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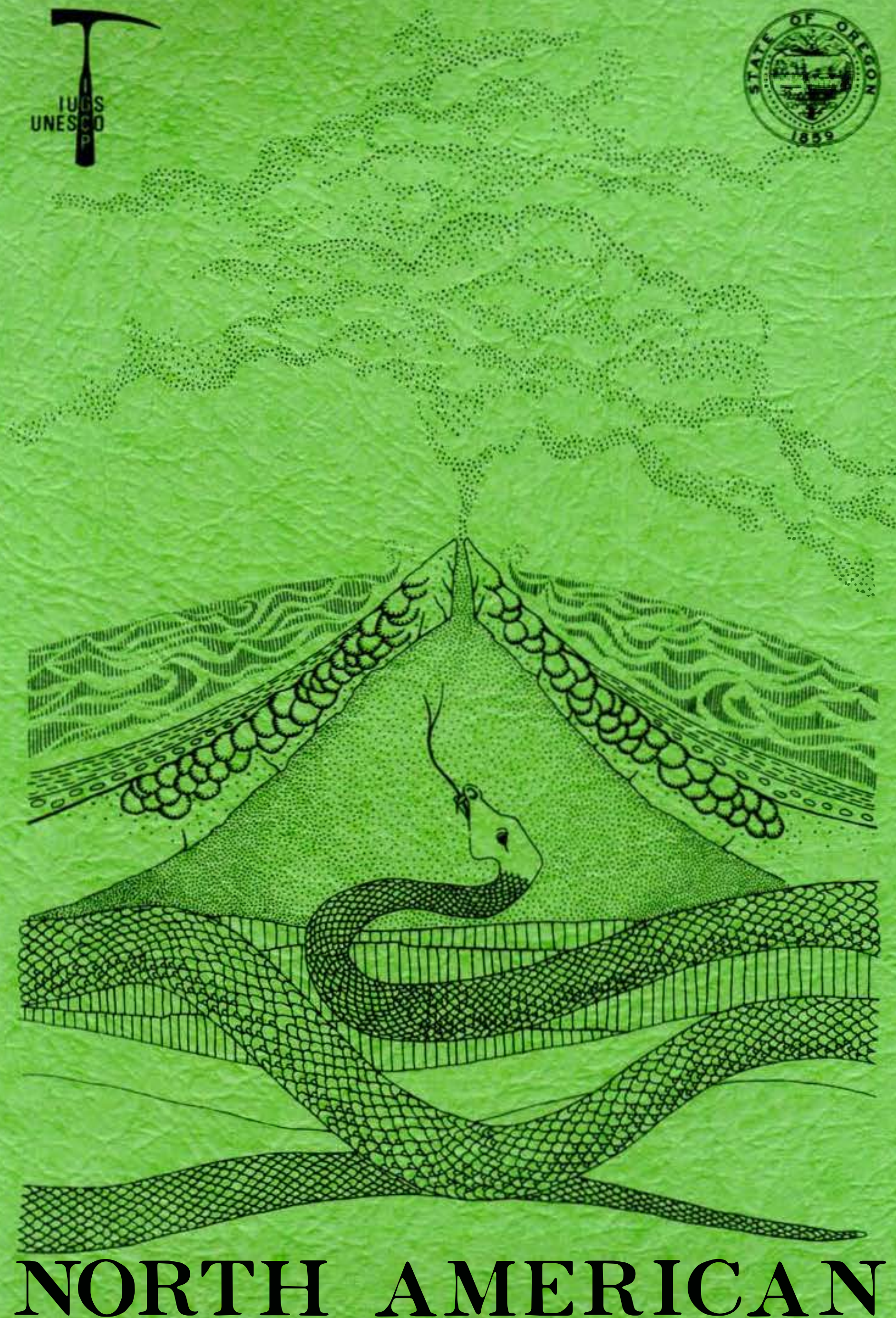

# NORTH AMERICAN OPHIOLITES

1977

STATE OF OREGON
DEPARTMENT OF GEOLOGY AND MINERAL INDUSTRIES
1069 State Office Building, Portland, Oregon 97201

BULLETIN 95

# NORTH AMERICAN OPHIOLITES

Edited by
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U. S. Geological Survey

The preparation of this report was sponsored by the International Geological Program (IGCP) as part of the North American meeting of the IGCP working group: "Ophiolites of continents and comparable oceanic rocks."


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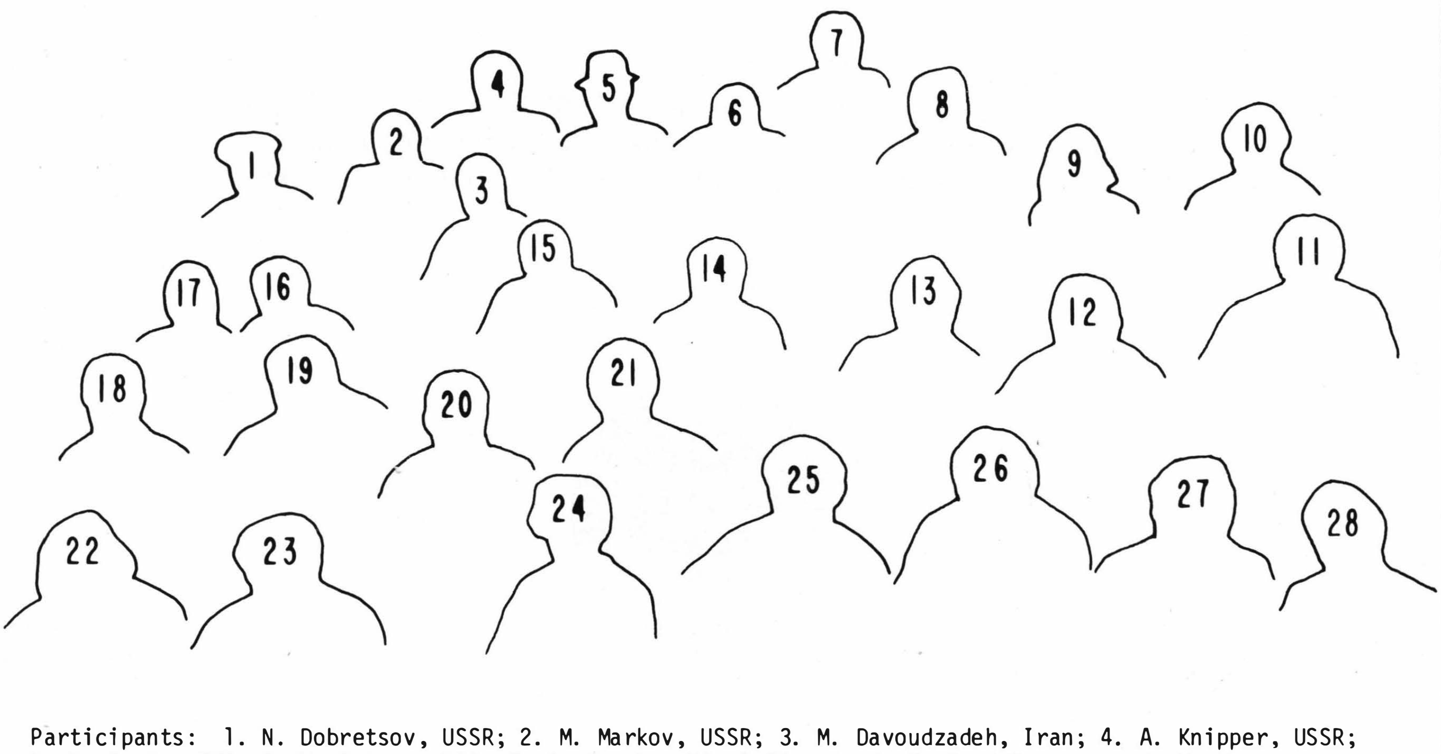


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Participants sitting on South Fork Mountain Schist at the last stop of the Klamath Mountains ophiolite excursion, 1977 (R. G. Coleman, photographer)

FOREWARD

During September 1977, the international project for the study "Ophiolites of continents and comparable oceanic rocks" carried out a series of field excursions and seminars in North America. This project is sponsored by the International Geological Correlation Program (IGCP), which is under the guidance of both the International Union of Geological Sciences (IUGS) and UNESCO. The IGCP projects have been organized to coordinate and stimulate international scientific cooperation in solving fundamental geologic problems in the exploration and appraisal of earth resoures.

The IGCP "ophiolite" project was begun in 1974 by Academician Peyve of the Geological Institute of the Academy of Sciences, USSR, with Dr. N. Bogdanov as Chairman, and has already convened two separate conferences. The first was a 21-day field trip held during October, 1975, in Iran to investigate the ophiolites of that country. During the second, in 1976, the ophiolite working group joined the 17th cruise of the USSR research vessel Dmitry Mendeleev to investigate deep trenches and island ophiolites of the western Pacific.

The North American excursions in September 1977 began in Newfoundland and then proceeded to Quebec where some ophiolites of eastern North America were studied. From there, the group traveled westward to the Stillwater complex in Montana in order to compare a continental igneous cumulate complex with the cumulate sequences found in ophiolites. The final phase of the North American excursions focused on the Klamath Mountains province in southwestern Oregon and northern California. These excursions concluded with a seminar at Stanford University. Geologists participating in the 1977 Klamath Mountains field excursions were from Canada, Columbia, France, Great Britain, Iran, Italy, Japan, Soviet Union, Switzerland, United States, and Yugoslavia.

The term ophiolite is used not for a single kind of rock but rather for a particular sequence consisting, from base to top, of peridotite, gabbro, diabase dike swarm, and pillow lava. These sequences are considered to be oceanic crust stranded on land. Ophiolites have major economic significance because some contain deposits of chromium, platinum, nickel, copper, and asbestos. Ophiolites have acquired increased geologic significance since the development of widespread interest in the theory of plate tectonics. Understanding of them is critical to the solution of questions regarding the evolution of the North American continent and particularly the origin of ore deposits in the Klamath Mountains. This bulletin will provide background for participants of the 1977 field excursions and also will serve as a general reference on North American ophiolites for others interested in economic development of an scientific research on these rocks.

We are grateful to the National Science Foundation and to UNESCO for financial support of the Klamath Mountain field trip and to the U.S. Geological Survey for logistical support. Mary Donato, U.S. Geological Survey, was very helpful in the final assembly of these papers. The generosity of the Oregon State Department of Geology and Mineral Industries in publishing this bulletin will further the scientific aims of this IGCP project.

September 1977

Robert G. Coleman and William P. Irwin, Editors
U.S. Geological Survey, Menlo Park, California

The Oregon Department of Geology and Mineral Industries is pleased to take an active part in the production of this ophiolite publication. The term "ophiolite" embodies a relatively new set of earth science concepts and requires a worldwide perspective for proper interpretation. This book advances the understanding of Oregon ophiolites, not only by information presented in the articles on Oregon ophiolites, but also by data and concepts developed in the other articles. We are proud to assist in the efforts leading to this publication and are certain that this volume will lead to substantial progress in the understanding of ophiolites and, as a result, to practical benefits for mankind in the future.

December 1977

John D. Beaulieu, Deputy State Geologist
Oregon Department of Geology and Mineral Industries

The Oregon Department of Geology and Mineral Industries has agreed to publish these papers because the subject matter is consistent with the mission of the Department. The usual style and standards for this series have been modified to accomodate the style used by the editors of this bulletin. To facilitate timely distribution of information, camera-ready copy submitted by the editors has not been edited by the staff of the Oregon Department of Geology and Mineral Industries.

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## ILLUSTRATIONS



## LIST OF PLATES
(In Pocket)



## COVER DESIGN

The drawing on the front cover is the creation of Susan Engwicht, U.S. Geological Survey, and is an allegorical representation of the formation of ophiolite. The serpent is symbolic of the altered ultramafic rock serpentinite, which was named for its resemblance to snake skin. Also depicted are diabase dikes, shown as parallel vertical lines, and bulbous volcanic extrusions of pillow lavas.


INDEX MAP OF NORTH AMERICA SHOWING AREAS OF OPHIOLITE DESCRIBED IN THIS BULLETIN

# NORTH AMERICAN OPHIOLITES

## DISTRIBUTION AND TECTONIC SETTING OF OPHIOLITES AND OPHIOLITIC MELANGES IN THE APPALACHIAN OROGEN

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ABSTRACT

Ophiolites of highly allochthonous character and associated ophiolitic melange occur along the western margin of the Appalachian Orogen, i.e. the ancient continental margin of eastern North America (Humber Zone). The Bay of Islands Complex is a typical example of a transported ophiolite, and ophiolitic melanges such as the Milan Arm, Second Pond and Coachman's Melanges are all interpreted as related to ophiolite obduction. The melanges vary in structural style from undeformed thin sheets in the west to polydeformed and metamorphosed in the east.

Small mafic-ultramafic bodies in eastern parts of the Humber Zone occur along the entire length of the orogen. Some of these are blocks in ophiolitic melange and others are sited at major structural contacts. They are affected by the full range of deformations and metamorphism related to the destruction of the ancient continental margin of eastern North America. Most are probably dismembered parts of ophiolite suites, and their occurrence from Newfoundland to Alabama implies a similar structural history for western parts of the system.

In central parts of the Appalachian Orogen, ophiolites that represent vestiges of an ancient Iapetus Ocean (Dunnage Zone) occur along the Baie Verte-Brompton Line and they form the basement to island arc sequences farther east. Mafic-ultramafic complexes along eastern parts of the Dunnage Zone in the north may be blocks in melange, and some may represent diapiric intrusions.

Relationships in the northern Appalachians imply a major suture within the exposed parts of the southern Appalachians, either at the Brevard Zone or somewhere between the Brevard Zone and the Carolina Slate Belt.

Appalachian ophiolites are Late Cambrian and Early Ordovician in age, where dated, and they were emplaced and deformed during the Ordovician Taconic Orogeny. Silurian or younger ophiolites are unknown and the stratigraphic record for this period indicates largely terrestrial conditions without the presence of important continental margins or major oceans.

### INTRODUCTION

The interpretation of the on-land ophiolite suite of rock units as oceanic crust and mantle has brought new and considerable interest to the study of ophiolitic rocks in orogenic belts. Because the ophiolite suite of rock units relate in a total way to their place and mode of generation, their presence provides an important adjunct to the interpretation of the regional geology of the orogen in which they are found. In ancient orogenic belts, ophiolite suites occur as highly allochthonous structural slices transported across former continental margins, e.g. western Newfoundland, Oman, Zagros, Himalayas, or as imbricated and deformed slices in central parts of orogens, e.g. northeastern Newfoundland, Ballantrae, Indus Suture, and Pindus Zone. Ophiolitic melanges (Gansser, 1974), or chaotic rocks that contain blocks derived from the ophiolite suite of rock units are associated with allochthonous ophiolites (obduction) or they are related to the destruction of oceanic crust by its descent at oceanic trenches (subduction). The distribution and structural setting of ophiolitic melanges in orogenic belts are in most cases therefore as important to tectonic syntheses as the presence of the ophiolite suite itself.

The common occurrence of mafic-ultramafic rocks in the Appalachian Orogen was first pointed out by Hess (1939; 1955), and since then there have been numerous attempts to explain their origin. Most occurrences are interpreted now as ancient oceanic crust and mantle (Moores, 1970; Stevens, 1970; Church and Stevens, 1971; Dewey and Bird, 1971). Some are part of well-preserved ophiolite suites, e.g. Bay of Islands Complex, others are dismembered ophiolites, e.g. Advocate Complex, and still others form blocks in ophiolitic melange, e.g. Coachman's Melange. In a few cases, local examples are interpreted as intrusions (Chidester and Cady, 1972; Kennedy and Phillips, 1971) or mantle diapirs situated above subduction zones (Stevens and others 1974; Kean, 1974).

Although Hess suggested that there were two ultramafic belts in the Appalachian Orogen, new data require a revision of this view and indicate a wide spectrum of occurrences in a variety of different tectonic settings. Recent geologic syntheses of the Appalachian Orogen are based upon the recognition of contrasting tectonic-stratigraphic zones across the system. Nine zones are defined in the Canadian Appalachians (Williams and others 1972; 1974) and these have been amalgamated for purposes of broad correlation into five zones that are extrapolated along the full length of the system from Newfoundland to Alabama (Williams, 1976). From west to east these zones are given local names in the northern Appalachians, viz: Humber, Dunnage, Gander, Avalon, Meguma (plate 1). Ophiolitic rocks are restricted almost entirely to the Humber and Dunnage Zones.

The model for the development of the Appalachian Orogen follows the suggestion of Wilson (1966) and involves the generation and destruction of a late Precambrian - early Paleozoic Iapetus Ocean (Dewey, 1969, Bird and Dewey, 1970; Stevens, 1970; McKerrow and Cocks, 1977; etc.). The Humber Zone records the development and destruction of an Atlantic type continental margin of eastern North America (Williams and Stevens, 1974). It contains the best examples of transported complete ophiolite suites (e.g. Bay of Islands Complex; Williams, 1973) as well as a variety of ophiolitic melanges in various states of structural complexity. The Dunnage Zone represents the former site of the Iapetus Ocean. In places it exhibits well-developed ophiolite suites (e.g. Betts Cove Complex, Upadhyay and others 1971; Thetford Mines ophiolites, Laurent, 1975), commonly overlain by thick island arc volcanic sequences. As well, it contains melanges that possibly relate to subduction (e.g. Dunnage Melange, Kay, 1976; Fournier Complex, N. Rast, pers. comm. 1975), and in the northeast a belt of mafic-ultramafic complexes that are either blocks in melange or mantle diapirs. The Gander and Avalon Zones developed upon continental crust and lay to the east and southeast of the Iapetus Ocean. Neither contains well-preserved examples of a complete ophiolite. The Meguma Zone may represent the eastern continental margin of an ancient ocean that lay to the east of the Avalon Zone (Schenk, 1971). Like the Gander and Avalon Zones, it too is devoid of ophiolite suites.

In the Canadian Appalachians, the boundary between the Humber and Dunnage Zones is marked by the occurrence of ophiolites in a steep structural belt. These ophiolites can be traced as discontinuous bodies from Baie Verte, Newfoundland, to Brompton Lake, Quebec. Accordingly, the steep ophiolite zone has been termed the Baie Verte-Brompton Line (St. Julien and others 1976). It is an important structural junction in the Northern Appalachians and its ophiolites are host to the asbestos deposits that make this zone the world's richest asbestos belt. In places where the ophiolitic rocks are absent, the Humber-Dunnage boundary is marked by faults that separate Humber Zone metamorphosed clastics (west) and Dunnage Zone less metamorphosed volcanic rocks (east). In other places, the boundary is hidden by Silurian and younger cover rocks, e.g. Gaspé Peninsula (plate 1).

South of the Canada-United States border, the projection of the Baie Verte-Brompton Line is marked by a zone of small isolated ultramafic bodies in Vermont that extends all the way southward to Staten Island, New York. Farther South, the Baltimore Gabbro Complex (Crowley, 1969) of Maryland lies at or near the continuation of the same structural zone, and the zone may be marked by local occurrences of ophiolitic melange in the James River Synclinorium (Brown, 1976). From there, it projects between the Grenvillian basement rocks of the Blue Ridge and Sauratown Mountains, and appears to continue farther south along the Brevard Zone (Hatcher, 1972). In the northern Appalachians, the Baie Verte-Brompton Line marks an ancient continent-ocean interface and it is the most westerly possible root zone for allochthonous ophiolites found farther west. If this same structural zone continues southward, as proposed, the Brevard Zone of the southern Appalachians is an important suture marking the site of the former Iapetus Ocean.

The distrubution and tectonic setting of ophiolitic rocks in the Humber and Dunnage Zones form the basis of the discussion that follows. No attempt is made to describe each individual ophiolite occurrence. Instead, a description of the regional geology and extrapolations along the length of the system for each kind of ophiolite occurrence is followed by a description of a type example. Most examples are taken from the northern Appalachians because of a greater familiarity to the authors and because southerly examples are treated by other contributions to this volume (see papers by Laurent and Morgan).

## OPHIOLITES AND OPHIOLITIC MELANGES OF THE HUMBER ZONE

The Humber Zone consists of a crystalline Grenvillian basement overlain by a thick clastic sequence with associated volcanic rocks, and a prominent Cambro-Ordovician carbonate sequence. Ophiolites of highly allochthonous nature occur in western parts of the zone where they overlie relatively undeformed parts of the autochthonous carbonate sequence. These ophiolites are associated with transported sedimentary rocks and collectively they constitute allochthons emplaced during Middle Ordovician.

The carbonate sequence of the Humber Zone and correlative coarse limestone breccias in overlying structural slices are interpreted as bank and bank foot deposits, respectively, formed at the ancient continental margin of eastern North America (Rodgers, 1968). Underlying clastics that rest on Grenvillian gneisses formed a rise prism at the margin (Williams and Stevens, 1974), and transported ophiolites represent oceanic crust and mantle that lay farther east (Church and Stevens, 1971).

Examples of allochthonous ophiolites in westerly parts of the Humber Zone include the White Hills Peridotite of the Hare Bay Allochthon, the Bay of Islands Complex of the Humber Arm Allochthon, and the Mount Albert ophiolite of the Shick Shock Mountains (Williams, 1975). The Baltimore Gabbro Complex of Maryland occurs in a similar structural position but lies nearer the east boundary of the zone.

Ophiolitic melanges occur across the Humber Zone and they are particularly well-developed in western Newfoundland. Their formation is attributed to the transport of ophiolites from their root zone at the Baie Verte-Brompton Line, across the rise prism and carbonate bank successions, to their present positions. The best exposed and most extensive melanges form integral parts of the Humber Arm and Hare Bay Allochthons, e.g. Companion and Milan Arm Melanges (Williams, 1975), and comparable examples, though lacking ophiolitic blocks in most places, are associated with transported sedimentary rocks in Taconic klippen all the way southward to Harrisburg, Pennsylvania. These melanges are mainly thin subhorizontal sheets between other transported rocks that collectively lie above the carbonate sequence. Ophiolitic melanges occur also at the east margin of the carbonate terrane in Maryland and in western White Bay, Newfoundland, e.g. Second Pond Melange (Williams, 1977a). Farther east, ophiolitic melanges are associated with clastics of the rise prism at the eastern margin of the Humber Zone (e.g. Coachman's Melange, Williams, 1977b).

Deformation and metamorphism increase from west to east across the Humber Zone. Melanges associated with Taconic-type allochthons above the carbonate terrane have been little deformed since formation. These near the present easternmost exposures of the carbonate sequence vary from polydeformed and metamorphosed in Maryland to locally deformed and mildly metamorphosed in Newfoundland. Ophiolitic melanges at the eastern margin of the Humber Zone are everywhere polydeformed and metamorphosed and now bear little resemblance to occurrences farther west.

Small mafic-ultramafic bodies are common throughout the full length of the Appalachian Orogen in the belt of deformed clastic rocks at the eastern margin of the Humber Zone. Although locally interpreted as intrusions, some are clearly blocks in ophiolitic melange and other isolated bodies occur at structural contacts. All are affected by the full range of deformation and metamorphism that accompanied the destruction of the ancient continental margin of eastern North America. Examples in easterly parts of the Fleur de Lys Supergroup in Newfoundland are thought to occur at structural discontinuities (Williams and others, 1977), the Pennington Dike and nearby ultramafic bodies of the Eastern Townships of Quebec are sited at structural contacts (Pierre St. Julien, pers. comm. 1975), the continuous string of small ultramafic occurrences from Vermont to Staten Island marks a zone of nappes and imbricate slices (Barry Doolan, pers. comm. 1976), and many of the small ultramafic bodies of the eastern Blue Ridge from Virginia to Alabama may owe their presence to structural emplacement.

If these small mafic-ultramafic occurrences are intrusions, there is no apparent mechanism or obvious reason for their emplacement into an undeformed rise prism of clastic sediments. More likely, they represent blocks in melange and dismembered ophiolite at structural contacts. Their widespread occurrence within the deformed and metamorphosed clastic terrane at the eastern margin of the Humber Zone implies a similar early tectonic history for the full length of the western part of the Appalachian System.

The distinction between the Humber and Dunnage Zones is subtle in places where deformed ophiolitic melanges, which are incorporated structurally within the Humber Zone clastics, are juxtaposed with ophiolite suites along the Baie Verte-Brompton Line. The latter are a natural part of the Dunnage Zone, but the melanges are grouped in places with the Humber Zone clastic rocks and considered a natural part of local stratigraphic successions. This situation exists at the Baie Verte-Brompton Line in Newfoundland and it may be a common circumstance elsewhere. Stratigraphic studies in metamorphic terranes that include ophiolitic melanges or small ultramafic bodies of possible ophiolitic parentage should be made with extreme caution, as experience has shown that major structural disruptions have gone unnoticed in the polydeformed and metamorphosed rocks immeidately west of the Baie Verte-Brompton Line (Williams and others 1977; Williams, 1977b; Pierre St. Julien, pers. comm. 1976; Barry Doolan, pers. comm. 1976).

## Allochthonous complete ophiolite suites: The Bay of Islands Complex

The Bay of Islands Complex affords an excellent example of an allochthonous complete ophiolite suite that forms the highest structural slice of a composite allochthon in the western part of the Humber Zone. It is represented in four separate massifs, which from south to north are Lewis Hills, Blow Me Down, North Arm Mountain, and Table Mountain (pl. I). All lie in the same structural position and either represent separate transported bodies or erosional remnants of a once continuous slice. Two of the massifs (Blow Me Down and North Arm Mountain) display a completely developed ophiolite stratigraphy, but all four include the basal ultramafic unit.

The sequences of ophiolite units in the three northernmost massifs are disposed in synclines with northeast-trending subhorizontal axes and moderately to steeply dipping limbs. The present tectonic base of each massif is subhorizontal so that the ophiolite

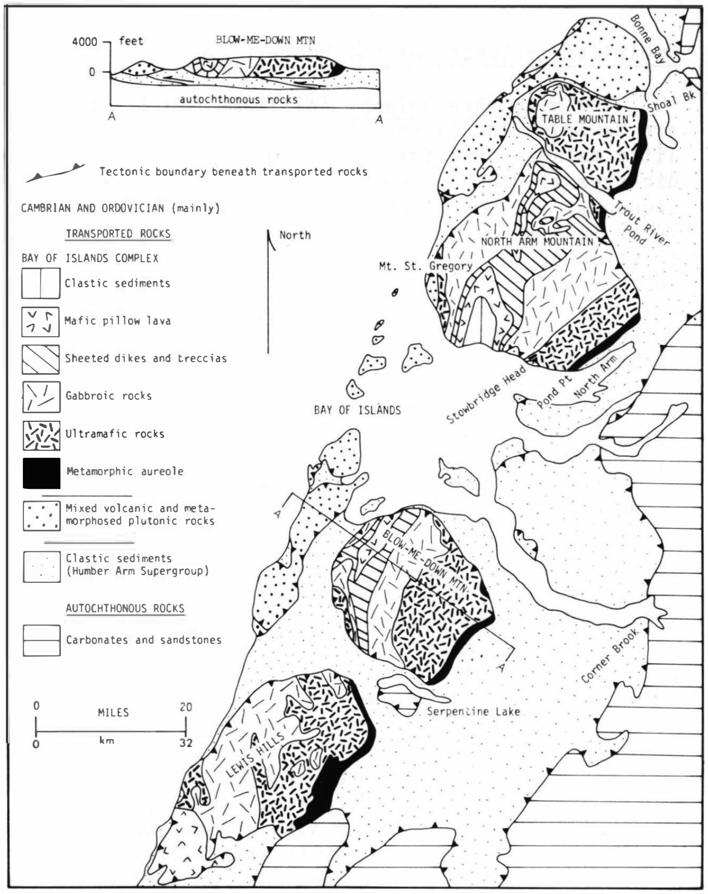


Figure 1: Geologic setting of the Bay of Islands Complex, Western Newfoundland (after Williams and Smyth, 1973).

units are structurally truncated at depth in much the same way as they are truncated at their top by the present erosional surface (see cross-section AA, fig. 1). A contact between the stratigraphic base of the ophiolite sequence and a dynamo-thermal aureole of supracrustal rocks, now frozen into the folded ophiolite slices, is interpreted to represent the actual zone of obduction where the hot oceanic plate moved across the continental margin. The contacts of latest emplacement of the structural slices are marked now by thin zones of shale melange with sedimentary, volcanic, and sparse gabbro and serpentinite blocks. These are the result of mass wastage and tectonic mixing that accompanied gravity sliding.

Trondhjemite from the Bay of Islands Complex has been dated isotopically at 504 m.y. $\pm$ 10 (Mattinson, 1976) and amphiboles from its dynamo-thermal aureole at 460 to 470 m.y. (Dallmeyer and Williams, 1975; Archibald and Farrar, 1976). The former gives the time of generation of the ophiolite suite as Late Cambrian. The latter indicates an Early Ordovician age for initial displacement and agrees with the sedimentologic evidence of ophiolite detritus in Lower Ordovician sedimentary rocks deposited during ophiolite transport (Stevens, 1970). An upper limit to the time of final emplacement is given by the Middle Ordovician age of the neo-autochthonous Long Point Formation (Bergström and others 1974). For more detailed descriptions the reader is referred to Smith (1958) and Malpas (this volume).

Ophiolitic melanges of the Humber Zone

The Milan Arm Melange (Williams, 1975) of northern Newfoundland is the clearest example of an ophiolitic melange that forms an integral part of a Taconic-type allochthon above the Humber Zone carbonate sequence. It structurally overlies autochthonous rocks in some places and intervening transported clastics of the rise prism (Maiden Point Formation) in other places. It is in turn structurally overlain by the ophiolitic White Hills Peridotite. The melange is similar in most respects to all of the shale melanges that separate structural slices of the west Newfoundland allochthons, however, it has a much wider variety of exotic blocks; some up to a kilometer or more across and resembling the largest slices that make up the west Newfoundland allochthons.

Its commonest blocks are serpentinized peridotite, mafic volcanic rocks, amphibolites, foliated gabbro, greywacke, diorite, and exceptionally coarse grained pyroxenite and hornblendite associated with tonalite and hornblende-biotite schist. Nephrite blocks are also known locally (R.K. Stevens, pers. comm. 1976). Most of these rock types can be matched directly with rocks in nearby structural slices, but a few are of unknown origin.

Many of the amphibolite, gabbro, and diorite blocks are encased in a relatively thin, hard rind of light grey calc-silicate alteration products (rodingite). The tough and resistant alteration halos form coastal wave-washed outcrop surfaces where the matrix shales are eroded. In some examples the rodingite alteration halos are surrounded in turn by a thinner serpentinite coating, implying that the rodingite represents an alteration zone between mafic rocks and serpentinite. These blocks appear therefore to have been once immersed in serpentinite or serpentinite melange, so that they are recycled where they now occur in a shale matrix.

Local serpentinite and gabbro blocks in melanges at the base of the west Newfoundland allochthons imply that the sequences of transported slices were emplaced as already-assembled allochthons (Stevens and Williams, 1973).

The recently recognized Coachman's Melange (Williams, 1977b) is an example of a polydeformed and metamorphosed ophiolitic melange that occurs at the eastern margin of the Newfoundland Humber Zone. It is closely associated with psammitic schists of the rise prism (Fleur de Lys Supergroup) and it has been affected by the full range of deformations recognized in nearby rocks. The melange occurs in a multitude of narrow zones that rarely exceed more than 50 m in structural thickness. If all occurrences represent a complexly folded single unit, exceedingly tight isoclines of more than 3 km amplitude affected easternmost local exposures of the rise prism.

The Coachman's Melange has a black pelitic matrix with conspicuous deformed and recrystallized ultramafic blocks now represented by bright green actinolite-fuchsite schist. Sedimentary blocks with ill-defined outlines are common everywhere, and in some places large serpentinized ultramafic blocks, foliated gabbro blocks, and marble are also known.

Actinolite-fuchsite schist occurs in lenses from 10 cm to 3 m in length and rarely more than 1 m in width. They exhibit minor folds and folded schistosities identical to structures in the surrounding schistose matrix and nearby psammitic schists. Pale green actinolite crystals are locally 2-4 cm in length, set in a fine-grained fuchsite-carbonate matrix. An ultramafic origin for these rocks is indicated by their mineralogy and because larger nearby ultramafic blocks are recrystallized to similar mineral assemblages at their margins. Interior parts of large ultramafic blocks are in places brecciated, and this feature predates both serpentinization and incorporation into the melange.

Recognition of the Coachman's Melange and an appreciation of its complex structural history bears upon one of the major problems of northern Appalachian geology, i.e. the timing of deformation and metamorphism within the rise prism in relation to the time of generation of nearby ophiolite suites and the time of their transport across an ancient continental margin. As is the case with other worldwide examples of ophiolitic melanges, the Coachman's Melange implies transport of oceanic crust across the rise prism represented by the local Fleur de Lys Supergroup. This transport is equated most reasonably with the emplacement of highly allochthonous ophiolites in western parts of the Humber Zone from an initial position to the east of the Baie Verte-Brompton Line. Similar structural histories for both the Coachman's Melange and nearby parts of the Fleur de Lys Supergroup indicate that the rise prism was undeformed at the time of melange formation and initial ophiolite displacement. This conclusion leads to a simple model for the place of origin and time of transport of ophiolitic complexes in western Newfoundland compared to the time of deformation and metamorphism in the intervening Fleur de Lys terrane (fig. 2). As well, it explains the marked structural contrasts between the Fleur de Lys Supergroup and nearby ophiolite suites, while implying a mechanism for deformation and metamorphism through ophiolite transport and structural loading at the ancient continental margin.

Ophiolitic melanges comparable to those at Coachman's Harbour, Newfoundland are unknown elsewhere in the Appalachian Humber Zone. Other occurrences are predicted because of structural similarities along the length of the system.

## OPHIOLITES AND OPHIOLITIC MELANGES OF THE DUNNAGE ZONE

Ophiolites and ophiolitic melanges are represented in the Dunnage Zone from Newfoundland to Virginia. Farther south, rocks typical of the Dunnage Zone are absent and the Humber Zone is bordered eastward by crystalline rocks of the Inner Piedmont (Hatcher, 1972). Relationships in the northern Appalachians predict a major suture in the southern Appalachians, either at the Brevard Zone or somewhere between this zone and Avalon Zone equivalents of the Carolina Slate Belt.

The most prominent ophiolite occurrences in the Dunnage Zone are found at its western margin along the Baie Verte-Brompton Line. In Newfoundland, examples can be traced from Baie Verte of the Burlington Peninsula, e.g. Advocate and Point Rousse Complexes (Williams and others 1977) to Glover Island of Grand Lake. From there, the Baie Verte-Brompton Line is ill-defined, but it is probably coincident with the Cape Ray Suture (Brown, 1973) farther south, and transported ophiolites at Cape Ray presumably root in this zone. In mainland Canada, volcanic rocks and deformed mafic-ultramafic rocks of the Fournier Complex, New Brunswick may lie at or near the Baie Verte-Brompton Line Farther west, the line is

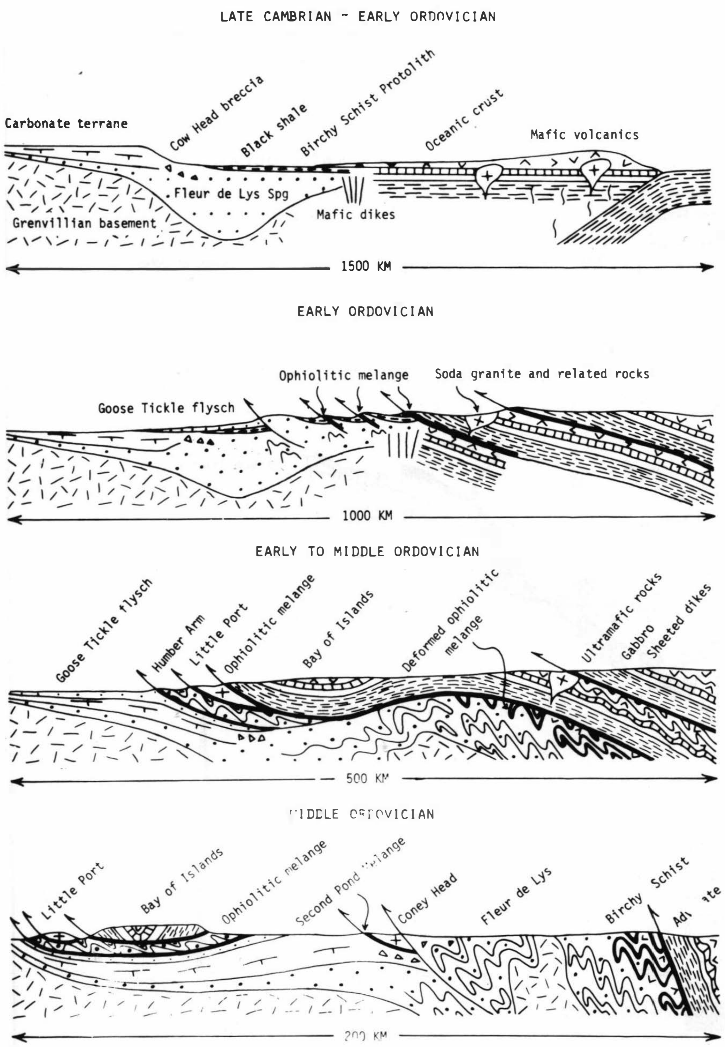


Figure 2: Model for the development of ophiolitic melange and the transport of ophiolites across the Humber Zone (after Williams, 1977b).

covered by Upper Ordovician rocks of the Metapedia Belt and Silurian rocks of the Gaspé Synclinorium. Where the Baie Verte-Brompton Line reappears in the Eastern Townships of Quebec, it is marked by major ophiolite occurrences along most of its length.

The Baie Verte-Brompton Line is a zone of intense deformation and ophiolites along its length are in most places thrust-imbricated and penetratively deformed. The ophiolite suites are mainly steeply-dipping to overturned, but their stratigraphic sections face eastwards. In Newfoundland, the Baie Verte-Brompton Line is a zone of marked gravity gradient with negative anomalies on its western continental side, and positive anomalies across oceanic rocks to the east (Miller and Deutsch, 1976).

East of the Baie Verte-Brompton Line, separate ophiolite occurrences are common throughout central parts of the Dunnage Zone. These form the basement to thick volcanic arc sequences that are locally dated as Lower Ordovician, thus defining an upper age limit for the underlying ophiolite suites. In Newfoundland, the Betts Cove Complex is probably the best known example, but sheeted dikes occur throughout the central volcanic terrane (Strong, 1972) and gabbros and ultramafic rocks occur locally at South Pond, Brighton and Gull Island of Cape St. John. In Maine, mafic-ultramafic rocks along the southern margin of the Chain Lakes massif are interpreted as ophiolites (G.M. Boone, pers. comm. 1975). Their occurrence and relationships to nearby crystalline rocks are still poorly understood.

Olistostromes that contain ophiolitic blocks occur along the east margin of the Thetford Mines ophiolite belt, e.g. St. Daniel Formation (St. Julien and Hubert, 1975), and megaconglomerates with outcrop size gabbro and penetratively deformed and altered ultramafic blocks occur along the east margin of the Advocate Complex in Newfoundland. The significance and time of deposition of these rocks is still poorly understood, but in Newfoundland, deposition post-dates earliest deformations in nearby ophiolites and predates Silurian volcanism. Farther south in the New England Appalachians, small metamorphosed ultramafic bodies in sulphidic schists of the Partridge Formation may represent blocks in olistostrome rather than small intrusions.

The Dunnage Melange (Kay, 1976), which lies to the east of the central Newfoundland island arc terrane, is not itself ophiolitic but a similar melange, 20 km eastward at Carmanville, contains sparse ultramafic blocks. The Carmanville melange may represent a subsurface continuation of the Dunnage. The Dunnage Melange has been interpreted as an oceanic trench fill (Dewey and Bird, 1971; Williams and Hibbard, 1976; Kay, 1976). If so, then ophiolitic melange would be expected in this position.

A prominent belt of mafic-ultramafic complexes along the eastern margin of the Dunnage Zone in Newfoundland, and east of the Dunnage Melange, remains poorly understood in present models for the development of the Appalachian Orogen. Some occurrences contain ultramafic rocks, gabbros and volcanic rocks that collectively are reminiscent of an ophiolite suite, e.g. Pipestone Pond (Kean, 1974). Others are mainly clinopyroxenite bodies in structural contact with surrounding dark shales, e.g. Gander River Belt, and still others may represent differentiated intrusions of mantle derivation, e.g. Great Bend of Gander River (Stevens and others 1974). At one locality on the north shore of Gander Lake, penetratively deformed ultramafic rocks are unconformably overlain by conglomerates of probable Caradocian age, thus defining an upper age limit for some of these occurrences.

Transported ophiolites of the Humber Zone, ophiolites at the Baie Verte-Brompton Line, and ophiolites that form the basement to island arc sequences across the Dunnage Zone may all relate to a single cycle of ophiolite generation (as summarized in Figure 2). This simple view is contrasted with an earlier interpretation that relates each ophiolite belt to an equal number of small ocean basins that formed, at least in part, after deformation and metamorphism of the Humber Zone continental rise prism (Dewey and Bird, 1971; Kennedy, 1975; Kidd, in press).

## Ophiolites at the Baie Verte-Brompton Line

Ophiolites along the Baie Verte-Brompton Line are bounded to the west by polydeformed and metamorphosed clastic rocks of the Humber Zone, and they are bounded to the east by olistostromes and volcanic sequences. Several occurrences in Newfoundland and Quebec are overlain by thick volcanic sequences, which are similar to volcanic arc sequences found above ophiolites farther east. The Advocate and Point Rousse Complexes along the Baie Verte-Brompton Line

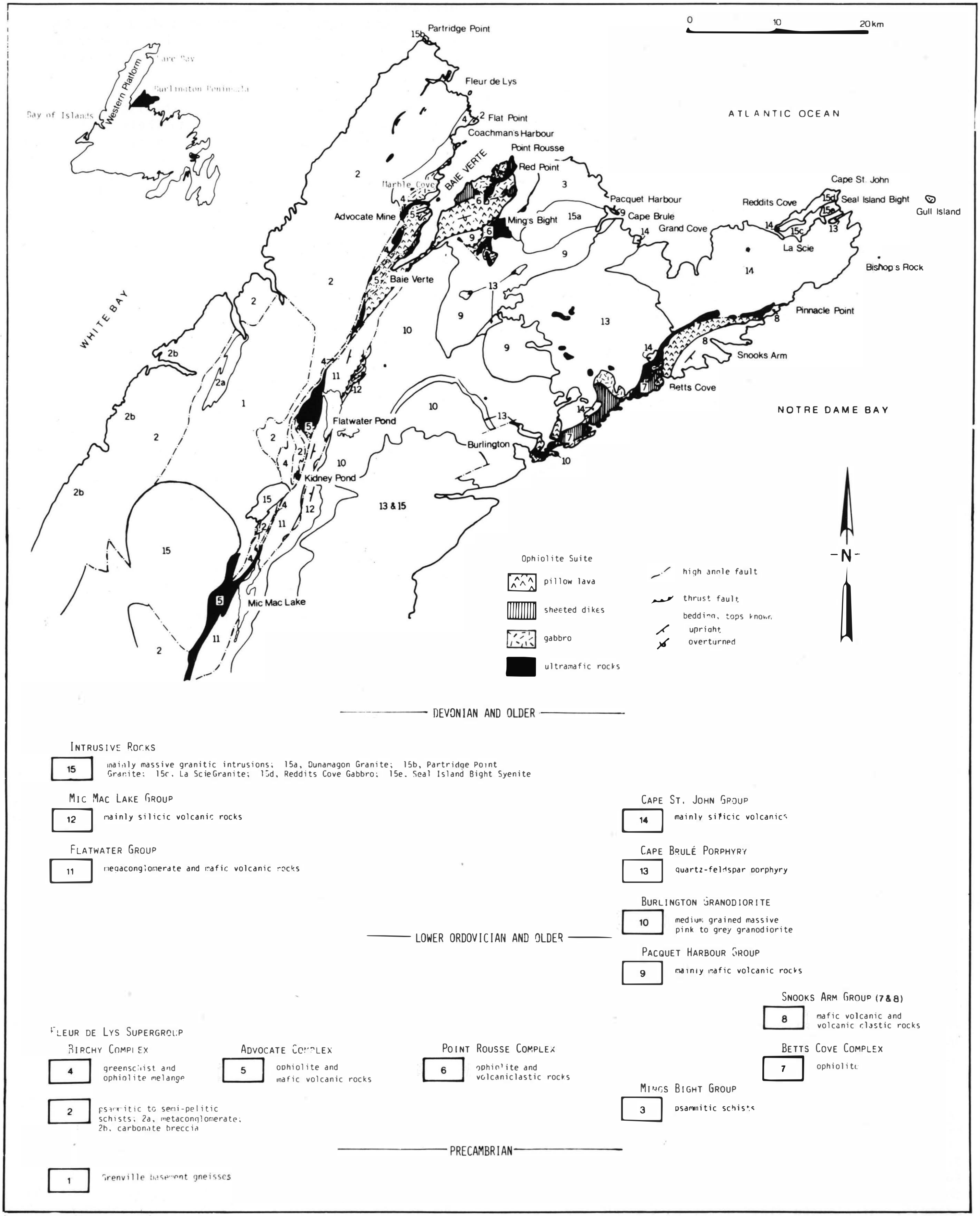


Figure 3: Distribution of ophiolites at and near the Baie Verte-Brompton Line, Burlington Peninsula, Newfoundland (After Williams and others, 1977).

in northern Newfoundland are taken as a type example and their distribution and relationships to nearby groups are summarized in Figure 3.

