18}O \approx +5$ to $+6$; if so, why should it now have a $\delta^{18}O = 0$? This problem has recently been treated by Muehlenbachs and Clayton (1976), who have made material-balance estimates of the amounts of high- ^{18}O rocks formed by submarine weathering and the amounts of low- ^{18}O rocks formed by hydrothermal circulation of ocean water at a mid-ocean ridge spreading center. They were able to demonstrate that the $\delta^{18}O$ value of ocean water is probably "buffered" at a value of about zero by the above processes, and that it would only require about 200 m.y. of sea-floor spreading for the ocean to be changed from $\delta^{18}O = +6$ to $\delta^{18}O = 0$. In conjunction with the previous arguments that the $\delta^{18}O$ of the ocean has remained roughly constant for the past 2.5 b.y., this would therefore imply that subduction processes may have begun on Earth as early as 3.0 b.y. ago.

(3) All of the isotopic shifts discussed above had to occur very early in the Earth's history, inasmuch as there have apparently been no significant isotopic changes in the oceans during the last 2500 m.y.; this therefore implies that the oceans and hydrosphere had to form in their entirety very early in the Earth's history during a massive outgassing. Such a conclusion is in conflict with the ideas presented in a widely-quoted paper by Rubey (1951), but it is in agreement with the arguments put forth by Fanale (1971) concerning this problem.

5. CONCLUSIONS

As far as the present writer is concerned, probably the most important and unique kind of information that has been derived from stable isotope studies of igneous rocks is the concept that many magmas may be strongly affected by widespread melting or assimilation of hydrothermally altered roof rocks above a magma chamber. In most volcanic-plutonic complexes, the volcanic rocks typically have chemical and isotopic properties that are nearly identical to the plutonic magmas which intrude them, because all these igneous rocks commonly have similar ages and are often genetically related. Thus, in such areas there are very few geochemical criteria that can distinguish between primary magmas and magmas formed by melting or large-scale assimilation of roof rocks. However, the $\delta^{18}\text{O}$ and δD values of the hydrothermally altered volcanic rocks are practically always markedly different from the values in primary magmas of deep-seated origin, thus making the above kinds of distinctions very obvious.

Composite volcanic-plutonic terranes represent an ideal situation for developing a variety of permutations and combinations of magma-rock-water interactions, because: (1) such complexes typically display favorable structural features such as strong jointing, ring-fracturing, caldera collapse features, explosive brecciation, resurgent doming, radial fracturing, and cone-sheet formation; and (2) the hydrothermally altered rocks are usually full of minerals such as turbid, readily-exchangeable feldspars, clays, and chlorites, and hence the rocks are water-rich and very reactive. When a new batch of magma intrudes this kind of environment, there is certain to be large-scale stoping and assimilation of roof-rocks, with attendant dehydration; this can cause the upper parts of the magma chamber to become saturated with H_2O . In this water-rich environment, partial melting and isotopic exchange between the country rocks and the magma should be an important process, resulting in the formation of low- ^{18}O magmas.

Localities where low- ^{18}O magmas have been extensively developed, as for example is the case for many young tholeiitic basaltic lavas on Iceland (Muehlenbachs et al., 1974), ought to be geochemically investigated more thoroughly with the above concept in mind. Even though only the $\delta^{18}O$ and δD effects will be readily measurable, such processes are probably also accompanied by exchange of other isotopic species, as well as certain cations (e.g. Na and K). This conceivably could produce subtle variations in other geochemical parameters (e.g. trace element contents, radiogenic isotopes, etc.).

Although the meteoric-hydrothermal systems will not appreciably change the chemistry of a hydrothermally altered volcanic section (aside from the greatly increased H_2O

contents and changed $\delta^{18}\text{O}$ and δD values), in analogous submarine volcanic-plutonic complexes there surely should be some profound chemical effects as well. For large water/rock ratios, the much more saline, heated ocean waters can radically affect the chemical compositions of the rocks with which they exchange, notably causing a drastic increase of the Na/K and Na/Ca ratios in spilitized basalts, for example. Thus, if the same processes of partial melting and assimilation of hydrothermally altered roof rocks take place in oceanic volcanic-plutonic complexes, the consequences should be apparent even without the aid of stable isotope studies. Favorable conditions are so clearly present at mid-ocean ridge spreading centers and in ophiolite complexes that there can be little doubt that magmas analogous to the low- ^{18}O magmas of continental volcanic-plutonic complexes should be developed on at least a local scale. The obvious candidates in the marine environments are the strongly sodic igneous rocks such as oceanic trondhjemites and quartz keratophyres; these are probably the equivalents of the potassic granophyres of continental volcanic-plutonic complexes. Thus, the markedly contrasting chemical compositions of igneous rocks in continental environments versus those from oceanic or island-arc environments (e.g. see Coleman and Peterman, 1975) may simply be due to a complicated series of overlapping processes that include submarine volcanism, emplacement of a sub-volcanic magma chamber or sheeted-dike complex, which in turn leads to hydrothermal alteration (spilitization) by convectively circulating ocean waters, followed by large-scale assimilation and/or

partial melting of the Na-rich, hydrothermally altered roof rocks above the major magma chambers. Note that in addition to the SiO_2 -rich partial melts, such submarine processes could also lead to spilitic basalt magmas; thus, trondhjemites conceivably might also form by fractional crystallization of such "contaminated" basaltic magmas.

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Table 1. Comparison of the isotopic data obtained along two sample traverses, 700 km apart, across the Coast Range batholith of British Columbia. (After Magaritz and Taylor, 1976b, 1976c)

	Zone I of Prince Rupert Traverse	SW 1/3 of Skagway Traverse	Zone II of Prince Rupert Traverse	NE 2/3 of Skagway Traverse
Number of Samples	7	8	12	12
<u>$\delta^{18}\text{O}$ quartz</u>				
Average	8.4	10.9	7.6	8.8
Range	7.8 to 9.6	9.2 to 11.5	3.1 to 10.8	6.0 to 10.9
<u>$\delta^{18}\text{O}$ feldspar</u>				
Average	7.8	8.9	4.4	5.2
Range	6.8 to 8.9	7.5 to 10.3	-2.9 to 7.3	-4.0 to 10.0
<u>$\Delta^{18}\text{O}$ quartz-feldspar</u>				
Average	1.2	1.8	3.3	3.7
Range	0.1 to 1.9	1.0 to 2.8	1.3 to 8.1	0.4 to 10.5
<u>δD mafic minerals</u>				
Average	-79	-106	-143	-137
Range	-68 to -95	-79 to -144	-106 to -165	-115 to -167
<u>Water/rock ratio</u>				
Average	.001	.02	.2	.4
Range	.0007 to .002	.000 to .03	.03 to 2.4	0.3 to 1.4

FIGURE CAPTIONS

- Figure 1. Plot of δD vs. $\delta^{18}O$ for volcanic rocks, gabbros, trondhjemites, and various vein minerals and sedimentary rocks from the Franciscan formation, California, and two samples from the Troodos ophiolite, Cyprus (data from Magaritz and Taylor, 1974; 1976a; Wenner and Taylor, 1974). The numbers in the upper right corner indicate the latitudes ($^{\circ} N$) of the sample localities in California. Note that some of the chert samples are low in deuterium because they contain extremely Fe-rich hydrous minerals (see Suzuoki and Epstein, 1976).
- Figure 2. Generalized geologic map of western Scotland (Stewart, 1965), showing the islands of the Inner Hebrides, and the approximate outer boundaries of low- ^{18}O rocks at 3 Tertiary plutonic centers (after Taylor and Forester, 1971; Forester and Taylor, 1976a; 1976b).
- Figure 3. Generalized geologic map of part of the Isle of Skye (see Stewart, 1965), showing sample localities and contours of whole-rock $\delta^{18}O$ values for all basalts, agglomerates, tuffs, and pre-Red Hills basaltic dikes, based on the data of Forester and Taylor (1976a).
- Figure 4. Semilogarithmic plot of $\delta^{18}O$ vs. distance for all basaltic and andesitic country rocks lying outside the central plutonic complex on the Isle of Mull (stippled pattern), except that the two open circles represent samples from country-rock screens near the edge of the central complex; the distance given is the distance in km to the nearest intrusive contact. The curves labelled shales, sandstones, and basalts represent analogous generalized profiles for various types of country rocks in the vicinity of the plutons on Skye (data from Forester and Taylor, 1976a, 1976b).

Figure 5. Plot of $\delta^{18}\text{O}_{\text{quartz}}$ (solid dots) and $\delta^{18}\text{O}_{\text{feldspar}}$ (open dots) vs. sequence of intrusion for the Skye granites (after Forester and Taylor, 1976a). The age sequence is from Stewart (1965). For comparison, the $\delta^{18}\text{O}$ values of quartz and feldspar of most normal granites throughout the world are typically at least as high as the horizontal bands shown on the figure. The symbol m indicates the mineral is from a miarolitic cavity.

MGE = Maol na Gainmhich epigranite, GE = Glamaig epigranite;
 BDME = Beinn Dearg Mhor epigranite, LAE = Loch Ainort epigranite,
 SPE and SPF = Southern Porphyritic epigranite and felsite, respectively,
 MS = Marscoite Suite (hybrid rocks), ME = Marsco epigranite,
 GBME = Glas Beinn Mhor epigranite, BCroG = Beinn na Cro granite,
 AFG = Allt Fearnna granite, BDG = Beinn an Dubhaich granite,
 BCG = Beinn na Caillich granophyre, and CSG = Creag Strollamus granite.

Figure 6. Semilogarithmic graph showing the extremely uniform $\delta^{18}\text{O}$ values of oxygen extracted during step-wise fluorination treatments of euhedral single crystals of β -morphology quartz from a sample of the Southern Porphyritic Epigranite on Skye (data of Forester and Taylor, 1976a).

Figure 7. Plot of $\delta^{18}\text{O}$ values of hydrothermally altered rocks vs. calculated water/rock (W/R) ratio for the open system and closed system models discussed in the text, assuming initial values of $\delta_{\text{Rock}}^i = +6.5$ and $\delta_{\text{H}_2\text{O}}^i = -14$.

- Figure 8. Plot of calculated values of $\delta D_{\text{biotite}}$ and $\delta^{18}O_{\text{feldspar}}$ that would be obtained during the hypothetical meteoric-hydrothermal alteration of a typical granodiorite (25% quartz, 50% plagioclase, 15% K feldspar, 10% biotite) at varying water/rock (w/r) ratios, assuming the initial conditions shown on the diagram.
- Figure 9. Map of western North America showing the locations of the Cordilleran granitic batholiths discussed in the text.
- Figure 10. Plot of δD and $\delta^{18}O$ of various coexisting minerals from granitic plutons of the Skagway traverse, as a function of distance NE and SW of Skagway (after Magaritz and Taylor, 1976c). The stippled areas are quartz diorite plutons and the blank areas are quartz monzonites. The K-Ar ages of various dated samples along the traverse are also shown.
- Figure 11. Graph showing the δD values of all minerals and whole-rock samples analyzed by Magaritz and Taylor (1976b) from the Prince Rupert-Prince George traverse across the Coast Range batholith (Zones I and II) and various country rocks up to 250 km east of the batholith (Zones III and IV). Also included are analyses of serpentinites from Wenner and Taylor (1974). The vertical lines indicate data on coexisting minerals, and the dashed lines join the δD values of samples from plutonic host rocks with those of the dikes or porphyries that cut them.
- Terminology: Gn. = gneiss complex; QD = quartz diorites; GD = granodiorites; Dikes = basaltic and diabase dikes; Volc. = volcanic rocks; QM = quartz monzonites; Porph. = granite porphyry intrusions; Sed. = sedimentary rocks; Serp. = serpentinites; Blsch. = blueschists. Note that the whole-rock samples of the 2 blueschist samples are low in deuterium only because they contain lawsonite, a mineral that has some very easily exchangeable H_2O in its crystal structure (Magaritz and Taylor, 1976b).

- Figure 12. Graph showing the $\delta^{18}\text{O}$ values obtained by Magaritz and Taylor (1976b) for the oxygen extracted during step-wise fluorination treatments of quartz separates from granodiorite sample 67 from Zone II and quartz monzonite sample 24 from Zone III in central British Columbia. These isotopically zoned quartz samples contrast sharply with the data from quartz in the Skye epigranite shown on Fig. 6.
- Figure 13. Generalized map of the Boulder batholith, Montana (see Tilling et al., 1968), showing sample localities and δD contours for biotites and hornblendes (\pm chlorite) analyzed by Sheppard and Taylor (1974).
- Figure 14. Generalized geologic map of the Idaho batholith (after Armstrong, 1975) showing the contours of re-set K-Ar ages and the locations of the traverse projections indicated in Fig. 16 and 17. The dotted line on the west side of the batholith is the $^{87}\text{Sr}/^{86}\text{Sr}$ boundary of Armstrong et al. (1976) discussed in the text.
- Figure 15. Generalized geologic map of the Idaho batholith, showing sample localities and δD contours for biotites and hornblendes (\pm chlorite) analyzed by Taylor and Magaritz (1976). Note the general correspondence between the areas of low δD values on this figure and the young K-Ar "ages" on Fig. 14.
- Figure 16. Graph showing how the $\delta^{18}\text{O}$ values of feldspar and the δD values of biotite and hornblende (\pm chlorite) change as a function of distance, based on a projection of the data of Taylor and Magaritz (1976) onto the Central Section traverse across the Idaho batholith. The bend in the section (see Fig. 14) is located at a distance of 132 km on the arbitrary scale shown above.

