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BEAR MOUNTAIN IGNEOUS COMPLEX, KLAMATH MOUNTAINS, CALIFORNIA: AN ULTRABASIC TO SILICIC CALC-ALKALINE SUITE

by

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ABSTRACT

The Bear Mountain igneous complex, Klamath Mountains, California, can be divided into distinct lithologic suites (order according to apparent relative age): 1) satellitic masses of clinopyroxene-rich ultramafic and gabbroic rocks with subordinate dunite and hornblende-plagioclase pegmatoid; 2) two-pyroxene-biotite diorite and monzodiorite; 3) heterogeneous hornblende-rich rocks varying from gabbro to diorite; 4) leucocratic rocks, chiefly consisting of biotite tonalite and granodiorite; and 5) late dikes (mafic to felsic). Elongate masses of unit (1) flank a composite pluton consisting of units (2-4), while the late dikes (unit 5) intrude the adjacent country rocks. The rocks of the complex invaded an ophiolite allochthon during the Late Jurassic Nevadan orogeny, and well-defined contact aureoles surround the complex. Lower greenschist facies rocks, chiefly metabasalt, impure siliceous metasedimentary rocks, and serpentinitized peridotite, have been dynamothermally metamorphosed to mineral assemblages indicative of hornblende-hornfels facies and locally pyroxene-hornfels facies. The emplacement of the igneous complex was chiefly by forcible shouldering aside, although local tectonic features such as faults in the ophiolite allochthon were instrumental in the emplacement history.

The ultramafic and gabbroic rocks are interpreted as crystal cumulates of a fractionated basaltic magma. Mineral compositions and whole-rock chemical characteristics of the proposed cumulates suggest that the Mg/Fe ratio of the parental basaltic liquid was high. The activity of silica was low, while water vapor pressure apparently increased through time until it was moderately high during the late magmatic stage. These cumulates were subsequently remobilized during lateral tectonic compression and emplaced higher in the

crust as hot, semisolid aggregates.

A diverse array of data including pyroxene compositions, major-, minor-, and rare-earth-element abundances, and field relations suggest that the two-pyroxene-biotite diorite/monzodiorite unit was consanguineous with the clinopyroxene-rich ultramafic and gabbroic rocks. The diorite/monzodiorite unit, therefore, is an intermediate differentiate of an early primitive basalt. Furthermore, major-, trace-, and rare-earth-element data characteristic of the diorite/monzodiorite unit indicate strong similarities to low-Si andesite and clearly suggest a calc-alkaline affinity.

Age relations indicate that the hornblende-rich and leucocratic units are younger and represent the intrusion of other magmas into the same igneous locus. Petrographic and geochemical data from the hornblende-rich unit suggest crystallization from hydrous magmas similar in composition to high-Al basalt and basaltic andesite. The leucocratic suite, consisting chiefly of calc-alkaline tonalitic rocks, is similar to other quartz-rich felsic rocks widespread throughout the Klamath Mountains-western Sierra Nevada. The available petrographic and geochemical data are consistent with formation of these rocks by either fractional crystallization of a wet basaltic magma or partial melting of amphibolite or eclogite.

The Bear Mountain igneous complex is an example of a diverse but distinctive association of ultrabasic to silicic rocks which characterize numerous plutonic complexes in the Klamath Mountains - western Sierra Nevada. These intrusive complexes invade older ensimatic rocks and appear to define the roots of a complex, Middle to Late Jurassic calc-alkaline magmatic arc. The ultramafic and gabbroic rocks characteristic of this plutonic association are similar to Alaskan-type complexes but differ in detail. More significantly, these rocks are important clues to the composition of early magmas as well as the complex processes operative in reservoirs that form the core of calc-alkaline magmatic centers.

INTRODUCTION

Ultramafic rocks in Phanerozoic orogenic belts are commonly classified as members of either two distinct associations: alpine-type (including ophiolitic assemblages) or Alaskan-type (Irvine, 1974). Several reviews (Wyllie, 1969, 1970; Jackson and Thayer, 1972; Naldrett and Cabri, 1976) have attempted to summarize critical differences between these associations. For example, the ultramafic rocks of Alaskan-type complexes commonly include varieties characterized by an abundance of calcic clinopyroxene. In these lithologies hornblende is also common; plagioclase and orthopyroxene are rare or absent (Irvine, 1974). Magnetite is locally an important varietal mineral. Although the ultramafic rocks of the ophiolite assemblage are variable, harzburgite-dunite tectonite is a widespread and ubiquitous element (Coleman, 1977). Clinopyroxene-rich ultramafic rocks are scarce and typically restricted to the cumulate portion of the plutonic suite.

The commonly inferred emplacement histories of the associations are another clearly contrasting aspect. The details of the emplacement histories of Alaskan-type complexes vary depending on locality, but most investigators believe that the intrusion of a hot, ultramafic mush was a key event in the evolution of these complexes (e.g., Findlay, 1969; Murray, 1972; Irvine, 1974). Many studies have documented the final, relatively cold, tectonic emplacement of serpentinitized alpine-type peridotite-dunite, although an earlier, high-temperature, ductile deformation history has also commonly been demonstrated for these bodies (e.g., Loney and others, 1971). In regard to petrogenesis of these contrasting orogenic ultramafic suites, Taylor and Noble (1960) and Taylor (1967) have argued that ultramafic magmas, the products of fractional fusion of upper mantle rock, were involved in the evolution of Alaskan-type complexes. In contrast, recent studies (e.g., Murray, 1972; Irvine, 1974) have favored a basic melt as the parental magma of the Alaskan-type complexes and consequently conclude that the ultramafic rocks are cumulates derived by fractional crystallization. The ultramafic tectonite of ophiolite assemblages, on the other hand, is commonly considered to be depleted or partially depleted upper mantle rock, while the cumulate ultramafic rocks are derived by fractional crystallization of a basaltic magma.

Nevertheless, other ultramafic rock associations have locally been recognized in Phanerozoic orogenic belts, but distinctive criteria are not well established. Among the orogenic ultramafic suites is the association of ultramafic-mafic rocks with felsic plutonic complexes (Joplin, 1959; Mullan and Bussell, 1977). Some of these occurrences, for example, the ultramafic-mafic rocks near Emigrant Gap, Sierra Nevada, California, have been considered a variant of the Alaskan-type complexes (James, 1971). Other occurrences (e.g., the ultramafic to mafic rocks associated with granitic intrusions in Scotland and Donegal) have been classified as members of the appinite suite (e.g., Nockolds, 1941; Joplin, 1959; French, 1966; Hall, 1967; Pitcher and Berger, 1972).

Clinopyroxene-rich ultramafic rocks[1], that commonly vary from olivine clinopyroxenite or hornblende-olivine clinopyroxenite to wehrlite, and subordinate dunite form numerous small intrusive masses in the Klamath Mountains-Sierra Nevada batholithic belt (Fig. 1). These ultramafic masses commonly flank large felsic batholiths (for example, the Emigrant Gap complex - James, 1971), but isolated masses are also known (Lower Coon Mountain, Gasquet quadrangle, Klamath Mountains, California - Cater and Wells, 1953). Associated with these ultramafic rocks are a distinct suite of mafic to intermediate rocks including:

'ig. 1

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[1] Nomenclature according to IUGS classification.

olivine gabbro, gabbro, two-pyroxene gabbro, two-biotite diorite/monzodiorite, and hornblende gabbro/diorite. Associated, but always younger, silicic rocks apparently vary along the belt; in the Klamath Mountains tonalite-trondhjemite is common, while in the Sierra Nevada more potassic rocks such as granodiorite and granite as well as tonalite are present. The detailed delineation of the petrogenetic history of these ultramafic-mafic intrusions and associated rocks is important in regard to comparisons with other orogenic ultramafic suites (i.e., alpine-type and Alaskan-type).

The Bear Mountain igneous complex, west-central Klamath Mountains northwestern California (Fig. 1 and 2), is a well-exposed and lithologically diverse example of one of these ultrabasic to silicic complexes. The purpose of this article is to summarize field, petrographic, and geochemical data on the Bear Mountain complex. The conclusions derived from the study of this complex are, at least in part applicable to the belt as a whole and may relate to analogous plutonic suites in other orogenic terranes. ig. 2

REGIONAL GEOLOGIC SETTING

The Klamath Mountains province consists of an imbricate sequence of arcuate lithologic belts bounded by east-dipping regional thrust faults. The metaigneous and metasedimentary rocks that compose these belts are chiefly fragments of ophiolites, island-arc complexes, and polymict melanges juxtaposed as a tectonic collage (Davis and others, 1978; Hamilton, 1969, 1978). Despite complex stratigraphy and structure, distinct subprovinces and terranes have been recognized throughout the Klamath Mountains (Irwin, 1960, 1966, 1972, 1977a,; Davis, 1966, 1968; Davis and others, 1965, 1978). Reconnaissance studies originally suggested a subdivision into four terranes on the basis of age and metamorphic and tectonic history. From east to west these are the eastern Klamath belt, central metamorphic belt, western Paleozoic and Triassic belt, and western Jurassic belt (Irwin, 1960, 1966). More detailed studies indicated that further subdivision is necessary, especially of the western Paleozoic and Triassic belt (Irwin, 1972). Furthermore, many new fossil localities within the western Paleozoic and Triassic belt, especially radiolarian-bearing cherts and tuffs, have yielded Jurassic ages rendering the original designation obsolete (Irwin, 1977a; Irwin and others, 1977; Irwin and others, 1978). The recent data now suggest that during the Jurassic Period the paleogeography of this segment of the western North American Cordillera was very complex: the area apparently consisted of multiple coeval volcanic arcs and associated basins, perhaps analogous to the present-day Indonesian region (Irwin and others, 1977; Hamilton, 1977, 1978). During the Middle to Late Jurassic, intrusive plutonic rocks extensively invaded the region but are especially concentrated in the central metamorphic and so-called western Paleozoic and Triassic belt. Some of these plutons (e.g., Ironside Mountain batholith) were emplaced into coeval volcanic rocks (Charlton, 1978). Other plutonic complexes, such as the Bear Mountain igneous complex, invaded older metaigneous and metasedimentary terranes (Snoke, 1977). According to Davis and others (1978), a broad, complex magmatic arc spread across the Klamath Mountains region during the Middle to Late Jurassic; presumably a convergent plate boundary with an east-dipping subduction zone lay to the west. The plutonic igneous rocks that are manifestation of this Middle to Late Jurassic arc range in composition from ultramafic to granitic but quartz diorite and tonalite are predominant (Hotz, 1971). The intrusions are commonly elongated parallel to the lithologic belts suggesting a tectonic influence on their emplacement. Both structural and mineralogical contact metamorphic effects are conspicuous around many of the intrusions.

GENERAL FIELD RELATIONS

Lithologic units and age relations

The Bear Mountain igneous complex can be divided into four distinct, mappable lithologic units (order according to apparent relative age): 1) satellitic masses of clinopyroxene-rich ultramafic and associated gabbroic rocks, 2) two-pyroxene-biotite diorite/monzodiorite, 3) heterogeneous hornblende-rich rocks varying from gabbro to diorite, and 4) a late felsic suite which chiefly includes biotite tonalite but also minor granodiorite and granite.

Elongate clinopyroxene-rich ultramafic-mafic intrusions (unit 1), designated the Clear and Cedar Creek bodies and the Blue Ridge body, flank a gabbro-diorite pluton (units 2, 3, and 4) on the east and west, respectively (Fig. 2). Maxson (1931) originally applied the name Bear Mountain batholith to the pluton. This name is retained, but here modified to Bear Mountain igneous complex so as to include the satellitic ultramafic-mafic intrusions. Throughout this report, the term Bear Mountain pluton refers to the main gabbro-diorite intrusion.

The ultramafic and associated rocks appear to be the oldest unit in the igneous complex. They are intruded by hornblende-rich and leucocratic rocks, and inclusions of ultramafic-mafic rocks locally occur in these younger intrusive phases. The age relations between the ultramafic-mafic rocks and the two-pyroxene-biotite diorite/monzodiorite unit is less certain. At no locality within the study area are these two units in contact. However, ultramafic dikes, petrographically similar to the Blue Ridge intrusion, are contact metamorphosed and deformed in the aureole of the Bear Mountain pluton near the western margin of the intrusion (Fig. 2 - locality A, Fig. 3). This superimposed metamorphism and deformation clearly is related to the emplacement of younger magmas. Amphibolitized pyroxene-biotite diorite is the nearest plutonic unit intruding the country rock at this locality. These field relations, therefore, suggest that the emplacement of the ultramafic dikes pre-dated the pyroxene-biotite diorite unit as well as the other units of the Bear Mountain pluton. Fig. 3

Age relations between the two-pyroxene-biotite diorite/monzodiorite unit and the hornblende-rich rocks are commonly difficult to establish clearly. However, at a few localities hornblende-rich plutonic rocks intrude the pyroxene-biotite diorite/monzodiorite unit. A clear example is exposed on the ridge southwest of Bück Lake (Fig. 2 - locality B). At this locality the following data indicate that the hornblende-rich rocks are younger than the pyroxene-biotite diorite/monzodiorite unit:

1. dikes and intrusive masses of hornblende-biotite quartz diorite within two-pyroxene-biotite diorite;
2. intrusive breccias (i.e., xenolithic clasts of two-pyroxene-biotite diorite within hornblende-biotite quartz diorite);
3. local contact metamorphism of the two pyroxene-biotite diorite;
4. large xenolith (approximately 3 x 8 m) of recrystallized two-pyroxene-biotite diorite within hornblende-biotite quartz diorite.

Age relations between the felsic rock suite and the other plutonic units which comprise the Bear Mountain igneous complex consistently indicate that the felsic rocks are younger. However, mafic dikes intrude all the major plutonic units and represent a widespread, late stage magmatic event.

K-Ar radiometric dates for two rocks from the Bear Mountain pluton were determined in the laboratories of the U.S. Geological Survey in Menlo Park (Jarel Von Essen, written comm., 1970). The analytical data are summarized in Table 1.

Table 1

Contact metamorphic effects

Contact metamorphic effects exclusively related to the emplacement of the Blue Ridge intrusion occur continuously along its western contact. Granular textures, attributed to contact metamorphism, locally extend outward for up to 300 m from the ultramafic-country rock contact. Hornblende-hornfels facies rocks are typical within 150 m of the contact while pyroxene hornfels facies assemblages occur locally within 15 m of the contact.

Contact metamorphic rocks form a band of variable width around the Bear Mountain pluton. Metabasaltic rocks within this contact aureole were separated into two zones based on megascopic textural changes (Fig. 2). Inner zone metabasalts are characterized by macroscopic hornblende and typically a penetrative foliation or compositional layering. Hornfelsic rocks are scarce, while hornblende schist, streaky, veined, and migmatitic amphibolite, and locally basic gneiss are common. The outer zone metabasalts include dark, grayish-green to greenish-black fine-grained semi-hornfelses, as well as various massive, semi-schistose, or cataclastic greenstones exhibiting partial to total recrystallization. Along the eastern and western margins of the pluton, metabasaltic rocks are entirely within the inner zone; while along the northwestern and northern margins lineated hornblende schists of the inner zone grade into massive and weakly schistose metabasalts of the outer zone.

Hornblende-hornfels facies rocks predominate in the inner aureole but pyroxene-hornfels facies assemblages have been found in scattered areas immediately adjacent to the intrusive contact. Hornblende-hornfels facies metabasaltic rocks persist beyond the boundary between the texturally-defined inner and outer zones but relict primary features (for example, intergranular texture, mantled or partially reacted phenocrysts, and amygdules) become increasingly apparent with increased distance from the intrusive contact. Table 2 is a summary of the metamorphic mineral assemblages in the outer and inner aureole of the Bear Mountain pluton as well as recognized contact metamorphic mineral assemblages along the western contact of the Blue Ridge ultramafic intrusion.

able 2

Structural relations between the Bear Mountain igneous complex and the country rock

The Blue Ridge intrusion is roughly concordant with the regional structural grain of the country rock units. However, the termination of the upper metavolcanic unit south of Hurdygurdy Butte (Fig. 2) implies local truncation of country rock contacts during emplacement. The eastward dip of the contact rock foliations and the eastward plunge of local linear structures along the western margin of the intrusion suggests an eastward-dipping contact. Contact relations along the east margin are complicated by the emplacement of the younger Bear Mountain pluton, but available data suggest a steep, east-dipping to subvertical boundary.

Interpretation of the subsurface geometry of the smaller Clear Creek and Cedar Creek bodies is more difficult. Contact relations along the narrow northern extension of the Cedar Creek body are locally complicated by dikes and sills of ultramafic rocks in contact-metamorphosed metabasalt. At several localities, low-dipping screens of amphibolite are interlayered with ultramafic rocks and younger mafic to felsic dikes intrude both lithologies. However, country rock foliations immediately north of the Clear Creek body are either vertical or dip steeply to the southwest. The shape of the Cedar Creek intrusion, in particular its bend up northwest-flowing Cedar Creek, is suggestive of a moderately steep, eastward dip.

The strong deflection of the adjacent country rock around the margins of the Bear Mountain pluton indicates that the principal mechanism of emplacement was by the forceful intrusion of hot, mobile magmas. Additional supporting evidence includes: 1) the concentric foliation pattern throughout the intrusion, 2) transverse, marginal faulting, 3) outward, low-dipping joints in both the contact aureole and the adjacent intrusive rocks, and 4) small-scale isoclinal folding in the inner zone of the contact aureole.

Coupled with the forcible shouldering aside of country rock was the upward extension or stretching of the metamorphic envelope as the pluton lifted and expanded its walls during the final stages of intrusion. Along the northern margin obvious extensional features include crystal lineations; outward, low to moderate-dipping joints; and local boudinage structures. Lensoidal serpentinite masses occur along or near the contact separating the inner and outer zones of the contact aureole and are interpreted to be local tectonic injections between a plastically deforming inner aureole and a more rigid outer zone. The considerable plasticity of the inner aureole is manifested by small-scale isoclinal folds whose axial surfaces parallel the intrusion contact.

The evidence suggests that at least the inner zone of the aureole was coupled to the upward and expanding plutonic intrusion. However, locally sheared contacts between country rock and pluton indicate that this unity was not maintained throughout the entire emplacement history of the pluton.

Several faults, present in the country rock prior to intrusion of the Bear Mountain igneous complex, were important in the emplacement history. For example, the South Siskiyou fault experienced reactivation during the emplacement of the pluton, particularly during late stage northwest bulging. Other pre-emplacement faults may have played an even more significant role in the localization of the intrusive magmas. Although the Bear Mountain pluton is immediately on strike with several faults north-northeast of Doe Flat, they cannot be traced into the pluton. These faults are part of a complex system that developed during the emplacement of the Preston Peak ophiolite (Snoke, 1977). Some of these fractures apparently were of the magnitude to partially influence the locus of later intrusive magmas.

ROCKS OF THE INTRUSIVE COMPLEX

Ultramafic and gabbroic rocks

Lithologic variation and field characteristics. Clinopyroxene-rich rocks varying from wehrlite to hornblende-olivine clinopyroxenite constitute over 90% of the exposed rock in the Blue Ridge, Clear Creek, and Cedar Creek ultramafic-mafic masses. Most of the ultramafic rocks in the center of the Blue Ridge intrusion are wehrlites or subordinate dunite, while olivine clinopyroxenite and hornblende-olivine clinopyroxenite predominate near the margins of the intrusion. The ultramafic rocks of the Clear and Cedar Creek intrusions are chiefly hornblende-olivine clinopyroxenites. Scarce hornblende-magnetite clinopyroxenite occurs in the Clear Creek body (Table 3). Table 3 Gabbroic rocks are limited to the Blue Ridge intrusion and are common as discontinuous masses near the western contact. At several localities there is a gradation from olivine clinopyroxenite to melanocratic gabbro by the incoming of interstitial plagioclase. Other gabbro occurrences show no gradation and appear to be discrete pods within the ultramafic rocks. Hornblende-anorthite pegmatite is widely distributed as dikes and irregular masses in the clinopyroxene-rich ultramafic rocks and, in particular, within the Blue Ridge intrusion.

The clinopyroxene-rich ultramafic rocks commonly form massive, blocky otucrops characterized by a knobby, pyroxene-studded, orangish crust. Magmatic flowage features such as compositional layering or grain shape orientation are scarce. The intense penetrative fabric characteristic of alpine-type ultramafic rocks is totally absent. In a few exposures, graded layering is apparent, but it is not as well defined or continuous as the rhythmic layering characteristic of stratiform intrusions. Dunite-wehrlite breccia is locally developed in the Blue Ridge body; the dunite occurs as irregular masses in peridotite and is often transected by thin pyroxenite veinlets. Mechanical disruption of consolidated dunite and incorporation into a peridotitic crystal mush appears to be the mechanism of formation. Similar breccias have been found by James (1971) in the ultramafic rocks near Emigrant Gap, Sierra Nevada, and Mossman (1973) has described a "mixed dunite-wehrlite unit" from the Greenhills ultramafic complex, Southland, New Zealand, which may be analogous.

Petrography. The dunite is nearly pure olivine but diopside and chromian spinel are common accessory minerals. Serpentine and magnetite are ubiquitous secondary minerals. The dunite is allotriomorphic-granular with grain size ranging from 1 to 3 mm. A grain shape fabric is apparent in some dunites in hand specimen as well as thin section. Large olivines define the grain shape orientation. The grain boundaries of these larger crystals often locally interlock producing an embayed texture. Commonly the large grains display undulose extinction and deformation bands. Smaller olivines are equant and unstrained. Aggregations of the small olivines have a granoblastic-polygonal texture and indicate local recrystallization of the large, strained olivine grains. Other dunites appear essentially structureless and appear to have experienced grain growth during their microstructural development.

The clinopyroxene-rich ultramafic rocks are allotriomorphic- to hypidiomorphic-granular and consist essentially of diopside, olivine, and hornblende in varying proportions. Pale green diopside, varying from 60 to 90 modal per cent, forms subhedral to anhedral grains of various sizes. The long dimensions of some large clinopyroxenes may exceed 6 mm. Typically the average grain size is 2-4 mm. Pyroxene-pyroxene grain boundaries are often irregular and interlocking. In some sections smaller grains lie along the boundaries of the large grains suggesting partial recrystallization. Exsolution intergrowths parallel to (100) are commonly present in diopside grains of the hornblende-olivine clinopyroxenite.

Olivine often occurs in irregular aggregations of anhedral grains. Size and deformation effects are variable, many olivine grains are equant (ave. grain size = .5 - .75 mm) and are strain-free or exhibit only weak undulose extinction. Other olivine grains are considerably larger (commonly the long dimension is 2 to 3 mm) and irregularly shaped. These grains invariably show deformation bands and strong undulose extinction.

Hornblende forms as either irregular patches (occasionally poikilitic) rimming clinopyroxene or as tiny replacement inclusions within pyroxene grains. Commonly the tiny inclusions are in optical continuity with a large hornblende patch outside of the grain. The hornblende is pleochroic; X-pale yellow, Y-light brown, Z-brown.

