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PETROCHEMICAL EVIDENCE

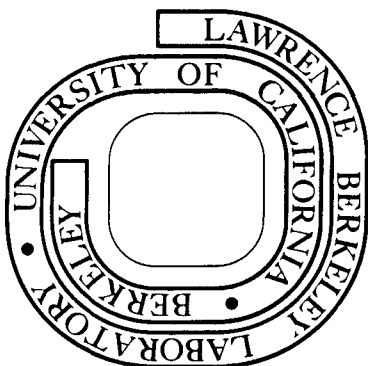
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THE PRESTON PEAK OPHIOLITE, KLAMATH MOUNTAINS, CALIFORNIA, AN IMMATURE ISLAND ARC: PETROCHEMICAL EVIDENCE¹

By ARTHUR W. SNOKE², H. R. BOWMAN³, and A. J. HEBERT³

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ABSTRACT

Major and trace element contents (including rare earth elements) have been determined on a suite of rocks from the Preston Peak ophiolite, Klamath Mountains, California. These geochemical data combined with field relations and petrography suggest that the Preston Peak ophiolite is a remnant of a Permian-Triassic immature island arc. The lowest unit in the ophiolite is a depleted peridotite tectonite which was subsequently intruded and overlain by mafic rocks as the ophiolite developed. Diabase and diabase breccia, the most common constituents of the mafic complex, have undergone greenschist facies metamorphism and related chemical migration. Nevertheless, low TiO₂ contents as well as low Cr values suggest similarity with island arc tholeiite rather than abyssal tholeiite. Xenoliths of clinopyroxenite and gabbro, locally common in the diabasic rocks, appear to be cumulates developed in small magma chambers that were later disrupted by intruding younger basaltic magmas. Evidence of extensive differentiation is lacking and apparently much of the fine-grained mafic rocks of the ophiolite represent primary magmas. The most leucocratic and siliceous rock in the suite is a Na-rich quartz diorite which is enriched in light rare earth elements and probably represents an independent magma type rather than a differentiate.

INTRODUCTION

Extensive terranes of ophiolitic rocks are widely developed throughout the Klamath Mountains of northwestern California and southwestern Oregon (Coleman and Irwin, 1974; figure 1). The ages of these complexes are uncertain; but minimum ages, based on overlying fossiliferous strata, decrease oceanward (east to west) across the province (Irwin, 1973). The ophiolites are concentrated along the boundaries between major lithic belts (figure 1) and are bounded along their basal contact by a fault or fault system of regional extent (Irwin, 1964; Davis, 1968). The emplacement of the ophiolites is not fully understood although an accretionary phenomenon related to an evolving series of east-dipping subduction zones is often suggested (Hamilton, 1969; Burchfiel and Davis, 1972, 1975; Coleman and Irwin, 1974). In the Preston Peak area, Del Norte and Siskiyou Counties, California (figure 1) a particularly well-exposed ophiolite forms the base of the western Paleozoic and Triassic belt of Irwin (1960). Field relations (Snoke, 1972, in press) indicate that the complex has a pseudo-stratigraphic aspect ranging upward from a basal sheet of peridotite tectonite through a mafic complex chiefly of metadiabase and metadiabase breccia and overlain by metabasaltic and metasedimentary rocks (figure 2).

An important conclusion drawn from field and petrographic studies of the Preston Peak ophiolite (Snoke, 1972, in press) is that the basal tectonitic ultramafic rocks and the overlying constructional pile have undergone different crystallization histories. This is indicated by the sharp metamorphic contrast between the tectonitic ultramafic rocks and the non-schistose lower greenschist facies metadiabase and metagabbro. The local preservation of intrusive contacts between younger mafic rocks and tectonitic peridotite substantiates this hiatus in the development of the ophiolite. Furthermore, widespread mafic dikes with chilled margins indicate that the ultramafic rocks have acted as country rock rather than source rock for the basaltic magmas.

The purpose of the present study was to make a petrochemical reconnaissance of the Preston Peak ophiolite to define its chemical character and variation. These new chemical data along with available field evidence suggest that the ophiolite may represent a remnant of an immature island arc. However, these data are not unequivocal, and the uncertainties of this interpretation are enumerated.

PETROGRAPHY

The analyzed rocks from the Preston Peak ophiolite can be divided into four distinct groups:

1. Serpentinized tectonitic peridotite (alpine-type)—two analyses (major and trace elements).
2. Diabase and diabase breccia—six analyses (major elements), nine analyses (trace elements).
3. Coarse-grained xenoliths in diabase—two analyses (major and trace elements).
4. Nonporphyritic biotite quartz diorite—one analysis (major and trace elements).

Tectonitic Peridotite

The analyzed alpine-type peridotites are massive and un-sheared but extensively serpentinized. However, relict primary minerals are common and indicate that the original mineral assemblage was Mg-rich olivine, enstatite, diopside, augite, and chromian spinel. In addition to serpentine-group minerals, other common alteration minerals are talc, chlorite, tremolite, and magnetite. Trace amounts of carbonate and brucite are also present.

A relict mesoscopic fabric is preserved in these rocks and is defined by a planar arrangement of ellipsoidal aggregates of olivine and pyroxene. In sample PP-225 distinct pyroxene-rich segregation bands (about 5 mm thick) are well developed. The relict primary mineral assemblage and the penetrative fabric indicate that these rocks have undergone high-temperature recrystallization and flowage. Similar ultramafic rocks form the basal part of other Klamath Mountains ophiolites (Lipman, 1964; Himmelberg and Loney, 1973; Goullaud, 1975).

Diabase and Diabase Breccia

Diabase and diabase breccia constitute more than 95 per cent of the mafic complex. Other common components are gabbroic and clinopyroxenite xenoliths (figure 3) and dikes and small masses of aphanitic greenstone. All these rocks are non-schistose but have undergone low-grade greenschist facies metamorphism. The typical mineral assemblage is altered plagioclase, actinolitic amphibole, epidote, chlorite, and leucoxene (e.g., PP-376, figure 3B). Other rocks have been contact metamorphosed by younger and unrelated Late Jurassic plutons. These rocks consist chiefly of oligoclase-andesine plagioclase and green hornblende; they typically lack or contain only minor quantities of chlorite and epidote. Although several diabase specimens have undergone incipient contact metamorphism (PP-317, PP-320, S-13B), all the analyzed diabases retain their original igneous texture and