Figure 17. Graph showing how the $\delta^{18}\text{O}$ values of quartz and feldspar and the δD values of biotite and hornblende (\pm chlorite) change as a function of distance, based on a projection of the data of Taylor and Magaritz (1976) onto two E-W traverses across the Atlanta Lobe of the Idaho batholith (see Fig. 14). The data-points for the north traverse are connected by solid lines and those for the south traverse by dashed lines.

Figure 18. Plot of δD of biotite and hornblende (\pm chlorite) vs. $\delta^{18}\text{O}$ of feldspar for samples from Atlanta and Bitterroot Lobes of the Idaho batholith (Taylor and Magaritz, 1976) and the Boulder batholith (Sheppard and Taylor, 1974). Also shown are generalized fields that encompass data on analogous samples from two traverses along the Skagway and Prince Rupert traverses across the Coast Range batholith of British Columbia (Magaritz and Taylor, 1976b, 1976c), and the field occupied by most samples of deep-seated, isotopically Normal Granitic Rocks.

Figure 19. Histogram comparing the proportions of granitic rocks in each of the Cordilleran batholiths that have δD values of biotite and hornblende (\pm chlorite) that are (1) greater than -85 , (2) between -85 and -120 , and (3) less than -120 , based on data shown above in Figs. 10, 11, 12, 16, 17, and 18, and previous data by Godfrey (1962), Levin and Friedman (1962), Taylor and Epstein (1966), and Turi and Taylor (1971).

Figure 20. Generalized geologic map of Mahé, the principal island of the Seychelles Group (modified after Baker, 1963), showing the variation of $\delta^{18}\text{O}$ and δD (Q = quartz, K = alkali feldspar). Most of the island is composed of a coarse-grained hornblende microperthite granite with an extremely uniform isotopic composition, as shown on the figure. The only exceptions are some locally distinct facies distributed along the coastal areas: (a) a higher- ^{18}O granite border zone, remnants of which locally remain along the NW and E coasts; (b) a lower- ^{18}O facies with very large $\Delta^{18}\text{O}_{\text{Q-K}}$ values, recognized only along the NE coast. The granite porphyry bodies (shown in solid black), although texturally very distinct, are isotopically similar to the higher- ^{18}O granite facies ($\delta^{18}\text{O}_{\text{Q}} = 5.3$ to 6.7 ; $\delta^{18}\text{O}_{\text{K}} = 3.8$ to 4.7 ; $\delta\text{D} = -82$ to -95). The dotted contacts shown on the map are very indefinite, but they do suggest that the present outline of the island probably approximates the original dimensions of the main Mahé pluton.

Figure 21. A simplified geologic map of the Precambrian rocks in the St. Francois Mtns., S.E. Missouri, showing the locations and $\delta^{18}\text{O}$ values of quartz and feldspar mineral separates from all the granitic and rhyolitic rocks (after Wenner and Taylor, 1976). The numbers enclosed in rectangles indicate analyses of basalts and diabbases. Mineral $\delta^{18}\text{O}$ values are subdivided into coarse (≥ 1.0 mm) and fine (≤ 0.3 mm) grain sizes. The coarse-grained quartz (Q) and feldspar (F) represent minerals from coarse-grained plutonic rocks or large phenocrysts from finer-grained rocks, whereas the fine-grained quartz (Qm) and feldspar (Fm) are from the groundmass of the rhyolitic volcanic rocks. The systematic pattern of $\Delta_{\text{quartz-feldspar}}$ fractionations is illustrated by contours at $\Delta = +1.0, 0.0, -1.0,$ and -1.5 . Also included in this figure are the δD values for muscovite (M) and chlorite (C), shown as smaller, italicized numbers.

Figure 22. Generalized geologic map of part of Swaziland and the Barberton region, eastern Transvaal, South Africa (after Hunter, 1971; Viljoen and Viljoen, 1969a), showing the sample localities studied by Taylor and Magaritz (1975). These various granitic rocks range in age from 2.6-2.7 b.y. for the Young Granite Plutons to >3.3 b.y. for the Tonalite Domes of the Transvaal and the Ancient Gneiss Complex of Swaziland (Allsopp et al., 1969; Anhaeusser, 1973; Hurley et al., 1972). The age of the latter rocks relative to the Swaziland Sequence is still a matter of controversy, but Hunter (1970) believes that the Ancient Gneiss Complex comprises the oldest rocks of the craton. Note that samples 63 and 64 come from the Mpageni granite, 67 and 66 from the Salisbury Kop pluton, 85 from the Dalmein pluton, and 88 from the Bosmankop syenite (isotopic data on these plutons are delineated on Fig. 23).

Figure 23. Plot of δD vs. $\delta^{18}O$ for samples from the localities shown in Fig. 22 (after Taylor and Magaritz, 1975). Note that the hydrogen isotopic compositions were determined on mafic mineral concentrates (hornblende \pm biotite \pm chlorite), and thus essentially represent the whole-rock δD value (minus any sericite alteration of the feldspars removed during mineral separation). Nels = Nelspruit migmatites, and the antigorite sample is from a peridotitic komatiite in the Onverwacht sequence of the Barberton greenstone belt (Viljoen and Viljoen, 1969b).

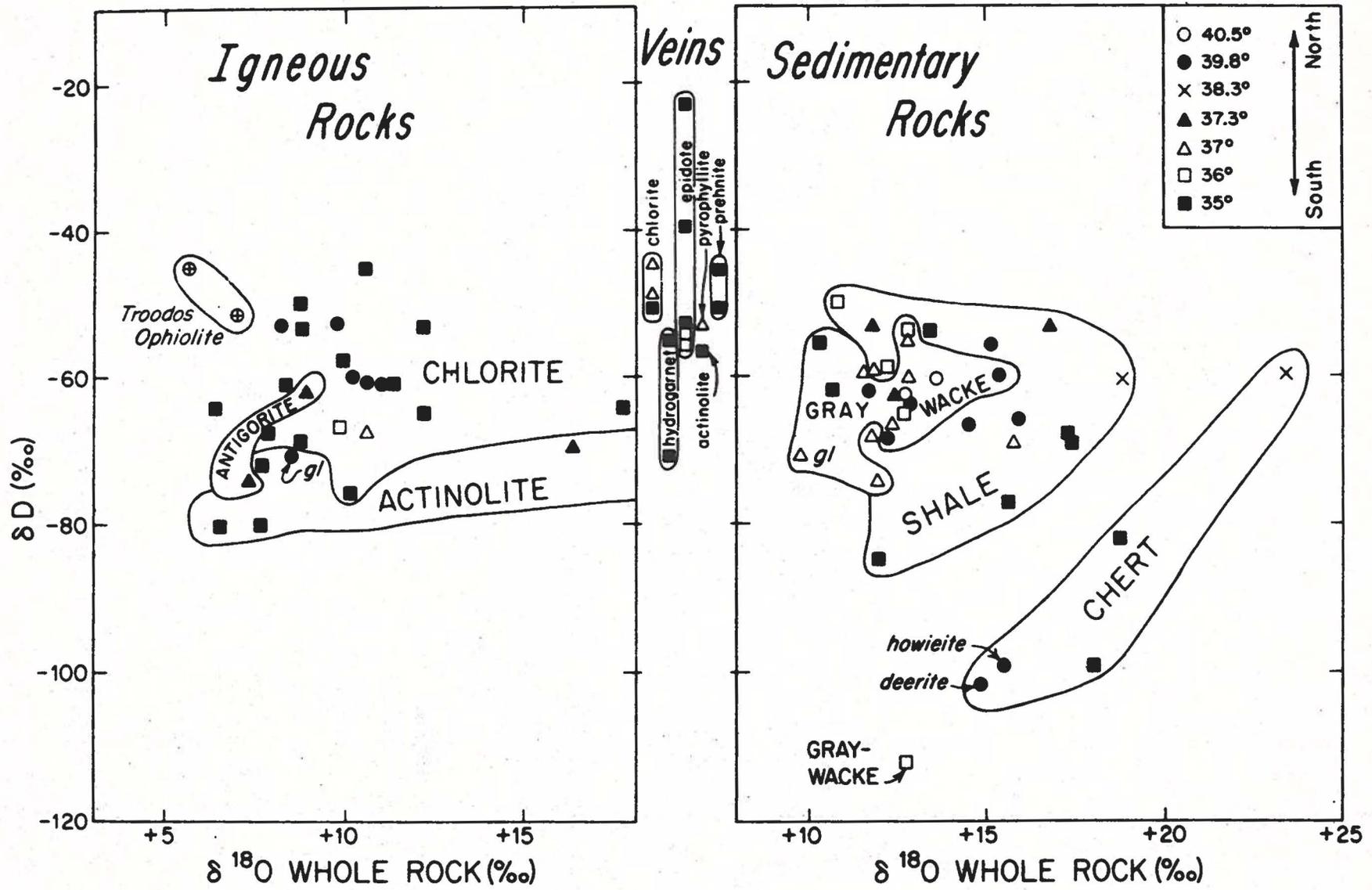


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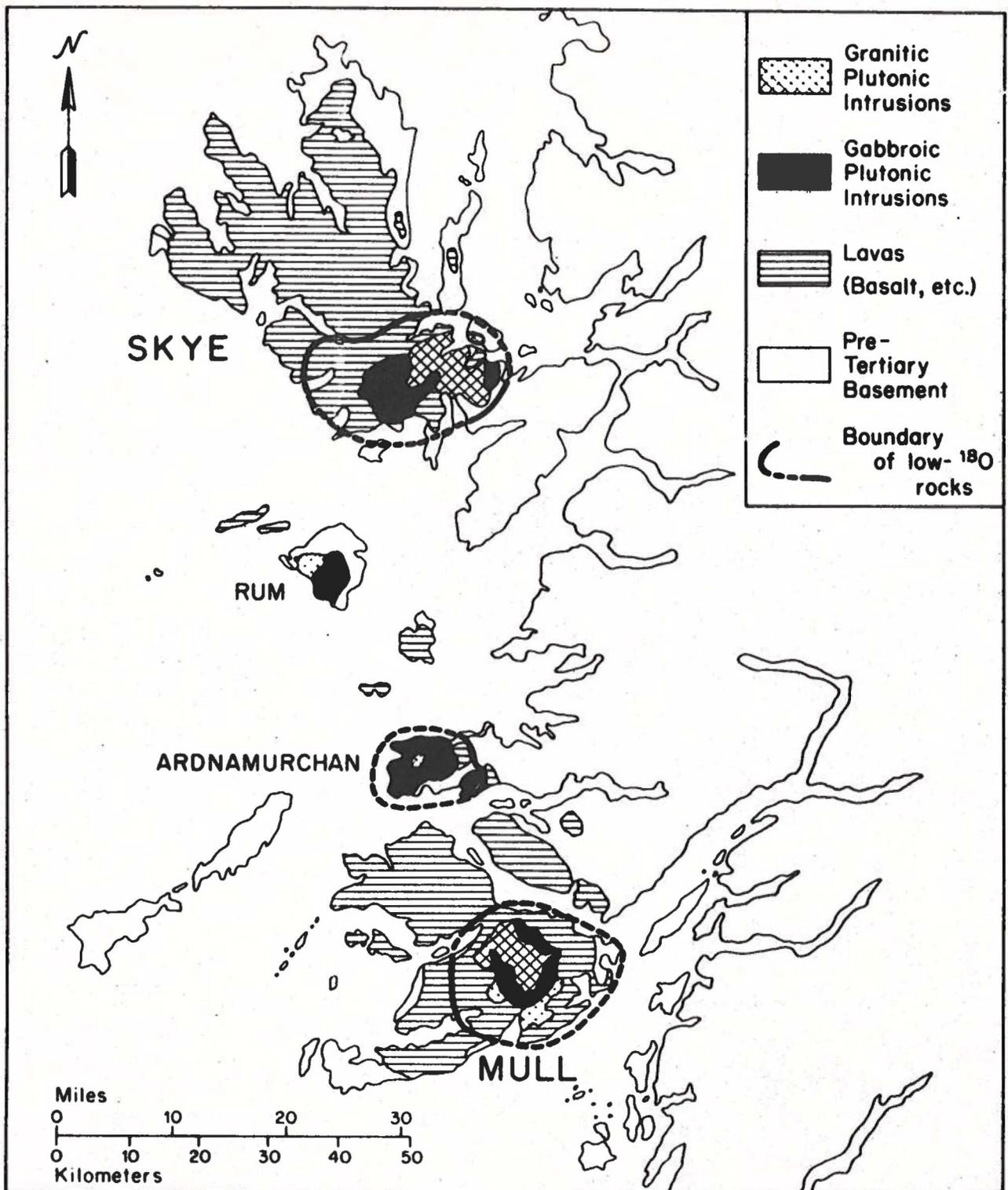


Figure 2.