Trace amounts of orthopyroxene as reaction rims on olivine occur in many of the rocks. In some hornblende-olivine clinopyroxenites, bronzite forms discrete grains occasionally amounting to 3 or more modal per cent.

Thin exsolution lamellae parallel to (100) occur in the orthopyroxene grains. Bent and kinked orthopyroxenes have been seen.

Subhedral flakes of orangish-brown biotite (probably magnesium-rich) occur in trace amounts in several of the hornblende-bearing clinopyroxenites. Although the biotite may be a deuteric mineral, its shape, composition, and association suggest it could be a late magmatic phase produced when water and alkali contents were at their maximum. Interstitial plagioclase occasionally is present.

In addition to these primary phases, a variety of secondary minerals are found in many of the rocks. They have been summarized in the explanation accompanying the modes (Table 3).

The gabbroic rocks are unfoliated to well-foliated and show a tremendous range in color index and modal mineralogy (Table 3, e.g., compare 198 and 328). In the nonfoliated gabbroic rocks, weakly zoned bytownite forms irregular, interstitial patches between subhedral olivine and clinopyroxene. Small plagioclase inclusions in pyroxene suggest that these minerals were crystallizing together at some point. Pyroxene grains also include rounded olivines, but pyroxene never occurs in olivine. Olivine grains are distinctly smaller (average 1-2 mm) than coexisting diopside crystals (average 3-4 mm). The olivine is fresh and shows slight strain effects. Brown hornblende forms irregular patches including plagioclase as well as olivine and pyroxene. Tiny hornblende replacement inclusions are sieved throughout the clinopyroxene grains. The hornblende paragenesis in these gabbroic rocks is identical to the hornblende-olivine clinopyroxenites.

In some foliated gabbros the rock fabric appears to be the product of grain orientation during flowage of a largely liquid magma. In other gabbros, the early igneous flow texture has been overprinted by solid-state deformation effects such as deformed lamellae and recrystallization along grain boundaries. In still other rocks, subsolidus deformation effects are widespread suggesting that locally late igneous flow has persisted in a nearly solidified rock (i.e., high-temperature, subsolidus protoclastic deformation). Plagioclase grains in these protoclastic gabbros are recrystallized to a micromosaic of unstrained xenomorphic grains, while remnants of the original grains are strongly deformed. Pyroxenes in these rocks are bent and twisted, while coexisting olivine exhibits undulose extinction and deformation bands. In summary, the foliated gabbros show a spectrum in texture from igneous flow to protoclasis with an apparent continuum between these processes.

Two-pyroxene-biotite diorite and monzodiorite

Lithologic variation and field characteristics. Two-pyroxene-biotite diorite, monzodiorite, and amphibolitized equivalents occur as discontinuous bodies in the western half of the Bear Mountain pluton (Fig. 2). We refer to these three large bodies from north to south as follows: Buck Lake, Smith River, and Jedediah Mountain bodies.

Pyroxene-mica diorite and monzodiorite and their amphibolitized equivalents are either medium - or coarse-grained rocks characterized by a prominent plagioclase foliation. A gradation between medium - and coarse-grained pyroxene-mica diorite/monzodiorite occurs in some areas, but sharp intrusive(?) contacts between the medium - and coarse-grained diorites have been observed both in the Buck Lake and Smith River bodies. Medium-grained pyroxene-mica diorite outcrops to the exclusion of the coarse-grained variety in the Jedediah Mountain body.

The pyroxene-mica diorite and monzodiorite are amphibolitized along the margins of all three dioritic masses as well as near fractured zones. Where amphibolization has been extreme, the rocks are almost indistinguishable from the hornblende diorites and gabbros, but the abundance of biotite is a valuable clue to the original mineralogy.

Fresh pyroxene-mica diorite and monzodiorite often appear unusually dark (average ci 30-32) because of the abundance of dark gray or amber plagioclase. Potash feldspar, which may amount to 10 per cent or more in some monzodiorites, is practically indistinguishable in unstained hand specimens. Table 4 summarizes modal analyses for some representative pyroxene-mica diorites and monzodiorites as well as two amphibolized pyroxene-mica diorites. able 4

<u>Petrography</u>. The typical two-pyroxene-biotite diorite and monzodiorite has a hypidiomorphic-granular texture with a prominent plagioclase fabric. The mafic minerals commonly occur in clots in the coarse-grained monzodiorites, while they are more evenly dispersed in the medium-grained rocks. Post-fabric biotite and potash feldspar form patchy poikilitic crystals typically engulfing pyroxene and plagioclase, respectively.

Plagioclase forms subhedral tablets parallel to (010). In the coarse-grained pyroxene-biotite monzodiorites, plagioclase grains are commonly 4-5 mm long; but some crystals may be longer than 8 mm. The plagioclase tablets in the medium-grained diorites range from 1-3 mm, with some grains as long as 5 mm.

Potash feldspar forms large poikilitic patches surrounding and preferentially replacing plagioclase. Evidence of the replacement of plagioclase are: 1) isolated islands of plagioclase in potash feldspar, 2) irregular potash feldspar embayments in subhedral plagioclase tablets, 3) myrmekite corrosion of plagioclase in contact with potash feldspar, and 4) the patchy post-fabric, sprawling growth of potash feldspar. The potash feldspar is typically perthitic .

Quartz is a common but minor constituent of the unaltered pyroxene-biotite diorite and monzodiorite, forming irregular interstitial grains often associated with biotite. In rocks in which biotite content is low, quartz may not occur at all, even though potash feldspar content may exceed 8 per cent (e.g., 255, Table 4). This suggests that the unaltered pyroxene monzodiorites have a true monzonitic character, and quartz formed during late magmatic alteration.

Pale green augite forms subhedral crystals with c-axes oriented approximately in the plagioclase foliation. Crystals as long as 4 mm are common in the coarse monzodiorites. Augite in the medium-grained rocks form 1-2 mm prisms. Twinning parallel to (100) was observed in all the rocks. Brown tabular platelets, typically parallel to (001), produce a schiller effect in many augite crystals.

Hypersthene forms stubby euhedral to subhedral grains showing pinkish-red to grayish-green pleochroism. Hypersthene crystals are always smaller than coexisting augites. The c-crystallographic axes of hypersthene crystals, like augite, lie crudely in the fabric plane of the rocks. Small inclusions of plagioclase, opaque oxides, and less commonly augite occur in hypersthene, but the average orthopyroxene grain contains fewer inclusions than augite. Schiller structure defined by oriented brown platelets occurs in many hypersthene grains.

Textural relations imply that both augite and hypersthene crystallized together during most of the magma's history. The occurrence of interstitial augite and augite rims around hypersthene and perhaps the higher percentage of inclusions in augite and its larger size indicate that Ca-rich pyroxene continued to crystallize after precipitation of Ca-poor pyroxene had ceased.

Post-fabric reddish-brown biotite forms large poikilitic, optically continuous crystals that include plagioclase, hypersthene, augite, and apatite. Both textural and modal data indicate that biotite formed from pyroxene by late magmatic alteration, but isolated subhedral biotite in a fine-grained pyroxene-mica gabbro from the Jedediah Mountain body suggests that some biotite crystallized directly from the melt. Perhaps these early biotites acted as seed crystals from which poikilitic patches grew.

Magnetite and ilmenite are the most abundant accessory minerals in the rocks, averaging between 3 and 4 modal per cent. They commonly form amoeboidal patches in biotite, isolated subrounded grains in pyroxene, or irregular partitions between pyroxene grains. Opaque inclusions in plagioclase are always very small and occasionally euhedral. This textural evidence implies that the Fe-Ti oxides are both preferentially developed on or within iron-bearing silicates and that their maximum period of growth was late in the magma's crystallization history. The common association with biotite implies a late post-fabric formation, perhaps when fO_2 was at its highest. The formation of ilmenite is probably a product of oxyexsolution from an originally homogeneous Fe-Ti oxide during late stage crystallization or subsolidus reequilibration.

Apatite is a common accessory in the two-pyroxene-biotite diorite/monzodiorite unit. In some coarse monzodiorites, apatite crystals are as long as 2-3 mm. Zircon is an uncommon accessory, forming irregular, sometimes elongate crystals. Green hornblende is always less than 1/2 per cent in unaltered pyroxene-biotite diorite/monzodiorite even though the biotite content may be greater than 10 per cent. In rocks in which amphibolization is extreme green hornblende ubiquitously rims the pyroxene grains and in some cases may be the most abundant varietal mineral.

Hornblende-rich gabbroic and dioritic rocks

Lithologic variation and field characteristics. A heterogeneous suite of hornblende gabbros and diorites constitutes over 75 per cent of the mapped area of the Bear Mountain pluton. The following are common varieties in the hornblende-rich unit:

medium-grained hornblende-pyroxene gabbro and diorite (ci 30-45, average per cent An--andesine to labradorite)

coarse-grained hornblende-pyroxene gabbro (ci 55-75, labradoirite-bytownite)

medium-grained hornblende gabbro (ci 50-60, bytownite)

feldspathic hornblende gabbro and hornblende-biotite diorite (ci 20-30, andesine to labradorite)

medium-grained to coarse-grained hornblende and hornblende-biotite diorite (ci 40-50, andesine)

hornblende-biotite quartz diorite (ci 30-40, andesine)

Age relations between these rock types are uncertain and appear conflicting. The observed relations are as follows:

1) Coarse hornblende-pyroxene gabbro intrusive into medium-grained hornblende gabbro.

2) Feldspathic hornblende diorite/gabbro intrusive into medium- and coarse-grained hornblende-pyroxene diorite/gabbro (Devil's Punchbowl area).

3) Gradational (?) contact between medium- and coarse-grained hornblende-pyroxene diorite/gabbro (Devil's Punchbowl area).

4) In Clear Creek south of the intersection with Preston Creek, a large inclusion of feldspathic hornblende diorite in coarse-grained hornblende-pyroxene gabbro [in conflict (?) with age relation stated in (2)] .

Late stage amphibolization and hornblende pegmatite development are characteristic of the hornblende-rich rocks throughout the pluton. Amphibolization occurs on a variety of scales from small wispy streaks parallel to flow structure to large, somewhat irregular masses obliterating mesoscopic foliation. The development of hornblende pegmatite dikes and masses is spatially and perhaps genetically associated with amphibolization. The pegmatites exhibit great variation in texture, size, distribution, and mode of emplacement. Hornblende pegmatite occasionally occurs along contacts between different phases within the hornblende-rich unit. Comb layers of hornblende pegmatite as described by Walawender (1976) from the Los Pinos pluton, southern California, are found at these contacts but also locally occur as fracture fillings within distinct, relatively homogeneous hornblende-rich phases.

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Petrography. Table 5 summarizes the modal variation of rocks from the hornblende-rich unit of the Bear Mountain pluton. The rocks are characterized by intermediate to calcic plagioclase, abundant green and rarely brown hornblende, and variable amounts of pyroxene. Magnetite, ilmenite, and apatite are ubiquitous accessories. Modal biotite, quartz, and potash feldspar are always very minor and commonly absent. Despite this simple mineral assemblage, modal and textural variations are extreme.

A foliation often defined by elongate multi-grain hornblende (with pyroxene) aggregates and tabular plagioclase crystals is prominent in many of the rocks. Scarce linear structures are defined by elongate lenses of mafic minerals. Other hornblende gabbros have only a weak grain shape orientation and are essentially xenomorphic-granular.

Plagioclase crystals range from small equant grains lying along grain boundaries to well-formed tabular crystals averaging 2 - 2 1/2 mm and occasionally exceeding 4 mm. Plagioclase zoning is exceptionally variable, ranging from unzoned bytownite in medium-grained hornblende gabbro to normal and oscillatory zoned grains in the scattered hornblende-biotite quartz diorites. Mild protoclastic (?) strain is typically apparent in the plagioclase crystals of the well-foliated rocks. Characteristic features include bent twin lamellae, occasional fractured crystals, and the occurrence of small equant grains along crystal boundaries.

Hornblende in the gabbroic rocks commonly occurs in irregular multigrain clots commonly cored by relict pyroxene grains and spotted with opaque oxides. In some diorites and quartz diorites, hornblende crystals are subhedral and occasionally acicular, while relict pyroxenes are absent. In the same rocks brown biotite occurs in close association with hornblende, and textural evidence indicates the replacement of hornblende by the late magmatic biotite. All hornblende-plagioclase boundaries in the gabbroic rocks are highly irregular indicating that the replacement of original pyroxene was accompanied by a contemporaneous reaction with adjacent plagioclase.

Pyroxene grains occur as rounded, skeletal, or multi-grain relicts surrounded by hornblende. The hornblende may occur either as an optically continuous collar or as irregular aggregates of xenomorphic grains. Augite is the predominant pyroxene in all the examined rocks, but relict hypersthene grains can typically be found in any relict pyroxene-bearing thin section. In some cases their abundance suggests that the original clinopyroxene/orthopyroxene ratios were as low as 3 to 1 or even 2 to 1.

In the gabbroic rocks, opaque minerals are associated with mafic silicate minerals; in the dioritic rocks, they commonly occur as isolated grains, occasionally exceptionally abundant. Magnetite and ilmenite are the predominant opaque minerals, but Fe- and NiFe-sulfides are also present. Apatite is the most common non-opaque accessory and may account for 1 or more per cent of the rock. Spehne, zircon, occasional epidote, and scarce hematite were seen in some thin sections.

Leucocratic rocks

Variation, distribution, and contact relations. Leucocratic rocks occur as dikes and sheets and comprise less than 1 per cent of the studied plutonic terrane. Biotite tonalite and granodiorite are the most common rocks, although minor amounts of biotite-hornblende quartz diorite (ci $\sim$ 25) and biotite granite (ci usually 5) are occasionally found. Pegmatitic veins and selvages are sometimes associated with these rocks. The leucocratic rocks intrude all of the previously described plutonic units and locally form intrusive breccias.

The leucocratic rocks appear to have an uneven distribution throughout the pluton. Leucocratic dikes are scarce in the western portion of the pluton, particularly in the pyroxene-biotite diorite and monzodiorite unit. In the central area, between Devil's Punchbowl and Island Lake, many leucocratic injections are found in the gabbros and diorites of the hornblende-rich gabbro-diorite unit. North of Devil's Punchbowl, felsic intrusions are unknown, while in the Clear Creek area, along the eastern margin of the pluton, several leucocratic bodies have been found including a mappable mass (Wilderness Falls stock) near the intersection of Clear Creek with Preston

Creek (Fig. 2).

Petrography. Biotite tonalite and granodiorite, the most common leucocratic rocks, are medium-grained hypidiomorphic-granular. They are characterized by abundant quartz (typically 25 to 35 per cent), strongly oscillatory zoned plagioclase, and a low mafic mineral content (ci usually not exceeding 15). Potash feldspar occurs as irregular patchy replacement of plagioclase in the tonalites, while in the granodiorites it forms discrete anhedral grains. Biotite is the most common mafic mineral typically forming subhedral flakes, sometimes defining a foliation. Tonalitic rocks that comprise the Wilderness Falls stock (Fig. 2) are strongly foliated, perhaps related to their emplacement near the margin of the pluton. Bluish-green acicular hornblende occurs in a few more mafic tonalities but it is not common. Modal data for several representative rocks from the leucocratic suite are given in Table 6. able 6

Late dikes

A variety of dikes are widespread throughout the metamorphic aureole that surrounds the Bear Mountain igneous complex. Dike swarms are especially common in the country rocks adjacent to the Blue Ridge intrusion. Although the dike rocks are chiefly dark-colored and presumably mafic, scattered alaskitic dikes may also be part of the suite. Hornblende is the common mafic silicate mineral in the dark-colored dikes, but clinopyroxene or its pseudomorph is present in a few dikes. No attempt has been made to study the dikes systematically; however, their proximity to the complex and dominantly mafic composition suggest a magmatic affinity with the Bear Mountain igneous complex.

WHOLE ROCK CHEMICAL VARIATIONS

Table 7
Table 8

Chemical data from rocks of the Bear Mountain igneous complex are summarized in Table 7 and 8. The reported abundances were obtained by using X-ray fluorescence and instrumental neutron activation analyses; the methods have been described by Hebert and Street (1973) and Bowman and others (1973), respectively. The indicated errors for the neutron activation analysis data are based on counting statistics. 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.

Major elements

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Major element variations as well as proposed chemical relations between the lithologic units in the complex are displayed in Fig. 4 to 6. Each of these diagrams indicate three distinct groupings separated by well-defined gaps: 1) ultramafic and gabbroic rocks, 2) pyroxene-biotite diorite/monzodiorite-hornblende-rich diorite and gabbro, and 3) leucocratic rocks. Additional analyses would probably partially reduce the apparent gaps between the groupings (e.g., see Fig. 7 on which 23 additional analyses from similar plutonic complexes from the Klamath Mountains - Sierra Nevada have been plotted).

ig. 7

The high total Fe of the analyzed dunite (332) coupled with high CaO but low Cr_2O_3 as compared to alpine-type ultramafic tectonites (Coleman, 1977, Tables 1 and 2) is similar to dunitic rocks from Alaskan-type complexes (Taylor, 1967; Irvine, 1974). The hornblende-olivine clinopyroxenite (200), characterized by high CaO, moderate total Fe, and relatively low Al_2O_3, is similar to clinopyroxene-rich ultramafic rocks from both ophiolitic terranes (e.g., Bailey and Blake, 1974, Table 2; Hopson and Frano, 1978, Table 3) and Alaskan-type complexes (Taylor, 1967). However, the high total Fe and relatively high Al_2O_3 and TiO_2 contents of the hornblende-magnetite-olivine clinopyroxenite (390) are features distinctive of magnetite-hornblende clinopyroxenites from Alaskan-type complexes (Taylor, 1967), and similar rocks have not been reported from ophiolitic terranes. The gabbroic rocks associated with the ultramafites display considerable whole-rock chemical variation as indicated by the broad ranges in Al_2O_3, total Fe, MgO, and Na_2O displayed by the two analyzed samples (194 and 328).

Despite significant dissimilarity in mineralogy, the two-pyroxene-biotite diorite/monzodiorite (255, 361) and the hornblende gabbro/diorite (223, 278) units have very similar compositions that are essentially calc-alkaline in nature. The high Al_2O_3 and CaO contents and the relatively low TiO_2 abundances (1.0 - 1.3 wt %) are features distinctive of the calc-alkaline association. The major element data coupled with trace element abundances indicate that the most probable volcanic equivalents of these plutonic rocks are high-Al basalt and basaltic andesite (Taylor, 1969). Furthermore, the relatively high K_2O content of the pyroxene-biotite monzodiorites perhaps suggests potassic variants of low-Si andesite (Gulson and others, 1972). Although H_2O+ was not determined for any of the analyzed rocks, the hornblende-rich rocks clearly crystallized from more hydrous magmas than the pyroxene-biotite diorite/monzodiorite unit. The chemical and petrographic data, therefore, suggest that K_2O content and the activity of H_2O during crystallization were the chief factors responsible for the contrasting mineralogy of the units.

The analyzed biotite tonalite, an example of the leucocratic suite, is characterized by high SiO_2, relatively high Al_2O_3, low TiO_2 and MgO, moderately low total Fe and CaO, and $Na_2O/K_2O > 2$. These chemical characteristics are common features of tonalitic-trondhjemitic rocks widespread in the western Sierra Nevada and Klamath Mountains (Compton, 1955; Larsen and Poldervaart, 1961; Hotz, 1971).

Minor elements

The ultramafic and gabbroic rocks exhibit considerable variation in minor element abundances as indicated in Figure 8. From these data the following relations are apparent: Fig. 8

1) Cr is most abundant in the analyzed dunite (332) and hornblende-olivine clinopyroxenite (200) but is distinctly low in abundance in the hornblende magnetite-olivine clinopyroxenite (390);
2) Co and Ni are maximum in the dunite (332), samples 200, 390, and 194 all have similar but distinctly lower abundances, and the leucocratic olivine gabbro (328) is even more impoverished in these elements;
3) the hornblende-magnetite-olivine clinopyroxenite (390) is strikingly enriched in V but is also characterized by high Mn and Sc contents.

Trace element abundances characteristic of the pyroxene-biotite diorite/monzodiorite and the hornblende gabbro/diorite units further substantiate the overall similarity in chemical components of the units (Fig. 8). The low Cr, Co, and Ni contents are characteristic of the calc-alkaline association; the Mn, Sc, and V values are typical of high-Al basalt and andesite (Taylor, 1969).

The abundances of Ba, Cs, Rb, and Sr are higher than typical for calc-alkaline rocks with similar SiO_2 contents.

The biotite tonalite is depleted in Cr, Co, Ni, Sc, and V while Ba, Rb, Th, and U abundances are higher than in the other analyzed rocks.

Rare earth elements

The ultramafic and gabbroic rocks are characterized by low rare earth element (REE) abundances in comparison to the other lithologies which comprise the Bear Mountain igneous complex (Table 8). The REE distribution in respect to Leedey chondrite (Fig. 9) for the ultramafic and gabbroic rocks is distinctive in that La/Yb ratios are low (1.85 - 4.72) and the patterns are relatively uninflected or display only a weak light rare earth elements enrichment relative to the heavy rare earth elements (HREE). In contrast, LREE depletion is a well-documented characteristic of ocean-floor basalt as well as some basaltic rocks from marginal basins and primitive island-arcs. However, rocks of the island-arc tholeiitic series generally tend to display flat REE abundance patterns, parallel but enriched relative to chondrites (Jakes and Gill, 1970). Kay and Senechal (1976) have also shown that LREE depletion relative to HREE is a common characteristic of many igneous rocks from the Troodos ophiolite including a possible late stage silicic differentiate. These various REE characteristics suggest that the ultramafic and gabbroic rocks in the Bear Mountain igneous complex were not derived either by the differentiation of ocean-floor basalt or island-arc tholeiite. The important question of the parental magma for the ultramafic and gabbroic rocks will be further evaluated in a later part of the article.

ig. 9

If the ultramafic and gabbroic rocks crystallized from crystal mushes as suggested by the field and petrographic data, then the REE contents of these rocks depend essentially on two factors: 1) the composition of the parent magma and 2) the relative percentages of cumulative minerals and trapped liquid. Kay and Senechal (1976) suggest that the REE contribution of the various components of a mafic cumulate would be: interstitial magma > clinopyroxene > orthopyroxene, plagioclase > olivine. The total REE contents of the analyzed rocks roughly follow this scheme in that the dunite (332) is lowest in REE, and the leucocratic olivine gabbro (328), a plagioclase-rich cumulate, has a higher total REE content than the dunite but a lower total content than the clinopyroxene-rich rocks. Another noteworthy aspect is the small positive Eu anomaly in the leucocratic olivine gabbro (Fig. 9) due to abundant cumulate plagioclase.