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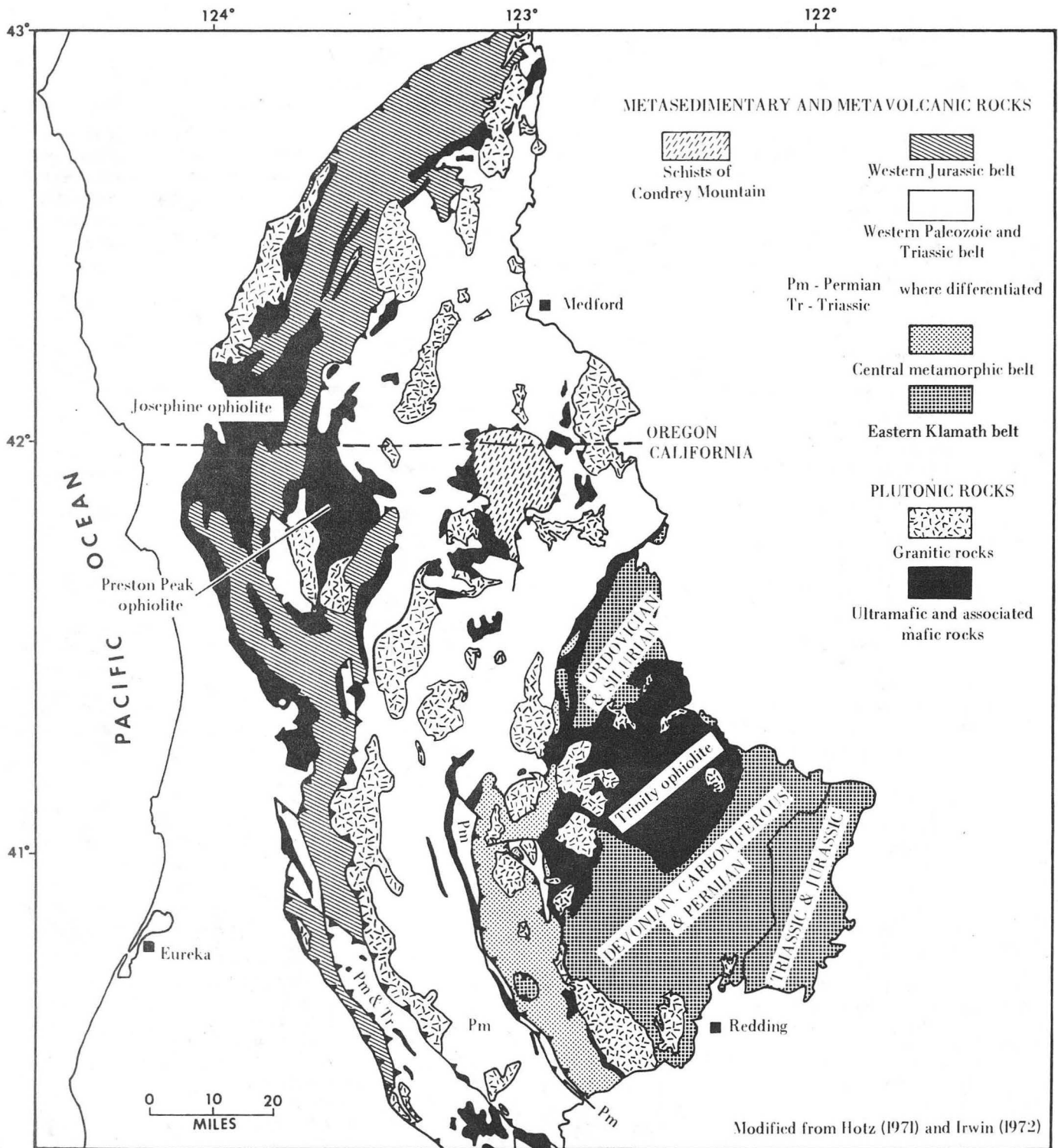


Figure 1. Generalized geologic map of the Klamath Mountains.

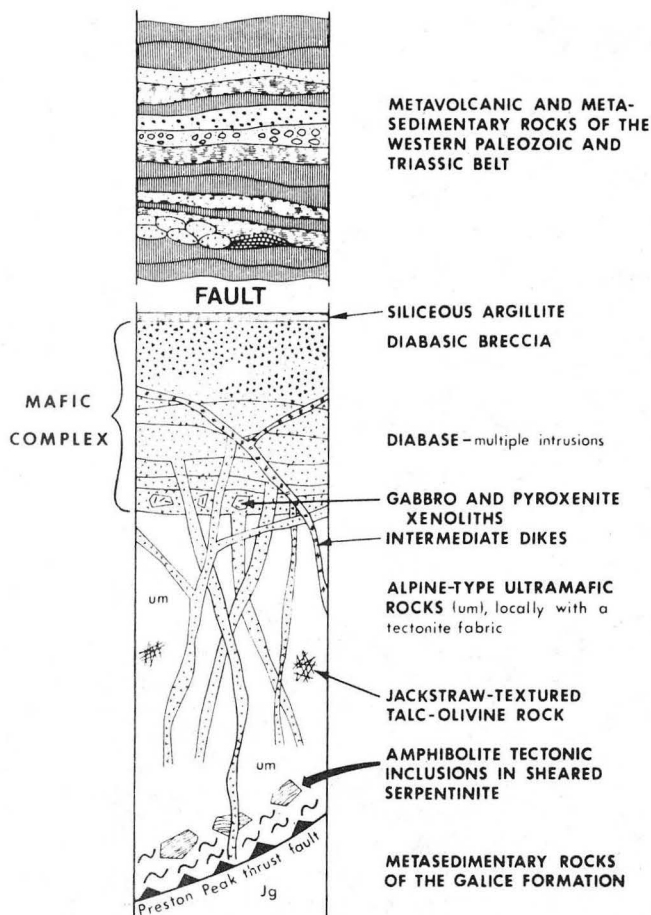


Figure 2. Schematic stratigraphic column of the Preston Peak ophiolite.

show little indication of extensive grain growth and recrystallization.

The diabase breccia (PP-381) consists of poorly sorted, angular fragments of mafic igneous rock. Diabase and aphanitic greenstone clasts are predominant, but microgabbro as well as porphyritic and vesicular basalt fragments are common. These clasts show no evidence of aqueous transport. The rock unit is interpreted to be pyroclastic breccia erupted from a nearby explosive vent. The coexisting mineral assemblage in the breccia indicates greenschist facies metamorphism.

Coarse-Grained Xenoliths

Gabbroic and less commonly clinopyroxenite xenoliths are found in metadiabasic rocks of the mafic complex. The xenoliths are apparently scattered throughout the mafic complex but appear to be somewhat preferentially concentrated near basal ultramafic-mafic rock contacts.

The xenoliths, like their host rock, have been extensively overprinted by greenschist facies metamorphism. The gabbro originally consisted of calcic plagioclase, augitic pyroxene, and opaque oxides (figure 3C). Plagioclase grains are now heavily saussuritized but retain their original igneous crystal shape. Pyroxene grains are pseudomorphed by fibrous green actinolitic amphibole and irregular chlorite patches. The primary minerals of the pyroxenites are likewise extensively altered. Original olivine grains are completely pseudomorphed by serpentine minerals, while pyroxenes are typically only partially replaced by amphibole. There is no indication that orthopyroxene was a primary mineral phase in any of the xenoliths.