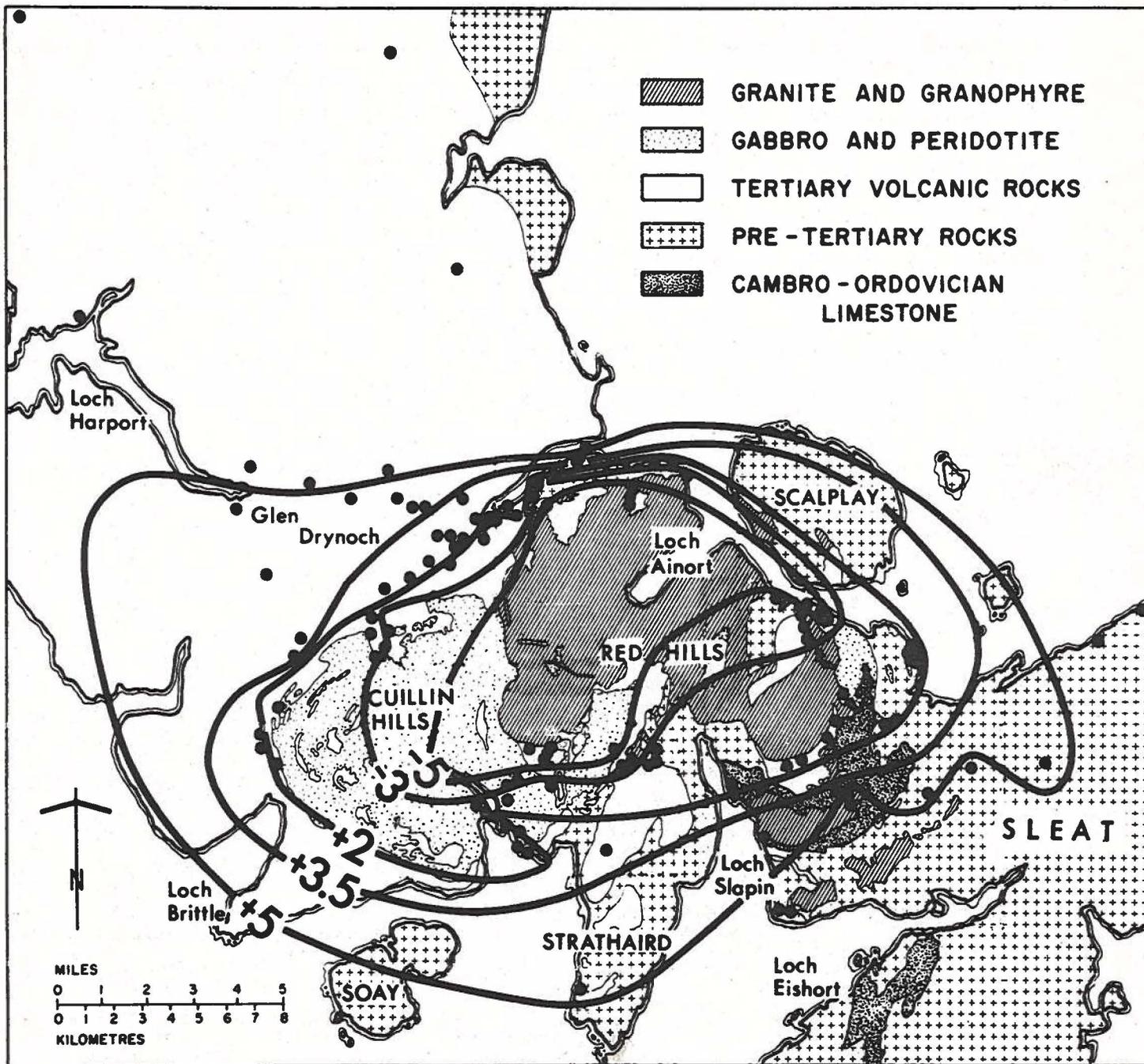
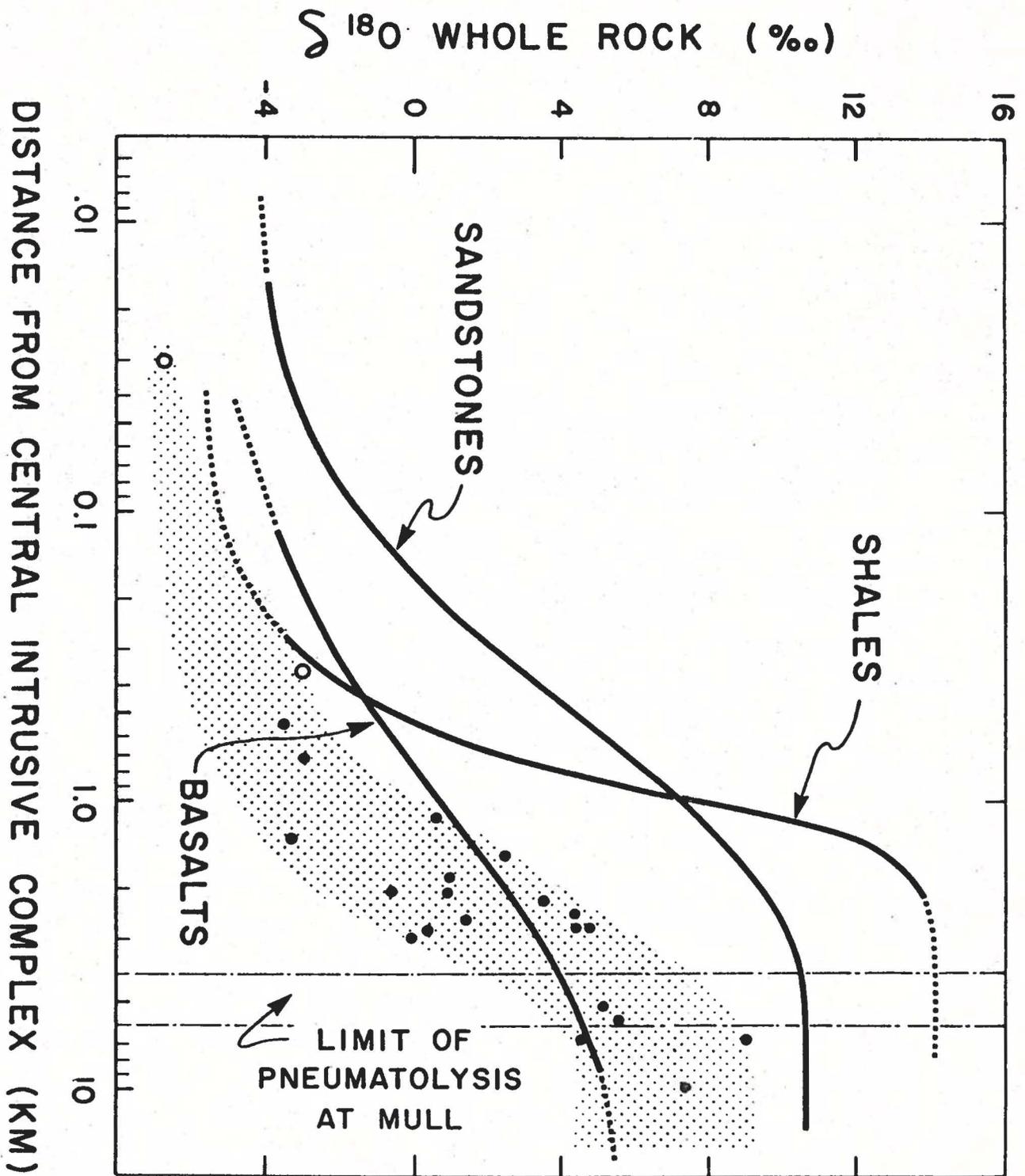


Figure 3.

Figure 4.



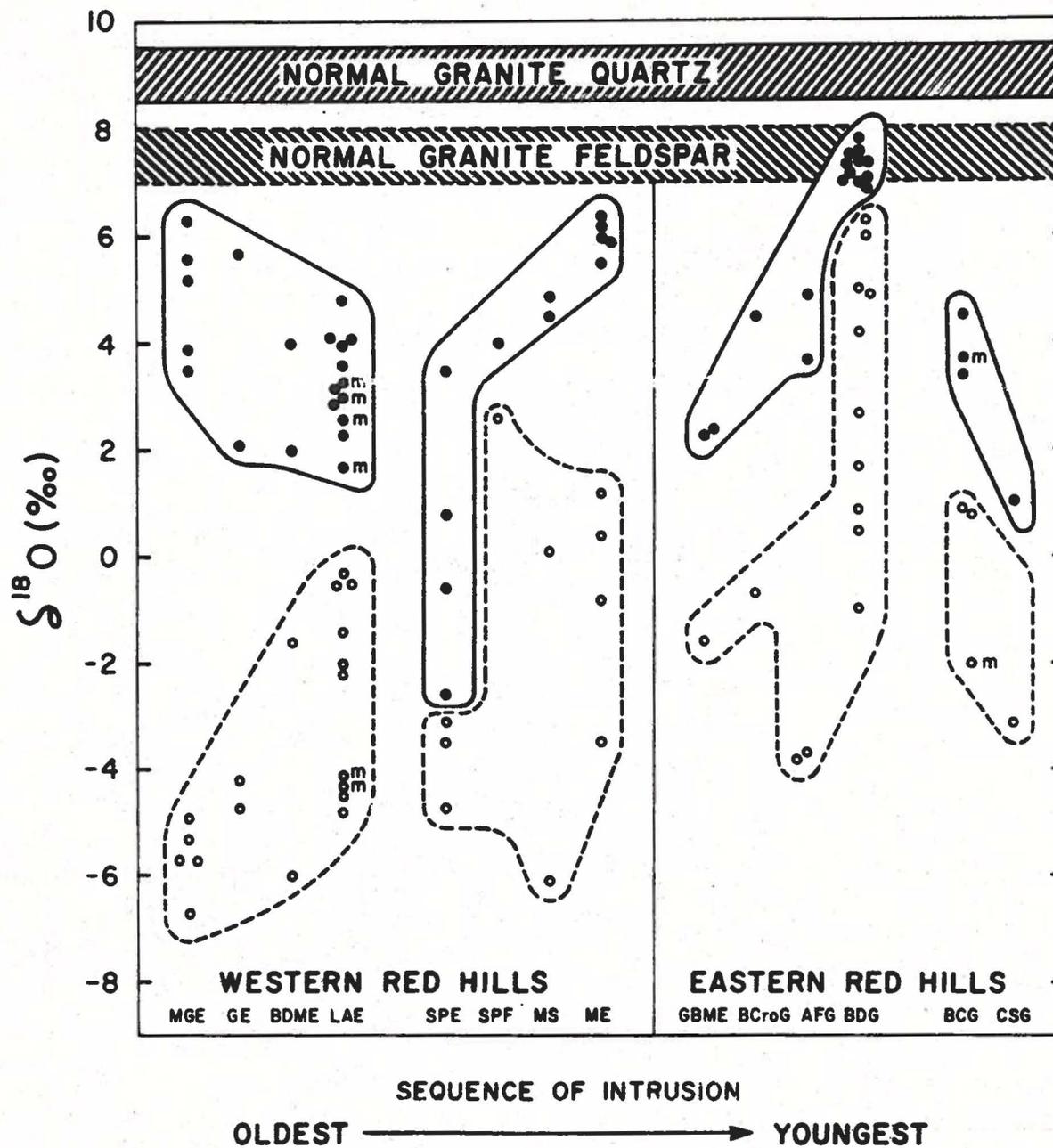


Figure 5.