The pyroxene-biotite diorite/monzodiorite and the hornblende gabbro/diorite units display LREE enrichment with flat patterns for HREE (Fig. 9). La/Yb ratios vary from approximately 5 to 8, and HREE concentrations are near or slightly less than 10 times Leedey chondrite. These REE characteristics are strikingly similar to features recently well documented in some high-Al basalts and andesites from the Chilean Andes (Thorpe and others, 1976; Lopez-Escobar and others, 1977).

The biotite tonalite (337) is strongly enriched in LREE (~69 times Leedey) relative to HREE (~2.6 times Leedey) and correspondingly has a high La/Yb ratio (37.43). No appreciable europium anomaly is apparent and the HREE define a relatively flat pattern (Fig. 9). These characteristics are typical of the high - Al_2O_3 tonalitic - trondhjemite suite as defined by Barker and Arth (1976).

MINERAL COMPOSITION VARIATIONS

Analyses of individual phases were performed at the California Institute of Technology with a MAC-5 electron microprobe interfaced to a PDP-8/L computer for control of the stage and crystal spectrometers and for on-line data processing (Chodos and others, 1973). Operating conditions were 15 kV and 0.05 microamps on brass; beam diameters were 1-5 μm for olivine, pyroxene, and spinel group minerals, and 10-30 μm for amphibole and feldspar.

Olivine

Olivine in the ultramafic rocks ranges from $Fo_{82.9}$ in dunite to $Fo_{73.2}$ in hornblende-magnetite-olivine clinopyroxenite. In the gabbroic rocks the olivine compositional range is from $Fo_{74.3}$ to $Fo_{68.6}$. Individual olivine grains are essentially unzoned, and the compositional range within a thin section is small. Olivine in the suite as a whole displays an Fe-enrichment trend (Fig. 10, Table 9), but a compositional gap is apparent approximately between Fo_{75-78}. The explanation of this gap is uncertain, but insufficient sampling is a likely possibility. NiO concentrations in the analyzed olivines are appreciably lower than common in olivine from alpine-type ultramafic rocks (e.g., Loney and others, 1971; Himmelberg and Loney, 1973), but similar to values reported by James (1971) from the Emigrant Gap complex, Sierra Nevada. Likewise, MnO contents (0.30-0.55 wt.%) in olivines from the Bear Mountain complex are similar to concentrations found by James (1971) in olivines from the ultramafic-mafic rocks of the Emigrant Gap complex. The low CaO contents (0.01-0.06 wt. %) in the analyzed olivines are characteristic of a plutonic crystallization environment (Simkin and Smith, 1970). Finally, judging from the available olivine compositional data on Alaskan-type complexes (Irvine, 1974), olivines

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from analogous lithologies in both the Bear Mountain and Emigrant Gap complexes are slightly more iron rich.

Pyroxenes

The compositional variations in the pyroxenes in the ultramafic and associated gabbroic rocks and the two-pyroxene-biotite diorite/monzodiorite unit are summarized in Tables 9, 10, and 11 and graphically displayed in Fig. 11.

able 10, 11, Fig. 11

The clinopyroxenes are calcic and vary from diopside-salite in the ultramafic and gabbroic rocks to calcic augite in the dioritic rocks. The high Ca/ (Mg+Fe) ratios characteristic of this suite are analogous to clinopyroxenes from Alaskan-type complexes (Irvine, 1974); ejected plutonic blocks associated with calc-alkaline volcanic centers such as the Soufriere, St. Vincent (Lewis, 1973a); several ultramafic-mafic plutonic complexes (James, 1971; Mossman, 1973); and alkaline basic rocks (Le Bas, 1962; Gibb, 1973). However, the low Al_2O_3 and TiO_2 contents of the clinopyroxenes from the Bear Mountain complex somewhat further discriminates between these occurrences. Clinopyroxenes from alkaline mafic rocks are characterized by high Al_2O_3 and TiO_2 contents. The clinopyroxenes from St. Vincent are also aluminous (Lewis, 1973b). The alumina content of clinopyroxenes from Alaskan-type complexes is apparently variable. Taylor (1967) reported that Al_2O_3 concentrations in the clinopyroxenes were commonly between 4 to 7 weight per cent, while more extensive recent studies suggest a spectrum from low alumina to high alumina clinopyroxenes in these complexes (Irvine, 1974). Perhaps the most similar clinopyroxenes are from the ultramafic-mafic rocks near Emigrant Gap, Sierra Nevada (James, 1971) and the Greenhills complex, New Zealand (Mossman, 1973), characterized by a calcic composition and low Al_2O_3 and TiO_2 concentrations.

Cr_2O_3 content in the clinopyroxenes from the ultramafic and associated gabbroic rocks of the Bear Mountain igneous complex is quite variable. Clinopyroxenes from the ultramafic rocks contain appreciable amounts of Cr_2O_3, which clearly reflect the partitioning of Cr into early formed Mg-rich pyroxenes (Seward, 1971). Cr_2O_3 concentrations in clinopyroxenes from the gabbroic rocks are low.

Na_2O is present in all the clinopyroxenes from the ultramafic and gabbroic rocks but is consistently low. MnO is very low in clinopyroxenes from dunite-wehrlite-olivine clinopyroxenite specimens but is distinctly higher in clinopyroxenes from the hornblende-magnetite clinopyroxenite and the gabbroic rocks.

The clinopyroxenes from the dioritic rocks share some important characteristics with the clinopyroxenes of the ultramafic and gabbroic rocks. These clinopyroxenes are also calcic with very low Al_2O_3, TiO_2, and Na_2O contents. On the other hand, the clinopyroxenes characteristic of the dioritic rocks are more Fe-rich and likewise have higher MnO concentrations. In that these clinopyroxenes crystallized from a far more differentiated magma than the clinopyroxenes from the ultramafic and gabbroic rocks, Cr_2O_3 is essentially nil. James (1971) reported similar but less Fe-rich clinopyroxenes from dioritic rocks near Emigrant Gap, Sierra Nevada.

Orthopyroxene is absent to sparse in the ultramafic rocks but is common in some gabbroic rocks and always present in the pyroxene-biotite diorite/monzodiorite unit. Bronzite is the Ca-poor pyroxene in the ultramafic and gabbroic rocks (Table 9 and 11, Fig. 11), while in the dioritic rocks orthopyroxenes range in composition from hypersthene to ferrohypersthene (Table 10 and 11, Fig. 11). These orthopyroxenes coexist with Ca-rich pyroxene, and coexisting pairs have been analyzed and are joined by tie lines in Fig. 11.

All the analyzed orthopyroxenes are characterized by relatively low alumina and very low TiO_2 and Na_2O concentrations. Exsolution lamellae are common in the Ca-poor pyroxenes from all the analyzed rocks but are especially well-developed in the dioritic rocks where intergrowth textures characteristic of inverted pigeonite have been recognized. Therefore, the microprobe analyses of the orthopyroxenes do not fully account for the original calcic content of the orthopyroxenes. MnO concentrations in orthopyroxenes from the ultramafic rocks are low, increase in the gabbroic rocks, and reach a maximum ($\sim$.65 to 1.0 wt %) in the dioritic rocks. Cr_2O_3 is present in orthopyroxenes from the ultramafic and melagabbroic rocks; its concentration in the more differentiated rocks is nil.

Oxide minerals

The compositions of the oxide minerals were determined in representative samples from 1) the ultramafic and associated gabbroic rocks, 2) the two-pyroxene-biotite diorite/monzodiorite unit, and 3) the hornblende-rich unit. The primary oxide minerals in these rocks appear to have been spinels ranging
ig. 12 in composition from Cr-Al-rich spinel to magnetite (Fig. 12 and Tables 9 and 10). However, ilmenite and green hercynitic spinel are also present in many magnetite-bearing samples. The ilmenite and hercynite are generally restricted to interstitial areas and are commonly lamellae or irregular patches contained within magnetite hosts; ilmenite and hercynite are rarely found as inclusions in olivine or pyroxene. Therefore, much of the ilmenite and hercynite in these rocks probably formed by late-stage magmatic and/or subsolidus oxidation and exsolution from primary Ti-rich magnetite grains. It follows that the primary compositions of the oxide minerals in these rocks may have been greatly

affected. However, oxide mineral inclusions in olivine and pyroxene appear to have been less altered than the interstitial grains, and systematic differences in primary oxide mineral compositions in these rocks are still discernable.

The primary spinels in the ultramafic and gabbroic rocks are significantly different from those in the two-pyroxene-biotite diorite/monzodiorite unit or hornblende-rich rocks. The spinels in the latter two lithologic units are essentially end-member magnetite and have only very low Ti, Cr, and Al abundances. In contrast, the primary spinels in the ultramafic and gabbroic rocks have high concentrations of Ti, Cr, and Al and define a distinct differentiation trend. Although there is little compositional variation in a single sample, the primary spinels in the ultramafic and gabbroic rocks, taken as a single suite, follow a differentiation trend from an intermediate Cr-Al-rich spinel to titanomagnetite. With increasing $Fe^{+2}/(Mg+Fe^{+2})$ there is a continuous enrichment in Fe^{+3} and a depletion in both Cr and Al. This trend is roughly correlative with decreasing Fo-content of coexisting olivine. If only spinels free of oxyexsolution features are considered, there is also a clear and steady increase in TiO_2 from 1.5 weight percent to >8.0 weight percent with increasing $Fe^{+2}/(Fe^{+2}+Mg)$. However, oxyreduction of interstitial magnetite has depleted the residual magnetite in TiO_2 and tends to obfuscate this Ti-enrichment trend. Cr/(Cr+Al) at first increases with increasing $Fe^{+2}/(Fe^{+2}+Mg)$ in spinels from dunite, wehrlite and some clinopyroxenite samples but covers a wide range in values in the spinels in most clinopyroxenites and gabbros. Some of this spread for the clinopyroxenites and gabbros is certainly due to errors in measuring the very low Cr and Al in these particular minerals. However, in a given rock, primary spinel inclusions in olivine or pyroxene tend to have higher Cr/(Cr+Al) values than do interstitial spinels, suggesting that Cr/(Cr+Al) is indeed decreasing during progressive crystallization, or, alternatively, that Cr/(Cr+Al) is lowered by late-state alteration of the interstitial spinels.

The compositions of the spinels in the Bear Mountain ultramafic rocks are most like those from Alaskan-type ultramafic bodies. Similar high Ti and Fe^{+3} concentrations are reported for spinels in Alaskan-type ultramafic complexes (Findlay, 1969; Irvine, 1973), although the primary spinels in the Bear Mountain rocks have somewhat higher Ti concentrations and a greater overall range in composition. Significantly, such high Ti and Fe^{+3} concentrations are not found in spinels of the other types of occurrences of plutonic ultramafic rocks--layered tholeiitic intrusions such as the Stillwater complex (Jackson, 1963), alpine-type ultramafic bodies such as the Burro Mountain peridotite (Loney and others, 1971), and ophiolitic ultramafic rocks (Coleman, 1977; Dick, 1977). Furthermore, the late-stage alteration of Ti-rich magnetite to hercynite + magnetite + ilmenite that is observed in the Bear Mountain rocks has also been reported in Alaskan-type ultramafic bodies (Irvine, 1973).

Amphiboles

Both primary magmatic amphibole and late deuteric amphibole are present in many rocks from the Bear Mountain igneous complex. However, these amphiboles are easily distinguished both petrographically and compositionally (the primary amphiboles are much more aluminous, alkali-rich, and silica-poor than the secondary amphiboles). Therefore, to relate the amphibole paragenesis to other pyrogenetic mineral phases we have concentrated our microprobe investigations on the primary amphiboles as developed in the ultramafic and gabbroic rocks, the two-pyroxene-biotite diorite/monzodiorite unit, and the hornblende-rich unit (Table 12 and Figure 13).

Table 12
Fig. 13

The primary amphiboles in the ultramafic and gabbroic rocks collectively preserve several compositional trends (Fig. 13). Mg/(Mg+Fe^{++}) decreases from

0.9 to 0.7 and is accompanied by increasing TiO_2 (0.7 to 3.3 wt. %), increasing Na_2O (1.05 to 2.48 wt. %), increasing K_2O (0.03 to 0.98 wt. %), increasing total Al_2O_3 (7.69 to 12.91 wt. %), increasing tetrahedrally coordinated Al, and decreasing SiO_2 (49.36 to 41.90 wt. %). Using the nomenclature of Leake (1978), with decreasing $Mg/(Mg+Fe^{++})$, the amphiboles range from edenite, through edenitic and pargasitic hornblende, to pargasite and ferroan paragasite (Fig. 13).

The amphiboles in an analyzed sample of two-pyroxene-biotite monzodiorite (361) are similar to those in the ultramafic-gabbroic suite except that they are significantly more Fe^{++}-rich (Fig. 13) and more potassic (K_2O = 1.5 wt. %). The amphiboles from a sample of hornblende-biotite meladiorite (278) are also significantly more Fe^{++}-rich than those in the ultramafic-gabbroic suite.

There is little chemical data available for amphiboles from Alaskan-type ultramafic complexes. Analyses published by Irvine (1974) for the Duke Island body have higher Al_2O_3 (14-16 wt. %) but are otherwise similar to analyses of magmatic amphiboles from the ultramafic and gabbroic rocks of the Bear Mountain complex.

Feldspars

Feldspar compositions determined by electron microprobe analyses from the Bear Mountain igneous complex are shown in Fig. 14, and ranges characteristic of various lithologies in the complex are summarized in Tables 9 and 10. Most plagioclase feldspars from the ultramafic-gabbroic suite are bytownite; grains are only weakly zoned, displaying a rather limited range in An content. All the analyzed plagioclases from the ultramafic-gabbroic suite are very depleted in K_2O but contain trace amounts of iron and TiO_2. Such calcic plagioclase feldspars are not characteristic of layered ultrabasic-basic intrusions such as the Skaergaard intrusion or the Stillwater complex but are rather common in

ig. 14

calc-alkaline volcanic suites and associated plutonic ejecta (Lewis, 1969). Furthermore, very calcic plagioclase feldspars have also been reported from mafic rocks within calc-alkaline batholithic terranes (e.g., the San Marcos gabbro, Peninsular Ranges batholith, see Miller, 1937) and from gabbroic pegmatoids in Alaskan-type complexes (Taylor and Noble, 1960; Taylor, 1967; Irvine, 1974).

Plagioclase feldspar from the pyroxene-biotite diorite/monzodiorite unit ranges from calcic to sodic andesine, although within a single thin section the range is considerably less. Petrographic examination indicates that these plagioclases are commonly normally zoned, although a coarse oscillatory zoning was detected in some grains. These data clearly demonstrate that a significant gap in plagioclase compositions exists between the ultramafic-gabbroic rocks and the pyroxene-biotite diorite/monzodiorite unit. Furthermore, the relatively high content of K_2O component in the plagioclase from the dioritic-monzodioritic rocks relative to plagioclase feldspar from the ultramafic-gabbroic rocks further substantiates the distinction between the two suites. Coexisting alkali feldspar, which is invariably perthitic, exhibits only a small range in composition (Fig. 14 and Table 10).

Plagioclase from the hornblende-rich unit was only studied from one specimen with the electron microprobe, but the data indicate an overlap with the compositions determined from the two-pyroxene-biotite diorite/monzodiorite unit (Fig. 14). Plagioclase compositions from a variety of samples from the hornblende-rich gabbro-diorite unit were determined using various universal stage techniques. These data indicate considerable compositional variation of plagioclase within the hornblende-rich unit. For example, some hornblende gabbros contain essentially unzoned calcic plagioclase (bytownite), while plagioclase within hornblende-biotite quartz diorites are characterized by complex normal and oscillatory zoning and range in composition from labradorite to sodic andesine.

EVOLUTION OF THE BEAR MOUNTAIN IGNEOUS COMPLEX

The dunite-wehrlite-gabbro association

The clinopyroxene-rich nature of the ultramafic-mafic intrusions that flank the Bear Mountain pluton is reminiscent of the Alaskan-type complexes as described by Taylor and Noble, 1960; Taylor, 1967; and Irvine, 1974. Other characteristics which support a petrologic kinship between these intrusions and Alaskan-type complexes are:

1) the gradation from wehrlite to hornblende-olivine clinopyroxenite,
2) the local occurrence of hornblende-magnetite clinopyroxenite,
3) the scarcity of orthopyroxene in both the ultramafic and mafic rocks,
4) Fe^{+3} - rich spinels,
5) mesoscopic evidence of crystal accumulation such as scarce size-sorted graded layers,
6) dunite-wehrlite breccias related to post-crystallization mechanical disruption,
7) considerable subsolidus recrystallization of the ultramafic rocks,
8) late hornblende-anorthite pegmatite,
9) hornblende- and pyroxene-hornfels facies contact metamorphism in the adjacent country rock terrane.

Similar characteristics were listed by James (1971) in her evaluation of the ultramafic-mafic rocks near Emigrant Gap, northern Sierra Nevada. However, she was careful to note that differences also existed between the Emigrant Gap ultramafic-mafic rocks and the Alaskan-type complexes. An especially critical feature is that the Emigrant Gap complex includes appreciable two-pyroxene-bearing felsic rocks ranging from diorite to granodiorite. These rocks are gradational into two-pyroxene gabbro and appear to be differentiates evolved from the same parent magma as the clinopyroxene-rich ultramafic rocks (James, 1971). The Bear Mountain igneous complex also includes two-pyroxene-bearing felsic rocks which appear to be genetically related to the clinopyroxene-rich ultramafic-mafic rocks.

The association dunite-wehrlite-gabbro is apparently widespread as judged from the occurrence of these rocks as xenoliths in volcanic rocks of contrasting tectonic regimes. For example, on Reunion Island, Indian Ocean, shield-forming olivine basalt lavas contain cognate xenoliths which include dunite, clinopyroxenite, wehrlite, feldspathic wehrlite, and various gabbroic rocks (Upton and Wadsworth, 1972). A distinct inclusion series in Hawaiian basalts include dunite, wehrlite, gabbro and has been interpreted to be crystal cumulates formed from a basaltic magma at relatively shallow depths (Richter and Murata, 1961; White, 1966; Kuno, 1969). Recently, DeLong and others (1975) have described dunite nodules in alkaline basalt from Kanaga Island, Alaska. These dunite inclusions have mineralogic affinities with the olivine-rich rocks found in Alaskan-type complexes. Therefore, the ultramafic-mafic intrusions that flank the Bear Mountain pluton are representatives of a rock suite that commonly appears to be the crystal cumulates of a fractionating basaltic magma.

However, cumulate layering is not common in the clinopyroxene-rich ultramafic-mafic intrusions that flank the Bear Mountain pluton. Furthermore, petrographic data indicate that the rocks of the flank intrusions have undergone considerable subsolidus flowage and recrystallization. The dunite-wehrlite breccias also imply significant late magmatic mobilization. Therefore, we argue that these data can best be reconciled by the diapiric re-emplacement of a hot, semisolid crystal slush. This interpretation parallels recent conclusions concerning the emplacement of the Alaskan-type complexes (Irvine, 1974).

The mobility of this crystal slush is documented by clinopyroxenite dikes in the adjacent country rock terrane (Fig. 3). Furthermore, the common occurrence of interstitial hornblende in the clinopyroxene-rich rocks, especially along the margins of the intrusions, suggest that a watery silicate intercumulus liquid was present during emplacement. Likewise, the abundance of hornblende-rich dikes in the adjacent country rock terrane and the hornblende-anorthite pegmatites in the intrusions are perhaps additional manifestations of the residual liquids squeezed from the mobile ultramafic crystal slush.

Finally, we note that the gravity studies by Kim and Blank (1973) delineate a strong gravity high over the north end of the Bear Mountain complex. Also, clinopyroxene-rich ultramafic and mafic xenoliths (see, e.g. 149, Table 3) occur at scattered localities in the interior of the pluton, a considerable distance from contacts with the ultramafic-mafic flank intrusions. These data, therefore, suggest that additional ultramafic-mafic cumulates are at depth, and that the flank intrusions are but a sample of the accumulative rocks that developed during the fractional crystallization of a basaltic parent.

Petrogenesis of the two-pyroxene-biotite diorite/monzodiorite unit

Major element (e.g., SiO_2, Al_2O_3, MgO, CaO, alkalis) and trace element (e.g., Cr, Co, Ni, REE) abundances indicate a distinct similarity between the two-pyroxene-biotite diorite/monzodiorite unit and low-Si andesite (Gulson and others, 1972). Perhaps the most important difference between the dioritic rocks and low-Si andesite is the higher concentration of total Fe in the plutonic intermediate rocks. This iron-enrichment is manifested in the composition of the mafic mineral phases (e.g., Fe-rich hypersthene to ferrohypersthene) and suggests limited availability of water during the early crystallization history of these rocks. Late alkali feldspar replacement, characteristic of some lithologies in this unit, implies widespread alkali-rich fluids during the late magmatic state. This late alkali enrichment suggests affinites and a transition to low-Si, high-K andesite (Gulson and others, 1972).

Our data suggest that the two-pyroxene-biotite diorite/monzodiorite unit is not a primary magma. Instead, the diorite/monzodiorite unit appears to be related to the ultramafic and gabbroic rocks by fractional crystallization. The critical observations in support of this conclusion are:

1. Similarities in modal mineral phases between the ultramafic-gabbroic rocks and the dioritic rocks (e.g., trace Ca-poor pyroxene in the ultramafic rocks and the presence of two-pyroxene gabbroic and dioritic rocks; trace biotite in the ultramafic-gabbroic rocks and the local monzonitic nature of the two-pyroxene-biotite dioritic rocks);
2. Similar characteristics of clinopyroxenes from the ultramafic-gabbroic rocks and the dioritic rocks (i.e., high CaO, low Al_2O_3 and TiO_2, and the progressive iron enrichment trend);
3. The association of late pyroxene-phyric mafic to intermediate dikes with the Blue Ridge ultramafic-mafic intrusion;
4. Relative emplacement chronology involving the initial rise of early accumulative rocks and the later emplacement of more differentiated, partially crystallized dioritic magmas;
5. Petrographic and geochemical similarities to plutonic suites that display nearly continuous modal and chemical variations from peridotite to diorite (e.g., the ultramafic to dioritic complex, near Emigrant Gap, Sierra Nevada -- James, 1971).

Therefore, we interpret the two-pyroxene-biotite diorite/monzodiorite unit to be derived by crystal fractionation of a primitive basaltic magma with possible accumulative rocks partially preserved as the ultramafic and gabbroic rocks that flank the Bear Mountain pluton.

Petrogenesis of the hornblende-rich gabbro-diorite suite

The textural and modal heterogeneity of the hornblende-rich rocks coupled with their multiple intrusion history hinder detailed analysis of the petrogenesis of this suite. However, several characteristics seem common to the various lithologies, and these aspects are considered most important in an evaluation of the magmatic evolution:

1. The abundance of hornblende;
2. The high Al_2O_3 content but low Cr, Co, Ni abundances; enrichment of LREE relative to HREE;
3. The many chemical similarities with the two-pyroxene-biotite diorite/monzodiorite unit (both major and trace elements);
4. The strong mesoscopic fabric development;
5. Widespread amphibolization and mafic pegmatoid development;
6. The overall chemical similarities to high-Al basalt-andesite.