Biotite Quartz Diorite

Intermediate dikes intrude metadiabase and serpentinized peridotite in several localities throughout the Preston Peak area. If all of these dikes are part of the ophiolite is uncertain, for Late Jurassic magmatic activity is widespread throughout the area. The analyzed biotite quartz diorite (PP-322) has undergone a low-grade metamorphic alteration characteristic of the ophiolitic rocks and is texturally distinct from any known Jurassic igneous rocks. This rock is considered to be an example of the leucocratic suite apparently common in ophiolitic assemblages (Coleman and Peterman, 1975).

The rock consists essentially of strongly zoned subtabular plagioclase (andesine to oligoclase), interstitial quartz, ragged biotite, and opaque minerals (figure 3D). Minor accessories include sphene and apatite. No potash feldspar was seen. Saussuritic alteration of plagioclase cores is common. Similarly biotite aggregates are often partially altered to chlorite and muscovite.

ANALYTICAL METHODS

The major elements were determined by X-ray fluorescence analysis using the method described by Hebert and Street (1973). The minor and trace element abundances were obtained using neutron activation analysis (N.A.A.) and the method has been described by Bowman and others (1973). The N.A.A. data are shown in table 2 and the indicated errors are based on counting statistics and are related more to precision than accuracy. A measure of the overall accuracies of these numbers can be obtained by comparing the present data with our measurements of BCR-1 Standard Rock (Mark and others, 1975). In the BCR-1 data we have included our calibration errors which are introduced by the standards used.

CHEMICAL RELATIONS BETWEEN MAJOR ROCK TYPES

Major element variations as well as chemical relations between the lithologic units of the ophiolite are summarized in figures 4–6. A plot of the analyses on an AFM diagram (figure 4) indicates that the rocks clearly fall into distinct groups compatible with field relations and petrography. A study of major element chemical variation in the Coast Range ophiolite of California (Bailey and Blake, 1974) indicate similar divisions between the major lithologies. A FeO^* vs. FeO^*/MgO plot (figure 5) shows somewhat similar divisions but also suggests an indication of iron enrichment in diabase (e.g., PP-320). This minor iron enrichment is also detectable on the AFM diagram (figure 4). A SiO_2 vs. FeO^*/MgO plot (figure 6) spreads out the diabase group based on their SiO_2 variation (47.9 to 52.4) and suggests slight SiO_2 enrichment, perhaps related to limited magmatic differentiation. As will be developed later in this paper, we argue that evidence of extensive fractional crystallization is not apparent in the ophiolite. Furthermore, PP-322 the most siliceous rock of the analyzed suite does not appear to be a late-stage differentiate but rather a sample of an independent magma type.

CHEMICAL CHARACTERISTICS OF MAJOR ROCK TYPES

Tectonitic Peridotite

As typical of many Cordilleran alpine-type ultramafic rocks (Thayer and Himmelberg, 1968; Himmelberg and Loney, 1973; Bailey and Blake, 1974), the analyzed tectonitic peridotite is characterized by low content of Al_2O_3 and CaO. The low CaO

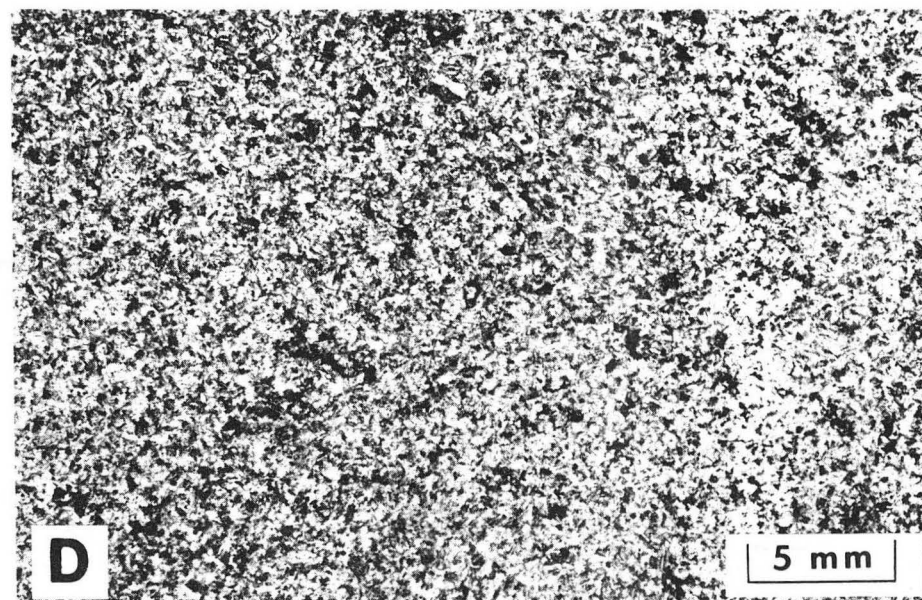
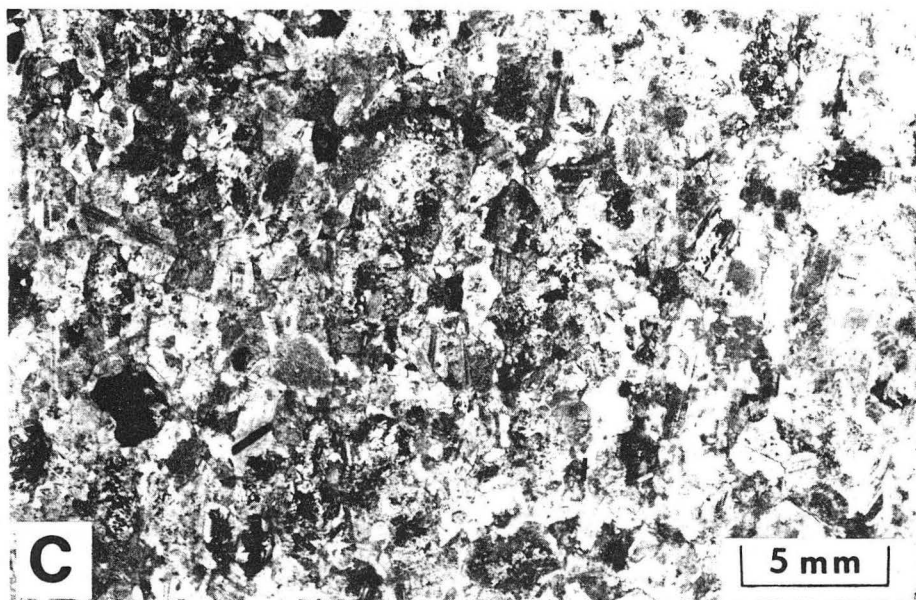
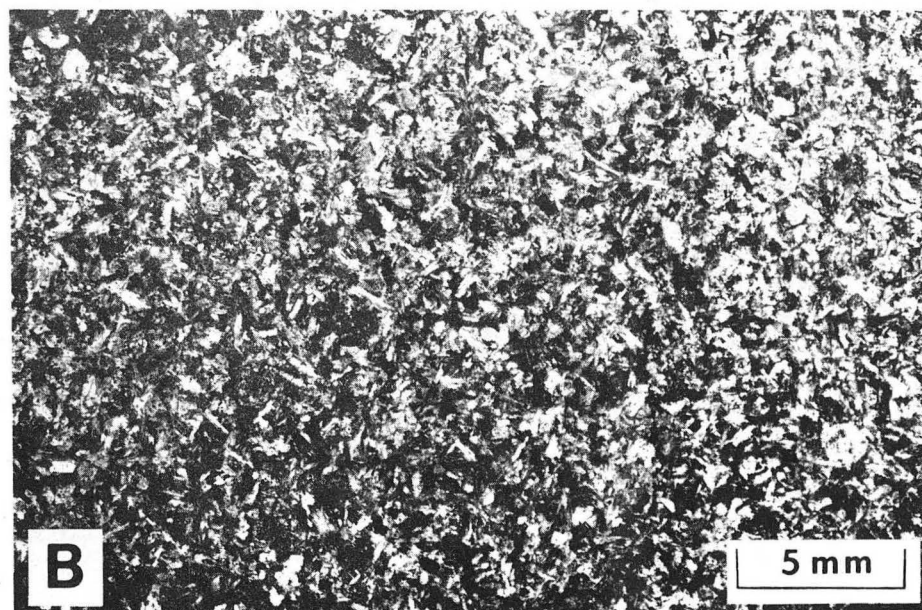
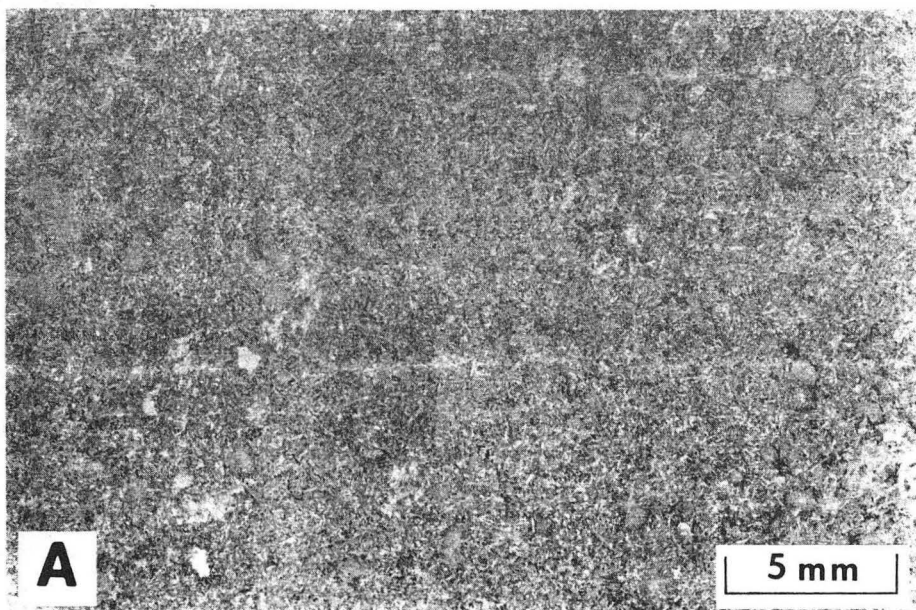