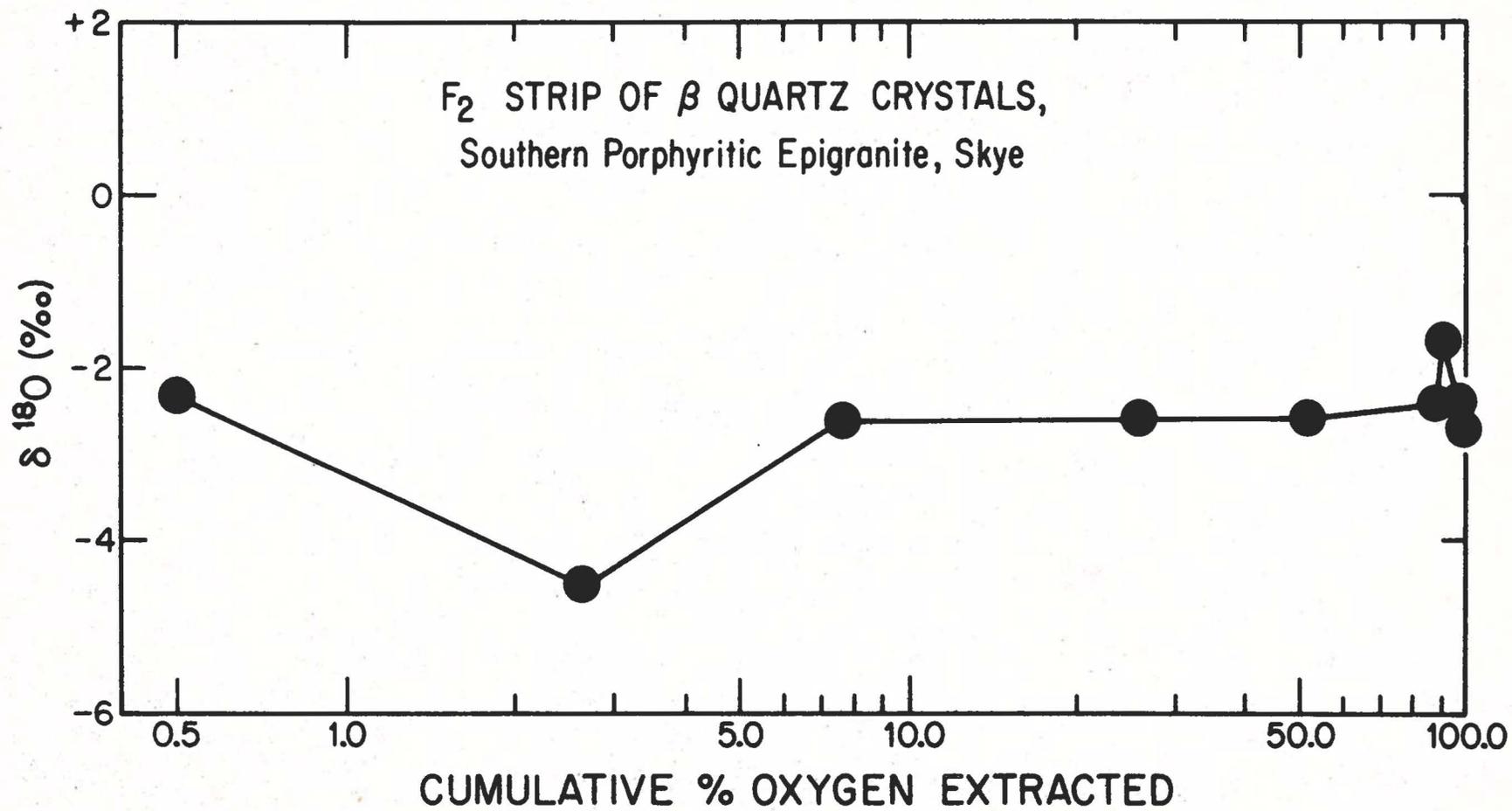


Figure 6.

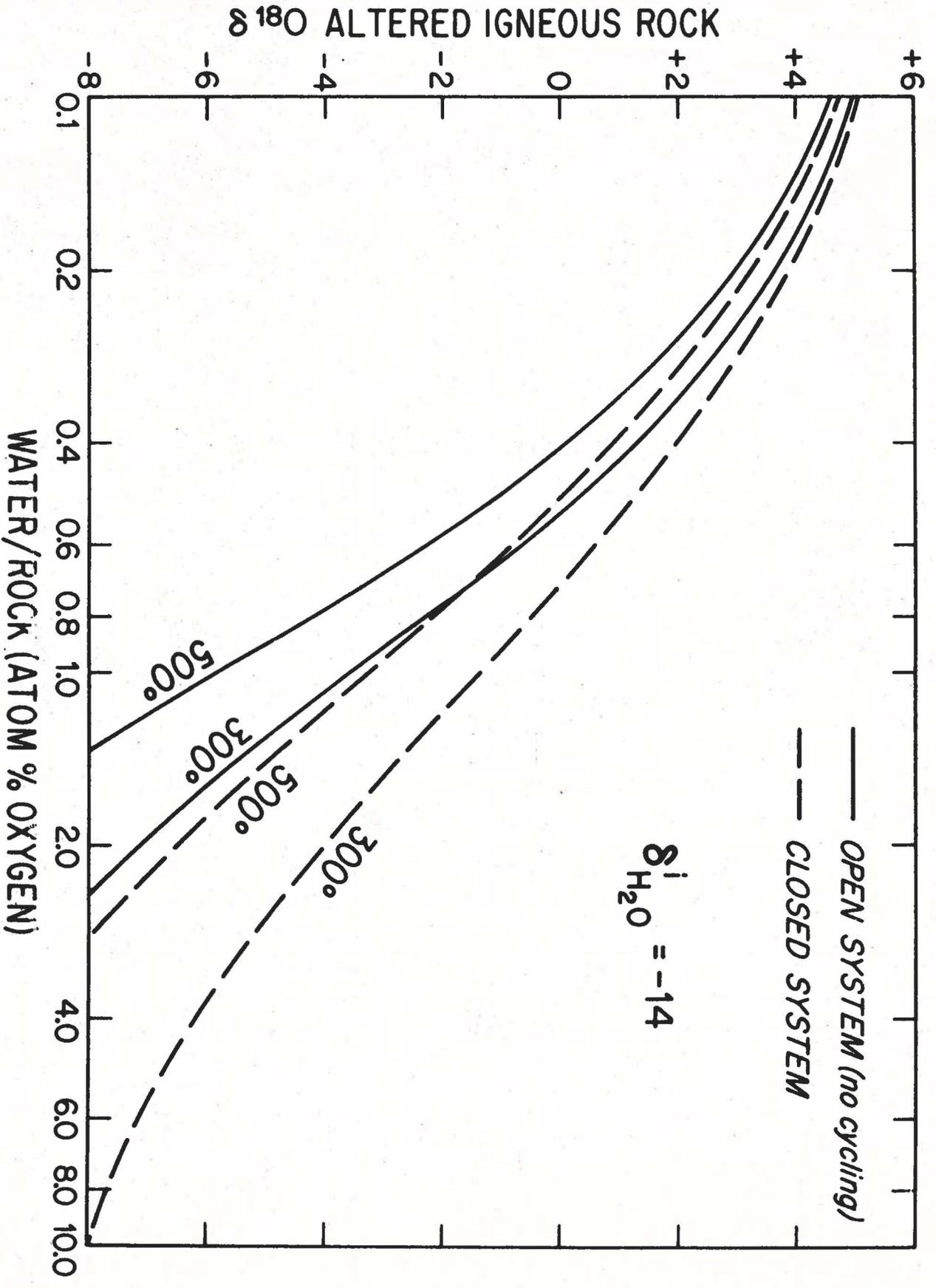


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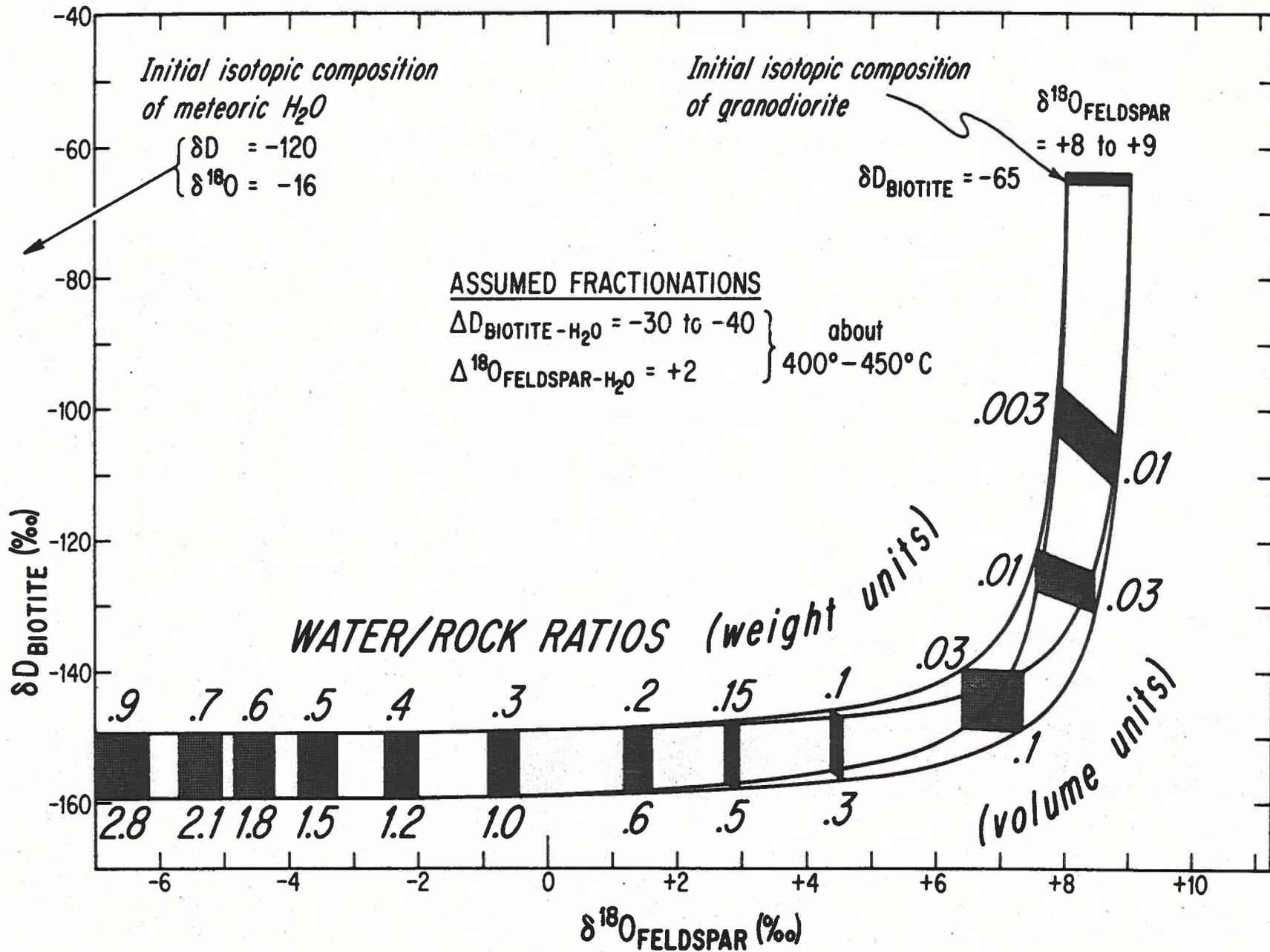


Figure 8.

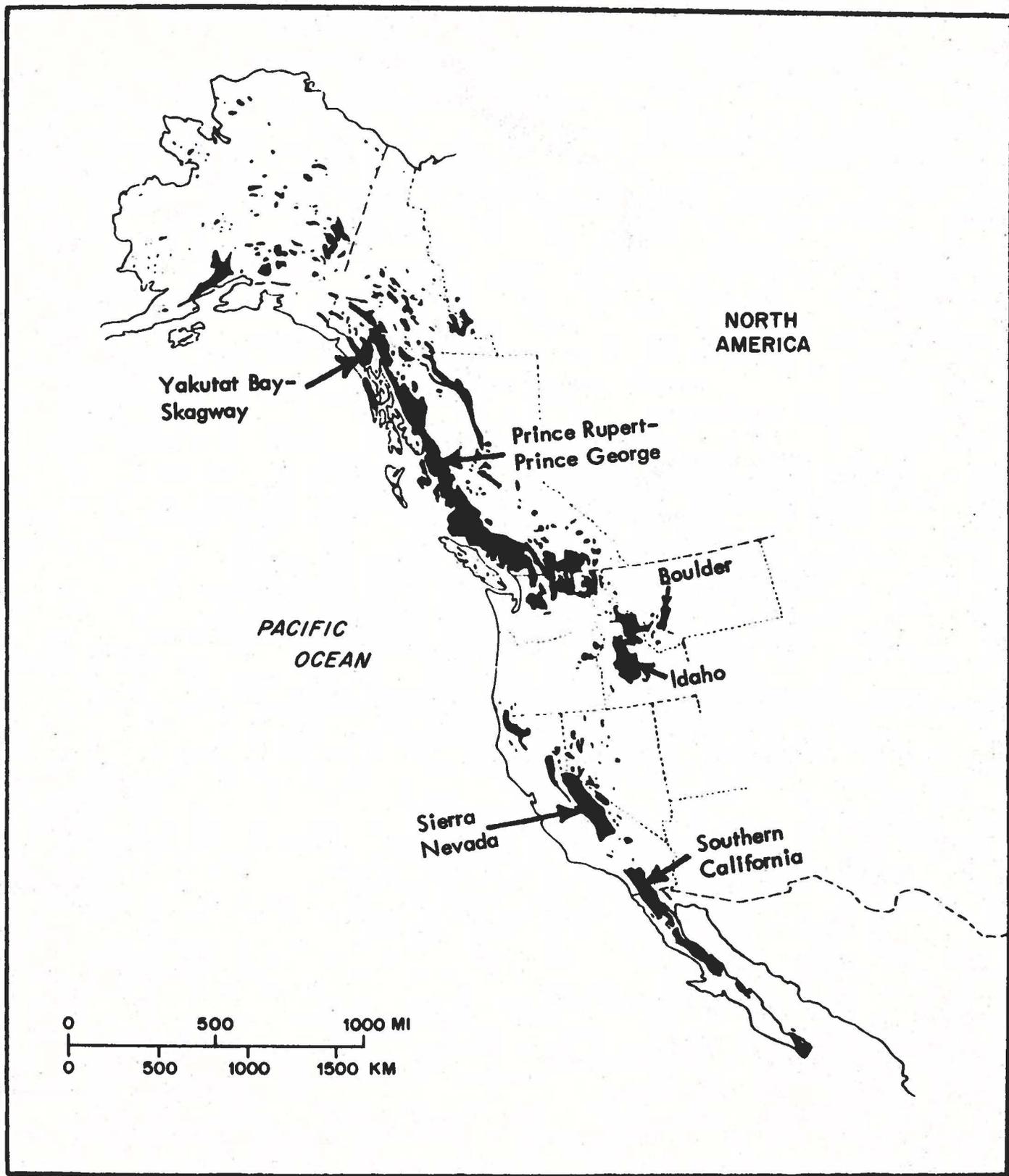


Figure 9.