The Advocate Complex can be traced from Baie Verte 60 km southward beyond Mic Mac Lake. It has a steep northeast-trending foliation in most places and it is cut by numerous steep northeast-trending shear zones that repeat its rock units. A large ultramafic body at its northwest margin near Baie Verte consists of intensely foliated and fractured serpentinites that are host to the Advocate asbestos deposit. Its gabbros vary from massive to intensely foliated and some are distinctive white-altered rocks with green fuchsitic smears, e.g. those at Marble Cove. Sheeted dikes are best exposed 1 km northeast of Baie Verte and well-preserved pillow lavas occur nearby.

Black and green slates, black pebbly slates and chaotic black slaty melange are also present in the Advocate Complex. Some of these rocks occur at tectonic contacts so that they are structurally co-mingled with the segmented ophiolite. Their chaotic character results, in part, from tectonic processes that accompanied imbrication of the ophiolite suites. Others may be depositional breccias. Some of these chaotic shale zones in the Advocate Complex resemble nearby chaotic rocks and black schists of the Coachman's Melange, so that all appear to mark significant structural dislocations.

The Point Rousse Complex is made up of several distinct structural blocks with their tectonic boundaries marked by foliated serpentinite or foliated carbonate-talc-fuchsite alterations of ultramafic rocks (Norman and Strong, 1975). Three separate blocks on Point Rousse Peninsula contain southeast-facing, overturned sections of gabbro, sheeted dikes and pillow lava. The most complete section occurs south of Red Point where a northwest-dipping, southeast-facing sequence of gabbro, sheeted dikes and pillow lava is followed by local chert beds and a thick section of volcaniclastic rocks, all southeast-facing. Locally, on the western side of the Point Rousse Peninsula, pillow lavas of the complex are intensely deformed and converted to greenschists. Southward thrusting of gabbros above the greenschists postdates the development of a steep foliation in the mafic volcanic rocks.

The Point Rousse Complex is less altered and deformed than the Advocate Complex, except in local zones of intense deformation. Farther east, the Betts Cove Complex is even less deformed. The regional distribution of ophiolite complexes across the Burlington Peninsula and the pattern of their deformation suggest westward imbrication of east facing ophiolitic suites with pervasive intense deformation in lower levels (Coachman's Melange, Advocate Complex) and less intense deformation and fewer deformed zones higher in the structural pile (Point Rousse Complex, Betts Cove Complex).

## Ophiolites beneath volcanic arc sequences

The Betts Cove Complex (fig. 4) is the clearest example of an ophiolite suite that forms the basement to a volcanic arc sequence. It consists of a basal ultramafic member overlain transitionally by a poorly developed gabbroic member, in turn overlain by a sheeted dike complex that consists of practically 100% mafic dikes (Upadhyay and others 1971). The sheeted dike complex is faulted against nearby mafic volcanic rocks, but locally the contact is gradational

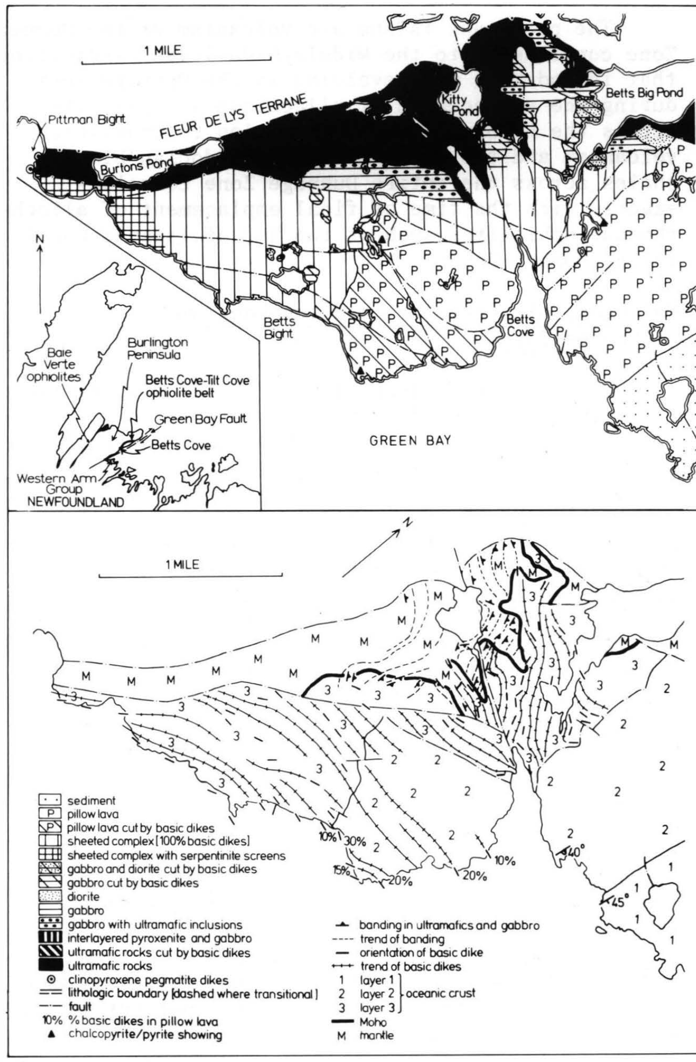


Figure 4: Geology of the Betts Cove Complex and its comparison to models of oceanic crust and mantle (after Upadhyay and others 1971).

across a narrow zone that shows a large decrease in the percentage of dikes over a short distance. Overlying Lower Ordovician rocks of the Snooks Arm Group are nearly 4 km thick and comprise a conformable sequence of pillow lavas, cherts, argillites, andesitic pyroclastic rocks and immature volcanic sediments. A lowermost pillow lava unit constitutes the upper part of the Betts Cove Complex so that a completely conformable transition exists from the ultramafic member of the ophiolite complex to the top of the thick overlying mainly volcanic succession.

Nearby correlatives of the Snooks Arm Group show a lithological evolution from lowermost pillow lavas upward through marine cherts and turbidites into pyroclastic rocks and volcaniclastic sedimentary rocks capped by limestone and subareal tuffs. The overall deep to shallow water lithic change is accompanied by geochemical changes in the volcanic stratigraphy from low potassium tholeiites at the base to calc-alkaline low-silica andesites toward the top that show progressive enrichment in $Al_2O_3$ and $K_2O$ and a decrease in CaO and MgO (Kean and Strong, 1975). Accordingly, the Dunnage Zone volcanic sequences are interpreted as analogues to modern island arcs and the relationships at Betts Cove leave little doubt that volcanic rocks of the Dunnage Zone, at least in westerly exposures, directly overlie oceanic crust.

The prominent island arc volcanism of the Dunnage Zone continued into the Middle Ordovician, indicating that island arcs were evolving in the Dunnage Zone during the emplacement of allochthonous ophiolites across the Humber Zone. A cessation to Ordovician volcanism and the deposition of Caradocian black shales across the entire Dunnage Zone coincides exactly with the time of final emplacement of allochthonous ophiolites such as the Bay of Islands Complex in the Humber Zone.

### Melanges, olistostromes and megaconglomerates of the Dunnage Zone

The Dunnage Melange, though devoid of ophiolitic components, is included in this analysis because of its similarity and possible subsurface continuity with melange at Carmanville that locally contains ultramafic and gabbroic blocks. As well, it is a prominent feature of the Dunnage Zone and its interpretation as an oceanic trench deposit bears heavily upon models proposed for the development of the northern Appalachians.

The melange is a strikingly heterogeneous deposit composed of blocks of mainly clastic sedimentary and mafic volcanic rocks enveloped in a dark scaly shale matrix. It is well-exposed along the rugged coast and clusters of islands of the Bay of Exploits, where it extends for 40 km along strike with a maximum outcrop width of 10 km (fig. 5). Its clasts vary in size from granules and cobbles to boulders and huge blocks up to a kilometre in diameter, thus producing a chaotic mosaic that contrasts sharply with nearby stratified volcanic and sedimentary rocks. Most blocks are indigeneous to nearby volcanic arc sequences and they can be matched with formations of the Exploits and Summerford Groups (Williams and Hibbard, 1976). Shale is much more important in the melange than in nearby terranes.

The Dunnage Melange overlies and interdigitates with the New Bay Formation of the Exploits Group in the southwest, and it has an apparent ghost stratigraphy comparable to that of the Exploits Group. Its matrix is Tremadocian (Hibbard and others 1977) and the melange is overlain by Caradocian black shales toward the northwest. These are succeeded by greywackes and Silurian conglomerates that are coarser and of shallower water deposition higher in the stratigraphic section. The sequence of units above the melange can be viewed therefore as representing the gradual sedimentary infilling of a marine trough, or an upward shoaling sequence built upon a melange basement.

A variety of small intrusions that are localized within the melange terrane are rare or absent in surrounding country rocks. These are mainly quartz-feldspar porphyries and related rocks, which in places contain numerous small mafic and ultramafic inclusions. Dated isotopically as Early to Middle Ordovician and exhibiting relationships suggesting contemporaneity with melange formation (Williams and Hibbard, 1976), the intrusions imply a direct magmatic linkage with deeper parts of the crust. Mafic and ultramafic inclusions in the porphyries indicate that the melange is underlain by a mafic-ultramafic substrate.

West of the Dunnage Melange, the main volcanic sequences of Notre Dame Bay are interpreted as island arc accumulations built upon oceanic crust. These are bordered to the southeast by mixed sedimentary and volcanic rocks of the Exploits Group, that interdigitate farther southeastward with the Dunnage Melange. The melange is interpreted therefore to occupy a fore arc area, based on the geographical distribution of these ancient elements and their similarity to that outlined for modern volcanic arcs (Dickinson, 1974; Karig, 1974). There is no evidence that the Dunnage Melange was ever buried in an actual subduction zone. It is most reasonably considered therefore as a trench-slope deposit that overlies an accretionary prism, analogous to the positioning of some modern melanges with respect to the arc and oceanic trench (Seely and others 1974; Karig and Sharman, 1975).

A zone of megaconglomerates and olistostromal melange with local ophiolitic blocks occurs along the east margin of the Thetford Mines ophiolite belt of Quebec. These rocks are known as the St. Daniel Formation and they rest conformably on basic volcanic rocks of the ophiolite suite. The unsorted rocks contain fragments of greywacke, quartz arenite, shale, siltstone, volcanic rocks, and outsize serpentinite blocks, all set in a dark grey to red and green shale matrix. The age of the St. Daniel Formation is unknown, more than that it is overlain by the Middle Ordovician Beauceville Formation.

The St. Daniel shale-melange facies has been interpreted as an offshore oceanic deposit equivalent to clastics of the Humber Zone rise prism, i.e. Rosaire and Caldwell Groups of Quebec (St. Julien and Hubert, 1975). It has been interpreted also as a subduction related melange (Laurent, 1975).

In Newfoundland, unsorted shale matrix megaconglomerates occupy a similar position to the St. Daniel Formation where they lie to the east of the Advocate Complex (included in Flatwater Group of Figure 3). These locally contain gabbro blocks up to tens of metres in diameter, a variety of sedimentary and volcanic clasts, granodiorite pebbles, altered and deformed ultramafic blocks, and rare semipelitic schist blocks. Deposition of the Newfoundland examples followed deformation in nearby ophiolites and deformation in sedimentary rocks of the rise prism to the west, i.e. Fleur de Lys Supergroup. These examples are interpreted therefore as coarse slump conglomerates derived mainly from deformed ophiolitic rocks at a destroyed continental margin. They are therefore not thought to be correlative with sediments of the rise prism or connected with subduction, as has been suggested for the Quebec examples.

### Dismembered ophiolites or mantle diapirs at the eastern margin of the Dunnage Zone

Mafic-ultramafic bodies along the east side of the Dunnage Zone in Newfoundland (Gander River Belt) have been interpreted as blocks in melange or as mantle diapirs related to subduction and intruded into the country rocks. One occurrence on the north shore of Gander Lake is overlain by conglomerate of presumed Middle Ordovician age, indicating an upper time limit for the age of some of these bodies.

The largest body at Pipestone Pond is approximately 16 km long and 5 km wide, and it is composed mainly of pyroxenite, gabbro, diorite and serpentinized equivalents. It is faulted against metasediments to the east and it is followed westward by volcanic rocks that may be an integral part of the plutonic complex.

A nearby occurrence at Great Bend of Gander

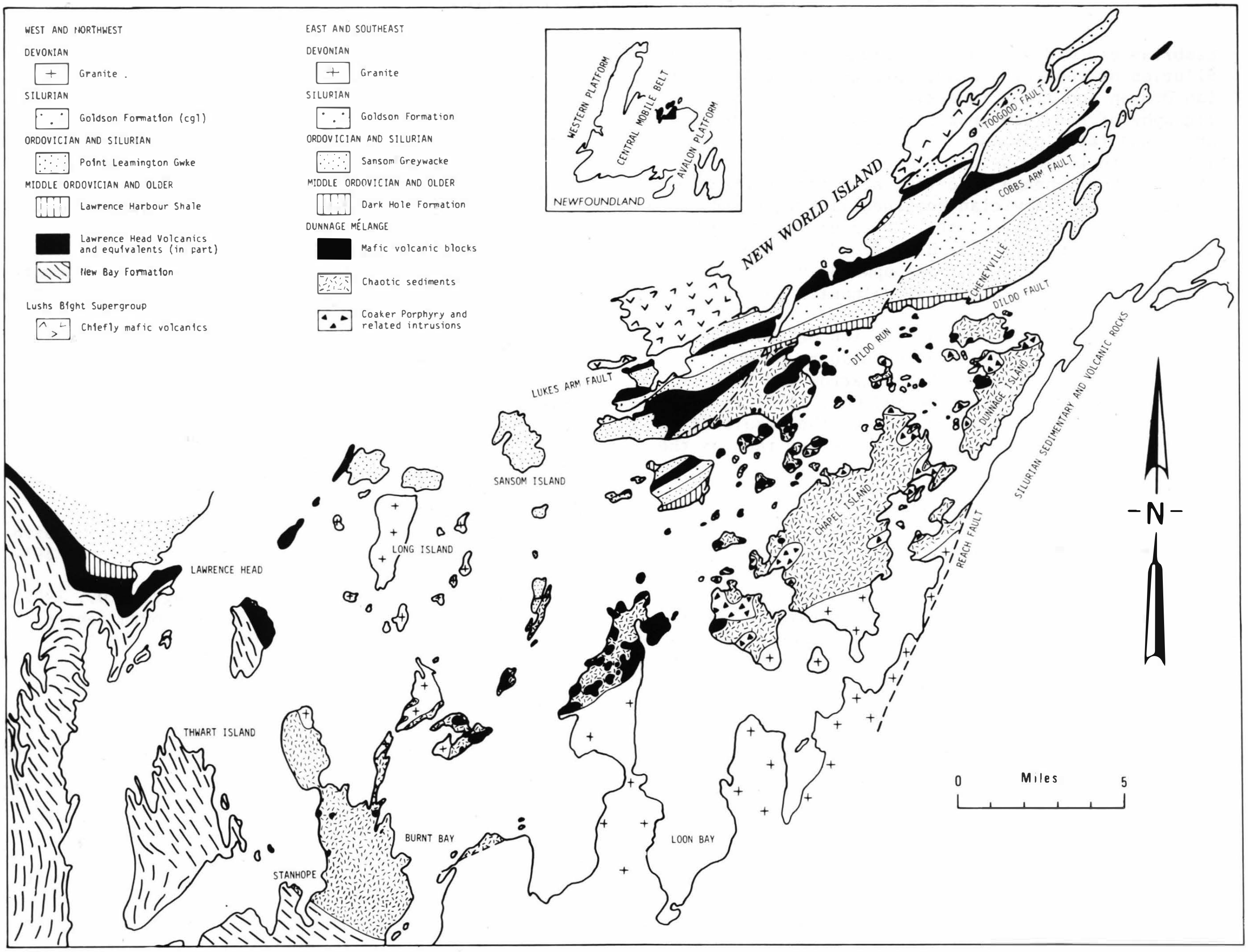


Figure 5: Internal characteristics and regional setting of the Dunnage Melange (after Williams and Hibbard, 1976).

River is composed of dunite and peridotite with an outer zone of gabbro, anorthosite and minor serpentinite. It is circular in shape and surrounded by hornfels of probable Silurian age. This body is most likely an intrusion (Kean, 1974).

Numerous circular, oval and elongate ultramafic bodies occur along the shores of Gander Lake and extend northward along the east side of Gander River. These range in size from a kilometre to less than a metre in diameter and extend over a distance of 60 km. They are mainly pyroxenites and gabbros with minor amounts of peridotite. Although described as intrusions, most contacts are tectonic.

## MAFIC-ULTRAMAFIC ROCKS OF THE SOUTHERN APPALACHIANS

In the Southern Appalachians (plate I), the greatest concentration of small mafic-ultramafic bodies lies to the west of the Brevard Zone. These occurrences are equated with similar occurrences farther north throughout the eastern parts of the Humber Zone. East of the Brevard Zone, small ultramafic bodies within the crystalline terrane of the Inner Piedmont are of unknown significance and have no northern counterparts, except perhaps for occurrences in metamorphosed easterly parts of the Dunnage Zone in Newfoundland. Farther east, in metamorphosed parts of the Carolina Slate Belt, small ultramafic bodies occur near the Georgia-South Carolina state line and in the Raleigh Belt of eastern North Carolina. All of these occurrences are of unknown significance and many of them may represent metamorphosed small intrusions rather than oceanic ophiolites.

## APPALACHIAN OPHIOLITES AND THE AGE OF IAPETUS

Stratigraphic analysis of the Humber Zone (Williams and Stevens, 1974) indicates rifting and the development of an ancient continental margin of Eastern North America in the late Precambrian with a stable margin existing well into the Ordovician. Destruction of this margin began in the Lower Ordovician and it was completed by Middle Ordovician. Ophiolite suites of the Appalachian Orogen are all of Late Cambrian to Early Ordovician age, where dated. Whether or not the preserved ophiolites represent the crust of a wide Iapetus Ocean, or a sampling of the floor of a related small ocean basin, remains debatable.

The Silurian and later development of the Appalachian Orogen presents a picture entirely different from that which prevailed during its

Cambrian and Ordovician development. Ophiolites of Silurian or younger age are unknown in the Appalachian Orogen and it is impossible to restore the stratigraphy of possible oceans and margins for the Silurian and later periods. An unconformity beneath Silurian rocks across the Humber Zone and westerly parts of the Dunnage Zone indicates the destruction of an earlier Ordovician continental margin and ocean basin. Elsewhere in central areas of the Dunnage Zone, where the stratigraphic record is complete, marine Ordovician shales pass upward into Silurian conglomerates and continental volcanics and red beds.

The common view that a Silurian or Devonian Iapetus Ocean closed in the Devonian to produce the Acadian orogeny (Dewey, 1969; McKerrow and Ziegler, 1971; Schenk, 1971; McKerrow and Cox, 1977) is based more on the premise that orogeny is the result of moving plates and closing oceans, rather than on the stratigraphic record.

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# PETROLOGY AND TECTONIC SIGNIFICANCE OF NEWFOUNDLAND OPHIOLITES, WITH EXAMPLES FROM THE BAY OF ISLANDS

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## ABSTRACT

Ophiolitic rocks of early Palaeozoic age occur at several localities in Newfoundland, Canada. These ophiolites have been interpreted as remnants of oceanic lithosphere which were preserved either as autochthonous or allochthonous terranes during orogenic episodes that led to the formation of the Appalachian Mountains.

The Bay of Islands is the best preserved of these ophiolite suites and can be compared with present day models of oceanic crust and upper mantle constructed from geophysical data and dredge hauls. Complete sections through the suite are present and indicate a genetic distinction between rocks formed in the mantle and those formed in the crust. The horizon separating these rocks is recognized on the basis of textures and mineralogy and chemistry, and is here called the 'petrological Moho' to distinguish it from the 'seismic Moho' which occurs above it. It is suggested that the formation of oceanic crust at spreading centres is attained by the intrusion of bodies of tholeiitic magma which are derived from partial melting of upwelling mantle material and which forms stratiform intrusive bodies and associated volcanics. The intrusions have a dunitic basal portion that grades downwards into tectonised residual harzburgitic mantle and upwards into layered gabbros. Pillow lavas and associated diabases represent partially differentiated portions of the magma, suggesting that modern abyssal tholeiites cannot be considered unfractionated primary magma.

Emplacement of the ophiolite suites took place in distinct stages involving thrusting and gravity sliding. During the early stages of emplacement a dynamothermal aureole was formed. Temperature conditions during the formation of this aureole suggests that not all the heat was derived by fractional heating but that the ophiolitic rocks themselves were hot during the early stages of obduction. The geothermal gradient realised is well above present day average oceanic geotherms, suggesting that either entirely different thermal conditions were present in the early Palaeozoic, or more reasonably that the ophiolites were obducted in a special tectonic environment. The model proposed is probably applicable to other ophiolite complexes.

## INTRODUCTION

Ophiolite suites composed of the association of ultramafic, gabbroic and basaltic rocks have been the source of controversy and discussion amongst geologists for several decades. Early workers (Benson, 1926; Steinmann, 1905, 1927 [etc.]), who first defined the ophiolite suite on the basis of their work in the Alps suggested that these rocks represented the intrusive and volcanic products of a eugeosynclinal environment that later became involved in the mountain building episode of the orogenic cycle. With recent theories of plate tectonics and seafloor spreading, ophiolites have assumed perhaps a more significant role in that they have been interpreted by a number of geologists as cross sections of ancient oceanic lithosphere (Dietz, 1963; Hess, 1964; Gass, 1968; Reinhardt, 1969). Williams and Smyth (1973) summarised the evidence to support such a view. Their most important points were:

1. Similarities exist in the gross physical characteristics of ophiolite suites with geophysical models of oceanic lithosphere.

2. The presence of 'on land ophiolite' rooted in oceanic lithosphere in Papua and New Guinea.

3. Lithologic similarities exist between ophiolite suites and rocks of the Macquarie Ridge where they are exposed on Macquarie Island. (This includes the presence of diabase dikes.)

4. Lithologic and chemical similarities exist between oceanic tholeiites and pillow lavas of the ophiolite suites.

5. Similar metamorphic mineral assemblages occur in oceanic rocks at mid-oceanic ridges when compared with ophiolites.

Ophiolitic rocks are thus thought of as pieces of oceanic crust and mantle trapped during the collision of continents and/or island arcs along consuming plate margins. This accounts for their association with the tectonically active parts of the earth, both past and present.

Ophiolites of early Palaeozoic age occur in the Appalachian mountains of the eastern United States and Canada, and are especially well preserved in Newfoundland (fig. 1). Of these Newfoundland occurrences, those of the Bay of Islands and Betts Cove are the most complete sequences (Williams, 1971; Williams and Malpas, 1972; Upadhyay et al., 1971). In the others only partial sequences remain (Norman and Strong, 1975; Malpas and Strong, 1975).

A number of models for the plate tectonic development of the Newfoundland Appalachians have been proposed (fig. 2) and each tries to indicate the origin of the ophiolitic rocks (Church and Stevens, 1971; Dewey and Bird, 1971; Kennedy, 1973). A recent model and the one that is preferred here is that of Strong et al. (1974) and is based upon structural, chemical and metallogenic considerations. This model depicts the closing of the early Palaeozoic proto Atlantic Ocean along an eastward dipping subduction zone that led to the collision of island arc and continent in Ordovician-Silurian times and the

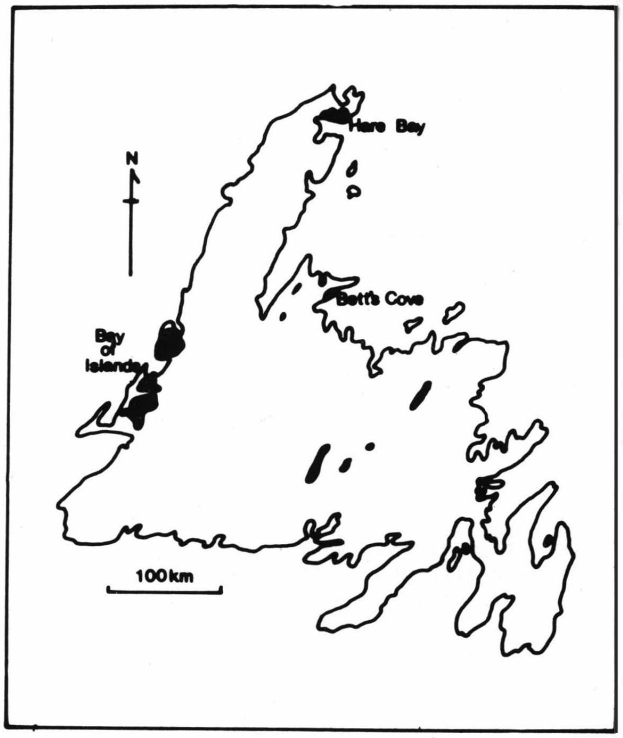


Fig. 1. Ophiolitic rocks in Newfoundland

simultaneous obduction of the Newfoundland ophiolites onto the western continental platform (fig. 3).

The writer has carried out regional studies on the Bay of Islands region over the last four years involving the remapping and reinterpretation of the ophiolites and a considerable amount of petrological and mineralogical analysis. Observation of the regional setting of these rocks and other rock types associated with them suggests that if correlation is to be made between the ophiolites and oceanic crust, then special consideration must be given to their mode of formation and their later emplacement under the following headings:

1. Regional Geology
2. Stratigraphy and chemistry of the ophiolitic rocks
3. Origin of the metamorphic rocks associated with the ophiolites
4. Origin of the clastic sediments associated with the ophiolites

## Regional Geology

In Western Newfoundland a Precambrian crystalline basement is unconformably overlain by several transported rock assemblages, all essentially coeval and comprising a number of separate and distinct subhorizontal slices emplaced during the Middle Ordovician. The Bay of Islands area itself (fig. 4) is underlain by a succession of transported slices that can be broadly subdivided into two main groups: a group of structurally lower slices that are composed

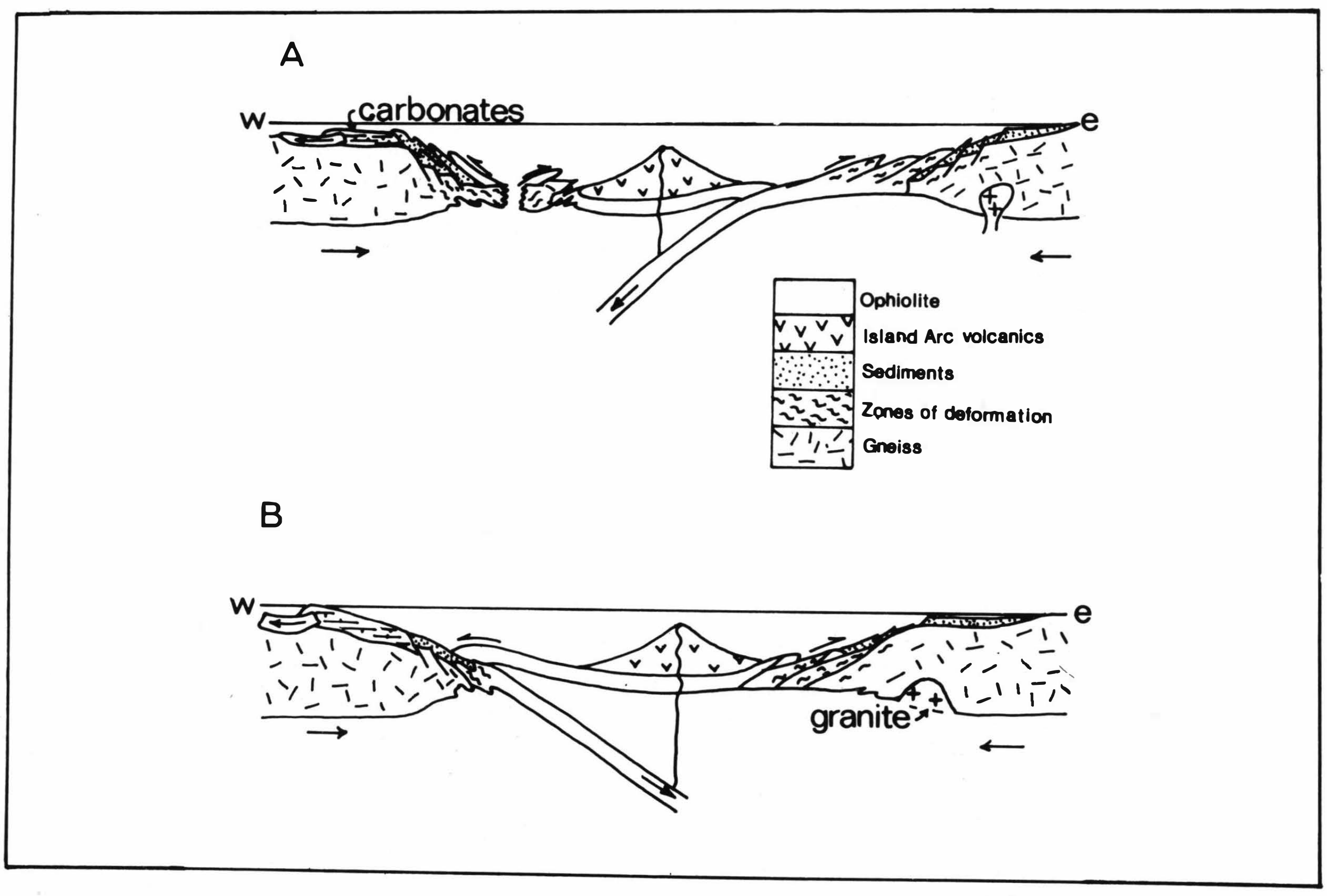


Fig. 2. Plate tectonic models applicable to Newfoundland. A. after Dewey and Bird (1971) and Kennedy (1973)
B. after Church and Stevens (1971)

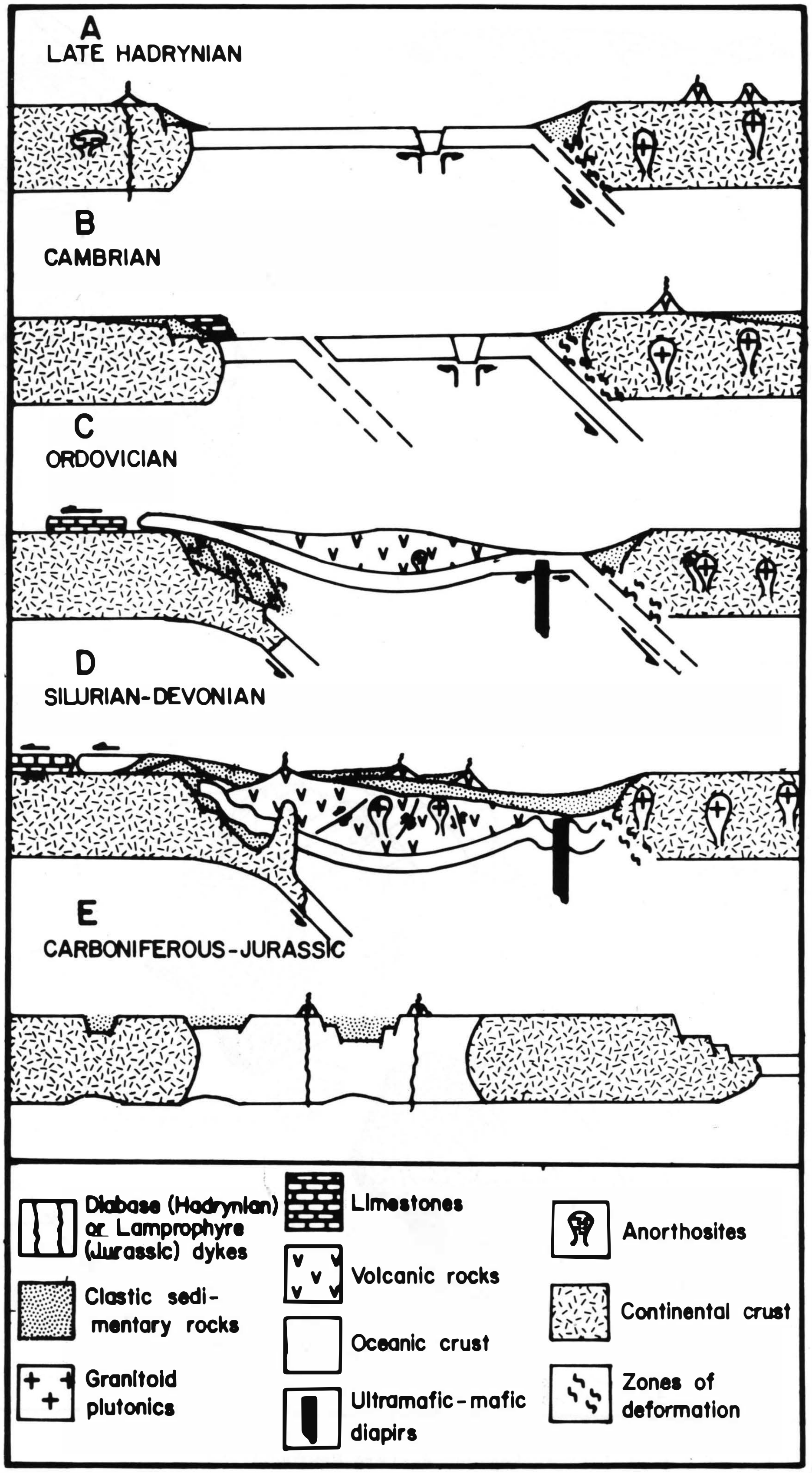


Fig. 3. Presently accepted model after Strong <u>et al</u>., (1974)

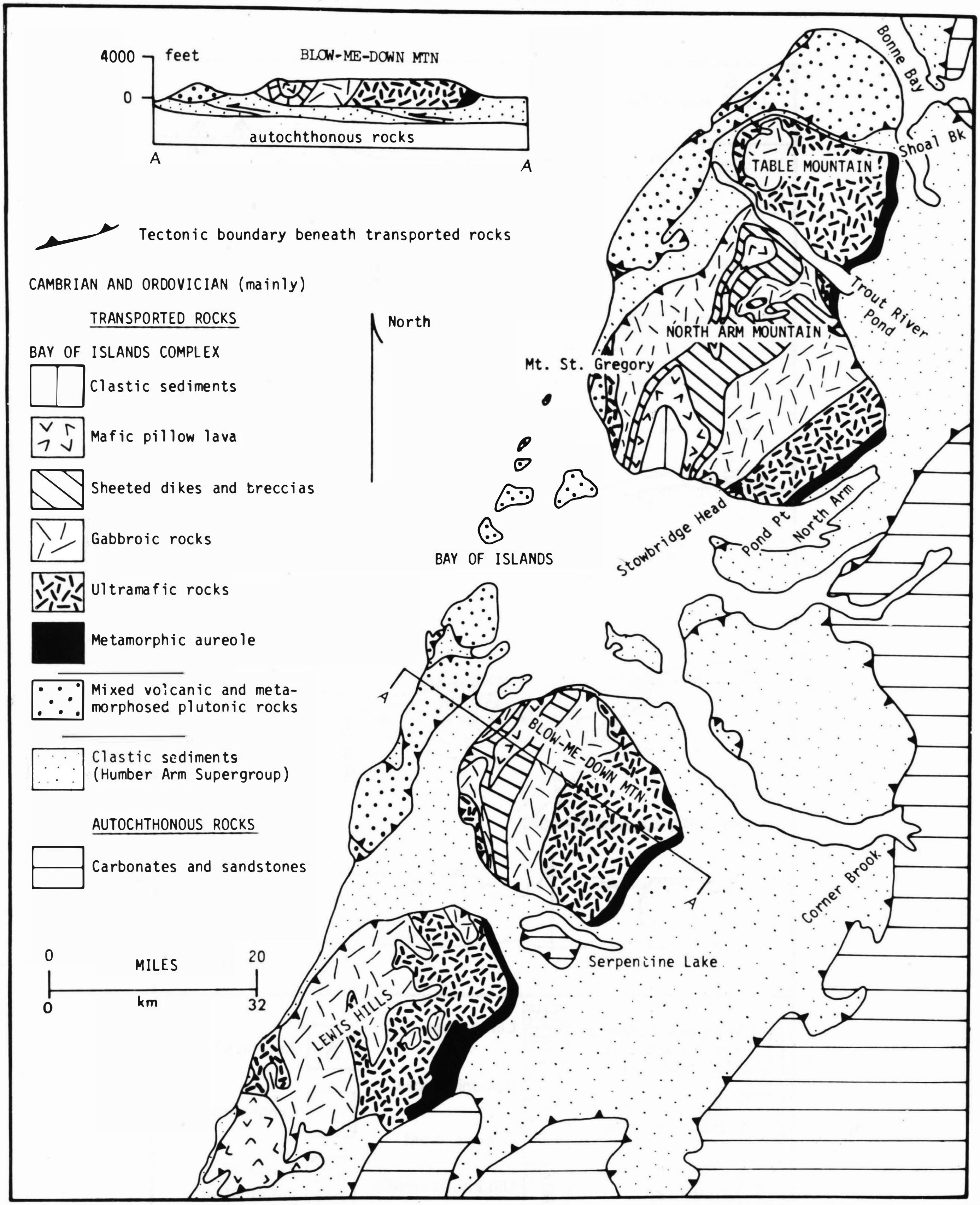


Fig. 4. Geological map of Bay of Islands Region, Western Newfoundland.

of Cambrian and lower Ordovician clastic sedimentary rocks and a group of higher slices that are for the most part composed of igneous rocks. These higher slices can be subdivided into at least four distinct rock groups, each group being represented in one or more superposed slices. In most places, all of these structurally higher slices directly overlie transported clastic sedimentary rocks, but locally they overlap and partially overlie one another. When the latter is the case, the rock groups maintain a consistent order of structural stacking.

One of the most extensive rock groups amongst the higher slices consist of foliated gabbro, amphibolite and foliated granite with complex internal relationships that are cut by poorly developed sheeted dykes that are themselves inseparable from undeformed pillowed lavas that also form part of the same slice. The foliated gabbros, amphibolites and granitic rocks have no direct correlatives in the main ophiolite mass which lies in the slice above, and are referred to as the Little Port Complex (Williams and Malpas, 1972). The main ophiolite slice has been called the Bay of Islands Complex. However, the rocks of the Little Port Complex and the Bay of Islands Complex all resemble rocks associated with the Troodos Massif of Cyprus in almost every important feature including the complicated tectonic history and the same primary ophiolite stratigraphy. They also include similar massive sulphide deposits (Duke and Hutchinson, 1973). It has also been observed that in the ophiolites of the Bay of Islands Complex there is a downward change of characteristic minerals from celadonite through chlorite in the pillow lavas to actinolite in the sheeted dikes and green hornblende in the gabbro (Williams and Malpas, 1972) as described by Gass and Smewing (1973) for the Troodos Complex. The third slice in the higher igneous series, the Skinner Cove Slice, consists of alkaline volcanics and associated dikes that have been interpreted by Strong (1974) as off-ridge volcanic rocks associated with the ophiolites and directly comparable to the Upper Pillow Lava series of the Troodos Complex.

Thus, according to the Lower Palaeozoic and Precambrian stratigraphy the evolution of western Newfoundland is interpreted as relating to the development of an early Palaeozoic continental margin (Williams and Stevens, 1974).

## Stratigraphy and Chemistry of Ophiolite Rocks

The ophiolite suite of the Bay of Islands Complex (fig. 5) is represented in four separate massifs, which from north to south are, Table Mountain, North Arm Mountain, Blow Me Down Mountain and Lewis Hills. All lie in the same structural position and represent either separate transported bodies or the erosional remnants of a once continuous slice. Two of the massifs (Blow Me Down and North Arm Mountain) display a completely developed ophiolite suite, but all four include a basal ultramafic unit and attached dynamothermal amphibolitic aureole. The sequences of ophiolite units in the three northernmost massifs are disposed in synclines with northeast trending, subhorizontal axes and moderately to steeply dipping limbs. The tectonic base of each massif is subhorizontal since the thrust faults delineating the base of the slices truncate the ophiolites at depth.

### Lherzolite:

Lherzolites, consisting of the mineral assemblage olivine, clinopyroxene, orthopyroxene, ceylonite and minor corundum, with metamorphic hornblende and Ti-phlogopite, form the basal member of the ultramafic sequence and are exposed immediately above the amphibolitic aureole of Table Mountain. Tectonic fabrics in the form of stretched and aligned, altered orthopyroxene phenocrysts and triple point junctions between ground mass olivines, are exhibited throughout, although the latter are generally masked by serpentinisation. Veins of bastite-enstatite both display the fabrics and cut the fabrics, and they are thought to represent the products of filter pressing mechanisms during tectonism and are part of the original magmatic activity of the ultramafics. Thus there is an indication of continuous deformation and igneous activity, processes that would result during the upwelling of mantle material below a spreading centre. The fabrics are consequently interpreted as mantle tectonite fabrics.

The equilibrium coexistence of clinopyroxene, orthopyroxene and olivine with an aluminous phase, in this case spinel, can be used to define the pressure and temperature conditions under which these minerals crystallised (O'Hara, 1967). The pressures and temperatures indicated are of the order of 20 kb and 1200$^{o}$C, suggesting a depth of formation of approximately 60 km or possibly less if tectonic overpressures were active (Malpas, 1977). The spinel lherzolite may therefore represent partially depleted upper mantle, in agreement with models of upper mantle composition suggested by O'Hara (1968) and Ringwood (1969).

### Harzburgite:

Harzburgitic rocks, with minor amounts of dunite and enstatolite, make up the major part of the ultramafic rocks, to a maximum of 4 km on Table Mountain. The mineral assemblage of these rocks is olivine, orthopyroxene and chrome spinel. Very little clinopyroxene has been recognized in this zone, the lower contact with the lherzolites being marked by the virtual disappearance of this phase. Mantle tectonite fabrics are present throughout the harzburgites, together with pyroxenite (clinopyroxene and orthopyroxene) veins which are more common toward the base of the zone. Dunite veins and layers increase in abundance towards the top of the harzburgites, where they exhibit branching and cross-cutting relationships. The dunite veins are commonly associated with traces of chromite.