These characteristics suggest crystallization from a wet magma, probably varying from high-Al basalt to andesite in composition.

As previously indicated, the hornblende-rich rocks locally appear to be contemporaneous with the two-pyroxene-biotite diorite/monzodiorite unit but in other areas are clearly younger. Therefore, the petrogenetic relations between the two units are complex. Certainly some of the hornblende-rich rocks could be "wet" differentiates from a more primitive basaltic magma. However, a substantial portion of the hornblende-rich suite apparently crystallized from a primary hydrous high-Al basaltic magma. Several field and petrologic relations support this conclusion:

1. The abundance of hornblende-rich rocks compared to the other plutonic units;
2. Local contact metamorphism of the two-pyroxene-biotite diorite/monzodiorite unit by lithologies of the hornblende-rich suite;

3. The apparent high water activity but low K_2O content of the magmas from which the hornblende-rich rocks crystallized;

4. Mineralogical variations aberrant from the ultramafic to dioritic suite such as scarcity of Ca-poor pyroxene, low biotite content, and virtual absence of alkali feldspar.

Recently, several authors have suggested that high-Al basalt could be the parent for some of the ultramafic and mafic rocks associated with calc-alkaline complexes (Lewis, 1973a; Saleeby, 1975; Erikson, 1977; Walawender and Smith, 1980; Perfit and others, 1980). A high-Al basalt parent for the postulated accumulative rocks in the Bear Mountain igneous complex seems unlikely in light of the generally low plagioclase content of the early ultramafic and gabbroic rocks. Relatively high water pressure during crystallization would inhibit the crystallization of plagioclase in lieu of Ca-rich pyroxene (Yoder and Tilley, 1962, Lewis, 1969, Murray, 1972). However, an abnormally high water pressure environment seems to be precluded for the early history of the Bear Mountain igneous complex. The two-pyroxene-biotite diorite/monzodiorite unit, which clearly relates to the accumulative rocks, did not crystallize under relatively high pH_2O as indicated by its modal mineralogical variations, low-Al clinopyroxenes, and moderate Fe-enrichment. Furthermore, the accumulative rocks are not composed of hydrous cumulus phases; hornblende and scarce biotite are present in clinopyroxenitic rocks but absent in the earlier cumulates. The hornblende paragenesis is always late magmatic and replacive, therefore, suggesting the crystallization of a hydrous, intercumulus silicate liquid and not primary cumulus crystallization.

In summary, wet high-Al basaltic magma was a major component of the Bear Mountain igneous complex, but its emplacement seems to be preceded by the crystal fractionation of a more mafic, primitive basaltic magma.

Petrogenesis of the leucocratic rocks

The leucocratic rock suite is volumetrically very minor compared to the abundance of ultramafic, gabbroic, and dioritic rocks that comprise the Bear Mountain igneous complex. Furthermore, field relations consistently indicate that the leucocratic magmas were intruded late in the evolution of this igneous center. When compared to other similar plutonic complexes in the Klamath Mountains (e.g., Russian Peak batholith-Romey, 1962; Deadman Peak pluton - Holdaway, 1962) the paucity of quartz-rich rocks within the Bear Mountain complex is striking. Although mineralogical and chemical variations are clearly evident among the lithologies of the suite, the leucocratic rocks are chiefly tonalitic with biotite or biotite and hornblende as varietal minerals. The limited chemical data indicate characteristics typical of the calc-alkalic tonalite-trondhjemite association: high SiO_2; $Al_2O>15\%$; $Na_2O/K_2O>2$; low concentrations of Cr, Co, Ni, Sc, V; enriched LREE relative to HREE and no europium anomaly.

Trondhjemite-trondhjemitic tonalite is apparently a common component of plutonic complexes that intruded an ensimatic basement (e.g., Compton, 1955; Larsen and Poldervaart, 1961; Davis, 1963; Lipman, 1963; Kelser and others, 1977; Size, 1979). Recent geochemical studies of high- Al_2O_3 tonalite-trondhjemite indicate that hornblende fractionation plays a major role in the origin of such magmas (Arth and Barker, 1976; Arth and others, 1978). These studies indicate that two evolutionary models are consistent with the geochemical data;

1. Fractional crystallization of a wet basaltic magma or
2. Partial melting of amphibolite or eclogite.

Distinguishing between which of these models was operative in a particular plutonic complex is difficult, as demonstrated by the detailed studies of Arth and others (1978). The petrographic and geochemical data from the Bear Mountain igneous complex is compatible with either model. The small volumetric proportion of leucocratic rocks in the complex is supportive of a fractional crystallization model. However, such factors such as depth of erosion and level of plutonic emplacement mollify any argument of consanguinity based on volumetric proportions as presently exposed. The partial melting hypothesis of amphibolite or eclogite is attractive in light of the widespread evidence of Middle to Late Jurassic subduction in the region (Davis and others, 1978; Hamilton, 1978). Arth and Hanson (1972) studied an analogous trondhjemite from the Craggy Peak pluton, east-central Klamath Mountains, California, and concluded that the partial melting of eclogite or amphibolite of subalkaline tholeiitic protolith was the most probable mechanism of magma genesis in accord with their major and trace element geochemical data. Furthermore, they suggested such conditions could be achieved along a subduction zone where basalt or gabbro is transformed to amphibolite or eclogite and subsequently melted at mantle depths. More recently, Arth and Barker (1976) concluded that either crystal fractionation or partial melting mechanisms involving hornblende fractionation must occur at depths less than 60 km. However, experimental studies by Wyllie and others (1976) suggest that the required H_2O content for the generation of tonalite magma under mantle conditions is unrealistically high. Their data suggest that models involving the genesis of tonalite-trondhjemite magmas along subduction zones require careful evaluation. The available restrictions on tonalite-trondhjemite magma genesis suggest anatexis of an amphibolitic source rock under crustal rather than mantle conditions. Such crustal anatexis could be "triggered" by the intrusion of gabbroic magmas derived during subduction of oceanic lithosphere beneath the arc terrane. Finally, none of the above comments preclude the previously suggested alternative of hornblende-controlled crystal fractionation of a hydrous basaltic liquid (presumably high-Al basalt in the case of the Bear Mountain igneous complex). The origin of the leucocratic rocks in the Bear Mountain complex is, therefore, not resolvable within the framework of the available data.

Emplacement History

The previous petrogenetic discussions suggest that multiple magmas were responsible for the evolution of the Bear Mountain igneous complex. The widespread deformation of the country rocks adjacent to the complex indicate the importance of forceful intrusion during the emplacement history. Local structural features such as faults within the country rock terrane apparently also influenced the development of the magmatic locus.

An initial basaltic magma is required as indicated by the early accumulative rocks enriched in magnesian olivine, calcic clinopyroxene, and Cr-Al-rich spinel and locally calcic plagioclase feldspar. This basaltic magma differentiated at depth forming an accumulative crystal slush which was subsequently emplaced higher in the crust as a crystal-liquid aggregate during an episode of lateral compression within the magmatic arc. The hot crystal slush was emplaced as elongated masses apparently dipping inward toward the main subjacent cumulate reservoir. The remobilization of gravity differentiated cumulates has been suggested by Irvine (1967, 1974) and Findlay (1969) for Alaskan-type complexes. Likewise James (1971) suggested that the ultramafic to dioritic rocks near Emigrant Gap, Sierra Nevada, were emplaced as magmatic crystal slushes. The following field observations support a hot, crystal mush emplacement model:

1) protoclastic deformation and recrystallization of ultramafic-mafic rocks locally near the margins of the intrusions;

2) strong deformation of the country rock along the western margin of the Blue Ridge intrusion;

3) synplutonic deformation features such as dunite-wehrlite breccias;

4) the scarcity of cumulate layering or other igneous sedimentation features;

5) ultramafic dikes emplaced into the adjacent country rock.

Intermediate differentiates also related to crystal fractionation of an early mafic basaltic magma subsequently rose as partially crystalline magmas. The rocks that crystallized from these magmas are commonly characterized by a strong grain shape orientation of the early pyrogenetic minerals suggesting an emplacement episode dominated by nearly solid igneous flow. Post-fabric biotite and alkali feldspar attest to widespread upward alkali transfer during the late magmatic stage. The igneous center was then extensively invaded by hydrous high-Al basaltic and basaltic andesite magmas which crystallized as the widespread hornblende-rich gabbro/diorite unit. Multiple emplacement episodes are suggested by the complex field relations within the hornblende-rich unit. Furthermore, local magmatic differentiation via crystal fractionation of these hydrous mafic magmas may account for local lithologic variants. High silica, felsic magmas were subsequently emplaced as numerous dikes and small intrusive masses. If these felsic magmas are differentiates from the hydrous mafic magmas is uncertain, an alternate interpretation is that the leucocratic rocks represent a separate magmatic episode. Still later, swarms of mafic to felsic dikes invaded the plutonic complex as well as the adjacent country rocks. A summary of the plutonic chronologic scheme is presented in Fig. 15.

Coupled with the multistage emplacement history, deduced from exposed field relations, are petrographic and geochemical data which suggest important genetic linkages throughout the rock suite. When all the data are considered concurrently, several inconsistencies require explanation:

1) the emplacement of accumulative rocks (ultramafic and gabbroic rocks) prior to the proposed differentiates (two-pyroxene-biotite diorite/ monzodiorite); and
2) the striking geochemical similarities between the two-pyroxene-biotite diorite/monzodiorite and the hornblende-rich gabbro-diorite unit, despite apparent dissimilarities in emplacement history and modal mineralogy.

Furthermore, the order of emplacement of successive magmas indicates a compositional trend from mafic to silicic (except for the late mafic dikes) and an increase in the apparent H_2O activities of the liquids.

Clearly an explanation of the above relationships is crucial to the proper understanding of the petrogenetic evolution of the complex. However, such an explanation is not immediately obvious from the present distribution of rocks which comprise the complex, and radiometric age data are too few for definitive conclusions. In this light, one viable explanation would be the storage and differentiation of primitive magmas in a lower-level reservoir prior to emplacement higher in the crust. Magmas at any stage of differentiation could be tapped off and mobilized upward. Likewise the reservoir could periodically be replenished by the influx of fresh mantle-derived melt. Accordingly, the proposed accumulative rocks and differentiates need not to be derived from the same initial batch of magma. Furthermore, if the bulk chemical characteristics of the replenishment magmas are similar to the initial batch, apparent petrogenetic linkages can be maintained despite derivation from successive magma batches. On the other hand, over an extended period of several millions of years, fundamental geochemical parameters could change within the lower-level reservoir (e.g., the activity of certain elements or H_2O). Such changes through time may account for the overall similarity but important differences characteristic of the two-pyroxene biotite diorite/monzodiorite unit and the hornblende-rich gabbro diorite unit. The rejuvenation of mafic magmatism via the lower-level reservoir into the upper crustal intrusive locus is also a feasible process to account for the late mafic dikes.

Parental basaltic magma

Numerous aspects of the Bear Mountain igneous complex indicate that this rock series is a calc-alkaline association:

1. SiO_2 enrichment and modal hypersthene and/or quartz in differentiated rocks;
2. limited iron enrichment;
3. abundant intermediate rocks (aluminous basaltic and andesitic compositions);
4. LREE enrichment of intermediate and silicic rocks;
5. low Cr, Co, and Ni as well as TiO_2 abundances in the intermediate rocks;
6. hornblende and biotite as varietal minerals in many lithologies;
7. K_2O enriched rocks such as monzodiorite;
8. very calcic plagioclase in the ultramafic-gabbroic rocks.

The origin of the calc-alkaline suite is an old and complex petrologic problem which has stimulated the development of a variety of petrogenetic models. Many summaries are available (e.g., Boettcher, 1973; Ringwood, 1974) and, therefore, the details of various models will not be reiterated. The fractional crystallization of basaltic magma is a concept of ancient vintage (Bowen, 1928) that has commonly been associated with the evolution of the calc-alkaline suite. The broad application of this mechanism to explain the compositional spectrum of the calc-alkaline suite is invariably complicated by a variety of difficulties which challenge its common viability. However, field, petrographic, and geochemical data from the Bear Mountain igneous complex indicate that an early basaltic magma played a significant role in the magmatic evolution of this calc-alkaline suite. The composition of this parental basaltic magma is uncertain, but a variety of petrologic constraints can be imposed from our available data:

1. Magnesian olivine indicates a high Mg/Fe ratio;
2. Abundant early calcic clinopyroxene implies a calcic melt;
3. The absence or scarcity of orthopyroxene in the ultramafic and gabbroic rocks suggests a low activity of silica;
4. Late magmatic hornblende in the ultramafic and gabbroic rocks suggests an increase in the activity of H_2O during cyrstallization;
5. Trace biotite in the ultramafic and gabbroic rocks as well as the monzonitic nature of the proposed differentiated intermediate rocks suggest an alkalic affinity;
6. Spinel compositions suggest relatively high H_2O activity or low silica activity;
7. REE abundances and distribution patterns for the ultramafic and gabbroic rocks (the proposed cumulates) suggest a parental magma enriched in LREE as compared to ocean-floor basalt or island-arc tholeiite.

These characteristics suggest a mildly alkaline basalt magma enriched in magnesia and lime and apparently further characterized by a low silica activity and moderate pH_2O. These hallmarks are <u>not</u> characteristics usually ascribed to the common basalt types such as ocean-floor basalt, island-arc tholeiite, or high-Al basalt. Although the hypothetical parental basaltic magma must be uncommon, a recent description by Perfit (1978) of "mafic basalt" from the Aleutian Islands, Alaska, suggests such magmas do occur in some magmatic arc terranes. Furthermore, Perfit (1978) argues that such primary basaltic melts are appropriate parental magmas for associated high-Al basalts and andesites. The required differentiation would presumably occur in the crust immediately under the volcanic center. Picritic basalt, conspicuously porphyritic and containing magnesian olivine (Fo_{88}) and diopsidic pyroxene, has been described by Stanton and Bell (1969) from the New Georgia Group, British Solomon Islands Protectorate. This highly mafic lava is part of a volcanic arc suite ranging from picritic and ankaramitic basalts to high-alumina basaltic andesites. Picritic magmas have also been reported from the Lesser Antilles, but these lavas only account for a small percentage of the presently exposed volcanic terrane (Brown and others, 1977). The petrochemical character as well as differentiation history of such primitive basaltic magmas is not well-known. Furthermore, these lavas have been commonly overlooked in calc-alkaline petrogenetic schemes in lieu of more common magma types. The abundance of olivine and olivine-clinopyroxene rich accumulative rocks characteristic of the Bear Mountain complex as well as analogous suites in the Klamath Mountains-Sierra Nevada suggests that such primitive mafic basalts do play an important role in the evolution of some calc-alkaline rock series.

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LIST OF FIGURE

Figure 4 - AFM diagram.

Figure 5 - FeO^* *versus* $\mathrm{FeO}^*/(\mathrm{FeO}^* + \mathrm{MgO})$ plot. Symbol identification is the same as in Fig. 4.

Figure 6 - $\mathrm{SiO_2}$ *versus* $\mathrm{FeO}^*/(\mathrm{FeO}^* + \mathrm{MgO})$

Figure 7 - AFM diagram showing geochemical data from peridotitic to dioritic intrusive plutonic complexes, Klamath Mountains and western Sierra Nevada. Data sources: Holdaway (1962), Romey (1962), Hotz (1971), and James (1971).

Figure 8 - Bar diagrams showing abundance of selected trace elements in analyzed rocks from the Bear Mountain igneous complex.

Figure 9 - Leedey normalized REE patterns for various rocks from the Bear Mountain igneous complex.

Figure 10 - A plot of wt. % Ni *versus* Fo content in analyzed olivines from ultramafic and gabbroic rocks of the Bear Mountain igneous complex. Key to symbols: ▲ = dunite, 332;

◆ = wehrlite, 412; ◍ = hornblende-olivine clinopyroxenite, 200 & 417; ▩ = hornblende-magnetite-olivine clinopyroxenite, 390; ○ = olivine melagabbro/gabbro, 194, 411, 421, 424; □ = leucocratic olivine gabbro, 328.

Figure 11 - Pyroxene compositions from the ultramafic and gabbroic rocks and the two pyroxene-biotite diorite/monzodiorite unit plotted on overlapping pyroxene quadrilaterals and triangular diagrams of .5Ti-Al-.5Cr. On the pyroxene quadrilaterals the pyroxene compositions have been normalized in terms of seven end-member components (see Table 11) and are plotted as altitudes of triangles. The point corresponding to Wo-En-Fs plots in the center of the triangle (i.e., one third up the height of the altitude). The sum of the remaining components equals the altitude of the triangle.

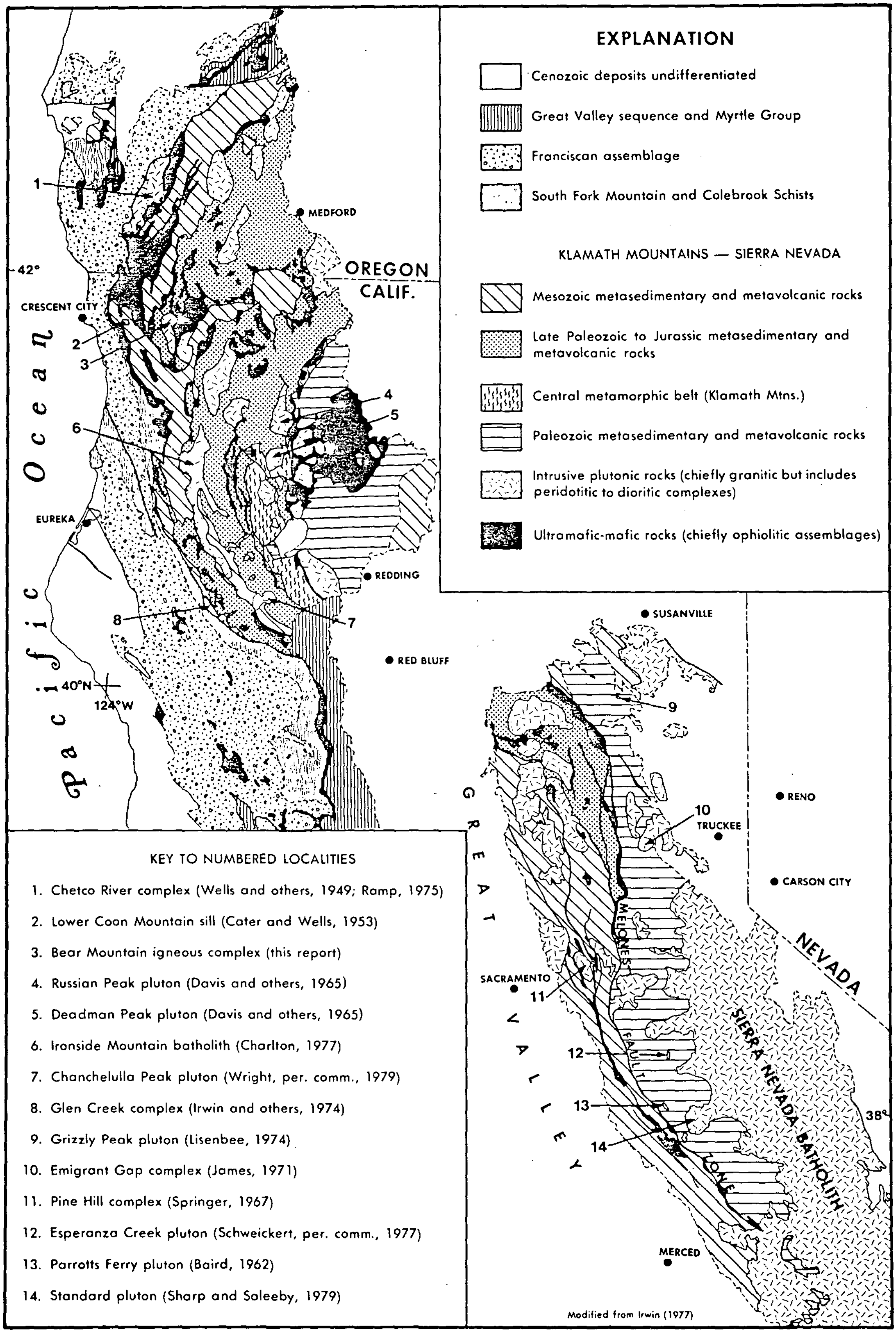
Figure 12 - Plots of $Fe^{+++}/Fe^{+++} + Cr + Al$ versus $Fe^{++}/Fe^{++} + Mg$ (left) and $Cr/Cr + Al$ versus $Fe^{++}/Fe^{++} + Mg$ (right) for spinel-group minerals, Bear Mountain igneous complex.

Figure 13	- A plot of $Mg/Mg + Fe^{++}$ *versus* catatoms Si for amphiboles from the Bear Mountain igneous complex. Amphibole nomenclature is after Leake (1978). Symbol identification is the same as in Fig. 19.
Figure 14	- Overlapping Or-Ab-An diagrams for feldspars, Bear Mountain igneous complex.
Figure 15	- Petrogenetic schemes for the magmatic evolution of the Bear Mountain igneous complex.

LIST OF TABLES

Table 8	- Trace element abundances of rocks from the Bear Mountain igneous complex (in parts per million).
Table 9	- Compositional range of analyzed mineral phases in the ultramafic and associated gabbroic rocks, Bear Mountain igneous complex.
Table 10	- Compositional range of analyzed mineral phases from the two-pyroxene-biotite diorite/monzodiorite unit, Bear Mountain igneous complex.
Table 11	- Representative analyses of pyroxenes.
Table 12	- Representative analyses of amphiboles.

Fig. 1 — Snoke + others



EXPLANATION

☐ Cenozoic deposits undifferentiated

▥ Great Valley sequence and Myrtle Group

⛆ Franciscan assemblage

⬚ South Fork Mountain and Colebrook Schists

KLAMATH MOUNTAINS — SIERRA NEVADA

▧ Mesozoic metasedimentary and metavolcanic rocks

▒ Late Paleozoic to Jurassic metasedimentary and metavolcanic rocks

▥ Central metamorphic belt (Klamath Mtns.)