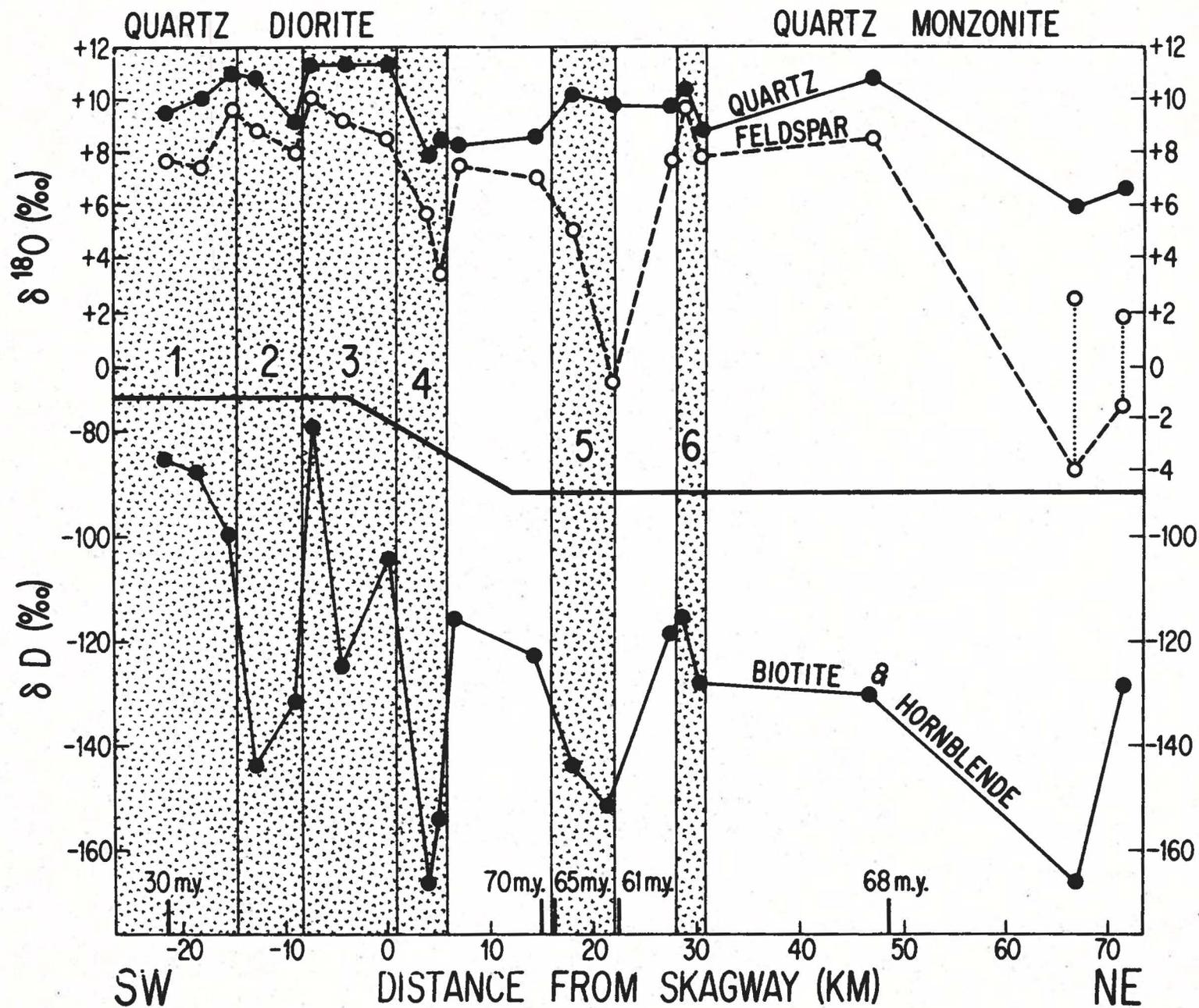


Figure 10.

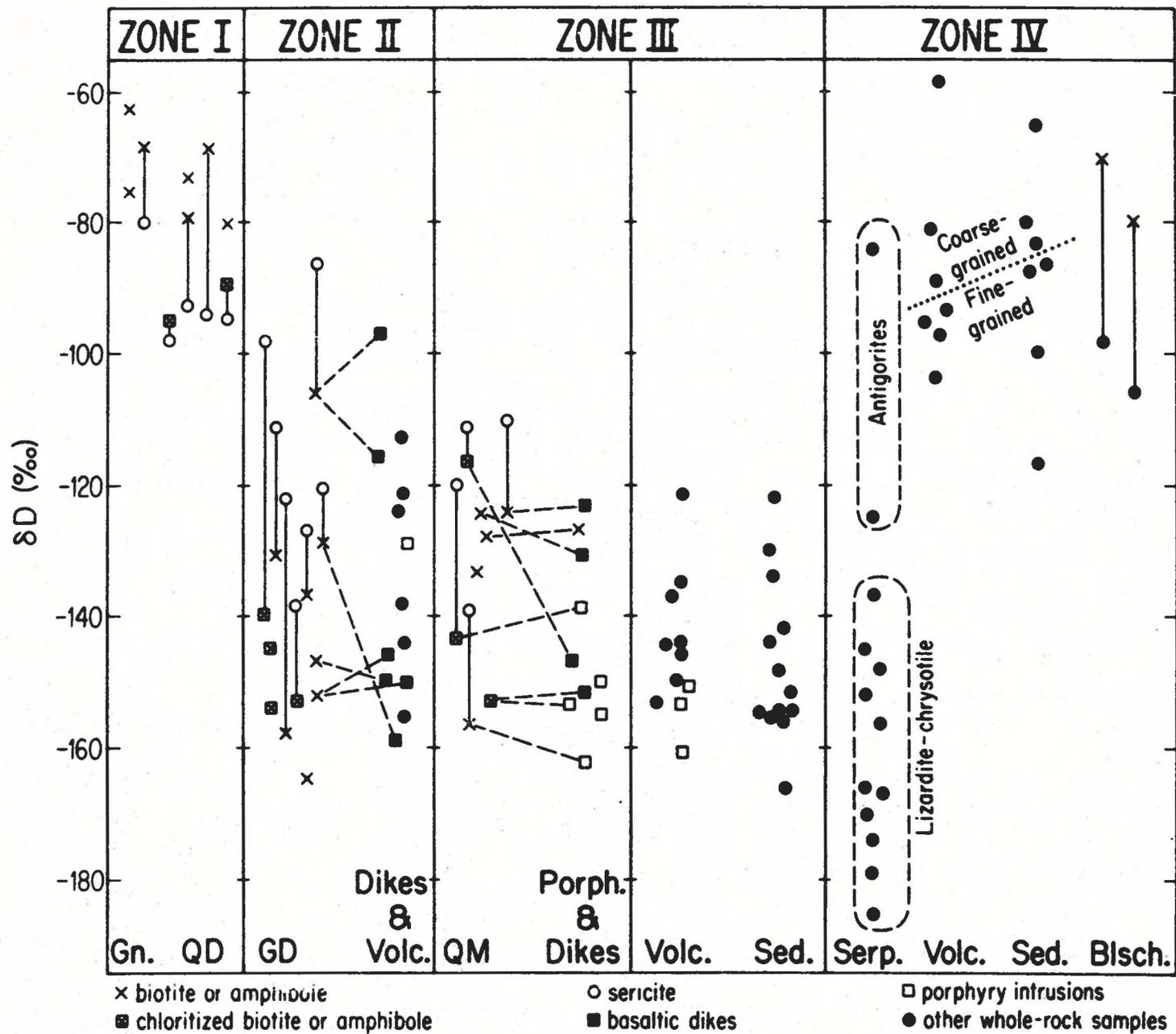


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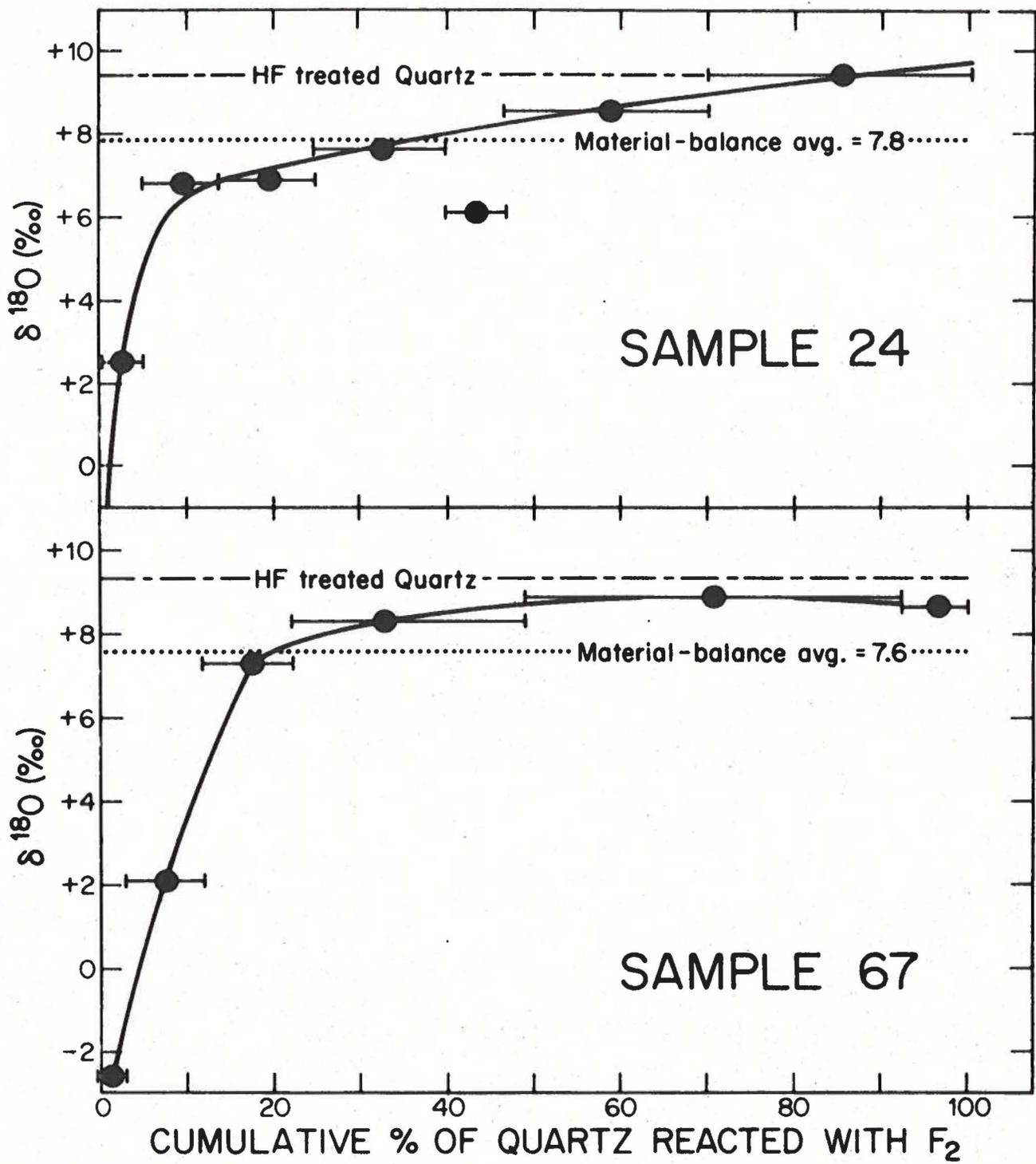


Figure 12.