### Dunite:

Although dunites are present to some extent throughout the ultramafic pile, a zone of almost pure dunite reaching an approximate thickness of 350 m is present above the harzburgite zone. However, an exact thickness of this zone cannot be defined since both upper and lower contacts are gradational. The lower contact is in some places marked by lenses of clinopyroxenite and podiform chromite, both of which are minor minerals in the dunite. Plagioclase increases in abundance as an intercumulus phase higher in the zone. Cumulate textures are discernible throughout the dunite zone, in contrast with the mantle tectonites in the rocks below. Although variably serpentinised often to as much as 65%, the olivines still exhibit well developed cumulate growth with both cumulus and intercumulus chromite. In hand specimen both the chromite and the plagioclase show distinct grading of mineral proportions where they occur as cumulus phases. The

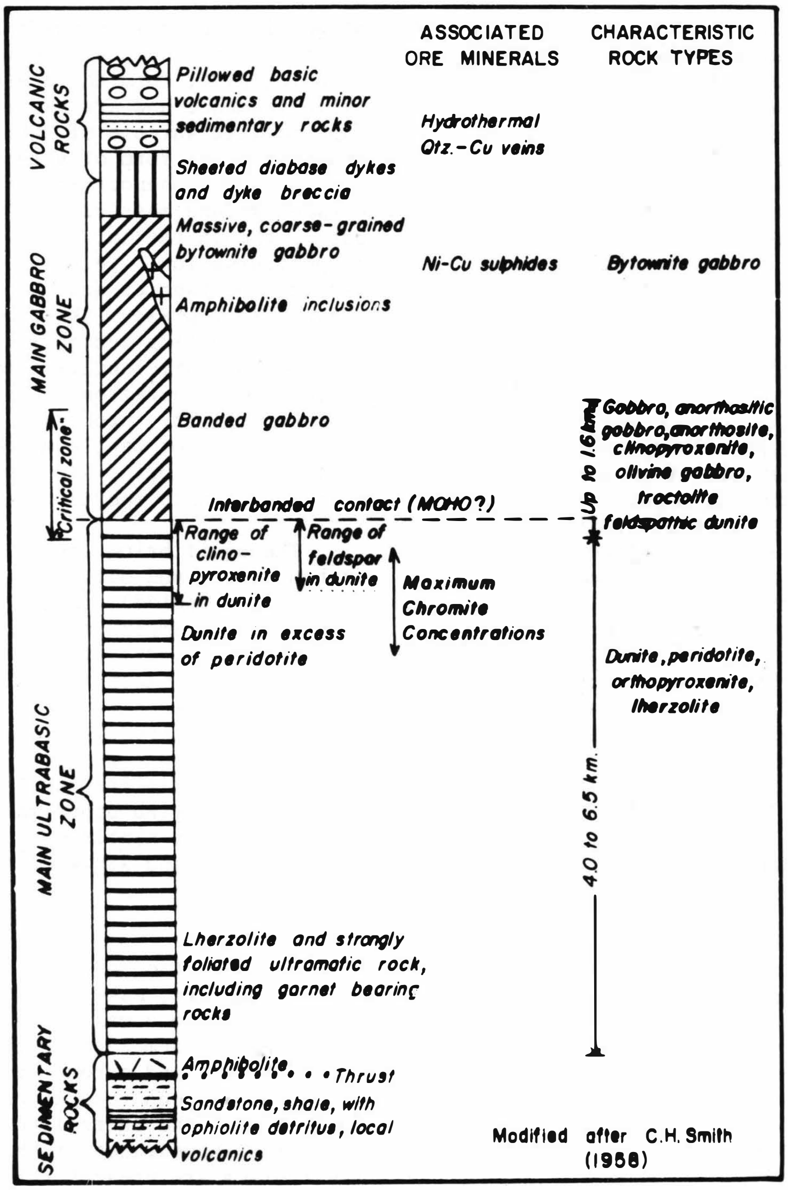


Fig. 5. Diagrammatic vertical section of Bay of Islands Complex ophiolite suite.

transition between the dunites with cumulus textures and harzburgites with tectonic fabrics occur over a thickness of several metres. These textural features suggest a genetic distinction between these rock types, which is supported by further evidence cited below.

Critical Zone (Transitional Rocks):

Dunite becomes more feldspathic higher in the zone and is interbanded with norite, troctolite, and anorthosite over a thickness of 100-200 metres which Smith (1958) termed the Critical Zone. This zone forms a transition between the ultrabasic and overlying basic rocks. Layers range from centimetres to several metres thick and many also show distinct settling features. In some places there has been some post-accumulation deformation, but its restricted nature and the fact that the affected rocks are surrounded by undeformed rocks suggest that it may have taken place as a result of slumping of a crystal mush. The strike of the layering is quite variable and is in places interrupted by such slump folds. The dip of the layers may be much steeper than in adjacent ultramafic rocks and reversals of dip are not uncommon. The layers are lenses of individual rock types with limited extent rather than laterally continuous units. Mineral alignment can be attributed to primary igneous flow lineation and is comparable to similar features noticed by the writer in the tectonically undeformed allivalites of Rhum, Inner Hebrides. The environment of formation seems to have been one of rapid deposition, possibly by similar turbidity currents to those envisaged by Wager (1939) for the Skaergaard Intrusion.

Smith (1958) has classified the rocks of the critical zone on the basis of relative proportions of plagioclase (bytownite An 75), clinopyroxene (diopsidic augite) and olivine (Fo 85). Chemical variations within the minerals of the zone seem relatively limited and no cryptic layering has yet been clearly defined. Where feldspar is more abundant, harrisites and allivalites are found. Hydrogarnet has replaced plagioclase in most cases, its composition lying towards the tri-calcium hexahydrate and member in composition

Veinlets of coarse clinopyroxenite, both concordant with and cross-cutting the layering are found in some areas, generally toward the base of the critical zone. The clinopyroxene is almost pure diopside, $Ca_{47}Mg_{49}Fe_4$, in composition and is associated with a small amount of serpentinised olivine.

Gabbro:

Rocks of gabbroic composition are equal in abundance to the ultramafic rocks in the Bay of Islands Complex. The gabbros are layered throughout and sedimentary features such as graded bedding and cross bedding have been noted. In some places the direction of grading reverses rapidly which suggests that in these cases it results from "flow differentiation". In other places a pseudo-cross bedding arises by essentially solid state shear dislocations. Anorthositic gabbro is the most common and it grades into metagabbro by an increase in the clinopyroxene content and into anorthosites by an increase in the plagioclase. Hornblende gabbro, commonly showing a lineation of the amphiboles, also occurs in areas of North Arm Mountain and Table Mountain towards the top of the gabbro pile (Malpas et al., 1973). Olivine rarely occurs in the gabbro, but where it is present it appears to be a relatively early formed mineral, being enclosed in plagioclase. Generally it is serpentinised but fresh crystals are fairly iron-rich (Fo 77).

Actinolite and green hornblende are developed also as secondary minerals, presumably as a result of burial metamorphism, and prehnite also occurs as late cross-cutting veinlets.

Diorite:

Soda-rich dioritic rocks or quartz-diorites approaching trondhjemites in composition cut coarse-grained gabbroic rocks and occur as part of the diabase dike sequence on both North Arm Mountain and on Blow Me Down Mountain. These small bodies are associated with the higher gabbros and have also been found as veins and dikes near the Mount St. Gregory Copper deposits. No intrusions of diorite have been found in the ultramafic rocks. The diorites are generally pink to grey in colour and medium to fine grained. Plagioclase which is often altered, is generally albite or oligoclase, and forms zoned euhedral phenocrysts. There may be as much as 50% quartz in the rocks and accessory minerals include epidote, apatite, zircon and rare pyrite. These quartz diorites are interpreted as late stage differentiates of tholeiitic magma.

Diabase:

Sheeted diabase dikes are most extensive on Blow Me Down Mountain and they have also been recognized on North Arm Mountain. The dikes and associated brecciated dike rocks from both areas were previously interpreted as metavolcanic rocks that formed the roof of an intrusion and separated gabbros below from pillow lavas above (Smith, 1958).

The sheeted dikes are dark grey to greenish, medium grained massive rocks, in most cases equigranular. Some are plagioclase-phyric, especially in the centres of wide sheets. Chilled contacts are found in many exposures. Massive dike rocks from the complex generally consist of plagioclase and actinolite in equal proportions. In some cases the actinolite surrounds colourless remnant cores of clinopyroxene.

Breccias, consisting of close-packed angular dike rock fragments of varying sizes in a finely comminuted matrix of similar material, are at least as extensive as sheeted dikes within the complex. They represent fragmented dikes no longer exhibiting a planar disposition. Some fragments, although by no means the majority, are rounded and indicate some degree of abrasion. Movement between fragments cannot, however, have been too great since the distribution of coarse and finer grained fragments within most breccias preserves the original position of pre-existing chilled margins. The brecciation is restricted almost entirely to the diabase and does not affect the overlying pillow lavas or the underlying gabbro to any extent (Williams and Malpas, 1972). Brecciated dikes have been found cutting fresh, massive gabbro. These dikes had chilled against the gabbro before being brecciated, as shown by the distribution of fragments. The mechanism by which such cold brecciation was produced, without affecting the host rocks, is not yet fully understood, but was probably a result of fluid, possibly sea-water, migration (Williams and Malpas, 1972).

All dikes are metamorphosed to low amphibolite or greenschist facies, a result of burial rather than regional metamorphism associated with deformation. Actinolite, chlorite, epidote and zoisite are the major metamorphic minerals. Prehnite and pumpellyite have also been noted.

Pillow Lavas:

Pillow lavas are exposed on Blow Me Down Mountain and North Arm Mountain. The lavas are fed by the underlying dikes, are interbedded with siltstones higher up, and are sulphide bearing at the bottom of the pile. Thus they are comparable to the Troodos pillow lavas of Cyprus (Searle, 1972). Approximately 300 metres of pillow lavas are exposed and complete the igneous section of the ophiolite.

Sediments:

Sediments reaching a maximum thickness of 200 metres overlie the pillow lavas in a number of places. They are generally red-brown sandstones, red shales, and siltstones, although rare pebble conglomerates have been recorded.

The sediments are generally well bedded and sorted and tops are determinable from minor ripple drift lamination and load structures. The presence of volcanic fragments and the lack of potash feldspar makes these sediments lithologically distinct from sandstones underlying the thrust slices. The sediments do not resemble closely any lithologies found in the present day deep ocean basins or in troughs associated with oceanic ridges. Likewise, although manganese rich sediments are known to be associated with other Newfoundland ophiolites (Upadhyay, 1973) and are recorded from the ocean floor (Scott et al., 1974), no comparable lithologies have been found associated with the Bay of Islands ophiolites. The presence of detrital quartz and the feldspars and the coarseness of some sediment horizons suggest a relatively proximal source of possibly granodioritic nature.

Geochemistry:

Chemical analyses for both major and trace elements have been made using atomoc absorption spectrophotometry and X-ray fluorescence spectrophotometry. All rock types are represented except for the sediments.

The analytical methods, precision and accuracy are as described by Strong (1973). The results are presented here only in diagrammatic form but can be obtained from the author upon request. Before the calculation of norms and plotting of diagrams, totals were recalculated to 100% volatile free and adjusted for oxidation (calculate Fe''' with Fe'' following the procedure outlined by Irvine and Baragar, 1971).

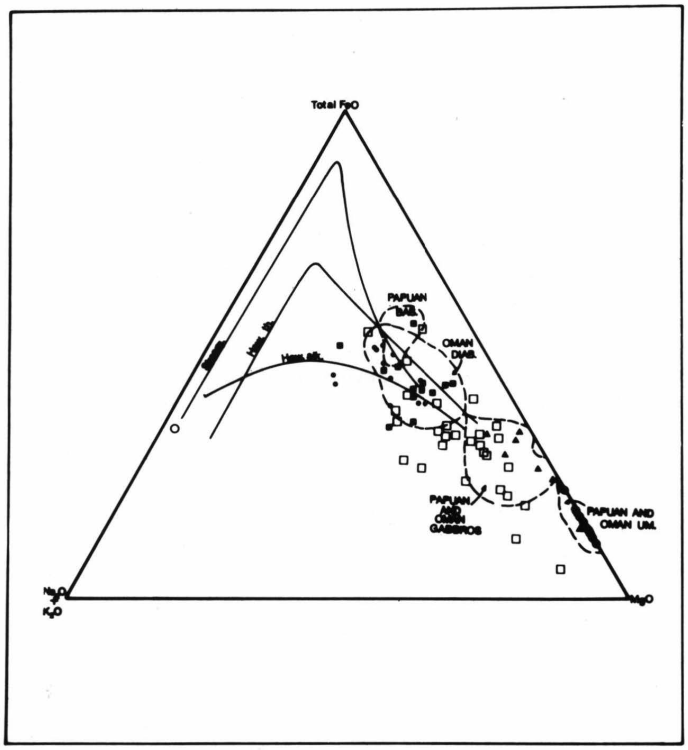


Fig. 6. F.M.A. plot for ● Dunites □ Gabbros △ Pyroxenites ■ Dykes ● Lavas ○ Diorites of Bay of Islands. Other recognised trends (Skaergaard, Hawaiian tholeiitic and alkaline) and ophiolitic rocks are shown.

On the FMA diagram (fig. 6) the Bay of Islands ultramafic-mafic fractionation trend is comparable to that for the Papuan and Oman ophiolites, and is similar to the general ophiolite trend given by Thayer (1967). However, the ultramafic-mafic trend is a little more complicated than it appears. There is a tendency for the gabbros lower in the sequence to plot across the diagram, i.e. away from the FeO apex. This trend could be produced by increased $Na_2O$ and $K_2O$ but these rocks do not show any appreciable enrichment in alkalies. Therefore, the differentiation trend is produced by fractionation so that $Fe_2O_3$ decreases with MgO in the ultramafic and lower gabbro, but sharply increases in the higher-level gabbro, dikes and pillow lavas. Fractionation of olivine was followed by crystallization of clinopyroxene. Plagioclase crystallization enriched the later liquids in iron as evidenced by the sudden rise in $Fe_2O_3$. The result of this iron enrichment was crystallization of magnetite and titaniferous magnetite in the later stages. Norman and Strong (1975) have pointed out similar features in other Newfoundland ophiolite suites, and have noted the close resemblance between the mafic rocks of these suites and oceanic tholeiites.

Figure 7 shows $Cr_2O_3$ and NiO values plotted for ultramafic alpine peridotites and ultramafic portions of tholeiitic stratiform intrusions represented by the Muskox intrusion, Northwest Territories, Canada. The line effectively separates olivine peridotites from ultramafic and mafic rocks of stratiform intrusions. Similar plots for the Bay of Islands (additional data from Irvine and Findlay, 1972), show that a distinction can also be made between rocks above and below the harzburgite-dunite boundary, the latter correlating with rocks of stratiform complexes and the former with alpine peridotites.

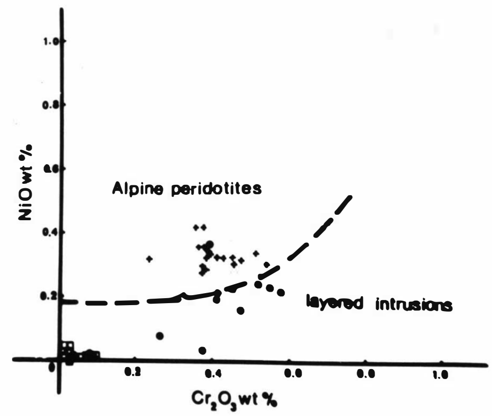


Fig. 7. $Cr_2O_3$ vs NiO for □ gabbros ● cumulate ultramafics + tectonite ultramafics.

| Rock Type | | Fo content of olivines |
|---|---|---|
| Olivine Gabbro | | 75.1 |
| Dunite | | 81.1 |
| Dunite | CUMULATES | 87.5 |
| Dunite | | 89.1 |
| Harzburgite | | 91.6 |
| Harzburgite | TECTONITES | 91.4 |
| Harzburgite | | 93.0 |
| Lherzolite | | 89.7 |
| Lherzolite | | 89.8 |

Table 1. Fo contents of olivines from Bay of Islands Complex

Unaltered cores of olivine crystals have been analyzed from most parts of the sections by electron microprobe, Table 1. Sharp decreases in the forsterite content are observed at the harzburgite-dunite boundary on both Table Mountain and North Arm Mountain. The olivines in the lherzolites and harzburgites have compositions of the order of $Fo_{92}$, while a lower average composition of $Fo_{88}$ occurs in the dunite and higher rocks.

Chromites also differ across this boundary. A red chrome-spinel is the major chrome bearing phase in the harzburgites, but opaque black chromite appears in the dunites. Major concentrations of this latter type occur in podiform bodies and as a major cumulate phase at the base of the dunite zone. Chromite unit cell dimensions may also show significant differences due to the amount of $Fe^{+3}$ solid solution (Malpas and Strong, 1975).

Clinopyroxenes become a major phase in the dunite zone, so much so as to often form distinct bodies of clinopyroxene. The introduction of this phase also makes the boundary between the harzburgite and dunite.

The significance of the contact between the harzburgite and overlying dunite is that it must represent the junction between rocks of mantle origin and rocks formed by crystal settling under crustal conditions in relatively high level stratiform intrusions (Irvine and Findlay, 1972, have suggested such a model for the Bay of Islands but did not include the dunites as part of the crustal rock). The horizon can therefore be called the 'petrological moho' or the genetic distinction between crust and mantle (Malpas, 1973). Such a distinction has been anticipated by Greenbaum (1972) for the Troodos complex. The Critical Zone, some 125-500 metres higher in the succession, represents the moho as defined seismically since it is here that major density changes take place. There is some indication that the petrological moho may also be defined seismically since weak discontinuities have been noted below the present oceanic moho to a depth of about 600 metres (M. J. Keen, R. Moberly, pers. comm., 1972).

## Metamorphic Rocks Associated with the Ophiolite Suites

Metamorphic aureoles that underlie alpine peridotites or ophiolite suites are common in many parts of the world (Challis, 1965; Dickey, 1970; Green, 1964; Karamata, 1968; Loomis, 1972; MacGregor, 1964; MacKenzie, 1960; Smith, 1958). The absence of aureole rocks in many other examples can be considered as a result of structural omission rather than a primary absence, for most ophiolite suites and alpine peridotites are fault bounded.

Metamorphic rocks occur at the base of all of the ophiolite masses forming the Bay of Islands complex. These metamorphic rocks are considered part of the Bay of Islands slice assemblage and are commonly referred to as the "basal aureole" (Malpas et al., 1973; Williams and Smyth, 1973). Aureole rocks occupy a constant level at the stratigraphic base of the exposed ophiolite sequence and are not related to the present structural base of the slice which is marked by serpentinite melanges. Where the aureole is best developed it has an overall structural thickness of approximately 130 metres and grades downwards from a foliated pyroxene amphibolite at the contact through garnetiferous amphibolite, greenschist, garnetiferous phyllite into an argillite. Rodingitic rocks, developed during later calcium metasomatism are found in places.

Recent discussions by MacTaggart (1972) and Moores and Raymond (1972) contrast earlier and more recent views on the origin of alpine peridotites which need reconciliation. On the one hand, the peridotites are interpreted as orthodox crustal intrusions as evidenced by the surrounding aureoles (MacTaggart, 1972), and on the other they are interpreted as mantle and dismembered parts of ophiolite suites, based for the most part on the analogy between the ophiolitic suite and oceanic crust and mantle (Moores and Raymond, 1972). The first interpretation negates the second, and the second makes no attempt to reconsider the field evidence upon which the first is based. This failure of recent plate tectonic models to account for the metamorphic aureole rocks has the effect of reducing the otherwise clear analogy between ophiolites and oceanic crust and mantle. If alpine peridotites are mantle derived and are in most cases simply the ultramafic fraction of ophiolite suites (now dismembered), then their surrounding metamorphic rocks cannot represent conventional thermal aureoles. Rather they must have a significance that relates to obduction or later tectonism and metamorphism. Such a hypothesis, relating the genesis of the Bay of Islands complex aureole rocks to emplacement of the thrust slices, has been proposed by Malpas et al. (1973) and Williams and Smyth (1973). The constant setting of the aureole, its narrow uniform width, its constant lithologies and its structural and metamorphic characteristics can be explained by a plate tectonic model that envisages the aureole as a result of obduction of oceanic crust and mantle onto a continental margin. According to this model, supracrustal rocks are overridden by a subhorizontal sheet of forcefully expelled oceanic crust and mantle that is everywhere detached at approximately the same level (3-6 km within the mantle or the thickness of the ophiolites' ultramafic unit). The aureole thus evolves as a contact dynamothermal aureole and acquires its structural style and metamorphic mineralogy during early stages of expulsion. Having established such a model, several points arise. The first is that the direction of strike of the ridge or spreading centre at which the ophiolites were presumably originally produced can be gleaned from the attitude of the sheeted dikes in the transported complexes which were presumably parallel to the ridge when emplaced (Williams and Malpas, 1972). Furthermore the direction of earliest transport (presumably at an angle to the continental margin), with respect to the direction of strike of the ridge, can be ascertained by comparing the attitude of the sheeted dikes with the direction of transport indicated by the vergence and facing directions of early recumbent structures in the aureole rocks. It appears that these directions are at a high angle to one another, suggesting that the ridge possibly intersected the continental margin.

The second problem is that the mechanics of displacing oceanic crust and mantle onto a continental margin are either completely unknown, or at the most, poorly understood (Coleman, 1971; Moores and MacGregor, 1972; Malpas and Stevens, in prep.). The heat for metamorphism of the aureole protoliths could be derived from two major sources. Firstly, heat in the mantle rocks, and secondly, heat derived during obduction. Calculations of partition coefficients of $Fe^{2+}$ and $Mg^{2+}$ between coexisting garnets and pyroxenes in the pyroxene amphibolite suggests that temperatures of 800° - 850°C were attained at the thrust contact. Calculations of the amount of heat that could be produced by friction during thrusting suggests that maximum temperature increases of 200°C might be expected in ideal conditions. Thus, the remaining heat must have been derived from the ultramafic rocks, suggesting that they were at a temperature in the order of 1100°C - 1200°C. Such temperatures could only occur at depths of 3 km in the mantle in two oceanic environments today. The first is on or close to a spreading centre, and the second and less likely is in a marginal basin where high heat flow exists as a result of igneous activity above subduction zones. Naturally, the heat source, the factors controlling heat distribution, and the attitude of geoisotherms are all recurring problems in most analyses of regional metamorphic terranes, and whether or not such a model is valid rests largely upon these poorly understood conditions. However, it can be generally concluded that the ophiolites were obducted not too long after their formation whilst they still retained a high heat content. They are unlikely to have been obducted on the edge of a large ocean basin at a distance from the spreading centre.

Sediments Associated with the Ophiolite Suite

Sediments penetrated by drill holes during the JOIDES deep-sea drilling project are typically calcareous ooze, partially chertified turbidites or deep sea muds. The radiolarian ribbon cherts commonly overlying basalts of ophiolite sequences are not found. In a few localities this overlying radiolarian chert may be thin or represented by chloritic-shales and may often be overlain by either limestones or turbiditic sandstones. Where fossil dating is possible the ages of the cherts and limestones associated with the uppermost basalts are the same or only slightly greater than that of the oldest flysch deposits of the orogen in which the ophiolite has been incorporated. Furthermore, in Newfoundland and elsewhere, the ophiolites are emplaced directly onto this flysch. This suggests a single episode of generation and initial emplacement of ophiolites, which would only be predicted by the plate tectonics model when the site of formation was extremely close to the site of obduction, as is the case for instance where a ridge system strikes directly into a continental margin to form a triple point junction, or in a marginal basin setting.

Conclusions

From this survey of western Newfoundland ophiolite stratigraphy and petrology certain conclusions may be drawn:

1. The ophiolite suites represent cross-sections of oceanic crust and upper mantle.

2. The genetic distinction between those rocks formed in the mantle and those formed under crustal conditions is marked by textural and chemical changes and can be called the "petrological moho".

3. Differentiation trends suggest that the basaltic proportions of the ophiolite are a result of fractionation and are not primary magmas. This may well hold true for oceanic tholeiites. Their constancy of composition is due to their shallow, low pressure eutectic crystallization.

4. The ophiolites were hot when emplaced, and not very old. The attitude of the spreading centre with regard to direction of obduction, the temperature required for production of the metamorphic aureole, and the presence of orogenic flysch deposits of the same age as the ophiolites all evidence this. The only feasible plate tectonic models that can explain these phenomena suggest that the ophiolites were produced in a marginal basin and almost immediately obducted onto the continental margin or, less likely, produced at a spreading centre close to a ridge/continental margin triple point junction and obducted almost immediately.

Such a model explains why:

(1) ophiolites are formed only once, early in orogenic development.
(2) they are about the same age as the oldest flysch in the orogen.
(3) contact metamorphic aureoles exist at the bases of the ophiolite slices.
(4) associated sedimentary rocks may not be of the 'typical' deep ocean basin type.

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# OPHIOLITES FROM THE NORTHERN APPALACHIANS OF QUEBEC

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## ABSTRACT

In southern Quebec the ophiolites mark a major suture connected to the Early Paleozoic southeastern boundary of the Canadian shield. They are regarded as fragments of the lapetus (Proto-Atlantic) oceanic lithosphere emplaced by thrusting upon the pre-Ordovician Notre Dame schists in Early Ordovician time.

Ultramafic and gabbroic cumulates overlying harzburgite tectonite are similar to the cumulates of the Mesozoic Vourinos ophiolite (Greece). Cumulate, shallow intrusive and extrusive rocks of the Quebec ophiolites derive from a low-Ti and low-K tholeiitic magma of the type common to the mid-ocean ridges. Lavas and immature volcanic sediments of assumed Early Ordovician age and volcanic island arc origin overlap stratigraphically the ophiolitic suite.

## INTRODUCTION

The Cambro-Ordovician rocks of the Quebec Appalachians form three northeasterly trending zones parallel to the Early Paleozoic margin of the Canadian shield. They are believed to represent the sedimentary and volcanic record of the shelf, continental slope and rise deposits of an Atlantic-type continental margin (Bird and Dewey, 1970; Williams and Stevens, 1974; St. Julien and Hubert, 1975). From the northwest to the southeast the three zones are, respectively, the St. Lawrence autochthonous domain, the Appalachian allochthon and the internal Notre Dame zone (fig. 1).

The Precambrian crystalline basement of the Grenville province (average isotopic dates of 950 my) and its sedimentary cover constitute the foreland of the Appalachian orogen. The sedimentary rocks overlapping the Precambrian are mainly a Cambro-Ordovician sandstone-limestone assemblage and a late Middle Ordovician flysch. They have been deposited in shallow marine waters on the St. Lawrence platform (Clark, 1972).

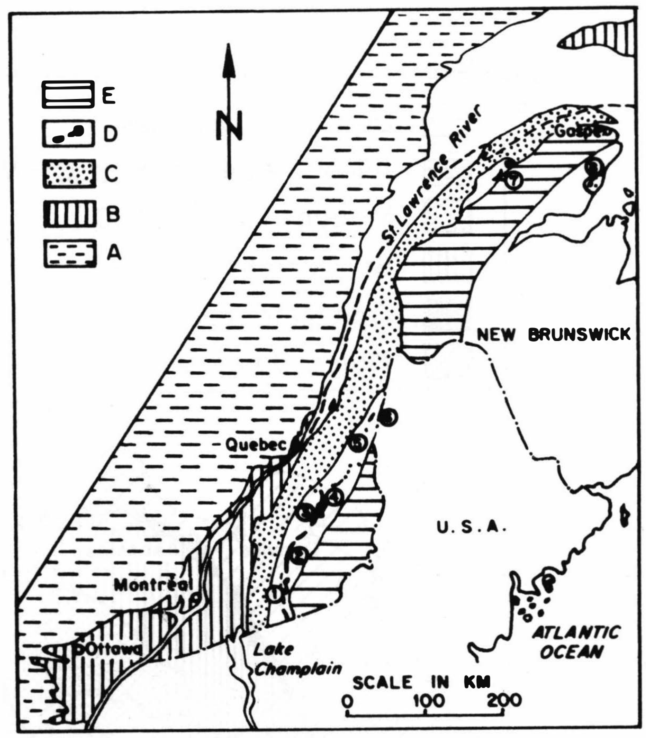


Figure 1. Map of the Appalachians of southeastern Quebec showing the main structural zones and the major ophiolite localities: (A) Precambrian crystalline basement. (B) sedimentary cover of the St. Lawrence platform, (C) external domain or Appalachian allochthon, (D) internal domain including Notre Dame schists and ophiolitic belt, (E) Siluro-Devonian rocks of the Gaspé-Connecticut Valley synclinorium; (1) Orford, (2) Asbestos, (3) Thetford Mines, (4) East Broughton, (5) St. Fabien, (6) St. Omer, (7) Mount Albert, (8) Port Daniel River.

The external domain, or Appalachian allochthon, consists of a Cambro-Ordovician sandstone - shale - limestone conglomerate assemblage and a calcareous early Middle Ordovician wildflysch (Osborne, 1956; Riva, 1972; St. Julien, 1972). These marine rocks have presumably been deposited on the continental slope of the ancient continental margin. They started sliding northwestwards in early Middle Ordovician time forming imbricated thrust slices and nappes which progressively moved towards the continent until they came to rest on the St. Lawrence platform by the beginning of the Upper Ordovician (Riva, 1972; St. Julien and Hubert, 1975).

The Appalachian allochthon grades southeastwards into the metamorphic and complexly deformed internal zone of Notre Dame (schist belt), which consists of metaclastic and metavolcanic rocks of greenschist facies that have accumulated in Late Precambrian and Cambrian time along the southeastern margin of the Canadian shield. Today these rocks constitute the Sutton and Notre Dame anticlinoria. They are underlain by a rejuvenated gneissic, Grenville-like basement which crops out towards the southwest in the Green Mountains of Vermont (Cady, 1969), and are faulted against the ophiolitic complexes to the southeast.

The ophiolitic complexes are regarded as tectonically emplaced fragments of the Proto-Atlantic (Iapetus) oceanic lithosphere (Laurent, 1973), or the remnants of the crust and mantle of small marginal or inter-arc basins. They were emplaced by thrusting upon the ancient continental margin where they occupy a constant structural position, between the underlying pre-Ordovician Notre Dame schists to the northwest and the overlying Ordovician St. Daniel breccia and Magog flysch to the southeast (fig. 2). The St. Daniel Formation is an important stratigraphic and structural marker of probable late Early Ordovician age occurring to the southeast of and within the ophiolitic belt. It is a chaotic deposit containing blocks of pre-Ordovician schists and, locally, ophiolitic rocks all in a black shaly matrix (St. Julien, 1968). The St. Daniel Formation is unconformably overlain by the basal beds of the well dated Magog flysch (Lamarche, 1972), which contains Middle Ordovician graptolites of the Nemagraptus gracilis and Diplograptus multidens Zones (Riva, 1974).

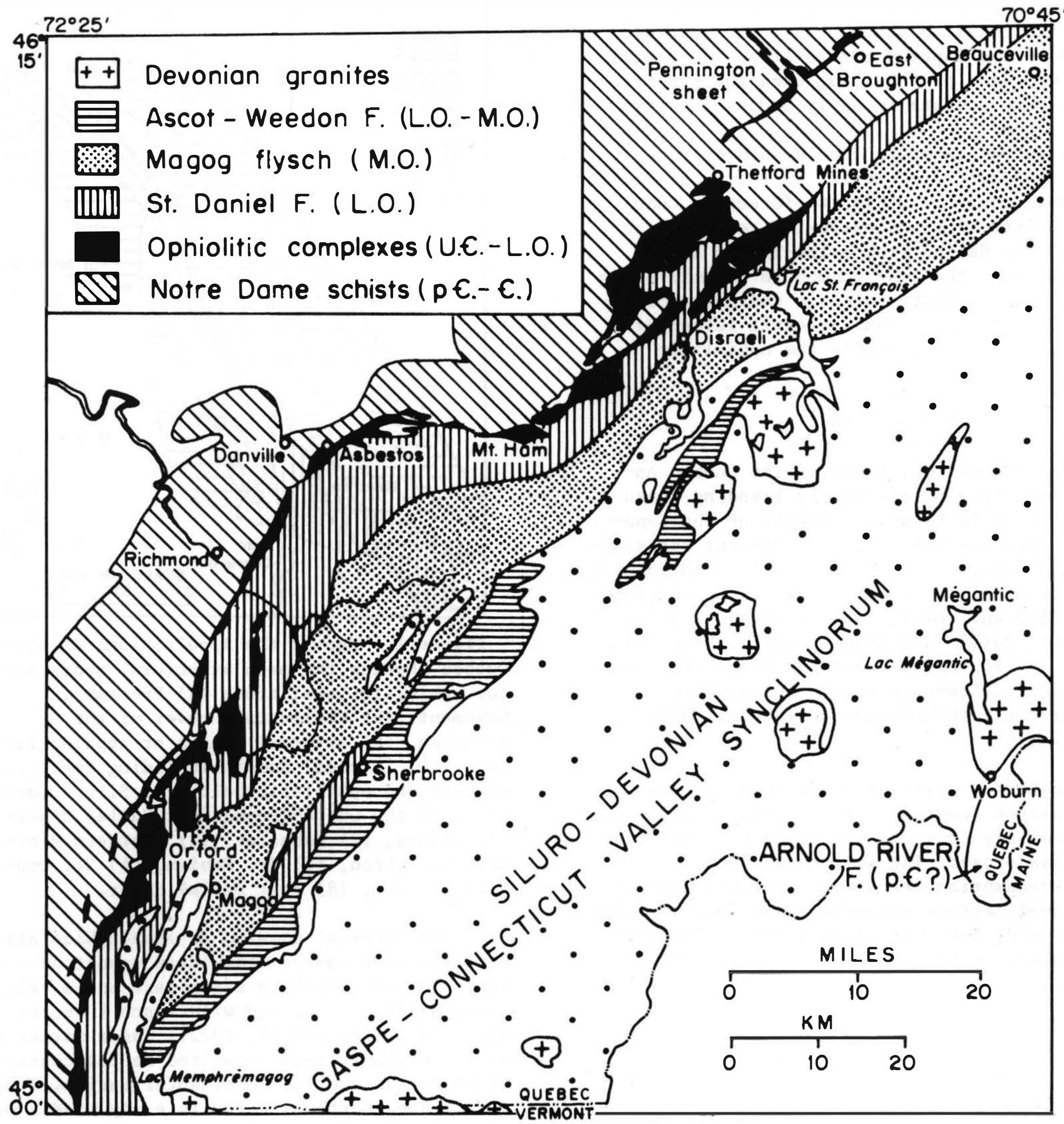


Figure 2. Sketch map of the Appalachians of Southeastern Quebec, modified from St. Julien (1967).

Field evidence shows that the polygenic St. Daniel Formation postdates the emplacement of the ophiolites as well as the main deformation and metamorphism of the pre-Ordovician Notre Dame schists, but predates the sedimentation of the Middle Ordovician Magog flysch. Hence, the St. Daniel Formation indicates that the ophiolitic complexes were emplaced by the end of the Lower Ordovician, and also suggests, since blocks of pre-Ordovician schists are associated with blocks of ophiolitic rocks, that the tectonic emplacement of the ophiolitic complexes is correlatable with the main deformation and metamorphism of the pre-Ordovician Notre Dame rocks. The orogenic episode is referred to as the Burlingtonian orogeny in western Newfoundland (Kennedy, 1975), where similar tectonic and stratigraphic relations are observed within the Fleur de Lys zone C between the Baie Verte ophiolites and the pre-Ordovician Fleur de Lys rocks (Williams et al., 1974).

## OCCURRENCES

The ophiolitic belt can be best described as a zone of tectonic mélange whose backbone is formed by the large ophiolitic complexes of Thetford Mines, Asbestos and Orford. Many smaller bodies of ophiolitic rocks as well are scattered along the complex boundary zone separating the pre-Ordovician Notre Dame schists from the Ordovician St. Daniel breccia and Magog flysch. Also lenses and screens of peridotite, such as the Pennington sheet of East Broughton (fig. 2 and 4), are found near the boundary with the ophiolitic belt but in thrust faults within the pre-Ordovician schists (St. Julien, 1972; Laurent, 1975a).

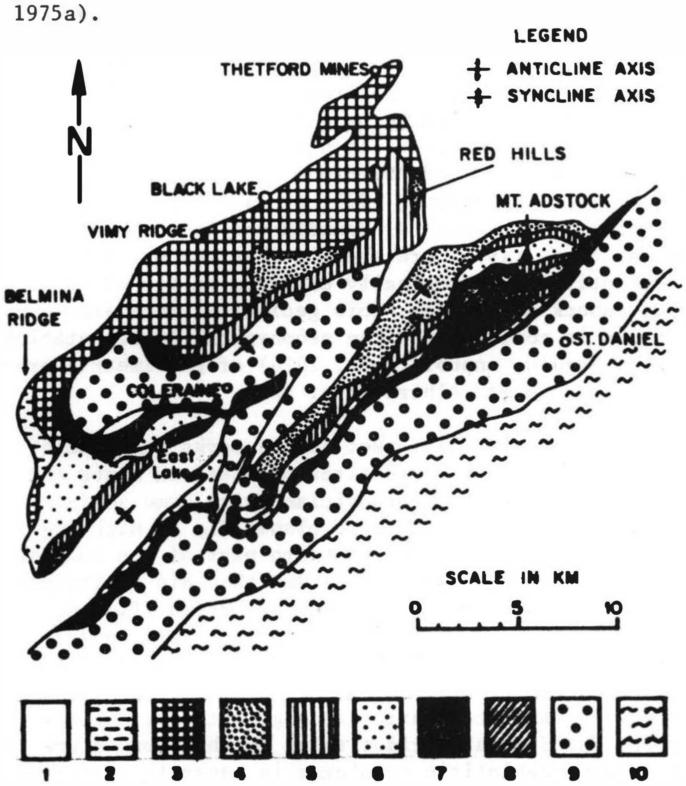


Figure 3. Map of the Thetford Mines ophiolitic complex: (1) pre-Ordovician Notre Dame schists, (2) amphibolite, (3) harzburgite tectonite, (4) dunite and wehrlite, (5) pyroxenite, (6) gabbro, (7) pillow lava, (8) chert and argillite, (9) St. Daniel Formation, (10) Middle Ordovician Magog flysch. See fig. 4 for cross section.

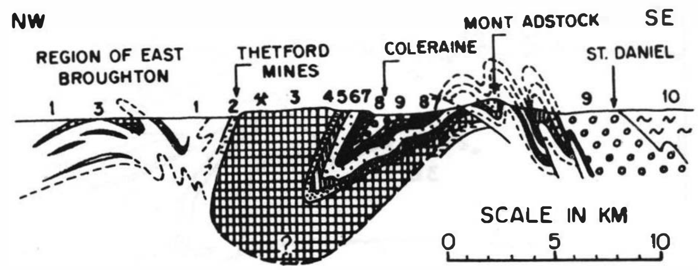


Figure 4. Cross section through the central and thickest part of the Thetford Mines ophiolitic complex. See fig. 3 for key to numbers.

The large ophiolitic complexes occur as stratified sheets bounded by faults, and they appear to have been tectonically emplaced into their present setting as solid masses. They are also internally deformed and are folded together with the underlying pre-Ordovician schists. Ophiolite suites vary from complete to extensively dismembered. When suites are complete, for example at Thetford Mines (fig. 3 and 4), the stratified sheet is divisible into lower and upper structural and lithologic units. The lower unit consists of harzburgite tectonite underlain by a thin sole of serpentinite and amphibolite. The upper unit is stratiform and consists of ultramafic and layered gabbroic cumulates followed by metagabbro which, in turn, is capped by metadiabase, tholeiitic metavolcanics and red chert and argillite. The basal peridotite contains a group of rootless granitic dikes which have yielded Early Ordovician K-Ar ages (Poole et al., 1963). Lower Ordovician (?) lavas and immature volcanic sediments overlap stratigraphically the ophiolitic suite (Laurent and Hebert, 1977). A similar succession is dated as Lower Ordovician in the Betts Cove ophiolitic complex of Newfoundland (Upadhyay et al., 1971). Other ophiolitic suites are dismembered at various extents. The Asbestos complex lacks most of the upper sequence of metagabbro, diabase and volcanics (fig. 5 and 6). At Orford, the ophiolitic suite is complete but the upper unit, from the ultramafic cumulates to the volcanic top, is detached from its basal peridotite tectonite and separated from it by a zone of tectonic mélange (fig. 7). The Mt. Albert body in the Shickshock Mountains of Gaspé Peninsula (fig. 1) consists only of peridotite tectonite.

Evidence regarding the allochthonous character of the ophiolites of southern Quebec is provided by their relationship with the pre-Ordovician country rock, their structural context and geophysical data. Data from gravity surveys indicate that even the largest peridotite bodies of the ophiolitic belt, such as those of the Thetford Mines area, extend to depths of not more than 5 to 10 kilometers. Figure 8a shows a gravimetric profile across the Thetford Mines ophiolitic complex and its geophysical interpretation (M.K. Seguin, unpubl. data, 1976) which compares reasonably well with the geological interpretation (fig. 4) and the aeromagnetic data (Seguin, 1977). The local gravimetric anomaly caused by the occurrence of the Thetford Mines ophiolitic complex is superimposed on the broader regional gravimetric anomaly caused by the Sutton and Notre Dame anticlinoria.

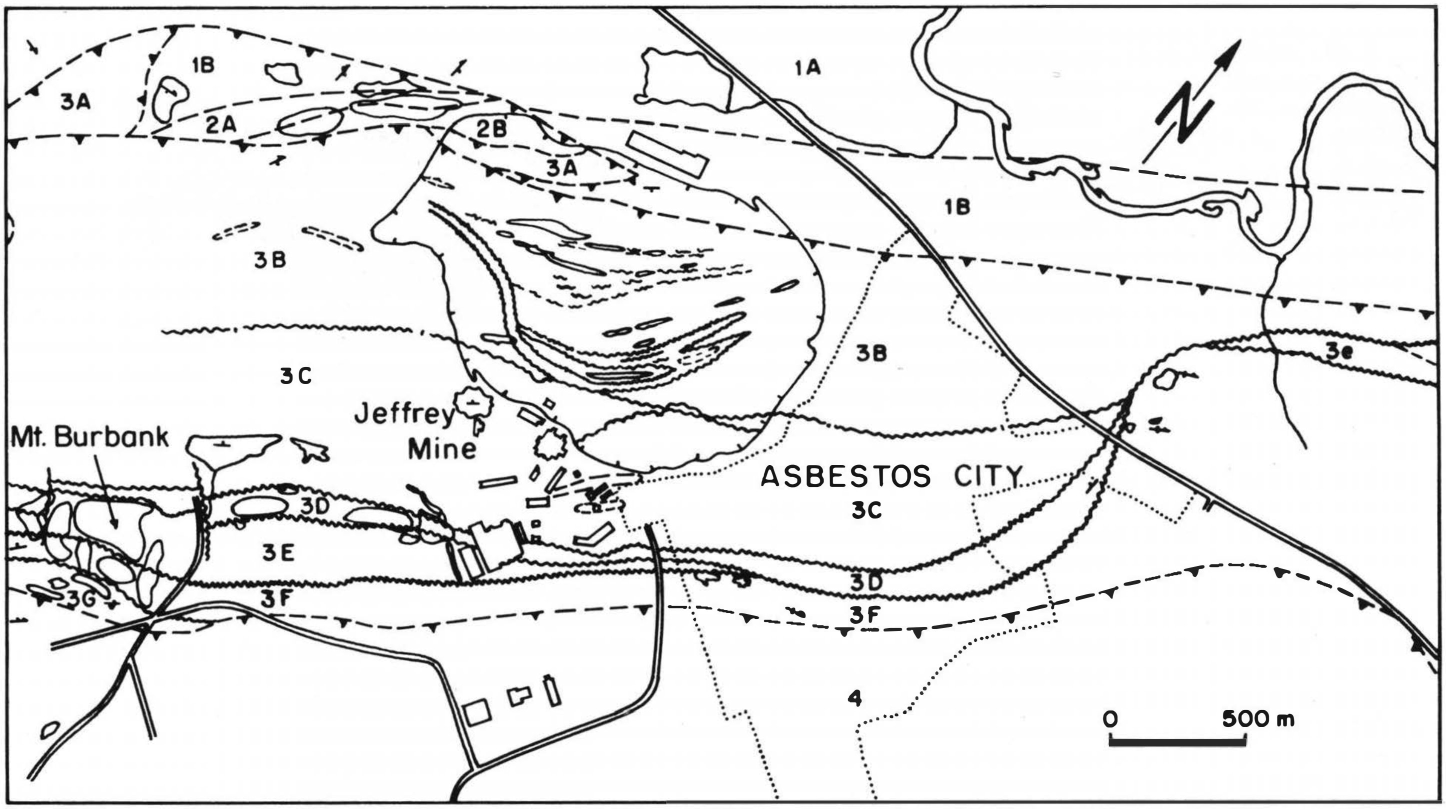


Figure 5. Map of the Asbestos ophiolitic complex compiled by R. Hébert from Lamarche (1973): (1) pre-Ordovician Notre Dame metagreywacke (A) and phyllite (B); (2) amphibolite layered (A) and schistose (B); (3) ophiolitic complex with serpentinite (A), harzburgite (B), dunite (C), pyroxenite (D), gabbro (E), metagabbro and metadiabase (e), pillow lava and volcanic breccia (F), and red chert and argillite (G); (4) breccias and shales of the St. Daniel Formation.