Figure 3. Photomicrographs of analyzed rocks from the Preston Peak ophiolite. (A) Porphyritic metadiabase (PP-384) with chlorite pseudomorphs after pyroxene. (B) Relict pyroxene-bearing metadiabase (PP-376). (C) Metagabbro (PP-374) with relict hypidiomorphic-granular texture. (D) Biotite quartz diorite (PP-322).

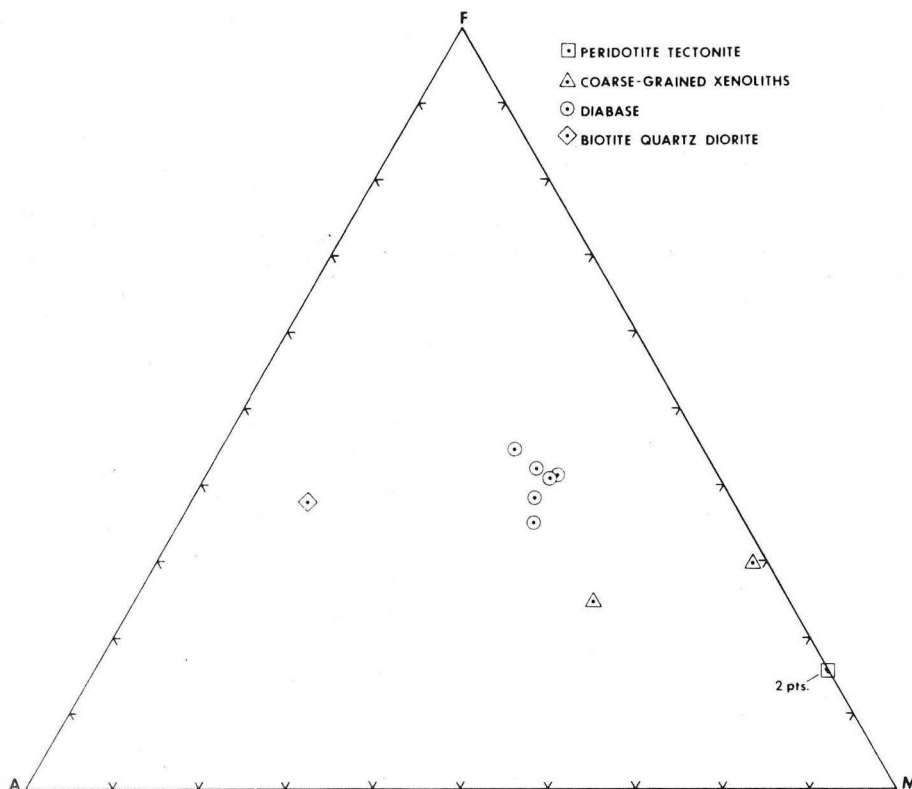


Figure 4. AFM diagram showing peridotite tectonite, coarse-grained xenoliths, diabase, and biotite quartz diorite.

as well as the SiO_2 values also reflect the extensive serpentinization that the rock has undergone (Coleman, 1967). When normalized on a water-free basis SiO_2 contents are 45.0 and 45.7 for specimens PP-225 and PP-401A, respectively. Normalized total iron to MgO ratios are .188 and .190, respectively. Both samples are very depleted in alkalis as well as almost all trace elements except Cr and Ni. Minor amounts of Co, V, Zn, and Sc are also present. These characteristics are in accord with the concept that the peridotite tectonite at the base of ophiolite is a residue remaining after the removal of basaltic liquids during partial melting of primitive upper mantle rock (Coleman, 1971).

Diabase and Diabase Breccia

The restricted range in silica content ($\sim 48\text{--}52.5$) and the high $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios are the most distinctive major element characteristics of this rock group. The sodic-rich character of these rocks coupled with a wide CaO range (5.8–10.3) are earmarks of mafic rocks which have undergone extensive spilitic alteration. This spilitic character, therefore, makes comparison with unaltered volcanic suites difficult if not impossible for almost all the major elements. An exception is TiO_2 , which is thought to be unaffected by low temperature alteration (Pearce and Cann, 1971, 1973). As indicated in table 1 and displayed on figure 7, TiO_2 abundances fall within a restricted range (0.70–1.20). Significantly, this range is clearly within the TiO_2 distribution in immature island arc tholeiite (Jakes and Gill, 1970; Ewart and Bryan, 1972; Miyashiro, 1974) but is lower than the average ocean-floor (abyssal) tholeiite (Cann, 1971).