▤ Paleozoic metasedimentary and metavolcanic rocks

⬚ Intrusive plutonic rocks (chiefly granitic but includes peridotitic to dioritic complexes)

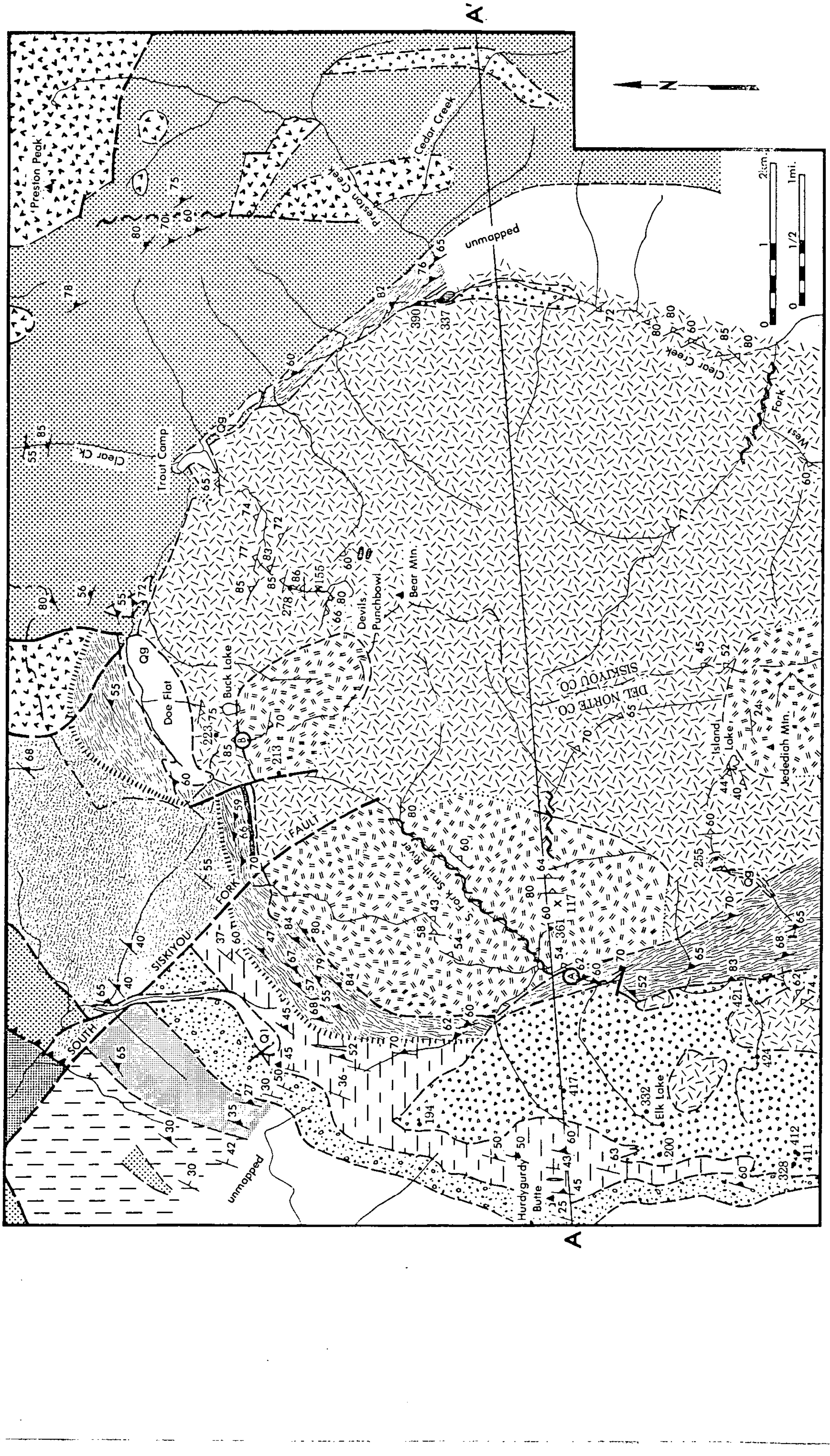
■ Ultramafic-mafic rocks (chiefly ophiolitic assemblages)

KEY TO NUMBERED LOCALITIES

1. Chetco River complex (Wells and others, 1949; Ramp, 1975)
2. Lower Coon Mountain sill (Cater and Wells, 1953)
3. Bear Mountain igneous complex (this report)
4. Russian Peak pluton (Davis and others, 1965)
5. Deadman Peak pluton (Davis and others, 1965)
6. Ironside Mountain batholith (Charlton, 1977)
7. Chanchelulla Peak pluton (Wright, per. comm., 1979)
8. Glen Creek complex (Irwin and others, 1974)
9. Grizzly Peak pluton (Lisenbee, 1974)
10. Emigrant Gap complex (James, 1971)
11. Pine Hill complex (Springer, 1967)
12. Esperanza Creek pluton (Schweickert, per. comm., 1977)
13. Parrotts Ferry pluton (Baird, 1962)
14. Standard pluton (Sharp and Saleeby, 1979)

Modified from Irwin (1977)

Figure


A'
A
N
Preston Peak
Preston Creek
Cedar Creek
unmapped
Clear Ck
Trout Camp
Qg
Doe Flat
Buck Lake
Devils Punchbowl
Bear Mtn.
SISKIYOU CO
DEL NORTE CO
Island Lake
Jedediah Mtn.
Clear Creek
West Fork
South Fork Smith River
SOUTH FORK SISKIYOU FAULT
Ql
Elk Lake
Hurdygurdy Butte
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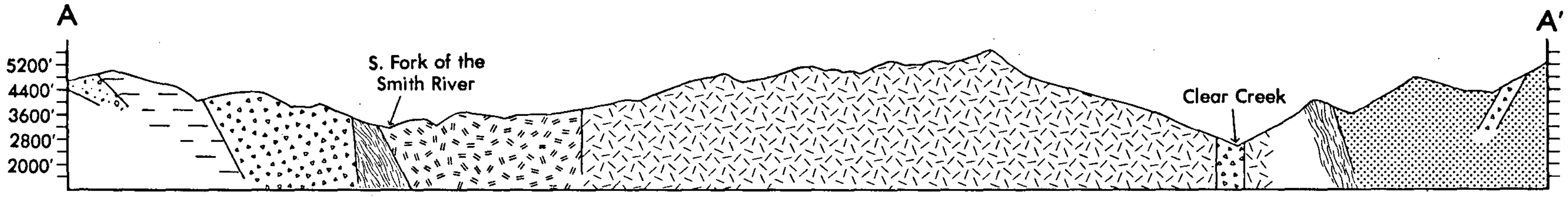


Base from parts of Preston Peak (1956) and Gasquet (1951) 15 min. quadrangles -- U.S. Geological Survey

SURFICIAL DEPOSITS

Quaternary

- Ql Landslide
- Qg Glacial deposits

BEAR MOUNTAIN IGNEOUS COMPLEX

Late Jurassic

- ▨ Leucocratic rocks (quartz diorite to granodiorite)
- ▨ Hornblende-rich rocks (gabbroic to dioritic)
- ▨ Two-pyroxene-biotite diorite and monzodiorite
- ▨ Ultramafic and gabbroic rocks

METASEDIMENTARY ROCKS

Late Jurassic

- ▨ Galice Formation

PRESTON PEAK OPHIOLITE

- ▨ Chiefly hornblende schist, amphibolite, and mafic gneiss of the inner zone of the contact aureole; siliceous metasedimentary rocks differentiated in black

PRESTON PEAK OPHIOLITE (continued)

Early Mesozoic?

- ▨ Upper metavolcanic unit
- ▨ Conglomerate-grit unit
- ▨ Argillite unit
- ▨ Lower metavolcanic unit

Bear Basin Road sequence (Snoke, 1977)

- ▨ Mafic complex (chiefly metadiabase)
- ▨ Alpine-type ultramafic rocks (chiefly serpentinitized harzburgite and dunite)

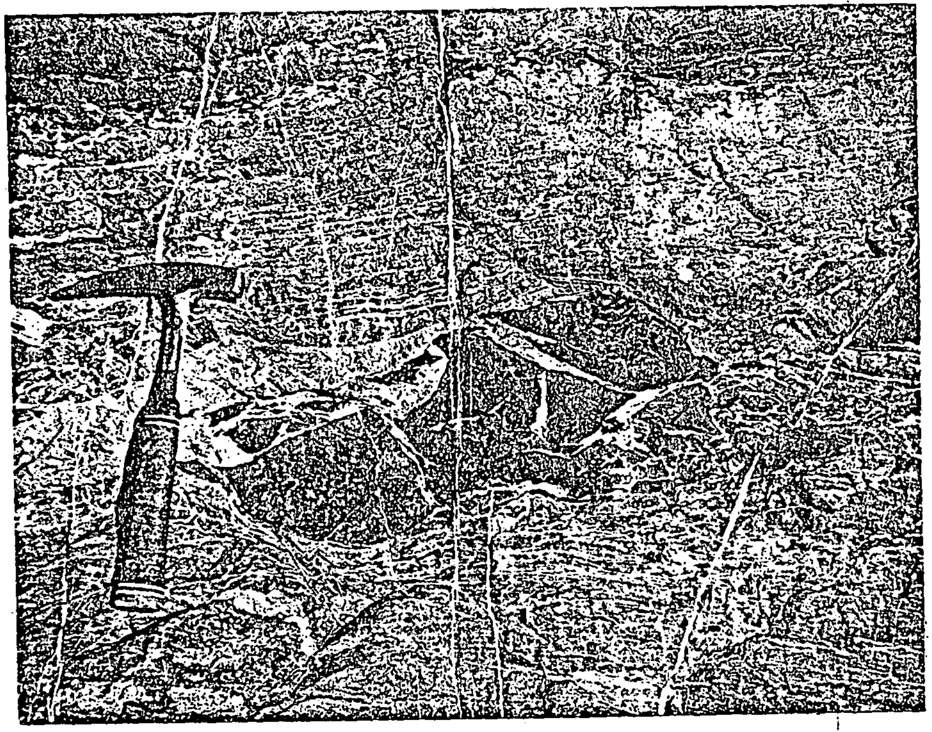
- ▨ Metavolcanic and metasedimentary rocks undifferentiated

—— — — ·····	Contact, dashed where approximately located, dotted where inferred or concealed
IIIIIIIIIIIIIIII	Gradational contact
——— — —	Fault, dashed where approximately located
~ ~ ~ ~	Zone of fractured rocks
▲ ▲ ▲	Thrust fault, dashed where approximately located; sawteeth on upper plate
60 ⊥	Strike and dip of bedding
45 ▲	Strike and dip of foliation in metamorphic and alpine-type ultramafic rocks
55 △	Strike and dip of foliation in intrusive plutonic rocks
416 ●	Sample locality for whole-rock geochemical analyses or microprobe analyses
117 x	Sample locality for K-Ar age determination
——	Drainage

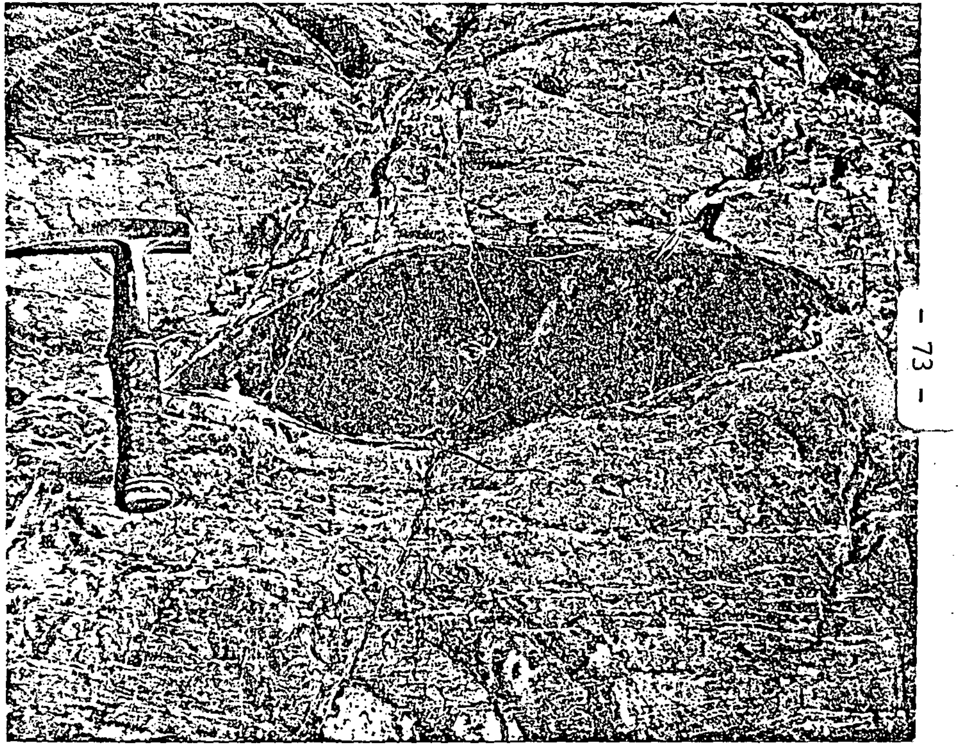
Fig. [illegible] (continued)