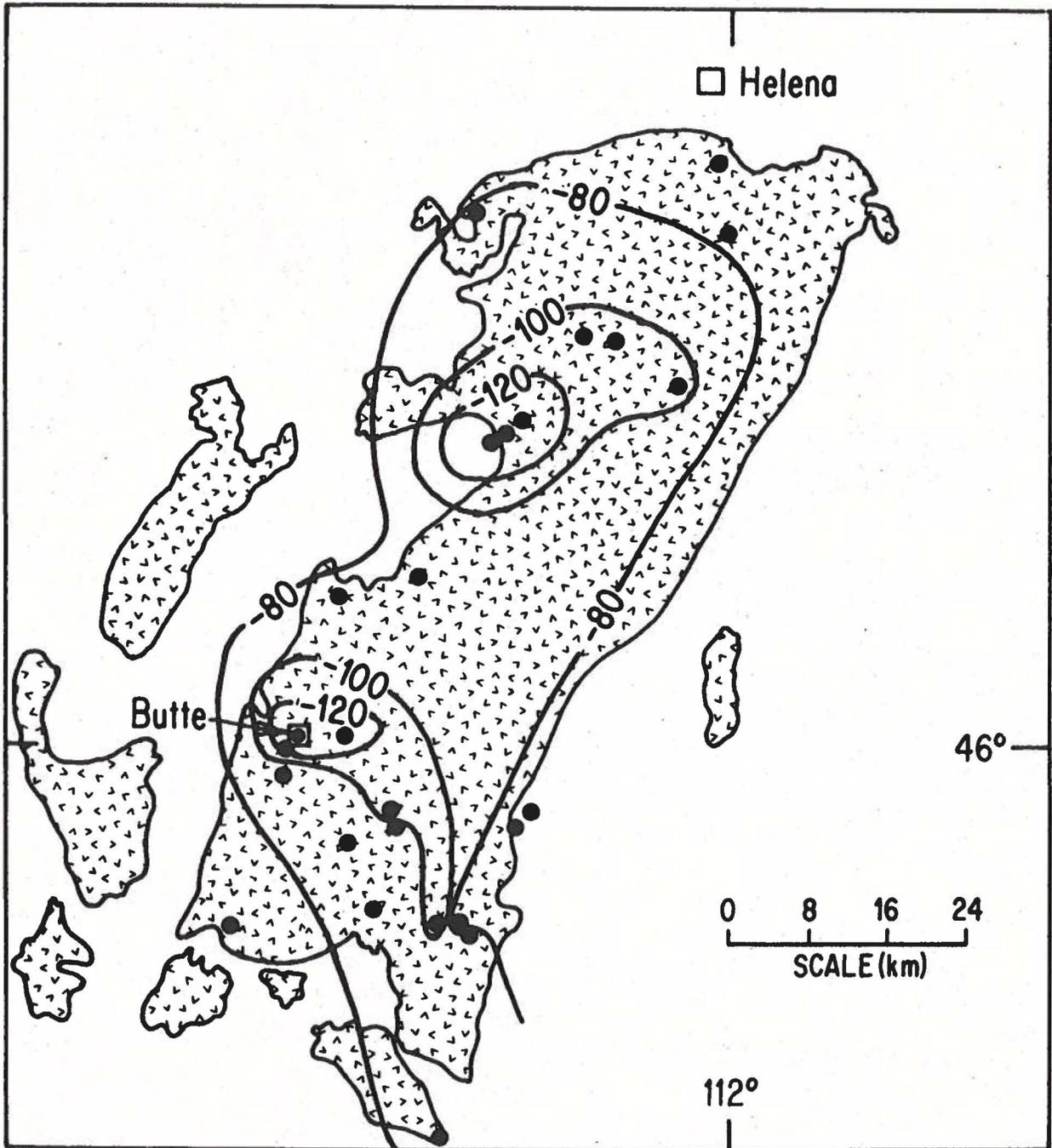


Figure 13.

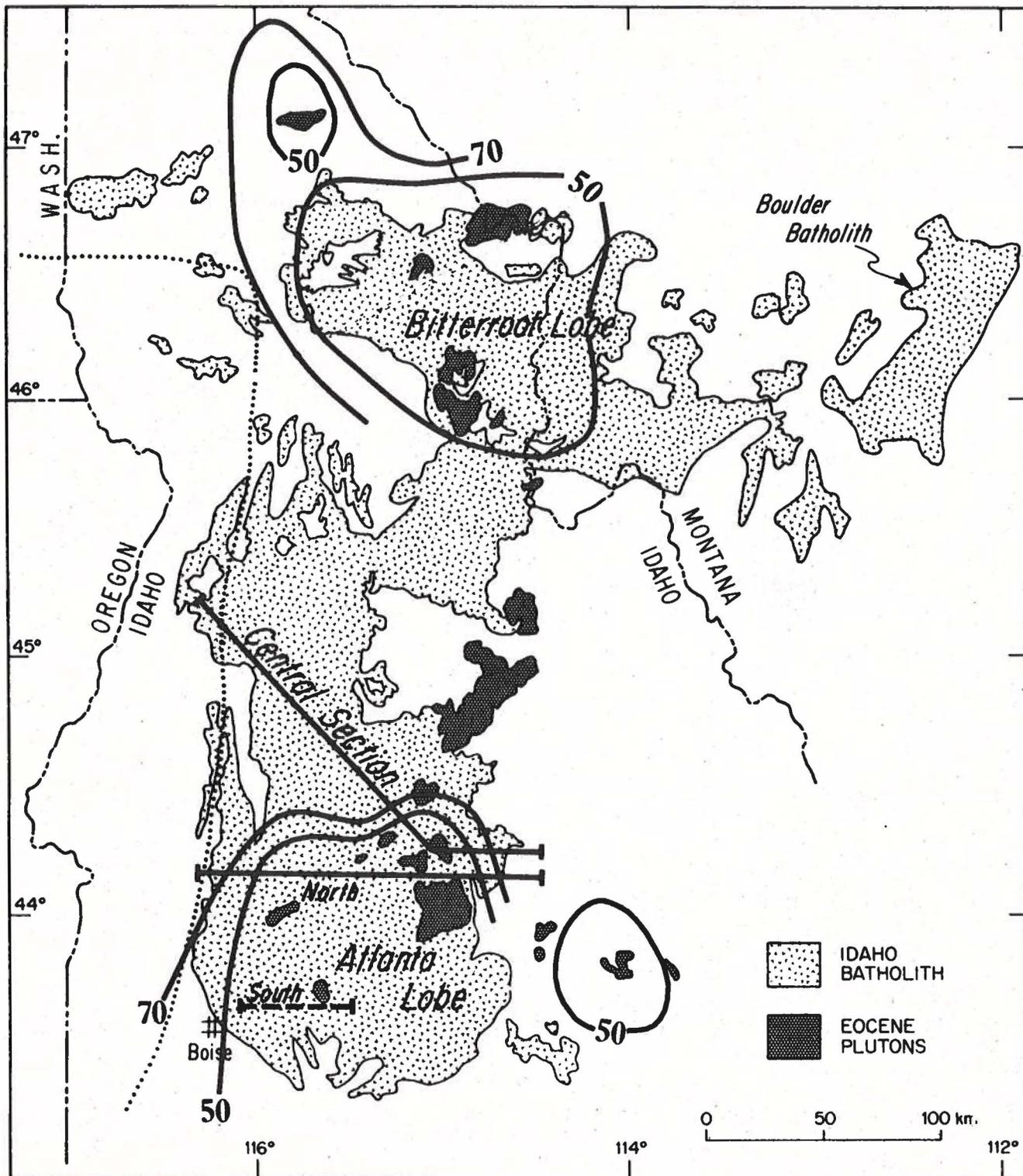


Figure 14.