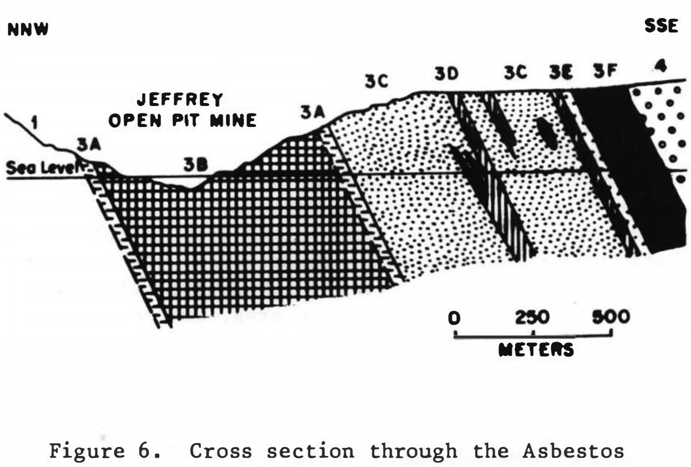


Figure 6. Cross section through the Asbestos ophiolitic complex at Jeffrey Mine, Johns-Manville Co. Ltd. See fig. 5 for key to numbers.

The extraction of the residual anomaly of Thetford Mines was obtained by M.K. Seguin through the use of a second regional trend. In the Quebec Appalachians there is no evidence that any peridotite is directly rooted in the mantle. The Mt. Albert body which still recently was regarded as an in situ ultramafic intrusion (MacGregor, 1962; MacGregor and Smith, 1963) appears to be a thrust sheet somewhat similar to the White Hills peridotite sheet of western Newfoundland. S. Biron and A. Vallières (in St. Julien and Hubert, 1975, p. 343) and structural studies in progress (J. Beaudin and St. Julien, pers. comm., 1976) are showing that the Mt. Albert peridotite is tectonically emplaced on top of the Shickshock allochthon. This geological interpretation is confirmed by the geophysical interpretation (M.K. Seguin, unpubl. data, 1976) of gravimetric profiles across Mt. Albert (fig. 8b).

A system of east-west faults marks the contact between the basal peridotites of Thetford Mines and Asbestos and the pre-Ordovician Notre Dame schists. At Thetford Mines the fault planes dip at high angle northwards (because the northern part of the complex is overturned to the south), and they dip southwards at Asbestos and southeastwards at Orford (fig. 4, 6 and 7). In the contact zone chlorite, talc, quartz and carbonates have been formed at low temperature, after the emplacement of the ophiolites, through hydrothermal metasomatic exchange with the country rock. The frontal peridotite or northwestern flank of the large ophiolitic complexes is severely shattered and cut by shear zones, which are highly serpentinized and often rich in asbestos ore (Riordon, 1953).

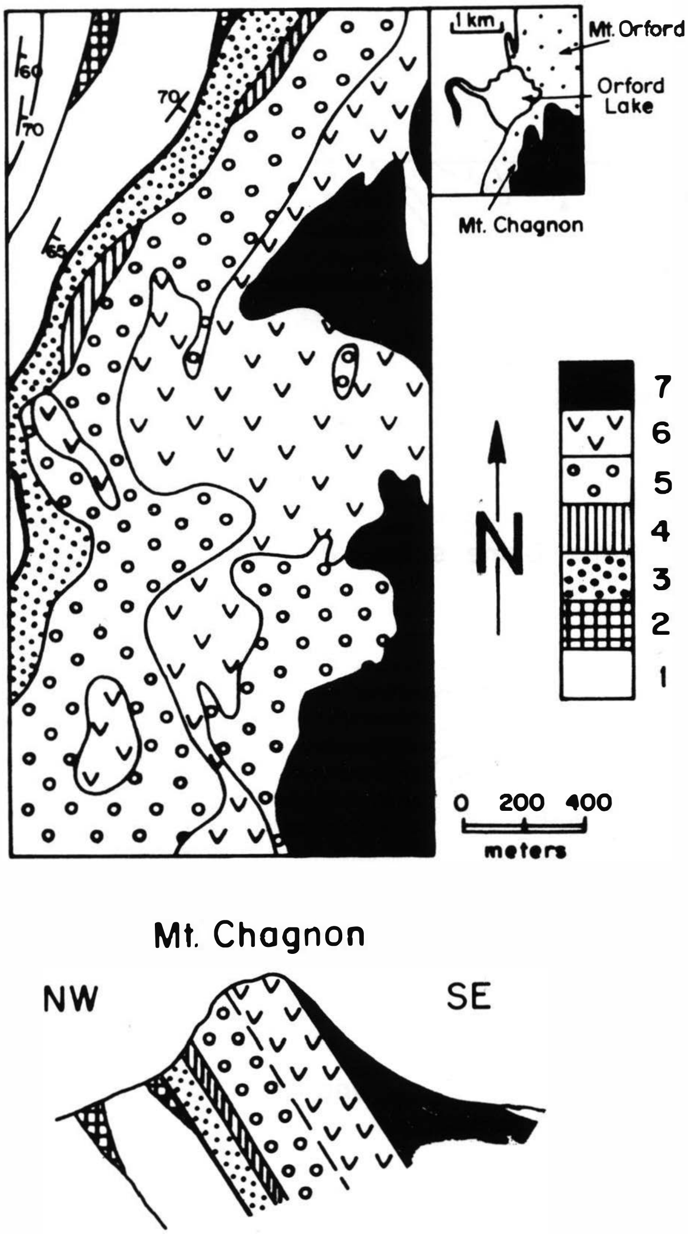


Figure 7. Map of the Mount Chagnon, part of the Orford ophiolitic complex. Geology and cross-section modified from De Römer (1963): (1) pre-Ordovician Notre Dame schists, (2) harzburgite, (3) dunite and wehrlite, (4) pyroxenite, (5) layered gabbro, (6) metagabbro, quartz diorite and plagiogranite, (7) pillow lava and volcanic breccia.

The floor, or basal plane limiting the peridotite body at depth is formed by a sole of schistose serpentinite. Some foliated hornblende amphibolites are locally associated with the basal serpentinite sole at Thetford Mines, East Broughton, Asbestos and Mt. Albert. Although discontinuous and not extensive, the occurrence of these rocks is significant. They are localized in the same position as the foliated ophiolitic amphibolites of western Newfoundland which are viewed by Church and Stevens (1971), and Williams and Smyth (1973) as contact dynamothermal aureoles related to the transport of the ophiolites.

Unlike many Alpine ophiolites, the ophiolites of the Quebec Appalachians are not overprinted by blueschist or other high pressure metamorphism. The rocks have partly recrystallized in the upper greenschist regime of low pressure and moderate temperature which suggests that they have not been subducted into a trench but were directly emplaced onto the continental margin by obduction. Doolan et al. (1973) and Trzcienski (1976) have reported the presence of metamorphic sodic amphiboles, mainly of magnesioriebeckite and crossite composition within a greenschist mineral assemblage of pre-Ordovician volcanics from northern Vermont and southern Quebec. It seems likely that these metabasaltic volcanics, which occur at or near the base of the pre-Ordovician schists underlying the ophiolitic complexes, have recrystallized at rather high pressures in a regime representing a possible transition from greenschist to blueschist facies. The conditions may be due to overpressure related to thrusting of the ophiolites. In areas where they are not overlain by the ophiolites the Caldwell metabasalts, which occur at the top of the pre-Ordovician schists, are metamorphosed in the epidote-pumpellyite facies and their structures and textures are well preserved. On the other hand the Caldwell metabasalts are metamorphosed in the greenschist facies and strongly deformed and recrystallized in areas where they have been tectonically overlain by the ophiolites. This is for example the case of the Caldwell metabasalts bordering the front of the Thetford Mines ophiolitic complex between the localities of Thetford Mines and Black Lake (fig. 3).

Dioritic rocks and a younger group of quartz monzonite (De, 1972) were emplaced as stocks and dikes within the peridotites of Thetford Mines, Asbestos and Orford. These granitic intrusions appear to have no roots and no granitic analogs are known to intrude the underlying pre-Ordovician schists in the vicinity of the ophiolites. This shows that these intrusive rocks have been emplaced within the peridotites before the ophiolitic complexes had reached their present position.

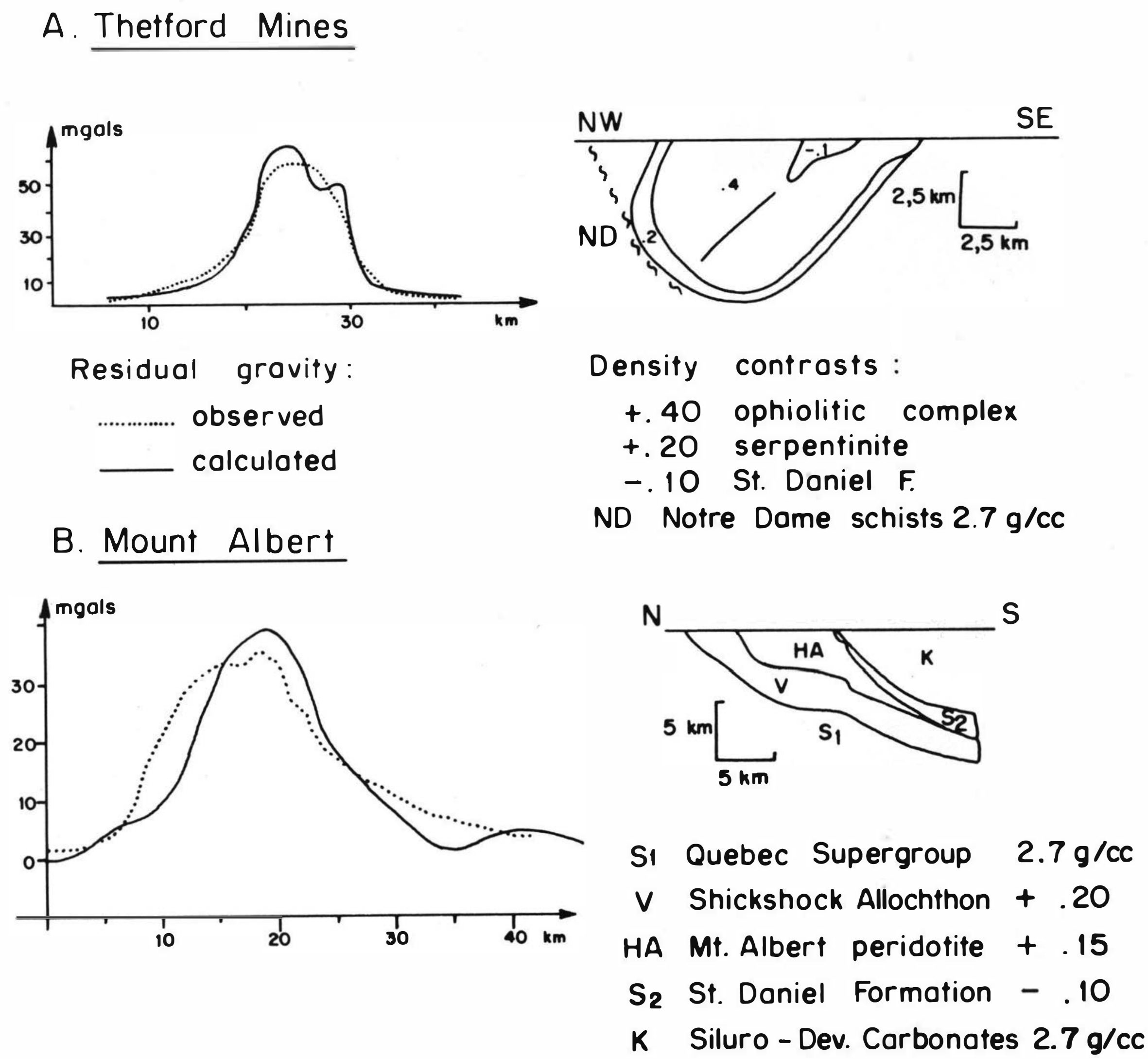


Figure 8. Gravity data and geophysical interpretation of the Thetford Mines ophiolite and Mount Albert peridotite body according to unpublished work of M.K. Seguin (1976).

The ophiolitic belt parallels the regional northeast trending structures of the Cambro-Ordovician belts for about 300 kilometers in southern Quebec marking a major suture connected to plate boundary, here the Early Paleozoic southeastern boundary of the Canadian shield (fig. 2). To the northeast the suture is believed to extend through Newfoundland. Recently St. Julien et al. (1976) and H. Williams (pers. comm., 1976) have emphasized the regional continuity of this tectonic feature which they have termed the "Baie Verte - Brompton line" (Baie Verte in western Newfoundland, Brompton in southern Quebec). To the southeast, towards the United States the ophiolitic belt is covered unconformably by the younger Middle Ordovician Magog flysch and the Siluro-Devonian limestones and sandstones of the Gaspé - Connecticut Valley synclinorium.

## LITHOLOGY

Table I summarizes the stratigraphic sequence of the ophiolites of the Quebec Appalachians. Next a description of the rock units and a short interpretation of their petrology are given.

### Peridotite tectonite

The lower unit is ultramafic in composition and consists mainly of harzburgite grading locally to dunite; constituent minerals are: olivine $Fo_{89-92}$ (65-95 %), orthopyroxene $En_{92-94}$ with exsolution lamellae of clinopyroxene, and less than 1 % chromian spinel (MacGregor, 1962; MacGregor and Smith, 1963; Kacira, 1971; Laurent, 1975b). Most harzburgite is homogeneous and strongly foliated, but some is layered with alternating orthopyroxenitic and dunitic bands on the scale of cm or dm. Dunite and orthopyroxenite also form veins, dykes and irregular bodies intrusive into the main harzburgitic body. The layering is rapidly anastomosed and reveals that the peridotites are complexly folded. Their tectonic foliation is parallel or slightly oblique to the layering. The latter, possibly, is the relic of a structure originated by one or several episodes of partial melting and high grade metamorphic differentiation that must have occurred within the mantle before the peridotite had acquired its main tectonic foliation.

The most significant feature of the harzburgite is its tectonite fabric which is expressed on the outcrops by a foliation defined as the plane of preferential orientation of the orthopyroxenes and recrystallization mosaics of olivine. Under the microscope the texture is blastomylonitic with orthopyroxene phenoclasts 1 to 30 mm long set in a foliated microblastic matrix of olivine grains of two distinct sizes. Orthopyroxenes are bent and elongated, and present deformation lamellae parallel to (100) and kink bands. Large olivine grains, up to 5 mm in size, are strained and granulated. Mosaics of much finer-grained polygonal olivine grains, with many triple-grain junctions, are observed enclosing orthopyroxene phenoclasts and spinels (see illustrations in Laurent, 1975b). In some occurrences the peridotite is so strongly foliated that its fabric is nematoblastic, with elongated olivine grains 8 times longer than their width and orthopyroxene phenoclasts up to 12 times longer than their width, both constituents presenting sharp deformation lamellae. An excellent example of this type of foliated and lineated tectonite is the fresh Mt. Albert harzburgitic dunite of the Lac du Diable area. In other occurrences, at or near the base of peridotite bodies, mylonites are developed in which most of the olivine has recrystallized. Ave'Lallemant and Carter (1970), Nicolas et al. (1972), Green and Radcliffe (1972), and others have shown that tectonite fabrics of peridotites are likely to result from an episode of slow and extensive plastic deformation and recrystallization under lower crust or upper mantle pressure and temperature conditions. It is believed that foliation, blastomylonitic fabric and mineral textures characteristic of solid state deformation and annealing have formed during the ascent of the peridotite from the mantle and its accretion within the oceanic lithosphere.

The peridotite contains a varied assemblage of rootless intrusive rocks. Three groups can be distinguished on the base of differences in composition, stage of alteration and deformation. The oldest group consists of isolated lenses of massive rodingite probably deriving from anastomosed dikes of gabbro metasomatically replaced by calcium-rich silicates during the serpentinisation of the peridotite. The second group consists of small bodies of hornblende-biotite diorite strongly deformed, altered and partly rodingitized (De, 1972). These two first groups are genetically associated with the ophiolitic suite. The third group consists of a younger complex of quartz monzonite dikes still relatively fresh and massive, although their margins are rodingitized (De, 1972). The peridotite was already serpentinized at the time of the monzonite intrusion since the serpentinite at the contact with monzonite was dehydrated and converted into chlorite, diopside, talc and anthophyllite (Cooke, 1937; De, 1972).

Fresh harzburgite is rare and most of the peridotites are partly or completely serpentinized. Two main episodes of serpentinization are recognized: a first one resulting from the partial or complete replacement of the ferromagnesian silicates by the lizardite-bastite (± magnetite-awaruite) assemblage; and a second one leading to the formation of asbestos veins (chrysotile-brucite-magnetite) which occupy late stage dilation fractures cutting across all other peridotite structures and taking various orientations (Cooke, 1937; Riordon, 1953). Furthermore, several generations of antigoritic serpentine, picrolite, brucite and chrysotile slipfibers mantle the fracture and fault surfaces, and are of dynamometamorphic origin. I have assumed that the peridotite was first hydrated and pervasively serpentinized in reducing conditions after having cooled to a temperature lower than 400°C in the oceanic environment (Laurent, 1975b). Alternatively, serpentinization may have taken place during low grade regional metamorphism at the time of transport and folding of the ophiolites. This, however, is unlikely since the granitic dikes predate the tectonic emplacement of the ophiolites and have intruded a peridotite already serpentinized. During the final stage of the tectonic emplacement and folding of the ophiolites, asbestos veins started growing in dilation fractures throughout the peridotite, and through the action of oxygen-rich waters. Conditions were different in the Mt. Albert body. This peridotite was probably not serpentinized at the time of its tectonic emplacement but was later partly altered to a talc-tremolite-magnesian chlorite-serpentine assemblage (see Aumento, 1970), an episode perhaps related with the thermal metamorphism caused by the intrusion of Devonian granites within the Shickshocks.

The amphibolites attached to the serpentinite sole at the floor of the peridotite may represent a metaigneous member of the ophiolitic suite or a metamorphic facies developed in the country rock at the contact with the peridotite. Analyzed amphibolites have a composition near that of the gabbros and olivine metatholeiitic lavas of the ophiolitic suite; in particular they are much too rich in Mg, Cr and Ni to derive from the metavolcanic and metasedimentary country rocks (table 2). X-ray diffraction study

TABLE 1

ROCK UNITS OF THE APPALACHIAN OPHIOLITIC BELT OF SOUTHERN QUEBEC

| | | |
|---|---|---|
| Arc island (?) assemblage | (up to 600 m thick) | Pillowed metabasalts, meta-andesites, volcaniclastic tuffs, breccias and mudstones |
| UPPER UNIT (up to 2800 m thick) | Extrusives and cover (up to 600 m thick) | B. Red chert and argillite<br>A. Pillowed metabasalts of tholeiitic and picritic composition |
| | Shallow intrusives (up to 700 m thick) | Non-cumulate metagabbro, overlain by meta-diabase and intruded by swarms of aphanitic diabase dikes and stocks of quartzdiorite |
| | Cumulates (up to 1500 m thick) | C. Gabbroic Zone: Layered gabbro<br>B. Pyroxenitic Zone: Olivine pyroxenites, pyroxenites and plagioclase pyroxenites<br>A. Dunitic Zone: Dunite with wehrlite and chromitite |
| LOWER UNIT (up to 5000 m thick) | Main body | Peridotite tectonite (harzburgite $\pm$ dunite) |
| | Basal sole | Serpentinite and amphibolite |

indicates that the amphibole is an hornblende whose composition and structure are identical in all localities, an observation difficult to explain by contact metamorphism. I prefer to believe these amphibolites are of igneous origin and, alike amphibolites found in oceanic fracture zones, were formed in relation with the processes of oceanic lithosphere fragmentation that must have necessarily occurred before the ophiolites were thrust onto the continental margin. It is evident, however, that the amphibolites have been affected by deformation and retrograde metamorphism during and after their emplacement with the ophiolites, as observed in the field and shown by fabric analysis (Williams and Smyth, 1973; Atkinson, 1976) and Argon isotopic data. In spite of the identity of the amphibole analyzed their K-Ar ages vary along the belt in probable relation with the intensity and duration of the metamorphic overprints (Laurent and Vallerand, 1974). Due to the complex history of the rock and its very low potassium content, ages obtained are scattered from the Middle Cambrian to the Upper Ordovician and are not reliable.

## Cumulates

Cumulate rocks are bedded and stratigraphically ordered from bottom to top in three zones, respectively of dunitic, pyroxenitic and gabbroic composition. Cumulate dunites at the floor of the cumulate sequence are in sharp contact with the harzburgite tectonite. This contact is always faulted, serpentinized and altered. The magnetic susceptibility of cumulate dunite averages $4 \cdot 10^{-3}$cgs emu cm$^{-3}$ and that of underlying harzburgite $8 \cdot 10^{-4}$cgs emu cm$^{-3}$ (Seguin, 1976). The much higher susceptibility of dunite is caused by its higher content of chromite and magnetite, the olivine being always entirely serpentinized. These two rocks having distinct susceptibility ranges, their contact can be ascertained by the magnetic methods.

Cumulate rocks have granoblastic textures and are not foliated in spite of the fact that they are often tightly folded around axes parallel to the bedding. Dikes, without chilled edges, of clinopyroxenite in the lower part of the cumulate sequence and of gabbro in the upper one cut across the folds and are not affected by the folding. Magmatic breccias are frequently associated with disturbed beds. Since cumulate textures of the folded rocks are not obliterated, the folding must have occurred plastically at relatively high temperature. This suggests that the environment of the magma chamber was tectonically active during the sedimentation of the cumulates. Early formation of fractures, quickly cemented by pyroxenite or gabbro, shows prolongation of the tectonic activity after the cumulates had cooled enough to undergo brittle failure. Locally, numerous dikes of diabase cut through all structures. Because of frequent folding actual thickness of the cumulates is difficult to estimate and is always smaller than that of the measured sections. Furthermore the same stratigraphic succession extends for short distances only, and in places parts of the succession are faulted out.

The lower part of the dunitic zone consists of cumulate cycles beginning with dunite and terminating with wehrlite. Beds of olivine adcumulate are fine-grained and isomodal, several tens of meters thick, while olivine cumulates with intercumulus clinopyroxene are thin and layered. In the upper part of the dunitic zone, chromite becomes a major cumulus phase at the end of dunitic cycles forming chromite-rich ribbons and, locally, chromitite layers or lenses (see Irvine and Findlay, 1972, p. 113). Composition of olivine is not known because it is replaced by lizardite and chrysotile in mesh and hour-glass textures. The clinopyroxene is diopsidic and partly replaced by tremolite, talc, chlorite, clinozoïsite and iron oxides. In chromite-rich cumulates, chromite either occurs between the olivine grains as rudely octahedral grains, or as smaller inclusions restricted to olivine margins showing that crystal shapes of serpentinized olivine grains are disguised by postcumulus olivine overgrowth. Kacira (1971) has shown that chromite in the dunite cumulates of Thetford Mines is richer in iron and chromium (33 to 41 % Cr), and poorer in magnesium and aluminium than the chromian spinels (10 to 34 % Cr) in the harzburgite tectonite. Apparent thickness of the dunitic zone may reach 480 m at Asbestos, 200 m at Orford and about 250 m at Thetford Mines in the Lake Caribou area.

Above the dunitic zone the character of cyclic units changes. They become much thinner and better developed. In the pyroxenitic zone, beds vary in thickness from a few centimeters to a maximum of about 20 m. Cumulate cycles are formed by dunite or wehrlite followed, at a phase contact, by olivine websterite and websterite which grade to clinopyroxenite. The orthopyroxene hypersthene $En_{70-90}$ largely replaced by lizardite, joins olivine as a cumulus phase in olivine websterites where the clinopyroxene occurs as intercumulus. In websterites, cumulate orthopyroxene can form up to 50 % of the rock with cumulus and postcumulus clinopyroxene forming the rest. Higher in the pyroxenitic zone, olivine and orthopyroxene progressively disappear as cumulus phases while postcumulus plagioclase appears. Cycles are formed by thick beds of adcumulate clinopyroxenite and thinner layers of orthocumulate plagioclase clinopyroxenite, the calcic plagioclase being rodingitized that is to say replaced by hydrogarnet or prehnite and clinozoïsite. Apparent thickness of the pyroxenitic zone reaches up to 500 m at Thetford Mines in the Red Hills area, but only 70 m at Asbestos and about 200 m. at Orford.

Cycles of the gabbroic zone combine three major cumulus phases: diopsidic augite (partly uralitized), hornblende, and labrador (partly prehnitized or saussuritized). Olivine with orthopyroxene re-appear in rare occurrences. Cumulate gabbros form layered beds of 10 cm to 1 m thick consisting of fine-grained mafic gabbro at the base grading upwards into a slightly coarser grained gabbro or hornblende gabbro; tops of beds are leucogabbroic. They are invaded by patches of hornblende gabbro pegmatite and other gabbroic rocks highly variable, on the scale of dm and m, in composition (from mafic to felsic) and grain-size (from pegmatitic to finely grained). These later phases must have solidified in an environment where temperature, confining pressure and gaz fugacity were undergoing rapid changes. Evidence of plastic folding is generally lacking in the gabbroic zone. Instead, beds of cumulate gabbro which not rarely are fresh and unmetamorphosed may locally be

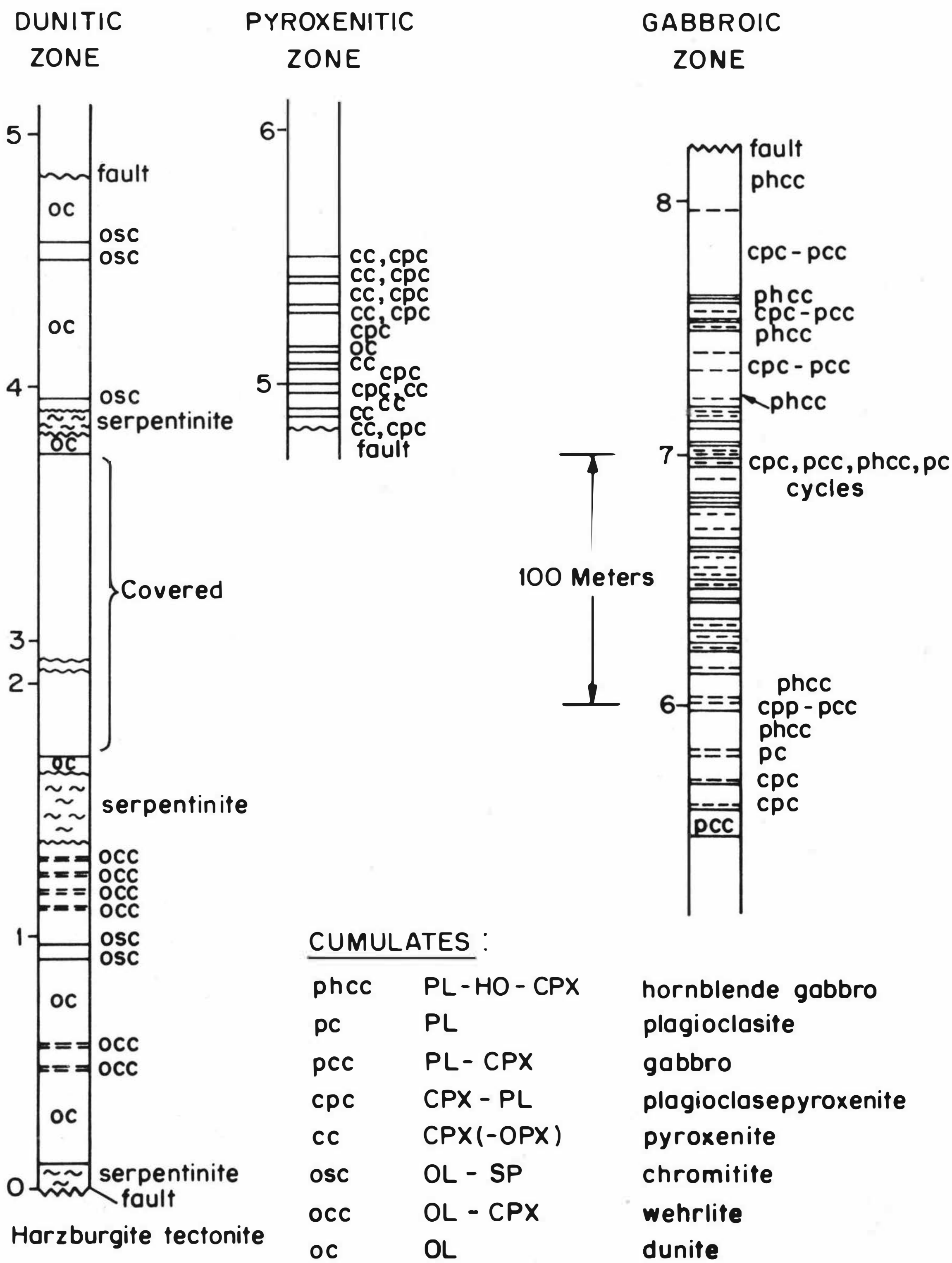


Figure 9. Columnar section of the stratiform dunite-pyroxenite-gabbro sequence in the Asbestos ophiolitic complex. Section studied and compiled by R. Hébert (Ph.D. thesis in preparation, Laval University).

cataclased and recrystallized displaying the dynamometamorphic structure of "flaser" gabbros. Thickness of the cumulate gabbroic zone is 250 m. at Asbestos (Burbank Hill), and up to 500 m at Orford and at Thetford Mines in the Mount Adstock and East Lake areas.

Hess (1960), Wager et al. (1960), Jackson (1961, 1971), Irvine (1970, 1974), and others have shown that cumulate ultramafic and mafic rocks from layered intrusions, like the Stillwater, Skaergaard, Muskox and Duke Island intrusions, are precipitates from basaltic magmas. The stratigraphic ordering of the cyclic units reveals the order of crystallization and the trend of gradual differentiation of the magma. In cumulates of the ophiolites of southern Quebec, the order of crystallization appears to be olivine (plus chromite), olivine plus clinopyroxene, olivine plus chromite, olivine plus orthopyroxene plus clinopyroxene, orthopyroxene plus clinopyroxene, clinopyroxene plus plagioclase plus hornblende. The late appearance of plagioclase in the sequence in contrast with its early appearance together with olivine in the Bay of Islands complex of western Newfoundland (Irvine and Findlay, 1972) is noteworthy. The cumulate succession described and illustrated by the composite section from Asbestos (fig. 9) is very similar to the cumulate units of the Vourinos ophiolite described and illustrated by Jackson et al. (1975). This agrees well with my earlier suggestion (Laurent, 1975a, p. 450) that the ophiolites of southern Quebec are close to the Vourinos type, an idea that I was recently able to confirm in visiting the Vourinos area.

## Shallow intrusives

Stratigraphically above the cumulates comes a complex zone consisting of metamorphosed gabbro and diabase intruded by swarms of mafic dikes and by silicic rocks.

Metagabbros form sheet-like bodies; they are mediumgrained with subophitic and ophitic textures and recrystallized in the greenschist facies with the actinolite-chlorite-clinozoisite-calcite-albite-quartz assemblage. Although relics of labradorite and augitic clinopyroxene are not uncommon, they do not contain identifiable olivine or orthopyroxene and are overlain by fine-grained rocks of diabasic texture and identical greenschist mineralogy. This complex constitutes the roof of the phaneritic rocks of our ophiolites. It is intruded firstly by a large number of dikes of aphanitic diabase bearing in places Cu-sulfide mineralization, and by a smaller number of dikes of keratophyre; and then by stocks and sills of quartz diorite and quartzrich granophyric rocks (plagiogranites). These latter silicic rocks intrude, engulf and assimilate the metagabbroic and, to a lesser extent, the metadiabasic rocks. At its contact with the metagabbros, the base of the diabasic unit is locally recrystallized in the amphibolite facies. Diabases grade upwards into pillowed flows of metatholeiite. The relations described are well displayed at the top of Mount Ham, near Ham Sud in the Thetford Mines area, and are illustrated here in figure 10. Shallow intrusives are up to 500 m thick at Thetford Mines, and up to 700 m thick at Orford where they have been described by De Römer (1963).

The metadiabases forming the floor of the ophiolitic extrusive sequence are interpreted as the first formed crust. I imagine that the non-cumulate gabbro sheets were emplaced under the first extrusive rocks and caused their recrystallization through contact metamorphism. Later, after the first formed crust had thickened, a magma chamber enclosed between a roof of gabbro, diabase and volcanics and a floor of mantle peridotite slowly filled itself with ultramafic and mafic cumulates. It is likely that the late silicic intrusions represent magmatic differentiates of the cumulate succession as it appears to be the case for Vourinos (Jackson et al., 1975). And it is also likely that brittle deformation and intrusions of diabase dike swarms, later followed by silicic differentiates are responsible for the main metamorphism of the non-cumulate gabbros.

## Extrusives

This succession is divided into two groups: a) a lower group consisting of pillowed metabasalts with a sedimentary cover of red chert and argillite, and b) an upper group of Early Ordovician age (?) consisting of metabasaltic and meta-andesitic lavas, breccias, siliceous volcaniclastic tuffs, and mudstones. The age of the lower volcanic group should indicate the time of formation of the ophiolites but is unknown. However, stratigraphic relations and paleomagnetism suggest a pre-Early Ordovician age. Preliminary paleomagnetic tests have yielded an uncertain Cambrian paleopole position at 146° E - 13° N, reverse polarity (Seguin, 1976). The total thickness of the two volcanic groups and their sedimentary cover is about 1000 to 1200 meters. As the petrology and magnetic signatures of these volcanic rocks have already been described in details (Seguin and Laurent, 1975; Laurent and Hebert, 1977), only a summary of their features will be given here.

The lower group is composed of metabasaltic pillow lavas and autoclastic breccias of upper greenschist metamorphic grade derived from tholeiites and olivine tholeiites. The two types of lava are interbedded and grade upwards into reddish, hematitic pillows interfingering and interbedded with hematite-rich radiolarian argillite separating the lower volcanic group from the upper group. Structures of pillows and lava tubes, their concentric textural zoning from a variolitic margin to the core, as well as quenched olivine and plagioclase microphenocrysts are locally well preserved. The mineral assemblage is actinolite-chlorite-epidote-albite-quartz-calcite (± magnetite). The first type of lava is rich in chlorite and tholeiitic in composition; the second type is rich in actinolite and contains abundant chlorite pseudomorphs after olivine microphenocrysts and phenocrysts, its composition being olivine tholeiitic. Chemical data show that the olivine metatholeiite is relatively high in Mg, Cr and Ni compared to the metatholeiite, and that both types of lava are very poor in K, Ti and P and metasomatically enriched in silica, soda and water (table 2). It was found that radial variations of magnetic properties in these pillow lavas are similar to the oceanic H-type pillows described by Marshall and Cox (1971), the intensity of their remanent magnetization as well as their Koenigsberger ratio decreasing from the margin to the center.

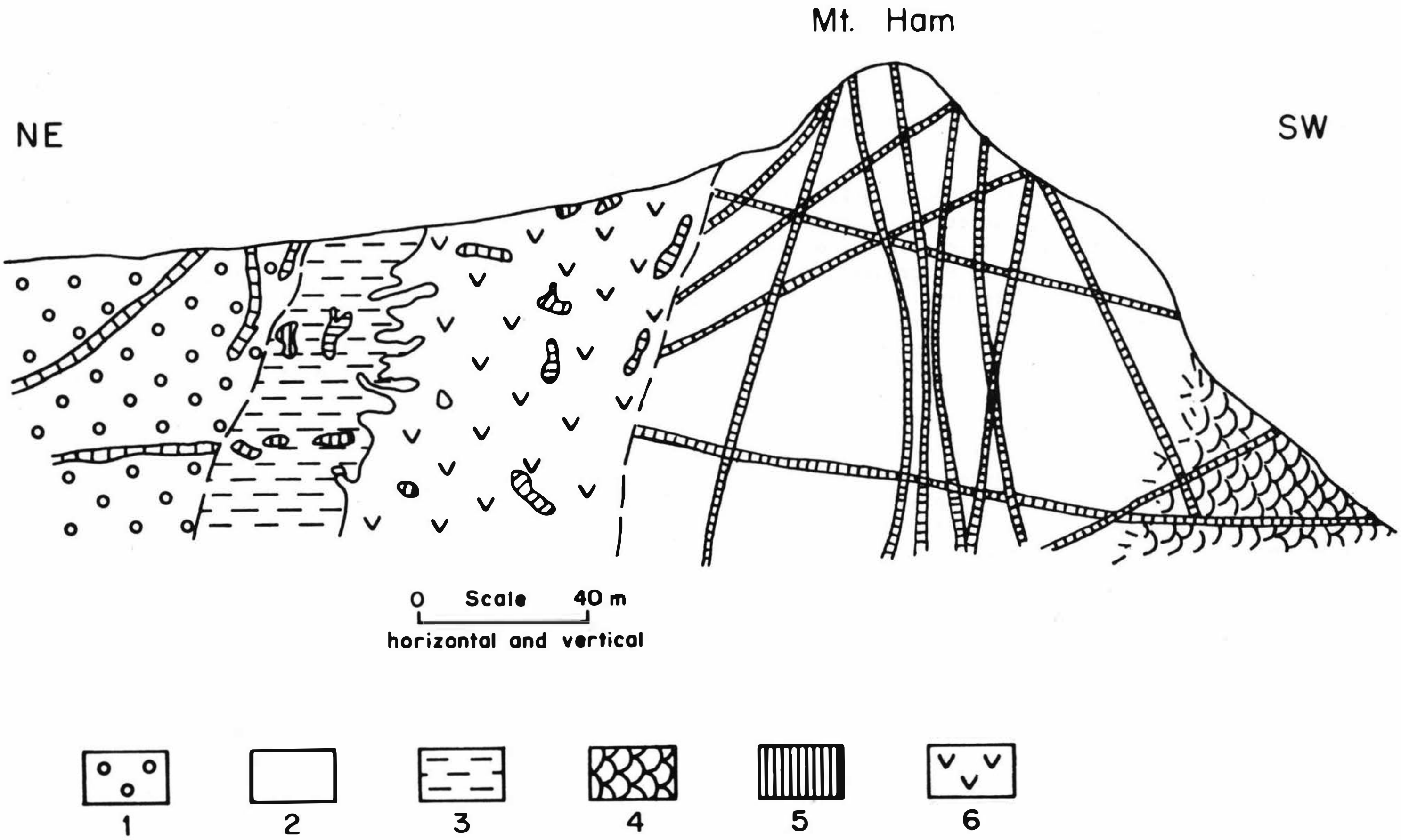


Figure 10. Cross section through the top of Mount Ham showing the relations between the upper gabbros and the lava of the extrusive sequence of the Thetford Mines ophiolitic complex. Geology according to detailed mapping by R. Hébert (Ph.D. thesis in preparation, Laval University): (1) metagabbro, (2) metadiabase, (3) amphibolitic metadiabase, (4) pillow lava, (5) aphanitic diabase dikes, (6) quartz diorite and plagiogranites.

Since radial variations of magnetic properties in pillows appear to be dependent upon the size and distribution of the NRM carriers, a function of the cooling rate of lavas, similar patterns may indicate similar physical environments. Therefore, I have assumed that this group of lavas and sediments has for analogs the oceanic layers 1 and 2 of present day oceanic crust (Fox et al., 1973). According to plate tectonic models showing ophiolite as oceanic crust generated in oceanic ridges (Coleman and Irwin, 1974), it can be postulated that the metabasalts of the lower volcanic group were extruded along the rift valley of an active ridge of the past Proto-Atlantic ocean. Composition of the sedimentary cover of this group is typical of deep-sea conditions, which agrees well with an oceanic ridge environment. This, features of the ophiolites and other paleotectonic and paleogeographic characteristics of the Quebec Appalachians do not speak in favor of an origin in small marginal or inter-arc basins as proposed by Dewey (1974) and Kennedy (1975) for the Newfoundland ophiolites.

The upper volcanic group is characterized by a suite of metabasaltic and meta-andesitic vesicular pillowed lavas, pyroclastic agglomerates, polygenic breccias, acidic volcaniclastic tuffs and mudstones which were built and deposited on the oceanic crust or ophiolite. Ophiolitic rocks and red cherts are reworked in breccias interbedded within the group. Clinopyroxene and more rarely plagioclase are locally well preserved although the rocks have recrystallized in a regime of low pressure and moderate temperature, at the limits of the prehnite-pumpellyite - upper greenschist facies, and are spilitized. They contain numerous veins of carbonates, quartz, chlorite, epidote, prehnite, pumpellyite, axinite and disseminated sulfides, mainly pyrite and pyrrhotite. The upper volcanic group may result from a volcanic activity in off-ridge oceanic environment, for example in an island-arc where brief periods of intensive volcanism were followed by structural uplift and erosion and then by subsidence and deposition of sediments. Analogs are found today in young volcanic island arcs such as the Tonga islands and the New Hebrides islands in the Pacific, where Ewart and Bryan (1972, 1973), and Mitchell and Warden (1971) have described similar rock assemblages.

The original chemical composition of the volcanic rocks forming the cap of the ophiolites has been significantly modified through metasomatic hydrothermal alterations and metamorphic recrystallization (Laurent and Hebert, 1977). Hence, a direct comparison between the chemistry of fresh volcanic rocks from the oceans and the chemistry of the ophiolitic lavas may well turn out to be meaningless. The same remark can be formulated for most of the phaneritic rocks.