Fig. 3

TOP



A



B

Figure . Snoke and others

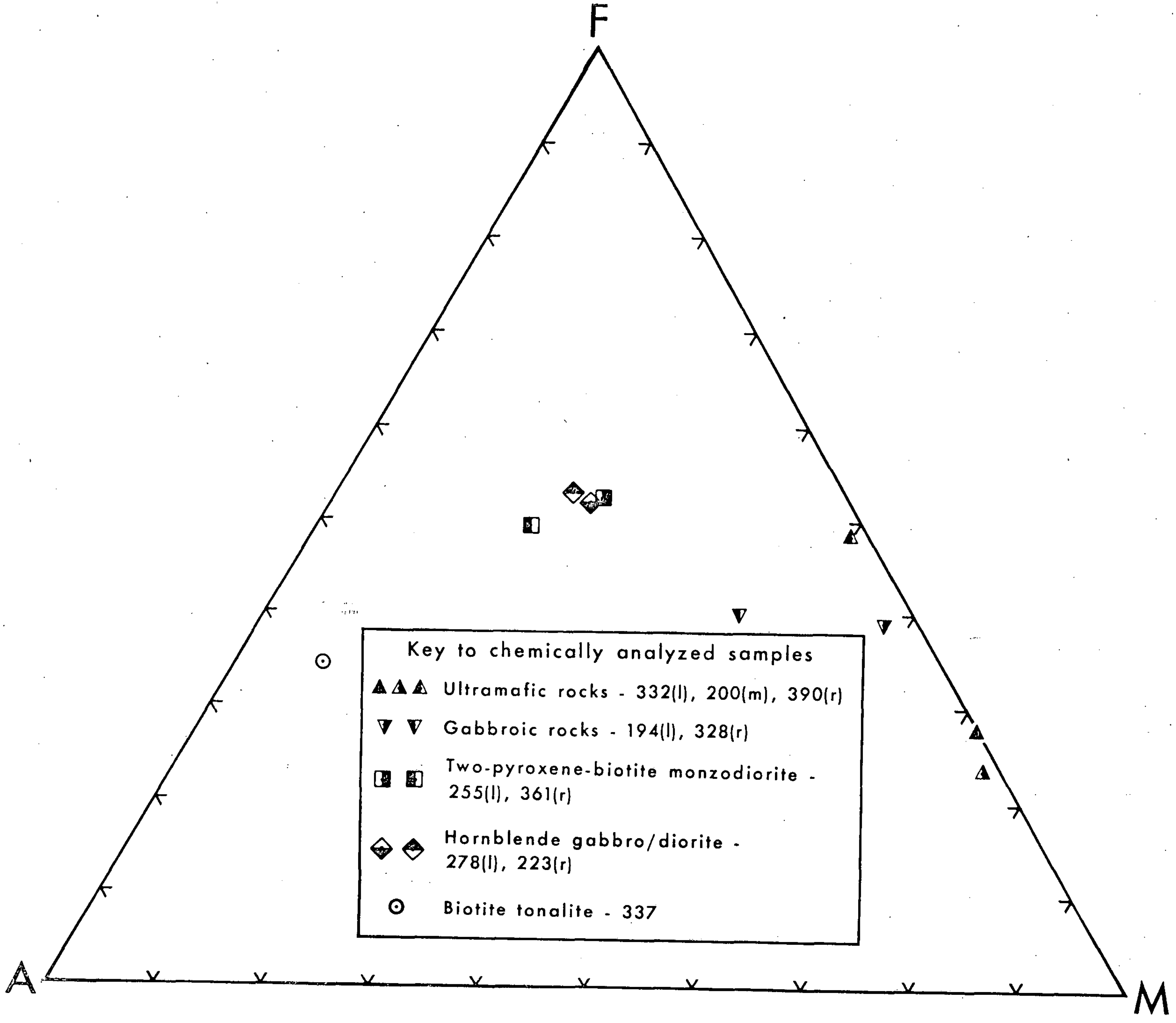


Fig. 4

Fig. 5 — Snoke and others

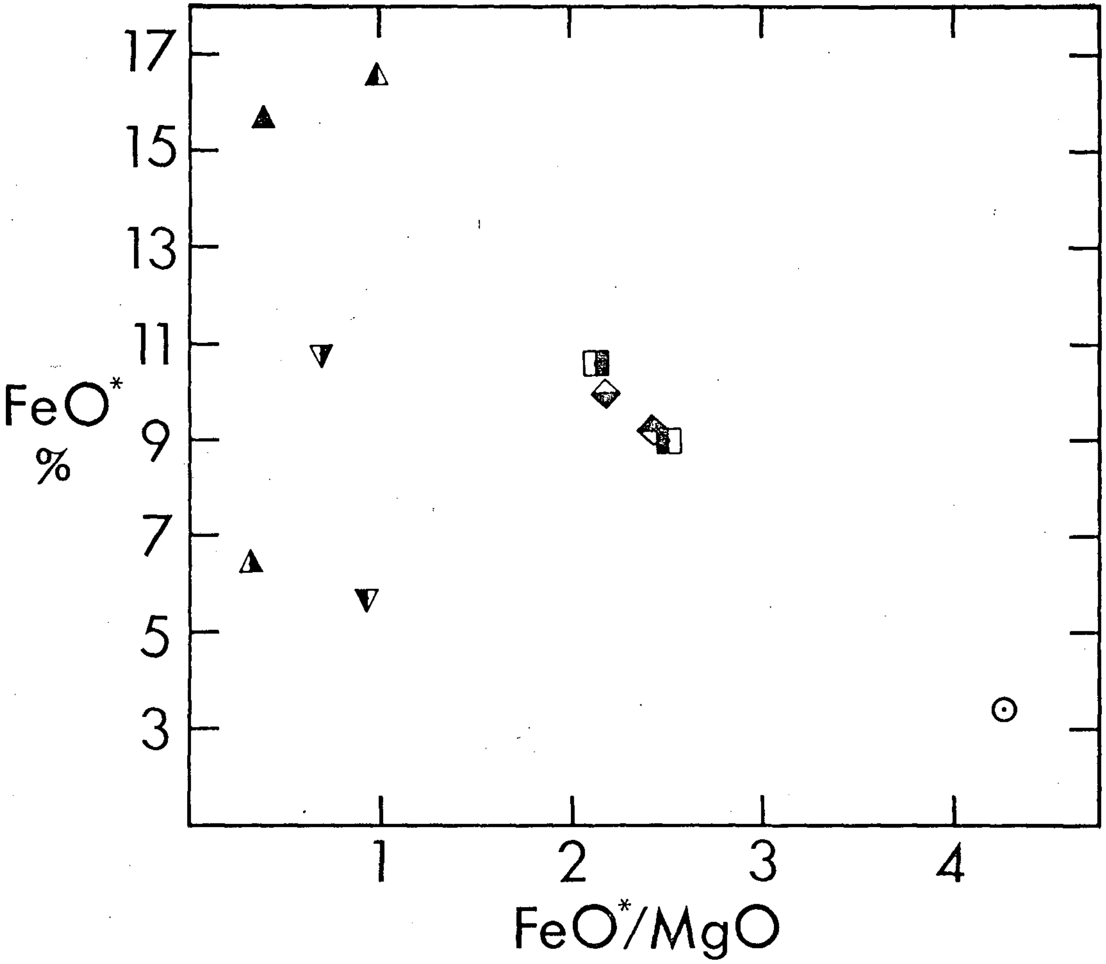


Figure 6 - Smoke and others

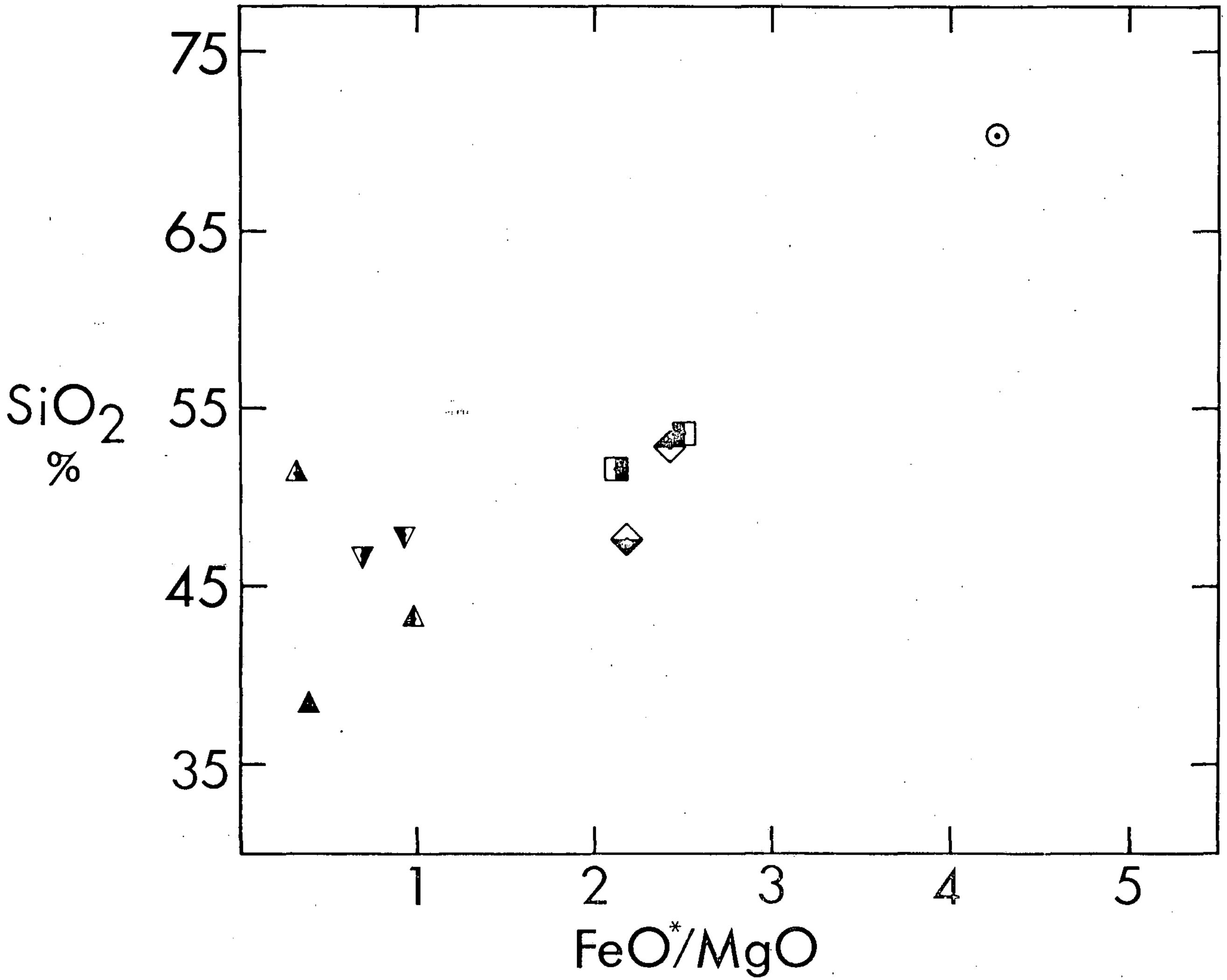


Fig. 7 - Snoke and others

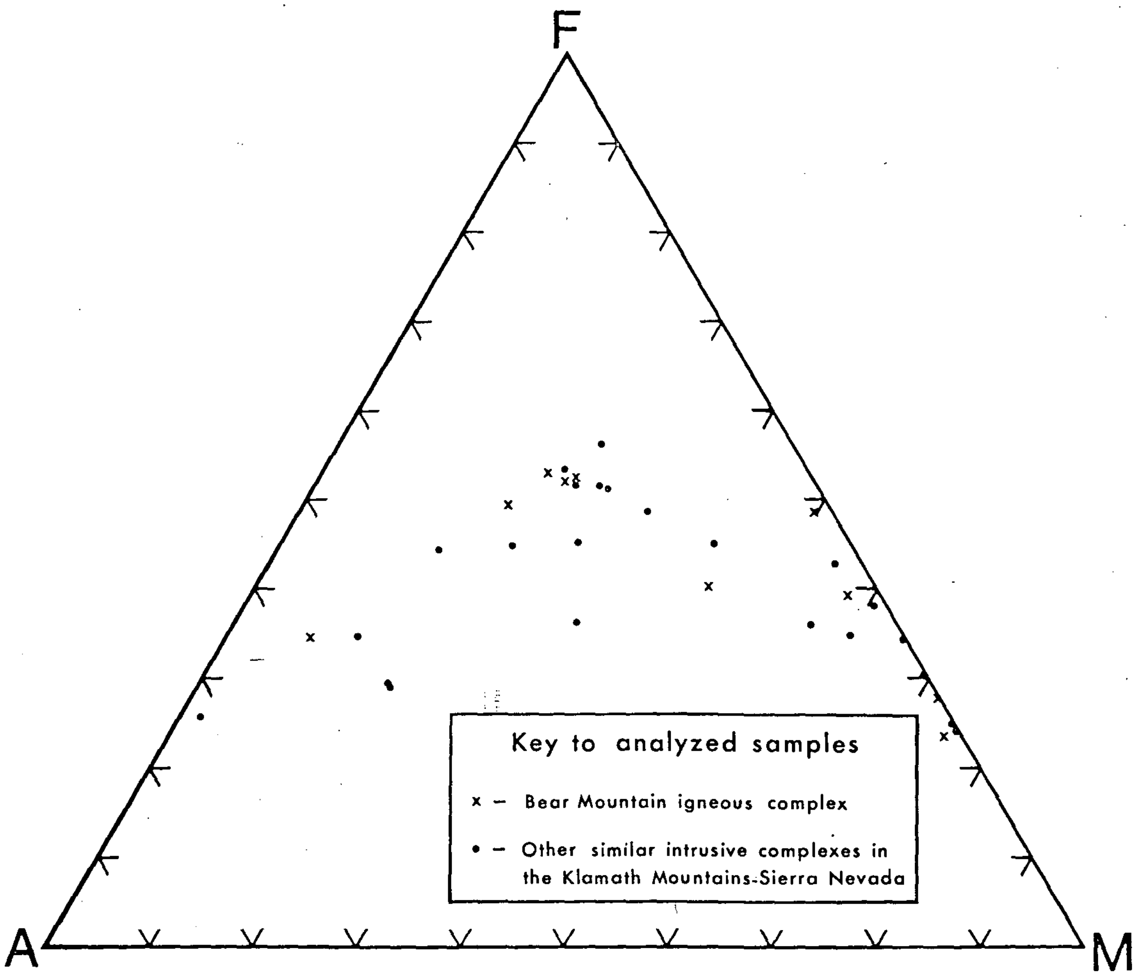
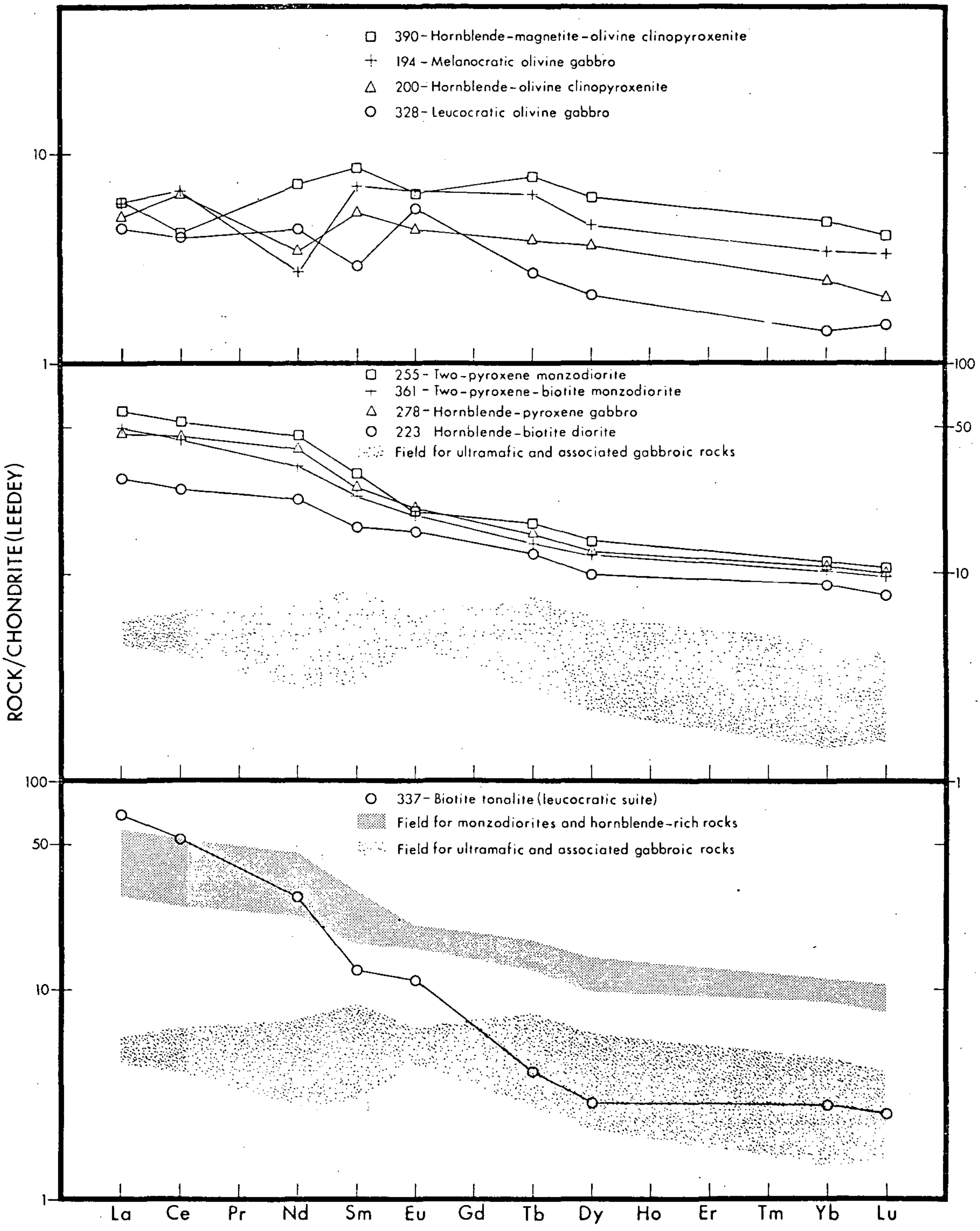
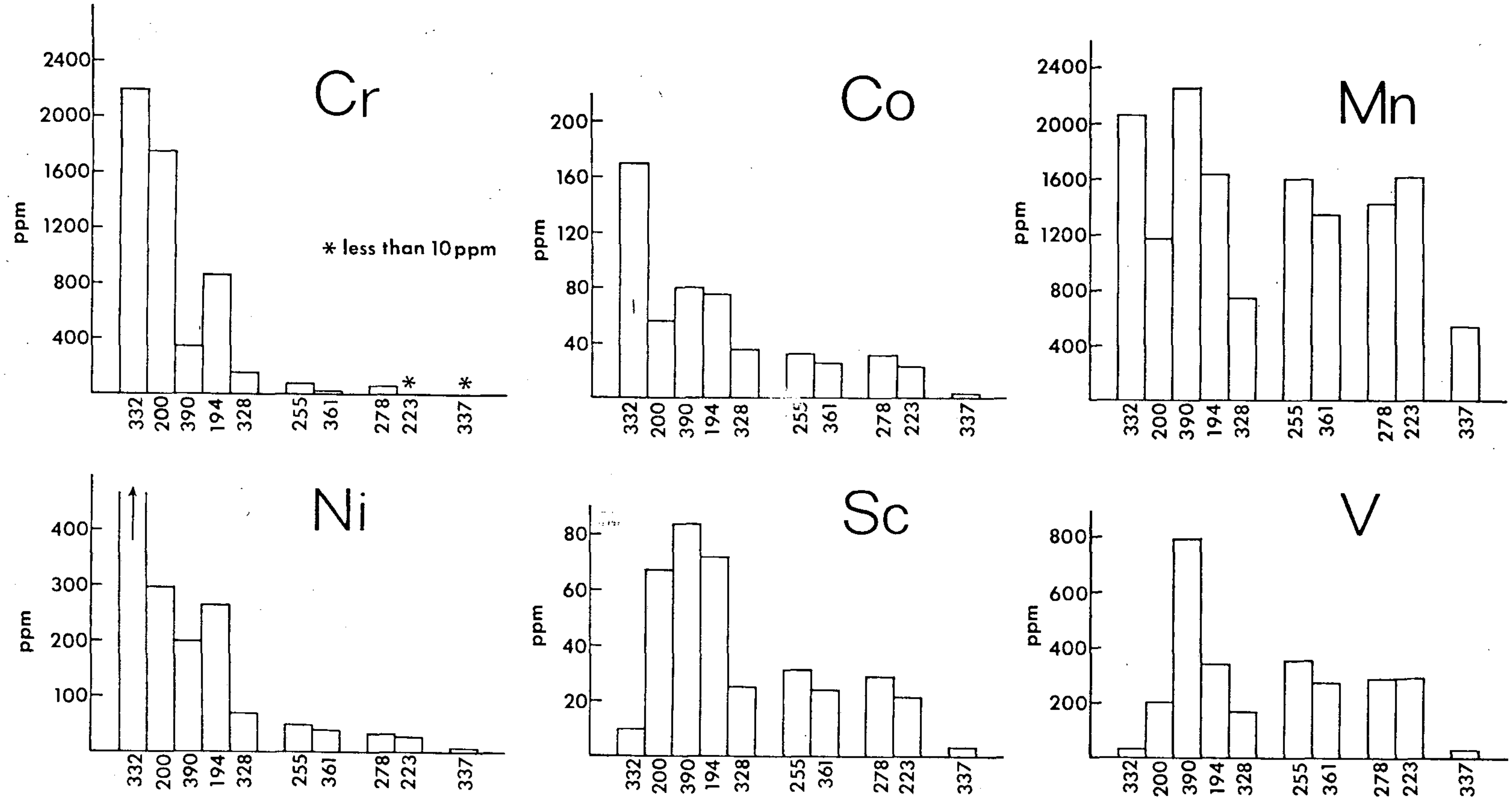


Fig. 7 - Snoke + others





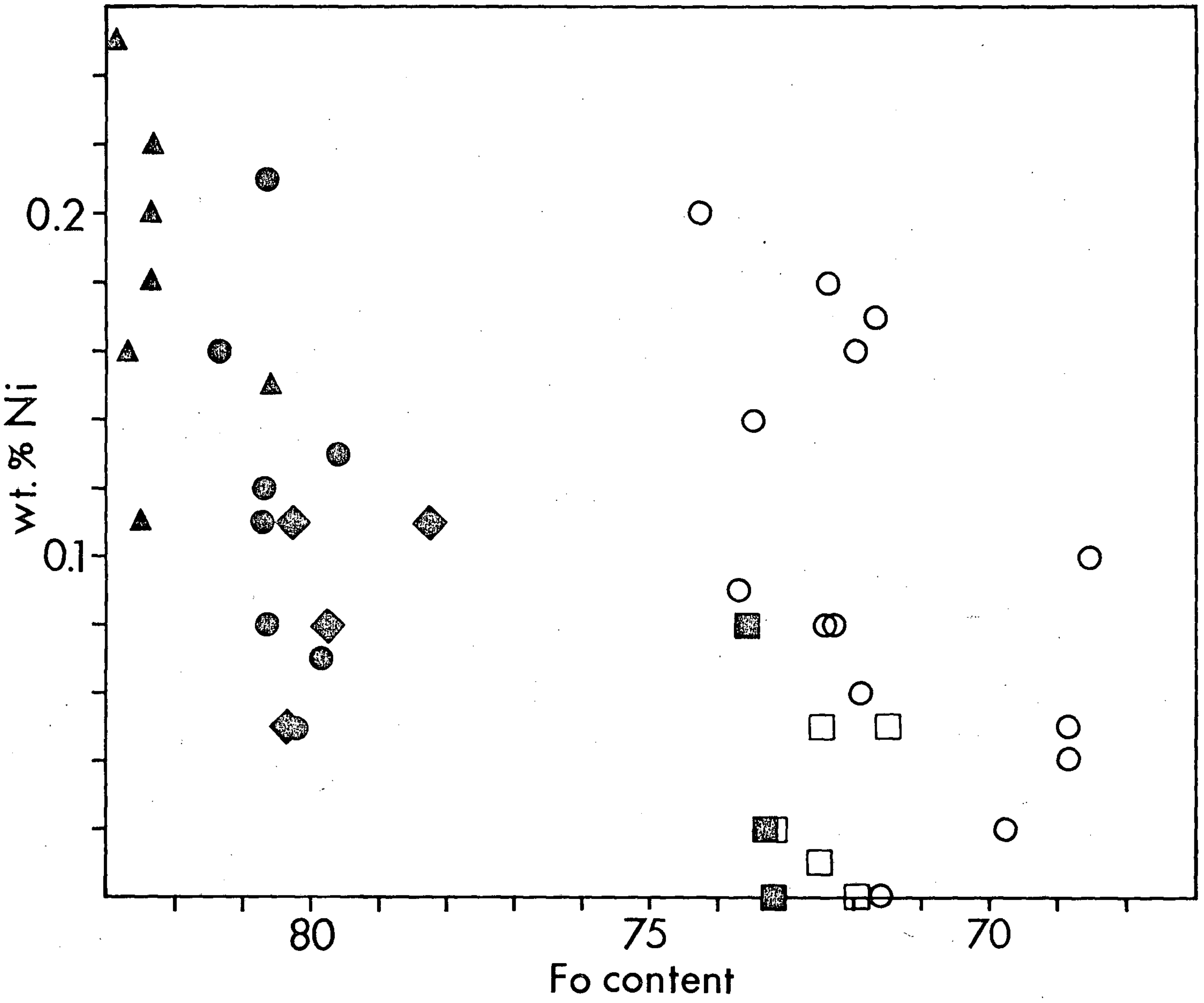
KEY TO CHEMICALLY ANALYZED SAMPLES

332- DUNITE
200- HORNBLENDE-OLIVINE CLINOPYROXENITE
390- HORNBLENDE-MAGNETITE-OLIVINE CLINOPYROXENITE
194- MELANOCRATIC OLIVINE GABBRO
328- LEUCOCRATIC OLIVINE GABBRO
255- TWO PYROXENE MONZODIORITE
361- TWO PYROXENE-BIOTITE MONZODIORITE
278- HORNBLENDE-PYROXENE GABBRO
223- HORNBLENDE-BIOTITE DIORITE
337- BIOTITE TONALITE

Fig. 9 - Snoke + others

Fig. 10

Snake + others, Fig.