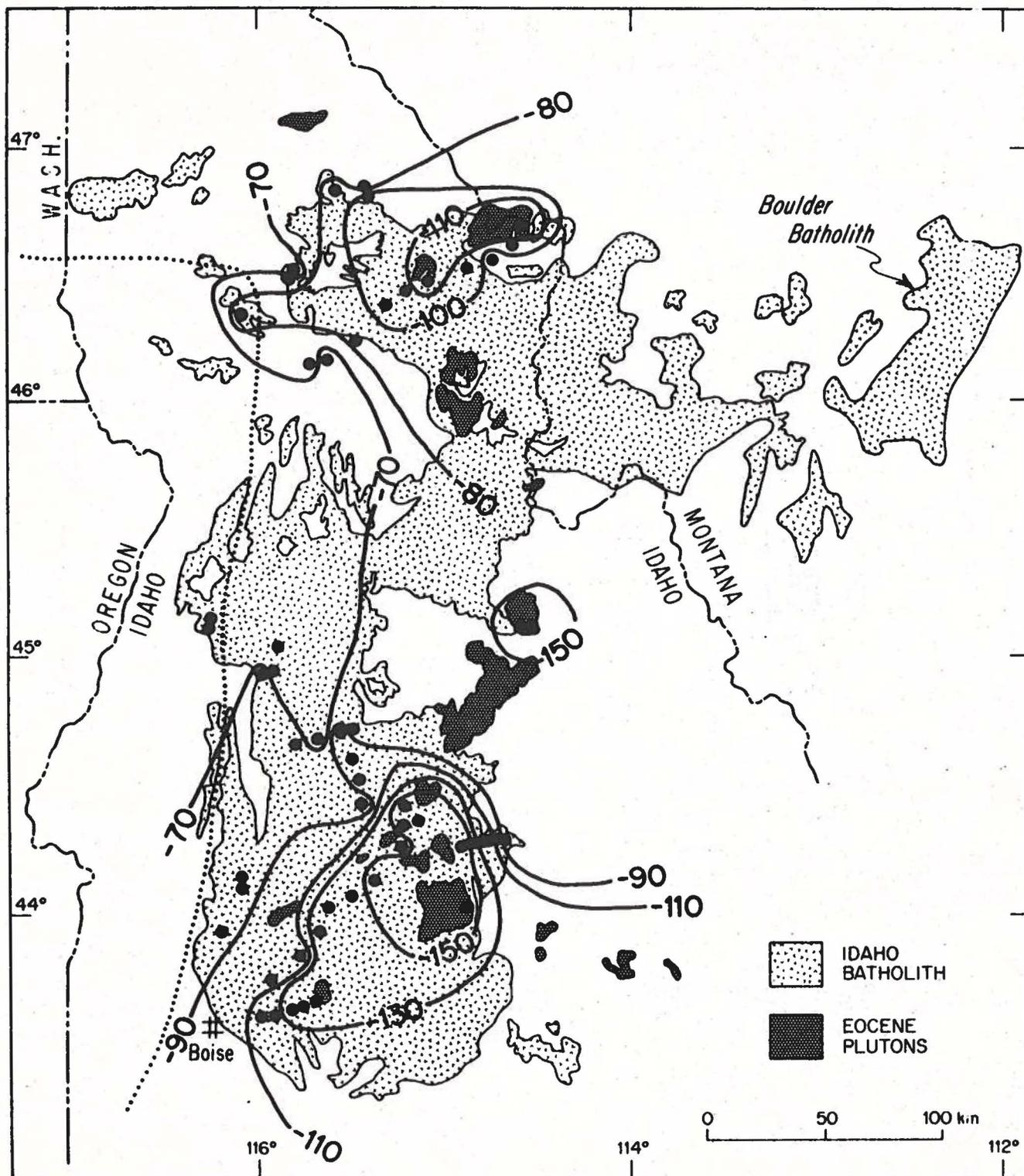


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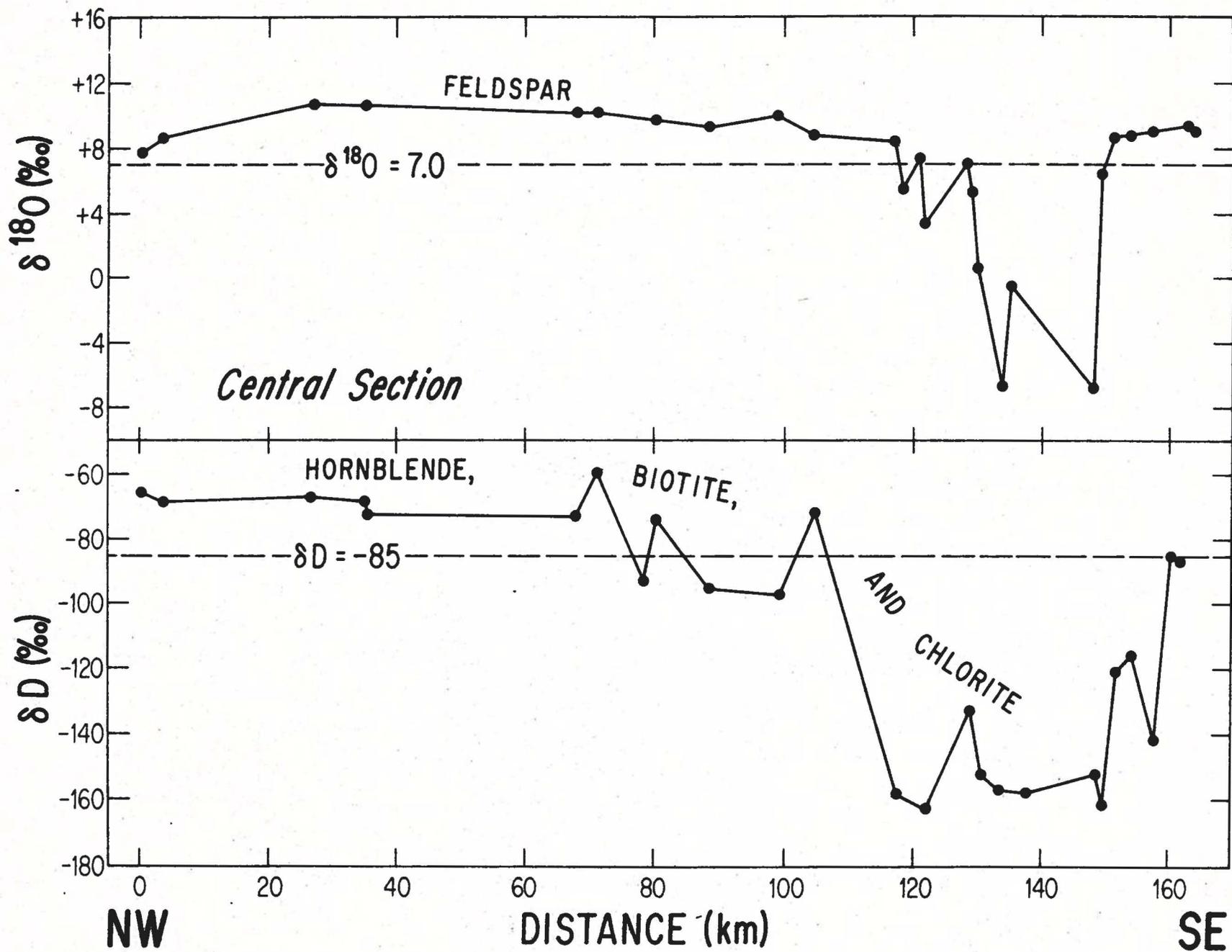


Figure 16.

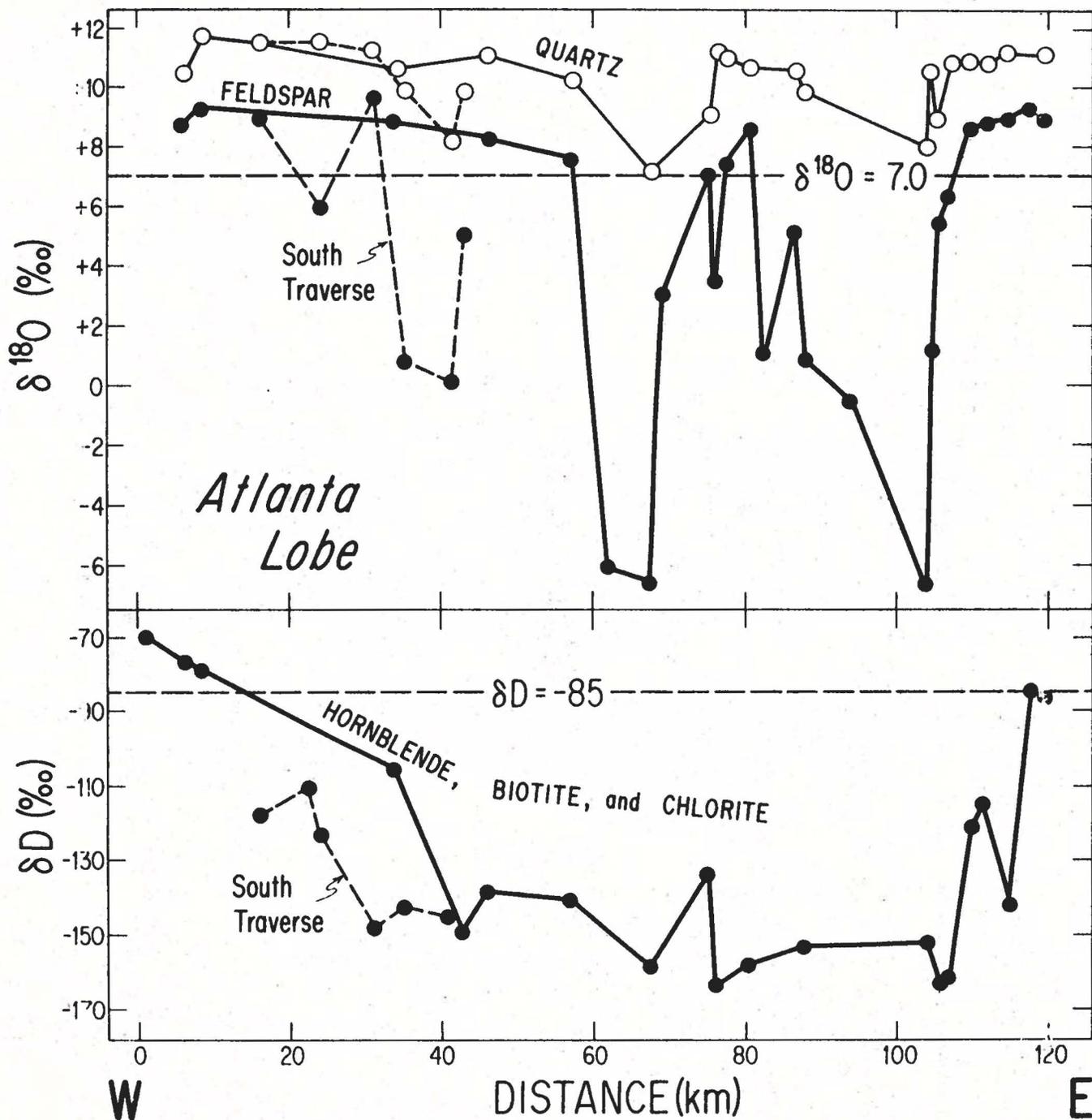


Figure 17.

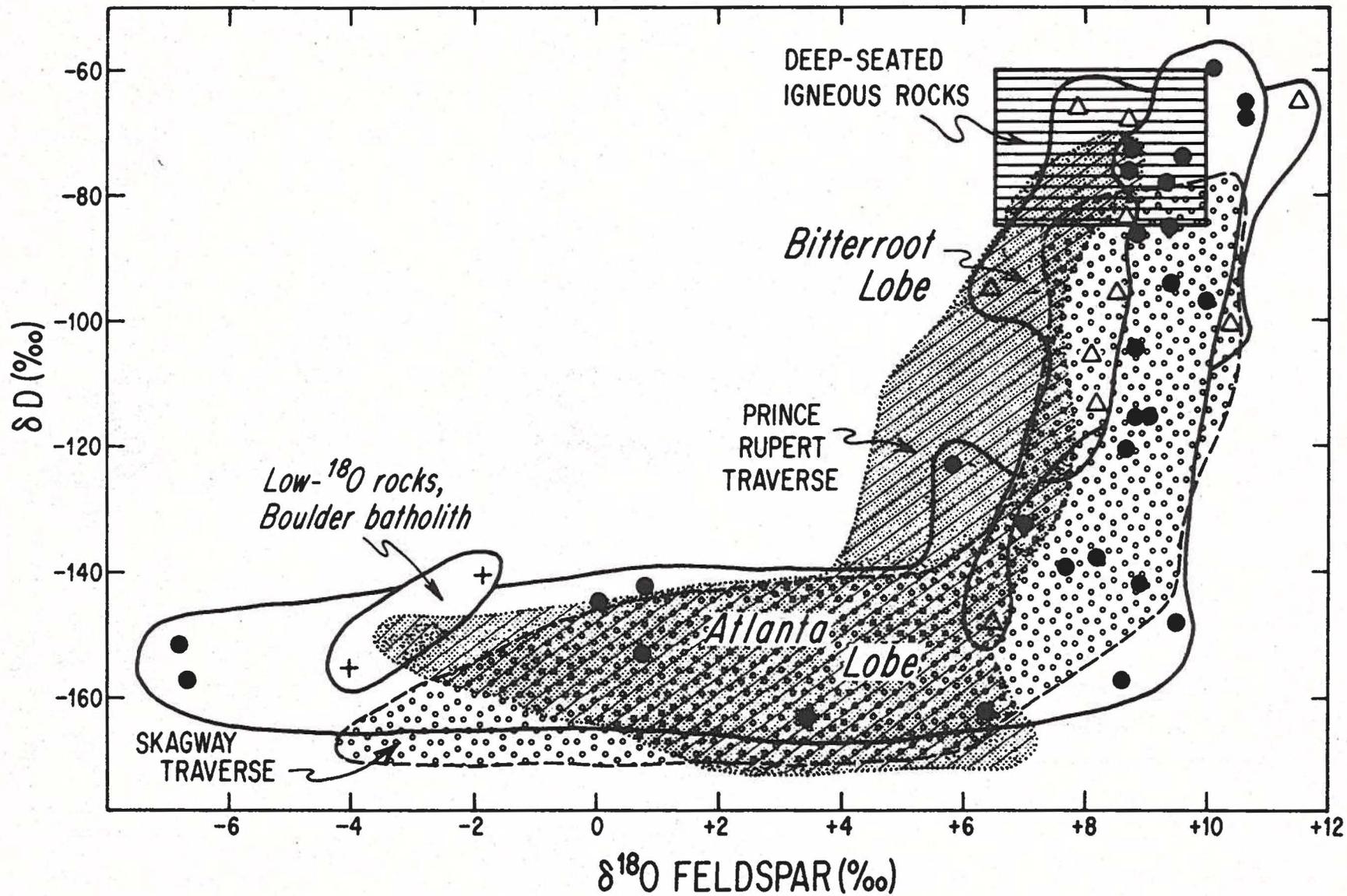


Figure 18.

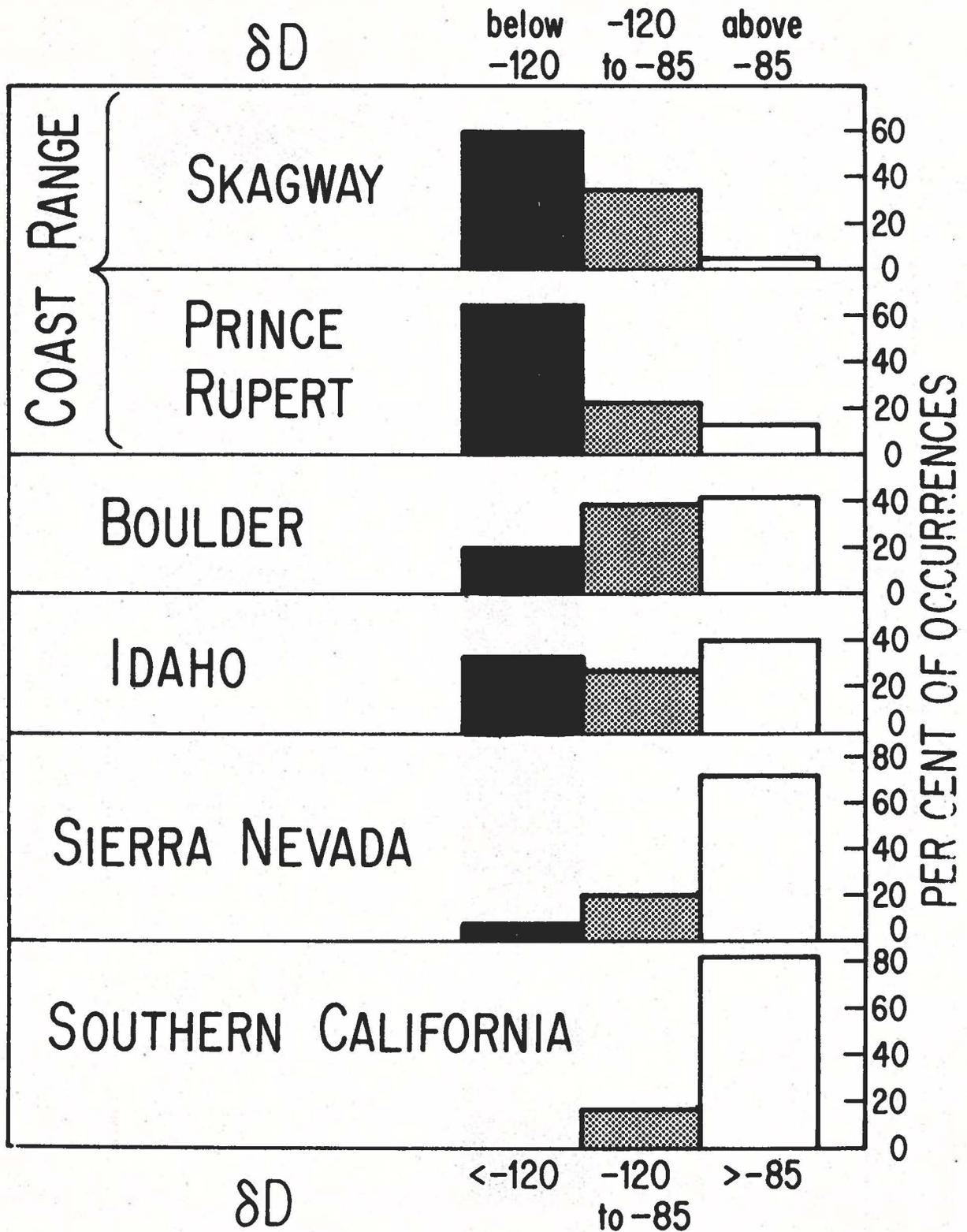


Figure 19.