TABLE 2 : SELECTED ANALYSES FROM THE THETFORD MINES OPHIOLITIC COMPLEX

A. Lower unit

Average composition of harzburgite (normalized to 100% without $H_2O$; see Laurent, 1975b): H

| | Wt. % | | | |
|---|---|---|---|---|
| $SiO_2$ | 44.22 | CaO | 0.53 | Mode: |
| $Al_2O_3$ | 0.65 | $Na_2O$ | 0.13 | OL 85%; OPX 13.5%; CPX (in OPX) 1.0%; SP 0.5%. |
| FeO | 9.06 | $K_2O$ | 0.05 | Norm: |
| MgO | 44.57 | NiO | 0.23 | Olivine $Fo_{90}$ 74.95%; Orthopyroxene $En_{90}$ 21.84%; |
| MnO | 0.14 | $Cr_2O_3$ | 0.42 | Clinopyroxene diopside 1.80%; Chromian spinel 0.5%. |

Garnet-bearing amphibolite (in Wt. %): A

| | | | | | | | |
|---|---|---|---|---|---|---|---|
| $SiO_2$ | 42.52 | FeO | 8.35 | $Na_2O$ | 1.05 | $H_2O^-$ | 0.02 |
| $TiO_2$ | 1.12 | MgO | 11.15 | $K_2O$ | 0.19 | Total | 99.47 |
| $Al_2O_3$ | 16.33 | MnO | 0.19 | $P_2O_5$ | 0.11 | Cr(ppm) | 563 |
| $Fe_2O_3$ | 2.65 | CaO | 13.22 | $H_2O^+$ | 2.50 | Ni(ppm) | 208 |

B. Upper Unit

| | Extrusives | | Shallow Intrusives | | | Cumulates | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| In Wt. % | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 |
| $SiO_2$ | 51.08 | 53.35 | 48.75 | 48.07 | 71.45 | 50.93 | 47.97 | 50.11 | 43.05 | 34.33 |
| $TiO_2$ | 0.07 | 1.09 | 2.06 | 0.15 | 0.18 | 0.03 | 0.04 | 0.29 | 0.02 | 0.00 |
| $Al_2O_3$ | 11.60 | 14.50 | 15.98 | 17.57 | 15.66 | 12.94 | 11.21 | 1.60 | 1.23 | 1.04 |
| $Fe_2O_3$ | 1.91 | 3.84 | 4.52 | 2.48 | 0.60 | 1.32 | 1.54 | 2.36 | 6.24 | 5.38 |
| FeO | 6.70 | 6.86 | 7.77 | 6.05 | 0.95 | 4.61 | 4.81 | 3.52 | 3.36 | 3.03 |
| MgO | 13.98 | 5.49 | 5.91 | 9.94 | 0.83 | 15.08 | 16.27 | 24.12 | 29.86 | 40.04 |
| MnO | 0.15 | 0.17 | 0.21 | 0.14 | 0.03 | 0.12 | 0.13 | 0.11 | 0.15 | 0.13 |
| CaO | 6.91 | 4.23 | 7.27 | 6.55 | 1.29 | 10.39 | 13.18 | 14.84 | 6.95 | 0.00 |
| $Na_2O$ | 2.21 | 5.12 | 3.60 | 2.99 | 3.39 | 1.92 | 0.30 | 0.20 | 0.09 | 0.02 |
| $K_2O$ | 0.20 | 0.16 | 0.58 | 0.96 | 3.78 | 0.84 | 0.09 | 0.01 | 0.01 | 0.00 |
| $P_2O_5$ | 0.02 | 0.07 | 0.08 | 0.00 | 0.23 | 0.01 | 0.00 | 0.01 | 0.01 | 0.00 |
| $H_2O^+$ | 3.91 | 3.82 | 2.54 | 4.17 | 1.37 | 2.12 | 3.86 | 2.82 | 8.63 | 13.91 |
| $H_2O^-$ | 0.30 | 0.20 | 0.35 | 0.26 | 0.10 | 0.13 | 0.09 | 0.18 | 0.05 | 0.52 |
| $CO_2$ | 0.42 | 0.63 | nd | 0.32 | 0.06 | nd | nd | nd | nd | nd |
| Total | 99.46 | 99.53 | 99.62 | 99.65 | 99.92 | 100.44 | 99.49 | 100.17 | 99.65 | 98.40 |
| Cr(in ppm) | 1061 | 119 | 80 | < 200 | nd | 414 | 683 | 2300 | 3300 | $Cr_2O_3$(%)1.20 |
| Ni(in ppm) | 277 | 28 | 50 | 25 | nd | 180 | 203 | 306 | 650 | 1920 |

Key to numbers:

1, Olivine metatholeiite (average of 5 analyses); 2, Metatholeiite (average of 7 analyses); 3, Metadiabase; 4, Metagabbro; 5, Quartz diorite; 6, Hornblende gabbro; 7, Gabbro; 8, Clinopyroxenite; 9, Olivine pyroxenite; 10, Chromite-bearing dunite.

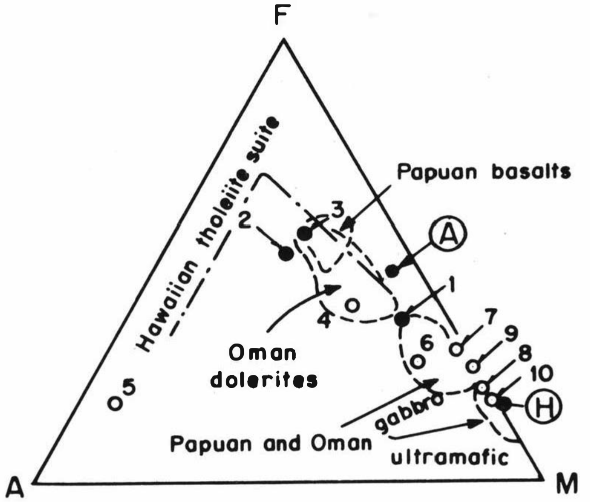


Figure 11. MFA diagram. See table 2 for key to numbers and letters.

Our chemical data (table 2, fig. 11) indicate that the ophiolitic suite is chemically characterized by high Mg and very low K, Ti and P contents, which is similar to the chemical features of the Bay of Islands and Baie Verte ophiolites of Newfoundland. As shown by the works of Irvine and Findlay (1972) and of Norman and Strong (1975) the parental magma in both cases was a low-Ti and low-K tholeiitic magma of the type common to the mid-ocean ridges.

## OPHIOLITE EMPLACEMENT

The Caldwell Group ends the pre-Ordovician Notre Dame sequence. It consists of great thicknesses of metagreywackes, slates and pillowed metabasalts. The deposition of this group may have marked the beginning of subduction at the northwestern margin of the Proto-Atlantic ocean. To the northeast of Thetford Mines the Caldwell rocks are overturned to the southeast and form the lower flank of a large recumbent fold which has originally slid in a southeasterly direction (St. Julien and Hubert, 1975, fig. 1 and 3). This movement directed towards the oceanic basin is believed to have been caused by the subsidence of the continental margin of southeastern Canada. Hence, as recorded by the geology of Notre Dame zone, emplacement of the ophiolites was preceded by subsidence of the continental margin, by basaltic submarine volcanism and by folding. Intrusive and extrusive calc-alkaline rocks of Late Cambrian and Early Ordovician age are absent from the Canadian continental margin; in contrast, Early Ordovician granitic intrusives and upper calc-alkaline volcanics are present in ophiolitic complexes. This suggests that the subduction zone dipped oceanwards at the time of the Burlingtonian orogeny. Kennedy (1975) in Newfoundland has ascribed this orogeny to the closing of a marginal basin. I prefer to postulate that the Canadian continental margin during the accumulation of the Caldwell greywackes and submarine basalts was part of a downgoing slab which reached the vicinity of the subduction zone. Deformation and metamorphism of the pre-Ordovician Notre Dame rocks are understood as resulting from the attempted subduction of the buoyant continental margin. According to Coleman (1971), Dewey and Bird (1971), and others the model involves obduction of ophiolite sheets from the leading edge of the overriding oceanic plate. Island arcs at various stages of evolution may be built on this leading edge, the association of oceanic crust and island arc sequences being actually similar to the stratigraphic suite of the Northern Appalachian ophiolites. Elliott (1976) has shown that ophiolite thrust sheets are mechanically and geometrically alike thrust slices made of sediments, and that they must be emplaced onto continental margins by gravitational downsurface slope stresses. This implies that gravitational forces are dominant and that a paleoslope is required to produce ophiolite obduction, while continent-continent and continent-island arc collisions, or other compressive processes, are not.

Following the Burlingtonian orogeny, the subduction zone may have undergone a "flip" in direction and dipped towards the Canadian continental margin at the time of the Taconian orogeny (St. Julien and Hubert, 1975), and of the building of the late Lower Ordovician and Middle Ordovician island arc volcanic sequences of southern Quebec and northern Maine (Hynes, 1976).

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# THE BALTIMORE COMPLEX, MARYLAND, PENNSYLVANIA, AND VIRGINIA

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## ABSTRACT

The assemblage of mafic and ultramafic rocks, herein called the Baltimore Complex (not to be confused with Baltimore Gneiss of Precambrian age and in the same area), is one of the largest associations of such rocks in the Appalachians within the United States. Recent work on the complex has led the author and others to conclude that these rocks are part of an ophiolite sequence that was disrupted during emplacement and subsequently severely deformed and metamorphosed. In terms of an ophiolite stratigraphy, the complex includes the following units: (A) Serpentinized peridotite including dunite with podiform chromitite and pyroxenite. (B) Gabbro, commonly norite, containing ultramafic cumulus layers of "lherzolite" and "websterite" and at many places metamorphosed to an epidote-amphibolite containing serpentinite layers. (C) Quartz gabbro and diorite containing hornblende and relict pyroxene. Units A, B, and C can all be considered as parts of a cumulus sequence characterized by the disappearance of olivine before the crystallization of cumulus plagioclase and clinopyroxene and by the appearance of postcumulus hornblende in the more differentiated rocks. (D) Volcanic rocks which include an association of pillow basalt, keratophyre, and minor intrusions into the underlying gabbro in the form of plagiogranite dikelets. Sheeted dike complexes have not been described in these rocks. The Baltimore Complex was thrust westward with a broken formation of diamictite and boulder gneiss, members of the Wissahickon Formation. The geologic evidence suggests that the minimum age for emplacement was Ordovician, but conflicting data from U/Pb zircon ages suggest that the time of emplacement could have been as old as Early Cambrian or latest Precambrian.

## INTRODUCTION

The Baltimore Complex extends 150 km southeast from southeastern Pennsylvania across Maryland to northern Virginia and constitutes one of the largest assemblages of mafic and ultramafic rocks in the Appalachian system within the United States. Plate II (in pocket) is a generalized geologic map of this complex, various parts of which have been studied for many years. The large gabbro mass underlying much of the city of Baltimore was the subject of classic papers in petrology published nearly a century ago (Williams, 1886, 1890). The ultramafic rocks along the Pennsylvania-Maryland State line are well known for their chromite deposits, which were mined during much of the 19th century (Pearre and Heyl, 1960). Throughout the history of the investigation of these rocks, the emphasis has usually been on specific units or districts within the complex. Modification of the ophiolite concept in the past decade has led to a more integrated approach for the study of peridotite-gabbro-volcanic complexes. The Baltimore Complex is now being restudied and reinterpreted (Crowley, 1976; this paper) as an ophiolite, albeit one that has been severely battered by the ordeal of emplacement and subsequent deformation and metamorphism in the Appalachian orogen.

The general geology of the Maryland Piedmont has been summarized in the many county reports and maps published by the Maryland and U.S. Geological Surveys. By county, starting in the northeast, these include: Cecil County (Bascom, 1902; M.W. Higgins, unpub. data, 1977); Harford County (Southwick, 1969; Southwick and Owens, 1968); Baltimore County and City (Knopf and Jonas, 1925, 1929; Crowley, 1976); Howard County (Cloos and Broedel, 1940); Montgomery County (Cloos and Cooke, 1953; Hopson, 1964; Froelich, 1975); and in northern Virginia Fairfax County (A.A. Drake, unpub. data, 1977). An outstanding report in this series which summarizes much of the regional geology is the stimulating paper by C.A. Hopson (1964) on the crystalline rocks of Howard and Montgomery counties. Hopson has presented a coherent synthesis of the Maryland Piedmont and provides a logical starting place for a consideration of the geologic setting for the Baltimore Complex. Following many others, Hopson's regional stratigraphy includes a basement gneiss (the Baltimore Gneiss) having an age in excess of 1 billion years (Wetherill and others, 1968) unconformably overlain by the Glenarm Group (Knopf and Jonas, 1923; Southwick and Fisher, 1967; Higgins, 1972). The lowermost formation in the group is a shelf-type sequence of quartzite (Setters Formation), followed by a carbonate sequence (Cockeysville Marble) and an argillite (the lower part of the Wissahickon Formation). Upward in the section, the Glenarm Group includes a sequence deposited during tectonically active conditions and consisting of graywacke, boulder gneiss and diamictite, and volcanic rocks (the upper part of the Wissahickon and James Run Formations). Hopson carefully described the chaotic nature of boulder gneiss or diamictite units of the upper part of the Wissahickon and he correlated several units (the so-called "Sykesville" and "Laurel") of the Wissahickon Formation. These units were thought to define a major structure, the Baltimore-Washington anticlinorium in the Maryland Piedmont. This structure extends from Washington, D.C., north and east to the Susquehanna River in northern Maryland. Hopson used this stratigraphy and structure to work out a broad regional correlation showing a deep trough filled with Glenarm sediments, the thinner shelf sediments being facies equivalents of formations to the west. Inherent in his treatment of the regional geology is the principle of sedimentary facies correlations. Spatial relationships of all units, while significantly shortened by later deformation, are preserved more or less intact according to their original site of deposition. Hopson's interpretations have found wide acceptance, and the State geologic map of Maryland (Cleaves and others, 1968) presents his synthesis in a stratigraphic and sedimentary facies section. Regional deformation, metamorphism, and upward mobilization of the basement to form mantled gneiss domes along the crest of the Baltimore-Washington anticlinorium complete this general view of the Maryland Piedmont. The peridotite and gabbroic rocks of the Baltimore Complex were deformed and metamorphosed and were thought to have intruded the Wissahickon Formation late in the history of the trough sedimentation but generally before the onset of tectonism.

Since the publication of Hopson's studies in 1964, Cecil, Harford, Montgomery, Baltimore, and Fairfax Counties have been remapped, and aeromagnetic surveys of the entire region have been completed. The new data have resulted in a considerable modification of Hopson's synthesis, although unanimity among later workers by no means has been achieved. Lenses of diamictite within the Wissahickon Formation were found to be much more widely distributed than Hopson in 1964 believed them to be, and it is difficult now to use the pattern of such lenses to define a unique stratigraphic horizon to outline the Baltimore-Washington anticlinorium. Diamictite is widespread in northern Fairfax County, Virginia, and in Cecil and Harford Counties, Maryland. Much of the Wissahickon consists of lenses of incoherent unstratified material or "broken formation." Excavations for the Washington subway system indicate that many areas thought to be underlain by plutonic rocks are in fact Wissahickon diamictite containing large floating blocks or rafts of igneous-looking rock without any original contacts preserved (A.A. Drake, oral communication, 1976). The author is led to conclude that the Wissahickon is a terrane that may be blocked into large segments of coherent strata and broken strata, but that much of it may have little internal stratigraphic order. It is within this terrane that the Baltimore Complex is emplaced. The distribution, petrology, and structure of each of the major units of the complex will be described in the following paragraphs.

## PETROLOGY

In terms of ophiolite "stratigraphy," the Baltimore Complex consists of serpentinized peridotite and chromitite, several belts of gabbro, quartz gabbro and diorite, and metavolcanic rocks including basaltic pillow lava and keratophyre. Plagiogranite dikes and anastomosing dikelets are common in parts of the complex, but sheeted dike swarms have not been described.

### Ultramafic Rocks

Ultramafic rocks of the Baltimore Complex may be divided into two groups: those constituting a serpentinite belt which consists of altered dunite associated with podiform chromitite as well as pyroxenite, and those that are partially serpentinized layers of cumulate olivine and pyroxene within the gabbro part of the complex. It should be pointed out that most maps of the complex, especially in Baltimore County (Knopf and Jonas, 1925; Crowley, 1976) do not differentiate between these very different rocks. Stratigraphic correlations between dissimilar ultramafic rocks most certainly have led to errors in the interpretations of the structure. Serpentinized ultramafic layers within the gabbro are not associated with podiform chromitite and many preserve cumulate textures.

Serpentinized ultramafic rocks derived from dunite and pyroxenite form a continuous unit at the structural base of a gabbro mass for about 30 km from the northeast end of the complex, across the Susquehanna River, to Scarboro, Maryland, in Harford County (Plate II). Farther to the south, these rocks are not in contact with the main mass of gabbro, but form a long narrow septum 12 km long, striking away from the gabbro and surrounded by diamictite of the Wissahickon Formation. Still farther to the south, physical continuity in the serpentinite belt is lost, and ultramafic rocks form small isolated lenses and pods in the Wissahickon Formation. The more important of these detached serpentinites are at Jarrettsville, Soldiers Delight, Bare Hills, Hunting Hill, and Gaithersburg; the Jarrettsville, Soldiers Delight, and Bare Hills masses contain podiform chromitite deposits. The northeastern part of the belt near the Susquehanna River is referred to as the State Line area (McKague, 1964; Lapham and McKague, 1964; Pearre and Heyle, 1960) and is the site of many chromite deposits having a long history of exploitation during the 19th century. The podiform deposit forming the Woods mine produced approximately 96,000 tons of ore between 1828 and 1880 and is the largest chromite deposit ever mined in the United States.

Dunite is the parent rock for much of the serpentinite belt in the State Line area as well as in isolated ultramafic blocks including the northern part of the Soldiers Delight mass, and also the blocks at Bare Hills and Hunting Hill. Despite extensive serpentinization of dunite, much residual olivine and accessory chromite survive and constitute about 5 to 15 volume percent of most hand specimens of serpentinite. No bastite pseudomorphs after orthopyroxene have been reported and the author has seen no other relict phase in the altered dunites of the State Line area. Chemical analyses of these rocks (see Table 1) indicate a uniform composition and a high degree of serpentinization.

Pyroxenites, most often websterites, but occasionally enstatolite, are present as a nearly continuous layer between dunite and gabbro in the State Line area, and are prominant in many parts of the Laurel Belt and in the southern part of the Soldiers Delight mass. These rocks are derived from cumulate orthopyroxene with post cumulate orthopyroxene and clinopyroxene. Rocks with relict olivine and with post-cumulate plagioclase have been described both from the Laurel Belt, and in the State Line area, but these rocks are relatively uncommon. Chemical analyses of representative pyroxenites are given in Table 1.

Compositions of relict olivine have been reported by Pearre and Heyl (1960), McKague (1964), and by Southwick (1970). Olivine from chromitite and serpentinite near the Susquehanna River is highly forsteritic (McKague, 1964), having Fo of 94 to 97 mole percent. Although this Fo content is higher than that for olivine in most alpine-type complexes (Green, 1964), it is typical of olivine compositions coexisting with chromite in podiform deposits (Irvine, 1967). In Harford County, Southwick (1970) reported a more iron-rich olivine having Fo of 86 to 92 percent. Both McKague and Southwick used X-ray and light optics for composition determinations. Average grain size of olivine appears to be about 3.6 mm, maximum length, although much larger grains infrequently occur (McKague, 1964). The author has seen olivine grains 50 mm long in drill core from the Woods mine. Extensive deformation and serpentinization have contributed to a decrease in the average grain size.

Very little data on the composition of pyroxenes from pyroxenites are available. Southwick (1970) has reported the X-ray and optic properties of seven orthopyroxenes from pyroxenites from Harford County. En contents of orthopyroxene ranges from 74 to 83 percent; some samples show zoning with a core of En 83 and a rim of En 78. Herz (1951) reported a

TABLE ONE - SELECTED CHEMICAL ANALYSES FROM THE BALTIMORE COMPLEX

| Rock Type | Serpentinized Dunite | | | Pyroxenite | | | | Gabbro | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Sample No. | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 |
| $SiO_2$ | 40.06 | 40.7 | 40.00 | 43.87 | 53.21 | 55.3 | 48.91 | 44.76 | 43.69 | 48.3 | 44.08 |
| $Al_2O_3$ | 1.37 | 0.80 | 1.50 | 1.64 | 1.94 | 2.7 | 8.81 | 18.82 | 14.70 | 18.7 | 20.01 |
| $Fe_2O_3$ | 3.02 | 7.8 | 4.90 | 8.94 | 1.44 | 1.0 | 1.04 | 2.19 | 5.59 | 1.3 | 4.22 |
| FeO | 3.43 | 1.2 | 2.50 | 2.60 | 7.92 | 11.2 | 9.52 | 4.73 | 11.63 | 3.7 | 8.61 |
| MgO | 39.02 | 36.4 | 37.60 | 27.32 | 20.78 | 24.8 | 15.19 | 11.32 | 7.42 | 9.4 | 5.01 |
| CaO | --- | 0.09 | 0.31 | 6.29 | 13.12 | 2.8 | 14.69 | 14.58 | 9.34 | 16.6 | 11.68 |
| $Na_2O$ | --- | 0.03 | 0.10 | }0.50 | 0.11 | 0.20 | 0.64 | 0.89 | 1.94 | 0.62 | 1.24 |
| $K_2O$ | --- | nil | nil | | 0.07 | 0.32 | 0.10 | 0.11 | 0.32 | 0.06 | 0.15 |
| $H_2O^+$ | 12.10 | 11.2 | 12.20 | }8.72 | 0.87 | 0.36 | 0.59 | 2.53 | 1.71 | 0.73 | 1.90 |
| $H_2O^-$ | --- | 0.74 | 0.47 | | 0.14 | 0.07 | --- | --- | 0.11 | 0.10 | 0.11 |
| $TiO_2$ | --- | 0.02 | 0.06 | 0.12 | 0.26 | 0.26 | 0.37 | 0.13 | 3.26 | 0.17 | 2.24 |
| $P_2O_5$ | --- | 0.01 | 0.02 | --- | tr | nil | tr | --- | tr | 0.02 | 0.52 |
| MnO | 0.09 | 0.08 | 0.12 | 0.19 | 0.22 | 0.15 | 0.16 | 0.15 | 0.22 | 0.10 | 0.28 |
| $CO_2$ | --- | 0.01 | 0.11 | --- | 0.10 | 0.09 | --- | --- | 0.19 | 0.16 | --- |
| $Cr_2O_3$ | 0.20 | --- | --- | 0.44 | 0.20 | --- | 0.15 | 0.08 | --- | --- | --- |
| NiO | 0.71 | --- | --- | --- | 0.03 | --- | --- | --- | --- | --- | 0.01 |
| Sum | 100.00 | 99.1 | 99.89 | 100.63 | 100.41 | 99.3 | 100.17 | 100.29 | 100.12 | 100.0 | 100.06 |

| Rock Type | Gabbro (cont.) | | | Diorite | | Volcanic Rocks | | | | Plagiogranite | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Sample No. | 12 | 13 | 14 | 15 | 16 | 17 | 18 | 19 | 20 | 21 | 22 |
| $SiO_2$ | 45.41 | 48.02 | 50.1 | 55.16 | 59.7 | 49.7 | 53.4 | 61.6 | 75.67 | 76.36 | 77.94 |
| $Al_2O_3$ | 23.05 | 20.01 | 17.0 | 17.51 | 15.0 | 15.6 | 15.6 | 15.0 | 12.28 | 17.14 | 11.75 |
| $Fe_2O_3$ | 1.52 | 1.13 | 0.33 | 2.62 | 2.3 | 2.7 | 3.1 | 4.2 | 0.85 | }1.42 | 1.65 |
| FeO | 8.35 | 7.29 | 6.7 | 5.83 | 7.0 | 6.5 | 7.8 | 6.0 | 2.59 | | 0.93 |
| MgO | 5.89 | 10.05 | 11.1 | 4.35 | 2.6 | 8.9 | 5.1 | 2.0 | 0.37 | nil | 0.25 |
| CaO | 12.52 | 11.42 | 12.5 | 8.50 | 6.7 | 10.4 | 7.8 | 4.0 | 2.65 | 3.65 | 1.28 |
| $Na_2O$ | 0.76 | 0.51 | 0.93 | 1.83 | 2.4 | 3.0 | 4.4 | 4.8 | 3.63 | 1.89 | 4.66 |
| $K_2O$ | 0.32 | 0.05 | 0.12 | 1.08 | 0.64 | 0.29 | 0.32 | 0.12 | 0.78 | 0.07 | 0.76 |
| $H_2O^+$ | 1.52 | 0.57 | 0.33 | 2.01 | 1.2 | 0.89 | 0.78 | 0.53 | 0.29 | --- | 0.39 |
| $H_2O^-$ | --- | 0.10 | 0.12 | 0.18 | 0.07 | 0.31 | 0.02 | 0.08 | 0.12 | --- | 0.07 |
| $TiO_2$ | 0.62 | 0.23 | 0.27 | 0.64 | 1.7 | 0.73 | 0.73 | 0.95 | 0.29 | 0.11 | 0.13 |
| $P_2O_5$ | 0.13 | tr | 0.07 | 0.21 | 0.53 | 0.12 | 0.04 | 0.27 | 0.05 | --- | 0.03 |
| MnO | 0.09 | 0.18 | 0.15 | 0.15 | 0.24 | 0.10 | 0.03 | 0.18 | 0.18 | 0.02 | 0.05 |
| $CO_2$ | --- | 0.25 | 0.15 | nil | nil | 0.05 | 0.05 | 0.09 | tr | --- | tr |
| $Cr_2O_3$ | --- | 0.03 | --- | tr | --- | --- | --- | --- | --- | --- | --- |
| NiO | --- | 0.01 | --- | 0.01 | --- | --- | --- | --- | --- | --- | --- |
| Sum | 100.18 | 99.85 | 99.9 | 100.08 | 100.1 | 99.3 | 99.2 | 99.8 | 99.75 | 100.66 | 99.89 |

DATA SOURCES FOR TABLE I

1. Serpentinite; Broad Brook, Harford County; Leonard, 1901; analysis VI, p. 164
2. Serpentinite; Octoraro Creek, Cecil County; Higgins, 1975; unpub. analysis, Sample S-1-4 (USGS chemical analysis W-182773).
3. Serpentinite; Hunting Hill quarry, Montgomery County; Larabee, 1969; Table 1, no. 3, p. 10.
4. Serpentinized lherzolite; Johnnycake Road, Baltimore County; Williams, 1895; W54, p. 674.
5. Websterite; Oakwood, Cecil County; Leonard, 1901; analysis V, p. 159.
6. Bronzite pyroxenite; Susquehanna River, Cecil County; Southwick, 1970; Table 5, no. 1 sample WD-2, p.412.
7. Feldspathic pyroxenite; Orange Grove, Baltimore County; Williams, 1895; sample W-69.
8. Gabbro; Wetheredville, Baltimore County; Williams, 1895; sample W-170.
9. Hypersthene-augite gabbro; Patapsco River, Baltimore County; Hopson, 1964, Table 34, no. 1, p. 146
10. Metagabbro; Maryland House, Harford County; Southwick, 1969; Table 16, no. 3, p. 62.
11. Amphibole gabbro; Rising Sun, Cecil County; Leonard, 1901; Analysis III, p. 146.
12. Gabbro; Dublin, Harford County; Insley, 1928.
13. Norite; Oak Grove, Cecil County; Leonard, 1901; Analysis IV, p. 151.
14. Hypersthene gabbro; Susquehanna River, Cecil County; Southwick, 1970; Table 5, no. 2, sample C-79a, p.412.
15. Quartz diorite; Octoraro Creek, Cecil County; Leonard, 1901; Analysis II, p. 146.
16. Biotite-hornblende diorite; Mountain Hill, Harford County; Southwick, 1970; Table 5, no. 10, sample A-49b, p. 412.
17. Basalt; Northeast Creek, Cecil County; Higgins, 1971; Table 1, no. 2.
18. Basaltic metatuff; Susquehanna River, Cecil County; Higgins, 1971; Table 1, no. 1.
19. Cummingtonite-hornblende felsite; Susquehanna River, Harford County; Southwick, 1969; Table 14, no. 4, p. 58.
20. Felsite metatuff; Cecil County; Bascom, 1902; p. 138.
21. Plagiogranite dike; Woodberry quarry, Baltimore City; Hanan, 1976, analysis 33, sample Q8-3c, p. 14.
22. Quartz diorite; Patapsco State Park, Howard County; Hopson, 1964; Table 37, no. 1, sample H37-1, p. 159.

more magnesian orthopyroxene with compositions ranging from En 83 to En 90 percent in pyroxenites from Baltimore County. Clinopyroxene forms a significant part of many pyroxenites and Southwick reports that this mineral is also zoned.

Chromite from the principal mines of the State Line area and accessory chromite from serpentinite have been analyzed by Pearre and Heyl (1960) and by McKague (1964). All the chromite is high in Cr, having $Cr_2O_3$ greater than 52 weight percent and $Al_2O_3$ usually less than 12 weight percent. Deposits of podiform chromitite occur only within serpentinized dunite. Massive chromite is characteristic of the podiform deposits, although most deposits are somewhat brecciated. Nodular ore is present in the area but is not common. Chromite grains often contain enclosed euhedra of olivine and large olivine grains contain many euhedra of chromite, suggesting simultaneous crystallization.

Serpentinization of the dunite took place during at least three separate stages (McKague, 1964; Lapham and McKague, 1964). An earlier stage of chrysotile (and lizardite) formed a mesh texture replacing much olivine. This stage was followed by the development of antigorite, which forms the bulk of the serpentinite and which developed a strong regional foliation. Finally, a third period of serpentinization resulted in veins of chrysotile and carbonates following a crude fracture cleavage. Small ultramafic masses such as the one at Hunting Hill and contact zones of the State Line area with underlying schists are extensively replaced by talc-carbonate-chlorite assemblages. Such alteration is obviously postemplacement.

Alteration of pyroxenite was concurrent with the development of antigorite. Commonly, pyroxenite has been altered to a felted mat of tremolite with minor chlorite, talc, antigorite, and magnetite. These altered rocks are often massive and textures of the pyroxenite have been preserved. In highly deformed blocks such as the Soldiers Delight and Piney Branch, these rocks have been altered to tremolite schists with a strong preferred orientation of tremolite grains.

Regional variations in the serpentine mineralogy are probably a response to a great range of regional metamorphic conditions imposed upon the complex. Regional metamorphism varies from greenschist in the State Line area (chlorite-garnet-muscovite schists) to upper amphibolite facies (kyanite-staurolite-muscovite pelitic schists with a sillimanite overprint) in the area of Baltimore County (Cleaves and others, 1968). Variation in regional metamorphism may be reflected in changes in the regional serpentinite mineralogy. Carbonates are abundant in serpentinites within greenschist terrane, but the isolated serpentinite block at Bare HIlls in a kyanite-staurolite zone of regional metamorphism contains clinopyroxene partly replaced by tremolite. These minerals may result from the reactions:

serpentine+calcite → diopside+forsterite+fluid

and

serpentine+diopside → tremolite+forsterite+fluid.

Similar reactions were reported by Trommsdorff and Evans (1972) for progressive metamorphism of antigorite of Val Malenco. Investigations on the changes in serpentine mineralogy resulting from regional metamorphism are in progress by the author.

## Gabbro

Gabbro, long associated with the city and county of Baltimore (Williams, 1886; Knopf and Jonas, 1929) is distributed in several belts (Hopson, 1964; Crowley, 1976) (see Plate II). The Baltimore City and County area is within the Laurel belt; the northern gabbro associated with the State Line ultramafic rocks forms the Bel Air belt; and the large metagabbro mass near Havre de Grace, Harford County, is the Aberdeen belt (Plate II). In addition, substantial amounts of gabbro are present in the Soldiers Delight block in Baltimore and Howard Counties and in other isolated ultramafic blocks in Montgomery County (Froelich, 1975). The Piney Creek block, Fairfax County, Virginia, is essentially a melangé of gabbro and serpentinite; this block is terminated to the west by faults and by a Triassic-Jurassic basin (Drake, oral communication, 1977).

Throughout much of the Baltimore Complex, gabbro is altered to amphibolite that has a well developed foliation (Cohen, 1937). The entire mass at Aberdeen and the isolated masses in Montgomery County are thus altered. However, relict igneous textures survive in many places within the Bel Air and Laurel belts; the discussion will concentrate on these, bearing in mind that igneous textures have been obliterated in much of the complex.

Many years ago Knopf (1921) described the Bel Air section on the Susquehanna River as a gravity differentiated igneous sheet or lopolith not unlike (her conception of) the Sudbury complex. In a gross way, the succession on the Susquehanna River is from the serpentinized peridotite just described, to pyroxenite, to hypersthene gabbro containing high-level thin layers of cumulate peridotite, to quartz-bearing gabbro, and finally to quartz diorite. Throughout the succession, the layering and foliation generally dip to the south and form a structural package that might be expected of a differentiated sill. However, no complete structural picture of the Laurel belt (the classic area of the Baltimore gabbro) has emerged, despite considerable work by Cohen (1937) and later by Herz (1950) in the Patapsco aqueduct which runs through the center of the gabbro. Hopson (1964) emphasized that ultramafic rocks are concentrated on the margins of the Laurel belt in Baltimore County, but neither William's (1886) map nor Crowley's (1976) version bears this out, both showing large areas of serpentinite and olivine-bearing rock throughout the complex. Deformation of a layered complex that contained repetitive cycles of ultramafic and mafic layers, coupled with metamorphism and serpentinization, has resulted in a structural tangle which has yet to be unraveled.

Crowley's (1976) mapping shows that the Laurel and Bel Air belts of gabbro are not physically continuous, although they have been supposed to be structurally and stratigraphically equivalent. South of Little Gunpowder Falls, the mafic rocks of the Bel Air belt change along strike from gabbro to highly metamorphosed volcanic rocks. The exact boundary has not been mapped but is tentatively shown in Plate II. The large mass of metagabbro at Aberdeen is more problematic; it may be an entirely separate structural block.

Selected chemical analyses of gabbro are given in Table 1. The gabbro is characterized by very low $Na_2O$ and $K_2O$, low $SiO_2$, and high $Al_2O_3$ and CaO.

Data on the mineralogy of the gabbro have been summarized by Herz (1951), Hopson (1964), and Southwick (1970) using older chemical analyses, as well as optical and X-ray methods of investigation. Hanan (1976) has added electron microprobe studies of minerals from the Susquehanna River and from several localities in the Laurel belt.

Orthopyroxene ranges from En81 to En64 and clinopyroxene from $En_{46}Fs_9Wo_{45}$ to $En_{40}Fs_{16}Wo_{44}$ (Hanan, 1976). A large miscibility gap between pyroxenes suggests extensive equilibration down to temperatures of about 600° to 750°C and may indicate a period of granulite facies of metamorphism. If so, this metamorphism must have taken place prior to the emplacement of the gabbro in its present environment where the surrounding rocks range from greenschist-facies metamorphism (Susquehanna River section of the Bel Air belt) to upper amphibolite facies metamorphism (Laurel belt).

Southwick (1970), using optical methods, has made a study of orthopyroxene compositions in the Bel Air belt within Harford County. He reported a decrease in magnesium in orthopyroxene in gabbro going from the base to the top of the section, which suggests that the Bel Air belt has cryptic zoning. His range in composition of orthopyroxene is En 72 to En 47. He was able to map a boundary in the upper part of the Bel Air belt above which all orthopyroxene has an En of less than 60 percent. M.W. Higgins (unpub. data, 1977) has been able to extend this boundary across Cecil County, so that cryptic zoning may be characteristic of all of the Bel Air belt.

Plagioclase is very calcic, and much unaltered plagioclase is bytownite. Hanan (1976) reported that unaltered samples range from An 80 to An 90 in the Susquehanna River section. Hopson (1964) reported that plagioclase from the Laurel belt from least altered gabbro most commonly ranges from An 75 to An 80. The highly calcic nature of the plagioclase is a dominant factor in the whole rock chemistry, which is reflected by the low $Na_2O$, $K_2O$, $SiO_2$, and high $Al_2O_3$ content.

Much has been written about the alteration of the gabbro and the process of uralitization; the papers of Williams are a classic in this regard (1884, 1886). Readers are referred to his papers as well as the review by Hopson (1964) for a discussion of the alteration of the gabbro to amphibolite, the replacement of pyroxene by green amphibole, the problems of differentiating between relict igneous amphibole and metamorphic amphibole, and the alteration of plagioclase to clinozoisite.

Isotopic studies of $Sr^{87}/Sr^{86}$ ratios in gabbro have been made by Hanan (1976). He reported an average value of 0.7097 for least altered gabbro and 0.7100 for uralitized gabbro. He related the high $Sr^{87}/Sr^{86}$ in Baltimore gabbro to metamorphic alteration and to partial exchange of Sr and Rb with the Wissahickon rocks which have values of $Sr^{87}/Sr^{86}$ in excess of 0.7291. However, the lack of correlation of $Sr^{87}/Sr^{86}$ with grade of regional metamorphism and the small range of variation either in the altered samples or in the least altered samples all suggests that exchange with sea water (0.7093 ± 0.0007; Cox and Faure, 1974) may have been a contributing factor producing rather uniform $Sr^{87}/Sr^{86}$ values.

## Diorite

Within the Bel Air belt, the gabbro is bounded on its southeast side by a less mafic rock mapped by Southwick (1969) as quartz gabbro-diorite. These rocks are transitional with the gabbro, and the mapped contacts are only approximate. The diorite is rarely more than 2 km wide and it extends for approximately 25 km from Rising Sun in Cecil County to Thomas Run in Harford County. The rocks have been described by Leonard (1901) and Southwick (1970), and mapped in Cecil County by Higgins (unpub. data, 1977). The primary mineralogy consists of strongly zoned plagioclase, hornblende, augite, a variable amount of quartz, and small amounts of biotite, magnetite, and hypersthene. The strongly zoned plagioclase is notable; Southwick (1970) stated that the cores are bytownite ($An_{85}$) and the rims, andesine (An 45-60). The crystallization sequence appears to be cumulus plagioclase, magnetite, and pyroxene, with a postcumulus filling of hornblende, biotite, and quartz. Most of the rocks have been metamorphosed and the primary minerals have been partly replaced by chlorite, a pale amphibole, and epidote. Two representative analyses are given in Table 1.

The entire assemblage of rocks including dunite, pyroxenite, gabbro, and diorite, can be considered as a cumulus series derived from a single magma. However, an accurate model for the crystallization history of the complex cannot be prepared until a careful study has been completed. The general sequence of crystallization given in Table 2 is based on older reports and illustrations (chiefly Williams, 1886, 1890; Leonard, 1901; Herz, 1951), Hopson's review (1964) of the petrology of the Baltimore gabbro, and studies by the author.

The general characteristics of the crystallization scheme are the early disappearance of olivine and chromite before the appearance of cumulus plagioclase. Consequently, olivine gabbro (usually having partly resorbed olivine) is not abundant, and troctolite is absent, whereas a wide range of rocks have lherzolite and pyroxenite compositions. Clinopyroxene is late in appearing as a cumulus phase, and olivine-clinopyroxene rocks are not abundant; wehrlite is described in only one locality and clinopyroxenite, not at all. The appearance of more sodic cumulus plagioclase and postcumulus hornblende marks the development of the more differentiated quartz gabbro and diorite (Southwick, 1970). The sequence of crystallization differs markedly from that described by Jackson and others (1975) in the Vourinos ophiolite, Greece, where the sequence includes early olivine plus clinopyroxene and correspondingly abundant "rock types" of wehrlite and clinopyroxenite. However, the cumulus sequence olivine-orthopyroxene-clinopyroxene has been described for parts of the Muskox complex by Irvine and Smith (1967) and can be modeled by the "system" clinopyroxene-olivine-silica having magnesium end-member compositions determined experimentally by Kushiro and Schairer (1963).

## Volcanic rocks and plagiogranites

Volcanic rocks within the area of the Baltimore Complex have long been recognized in Cecil County (Bascom, 1902; Marshall, 1937). More recent work in Cecil County (Higgins, 1971, 1972), Harford County, (Southwick, 1969), and Baltimore County (Crowley, 1976) has extended the areas known to be underlain by volcanic rocks and has resulted in revision of the older formational names of gneiss and diorite, names that were applied to the rocks when little evidence was available on the origin of the rocks.

Table 2 -- Generalized crystallization sequence for the Baltimore Complex

(↓ denotes cumulate phase and → means post cumulus crystallization. ol-olivine, chrom-chromite, opx-orthopyroxene, cpx-clinopyroxene, pl-plagioclase, hbd-hornblende, mt-magnetite, bi-biotite, and q-quartz)

| Map Unit | Rock Type | Cumulate Type | Minerals |
|---|---|---|---|
| Serpentinized peridotite of the State Line district and (for pyroxenite) small areas within Baltimore Gabbro of the Laurel belt | dunite chromitite | olivine-chromite cumulate | ol↓ + chrom↓ → ol |
| | lherzolite | olivine-orthopyroxene cumulate | ol↓ + opx↓ → cpx |
| | pyroxenite (often websterite rarely enstatolite) | orthopyroxene cumulate | opx↓ → cpx<br>opx↓ → opx |
| | | orthopyroxene clinopyroxene cumulate | opx↓ + cpx↓ → cpx |
| Baltimore gabbro in the Bel Air and Laurel belts | feldspathic pyroxenite | orthopyroxene-clinopyroxene cumulate | opx↓ + cpx↓ → pl<br>opx↓ + cpx↓ → cpx + pl |
| | hypersthene gabbro norite | cumulate texture poorly defined | opx↓ + cpx↓ + pl↓ → cpx + pl + hbd |
| Dark diorite and quartz gabbro of Bel Air belt | diorite | cumulate texture poorly defined | cpx↓ + pl↓ + mt↓ → hbd + bi + q |

The James Run Gneiss in Harford County, defined by Southwick and Fisher (1967) and described by Southwick (1969), is a layered rock composed of beds that range from mafic amphibolite (containing about 54 wt. percent $SiO_2$) to a quartz-plagioclase rock (containing about 76 wt. percent $SiO_2$). The James Run contains no relict textures and all rocks are thoroughly recrystallized. Plagioclase has a wide range in composition; $An_{25}$ is common for felsic rocks and $An_{40}$ for mafic rocks (Southwick, 1969, p. 47). Chemical analyses of these rocks show a high $Na_2O/K_2O$ ratio for all samples regardless of other compositional variations.

Higgins (1971, 1972) has expanded the James Run to include all the closely associated, approximately contemporaneous metavolcanic and metavolcaniclastic rocks in the northeastern Maryland Piedmont. His extended definition of the James Run Gneiss, now the James Run Formation, is now added to the Glenarm Group and includes the unit informally called the Cecil County volcanic rocks described by Bascom (1902) and Marshall (1937). Within Cecil County, Higgins (1972) has divided the volcanic rocks into four informal units, which include metamorphosed rhyolite, dacitic tuff, andesitic tuff, and metamorphosed pillow basalt. Many of the rocks are thin to medium bedded and contain intercalated fine-grained amphibolite. In Cecil County, surviving relict textures consist of epidote-filled amygdules and plagioclase phenocrysts. Higgins (1971) has described well-preserved pillow structures in basalt in Cecil County and has concluded that the rocks were extruded into a relatively shallow-water environment. The Cecil County part of the James Run Formation shows the same compositional variation as the James Run Gneiss described by Southwick in Harford County. Mafic rocks have an average value of about 51 wt. percent $SiO_2$ and rhyolitic tuffs about 75 wt. percent $SiO_2$. The large $Na_2O/K_2O$ ratios suggested to Southwick (1969) that all of these rocks have undergone metasomatic alteration either before or during metamorphism. Representative analyses of the James Run Formation are given in Table 1.