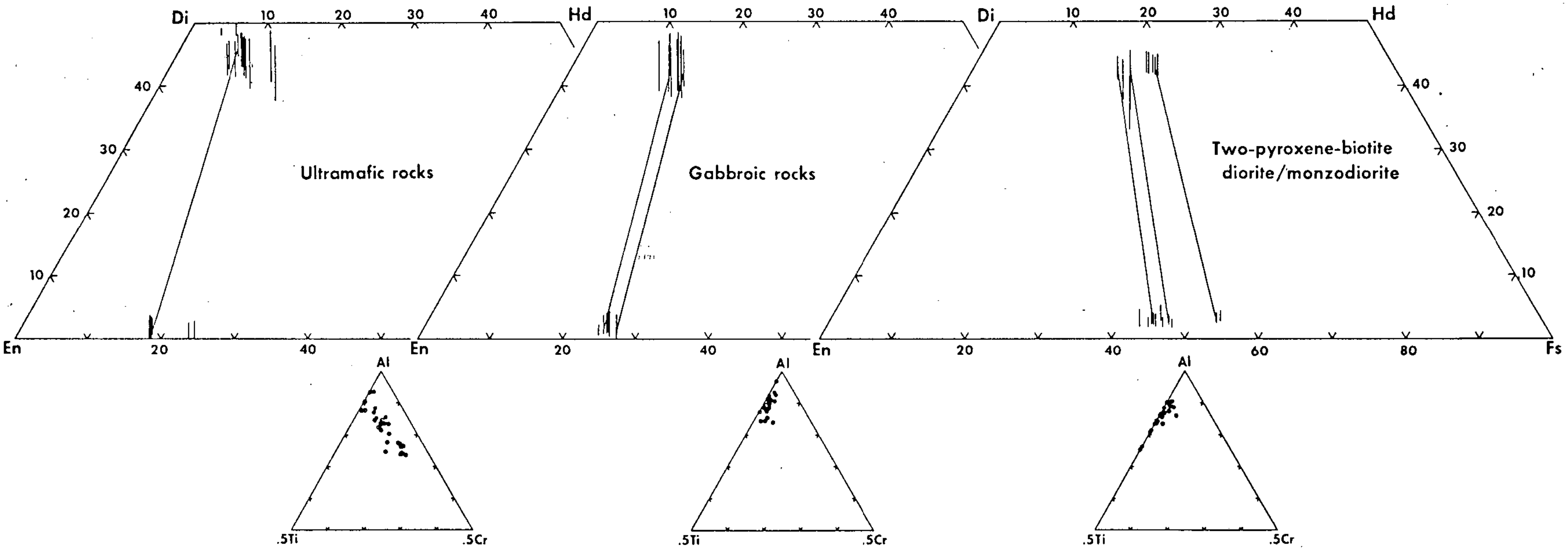


Fig. 11
Snoke + others

SPINEL GROUP MINERALS OF THE BEAR MOUNTAIN IGNEOUS COMPLEX

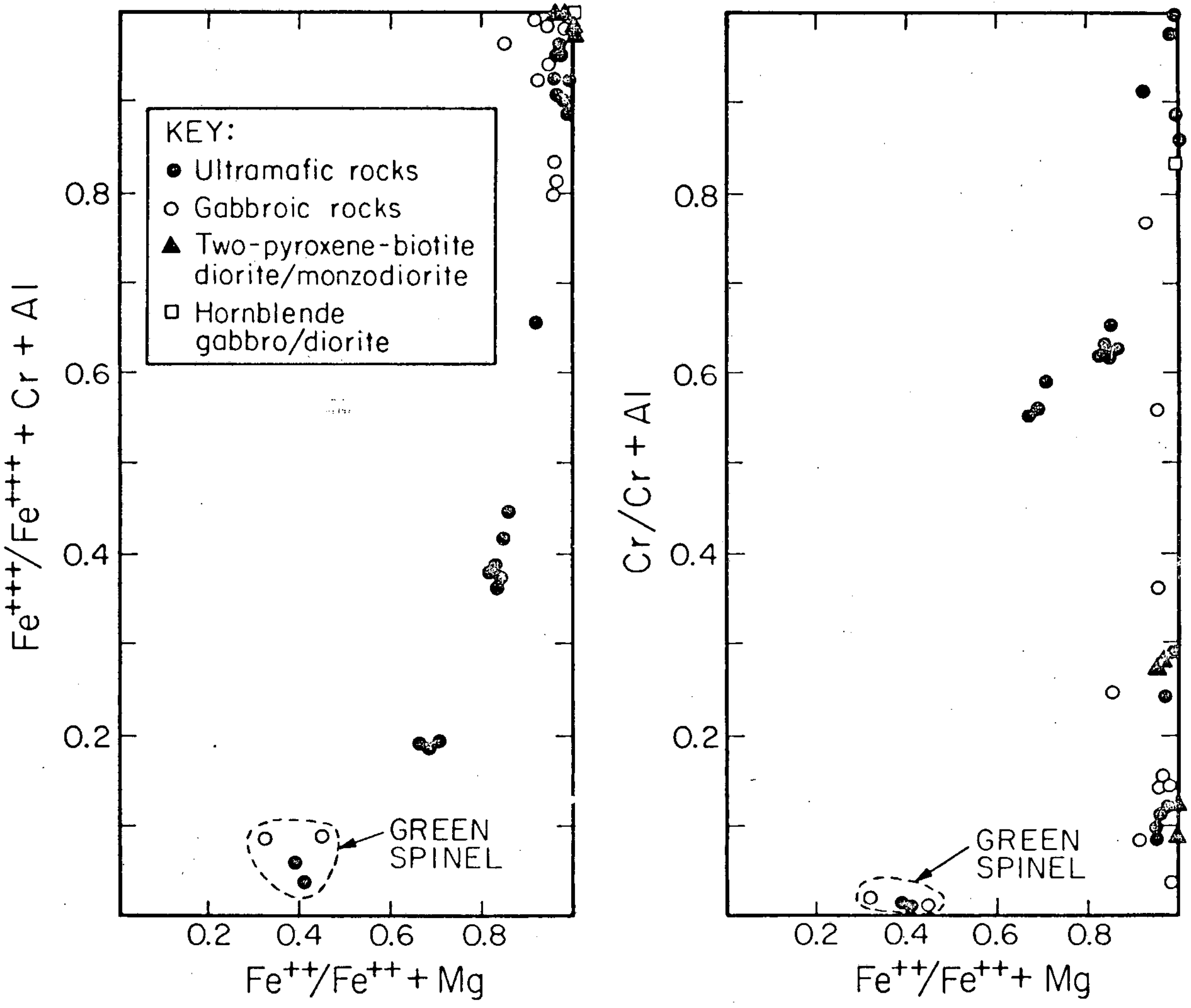


Figure [illegible].- Snoke and others

COMPOSITIONS OF AMPHIBOLES FROM THE BEAR MOUNTAIN COMPLEX

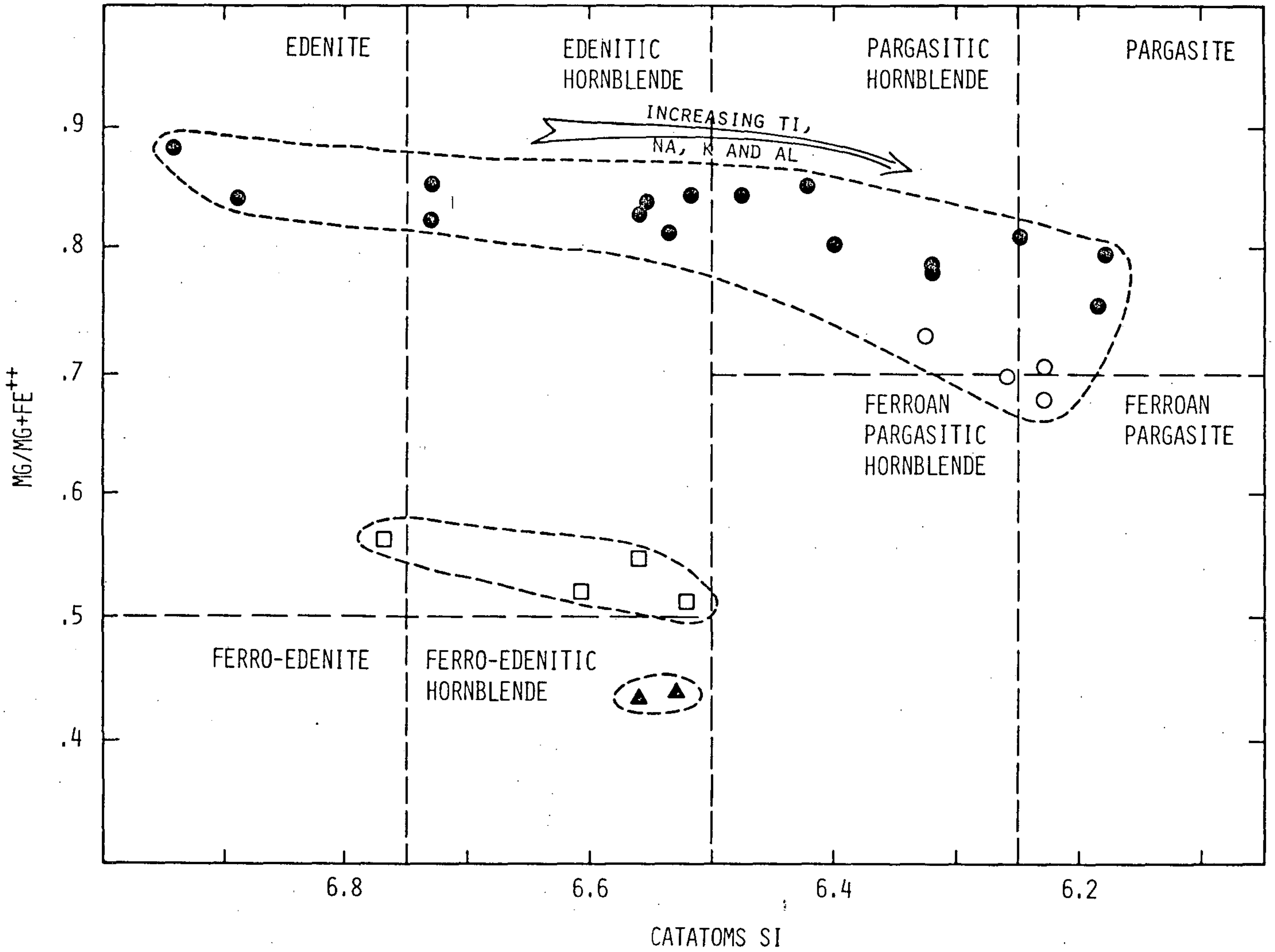
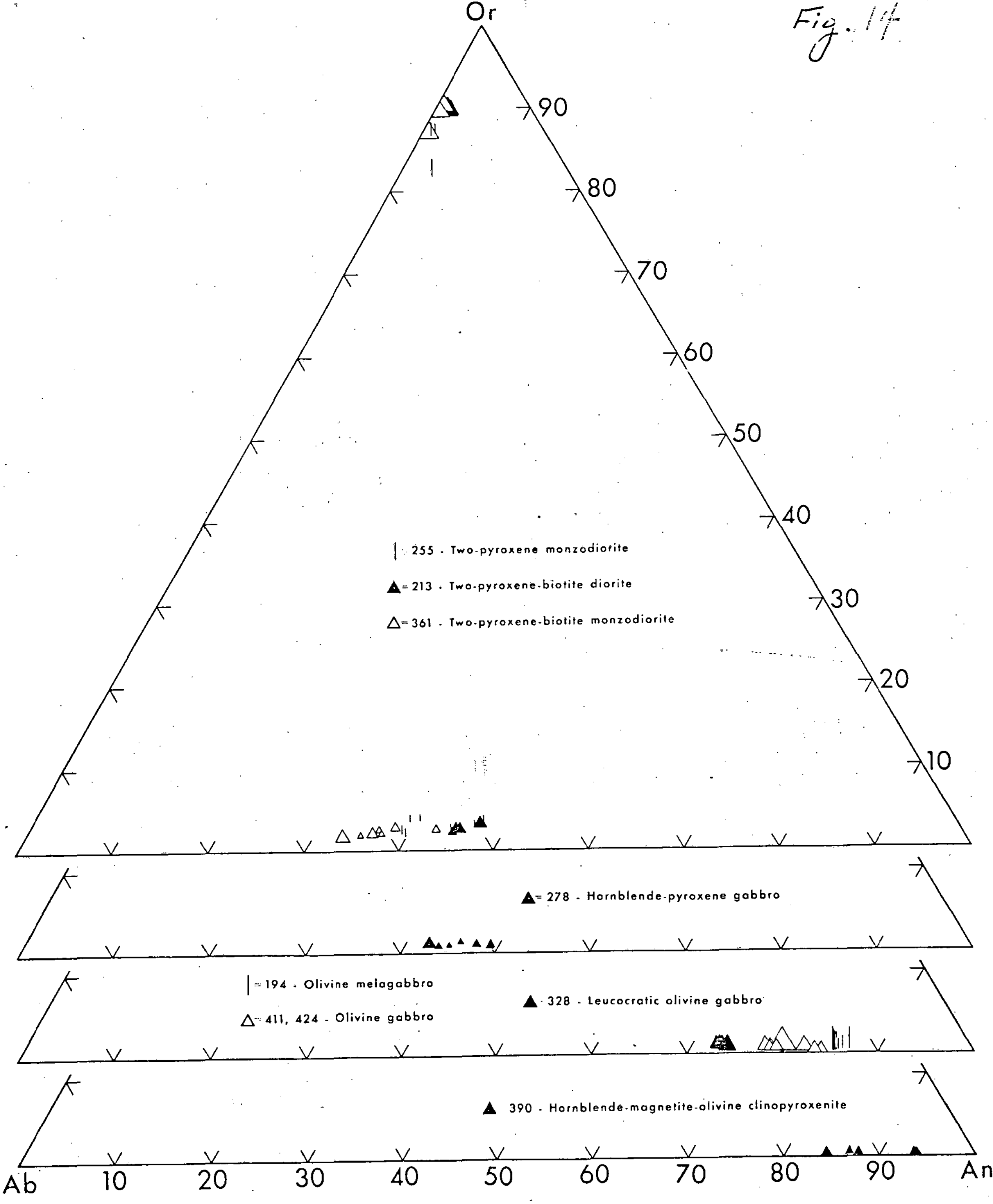
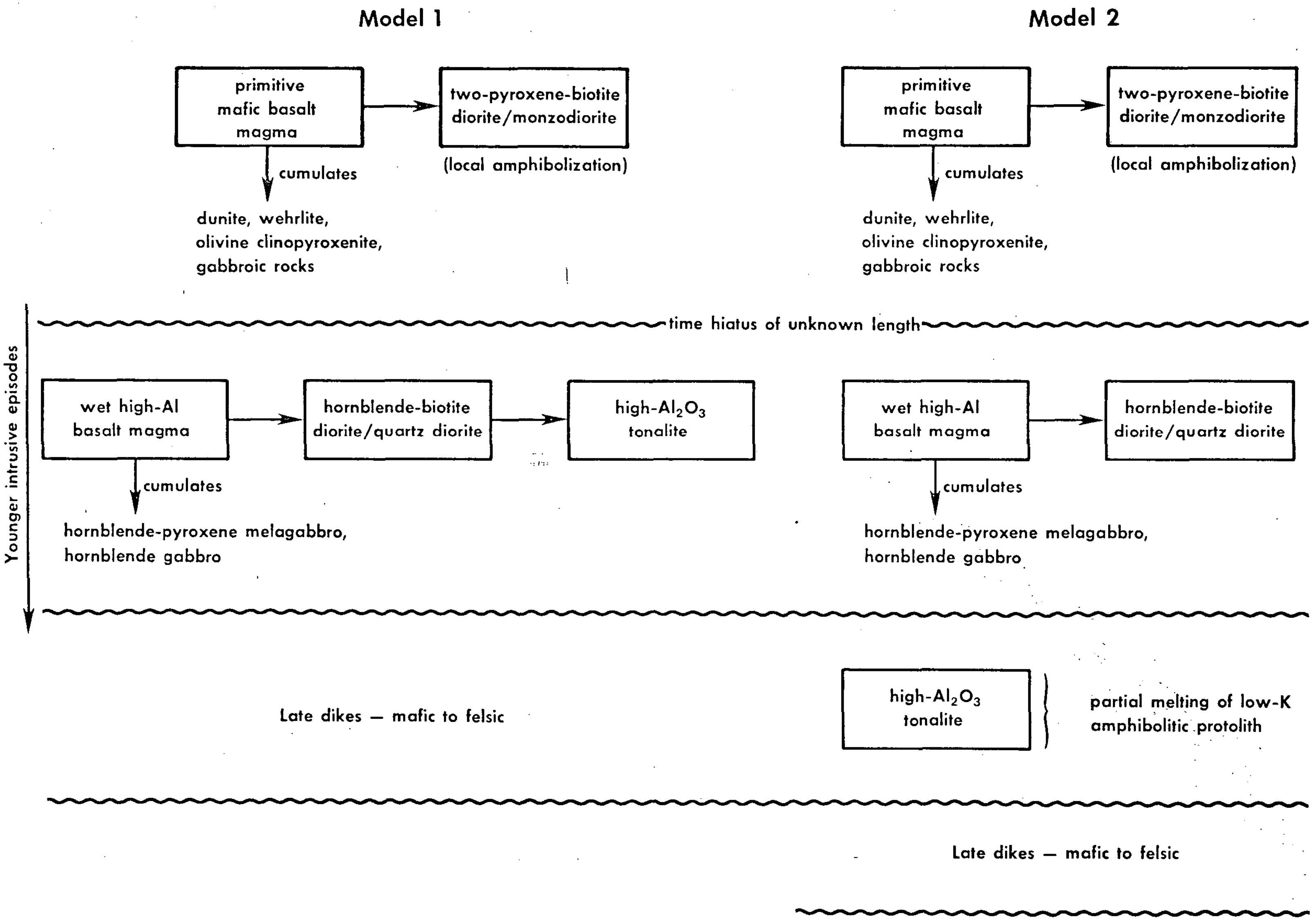


Figure 3 - Snoke and others

Snoke and others,
Fig. 14





Snoke and others, Fig. 15

TABLE I

Potassium - argon ages on rocks from the Bear Mountain igneous complex

Sample no.	Mineral	Percent K_2O	Ar^{40}_{rad} (moles/gm)	$\frac{Ar^{40}_{rad}}{Ar^{40}_{total}}$	Calculated Age (m.y.)
117	Biotite	8.91	1.95×10^{-9}	.93	143 ± 1
Two-pyroxene-biotite monzodiorite-- N41° 46' 43.7" Lat., W 123° 43' .7"Long.					
155	Hornblende	0.860 0.845] 0.852	1.642×10^{-10}	.84	126 ± 4
Hornblende-pyroxene gabbro-- N41° 48' 21.0" Lat., W 123° 40' 12.0" Long.					

Potassium measurements were done on an I.L. flame photometer using a lithium internal standard. Argon measurements were made using standard techniques of isotope dilution.

Ages were calculated using the following constants:

K^{40} decay constants: $\lambda_e = 0.585 \times 10^{-10}$ year $^{-1}$

$\lambda_\beta = 4.72 \times 10^{-10}$ year $^{-1}$

Potassium measurements: Lois B. Schlocker

Argon measurements and age calculation: Jarel L. Von Essen

Table 2

Representative metamorphic mineral assemblages in the contact aureole of the Bear Mountain igneous complex

Entry	Sample No.	Probable protolith	Q	PL	CC	MU	BI	CH	EP	HB	CPX	OPX	GA	CO	WO	SC	OL	TA	TR	SE	AP	SP	OX	SU	MISCELL.
Along the western contact of the Blue Ridge intrusion																									
1	331	basalt	-	X	-	-	-	Or	Or	Xb	X	-	-	-	-	-	-	-	-	-	0	-	X	-	a(r)
2	408	basalt	-	X	Or	-	-	-	Or	Xb	X	-	-	-	-	-	-	-	-	-	-	Or	-	-	a(r), hg(r)
3	416	siltstone	X	X	-	X	X	-	-	-	-	-	X	-	-	-	-	-	-	-	0	-	X	-	s(r)
4	419	polymictic conglomerate	X	X	X	-	X	X	Xz	Xg	X	-	-	-	X	X	-	-	-	-	X	X	X	X	
Outer aureole of the Bear Mountain pluton																									
5	195	basalt	X	X	X	-	-	-	X	Xpg	-	-	-	-	-	-	-	-	-	-	0	X	X	-	
6	314	siltstone	X	X	-	X	X	-	-	-	-	-	-	X	-	-	-	-	-	-	-	-	X	-	
7	366	serpentinized peridotite	-	-	-	-	-	X	-	-	-	-	-	-	-	-	X	X	X	X	-	-	X	0	
Inner aureole of the Bear Mountain pluton																									
8	201	basalt	X	X	-	-	-	-	-	Xbg	-	-	-	-	-	-	-	-	-	-	0	-	X	-	
9	28	basalt	X	X	-	-	-	Or	-	Xg	-	-	-	-	-	-	-	-	-	-	0	-	X	-	
10	126	basalt	X	X	-	-	X	Or	-	Xg	-	-	-	-	-	-	-	-	-	-	0	-	X	-	
11	259	basalt	X	X	-	-	-	Or	-	Xg	X	-	-	-	-	-	-	-	-	-	X	0	X	-	a(r)
12	267	basalt	-	X	-	-	-	-	-	Xb	X	X	-	-	-	-	-	-	Or	-	X	Or	X	-	hg(r)
13	214	siltstone	X	X	-	-	X	-	-	-	-	-	X	X	-	-	-	-	-	-	0	-	X	-	t
14	350	serpentinized peridotite	-	-	-	-	-	Xn	-	-	-	X	-	-	-	-	X	-	-	Xr	-	-	X	0	
15	351	serpentinized peridotite	-	-	-	-	-	X	-	-	-	-	-	-	-	-	X	X	X	Xr	-	-	X	0	

Q = quartz, PL = plagioclase, CC = calcite, MU = muscovite, BI = biotite, CH = chlorite, EP = epidote, HB = hornblende, CPX = Ca-rich pyroxene, OPX = Ca-poor pyroxene, GA = garnet, CO = cordierite, WO = wollastonite, SC = scapolite, OL = olivine, TA = talc, TR = tremolite, SE = serpentine-group minerals, AP = apatite, SP = sphene, OX = Fe-Ti oxides, SU = sulfides, MISCELL: a = actinolite, hg = green hornblende, s = sericite, t = tourmaline.
z = zoisite.
n = nickelian chlorite.
hornblende pleochroism (Z tint) is denoted as follows: pg = pale green, bg = blue-green, g = green, b = brown.
X = mineral present in amounts >1.0%.
0 = mineral present in amounts <1.0%.
- = mineral not present.
r or (r) = mineral species is a retrograde replacement after a prograde mineral.

Snoke and others, Table

Table 3
Representative modes of ultramafic and gabbroic rocks

	Ultramafic rocks										Gabbroic rocks					
	91	95	123	200	332*	342	344	390	412	417	194	328	410	421	424	149*
Olivine	10	4½	31	9	X	3	10	13	28½	15½	13½	10	6	11½	5½	13
Clinopyroxene	76	69½	61½	69	tr	86	82	48	65½	69½	42½	16	46	53	35½	38
Orthopyroxene	3	tr	--	3½	--	tr	tr	1½	tr	½	1½	2	4	tr	2½	tr
Hornblende	9	24½	2½	17	--	10	7	25	2½	12	15½	2	2½	7	11	13
Plagioclase	--	--	--	--	--	--	--	2½	--	--	24½	69	33½	26½	39	35
Opaque minerals	tr	tr	tr	tr	X	tr	tr	8	tr	½	1	1	2	tr	3½	½
Biotite	tr	tr	--	tr	--	--	--	--	--	--	--	tr	--	--	--	--
Miscell. Alter.#	2	1½	5	1½	X	1	1	2	3½	2	1	--	6	2	3	--

* = 332, serpentinized dunite; X = abundant constituent, tr = <1%, -- = not present; alteration is ubiquitous and consists chiefly of serpentine-group minerals, chlorite, and opaque oxides.

** = Olivine gabbro inclusion collected within the hornblende-rich unit approximately .7km north of Devil's Punchbowl.

tr = Trace, present in the thin section but not counted.

\# = Includes colorless clinoamphibole, chlorite, serpentine and very fine-grained opaque minerals (where not clearly associated with olivine), talc, carbonate.

The reported modal analyses were determined on a single standard petrographic thin section. About 800 points evenly spaced over the entire section were counted. Coarse grain size and irregular alteration of the ultramafic and gabbroic rocks limit the determined value to reasonable approximations.

Table 4
Representative modes of rocks from the two-pyroxene-biotite diorite/monzodiorite suite

	Coarse-grained monzodiorite								Medium-grained diorite & amphibolized equivalents*						
	39	117	232	238	239	240	255	362	213	237	252	253	275	45*	245*
Plagioclase	55	57½	54	54	58	56	57½	59	60½	58½	61	69½	67	56	51
Quartz	2½	1½	3½	2	2	2	--	1	½	1½	4½	2½	2	8½	9½
Alkali feldspar	12	9	9	7½	10	9	8½	9½	4½	3½	--	1½	tr	3½	3
Myrmekite	1	1	X	X	1	X	1½	½	½	½	--	--	tr	--	tr
Augite	9½	13	X	X	12½	X	14	14½	16½	15½	13	10½	12	3#	12½#
Hypersthene	7	8	X	X	7	X	11	7½	7½	6	2½	3	3½		
Biotite	10	6½	X	X	5½	X	2	5	6½	9½	13	10	12	11½	15
Hornblende	½	½	X	X	tr	X	--	tr	tr	tr	2½	tr	tr	15½	7
Opaque minerals	2	3	X	X	3½	X	4	3	3	4½	3	3	3½	2½	1
Apatite	1	½	X	X	½	X	1	½	tr	½	½	tr	½	tr	1
Color index	30	31	34	37	29	33	32	30	34	36	34	27	31	32	36

* = amphibolization and associated hydrous alteration of the two-pyroxene-biotite diorite/monzodiorite greatly changes the modal mineralogy of the rock. Pyroxene content is lowered by green hornblende replacement, while modal quartz and biotite increase and plagioclase and alkali feldspar decrease. Modal variations are extreme, and the reported values are only a sample of possible results.

= total pyroxene.

X = denotes that constituent is present but volume percent was not determined.

Modes on medium-grained rocks were done on a single standard petrographic thin section. Approximately 800 points evenly spaced over the entire thin section were counted. Reported modes on coarse-grained rocks were determined by a combination of thin section and slab analyses. Percent mafic minerals, plagioclase, alkali feldspar, and quartz were determined on a stained slab. At least 1200 points were counted. Mafic ratios as well as myrmekite content were determined by modal analysis of a thin section.

Table 5
Representative modes of rocks from the hornblende-rich suite

	111	155	223	230	249	254	268	273	278	336	338	339	341	367	370
Plagioclase	X	68	60½	51	56	56	78	70½	46	54	56	42	74	38	28
Quartz	X	--	4	--	--	--	--	--	--	--	--	--	--	--	--
Alkali feldspar	X	--	--	--	--	--	--	--	--	--	--	--	--	--	--
Hornblende	X	27½	X	X	X	35	X	22½	X	31	X	41	X	X	X
Pyroxene	--	½	X	X	X	3	--	½	X	½	X	--	--	--	X
Biotite	X	tr	X	--	--	--	X	1	X	2	--	--	X	--	--
Opaque oxides	X	3	X	X	X	6	X	4	X	2½	X	4	X	X	X
Apatite	X	1	X	X	X	½	X	1½	X	10	X	½	X	X	X
Miscell. access.	X	tr	X	X	X	tr	X	tr	X	tr	X	tr	X	X	X
Alteration*	X	--	X	X	X	--	X	--	X	--	X	12½	X	X	--
Color index	30	32	35½	49	44	44	22	29½	54	46	44	58	26	62	72

* = includes epidote, chlorite, muscovite, etc.
X = denotes that constituent is present but volume percent was not determined.

Modes on medium-grained rocks were done on a single standard petrographic thin section. Approximately 800 points evenly spaced over the entire thin section were counted. Reported modes on coarse-grained rocks were determined on a stained rock slab. Percent mafic minerals, plagioclase, alkali feldspar and quartz were determined, and at least 1200 points were counted.

Table 6
Representative modes of rocks from the leucocratic suite

	269	270	272	276	337
Plagioclase	51½	53½	47	51	45½
Quartz	30	30½	25½	37	32½
Alkali feldspar	5	1½	3	2	1½
Biotite	11½	12½	17	7½	15
Hornblende	--	--	7	--	--
Muscovite	tr	1	tr	2½	3½
Epidote	1	1	tr	tr	2
Opaque minerals	½	tr	tr	tr	tr
Miscell. access.#	tr	tr	tr	tr	tr
Miscell. alter.	½	tr	½	tr	tr

= Chiefly apatite but also includes very scarce zircon. Scarce garnet in specimen 270

The reported modal analyses were determined on a single standard petrographic thin section stained for both plagioclase and alkali feldspar. About 800 points evenly spaced over the entire section were counted.