Another element considered to be useful in distinguishing prototectonic igneous suites is Cr (Bloxam and Lewis, 1972). The average Cr value for the diabase-diabase breccia unit is 135 ppm (range—34 to 209 ppm). This value is less than typical abyssal tholeiite (e.g., Engel and others, 1965, Smewing and others, 1975), and is closer to values given for average island arc

tholeiite and calc-alkali basalt (Jakes and White, 1972). When these data are combined with the coexisting TiO_2 abundances, the plotted points form a trend from island arc tholeiite toward abyssal tholeiite (figure 7).

Field relations and petrography of the diabase suite indicate no evidence of major differentiation in place. Furthermore, the flat rare earth element (REE) distribution patterns characteristic of the diabase group (figure 8) also suggest that high-level fractionation was minimal. Such REE patterns are often considered characteristic of primitive magmas developed from partial melting (20 to 30%) of mantle material (Gast, 1968; Kay and others, 1970). However, the coarse-grained xenoliths and indications of slight Fe and/or Si enrichment suggest that differentiation was locally operative during part of the development of the mafic complex. We reconcile these two opposing views by arguing that the bulk of the mafic complex was built of primitive and unfractionated magma; however, local ponding of magma is responsible for the scattered evidence of differentiation. Apparently these small magma chambers were disrupted by younger intrusive basalt or occasionally tapped of their fractionated liquids. Judging from the paucity of coarse-grained xenoliths in the mafic complex, most of the magmas intruded high into the crust and cooled quickly as flows, dikes, or sills.

Many studies (e.g., Frey and others, 1968; Philpotts and others, 1969; Kay and others, 1970) indicate that REE abundances are unaffected by post-magmatic alterations and, therefore, are useful in characterizing magma types. However, substantial data (e.g., Jakes and Gill, 1970; Ewart and Bryan, 1972) also indicate that the flat REE patterns are characteristic of basalt from contrasting tectonic environments (i.e., abyssal tholeiite vs. island arc basalt). Therefore, although our REE data unquestionably suggest a tholeiitic protolith for the metadiabase, we cannot unequivocally indicate the site of origin. Kay and Senechal (1976) likewise pointed out this problem in their analysis of REE data on basalt from the Troodos ophiolite complex.

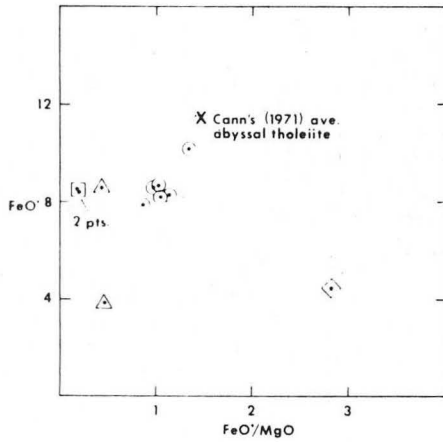


Figure 5. FeO^* versus FeO^*/MgO plot showing peridotite tectonite, coarse-grained xenoliths, diabase, and biotite quartz diorite. Rock types are designated with the same symbols as in figure 4. Cann's (1971) average abyssal tholeiite is shown for comparison.

Figure 6. SiO_2 versus FeO^*/MgO plot showing peridotite tectonite, coarse-grained xenoliths, diabase, and biotite quartz diorite. Rock types are designated with the same symbols as in figure 4. Cann's (1971) average abyssal tholeiite is shown for comparison.

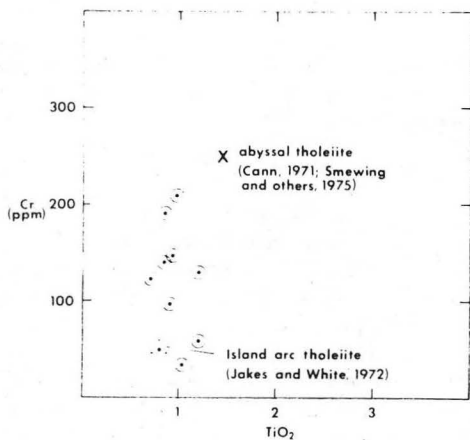
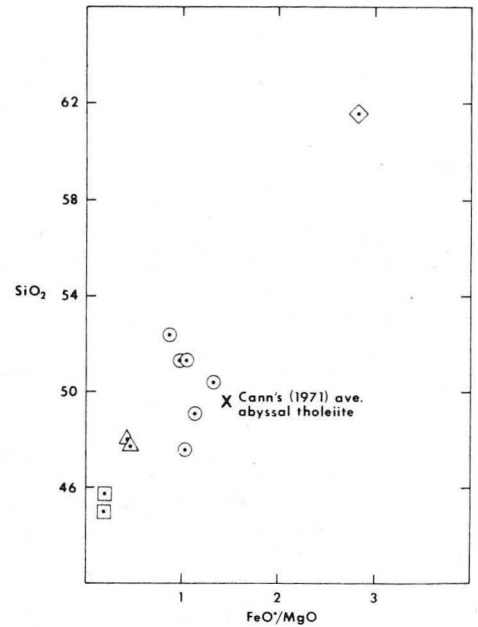


Figure 7. Cr versus TiO_2 plot for Preston Peak diabasic rocks, circles. Average values for abyssal tholeiite and island arc tholeiite are plotted for comparison.

Coarse-Grained Xenoliths

Although only two specimens of xenoliths were analyzed, the xenoliths appear to constitute a chemically distinct group within the ophiolite suite. In regard to major element content this is especially apparent in the AFM diagram (figure 4) on which the xenolith specimens plot toward the MgO corner. Likewise in figures 5 and 6, the xenoliths have distinctly lower FeO^*/MgO ratios than the other rocks with the exception of the peridotite tectonite. These data coupled with petrography and field occurrence suggest that these rocks are cumulates brought up by younger basaltic magmas.

The clinopyroxene specimen (PP-382) in comparison to the alpine-type peridotite tectonites has distinctly higher Al_2O_3 , CaO and TiO_2 contents, reflecting the abundance of clinopyroxene. The gabbro xenolith (PP-374) in comparison to the diabasic rocks is characterized by a higher Al_2O_3 and CaO content and lacks the high Na_2O/K_2O ratio.

The low REE contents of these rocks is consistent with a cumulate origin; however, the analyzed specimens are not as depleted in REE as many gabbros from the Troodos complex (Kay and Senechal, 1976) or from the Pindos suite (Montigny and others, 1973). Perhaps as suggested by Kay and Senechal (1976) for a similar Troodos gabbro, the Preston Peak xenoliths are accumulates which once contained a significant quantity of trapped liquid. According to Kay and Senechal (1976), the trapped liquid would contribute most of the REE. Therefore, because REE patterns from xenoliths and diabase are crudely parallel, we argue that both groups were derived from magmas with similar REE concentrations.