The James Run rocks are not in structural contact with the mass of the Bel Air gabbro in Harford and Cecil Counties, and the volcanic rocks are separated from plutonic rocks by a wide belt of diamictite of the Wissahickon Formation (Southwick, 1969; Higgins, unpub. data, 1977) and by later intrusions of the Port Deposit Granodiorite. The volcanic rocks are in contact with Aberdeen metagabbro and the James Run is separated from the Cecil County volcanic rocks by this gabbro. The structural relationship of the James Run to the Aberdeen metagabbro is not clear. Southwick's large-scale map (1969, pl. 4) of a critical area on the west side of the metagabbro suggests that the James Run is an overturned syncline flanking the gabbro and structurally above it. However, Crowley (1976, pl. 2) indicated that this contact may be a fault. The Cecil County volcanic rocks strike directly into the metagabbro; Southwick has interpreted the contact as intrusive, the gabbro being a later intrusion. The isolated patches of Cecil County volcanic rocks in the gabbro are roof pendants according to Southwick (1969). However, these rocks may be interpreted as having been unconformably deposited above the Aberdeen metagabbro and subsequently infolded. They now are outliers from the main mass of volcanic rocks.

The more southerly Laurel belt of gabbro in Baltimore City and County is bounded on the southeast side by leucocratic plagioclase-quartz gneiss having a high $Na_2O/K_2O$ ratio. This gneiss is intercalated with amphibolite. Knopf and Jonas (1929) considered that these rocks were plutonic and named them the Relay Quartz Diorite with the type locality at Relay on the Patapsco River. Hopson (1964) considered the

Relay Quartz Diorite to be a differentiation product of the Baltimore Gabbro. Higgins (1972) reinvestigated the Patapsco River section and stated that the Relay is strikingly similar to part of the James Run Formation, especially some of the more siliceous rocks. He did note that other parts of the formation more nearly resemble a plutonic quartz diorite. A contact of the Relay rocks with the gabbro is "nearly impossible to pick" (Higgins, 1972, p. 1006). Crowley (1976) considered the Relay rocks in Baltimore County to be a member of the James Run Formation. Part of that formation as mapped by Crowley probably does represent true plutonic diorite associated with the Bel Air mass.

Much of the Laurel belt of gabbro is intruded by quartz-plagioclase (plagiogranite) dikes which branch and anastomose, producing a pattern of net veining (Thayer, 1963). Hopson (1964) has suggested that these dikelets along with the Relay Quartz Diorite are the last differentiation product of the gabbro. However, the similarity in composition between plagiogranite dikes and leucocratic metavolcanic rocks (see table 1) suggests that the net veins were derived from the James Run volcanic rocks and are plutonic equivalents of them. Further evidence is given by the distribution of net veins. They are common in the Laurel belt of gabbro near the contact with the James Run volcanic rocks. In the Bel Air belt, the latest preserved differentiation products are the quartz gabbro and diorite; net veins are not described in these rocks.

It is difficult at this writing to make a quantitative assessment of the abundance and composition of the volcanic rocks and associated leucocratic dikes that form a part of the Baltimore Complex, because of the relatively recent remapping of many areas now considered to be underlain by volcanic rocks, and because of the lack of systematic published chemical data. Descriptions of the James Run Gneiss of Southwick (1969) show that the mafic and siliceous rocks are intimately associated. Elsewhere, in Baltimore and Cecil Counties, the more silicic rocks dominate. The silicic rocks show an affinity with plagiogranite of oceanic origin (Coleman and Peterman, 1975) rather than with granophyre (as suggested by Hopson, 1964) or with a basalt-rhyolite association (as suggested by Southwick, 1969). This affinity is suggested by the high silica, moderate alumina, and low iron, magnesia, and potash contents of these rocks. Further work on $Sr^{87}/Sr^{86}$ ratios might help to define the kinship of these rocks, but as Hanan (1976) has shown from analyses of two leucocratic rocks from the Laurel belt, regional metamorphism has apparently altered the isotopic ratios.

Leucocratic rocks are widely associated with ophiolites (see Coleman and Peterman, 1975, for a review of occurrences), and in a few sequences these rocks may exceed basaltic extrusive rocks in abundance (Bailey and others, 1970), as they apparently do in the Baltimore Complex.

A crustal section for the Baltimore Complex can be restored from data from the Bel Air belt, from structural data in the Susquehanna River section, and from drill data at the Woods mine. The section can be assembled in terms of an ophiolite succession. From the top of the succession down the thicknesses are as follows: volcanic rocks, 3 km (from M.W. Higgins, oral communication, 1977); diorite, 1.4 km; gabbro, 3.5 km; pyroxenite 0.4 km; and serpentinized dunite, 1.5 km exposed and terminated by a thrust fault. The crustal section is comparable with those of other ophiolite complexes as compiled by Moores and Jackson (1974) and Jackson and others (1975), except that the thickness of volcanic rocks in the Baltimore Complex is considerably greater that that for most documented ophiolites.

## EMPLACEMENT AND AGE

Almost all earlier reports considered the Baltimore Gabbro to be intrusive into the Glenarm metasedimentary rocks of the Maryland Piedmont (Williams, 1886; Leonard, 1901; Herz, 1951; Hopson, 1964; Higgins, 1972). Only the more recent investigations have attempted to look at the Baltimore Complex as a whole (Southwick, 1969, 1970; Crowley, 1976) and to discuss problems relating to the emplacement of the entire mass into the Glenarm rocks. Crowley has rather convincingly argued that the entire complex must be allochthonous. His arguments rest chiefly on the observation the the ultramafic-gabbro-volcanic sequence is found only in the Wissahickon Formation and that no gabbro dikes, sills, or other evidence of intrusion through the underlying rocks can be seen. His detailed mapping in Baltimore County shows that the Baltimore Complex rests on a heterogeneous assortment of metasedimentary and metavolcaniclastic rocks; hence, it cannot have been intruded as a conformable sill into the Wissahickon. Finally, Crowley has demonstrated that the internal contacts within the Laurel belt of gabbro are truncated by the basal contact of the mass, indicating a fault.

Within the Bel Air belt, the complex appears to be an intercalated packet of plutonic rocks within Wissahickon diamictite. If the plutonic rocks are fault bounded, then constraints on the geometry of the body indicate that the fault must be a low-angle (thrust) fault rather than a high angle fault.

In addition to the probability that the Baltimore Complex is allochthonous on the Wissahickon formation, it also seems probable that much of the Wissahickon and all of the James Run formation rocks may also be allochthonous (A.A. Drake, oral communication, 1977). The present outcrop distribution of members of the complex shown in Plate II may have all been part of a coherent ophiolite now broken into a series of structural blocks. However, correlation of all parts of the complex into a single coherent mass is not possible with the available data. A detailed study of the distribution of gabbro and serpentinite clasts in the diamictite should be undertaken in order to determine the extent and nature of the internal correlations within the complex. It should be emphasized that the present position of the Baltimore Complex does not mark a suture between colliding crustal plates. Instead, the complex has been transported to the west within the diamictite, and in the process, it has been dismembered and has foundered within the broken terrane of the upper Wissahickon Formation. Plate II shows the western limit of Wissahickon rocks containing tectonic blocks of serpentinite and metagabbro; this line whould approximate the western limit of allochthonous sheets containing the ophiolite.

The age of emplacement of the Baltimore Complex must be that of the deposition or emplacement of the Wissahickon Formation, inasmuch as detached blocks of the serpentinized base of the complex are found distributed throughout the formation (Crowley, 1976). The Wissahickon Formation is intruded by granodiorite plutons; $U^{238}/Pb^{206}$ ages from the Port Deposit, Ellicott City, and Occoquan plutons (see Plate II) are 322, 354, and 557 m.y. respectively. A list of all U-Pb zircon dates from this area has been given in Higgins and others (1977, table 1). The U-Pb zircon dates are all discordant, and their use has led to a considerable controversy over the age of the Glenarm (Higgins, 1972, 1976; Higgins and others, 1977; Seiders and others, 1975, 1976). The argument mainly revolves around the nature of plutonic rocks that may or may not intrude Glenarm rocks, the correlation of the Glenarm with other units, and the interpretation of the origin of the discordant ages of the zircons. Higgins and others (1977) have minimized the evidence from radiometric ages on the basis of the wide scatter of ages and have suggested a possible contamination of the zircons by older material derived from the continental basement. At this writing, the age of the Wissahickon rocks together with the enclosed Baltimore Complex has not been resolved, and the range of ages for the Wissahickon must be between Late Precambrian and Ordovician.

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# PRELIMINARY REPORT ON THE OPHIOLITES OF NORTHERN AND WESTERN ALASKA

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## ABSTRACT

In northern and western Alaska ophiolites have been mapped in three separate belts, herein named the western Brooks Range, Yukon-Koyukuk, and Rampart ophiolite belts. The ophiolite bodies in the western Brooks Range and Rampart belts appear to be broad synclinal remnants of allochthonous sheets. The narrow V-shaped Yukon-Koyukuk ophiolite belt, which is envisaged as the root zone for these allochthonous sheets, is composed of slablike ophiolite bodies that dip inward beneath the Cretaceous rocks of the Yukon-Koyukuk tectonic province. The Yukon-Koyukuk belt is bordered in the southern Brooks Range and on the Ruby geanticline by Precambrian(?) and Paleozoic metamorphic complexes which are chiefly of greenschist metamorphic facies but which locally contain "blueschist" mineral assemblages. The presence of small ophiolite klippen on the metamorphic complexes and the abundance of ophiolite debris in Cretaceous conglomerates along the borders of the Yukon-Koyukuk province support the view that these metamorphic complexes were once covered by ophiolite sheets.

The ophiolite assemblages in all three belts can be divided into two separate and distinct tectonic units: a lower unit composed of basalt, diabase, gabbro and chert, and an upper unit composed of ultramafic rocks and layered gabbro. The contact between the two units is generally interpreted as a thrust fault and locally is marked by a thin layer of garnet amphibolite. Potassium-argon determinations for the ultramafic-layered gabbro unit provide Middle and Late Jurassic ages, which are interpreted as reflecting the tectonic event involved in the emplacement of this unit onto continental crust. Fossil collections from the basalt-diabase-chert units in the Yukon-Koyukuk and Rampart belts are late Paleozoic in age. The basalt-diabase-chert unit in the western Brooks Range belt is not dated more closely than Devonian to Early Cretaceous.

## INTRODUCTION

Ophiolites are widely exposed in northern and western Alaska along three separate but probably genetically related belts: western Brooks Range, Yukon-Koyukuk, and Rampart (Plate III, in pocket). Although these three belts provide the most complete and extensive exposures of ophiolite assemblages in Alaska, they have received surprisingly little attention in the literature. The ophiolitelike character of the western Brooks Range belt was recognized first by Tailleur (1973a) and recently has been described more fully by Roeder and Mull (in press). In the Yukon-Koyukuk tectonic province, the presence of ophiolites has been mentioned by Patton and Miller (1970) and by Patton (1973), but few details have been published about their geologic setting and tectonic significance. The ophiolitic affinities of the rocks in the Rampart belt, so far as is known, have not been recognized previously in the literature.

This report briefly summarizes available information on the geologic features of the ophiolite assemblages and specifically documents previously unpublished isotope dating and fossil collections bearing upon their age. It should be emphasized that all of this information is based upon reconnaissance-scale mapping and study. At present only the broad features of these ophiolites are known, and much remains to be learned, particularly about their internal structure and petrography.

All three belts contain the major elements of a typical ophiolite assemblage, including serpentinized peridotite and dunite, layered gabbro, pillow basalt, diabase, and radiolarian chert. Possible sheeted dike complexes have been described by Roeder and Mull (in press) in the western Brooks Range belt but have not been recognized in the other two belts. All of the ophiolite sequences are tectonically dismembered and disturbed.

A particularly noteworthy feature of the ophiolites in all three belts is the relation between the ultramafic-layered gabbro complexes and the basalt-diabase-chert complexes. In each of the belts, these two complexes form discrete tectonic units that appear to have been juxtaposed by thrust faulting. Where the structure can be deciphered, the stacking order is invariably the reverse of that found in a normal ophiolite succession; i.e., the ultramafic-layered gabbro complexes structurally overlie the basalt-diabase-chert complexes. This reversed stacking order has been recognized in the western Brooks Range belt by Tailleur (1973a) and by Roeder and Mull (in press), in the Yukon-Koyukuk belt by Patton and Miller (1970), and in the Rampart belt by Brosgé and others (1974).

The similarities of these three ophiolite belts in structural style and in the apparent ages provided by K/Ar dating strongly suggest that they had a common orogenic history. This fact alone places severe constraints on any model that may be devised to explain the present tectonic framework of northern and western Alaska.

## REGIONAL GEOLOGIC SETTING OF THE OPHIOLITE BELTS

The ophiolite belts are situated in three separate and distinctive tectonic provinces: Brooks Range, Ruby geanticline, and Yukon-Koyukuk (Pl. III) The Brooks Range province is a broad Cretaceous orogen that stretches across northern Alaska from the Yukon border to the Chukchi Sea. The northern part of this orogen is a northward vergent imbricate thrust complex of Paleozoic miogeoclinal deposits. The southern part is a metamorphic core complex of Precambrian(?) and Paleozoic metasedimentary rocks that is intruded by several large granitic plutons of Cretaceous age. The Ruby geanticline province (Miller and others, 1959) is a narrow Cretaceous uplift that extends from the south edge of the Brooks Range southwestward across central Alaska to the Yukon-Kuskokwim Delta. The axis of the uplift is composed of metasedimentary rocks of Precambrian(?) and Paleozoic age that are widely intruded by Cretaceous granitic plutons. Flanking the uplift on the southeast are severely deformed but unmetamorphosed Paleozoic miogeoclinal deposits. The Yukon-Koyukuk province is a wedge-shaped Cretaceous depression filled with volcanic rocks and volcanogenic sedimentary rocks of Early and early Late Cretaceous age and intruded by granitic rocks of middle and Late Cretaceous age. Along the perimeters of the province are thick marginal conglomerates derived from the bordering ophiolites and metamorphic core complexes.

The Kaltag fault, a major northeast-trending strike-slip fault, appears to offset the southeastern margin of the Yukon-Koyukuk province and the Ruby geanticline province 130 to 150 km right laterally (Pl. III).

## OPHIOLITE BELTS

### Western Brooks Range Belt

*Distribution*. The western Brooks Range ophiolites extend in a broad belt from the Chukchi Sea to Howard Pass, a distance of about 350 km (Pl. III). Few map details of these ophiolites have been published, and the generalized outlines of the bodies shown in Plate III are taken from a regional-scale map compiled by Beikman and Lathram (1976). The ophiolites are distributed in five separate masses, the largest of which covers about 3,000 $km^2$. Four large ultramafic-layered gabbro complexes, the Siniktanneyak, Misheguk, Avan Hills and Asik bodies, make up about 25 percent of the ophiolite exposures.

*Geologic Setting and Lithologic Character*. All recent workers in the western Brooks Range agree that the ophiolite assemblages represent synclinal remnants of a northward vergent allochthonous sheet that had its roots along the northern margin of the Yukon-Koyukuk province (Snelson and Tailleur, 1968; Martin, 1970; Tailleur, 1973b; Roeder and Mull, in press). The ophiolite allochthon lies at the top of an imbricated stack of Devonian to Lower Cretaceous strata in which as many as six separate allochthonous sheets can be recognized. Total foreshortening as a result of thrusting is believed to be at least 250 km (Tailleur and Brosgé, 1970). Reconnaissance gravity and aeromagnetic data in the western Brooks Range generally support the view that the ophiolites are rootless (Barnes, 1976; Tailleur and others, 1967).

The ophiolite allochthon typically consists of a lower unit of basalt, diabase, and intercalated chert and an upper unit of ultramafic rocks and layered gabbro. The contact between the lower and the upper units is interpreted by Roeder and Mull (in press) as a thrust fault. The base of the allochthon commonly contains slices of Devonian carbonate rocks, most of which appear to be tectonic blocks but some of which show intrusive and extrusive contacts with basalt and diabase. The ultramafic rocks are composed chiefly of serpentinized peridotite and dunite that, at least locally, display a well-developed tectonite fabric (Tailleur, 1973a). The layered gabbro overlies the ultramafic rocks and in places is intruded by basalt dike swarms that have been interpreted by Roeder and Mull (in press) as possible sheeted dike complexes.

*Age Data*. Three K/Ar ages have been obtained from hornblende-bearing rocks in the ultramafic-layered gabbro complexes of the western Brooks Range (Pl. III, Table 1, nos. 1-3). Two of the age determinations yielded Jurassic ages (151 ± 15, 164 ± 7.2 m.y.) that agree closely with ages obtained from ultramafic-layered gabbro complexes in the Yukon-Koyukuk and Rampart belts. One of the samples (no. 2) is from a hornblende pegmatite dike intruding gabbro in the Siniktanneyak body, and the other (no. 3) is from a hornblende gabbro in the Misheguk body. The third determination (no. 1), obtained from an amphibolite which appears to be part of the Asik ophiolite body, gave a middle Paleozoic age (384 ± 55 m.y.). This age is considered anomalous and cannot be reconciled with the other two determinations or, for that matter, with any K/Ar ages obtained thus far from the ophiolites of northern and western Alaska. There is no evidence, stratigraphic or otherwise, to suggest that any of the ophiolite assemblage is older than Late Devonian.

The basalt-diabase-chert complexes in the western Brooks Range allochthon cannot be dated more precisely than Late Devonian to Early Cretaceous. The older limit is established by the extrusive and intrusive relations between the basalt and slices of Upper Devonian carbonate rocks that are incorporated in the base of the ophiolite allochthon. The younger limit is fixed by the abundant occurrence of basalt, diabase, and chert debris in Early Cretaceous flyschoid deposits in the Colville basin foredeep at the north edge of the Brooks Range (Tailleur and Brosgé, 1970; Roeder and Mull, in press).

### Yukon-Koyukuk Belt

*Distribution*. The wedge-shaped Yukon-Koyukuk province of west-central Alaska is bounded on at least two sides by ophiolites that dip inward beneath the Cretaceous volcanic and sedimentary rocks of the province (Pl. III). The ophiolites have been traced in a narrow but nearly continuous belt along the northern margin of the province for nearly 500 km and along the southeastern margin for another 400 km (Patton, 1973). The Kaltag fault on the Yukon River offsets the southeastern margin of the province and the ophiolite belt 130 to 150 km right laterally (Patton and Hoare, 1968). South of the fault, the Yukon-Koyukuk ophiolite belt occurs in an isolated group of hills situated west of the Kaiyuh Mountains near the Yukon River (Pl. III).

TABLE 1. *K-Ar AGES AND ANALYTICAL DATA FOR WESTERN BROOKS RANGE, YUKON-KOYUKUK AND RAMPART OPHIOLITE BELTS*

| Map no. (pl. III) | Field no. | Lat. (N) | Long. (W) | Rock type | Mineral | $K_2O$* (wt. percent) | $^{40}Ar_{rad}$ (mol/gm) $\times 10^{10}$ | % $^{40}Ar_{rad}$ | Calculated age† (millions of years) |
|---|---|---|---|---|---|---|---|---|---|
| 1 | 66ATr 56A | 67°24' | 162°23' | Amphibolite | Hornblende | 0.0588 ± 0.0019 (6) | 0.3626 | 57.0 | 384 ± 55 |
| 2 | 66ATr 76.2 | 68°20' | 158°15' | Hornblende pegmatite | --do-- | .131 (1) | .2976 | 42.0 | 151 ± 15 |
| 3 | 66ATr 152C | 68°17' | 160°32' | Gabbro | --do-- | .307 (2) | .7577 | 50.0 | 164 ± 7.2 |
| 4 | 73APa 242A | 65°58' | 151°53' | Garnet amphibolite | --do-- | .681 (2) | 1.651 | 89.6 | 161 ± 4.9 |
| 5 | 70APa 271 | 66°01' | 151°42' | Hornblende pegmatite | --do-- | .117 (2) | .2420 | 20.8 | 138 ± 8.0 |
| 6 | 73APa 254A | 65°56' | 152°07' | Hornblendite | --do-- | .054 (2) | .1218 | 44.0 | 149 ± 9.6 |
| 7 | 73APa 246A | 66°56' | 150°48' | Gabbro | --do-- | .479 (2) | 1.139 | 84.0 | 158 ± 4.6 |
| 8 | 71ABe 494Z | 66°52' | 148°28' | Amphibolite | --do-- | .192 (2) | .4479 | 71.2 | 155 ± 4.6 |
| 9 | 63ABe 202 | 67°30' | 145°45' | Diorite | Plagioclase | 1.12 (2) | 2.677 | 69.0 | 159 ± 6 |
| 10 | 63ARr 151 | 67°28' | 145°31' | Gabbro | Hornblende | .251 ± 0.0004 (4) | .6503 | 51.0 | 172 ± 8 |
| 11 | 65ABe 106CZ | 65°39' | 150°09' | --do-- | --do-- | .174 (2) | .5592 | 47.0 | 210 ± 6 |

Nos. 1-8, not previously published

Nos. 9, 10, Reiser and others, 1965

No. 11, Brosgé and others, 1969

*Mean value and, where more than two measurements were made, standard deviation. Number of measurements is in parentheses.

†$\lambda_\varepsilon = 0.572 \times 10^{-10}$ yr$^{-1}$; $\lambda_\varepsilon' = 8.78 \times 10^{-13}$ yr$^{-1}$; $\lambda_\beta = 4.963 \times 10^{-10}$ yr$^{-1}$; $^{40}K/K_{total} = 1.167 \times 10^{-4}$. Where more than one measurement was made on a sample, the age given is the weighted mean; weighting was by the inverse of the variance. The ± figures are estimates of analytical precision at the 68 percent confidence level. Previously published ages recalculated with these decay constants.

Narrow fault-bounded slivers of basalt and diabase also occur along the western boundary of the Yukon-Koyukuk tectonic province on the eastern Seward Peninsula and along the south shore of Escholtz Bay. However, the relation of these mafic rocks to the Cretaceous rocks of Yukon-Koyukuk province is obscure, and it is problematical whether they are part of the Yukon-Koyukuk ophiolite belt. Interpretation of the structural relations along the western margin of the Yukon-Koyukuk province is complicated by a north-trending belt of Late Cretaceous to early Tertiary folds and thrusts that has cut across and deflected the west and southwest middle Cretaceous trends of the province (Patton and Tailleur, in press).

*Geologic Setting and Lithologic Character*. The Yukon-Koyukuk ophiolite bodies are structurally complicated in detail but in gross aspect are interpreted as slablike bodies that dip 10° to 80° inward beneath the Cretaceous rocks of the Yukon-Koyukuk province. Computer modeling of gravity anomalies along the ophiolite belt appears to support this interpretation (Barnes, 1970).

The ophiolites are structurally underlain by a Paleozoic and Precambrian(?) metamorphic complex composed chiefly of pelitic schist, quartzite, and carbonate rocks of greenschist and locally amphibolite metamorphic facies. The contact between the ophiolites and the metamorphic complex typically is characterized by a broad zone in which the ophiolites and metasedimentary rocks are tectonically shuffled together and by a steep metamorphic gradient from little altered ophiolite through slate and phyllite grade rocks to schist and well-foliated greenstones.

The ophiolites are overlain nearly everywhere along the borders of the Yukon-Koyukuk province by coarse Cretaceous conglomerates, as much as 2,500 m thick. The lower part of the conglomerate sequence is made up of Lower Cretaceous marine flyschoid deposits composed largely of ophiolite debris, and the upper part is made up of Upper Cretaceous molassoid deposits derived from the metamorphic complex.

Glaucophane-bearing rocks have been found in the metamorphic complexes bordering the Yukon-Koyukuk province in the southern Brooks Range, Ruby geanticline, and Seward Peninsula (Forbes and others, 1971; Brosgé, 1975; Patton, 1975) (Pl. III). The glaucophane occurs both in the metasedimentary rocks of the metamorphic core complexes and in greenstone bodies along the margins of the ophiolite belts. Potassium-argon age determinations by Forbes and others (1977) for these "blueschists" have yielded late Precambrian ages in the Brooks Range, Paleozoic and Jurassic or Cretaceous ages in the Kaiyuh Hills segment of the Ruby geanticline, and Jurassic ages on the Seward Peninsula. Forbes and others (1977) conclude that neither the Brooks Range nor the Kaiyuh Hills "blueschist" belts are tectonically related to ophiolites, a conclusion that seems questionable in view of the distribution of these glaucophane-bearing rocks around the perimeter of the Yukon-Koyukuk province (Pl. III).

The Yukon-Koyukuk ophiolite bodies, like the western Brooks Range ophiolites, are composed of two lithologically distinct tectonic units that have been juxtaposed by thrust faulting. The structurally lower of these units is composed of pillow basalt, diabase, and gabbro with minor chert, argillite, volcaniclastic rocks, and limestone. Pillow structures, where examined in detail, generally appear to be right-side up. The structurally higher unit is made up chiefly of serpentinized dunite and peridotite tectonite overlain by layered gabbro. In a few places a thin layer of strongly banded and foliated garnet amphibolite and pyroxene granulite occurs at the base of the upper unit.

These two units are best exposed in the Kanuti River area along the southeastern margin of the Yukon-Koyukuk province (Patton and Miller, 1970) (Fig. 1). There the lower unit is as much as 2 km thick and rests in probable thrust fault contact on metasedimentary rocks of the Ruby geanticline. The upper unit, which is as much as 1.5 km thick, is thrust onto the lower unit and also onto the metasedimentary rocks. Both units are severely deformed internally but have a gross regional dip of 10° to 60° to the northwest beneath the Cretaceous rock of the Yukon-Koyukuk province.

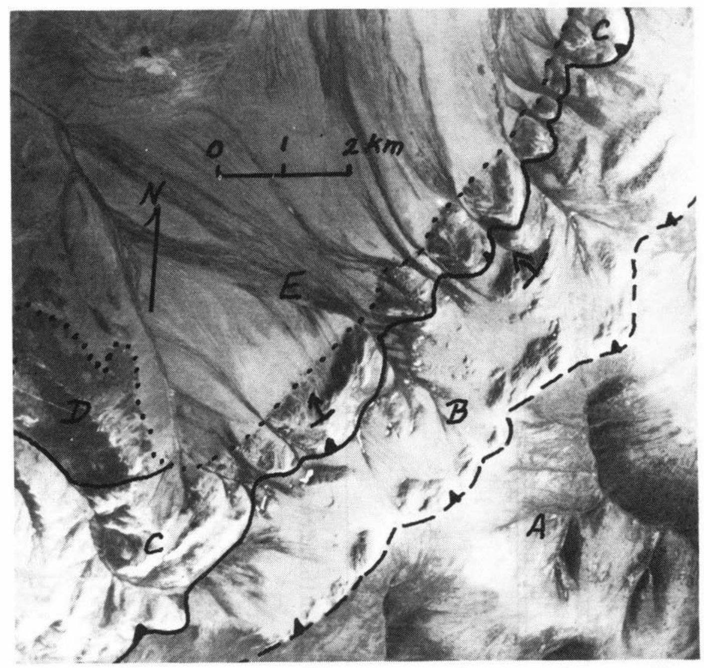


Figure 1. Aerial view of Yukon-Koyukuk ophiolite belt on southeastern margin of Yukon-Koyukuk province. Ophiolite belt (B,C,D) has regional dip to the northwest and rests in probable thrust-fault contact on metasedimentary rocks (A) of Precambrian(?) and Paleozoic age. Lower part of ophiolite assemblage (B) is a structural complex of basalt, diabase, gabbro, and argillite. Upper part is composed of serpentinized dunite and peridotite (C) and layered gabbro and peridotite (D). Upper part (C,D) is thrust faulted on lower part (B). Ophiolite belt is overlain unconformably by Cretaceous conglomerate and Quaternary surficial deposits in Kanuti River lowlands (E).

*Age Data*. Fossils of late Paleozoic age have been collected from the basalt-diabase-gabbro unit at three localities in the Yukon-Koyukuk belt (Pl. III, Table 2, nos. 1-3). All three collections are from thin carbonate lenses that appear to be in sequence with pillow basalt. In addition to these collections, tectonic and exotic blocks of fossiliferous Devonian carbonate rock have been found in a basalt-diabase-gabbro complex along the northern margin of the Yukon-Koyukuk province (Patton and others, 1968). Most of these carbonate blocks are fault bounded, but a few show intrusive and extrusive contacts with basalt and diabase and thereby clearly establish the age of the basalt-diabase-gabbro unit as post-Middle Devonian.

Four K/Ar hornblende ages ranging from 138 ± 8 to 161 ± 4.9 m.y. (Jurassic) have been obtained from ultramafic-layered gabbro complexes in the Yukon-Koyukuk belt (Pl. III, Table 1, nos. 4-7). Three of the dated samples (nos. 5-7) are from hornblende gabbro and hornblende-bearing dikes in the layered gabbro sequence in the upper parts of the complexes. The fourth sample (no. 4) is from a thin layer of garnet amphibolite beneath the dunite-peridotite sequence in the lower part of the complex.

## Rampart Belt

*Distribution*. The Rampart ophiolite bodies occur as broadly synformal masses that are crudely alined in a belt that extends 800 km from the eastern Brooks Range to the Kaiyuh Mountains (Pl. III). Four large masses, Christian, Hadweenzic, Rampart, and Kaiyuh, as well as several smaller bodies have been mapped along this belt. Some of these masses may join beneath the alluviated lowlands that border the Yukon River.

The Rampart belt generally parallels the southwest-trending segment of the Yukon-Koyukuk ophiolite belt and is separated from it by the 30- to 50-km-wide metamorphic complex of the Ruby geanticline. The presence of several small klippenlike masses of ophiolite on the Ruby geanticline suggests that the two belts were once connected as a single continuous sheet.

*Geologic Setting and Lithologic Character*. Structural interpretations of the ophiolites in the Rampart belt are handicapped by a combination of poor exposures and lack of detailed map information. All the masses appear to be synclinal remnants and to rest on Devonian and older strata, except at the extreme northeastern end of the belt where they rest locally on Mississippian strata. The contact between the ophiolites and the underlying rocks, however, is largely obscured by alluvial deposits and a thick cover of vegetation. In the few places where seen, it generally has been interpreted as a fault.

The Rampart ophiolites appear to have the same internal structural order that is found in the Yukon-Koyukuk and Brooks Range belts; i.e., the ultramafic-layered gabbro complexes lie structurally above and in thrust fault contact on the basalt-diabase-chert complexes.

The Christian mass (Pl. III) at the northeastern end of the belt consists of a large oblong synform of nonlayered gabbro, basalt, diabase, chert, shale and carbonate rocks within which are found two small structurally higher(?) bodies of layered gabbro and ultramafic rocks (Reiser and others, 1965). The contact at the base of the layered gabbro and ultramafic bodies is poorly exposed but is believed to be a thrust fault. Similarly, the Hadweenzic mass is made up of a broad sheet of basalt, gabbro, and chert that is overlain on the northeast side by a small synclinal klippe of layered gabbro and peridotite. At the base of the klippe are banded garnet amphibolites similar to those which occur on the sole of the ultramafic thrust sheets in the Yukon-Koyukuk ophiolite belt (Brosgé and others, 1974). The Rampart mass is made up entirely of basalt, diabase, chert, and volcaniclastic rocks, and no large bodies of ultramafic rocks or layered gabbro have been found. In the Kaiyuh mass, at the southwest end of the belt, six separate bodies of serpentinized dunite and peridotite, as much as 600 m thick, are alined along a northeast-trending belt near the center of a synform of basalt, diabase, and chert. Structural relations at the base of the ultramafic bodies are obscured by a dense cover of vegetation, but scattered exposures of banded garnet amphibolite suggest that the contact is a thrust fault.

TABLE 2. *FOSSIL COLLECTIONS FROM YUKON-KOYUKUK AND RAMPART OPHIOLITE BELTS*

| Map no. (pl. III) | Field No. | Lat. (N) | Long. (W) | Fossils | Source of information | Age assignment |
|---|---|---|---|---|---|---|
| 1 | 70APa 236 | 67°05' | 152°29' | Foraminifera, bryozoans, brachiopods | Patton and Miller, 1973 | Permian |
| 2 | 75Tr 140 | 67°09' | 150°28' | Foraminifera | Bird, 1977 | Pennsylvanian(?) |
| 3 | 73APa 255 | 65°53' | 152°13' | Corals | A. K. Armstrong, written communication, 1973 | Mississippian(?) |
| 4 | 65ABe 95<br>65ABe 99 | 65°41'<br>65°41' | 150°01'<br>150°03' | Foraminifera, bryozoans, pelecypod prisms<br>Bryozoans | Brosgé and others, 1969 | Permian(?) |
| 5 | 75APa 118<br>75APa 120 | 63°51'<br>63°53' | 155°45'<br>155°35' | Radiolarians<br>Radiolarians | Brian Holdsworth, written communication, 1977 | Mississippian(?) |

*Age Data*. Fossils of probable Permian age were found in carbonate and volcaniclastic rocks interlayered with flows in the Rampart mass (Pl. III).Table 2, no. 4), and radiolarians of probable Mississippian age were recovered from cherts in a small body of ophiolitic rocks southeast of the Kaiyuh mass (Pl. III, Table 2, no. 5). In addition, a K/Ar age of 210 ± 6 m.y. (Triassic) was obtained from a gabbro intrusion in the Rampart mass (Pl. III, Table 1, no. 11), and a K/Ar age of 159 ± 6 m.y. (Jurassic) from a diorite intrusion in the Christian mass (Pl. III, Table 1, no. 9).

Two K/Ar age determinations were made for samples from the ultramafic-layered gabbro complexes: 172 ± 8 m.y. (Jurassic) from a layered gabbro in the Christian mass (Pl. III, Table 1, no. 10) and 155 ± 4.6 m.y. (Jurassic) from garnet amphibolite at the base of the small klippe of ultramafic rock and layered gabbro in the Hadweenzic mass (Pl. III, Table 1, no. 8).

## SUMMARY OF FOSSIL AND K/AR AGE DATA

### Basalt-Diabase-Chert Complexes

Fossil evidence suggests a late Paleozoic age for the basalt-diabase-chert complexes in the Yukon-Koyukuk and Rampart belts. The age assignments in four of the five collections listed in Table 2 are queried because they cannot be refined to a specific period. However, it is unlikely that any of the four collections are younger than late Paleozoic. An age older than Devonian is also ruled out, at least for the Yukon-Koyukuk complex, because of the presence of exotic blocks of fossiliferous Devonian carbonates in the basalt.

The age of the basalt-diabase-chert complexes in the western Brooks Range cannot be established at present more closely than Devonian to Early Cretaceous.

The two K/Ar dates of 159 ± 6 m.y. (Jurassic) and 210 ± 6 m.y. (Triassic) (Table 1, nos. 9, 11) from diorite and gabbro bodies in the basalt-diabase-chert complexes of the Rampart belt probably represent separate and later intrusive events.

### Ultramafic-Layered Gabbro Complexes

Of the nine samples (Table 1, nos. 1-8, 10) from the ultramafic-layered gabbro complexes that have been analyzed by K/Ar method, eight yielded ages of Middle and Late Jurassic (172 ± 8 to 138 ± 8 m.y.). We interpret these Jurassic ages as reflecting the tectonic events involved when the complexes were emplaced onto continental crust rather than the time that they were formed in the upper mantle and lower oceanic crust. This interpretation seems to be supported by the fact that two of the ages (nos. 4 and 8) were derived from strongly tectonized amphibolite on the sole of the ultramafic thrust sheets. Also, the Jurassic ages fit comfortably into the tectonic history of northern and western Alaska. All evidence in this region points to a major reorganization of the tectonic framework beginning about the Middle Jurassic and culminating in the Early Cretaceous with the rise of the ancestral Brooks Range (Patton and Tailleur, 1964; Tailleur and Brosgé, 1970; Martin, 1970). Emplacement of the ophiolites appears to have been completed by the end of Early Cretaceous. This time limit is indicated by the abundance of ophiolite debris in upper Lower Cretaceous conglomerates bordering the Brooks Range and Ruby geanticline and by upper Lower Cretaceous granitic plutons on the Ruby geanticline that cut across the contacts of the ophiolite and the underlying metamorphic complex (Patton and Miller, 1973; Patton and others, 1977).

## ORIGIN OF THE OPHIOLITES

We interpret the western Brooks Range and Rampart ophiolite belts as remnants of the leading edges of allochthonous sheets, and the Yukon-Koyukuk ophiolite belt as the root zones of these sheets. The klippen-like bodies of the ophiolites on the Ruby geanticline and the abundance of ophiolite debris in the marginal conglomerates of the Brooks Range and Ruby geanticline support the notion that the metamorphic cores of the southern Brooks Range and Ruby geanticline were once covered by the ophiolites.

Two different tectonic schemes are offered to explain the origin of the Yukon-Koyukuk tectonic province and the emplacement of the ophiolites. The first, suggested by Patton, proposes that the wedge-shaped Yukon-Koyukuk province began as an intracratonic rift that opened in late Paleozoic time and by Jurassic time had widened to an ocean basin at least 600 km wide. From Middle Jurassic to Early Cretaceous time, the rifted basin was subjected to strong compression, which resulted in thrusting or obduction of the ophiolites onto the borderlands, possibly from short-lived subduction zones along the northern and southeastern margins.

The second scheme, suggested by Tailleur, argues that the present V-shape of the northern and southeastern margins of the Yukon-Koyukuk province is a relatively young feature and the result of clockwise rotation by oroclinal bending. Prior to Late Cretaceous time, these two margins are envisaged as a nearly straight continental margin, approximately paralleling the present-day southern edge of the Brooks Range. The ophiolites were obducted onto this continental margin during convergence with an oceanic plate during Middle Jurassic to Early Cretaceous time. The Rampart part of the obducted sheet was subsequently rotated to its present position by oroclinal bending. According to this scheme, the western Brooks Range ophiolites and Rampart ophiolites are regarded as part of a single allochthonous sheet, whereas in the first scheme they are believed to be remnants of two separate sheets emplaced in opposite directions.

A detailed discussion of each hypothesis is beyond the scope of this report. Suffice it to state that, while the second offers a simpler explanation for emplacement of the ophiolites, it is seriously complicated by time and space problems attendant upon oroclinal bending of as much as 130°.

## ACKNOWLEDGMENTS

We wish to express our thanks to our colleagues D. L. Jones and J. M. Hoare for their reviews and discussions of this report and to Dietrich Roeder and C. G. Mull for kindly providing us with an advance copy of their paper "Tectonics of Brooks Range ophiolites" (in press).

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# OPHIOLITIC ASSEMBLAGES IN THE CANADIAN CORDILLERA

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## ABSTRACT

Ophiolites in the Canadian Cordillera occur almost entirely in two belts of late Paleozoic and Mesozoic strata that lie 250 to 500 km east of the present continental margin. The western belt, called the Cache Creek-Bridge River assemblage, appears to have been emplaced in part in latest Triassic and in part in latest? Jurassic time. The eastern belt, known as the Eastern assemblage, lies in places with stratigraphic contact on Devonian and older miogeoclinal strata and elsewhere is allochthonous. Emplacement of the allochthonous component was probably in the Late Triassic. Other possible ophiolites are of pre-Permian age in the Saint Elias Mountains, southwestern Yukon, and Eocene age on southern Vancouver Island.

## INTRODUCTION

This brief summary gives the location, nature, and setting of rocks in the Canadian Cordillera considered by the writer to be ophiolites. 'Ophiolite' is used for mafic and ultramafic associations with the characteristics outlined in the report of the first Penrose Field Conference (Geotimes, December, 1972). In the Canadian Cordillera such rocks are mainly in two belts of upper Paleozóic and Mesozoic strata, the Cache Creek-Bridge River and Eastern assemblages, that lie between 250 km and 500 km east of the Pacific Ocean in the Intermontane Belt and Omineca Crystalline Belt (Figure 1). Volcanic rocks in these belts are entirely mafic and commonly associated with alpine-type ultramafics. They contrast with coeval volcanosedimentary strata elsewhere in the western Cordillera that contain volcanics ranging in composition from basalt to rhyolite, with no associated alpine ultramafics (Monger, in press). Smaller areas of possible ophiolites are of Eocene age on southernmost Vancouver Island, southwestern British Columbia (Muller, 1977), of pre-Permian age in the Kluane area, southwestern Yukon (Read and Monger, 1975) and, (extraterritorially) of Jurassic age in the San Juan Islands, northwestern Washington (Whetten and others, 1976; Brown, this volume). Other mafic and ultramafic associations in the western Cordillera, namely zoned, intrusive, Alaskan-type ultramafic bodies and probably related Lower Cretaceous alkaline basalt in southeastern Alaska (Irvine, 1973) and comparable associations of Upper Triassic age in the Intermontane Belt (Irvine, 1976) are not discussed herein.

## CACHE CREEK-BRIDGE RIVER ASSEMBLAGE

Most of this assemblage is made up of the Cache Creek Group, except in the extreme south where it includes the partly younger Bridge River (or Fergusson) Group and its probable correlative the Hozameen Group. Lithologies in the Cache Creek Group in decreasing order of abundance are (1) ribbon chert and argillite, typically radiolarian-bearing, (2) massive, shallow water carbonate, (3) mafic volcanic rock (Figure 2a), mainly flows but locally breccia, and related diabasic or microgabbroic intrusions and (4) alpine-type ultramafics. The Bridge River and Hozameen Groups differ in that the amount of carbonate in them is relatively minor. Lithological units in this assemblage are discontinuous, in part due to originally podiform stratigraphy but in part to locally intense tectonic disruption. The stratigraphy is thus very difficult to establish. The age of the Cache Creek Group ranges from Early Mississippian to Late Triassic and that of the Bridge River Group from Middle Triassic to possibly mid-Jurassic. Carbonate in the Cache Creek Group is mainly Pennsylvanian and Permian, but locally is as old as Early Mississippian and possibly as young as Late Triassic. Fossils from chert are Pennsylvanian, Permian, Middle and Late Triassic, according to D.L. Jones and E. Pessagno (pers. comm. 1977). Middle Triassic fossils have been obtained from carbonate in the Bridge River Group and possible mid-Jurassic ones from chert (D.L. Jones, pers. comm. 1977). External contacts of this assemblage with other upper Paleozoic and lower Mesozoic sequences are commonly tectonic. The Cache Creek Group is intruded by mainly Mesozoic granites, some as old as 200 m.y., and generally overlain only by Cretaceous and Tertiary continental rocks, although in a few places possible Lower Jurassic strata overlie it. The Bridge River Group is overlain by marine Albian strata (H.W. Tipper, pers. comm. 1977) and possible Lower Jurassic rocks lie on volcanics that probably belong to the Hozameen Group (Anderson, 1976). Metamorphic grade of the assemblage is generally prehnite-pumpellyite or pumpellyite-chlorite, but blueschist is known from the Atlin Terrane, from the Stuart Lake belt (which yields 210-220 m.y. K/Ar ages), and from the type area (Monger, 1969; Paterson, 1973; Paterson and Harakal, 1974; J. Grette, pers. comm., 1976).

Paleontological and paleomagnetic arguments suggest that it is reasonable to interpret the Cache Creek and Bridge River Groups as fragments of the floor of the ancestral Pacific Ocean of, respectively, mainly late Paleozoic and early Mesozoic ages, that were trapped east of allochthonous, coeval, non-ophiolitic terranes now underlying parts of the Intermontane Belt, Coast Plutonic Complex and Insular Belt (see discussion in Monger, in press). Time of emplacement of the Cache Creek Group was probably in the Late Triassic and that of the Bridge River Group possibly in the late Jurassic or earliest Cretaceous.