TABLE 7

Major element abundances of rocks from the Bear Mountain igneous complex (in percents)

	Ultramafic rocks			Gabbroic rocks		Pyroxene-biotite monzodiorite		Hornblende gabbro/diorite		Biotite tonalite
Sample no.	332	200	390	194	328	255	361	278	223	337
SiO_2	38.4	51.3	43.1	46.7	47.9	51.6	53.6	47.6	52.8	70.4
TiO_2	.05	0.3	1.1	.5	.3	1.3	1.0	1.3	1.1	0.2
Al_2O_3	0.3	2.2	4.3	9.7	23.1	16.3	17.6	19.0	18.3	15.9
FeO*	15.6	6.4	16.5	10.8	5.7	10.6	9.0	10.0	9.2	3.4
MnO*	0.262	0.150	0.285	0.207	0.093	0.203	0.170	0.182	0.204	0.068
MgO	40.3	20.5	16.9	15.8	6.2	5.0	3.6	4.6	3.8	0.8
CaO	1.4	16.4	13.4	14.3	13.6	8.8	7.8	9.7	7.9	3.7
Na_2O*	0.06	0.30	0.43	0.67	2.06	3.20	3.62	3.62	3.55	3.82
K_2O*	0.1	0.2	0.2	0.2	0.2	1.4	2.0	1.0	0.8	1.8
Cr_2O_3*	0.321	0.256	0.050	0.126	0.023	0.011	0.003	0.008	0.001	0.001
Total	96.7	97.8	96.3	99.0	99.2	98.4	98.4	97.0	97.7	100.1

FeO = Total iron

* = Instrumental neutron activation analysis (Analyst-H.R. Bowman) otherwise X-ray fluorescene (Analyst-David W. Yam)

Snoke and others, Table 7

TABLE 8

Trace element abundances of rocks from the Bear Mountain igneous complex (in parts per million)

	Ultramafic rocks			Gabbroic rocks		Pyroxene-biotite monzodiorite		Hornblende gabbro/diorite		Biotite tonalite
Sample no.	332	200	390	194	328	255	361*	278*	223	337
Ba	<10	40±10	53±12	50±11	73±10	323±21	496±26	319	246±17	715±36
Co	170±2	56±1	80±1	75±1	35±1	32±.5	25±1.2	31±2	22.9±.4	3.4±.1
Cs	1.2±.2	.4±.3	.5±.4	.8±.3	.4±.2	.9±.2	1.2±.1	1.05±.15	1.1±.2	1.7±.2
Hf	<.2	.37±.10	.75±.12	.83±.12	.31±.06	3.1±.2	4±1.2	4.7±.6	2.0±.2	3.9±.3
Ni	1341±108	297±34	200±34	264±33	70±16	50±18	40±3	34±11	<29	9±7
Rb	43±16	28±13	37±17	40±15	<18	39±12	51±6	28±9	38±10	58±8
Sc	9.9±.2	67±1	84±1	72±1	25.2±.4	31.3±.5	24±1.5	29±3	21.6±.4	3.2±.1
Sr	<200	152±106	333±167	336±138	1106±285	808±215	942±23	1076±15	423±161	785±223
Th	<.6	.39±.26	.74±.30	.87±.28	<.3	2.0±.2	2.2±.4	1.6±.1	1.1±.2	6.8±.2
U	<.03	.09±.02	.10±.02	.11±.02	<.07	.67±.07	.8±.01	.64±.03	.51±.05	1.2±.1
V	37± 12	202±41	792±123	339±64	168±31	353±77	272±21	284±7	289±63	27±18
Zn	146±22	11±9	68±14	15±10	32±7	126±18	109	114±9	115±17	100±14
La	<.6	1.9±.4	2.2±.5	2.2±.5	1.7±.4	22.5±.7	18.8±.3	17.6±.8	10.7±.5	26.2±.6
Ce	<2.0	6.3±.7	4.1±.8	6.5±.7	4.0±.5	52.0±1.5	42.8±.5	44.1±1.7	24.8±.8	51.1±1.4
Nd	<1.6	2.5±1.1	5.2±1.2	2.0±1.2	3.2±.9	33.0±3	23.3±.6	28.2±1.9	16.3±1.8	20.0±2.0
Sm	.05	1.22±.03	1.99±.04	1.62±.03	.68±.01	6.95±.15	5.37±.15	5.97±.45	3.84±.08	2.88±.06
Eu	<.04	.38±.01	.57±.02	.58±.02	.48±.02	1.71±.0[illegible]	1.64±.02	1.78±.05	1.38±.04	.97±.03
Tb	<.1	.23±.07	.46±.08	.38±.07	.16±.04	1.03±.07	.82±.03	.92±.04	.73±.06	.24±.03
Dy	<.7	1.44±.14	2.4±.2	1.80±.16	.84±.12	5.6±.3	4.8±.2	5.0±.4	3.8±.2	1.13±.13
Yb	<.08	.62±.03	1.19±.04	.86±.04	.36±.02	2.82±.07	2.53±.04	2.71±.14	2.17±.05	.70±.02
Lu	<.02	.08±.01	.16±.02	.13±.02	.06±.01	.41±.03	.37	.38±.06	.30±.02	.10±.01
La/Yb	-	3.06	1.85	2.56	4.72	7.98	7.43	6.49	4.93	37.43
ΣREE	<5.19	14.67	18.27	16.07	11.48	126.02	100.43	106.66	64.02	103.32

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

* = Two samples from the same original hand specimen were analyzed as separate samples but subsequently averaged. The deviation from the mean is half the difference between the two determined values.

Snoke + others,
Table 8

TABLE 9

Compositional range of analyzed mineral phases in the ultramafic and associated gabbroic rocks, Bear Mountain igneous complex

	Olivine	Clinopyroxene	Orthopyroxene	Plagioclase	Spinel group
Ultramafic rocks					
Dunite(332)	$\mathrm{Fo}_{82.9-80.6}$	$\mathrm{Wo}_{45.1}\mathrm{En}_{47.3}\mathrm{Fs}_{7.5}$- $\mathrm{Wo}_{46.0}\mathrm{En}_{46.4}\mathrm{Fs}_{7.7}$	none	none	$\mathrm{Chr\text{-}sp}_{61.1-54.2};\mathrm{Ulv}_{4.2-4.0};\mathrm{Mag}_{34.7-41.8}$
Wehrlite(412)	$\mathrm{Fo}_{80.3-78.8}$	$\mathrm{Wo}_{46.8}\mathrm{En}_{46.3}\mathrm{Fs}_{6.9}$- $\mathrm{Wo}_{43.2}\mathrm{En}_{47.4}\mathrm{Fs}_{9.4}$	none	none	$\mathrm{Chr\text{-}sp}_{33.2};\mathrm{Ulv}_{9.8};\mathrm{Mag}_{57.0}$
Hb-olivine clinopyroxenite(200, 417)	$\mathrm{Fo}_{81.4-79.6}$	$\mathrm{Wo}_{43.9}\mathrm{En}_{49.8}\mathrm{Fs}_{6.3}$- $\mathrm{Wo}_{48.8}\mathrm{En}_{45.2}\mathrm{Fs}_{5.9}$	$\mathrm{Wo}_{.2}\mathrm{En}_{82.5}\mathrm{Fs}_{17.2}$	none	$\mathrm{Chr\text{-}sp}_{78.8-7.7};\mathrm{Ulv}_{3.6-1.2};\mathrm{Mag}_{17.6-91.1}$
Hb-magnetite clinopyroxenite(390)	$\mathrm{Fo}_{73.6-73.2}$	$\mathrm{Wo}_{44.2}\mathrm{En}_{43.9}\mathrm{Fs}_{11.9}$- $\mathrm{Wo}_{41.3}\mathrm{En}_{44.9}\mathrm{Fs}_{13.8}$	$\mathrm{Wo}_{.2}\mathrm{En}_{77.0}\mathrm{Fs}_{22.8}$	$\mathrm{Ab}_{5.7}\mathrm{An}_{94.3}\mathrm{Or}_{0.0}$- $\mathrm{Ab}_{15.1}\mathrm{An}_{84.8}\mathrm{Or}_{0.1}$	$\mathrm{Chr\text{-}sp}_{7.4-3.9};\mathrm{Ulv}_{13.1-7.1};\mathrm{Mag}_{79.5-89.0}$
Gabbroic rocks					
Olivine melagabbro (194,421)	$\mathrm{Fo}_{73.5-71.6}$	$\mathrm{Wo}_{42.5}\mathrm{En}_{46.6}\mathrm{Fs}_{11.0}$- $\mathrm{Wo}_{42.5}\mathrm{En}_{43.4}\mathrm{Fs}_{14.2}$	$\mathrm{Wo}_{.5}\mathrm{En}_{75.2}\mathrm{Fs}_{24.3}$- $\mathrm{Wo}_{1.0}\mathrm{En}_{73.8}\mathrm{Fs}_{25.2}$	$\mathrm{Ab}_{12.6}\mathrm{An}_{87.2}\mathrm{Or}_{0.2}$ $\mathrm{Ab}_{13.3}\mathrm{An}_{86.5}\mathrm{Or}_{0.2}$	$\mathrm{Chr\text{-}sp}_{28.7-6.5};\mathrm{Ulv}_{5.9-2.8};\mathrm{Mag}_{65.4-90.8}$
Olivine gabbro(411, 424)	$\mathrm{Fo}_{74.3-68.6}$	$\mathrm{Wo}_{44.7}\mathrm{En}_{43.8}\mathrm{Fs}_{11.5}$- $\mathrm{Wo}_{42.5}\mathrm{En}_{43.1}\mathrm{Fs}_{14.4}$	$\mathrm{Wo}_{.3}\mathrm{En}_{75.7}\mathrm{Fs}_{24.0}$- $\mathrm{Wo}_{.4}\mathrm{En}_{72.6}\mathrm{Fs}_{27.1}$	$\mathrm{Ab}_{15.3}\mathrm{An}_{84.5}\mathrm{Or}_{0.2}$- $\mathrm{Ab}_{20.8}\mathrm{An}_{79.0}\mathrm{Or}_{0.2}$	$\mathrm{Chr\text{-}sp}_{4.8-2.5};\mathrm{Ulv}_{9.5-0.3};\mathrm{Mag}_{85.6-97.1}$
Leucocratic olivine gabbro(328)	$\mathrm{Fo}_{73.3-71.5}$	$\mathrm{Wo}_{39.8}\mathrm{En}_{45.3}\mathrm{Fs}_{15.0}$	$\mathrm{Wo}_{.2}\mathrm{En}_{75.2}\mathrm{Fs}_{24.7}$	$\mathrm{Ab}_{24.8}\mathrm{An}_{74.7}\mathrm{Or}_{0.5}$- $\mathrm{Ab}_{25.7}\mathrm{An}_{73.6}\mathrm{Or}_{0.7}$	$\mathrm{Chr\text{-}sp}_{6.2-5.4};\mathrm{Ulv}_{18.3-14.0};\mathrm{Mag}_{75.5-79.7}$

Snoke + others, Table 9

TABLE 10

Compositional range of analyzed mineral phases from the two-pyroxene-biotite diorite/monzodiorite unit

	Orthopyroxene	Clinopyroxene	Plagioclase	Alkali feldspar	Spinel group
Two-pyroxene-biotite diorite (213)	$Wo_{1.9}En_{55.3}Fs_{42.7}$- $Wo_{1.8}En_{52.2}Fs_{46.0}$	$Wo_{42.6}En_{38.0}Fs_{19.5}$	$Ab_{49.3}An_{48.1}Or_{2.6}$- $Ab_{53.0}An_{45.2}Or_{1.8}$	$Ab_{10.2}An_{2.6}Or_{87.2}$- $Ab_{8.5}An_{.4}Or_{91.1}$	$Chr\text{-}sp_{3.1-1.9}$; $Ulv_{1.7-1.5}$; $Mag_{95.2-96.5}$
Two-pyroxene monzodiorite (255)	$Wo_{2.5}En_{53.4}Fs_{44.2}$- $Wo_{1.9}En_{50.9}Fs_{47.2}$	$Wo_{40.3}En_{38.6}Fs_{21.1}$- $Wo_{42.4}En_{35.8}Fs_{21.8}$	$Ab_{49.0}An_{47.9}Or_{3.1}$- $Ab_{58.4}An_{39.6}Or_{2.0}$	$Ab_{14.0}An_{2.6}Or_{83.4}$- $Ab_{11.1}An_{.8}Or_{88.1}$	$Chr\text{-}sp_{1.8}$; $Ulv_{2.1}$; $Mag_{96.1}$
Two-pyroxene-biotite monzodiorite (361)	$\sim Wo_{2.7}En_{44.2}Fs_{53.0}$	$Wo_{43.3}En_{33.6}Fs_{23.1}$- $Wo_{43.0}En_{32.2}Fs_{24.8}$	$Ab_{54.7}An_{43.2}Or_{2.1}$- $Ab_{64.7}An_{34.0}Or_{1.3}$	$Ab_{11.8}An_{0.0}Or_{88.2}$- $Ab_{8.6}An_{0.0}Or_{91.4}$	$Chr\text{-}sp_{1.0-0.8}$; $Ulv_{1.6-0.5}$; $Mag_{97.5-98.7}$

Snoke + others, Table 1

Table 11

Table : Representative Analyses of Pyroxenes

Weight Percent of Oxides

	1	2	3	4	5	6	7	8	9	10	11	12	13
SiO_2	52.49	52.90	53.03	55.47	51.31	51.12	54.04	51.56	53.50	50.85	52.37	51.30	52.05
TiO_2	0.29	0.37	0.31	0.14	0.64	0.63	0.16	0.50	0.16	0.74	0.17	0.22	0.23
Al_2O_3	2.07	2.39	1.99	1.25	3.40	2.85	1.35	3.00	1.55	2.93	1.13	0.73	1.15
Cr_2O_3	0.83	0.34	0.44	0.18	0.11	0.00	0.05	0.13	0.00	0.11	0.05	0.00	0.00
MgO	16.45	16.28	17.32	30.84	15.12	15.72	27.87	15.87	26.88	15.65	13.21	18.41	12.52
CaO	23.07	22.29	22.78	0.89	22.88	21.53	1.16	21.62	0.92	20.44	21.00	1.50	20.95
MnO	0.10	0.15	0.03	0.27	0.33	0.22	0.46	0.24	0.46	0.24	0.53	0.93	0.39
FeO	4.58	5.17	4.59	11.19	6.97	7.56	15.88	7.14	16.80	8.98	11.54	26.61	13.23
Na_2O	0.21	0.22	0.23	0.00	0.26	0.34	0.00	0.28	0.00	0.41	0.29	0.00	0.35
Total	100.09	100.11	100.69	100.23	101.02	99.97	100.97	100.34	100.27	100.35	100.29	99.70	100.87

Catatoms Normalized to 4

	1	2	3	4	5	6	7	8	9	10	11	12	13
Si	1.920	1.936	1.920	1.951	1.874	1.882	1.925	1.890	1.930	1.871	1.963	1.965	1.950
Ti	0.008	0.010	0.008	0.004	0.017	0.018	0.004	0.014	0.004	0.021	0.005	0.006	0.006
Al	0.089	0.103	0.085	0.052	0.146	0.124	0.057	0.130	0.066	0.127	0.050	0.033	0.051
Cr	0.024	0.010	0.013	0.005	0.003		0.002	0.004		0.003	0.002		
Mg	0.897	0.888	0.935	1.618	0.823	0.863	1.481	0.868	1.445	0.859	0.783	1.051	0.699
Ca	0.904	0.874	0.884	0.033	0.895	0.850	0.044	0.850	0.035	0.806	0.843	0.062	0.841
Mn	0.003	0.005	0.001	0.008	0.010	0.007	0.014	0.007	0.014	0.008	0.017	0.030	0.012
Fe	0.140	0.158	0.139	0.329	0.213	0.233	0.473	0.219	0.506	0.276	0.362	0.852	0.415
Na	0.015	0.016	0.016		0.018	0.024		0.020		0.029	0.021		0.025
$(Fe,Mn)_2Si_2O_6$	7.069	8.127	6.885	16.719	10.927	11.692	23.865	11.089	25.587	13.801	18.769	43.880	21.027
$Mg_2Si_2O_6$	44.343	44.309	46.014	80.[illegible]5	40.280	42.107	72.546	42.546	71.032	41.741	36.588	52.247	34.436
$Ca_2Si_2O_6$	42.269	41.154	41.504	0.243	40.591	39.006	0.747	38.854	0.123	36.701	41.048	2.242	40.788
$CaTiAl_2O_6$	0.776	1.008	0.826	0.361	1.707	1.714	0.416	1.350	0.413	1.994	0.476	0.635	0.635
$NaAlSi_2O_6$	1.450	1.546	1.599		1.776	2.334		1.929	0.024	2.819	2.077		2.505
$CaAl_2Si_2O_6$	1.720	2.879	1.925	1.961	4.414	3.138	2.277	3.874	2.813	2.621	0.890	0.996	0.600
$CaCrAlSiO_6$	2.374	0.977	1.248	0.501	0.304	0.009	0.149	0.358	0.009	0.323	0.152		0.009

14	15	16
51.25	51.77	51.17
0.22	0.11	0.30
0.60	0.74	0.59
0.00	0.02	0.00
17.92	11.23	15.16
1.54	20.76	1.63
0.72	0.42	1.05
28.48	14.69	31.32
0.00	0.27	0.00
100.73	100.01	101.22
1.954	1.974	1.977
0.006	0.003	0.009
0.027	0.033	0.027
	0.001	
1.019	0.638	0.873
0.063	0.848	0.067
0.023	0.013	0.034
0.908	0.469	1.012
1.019	0.020	
45.970	23.914	52.321
50.268	31.672	43.642
2.427	41.741	2.689
0.609	0.309	0.872
	1.976	
0.726	0.337	0.474
	0.052	0.001

1 Dunite - 332
2 Wehrlite - 412
3 Hornblende-olivine clinopyroxenite - 417
4 Hornblende-olivine clinopyroxenite - 200
5 Magnetite-hornblende-olivine clinopyroxenite - 390
6 Olivine melagabbro - 194
7 Olivine melagabbro - 194
8 Olivine gabbro - 424
9 Olivine gabbro - 424
10 Leucocratic olivine gabbro - 328
11 Two-pyroxene-biotite diorite - 213
12 Two-pyroxene-biotite diorite - 213
13 Two-pyroxene monzodiorite - 255
14 Two-pyroxene monzodiorite - 255
15 Two-pyroxene-biotite monzodiorite - 361
16 Two-pyroxene-biotite monzodiorite - 361

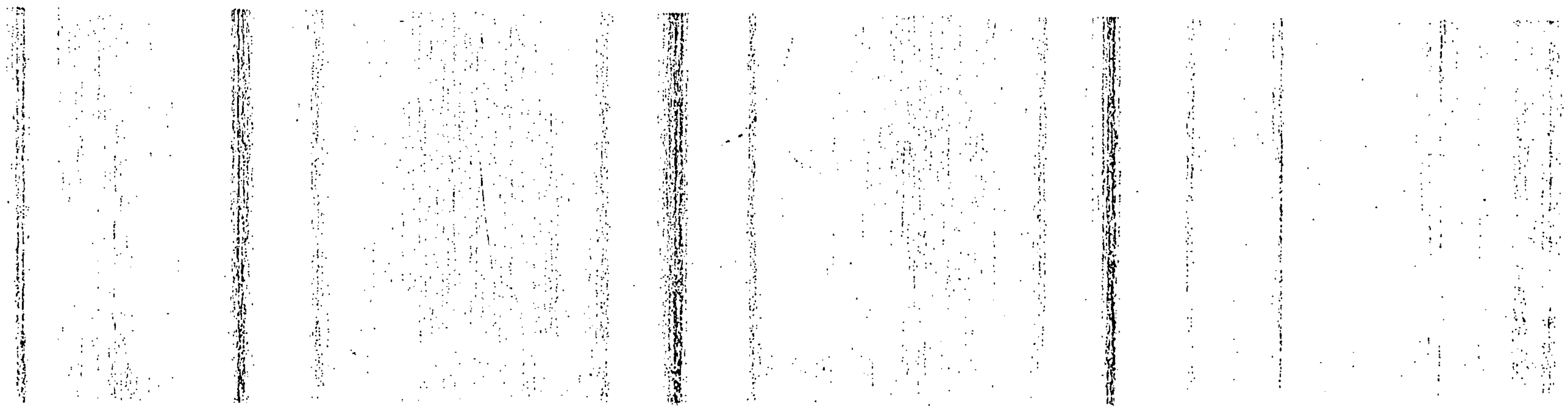


Table : Representative Analyses of Amphiboles

Weight Percent of Oxides

	1	2	3	4	5	6	7
SiO_2	47.43	45.48	42.90	42.63	42.84	44.00	43.03
TiO_2	1.53	1.61	2.46	2.96	1.63	1.47	1.97
Al_2O_3	9.24	10.89	12.25	11.43	11.22	10.11	9.48
Cr_2O_3	0.81	0.55	0.08	0.13	0.12	0.01	0.01
MgO	17.69	17.45	15.82	14.70	14.56	11.02	8.90
CaO	12.28	12.63	12.26	11.37	11.57	11.38	11.27
MnO	0.10	0.09	0.15	0.17	0.17	0.36	0.29
FeO	7.09	6.83	9.56	11.29	11.87	17.50	20.46
Na_2O	1.11	1.97	2.10	2.30	2.12	1.36	1.57
K_2O	0.27	0.74	0.69	0.88	0.56	0.86	1.41
F	0.00	0.08	0.00	0.00	0.00	0.00	0.00
Cl	0.04	0.03	0.03	0.06	0.03	0.06	0.58
Total	97.58	98.31	98.31	97.91	96.88	98.12	98.84

Catatoms – (Na+K) = 15*

	1	2	3	4	5	6	7
Si	6.733	6.478	6.181	6.261	6.327	6.560	6.559
Ti	0.163	0.173	0.267	0.327	0.181	0.165	0.226
Al	1.547	1.829	2.081	1.979	1.954	1.777	1.704
Cr	0.091	0.062	0.009	0.015	0.014	0.001	0.001
Mg	3.744	3.706	3.399	3.219	3.206	2.450	2.023
Ca	1.868	1.928	1.893	1.789	1.831	1.818	1.841
Mn	0.012	0.011	0.018	0.021	0.021	0.045	0.037
Fe	0.842	0.814	1.152	1.387	1.466	2.183	2.609
Na	0.306	0.544	0.587	0.655	0.607	0.393	0.464
K	0.049	0.134	0.127	0.165	0.106	0.164	0.274
F	0.00	0.036	0.014	0.000	0.000	0.000	0.000
Cl	0.010	0.007	0.007	0.015	0.008	0.015	0.150

1 Wehrlite - 412
2 Hornblende-olivine clinopyroxenite - 200
3 Hornblende-magnetite-olivine clinopyroxenite - 390
4 Olivine melagabbro - 194
5 Olivine gabbro - 424
6 Hornblende-biotite meladiorite - 278
7 Two-pyroxene-biotite monzodiorite - 361

*Calculated assuming all Na and K occupy the A site.

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