Biotite Quartz Diorite

The biotite quartz diorite specimen (PP-322) is the most siliceous of the analyzed rocks from the Preston Peak ophiolite, and it has the highest FeO^*/MgO ratio. It is characterized by a high Na_2O/K_2O ratio but also contains substantial CaO. These characteristics are typical of rocks of the keratophyre-trondhjemite suite, often an integral part of many ophiolite complexes (Coleman and Peterman, 1975).

The REE pattern (figure 8) for the quartz diorite (PP-322) shows enrichment in light REE (~50 times the Leedy chondrite) with a depletion of heavy REE. No Eu anomaly is apparent. This pattern is strikingly different from Troodos plagiogranite (Coleman and Peterman, 1975; Kay and Senechal, 1976) and appears similar to fractionated REE patterns in the Precambrian Saganaga tonalite and the Mesozoic Craggy Peak trondhjemite as reported by Arth and Hanson (1972). According to Arth and Hanson (1972), such patterns which show a depletion of heavy REE are consistent with an origin involving partial melting of amphibolite or eclogite at depths greater than 30 km. Barker and others (1976) have also documented Precambrian trondhjemitic rocks ($Al_2O_3 > 15$ wt%) with similar REE patterns. They favor a partial melting mechanism at greater than 50 km but point out that rocks with similar REE patterns could be produced by fractional crystallization at crustal levels. Regardless of the origin, our data indicate that leucocratic rocks associated with ophiolite may be more variable than previously emphasized (e.g., Coleman and Peterman, 1975).

EVALUATION OF THE PETROGENESIS OF THE PRESTON PEAK OPHIOLITE

Despite substantial new data on ophiolite in recent years, the site of origin of this enigmatic rock sequence remains a con-

troversial problem in petrology. At present, three models are commonly considered for the origin of ophiolite complexes:

1. They formed at an oceanic ridge in a major ocean basin (Coleman, 1971; Dewey and Bird, 1971; Moores and Vine, 1971).
2. They formed in a marginal basin behind an active island arc (Dewey and Bird, 1971; Blake and Jones, 1974; Dewey, 1974).
3. They represent the immature stage (i.e., the basal part) of an island arc complex (Ewart and Bryan, 1972; Miyashiro, 1973, 1975).

We concur with Kay and Senechal (1976) that whole rock chemical analyses including major and trace element data cannot unequivocally discriminate between the above three models. However, we hasten to add that our chemical data do indicate certain constraints for each lithologic group of the Preston Peak ophiolite:

1. The ultramafic tectonite is chemically and texturally distinct from all other rocks in the ophiolite. Its depleted character coupled with its metamorphic texture suggests that it is residual mantle rock.
2. The diabasic rocks show only minor fractionation and probably represent primary magmas that have crystallized in a subvolcanic environment. The diabase breccia unit is chemically similar to the massive, nonfragmental diabase and probably is a pyroclastic equivalent.
3. Chemical data coupled with field occurrence and petrography suggest that the coarse-grained xenoliths are cumulates rafted up by younger basaltic magmas. REE data indicate that they may be derived from magmas similar in composition to their enclosing diabasic hosts.
4. The fractionated REE pattern of the biotite quartz diorite is analogous to certain Na-rich calc-alkaline rocks. Therefore, this rock type probably represents a distinct magma type chemically unrelated to the mafic rocks of the ophiolite.

Detailed evaluation of the genesis of the Preston Peak ophiolite as deduced from field data is presented elsewhere (Snook, in press). We will not reiterate these statements but emphasize that many observations seem incompatible with known data and postulated models for modern spreading centers either in marginal or major ocean basins. Some of the key points are:

1. Metadiabase dikes in the basal peridotite appear to be feeders that in part nourished the overlying subvolcanic mafic complex. These dikes form sharp contacts with chilled selvages adjacent to tectonic ultramafic rocks. There is no evidence, such as magmatic segregations (Dickey, 1970), that the mafic magmas were derived from the surrounding depleted peridotite.
2. The tectonite fabric (i.e., metamorphic recrystallization) of the basal peridotite predates most of the mafic magmatism of the Preston Peak ophiolite. An exception is the amphibolite masses which occur in serpentinite melange along the basal fault contact (figure 2). These rocks may be remnants of oceanic crust upon which the mafic complex was built.
3. Mafic-ultramafic cumulate rocks, characteristic of many ophiolites and commonly postulated to be abundant at spreading centers (e.g., Moores and Jackson, 1974; Jackson and others, 1975), are scarce in the Preston Peak ophiolite. Perhaps this feature is indicative of a fast spreading ridge (Moores and Vine, 1971), but alternatively it may represent a fundamental difference in site of origin.
4. Recent investigations of Mid-Atlantic Ridge volcanism (Moore and others, 1974) indicate that a preponderance of pillow lava erupted quietly from fissure vents. Pillow lava is not abundant in the Preston Peak ophiolite; in fact pyroclastic mafic breccia is common and probably is evidence of local explosive volcanic centers.

These observations as well as certain enumerated aspects of the petrochemistry suggest to us that the young island arc model is a possible alternative. We can only speculate on what factors controlled the initiation of the proposed Preston Peak arc but perhaps late Paleozoic geanticlinal buckling as postulated by Burchfiel and Davis (1975) for the Trinity ophiolite is a viable explanation. Nevertheless, upper mantle peridotite tectonite and metamorphosed oceanic crust (amphibolite) served as the foundation upon which the arc was built. Initial magmas were basaltic (island arc tholeiite) and as the volcanic pile grew, younger magmas often didn't reach the surface but formed an extensive dike-sill complex. The fluidity and probable low pH_2O of the magma precluded extensive pyroclastic deposits during this