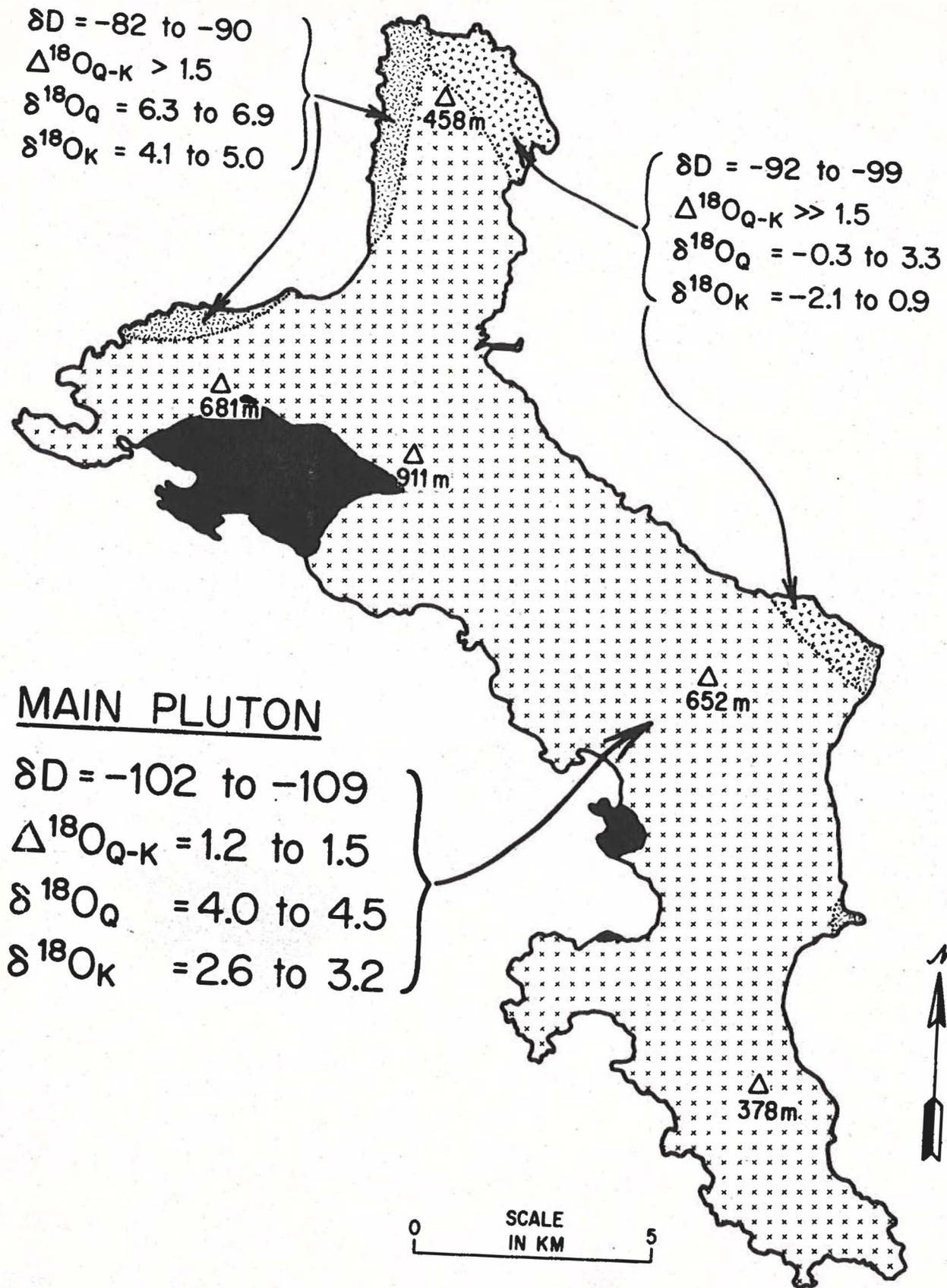


Figure 20.

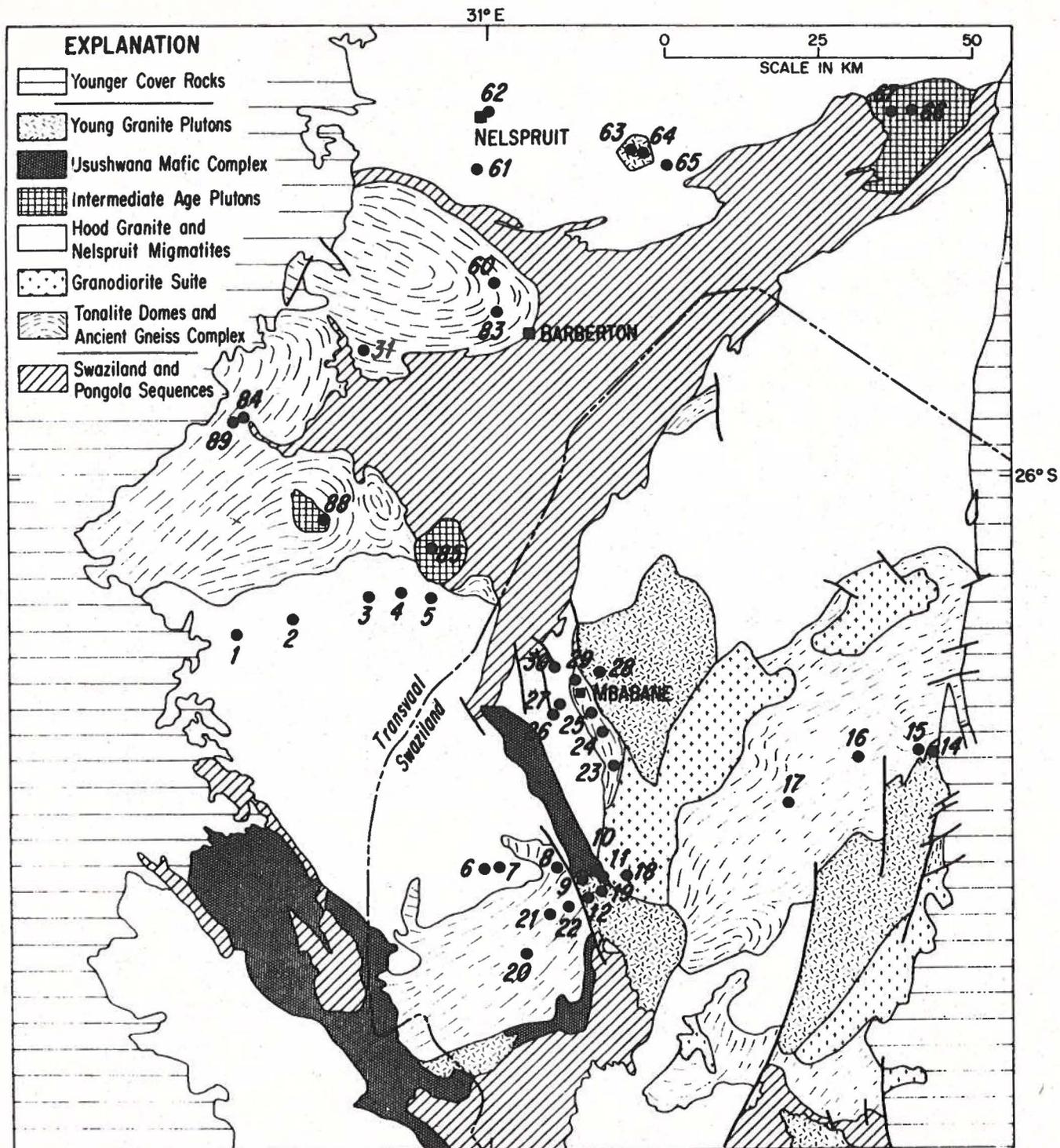


Figure 22.

- Chlorite
- ⊙ Chlorite + Biotite (or Amphibole)
- Slightly Chloritized Biotite (or Amphibole)
- Biotite (or Amphibole)

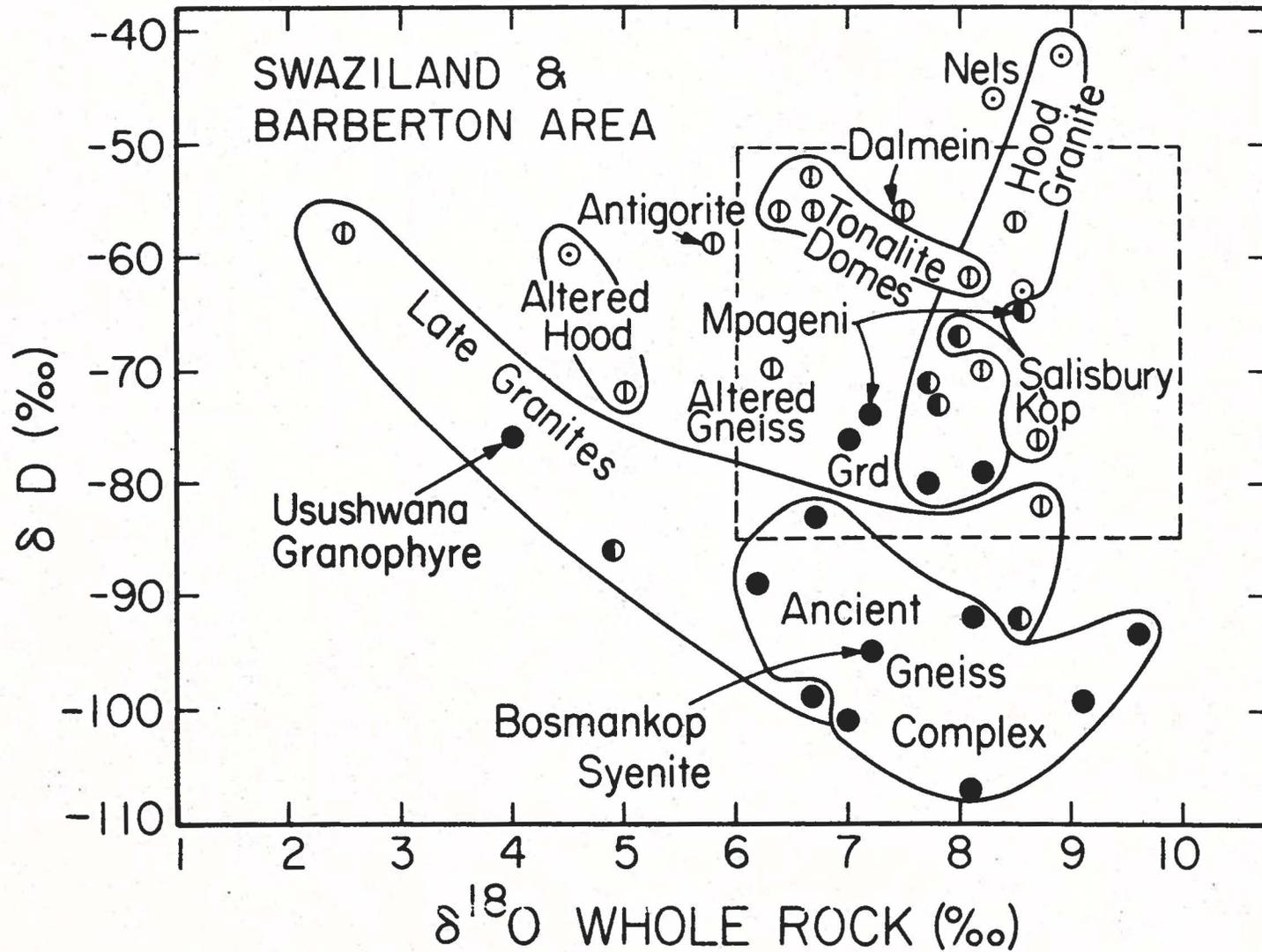


Figure 23.