The most widely exposed and probably best preserved strata in this assemblage form the Atlin Terrane, northwestern British Columbia (Figure 2b). This area is bounded by major faults. Those on the southwest are thrust and high-angle reverse

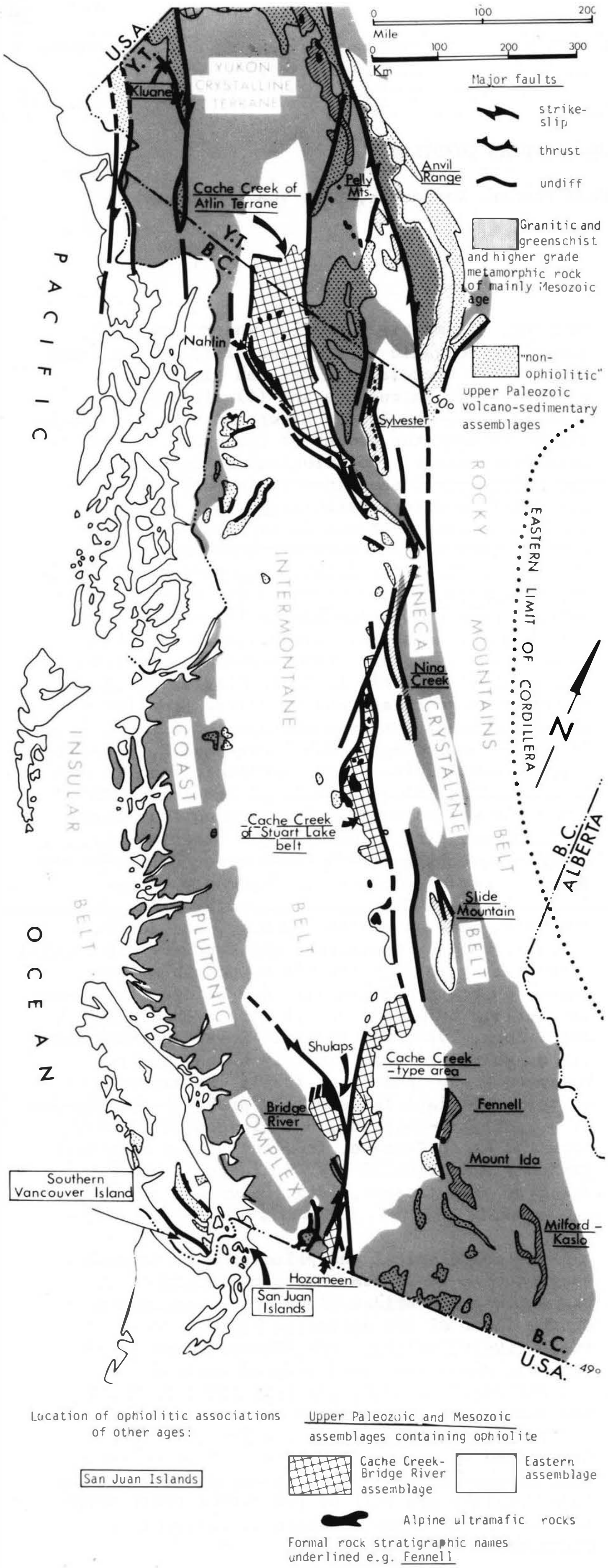


Figure 1. Distribution of assemblages containing ophiolites in the Canadian Cordillera.

faults of probable Late Jurassic age that bring Cache Creek rocks over and against Upper Jurassic, Upper Triassic and, locally, Permian strata. The ultramafic rocks are typically associated with the mafic volcanics, as noted by Aitken (1953). The large Nahlin ultramafic body (Figure 2c), exposed along the southwestern fault system, possibly represents the basement of the Cache Creek Group in this region. Ultramafic fault slices splay out from the Nahlin body and are intimately faulted with massive basalt that underlies Cache Creek sedimentary rocks. Within the main Nahlin body are small fault slices containing cumulative gabbro that intrudes basic volcanics. A rubidium strontium age of 325 $\pm$ 20 m.y. from the basalt (R.L. Armstrong, pers. comm., 1976), northeast of the ultramafic body, agrees well with the Late Mississippian age of fossils in tuffaceous limestone along the upper contact of the basalt. In places the basalt is overlain stratigraphically by thick (up to 2000 m) masses of shallow water reefoidal limestone, one example of which contains fossils of all epochs from the Late Mississippian to the Late Permian. These carbonate bodies are presumably Bahama-type banks or large atolls built on the volcanic basement. Northeast of the basalt-carbonate belt is mainly chert, that contains lenses of locally brecciated volcanics and carbonate. The latter are probably olistoliths, perhaps derived from the reefoidal carbonates and underlying volcanics to the southwest. Along the northeastern margin of the Atlin Terrane, chert is stratigraphically capped by basalt that in turn is overlain gradationally by shallow water Permian limestone. Thus, basalt in the Atlin Terrane appears to have been extruded both as part of an ophiolitic basement that in places was thick enough to have a cover of shallow water carbonate, and also as seamounts on chert that overlies this basement (Monger, 1977).

The Nahlin ultramafic body was originally mapped on a scale of 1:250,000 by Aitken (1959) and Souther (1971). Recently, J. Terry of the University of Lille did detailed mapping of parts of it for a comparative study of this body with ophiolites of the Pindus nappe in Greece; details on Figure 2c are mainly his mapping (Terry, 1977). From field work Terry finds that most lithologies of "classical" ophiolites are present but the sequence is disrupted. Much of the body is well-folidated peridotite, ranging in composition from peridotite with up to 40% pyroxene, to dunite almost devoid of pyroxene. Pyroxenite forms layers up to 25 cm thick throughout the peridotite and dunite is present as stringers, diffuse bands, and irregularly-shaped masses. The pyroxenite layering is deformed by probably more than one episode of isoclinal folding with the pervasive foliation parallel to the axial surfaces of these folds. The latest folds are large-scale, regular open structures. Numerous, sub-parallel diabasic intrusions up to 25 m wide, many of which are composite, cut the foliated ultramafics, and with increasing serpentinization and shearing become disrupted and form discrete, lensoidal bodies of rodingite. Very small areas of cumulates occur in fault slices within the main mass of foliated peridotite, and grade from peridotite, through pyroxenite and gabbro to trondjhemite. The last two lithologies intrude basic volcanics also preserved in the fault slices. Although local chilled contacts have been

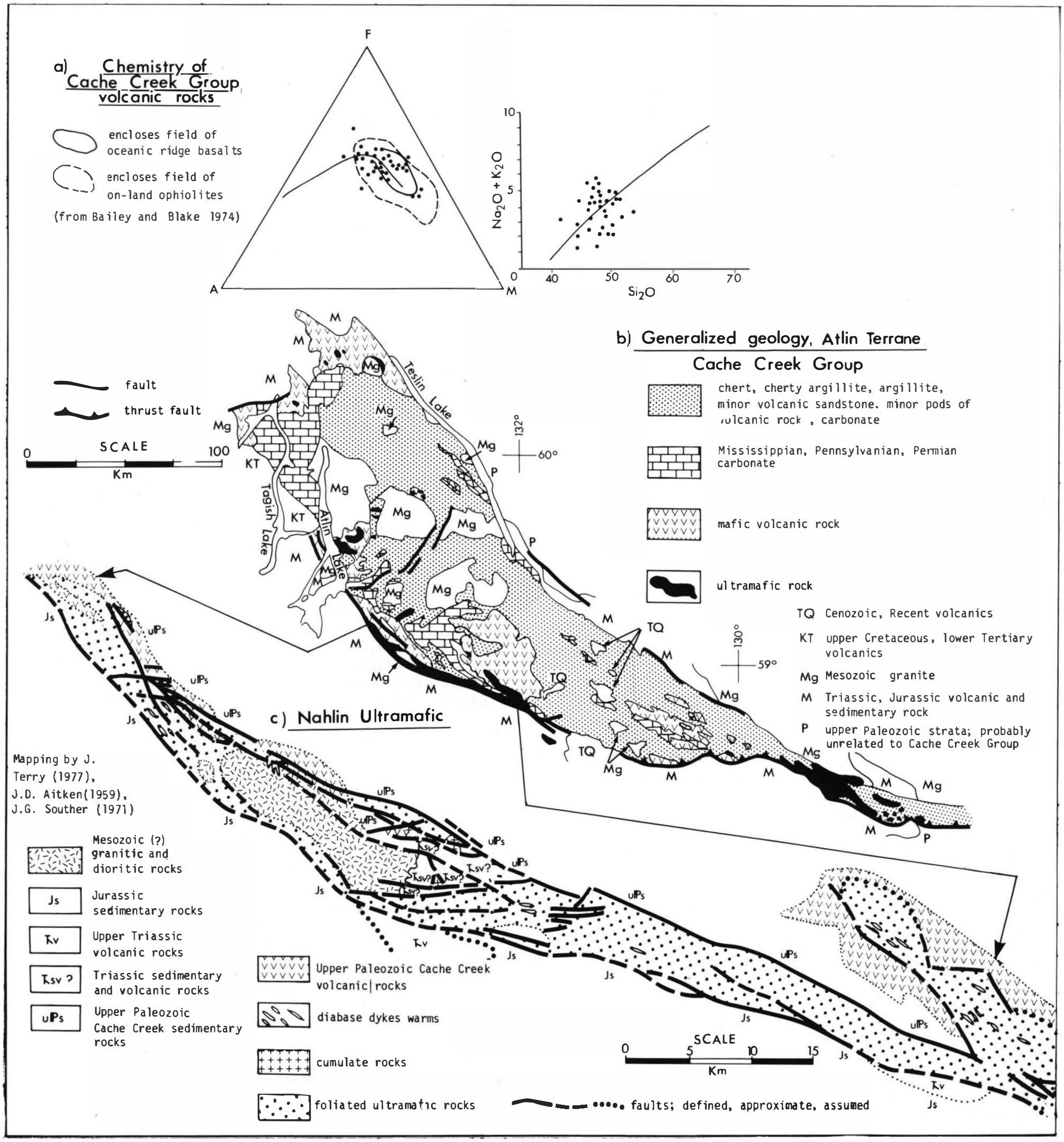


Figure 2. Cache Creek Group: (a) chemistry of volcanic rocks (from Monger, 1975 and unpublished; Paterson, 1973), (b) geology of the Atlin Terrane and (c) geology of Nahlin ultramafic body (from Terry, 1977).

observed in fine grained diabase, no consistent pattern of sheeted dykes is known.

Alpine-type ultramafics (Trembleur intrusions) and basic volcanics of the Cache Creek Group were mapped by Armstrong (1949, pp. 79-92) in the Stuart Lake belt, 700 km south southeast of the Atlin Terrane. An early detailed study of some of these ultramafic rocks was made by Little (1949) who reported variably serpentinized peridotites and dunites together with pyroxenites and norites. Also in this area, a succession of harzburgite with minor dunite and pyroxenite, gabbro and mafic volcanic rock, with poorly exposed contacts, was recognized as an ophiolite by Paterson (1973, p. 26). From the same general area Ross (1977) described the internal fabric of a partly serpentinized harzburgite and dunite and found evidence for three deformational episodes. He suggested the two earlier penetrative fabrics were probably of mantle origin and the latest one was produced during emplacement within the crust. In the Stuart Lake belt, ophiolites appear to have been emplaced and eroded prior to the latest Triassic. Serpentinite, chromite, and basic volcanic detritus is present in an uppermost Triassic (Monotis bearing) conglomerate adjacent to the Cache Creek Group (Armstrong, 1949, p. 53; Paterson, 1973, p. 37).

In the eastern margin of the Coast Plutonic Complex, southwestern British Columbia, chert containing possible mid-Jurassic radiolaria (D.L. Jones, pers. comm., 1977), argillite, pillow basalt, and minor limestone pods with Middle Triassic conodonts form the Bridge River (or Fergusson) Group (Roddick and Hutchison, 1973, pp. 2-3). Numerous alpine-type ultramafics are spatially associated with this group, and locally (tectonically?) cut nearby probable Upper Triassic clastic rocks as well. The relationship of these Upper Triassic rocks to the Bridge River Group is unknown.

The largest ultramafic body in the region, known as the Shulaps, was mapped by Leech (1953) on a scale of one mile to one inch. It lies between the Bridge River Group, to the southwest, and the major, Yalakom fault to the northeast. It consists of harzburgite, dunite and pyroxenite, with these three lithologies locally interlayered and cut in places by gabbroid dykes. On the southwest side of the body are exposed diopside pyroxenite, locally layered gabbro, and intermingled gabbro and pillow basalt. Recent studies by J. Nagel of the University of British Columbia (pers. comm., 1977) show a melange zone on this southwest side consisting of pervasively sheared serpentinite with a near-horizontal schistosity, and inclusions of volcanic, gabbroic and sedimentary rocks on all scales up to 150 m long. The ultramafic rocks appear to be tectonically emplaced within possible Upper Triassic strata, although J. Nagel (pers. comm., 1977) has found it difficult to correlate these strata with known Upper Triassic rocks in the region. The age of emplacement of the Shulaps ultramafic is pre-Early Cretaceous. Chromite grains, presumably derived from this body,are present in nearby sandstones that formerly were thought to be of Early Jurassic age (Leech, 1953, p. 39) but now are known to be Early Cretaceous (H.W. Tipper, pers. comm., 1977).

Wright (1974) made a detailed petrographic and structural study of the Pioneer ultramafic body which is exposed 30 km southwest of the Shulaps ultramafic. This ultramafic body lies along the contact between the Bridge River Group and downfaulted argillite, chert, greenstone conglomerate and breccia assigned to the Upper Triassic Noel Formation. It consists of partly serpentinized harzburgite with minor orthopyroxenite, dunite, and rare chromitite, and exhibits well-developed layering. Wright (1974, p. 119) concluded that this body is part of a dismembered ophiolite.

The Hozameen Group appears to be the south-southeastern continuation of the Bridge River Group across the Fraser River fault system. Separating it from Lower Jurassic clastic rocks to the east is a belt of serpentinite, gabbro, metadiorite and mafic volcanics known as the Coquihalla serpentine belt (Cairnes, 1930, p. 144a). In a recent note, Anderson (1976, pp. 443-446) reported a depositional contact between volcanics of the serpentine belt, that he concluded was oceanic lithosphere, and possible early Lower Jurassic strata of the Ladner Group to the east. The recently discovered Jurassic radiolaria in the Bridge River Group suggest a possible correlation with rocks in the San Juan Islands, mentioned below, as well.

## EASTERN ASSEMBLAGE

Rocks in the Eastern assemblage extend discontinuously from the Yukon boundary with Alaska in the north, at latitude 65°, almost to the British Columbia-Idaho boundary in the south (Figure 1). The assemblage consists of two divisions, the lower predominantly sedimentary, and the upper, mainly volcanic. The gross stratigraphy of this assemblage, is summarized in Figure 3a. The lower division is mainly fine-grained clastic rock and ribbon chert, with local sandstone, conglomerate, and carbonate. In most places, these rocks lie with stratigraphic or possibly structural contacts on shallow water and intertidal Devonian carbonate, which is the uppermost unit of the Cordilleran miogeoclinal succession in these places. In southeastern British Columbia, the assemblage is underlain by a terrane probably metamorphosed in Devonian time (P.B. Read, pers. comm., 1976). The age of the lower division ranges from Late Mississippian to Early Permian. Above the lower division is mainly basalt (Figure 3b), local diabase, gabbro and alpine-ultramafic rock. In the Anvil Range, Yukon Territory, and Nina Creek, central British Columbia the volcanics are Lower Permian. The Sylvester Group contains at least two volcanic assemblages; one, pre-Late Mississippian, the other, mid-Permian. Elsewhere, the volcanics are dated as post-Late Pennsylvanian and post-Late Mississippian, pre-Late Triassic. Metamorphism is predominantly in the prehnite pumpellyite grade, but locally in greenschist and, rarely, amphibolite grade. Contact relations between the upper and lower divisions vary. In places, as near Nina Creek and between the Milford and Kaslo Groups the two appear to be in normal stratigraphic relationship, with diabase and gabbro sills, feeders to the volcanic rocks, in the top of the underlying sedimentary succession.

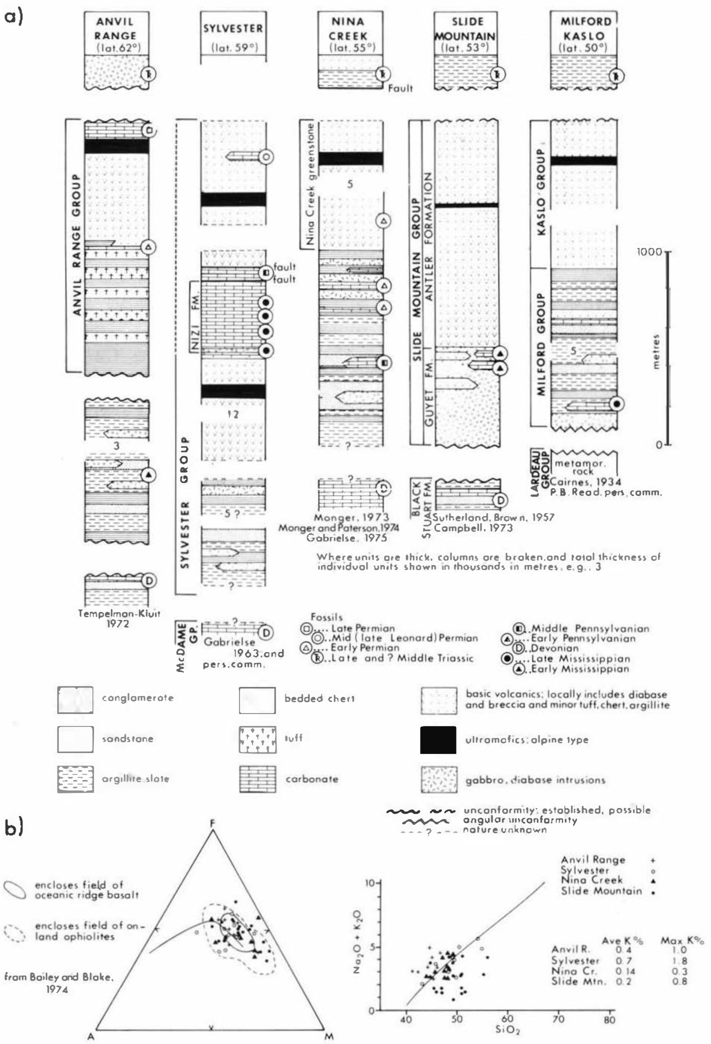


Figure 3. Eastern assemblage (from Monger, in press) (a) gross stratigraphy of the Eastern assemblage, (b) chemistry of volcanic rocks of the Eastern assemblage, from Tempelman-Kluit, 1972, Anvil Range; Gabrielse, 1963 and unpublished, Sylvester Group; Gabrielse, 1975, Nina Creek; K.V. Campbell, 1971, Slide Mountain.

Elsewhere, as in the Pelly Mountains, Yukon Territory, ultramafic and volcanic rocks lie above thrust faults on miogeoclinal strata (Tempelman-Kluit, 1976, pp. 99-100), or else the nature of the lower contact of the volcanics is uncertain (Campbell and others, 1973, p. 80). The assemblage can be interpreted as having formed along the western margin of the North American craton, in part on the extension of the craton, in part on oceanic crust. In southern British Columbia, where coeval arc-like rocks occur immediately to the west, there may have been a true marginal basin, but elsewhere only rocks of the Cache Creek Group, with different faunas, are known to the west. Where the rocks are in normal stratigraphic succession, there was apparently deepening of the basin followed by extrusion of mafic volcanics. Elsewhere, the ultramafic and mafic rocks appear to have been thrust right out of the basin, as in the Yukon, or else to be internally imbricated, so that the volcanics contain ultramafic slices. Tempelman-Kluit and others (1976) suggest the time of thrusting in the Yukon was in the Upper Triassic.

As in the case of the Cache Creek-Bridge River assemblage, relatively little detailed work has been done in recent years on the ophiolitic assemblages of the upper division. Gabrielse (1955, 1963) discussed ultramafics in the Sylvester Group, northern British Columbia, and described peridotite (hartzburgite and lherzolite), dunite and pyroxenite with ubiquitous chromite. A detailed study of one of these ultramafic bodies was made by Wolf (1965), who reported serpentinite, interlayered peridotite and dunite and variably rodingitized gabbroic dykes and sills. The contact between this body and mafic volcanic rocks is marked by an aureole over 100 m wide of amphibolite with gneissic textures. Further south, in east-central British Columbia, Hall-Beyer (1976) studied major and trace element geochemistry of the Slide Mountain basalt and concluded they were ocean floor tholeiites and komatiites. S. Montgomery (pers. comm. to R.B. Campbell, 1976) working on an ultramafic body in the metamorphic terrane between the Slide Mountain Group and Fennell Group, found a peridotite base, passing up into gabbro and overlying greenschists.

## KLUANE AREA

The upper Paleozoic rocks of the Saint Elias Range, southwestern Yukon and eastern Alaska Range form the Skolai Group, which consists of a lower basic to intermediate volcanic unit, the Permian and (?) Pennsylvanian Station Creek Formation, and an upper sedimentary unit, the Permian Hasen Creek Formation (Smith and MacKevett, 1970; Read and Monger, 1975). In the Kluane area, south-western Yukon, the typical Skolai stratigraphy with two formations is present only on the east side of the outcrop area. On the west side, at one locality, the Hasen Creek Formation non-conformably lies on coarse, locally pegmatoid uralitic gabbro and diabase. Nearby, the basal Hasen Creek consists of a conglomerate with gabbro and diabase together with local pyroxenite and probable peridotite clasts. Similar gabbroic rocks occur to the west in the eastern Alaska Range (E.M. MacKevett, pers. comm., 1976). Richter and Jones (1973) suggested that the upper Paleozoic volcanic rocks in the eastern Alaska Range are the remains of an island arc that was built upon oceanic crust. This crust may be the gabbro and diabase in the Kluane area.

## SAN JUAN ISLANDS

Ophiolites in the San Juan Islands, discussed elsewhere in this volume by Brown, are of mid-Jurassic age. They are mentioned here only because they may represent the correlative of the basement of a sequence of highly deformed radiolarian chert, argillite, greywacke, breccia, and minor pillow basalt of Late Jurassic to Early Cretaceous age that is exposed on the west coast of Vancouver Island and called Pacific Rim Complex by Muller (1973 and pers. comm.) and possibly the Bridge River Group.

## SOUTHERN VANCOUVER ISLAND

Basic volcanic rocks of Eocene age known as the Mechosin Volcanics, and associated gabbroic Sooke Intrusions, described by Muller (1977) represent the northerly extension into Canada of Olympic Peninsula geology. The volcanics are up to 3,000 m thick and show a progression from deep water pillow basalts at the base to subaerial flows at the top. Dykes are particularly abundant in the lower part of the section. The composition of the volcanics is predominantly tholeiitic basalt with minor alkali basalt. The Sooke Intrusions are mainly gabbro with minor quartz dioritic and trondhjemitic phases, and local hornblendite and aplite stringers. The gabbro is locally coarse-grained olivine gabbro and in places becomes a gneissic plagioclase amphibolite. Muller (1977, p. 292) considers these rocks to have the characteristics of oceanic crust. Although no ultramafics are exposed, other than slivers of serpentinite along the major fault that bounds these volcanics on the north, a large gravity high in the region may indicate they lie close to the surface.

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# OPHIOLITE ON FIDALGO ISLAND, WASHINGTON

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ABSTRACT

Pre-Tertiary rocks on Fidalgo Island in northwestern Washington occur in a stratigraphic sequence which from the bottom upward consists of serpentinite, layered gabbro, a plagiogranite dike complex, keratophyric and spilitic volcanic rocks, sedimentary breccia, pelagic argillite and terrigenous siltstones and sandstones. This suite is interpreted to be an ophiolite. Disrupted parts of the sequence occur widely in the San Juan Islands and adjacent mainland. The gabbroic and granitic parts of the section were formerly considered to be part of the Ordovician Turtleback Complex, exposed on Orcas Island. However, new radiometric dates indicate a Jurassic age for this rock, and it is renamed the Fidalgo Complex. The tectonic setting during formation of the ophiolite is an unresolved question; the pelagic sediments indicate an oceanic origin, but the silicic igneous rocks suggest an island-arc setting. When viewed in a regional context it is apparent that the Fidalgo ophiolite and associated rock units of northwestern

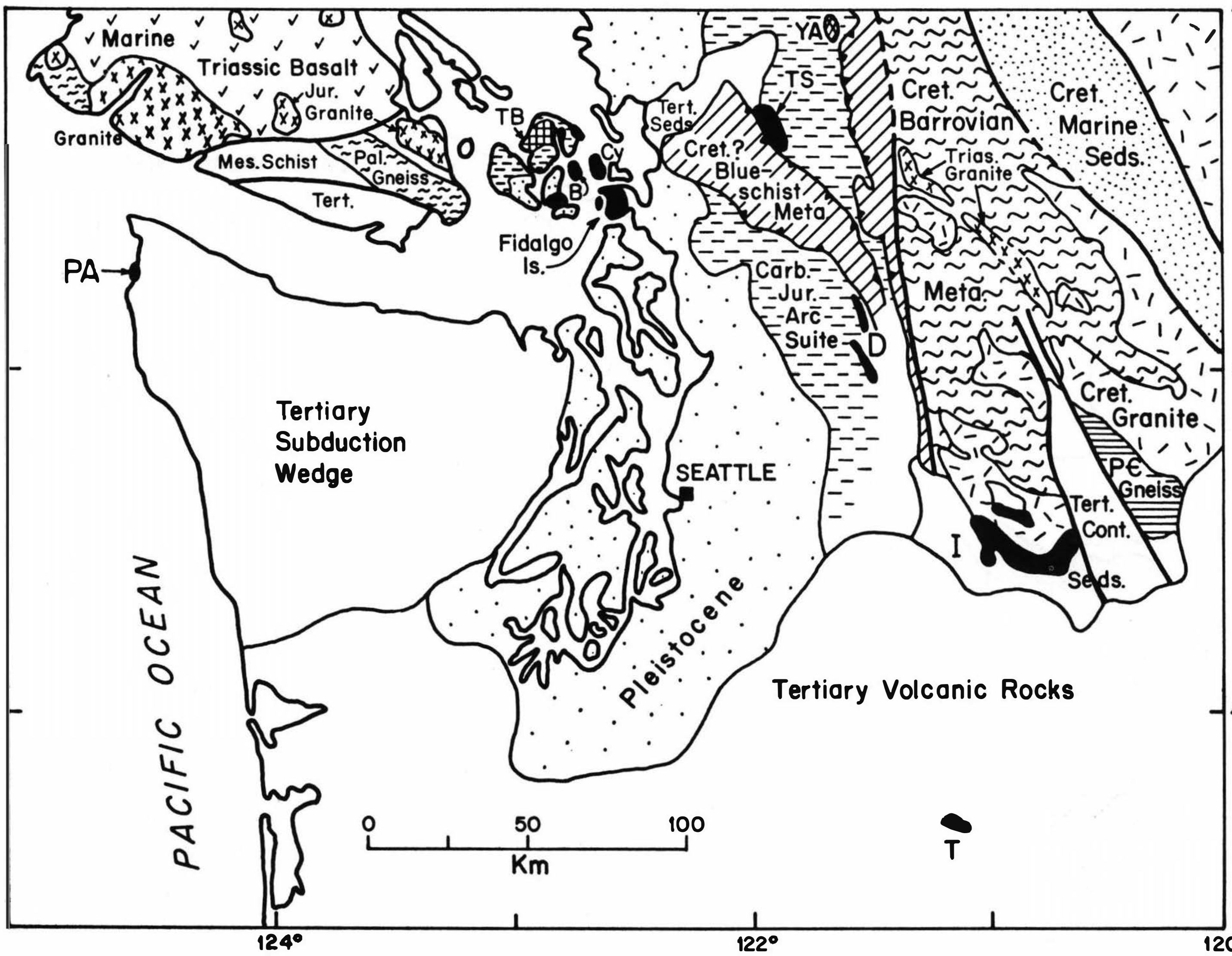



Fig. 1. Regional geologic setting of the Fidalgo ophiolite. PA=Point of the Arches ophiolite, Jurassic; TB=Turtleback Complex, Ordovician; B=Blakely Island ophiolite, Jurassic; CY=Cypress Island periodotite; TS=Twin Sisters dunite; YA=Yellow Aster Complex, Precambrian and Ordovician; D=Darrington ultramafic bodies; I=Ingalls periodotite/ophiolite, Jurassic; T=Tieton ophiolite, Jurassic. The map is simplified and interpretive. It is based on data from the following sources; Huntting and others, 1961; Misch, 1966; Mattinson, 1972; Snavely and others, 1972; Carson, 1973; Hopson and Mattinson, 1973; Muller and others, 1974; Tabor, 1975; Vance and others, 1975; and R. Armstrong, 1976, personal communication re/age of blueschist metamorphism. Heavy contact lines represent high-angle faults.

Washington and Vancouver Island have been greatly dislocated relative to one another if their formation and present distribution are to be explained by plate-tectonic theory.

## INTRODUCTION

This report concerns pre-Tertiary rocks on Fidalgo Island which are part of an assemblage of dioritic and gabbroic rocks exposed sporadically in the San Juan Islands, where they are called the Turtleback Complex (McClellan, 1927), and North Cascades of northwestern Washington, where they are known as the Yellow Aster Complex (Misch, 1966). The regional geologic setting of Fidalgo Island is shown on Figure 1.

The origin of the Turtleback and Yellow Aster Complexes has important bearing on the general problem of the tectonic evolution of northwestern Washington. They have been variously interpreted to be: (1) continental crust occurring as an extension of the North American Craton (Misch, 1966, 1973, 1974); (2) continental crust representing a microcontinent separate from the North American Craton (Vance, 1974); and (3) portions of an ophiolite possibly representing oceanic crust (Hopson and Mattinson, 1973; Brown and Bradshaw, 1974, 1975; and Brown and others, 1977).

An allied problem presented by these rocks is that of the petrogenesis of plagiogranite and keratophyre which constitute a large portion of the complex. Such silicic rocks occurring elsewhere in ophiolites are interpreted to have formed by fractional crystallization of gabbroic melt in an ocean ridge setting (Coleman and Peterman, 1975), or to be the result of calc-alkaline magmatism in an island arc environment (Miyashiro, 1973). The plagiogranite and keratophyre of the Fidalgo ophiolite are described in detail in a separate report (Brown and others, 1977).

The purpose of this report is to briefly describe the Fidalgo ophiolite, to present evidence relating to the tectonic setting of its origin, and to consider its significance to the regional geology.

## FIELD AND PETROGRAPHIC RELATIONS

__Introduction:__

Primary igneous and stratigraphic relations among differing lithologies of the Turtleback and Yellow Aster complexes are in most places poorly preserved because of faulting. However, the pre-Tertiary terrane on the northern part of Fidalgo Island is relatively undisturbed, and relations established there can be used to explain the correlative but disrupted rock suites exposed elsewhere. A geologic map of this area is given in Figure 2 and

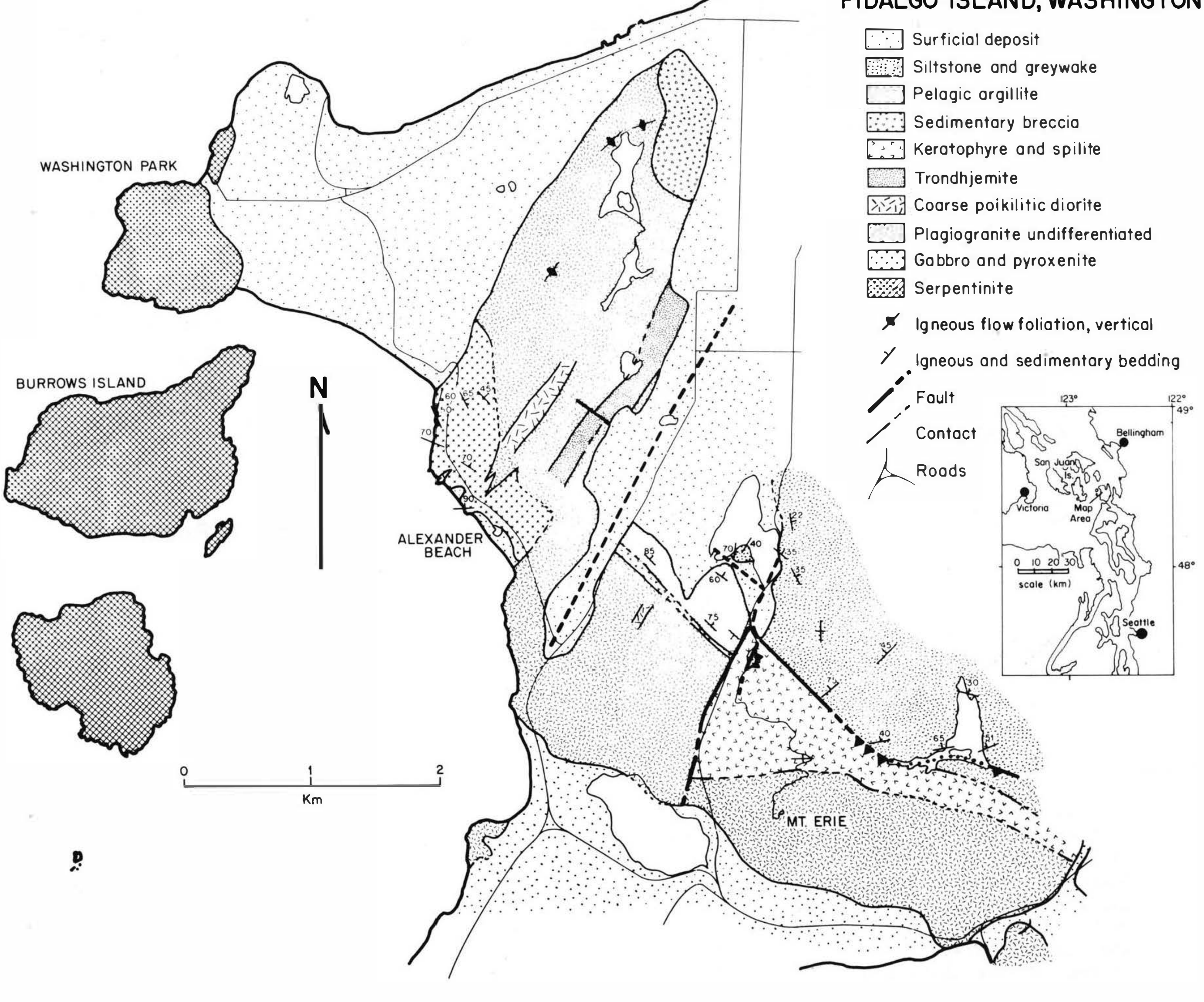
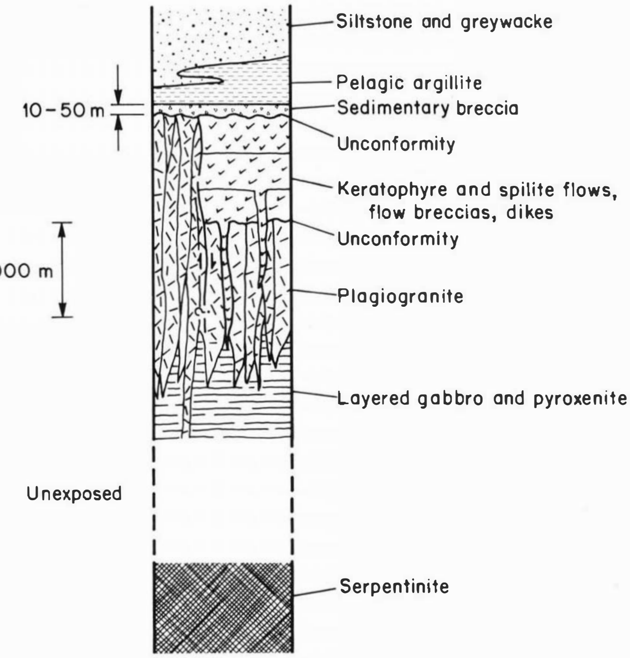



FIG. 2 GEOLOGY OF PART OF FIDALGO ISLAND, WASHINGTON

a somewhat idealized stratigraphic section in Figure 3.

FIG.3 STRATIGRAPHIC SECTION OF FIDALGO ISLAND MAP AREA




Serpentinite:

The stratigraphically lowest part of the complex is serpentinized peridotite which occurs at Washington Park and on Burrows and Allan Islands (Fig. 2). Similar ultramafic rock occurs widely in northwestern Washington, the most notable occurrences being at Cypress Island and the Twin Sisters (Fig. 1).

In previous studies the ultramafic rock of this region has not been linked with the gabbro, mainly because an undisturbed contact between the ultramafic rock and gabbro has not been observed. The genetic relation proposed here is based on circumstantial evidence, as follows: (1) There is close spatial association between serpentinite and gabbro throughout the region. The two rocks occur together in fault contact at many localities (e.g., Fidalgo Is., Blakely Is., South Chuckanut Mtn., Vedder Mtn.). (2) On Fidalgo Island exposures of the main unit of serpentinite are separated from that of layered gabbro by water and glacial drift; but cumulus layering and graded bedding in the gabbro indicate that the ultramafic rock lies stratigraphically below the gabbro, a relation suggestive that the two rock units are in place with respect to one another and represent the basal portion of an ophiolite.

The ultramafic rock on or near Fidalgo Island and Cypress Island consists of serpentinized harzburgite containing irregular and discontinuous layers of dunite and olivine chromitite (Raleigh, 1965). The layering is predominantly north to northwest striking and steeply dipping. Isoclinal folds are found in these layers. Planar or gently folded veins (or dikes) or dunite and pyroxenite cross-cut the layering and isoclinal folds. The peridotite has a well developed tectonite fabric defined by preferred orientation of olivine with a strong (100) maximum normal to girdles of (010) and (001) (Raleigh, ibid.). Attitudes of mesoscopic fold axes, mineral lineations and (100) olivine maxima are mutually parallel and vary from a gentle southwest to steep northeast plunge.

Raleigh (ibid) observed that textures in fresh peridotite on Cypress Island, are similar to those in the ultramafic zone of the Stillwater Complex, and on this basis he inferred a cumulate origin for the ultramafic rock of Cypress and Fidalgo Islands.

Gabbro:

The gabbro is exposed only over a small area on Fidalgo Island, near Alexander Beach. However, much gabbro occurs elsewhere in the region, as on Blakely Island, South Chuckanut Mtn., and Vedder Mtn. Virtually all of the gabbro on Fidalgo Island shows cumulus layering, and locally graded bedding. Slump structures, pyroxenite dikes and sills and gabbroic pegmatite dikes are also common features in this unit. The primary minerals are plagioclase, orthopyroxene and clinopyroxene. Olivine is not found, however, the gabbros are altered to an extent that all primary olivine could have been destroyed. Low-grade metamorphism of the gabbro has caused partial replacement of the primary phases by minerals of the greenschist facies: Ca-plagioclase is completely replaced by a fine-grained aggregate of albite + epidote; clinopyroxene is partially replaced by actinolite; and orthopyroxene is partially replaced by chlorite. Grain outlines of the original gabbroic minerals are still well preserved, and a cumulate texture with interstitial plagioclase can be observed in some specimens.

Locally, another type of alteration, predating the greenschist facies metamorphism, has been caused by intrusion of dikes of plagiogranite and hornblende basalt. The altered gabbro shows all stages of replacement of pyroxene by hornblende.

Dike Rocks:

The layered gabbro, gabbroic pegmatite, and pyroxenite dikes and sills are cross-cut by dikes of mafic to felsic rocks which intrude roughly normal to the gabbroic layering, and become increasingly abundant toward the upper part of the gabbroic section. In the middle and upper parts of the dike complex gabbro is absent. The dikes comprise a suite of lithologies intergradational in both texture and mineralogy, which includes hornblende gabbro, diorite, trondhjemite, albite granite, diabase, keratophyre and basalt. The diorite (56-58% $SiO_2$) predominates. In the basal part of the dike complex the dikes have sharp contacts with the gabbro, contain xenoliths of gabbro and have caused hydrothermal contact metamorphism of the gabbro as described above. Especially in the basal part of the dike complex, flow foliation is strongly developed in the dikes.

Plagioclase and hornblende are ubiquitous primary minerals in the dikes and quartz and Fe-Ti oxide minerals are common. Some dike rocks contain pyroxene as well as hornblende. K-feldspar is absent, even in the most silicic ($SiO_2$ = 75%) varieties, and the term plagiogranite (cf. Thayer, 1973; Coleman and Peterman, 1975) applies to much of the rock in this suite. On Figure 2 undifferentiated rock of the dike complex is broadly referred to as plagiogranite. Primary textures are hypidiomorphic granular, with quartz being the typical interstitial phase. Myrmekitic texture occurs in some of the finer-grained dikes of albite granite. Low grade metamorphism has caused partial development of typical greenschist facies minerals, as in the gabbro.

Volcanic Rocks:

Keratophyre and spilite comprise the volcanic suite. These rocks occur mainly as flows and flow breccias overlying the plagiogranite and other dikes. Pillows have not been found in the map area of Figure 2, but are common in correlative volcanic rocks elsewhere on Fidalgo Is. (e.g., Rosario Head) and in the San Juan Islands. Minor red chert occurs locally in brecciated flows. Keratophyric dikes are common in the upper part of the plagiogranite complex and may represent feeders for the volcanic pile.

Phenocrysts are plagioclase and clinopyroxene. The unaltered groundmass mineral is dominantly plagioclase, occurring together with quartz in the felsic varieties. Alteration minerals are chlorite, quartz, calcite, iron-rich epidote, prehnite and zeolites. Porphyritic, synneusis and trachytic textures are generally well preserved.

Sedimentary Rocks:

Sedimentary rock occurs unconformably on plagiogranite in part of the map area (Fig. 2). A contact relation between volcanic and sedimentary rock has not been observed. However, clasts of keratophyre and spilite occurring in the basal part of the sedimentary unit indicate the time sequence of the rocks.

Sedimentary rock directly overlying plagiogranite is coarse breccia composed entirely of fragments of plagiogranite and volcanic rock. The relative proportions of plutonic and volcanic fragments vary greatly over a distance of a few 10s of meters along strike. This unit ranges from 10 to 50 meters in thickness. It passes conformably, and sharply, upward to pelagic argillite. Fine-grained (metalliferous?, pelagic?) material occurs in the breccia matrix near the contact with argillite. The argillite is thin-bedded, brown or green and is more than 90% composed of clay-sized material and radiolaria. Sparse interbeds consist of greenish tuff, radiolarite, chert, carbonate sandstone with ultramafic minerals, and breccia containing clasts of plagiogranite and keratophyre. The mapped thickness of the unit is about 300 meters.

A thick unit of terrigenous siltstone and greywacke-type sandstone occurs in a general stratigraphic position above the argillite and igneous rocks. However, an unfaulted contact has not yet been found. The lowest beds are siltstone; higher in the section sandstone and even conglomerate appear. Clast lithologies are chert, siltstone, volcanic rock, plagiogranite and rarely epidote. Some beds are tuffaceous. Turbidite bedding is common.

## AGE OF THE ROCKS

Three K/Ar ages of hornblende in diorite have been determined for this report by R. Forbes. The data, summarized in Table 1, indicate that the plagiogranite dike complex crystallized at about 155±5 m.y. ago. A previously published U/Pb date on zircons from two plagiogranites on Fidalgo and Blakely Islands, obtained by Whetten and others (1976), is 170±10 m.y.

Preliminary study of radiolaria in the lowermost beds of pelagic argillite by Pessagno (pers. comm., 1976) indicates a Tithonian to Neocomain age (approximately 140 m.y.) for this part of the sedimentary section. Cretaceous radiolaria have been found in the siltstone greywacke unit by E. Artim (pers. comm., 1976) and Mulcahey (1974).

Table 1 K/Ar ages of hornblende in dike rocks of the Fidalgo ophiolite. Data obtained by R. Forbes, Geophysical Institute, University of Alaska.

| Sample | Rock Type & Location | $K_2O$ Wt% | 40/Ar moles/gm $x10^{-11}$ | Age± 1 sigma m.y. |
|---|---|---|---|---|
| 52-F35 | hornblendite Mt. Erie | 0.092 | 2.250 | 158.6±7.9 |
| 52-F82 | diorite Mt. Erie | 0.128 | 2.914 | 149.0±7.4 |
| 52-F84 | hornblende gabbro Alexander Beach | 0.400 | 9.457 | 153.6±4.6 |

The Fidalgo ophiolite suite thus appears to have originated in mid- to late-Jurassic time, with terrigenous sedimentation extending into the Cretaceous.