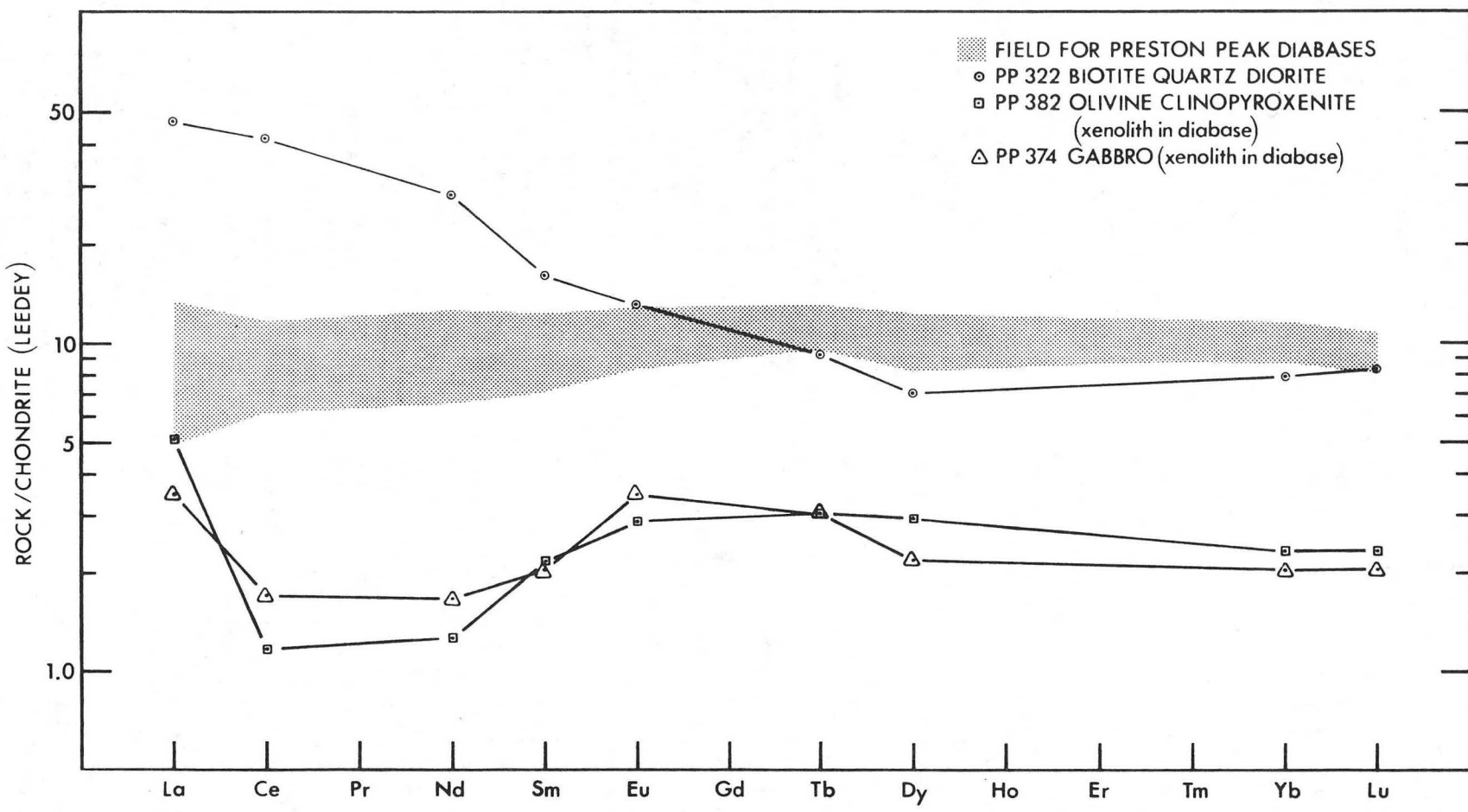
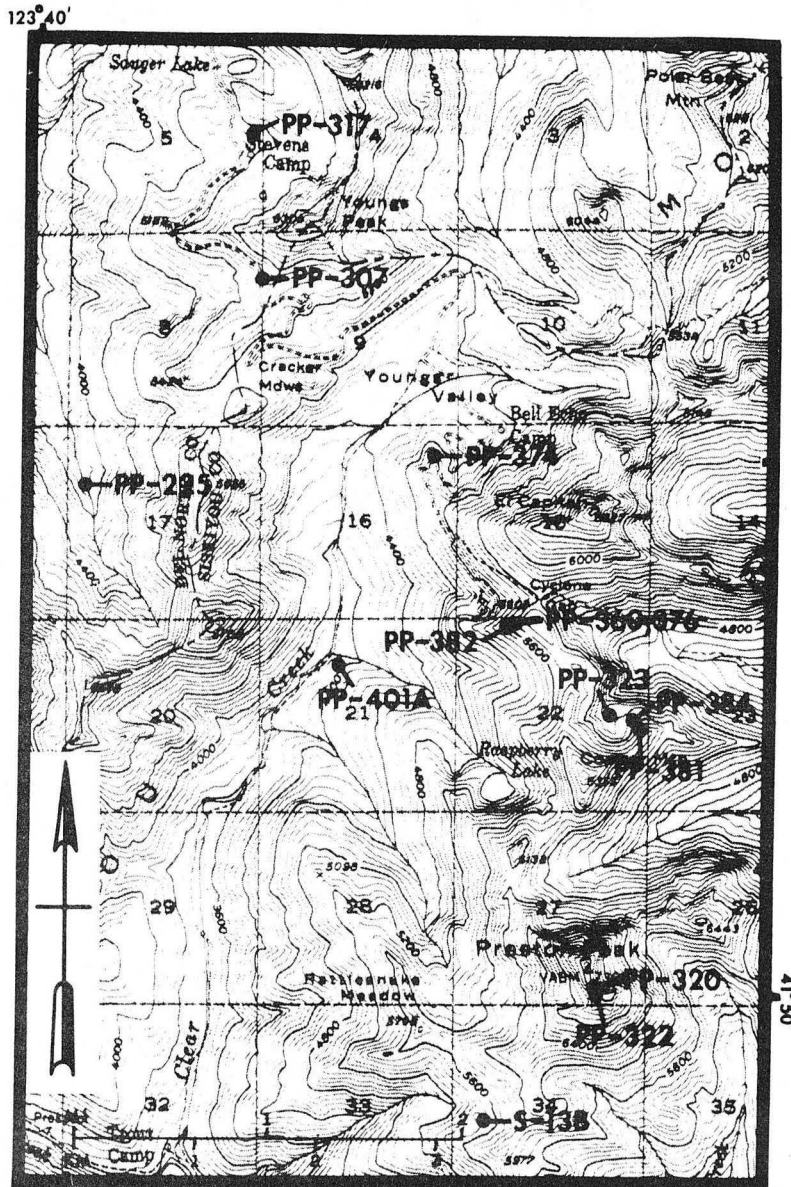


Figure 8. Leedeey normalized REE patterns for coarse-grained xenoliths, diabases, and biotite quartz diorite from the Preston Peak ophiolite.



Topography from Preston Peak 15min. quadrangle (1956)--U.S.G.S.

Figure 9. Map showing location of chemically analyzed samples.

stage, but local vent breccia occasionally developed. Small magma chambers formed but commonly were disrupted by younger intrusive magmas. As the arc matured, intermediate magmas represented by Na-rich quartz diorite became part of the igneous complex. Overlying coarse-grained, immature sedimentary rocks suggest that local relief became progressively more prominent as the arc matured. Intermediate hypabyssal and plutonic fragments are phenoclasts in these detrital rocks suggesting that calc-alkaline rocks in the Preston Peak ophiolite were more extensive than the present exposures indicate (Snoke, in press).

The proposed island arc model appears consistent with the available data on the Preston Peak ophiolite, and we know of no regional data to preclude our interpretation. In fact, we note that apparent calc-alkaline igneous rocks appear to form the upper parts of several other ophiolites in the Klamath Mountains (e.g., the Josephine and Trinity ophiolites). The Rogue Formation, described by Wells and Walker (1953) as predominantly meta-

andesitic flows and breccias, structurally and locally depositionally overlies in the Josephine ophiolite in southwestern Oregon. Analogous volcanic rocks forming the base of the Galice Formation appear to overlie the southern part of the Josephine ophiolite in California (Cater and Wells, 1953). In the southeastern Klamath Mountains, the southern part of the Trinity ophiolite is overlain by the Copley Greenstone, metabasalt to meta-andesite flows and breccias, and the Balaklala Rhyolite, predominantly keratophyre to quartz keratophyre flows and pyroclastic rocks (Kinkel and others, 1956). Therefore, we suggest that these ophiolites should be investigated as possible remnants of the underpinnings of an island arc complex.

ACKNOWLEDGMENTS

We thank Akiho Miyashiro and Eldridge M. Moores for their critical reviews of the manuscript.

Table 1. Partial chemical analyses of Preston Peak ophiolitic rocks (in percents).

	Tectonitic Peridotite		Olivine Clino- pyroxenite	Gabbro	Diabase						Quartz Diorite
	PP-225	PP-401A	PP-382	PP-374	PP-317	PP-320	PP-323	PP-360	PP-376	PP-384	PP-322
SiO	39.7 (45.0) [#]	38.8 (45.7)	48.0	47.8	51.3	50.4	51.3	47.9	49.1	52.4	61.6
TiO ₂ [*]	<.02 (.02)	<.02 (.02)	.12	.20	.97	1.20	.92	.90	.83	.85	.35
Al ₂ O ₃	.19 (.22)	.39 (.46)	4.8	21.8	15.2	15.4	15.2	16.8	15.6	14.9	17.3
FeO [*]	7.5 (8.5)	7.1 (8.4)	8.6	3.9	8.6	10.2	8.2	8.7	8.3	7.9	4.5
MgO	40.0 (45.3)	37.6 (44.3)	19.8	8.3	8.8	7.7	7.9	8.5	7.4	9.2	1.6
CaO	.81 (.92)	.99 (1.2)	13.2	12.5	7.9	7.0	10.3	10.2	10.3	5.8	5.3
Na ₂ O [*]	.02 (.02)	.02 (.02)	.53	2.8	5.0	4.4	3.8	3.7	3.9	5.3	4.7
K ₂ O	n.d.	n.d.	.01	.73	.08	.56	.11	.23	.12	.15	1.1
Total	88.24	84.92	95.06	98.03	97.85	96.86	97.73	96.93	95.55	96.50	96.45

FeO = Total iron.

n.d. = not detected.

= The number in parentheses is a recalculated value so as to sum 100%.

* = Instrumental neutron activation analysis (Analyst - H.R. Bowman) otherwise X-ray fluorescence (Analyst - A.J. Hebert).

Table 2. Trace element compositions of Preston Peak ophiolitic rocks (in parts per million).

	Tectonic Peridotite		Olivine Clino- pyroxenite	Gabbro	Diabase								Diabase Breccia	Quartz Diorite
	PP-225	PP-401A	PP-382	PP-374	PP-307	PP-317	PP-320	PP-323	PP-360	PP-376	PP-384	S-13B	PP-381	PP-322
Ba	<12	9±7	30±11	53±9	55±11	68±13	70±13	64±12	89±13	73±13	56±12	11±11	80±12	437±24
Co	113±2	102±1	92±1	26.4±.4	38.4±.6	38.6±.6	38.1±.6	38.7±.6	41.3±.6	38.5±.6	37.3±.6	43.0±.7	45.3±.7	2.5±.1
Cr	2602±91	2663±93	4298±150	649±23	34±3	209±8	59±3	147±6	98±4	140±5	191±7	123±5	130±5	2.3±1.3
Hf	<.12	<.12	.12±.10	.23±.07	1.9±.2	1.8±.2	2.1±.2	1.7±.1	1.9±.2	1.8±.2	1.6±.1	1.3±.1	2.0±.2	3.5±.3
Mn	1547±49	859±27	1205±38	573±18	1359±43	1276±41	1419±46	1302±41	1444±47	1244±40	1093±35	1250±40	7717±234	1008±32
Ni	2262±176	2259±177	914±75	165±20	32±18	101±21	55±20	72±18	75±18	74±21	84±19	84±19	124±20	5±9
Sc	9.4±.2	9.2±.2	53.0±.9	32.5±.5	39.0±.6	41.3±.7	40.6±.7	39.6±.6	40.2±.6	40.6±.7	36.7±.6	42.2±.7	38.4±.6	7.4±.1
V	64±27	44±11	168±30	119±26	279±45	291±51	405±69	295±51	288±53	324±54	291±49	322±59	173±44	66±19
Zn	69±12	74±12	54±12	13±6	63±11	94±15	82±13	64±11	98±15	86±14	60±11	64±12	115±17	88±12
Ba	<.5	<.5	1.9±.7	1.3±.6	3.9±.7	3.4±.5	3.7±.5	2.6±.5	6.4±.5	4.4±.5	2.9±.5	1.8±.5	5.0±.5	17.7±.9
Ce	<1.0	<1.0	1.1±.8	1.7±.5	10.7±.6	9.4±.4	9.4±.6	8.3±.5	14.5±.6	12.0±.6	7.6±.5	6.0±.5	11.5±.6	40.9±1.2
Nd	<1.4	<1.6	.9±1.2	1.2±.8	8.6±1.3	7.2±1.5	8.7±1.5	5.7±1.3	7.3±1.5	7.9±1.5	6.2±1.3	4.7±1.2	9.1±1.4	20.1±2.0
Sm	<.006	<.006	.50±.01	.47±.01	2.60±.06	2.29±.05	2.60±.05	2.13±.04	2.72±.06	2.44±.05	2.01±.04	1.63±.03	2.84±.06	3.69±.08
Eu	<.02	<.02	.25±.01	.30±.01	1.02±.03	.87±.03	1.00±.03	.90±.03	.95±.03	.92±.03	.77±.02	.56±.06	1.12±.03	1.13±.03
Tb	<.08	<.08	.18±.07	.18±.05	.61±.06	.63±.06	.72±.06	.65±.06	.65±.06	.61±.06	.57±.06	.56±.06	.77±.06	.55±.04
Dy	<.3	<.2	1.14±.15	.86±.12	4.09±.23	3.75±.23	4.33±.25	3.58±.21	3.52±.23	3.47±.21	3.27±.20	3.19±.20	4.79±.68	2.73±.19
Yb	<.04	<.04	.58±.03	.51±.02	2.62±.06	2.32±.06	2.70±.06	2.30±.06	2.22±.05	2.19±.05	2.16±.05	2.23±.05	2.90±.07	1.97±.05
La	<.02	<.02	.09±.02	.08±.01	.37±.02	.35±.02	.37±.02	.34±.02	.33±.02	.30±.02	.31±.02	.33±.02	.42±.02	.32±.02
REF	3.4	3.5	6.64	6.60	34.51	30.21	33.52	26.50	38.59	34.23	25.79	21.16	38.44	89.09

Analyst: H.R. Bowman (instrumental neutron activation analysis)

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