The Jurassic age of the Fidalgo ophiolite contrasts sharply with the Ordovician age of granitic rocks of the Turtleback Complex in the western San Juan Islands (Mattinson, 1972) and with Yellow Aster Complex, near Yellow Aster Butte (Mattinson, ibid.). Thus, the use of the name Turtleback Complex for all granitic and gabbroic rocks in the San Juan Islands must be abandoned and a new name established for the younger suite of plutonic and associated volcanic rocks. It is herein suggested that the name Fidalgo Complex be given for this rock suite, including all of the plutonic and volcanic rocks of the ophiolite. The name Turtleback Complex should be restricted to the older suite of plutonic rocks.

## CHEMICAL COMPOSITION OF ROCKS

Chemical analyses of plutonic, volcanic and sedimentary rocks of the Fidalgo ophiolite are reported in detail elsewhere (Brown, and others, 1977). The salient features of these data are as follows: (1) The volcanic rocks are consanguineous with the plagiogranite suite (including the diorite and hornblende gabbro). (2) The cumulus gabbros are not consanguineous with the volcanic rocks and plagiogranite. (3) All igneous rocks are low in $K_2O$; even the most silicic varieties ($SiO_2$ = 75%) generally have less than 0.5% $K_2O$. (4) The $K_2O$ content is variable but this variation occurs essentially irrespective of variation in other components such as $SiO_2$ or CaO. Normal crystal/melt fractionation would yield magmas with greater uniformity of variation of $K_2O$ to other elements. Thus it is suggested that $K_2O$ variation is not due to a normal process of fractional crystallization. (5) The chemical composition within individual igneous units (such as a dike or flow) is fairly uniform, suggesting that the compositions have not been substantially altered by metasomatism. (6) The layered gabbros are tholeiitic, being similar in composition to gabbros from other ophiolites. (7) The pelagic argillite is relatively rich in Mn, Co, Ni, and Cu, as are modern pelagic sediments. The overlying siltstone/greywacke unit contains less of these metals and is comparable to nearshore or epicontinental sediment.

## COMPARISON WITH OTHER OPHIOLITES

The Fidalgo ophiolite is virtually identical in age to other ophiolitic rock in the Washington Cascades (Ingalls peridotite and Tieton ophiolite, Hopson and Mattinson, 1973) and in the California Coast Ranges (Bailey and others, 1970; Hopson and others, 1975). These other Jurassic ophiolites, except for that at Point Sal, have not yet been described in detail so comparisons of the lithology and chemical compositions are not easily made. The general stratigraphic order of the Fidalgo ophiolite appears to be comparable to that of other ophiolites in this belt. Cumulus gabbros are found above ultramafic rock in virtually all ophiolites. The gabbros in all these ophiolites are tholeiitic and show chemical trends divergent from later intrusive and extrusive rocks (Brown and others, 1977; Bailey and Blake, 1974). A sheeted diabase dike-swarm, typical of Tethyan ophiolites (Moores and Vine, 1971), is absent. The dominant volcanic rock is keratophyre, not tholeiitic basalt as in the upper pillow lavas of Troodos, or the pillow basalts of Bay of Islands. Plagiogranite is abundant at Fidalgo Island, apparently more so than in other bodies of the Jurassic ophiolite belt (e.g., Point Sal, Hopson and others, 1975). However, the Canyon Mountain ophiolite (Triassic, in central Oregon) has a substantial amount of plagiogranite, comprising about 10% of the plutonic rocks (Thayer and Himmelberg, 1968). The mafic members of the plagiogranite suite at Fidalgo Island and in California Jurassic ophiolites (e.g., Page 1972) are gabbroic rocks containing hornblende instead of pyroxene. The pelagic sediment at Fidalgo Island appears identical in the field and in thin section to that in ophiolite bodies of California Coast Ranges. It is also comparable, as well as can be discerned from the literature, to pelagic sediments in some Tethyan ophiolites (Bonatti and others, 1976). However, the calcareous pelagic beds abundant in some ophiolites (Troodos, Moores and Vine, 1971; Papua, Davies, 1971) are very sparse at Fidalgo.

## ORIGIN OF THE FIDALGO OPHIOLITE

The tectonic setting of formation of the Fidalgo ophiolite is an unresolved problem. The relative abundance of plagiogranite and keratophyre in this body and the scarcity of these lithologies in suites of dredged or drilled samples from the sea floor poses difficulty for the interpretation that the complex is of oceanic origin. However, the pelagic sedimentary rocks have all the textural and chemical features of oceanic sediments which occur only far from an eroding land mass or active volcanic arc.

An argument supporting an island arc origin of the Fidalgo ophiolite can be made by comparison with Triassic rocks in eastern Oregon. A stratigraphic section of the Sparta, Oregon, area is given in Figure 4. The lower and middle parts of this section are very similar to the Fidalgo Complex. The upper part, however, differs in that (1) the volcanic section has an abundance of pyroclastic deposits and interlayered coarse sediment; and (2) the volcanic rock is overlain not by pelagic sedimentary rock, but limestone of apparent reef origin (Protska, 1963). The volcanic section contains some shale and radiolarian chert, but also locally conglomerate beds indicative of a nearby land mass. The albite granite appears to be cogenetic with the overlying keratophyre (Almy, 1977). The upper part of this section, at least the albite granite and overlying rocks, is appropriately interpreted to have originated in an island arc rather than an ocean ridge (cf., Protska, 1963; Brooks, 1976). The lower part of the complex could represent oceanic crust or plutonic rocks related to the arc volcanism.

If plagiogranite and keratophyre are associated with arc formation at Sparta, Oregon, the same may be true of the Fidalgo ophiolite. Conceivably, the pelagic sediments of the Fidalgo ophiolite could have been deposited on an extinct island arc which never developed a subaerial mass or even a substantial apron of volcanogenic debris. The answer to this problem will require further study, perhaps most profitably in the sedimentary part of the section.

In conclusion, the occurrence on Fidalgo Island of ultramafic tectonite at the base of the section and pelagic sediment at the top, indicates that the ophiolite formed either at an ocean ridge or in a poorly developed island arc on oceanic crust. It apparently did not form on continental crust or near a land mass.

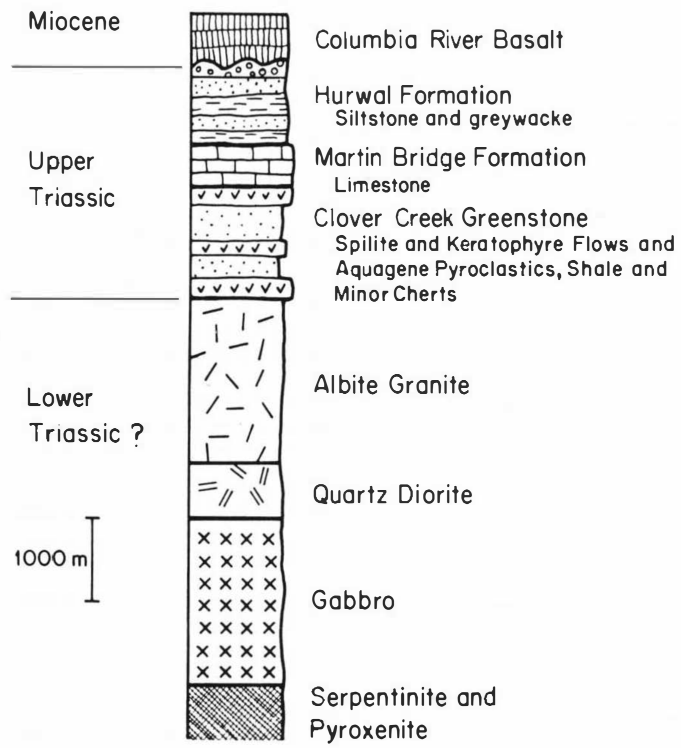



Fig. 4. Stratigraphic Section for the Sparta, Oregon area. Based mostly on data in Prostka (1963) and Almy (1977).

## REGIONAL SIGNIFICANCE OF THE FIDALGO OPHIOLITE

If the Fidalgo ophiolite is of oceanic origin, representing either an ocean ridge or an island arc, then its occurrence and that of like bodies should allow a delineation of the maximum possible westward extent of the North American plate in Jurassic times. Figure 1 shows the distribution of Jurassic ophiolite fragments in Washington, i.e., at Fidalgo Island, Blakely Island, Point of the Arches, Ingalls Peak and Tieton reservoir. Ultramafic masses, possibly related to ophiolite, on Cypress Island, the Twin Sisters and near Darrington are also shown. The occurrence of these bodies considered within the framework of other rock units of the region does not allow a simple reconstruction of the Jurassic continental margin or of the subsequent tectonic history. Interesting problems are posed by: (1) the ophiolitic material at Point of the Arches which lies far west of the other ophiolites and even west of a presumed Tertiary subduction complex; (2) the occurrence of Paleozoic schists and gneisses which appear to represent continental crust on Vancouver Island west of the Fidalgo ophiolite; and (3) the juxtaposition in the San Juan Islands of the Jurassic ophiolite with Ordovician plutonic rocks of the Turtleback Complex.

Of interest in the context of these geological complexities, which are not readily explained by a plate-tectonic model, is the recognition of major tectonic dislocations in the region. Whetton (1975) has defined a broad zone of intensely sheared rock in the San Juan Islands, which he terms melange. Major high angle faults occur at the southern end of Vancouver Island, striking east into the San Juan Islands (Muller and others, 1974). Danner (1977) has proposed that a major fault, the "Vedder Discontinuity", extends from the Fraser Valley of British Columbia into the San Juan Islands. Vance (1977) has mapped a fault, the "Orcas Thrust", which has carried Jurassic ophiolite of the Fidalgo Complex over the Turtleback Complex on Orcas Island. In the Western Cascades a 40-mile wide belt of blueschists appear to be an entirely allochthonous thrust plate (Misch, 1966). It is apparent that these structures, and perhaps others yet to be recognized, have greatly modified the plate-tectonic configuration within which the Jurassic ophiolite formed. Much interesting research remains to be done in this region before these problems will be solved.

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# OPHIOLITIC TERRANES OF CALIFORNIA, OREGON, AND NEVADA

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## ABSTRACT

Ophiolitic rocks are important components of the Sierra Nevada, California Coast Ranges, Klamath Mountains, and Ochoco-Blue Mountains, and occur sparsely in a vague zone from southwestern to north-central Nevada. The rocks occur with other volcanic and sedimentary oceanic rocks in belt-like terranes. These terranes are designated Paleozoic, Triassic, Lower or Middle Jurassic, and Upper Jurassic on the basis of the probable age of the associated ophiolitic rock. They are arranged side by side, and are sequentially younger from east to west (oceanward) although the complete sequence is not present at all latitudes. The ages assigned to the ophiolitic rocks are variously based both on isotopic ages of some igneous components of the ophiolite--the gabbro and plagiogranite--and on paleontologic ages of associated or overlying strata. Red radiolarian chert that commonly overlies or is interlayered with pillow lavas at the top of some ophiolites is proving to be particularly valuable for determining the age of an ophiolite. Also useful are the metamorphic ages of amphibolites that form a dynamothermal aureole at the base of some ophiolites, and the metamorphic ages of blueschists that are developed during regional tectonism.

Both lower and upper Paleozoic ophiolites are present in the Paleozoic terrane. The lower Paleozoic ophiolite is the Trinity ultramafic sheet of the eastern Klamath Mountains, and its probable correlative, the Feather River ultramafic body of the northern Sierra Nevada. The Trinity sheet contains gabbro and trondhjemite as old as 480 m.y. The northern part of the Trinity sheet is overlain by Orodovician sedimentary rocks, but elsewhere is overlain by Devonian and younger volcanic arc rocks. Amphibolitic rocks at the base of the Trinity ultramafic sheet are nearly 400 m.y. old, and tend to confirm that the Trinity sheet was the base of a tectonically active arc during the Devonian.

Upper Paleozoic ophiolite is present in the Kings-Kaweah belt of the central Sierra Nevada foothills and at Canyon Mountain in the John Day region of east-central Oregon. In both regions the ophiolite is in a mélange of oceanic rocks that include blocks of Permian limestone, some of which contain fossils of Tethyan faunal aspect. Plagiogranite in the ophiolite of the Kings-Kaweah belt yields an isotopic age of 300 m.y. Hornblende pegmatite that intrudes the Canyon Mountain Complex yields an isotopic age of 254 m.y., and an associated amphibolite yields 248 m.y.

Triassic ophiolite occurs in the so-called western Paleozoic and Triassic belt of the Klamath Mountains, in both the North Fork and Rattlesnake Creek terranes. The age is based on the presence of Upper Triassic radiolarians in red chert associated with the ophiolite. Isotopic ages for the ophiolitic rocks are not available, but blueschists in the belt are about 220 m.y. old. Some of the limestone blocks in the mélange contain upper Paleozoic fossils which at a few localities have Tethyan faunal affinities. Some of the mélange includes Lower Jurassic radiolarian chert and tuff. Probable extensions of the mélange are in the northern Sierra Nevada and in east-central Oregon.

Ophiolite of probable Early or Middle Jurassic age underlies island arc volcanic rocks and flysch in a well-defined belt along the western Klamath Mountains and Sierra Nevada. In the Klamath Mountains the ophiolite is mainly the Josephine Peridotite body and its extensions, which is overlain by the Rogue and Galice Formations of Late Jurassic age. In the Sierra Nevada the island arc volcanic rocks and flysch associated with the ophiolitic rocks are mainly the Logtown Ridge and Mariposa Formations. The Smartsville complex is thought to represent the upper level of an ophiolite that probably lies stratigraphically below meta-andesite correlative with the Logtown Ridge. At the latitude of the Tuolumne River, pillow lava and breccia of the Peñon Blanco Volcanics is the upper part of an ophiolite that includes nearby ultramafic and gabbroic rocks.

The Upper Jurassic (Coast Range) ophiolite is at the base of the Great Valley sequence of Upper Jurassic and Cretaceous flysch, and is in thrust fault contact with underlying regional blueschist and other rocks of the Franciscan assemblage. The ophiolite is discontinuously exposed for much of the length of the California Coast Ranges, and includes the well known Elder Creek, Red Mountain-Del Puerto, San Luis Obispo, and Point Sal localities. Correlative ophiolites are found to the north in the vicinity of Riddle, Oregon, and as far to the south as Baja California. Gabbros in the ophiolites have isotopic ages of 151 and 160 m.y., and the age of plagiogranite at Point Sal is 160 m.y. The regional blueschist at the base of the ophiolite was formed by eastward underthrusting of the Franciscan during the Early Cretaceous about 120 m.y. ago.

Fragments of ophiolite are part of the Franciscan assemblage throughout the length of the California Coast Ranges, and occur chiefly in zones of mélange. The source of the ophiolitic components of the mélange generally is not known. Some of the ophiolitic rock may be infaulted fragments of the structurally overlying Coast Range ophiolite. Others may be tectonically derived from an oceanic plate that was subducted beneath the Coast Range ophiolite and Great Valley sequence during the Cretaceous.

## INTRODUCTION

The ophiolitic rocks of the western United States occur mainly in a series of subparallel linear belts that follow a sinuous path northward thru the Pacific Coast region. In California they are exposed in the central and northern Coast Ranges and along much of the length of the Sierra Nevada, but attain their greatest development in the Klamath Mountains where they tend to form concentric west-facing arcs that lie across the California-Oregon boundary (Fig. 1).

Northward, most of the pre-Tertiary rocks are concealed by a blanket of young volcanic rocks, but important segments of belts of ophiolitic rocks are seen in the Ochoco-Blue Mountains (John Day and Baker areas) of east-central Oregon and in northwestern Washington. A few widely-spaced localities of sparse ophiolitic rocks form an interesting but vague and rather enigmatic belt in north-central Nevada. All of these regions except northwestern Washington will be reviewed briefly in this report for the purpose of documenting the ages and belt-like distribution of the ophiolitic and associated oceanic rocks where pertinent data are available. More elaborate descriptions of ophiolite of various parts of the belts are indicated by reference to the literature and are found in other papers in this volume.

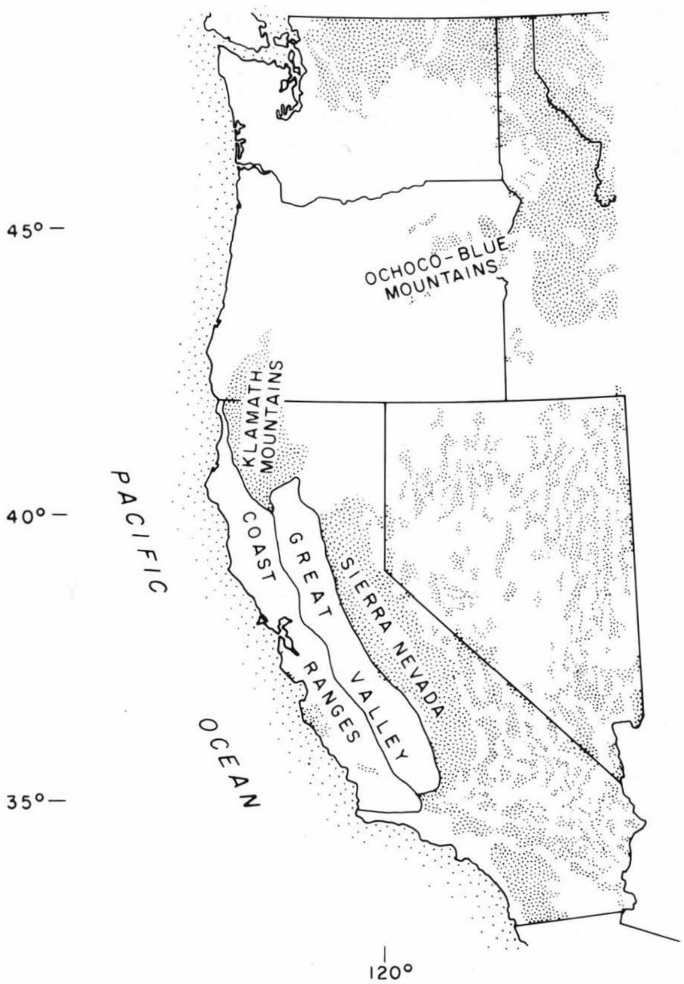



Figure 1 - Map of the Pacific Coast region showing the major geomorphic features referred to in the text. Nevadan and older rocks are indicated by stipple pattern.

The terranes of ophiolitic rocks of the Pacific Coast region generally have an obscure but obviously long and complex tectonic history. The ophiolites at most places are highly broken and dismembered, and rarely is an ophiolite sequence seen that is complete or nearly so. In most cases it is not clear whether the greatest deformation occurred before, during, or after the ophiolite became incorporated with the continental crust. In this regard, the distinction between a dismembered ophiolite, an ophiolitic mélange, or an olistostrome may not be recognized as readily in the field as in theory.

The Pacific Coast region includes the largest exposures of ultramafic rocks in North America. However, the greatest interest in the ultramafic rocks lies not in their abundance but rather in their unusual diversity in apparent age, in the geometry of their distribution, and in their structural succession. Distribution of the ultramafic rocks along the foothills of the Sierra Nevada had been established by the classic early geological surveys of the Mother Lode belt during the mid and late 1800's that closely followed the gold rush of '49. In the northern Coast Ranges and Klamath Mountains, the distribution of much of the ultramafic rock remained unknown until the area was mapped in reconnaissance during the 1950's. Essentially the only detailed studies of the ultramafic rocks were in regard to associated deposits of chromite, and these studies generally were not of a kind that related the ultramafic rocks to the regional geology. However, by the early 1960's it was recognized that the ultramafic bodies tend to mark boundaries between rocks of different sedimentary and metamorphic facies and are probably of great structural significance. During the early 1970's the ophiolite concept in its modern guise began to take hold in the Pacific Coast region, first in relation to the Coast Range ophiolite. Most of the ultramafic rocks of the Pacific Coast region are now regarded as parts of allochthonous slabs of ophiolite.

The ophiolitic rocks of the Pacific Coast are herein grouped into four generally subparallel linear terranes on the basis of their presumed age (Fig. 2). In addition to the ophiolites the terranes include large tracts of arc-related volcanic rocks and flysch, and these as well as the ophiolitic rocks occur both as mélanges and as relatively coherent blocks. The ophiolitic rocks of the terranes are successively younger from east to west (oceanward). The oldest are Paleozoic, succeeded on the west by Triassic, next by Lower or Middle Jurassic, and finally by Upper Jurassic. This complete age sequence is identified only at the latitude of the Klamath Mountains and the northern Sierra Nevada and may not exist at most other latitudes along the Pacific Coast.

The most direct available method of determining the age of formation of an ophiolite is the isotopic dating of the gabbroic and plagiogranitic rocks, but even this does not give the age of the associated ultramafic rocks nor does it take into consideration any intervals of time that may exist between different elements of the ophiolitic sequence, particularly the great hiatus that may exist between the constructional pile and the underlying metamorphic peridotite. In some instances the upper part of the ophiolite can now be dated paleontologically if the pillow lavas include or are capped by radiolarian chert or tuff. An upper limit to the age of some ophiolites can be determined by the paleontologic age of overlying or associated strata, which are commonly flysch or arc-related volcanic deposits, but these ages are interpreted to be the times of convergence of ophiolitic plates at an island arc or at a continental margin rather than the time of formation of the ophiolite. However, within broad limits these are useful data. Similarly, certain metamorphic rocks such as amphibolites and blueschists are thought to form during times of plate convergence, and in some instances these contribute to our interpretation of the age of an associated ophiolite. Other features that commonly serve to place upper limits on the age of the ophiolitic rocks are regional unconformities and intrusive plutonic rocks. Despite the obvious shortcomings and inherent vagueness in assigning ages to ophiolite belts, the belts thus dated nevertheless provide a systematic framework to guide or suggest further investigation as well as to emphasize certain tectonic aspects of the Pacific Coast region. Whether the age is the time of formation of the constructional pile at a mid-ocean ridge, or is the time of tectonic emplacement along an island arc or continental margin, the ophiolitic terranes of the Pacific Coast region seem to be sequentially younger from continent to ocean and support the idea that the continental border is built up by accretion of successive slices of oceanic crust and related rocks.

## PALEOZOIC OPHIOLITIC ROCKS

The terrane of Paleozoic ophiolitic rocks trends northwest along the southern Sierra Nevada foothills, curves northward toward the Taylorsville area of the northern Sierra Nevada region, and then swings northwest toward the eastern part of the Klamath Mountains where it constitutes a west-facing arcuate nappe. Paleozoic oceanic and ultramafic rocks are seen through windows in the broad cover of Cenozoic volcanic rocks in the Ochoco-Blue Mountains region of eastern Oregon, and also at scattered localities in north-central Nevada. The ophiolitic rocks do not represent the same part of the Paleozoic along all segments of the belt. Those in the eastern Klamath Mountains are early Paleozoic in age, as probably are some in the northern Sierra Nevada and in Nevada. Those in the southern Sierra Nevada and eastern Oregon are considered to be late Paleozoic in age.

### Klamath Mountains

The oldest known ophiolitic rocks of the Pacific Coast region are the Trinity sheet and related rocks of the eastern Klamath Mountains. The Trinity sheet is one of the largest exposures of ultramafic-mafic rock in North America, and is the easternmost of the several belts of ultramafic-mafic rock in the Klamath Mountains province. Its arcuate western boundary is virtually continuous for a length of more than 160 kilometers, and although the greatest width of outcrop is 50 kilometers, the sheet probably underlies all of the Paleozoic and younger strata of the eastern Klamath Mountains. The Trinity sheet consists mostly of tectonized harzburgite and dunite, both partly serpentinized, and is intruded by gabbro, pyroxenite, diabase, and plagiogranite. This lithic association led some geologists to consider the Trinity sheet an ophiolite (Mattinson and Hopson, 1972; Hopson and Mattinson, 1973; Irwin, 1973; and Lindsley-Griffin, 1973).

The Trinity ultramafic sheet was early recognized as a fundamental tectonic element of the Klamath Mountains owing to its widespread and consistent separation of rocks of differing histories (Irwin and Lipman, 1962). The upturned lip of the sheet is along the western side where the ultramafic rock overlies Devonian metamorphic rocks. The contact is somewhat irregular for it is folded on a large scale, in some places nearly isoclinally, and locally is offset by cross faults (Irwin, 1963; Davis and others, 1965). Fold axes tend to be parallel to the regional arcuate trend of the contact. Both compositional layering and foliation in the ultramafic rock are parallel to the contacts with the Devonian metamorphic rocks (Irwin and Lipman, 1962; Lipman, 1964). The exposed lip of the ultramafic sheet generally ranges from a few meters to a kilometer or more in thickness, and at a few places it pinches out entirely. Eastward the ultramafic sheet disappears beneath the Paleozoic strata, in the direction of its root zone. The thickness of the ultramafic sheet seems likely to increase eastward toward the root zone, if one judges from the considerable topographic relief and extremely broad exposure of ultramafic rock between the Yreka and Redding areas of Paleozoic strata.

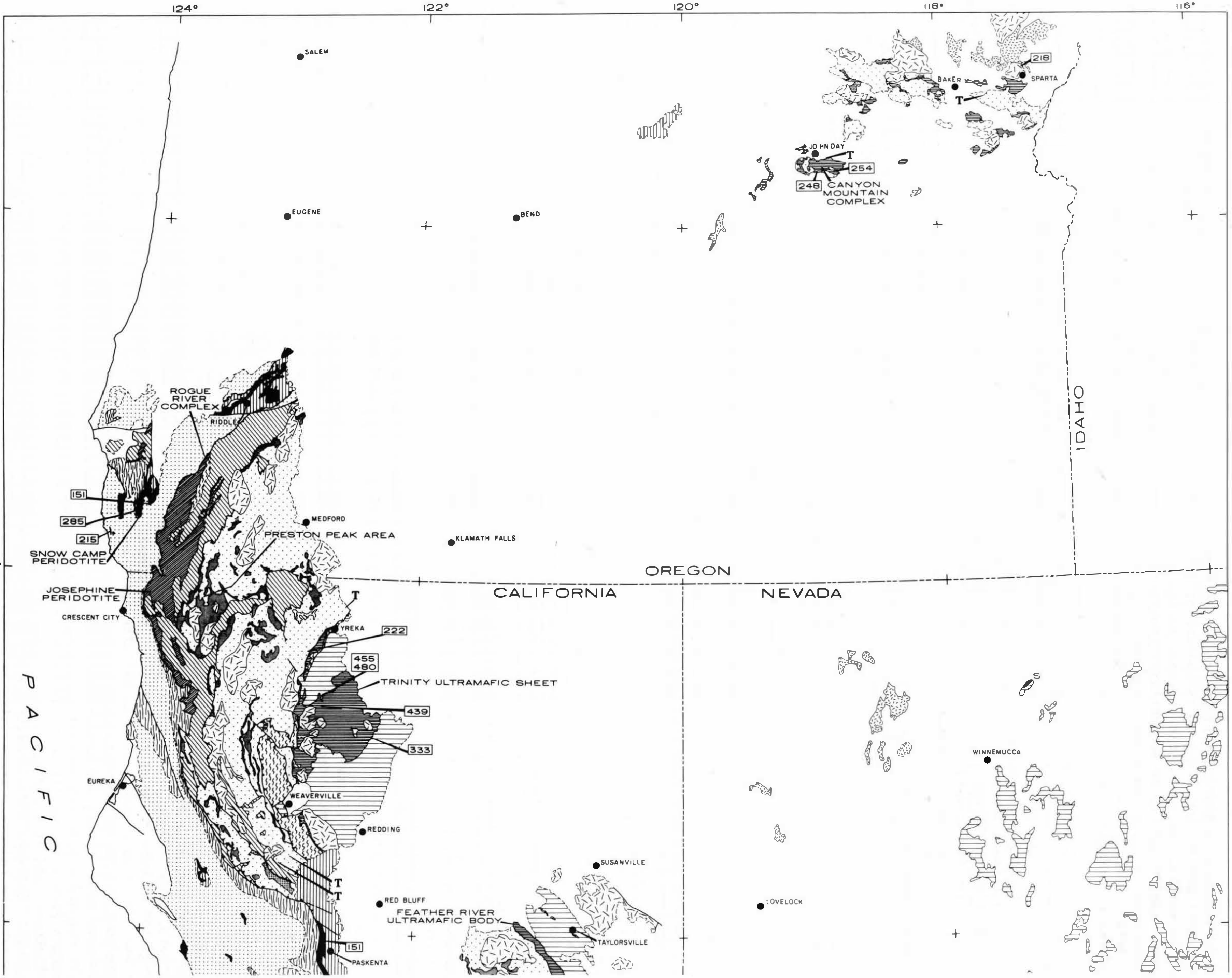
The ophiolitic nature of the Trinity ultramafic sheet has been studied in most detail near Callahan, at the south end of the Yreka area of Paleozoic strata, where the ultramafic rock is the basal part of a sequence that includes successively overlying zones of gabbro, mafic dikes, and, locally, mafic volcanic rocks (Lindsley-Griffin and others, 1974). The volcanic rocks are partly interbedded with and overlain by sedimentary rocks that have an Ordovician paleontologic age (Lindsley-Griffin and others, 1974). Early Paleozoic isotopic ages have been obtained for some of the gabbro and other rocks of the sheet (333-439 m.y., K-Ar, Lanphere and others, 1968; 455-480 m.y., Pb/U, Hopson and Mattinson, 1973), and these are compatible with the concept that the sheet represents oceanic crust on which the early Paleozoic strata of the eastern Klamath Mountains were deposited.

In the southern part of the eastern Klamath Mountains region the Trinity ultramafic sheet is overlain by the Copley Greenstone of Devonian age, which consists of intermediate to mafic volcanic flows, pillow lavas, breccias, and tuffs, and which is a minimum of 1,130 meters thick. The Copley is succeeded stratigraphically upward by a thick succession of arc-related volcanic and sedimentary rocks that range from the middle and upper Paleozoic into the Jurassic (Irwin, 1977).

The Devonian metamorphic rocks that underlie the Trinity ultramafic sheet on the west are the co-metamorphic Salmon Hornblende Schist and Abrams Mica Schist. The Salmon, a metamorphosed mafic volcanic unit, is the structurally lower of the two units. The Abrams is a dominantly metasedimentary unit that locally contains micaceous crystalline limestone. There is no stratigraphic evidence of the age of the two formations. However, Rb-Sr ages of approximately 380 m.y. are obtained from the Abrams (Lanphere and others, 1968), and K-Ar ages of 390-399 m.y. were measured by M. A. Lanphere on Salmon(?) from the Yreka area (P. E. Hotz, oral commun., 1976). These ages of metamorphism presumably represent the time that the Trinity ultramafic sheet overrode the Salmon and Abrams protoliths.

### Sierra Nevada

The Feather River ultramafic body of the northern Sierra Nevada is thought to occupy a tectonic position correlative to that of the Trinity ultramafic sheet of the Klamath Mountains (Davis, 1969). The Melones fault zone, along which most of the ultramafic bodies of the Paleozoic terrane occur, is a major structural element of the Sierra Nevada province. It trends southward for half the length of the Sierra Nevada, dividing the province longitudinally to where the fault is truncated at a low angle by the Sierra Nevada batholith. The strata east of the fault are mainly Paleozoic, including both lower and upper Paleozoic units. Those west of the fault include some upper Paleozoic but are mainly Mesozoic. The Paleozoic rocks of the southern and central part of the eastern side of the fault are generally assigned to the Calaveras Formation, and in the northern part to the Shoo Fly Formation and various other Paleozoic formations of the Taylorsville area. Although the Calaveras is considered upper Paleozoic, its age is not well documented paleontologically. Some indication of its age is shown by a pyroxene diorite pluton that forcibly intruded and contact metamorphosed the Calaveras and that has a U/Pb isotopic age of 259 m.y. (Morgan and Stern, 1977). A recent summary of the regional geologic units and possible tectonic evolution is given by Schweickert and Cowan (1975).


124°
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SALEM
EUGENE
BEND
BAKER
SPARTA
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JOHN DAY
254
248
CANYON MOUNTAIN COMPLEX
IDAHO
ROGUE RIVER COMPLEX
RIDDLE
151
285
215
SNOW CAMP PERIDOTITE
MEDFORD
PRESTON PEAK AREA
KLAMATH FALLS
OREGON
CALIFORNIA
NEVADA
JOSEPHINE PERIDOTITE
CRESCENT CITY
YREKA
222
455
480
TRINITY ULTRAMAFIC SHEET
439
333
PACIFIC
EUREKA
WEAVERVILLE
REDDING
SUSANVILLE
RED BLUFF
FEATHER RIVER ULTRAMAFIC BODY
TAYLORSVILLE
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LOVELOCK
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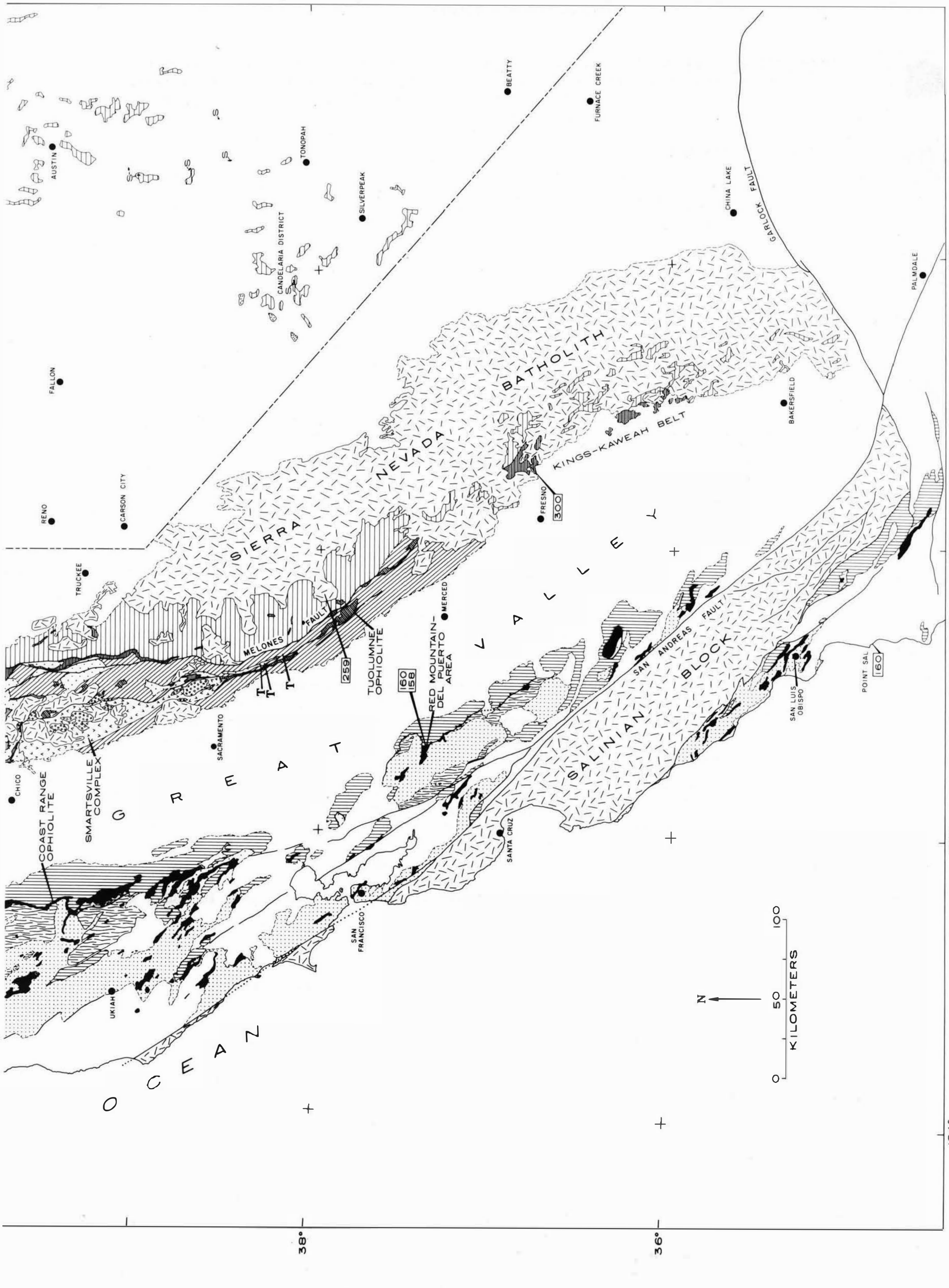
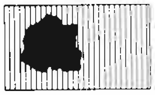
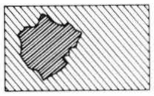



Figure 2 -- (Explanation on next page)

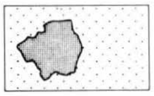
EXPLANATION FOR FIGURE 2



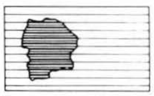
TERRANE CONTAINING UPPER JURASSIC OPHIOLITE: includes Great Valley sequence in California and Myrtle Group in Oregon: dominantly flysch of Late Jurassic and Cretaceous age: depositional on Coast Range ophiolite: ophiolitic rocks are indicated by dark irregular areas.



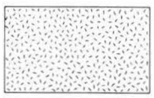
TERRANE CONTAINING LOWER OR MIDDLE JURASSIC OPHIOLITE: includes the Galice and Rogue Formations in Klamath Mountains and the Mariposa, Logtown Ridge, and related formations in the Sierra Nevada: mainly flysch and meta-andesite of a middle to Late Jurassic volcanic arc built on ophiolite: ophiolitic rocks are indicated by dark pattern in irregular area.



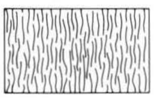
TERRANE CONTAINING TRIASSIC OPHIOLITE: includes rocks of western Paleozoic and Triassic belt of the Klamath Mountains, western areas of "Calaveras Formation" in the Sierra Nevada, and the Elkhorn Ridge Argillite and Burnt River Schist in eastern Oregon; mainly melange of upper Paleozoic and Mesozoic oceanic rocks: ophiolitic rocks are indicated by dark pattern in irregular area.



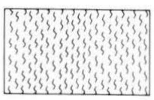
TERRANE CONTAINING PALEOZOIC OPHIOLITE: mainly arc-related lower and upper Paleozoic of eastern Klamath Mountains and northern Sierra Nevada, the Upper Paleozoic Calaveras Formation of central Sierra Nevada, and the lower and upper Paleozoic "western assemblage" of oceanic rocks of Nevada: ophiolitic rocks are indicated by dark pattern in irregular area.



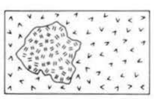
TRIASSIC VOLCANIC ARC: in northwestern Nevada and eastern Oregon



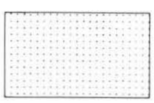
REGIONAL BLUESCHIST: includes South Fork Mountain Schist in California, Colebrooke Schist in Oregon, and associated semischists: age of metamorphism is about 120 m.y. (Early Cretaceous)



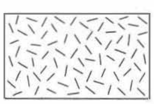
AMPHIBOLITIC METAMORPHIC ROCKS: includes Salmon Hornblende Schist and Abrams Mica Schist of Devonian metamorphic age in California, and Briggs Creek amphibolite of Mesozoic age in Oregon.



SMARTSVILLE COMPLEX: coherent, broadly antiformal block of oceanic crust: pillow lavas, with sheeted dikes, gabbro, and plagiogranite in the core of the antiform, suggest an ophiolite, but no ultramafic layer is exposed: may stratigraphically underlie Jurassic meta-andesite of Logtown Ridge Formation: gabbroic rocks are indicated by pattern in small irregular area.



FRANCISCAN ASSEMBLAGE: melange and associated regional slabs of relatively coherent graywacke and shale sequences; melange in part ophiolitic, including serpentinite, pillow lavas, keratophyre, radiolarian chert, graywacke, shale, conglomerate, minor foraminiferal limestone, low-grade (regional) blueschist, and knockers of high-grade blueschist. Coastal belt not shown.



GRANITIC ROCKS: virtually all are Mesozoic in age: Salinian block includes much metasedimentary rock of indeterminate age.

T Location of limestone containing fossils with Tethyan faunal affinity.

s Location of small serpentinite bodies in Nevada.

[15] Isotopic age referred to in text

Figure 2 - Map showing the ophiolitic terranes of California, Oregon, and Nevada. Compiled and modified after Brown and Thayer (1966), Burnett and Jennings (1962), Coleman (1972), Hotz (1971), Irwin (1977), Jennings (1975), Matthews and Burnett (1965), Ramp (1972), Rogers (1966), Saleeby (1975), Smith (1964), Stewart and Carlson (1974), Strand (1967), Strand and Koenig (1965), Vallier and others (1977), and Walker (1973).

The Feather River body is the most northerly of a number of ultramafic bodies in a belt that trends southward along the Melones fault zone to where it is truncated by the Sierra Nevada batholith. A series of isolated patches of ophiolite known as the Kings-Kaweah belt (Saleeby, 1975) represent a further extension of the ultramafic belt within the southern part of the Sierra Nevada batholith. The ultramafic bodies along the Melones fault zone and extension seem to be tectonically related, but not all are thought to represent the same part of the Paleozoic.

Somewhat similar to the Trinity ultramafic sheet, the Feather River body is flanked on the east by a terrane that includes lower Paleozoic strata, and although a continuous belt of Devonian metamorphic rocks is not present, as in the Klamath Mountains, there are a few small fault slices of possibly correlative metamorphic rocks adjacent on the west. Unlike the Trinity, the Feather River body is apparently bounded by steep faults. If truly correlative with the Trinity sheet, the Feather River ultramafic body is part of an early Paleozoic ophiolite.

The Feather River ultramafic body is described by Weisenberg and Avé Lallemant (1977) as a metamorphosed alpine peridotite and gabbro, with the peridotite having a steep northeast-trending tectonitic fabric that is overprinted by a steep northwest-trending foliation. The foliation formed during a second tectonic event that resulted in the formation of ultramafic and mafic schists that are concordant with and localized along the western boundary of the Feather River body. These geologists consider the formation of the schists and the emplacement of the ultramafic body to be syntectonic. They report that $Ar^{40-39}$ incremental heating data obtained on hornblende from the schist by John Sutter show a plateau at 236 m.y., which they regard as a cooling age related to emplacement of the ultramafic body along the western boundary fault during a collision of a Sierran-Klamath island arc with the continental margin during Permian and Triassic time.

According to Ehrenberg (1975) the Feather River ultramafic body includes peridotite and dunite. The rock consists of primary olivine, enstatite, diopside, and chromite, and secondary tremolite, chlorite, talc, anthophyllite, diopside, and serpentinite. The primary pyroxene of the peridotite is so thoroughly altered that it is not clear whether the original rock was harzburgite or lherzolite. Ehrenberg considers the evidence for the origin of the ultramafic body to be inconclusive as to whether the body intruded at a high temperature or as a cool slab derived from the upper mantle.

The Kings-Kaweah belt of ophiolite trends along the foothills of the southern Sierra Nevada in the vicinity of the Kings and Kaweah Rivers. It is a series of discontinuous patches of ophiolitic rocks that, along with Calaveras Formation, appear to be pendants or remnant inclusions in the Sierra Nevada batholith. The ophiolite includes serpentinized tectonitic dunite and harzburgite with metabasite dikes, overlain by cumulus pyroxenite, troctolite, anorthosite and gabbro, a diabase dike complex, albitite, hornblende tonalite, pillow basalt, keratophyre, chert, and marble (Saleeby, 1974). In the Kings River area, several tectonic blocks as large as 17.5 kilometers in length show a complete ophiolite sequence, but southward along the belt the ophiolite blocks decrease in size and most of the belt becomes a mélange with a serpentinite matrix (Saleeby, 1975). Concordia plots (U/Pb on zircon) on plagiogranite from the ophiolite have upper intercepts of 300 m.y., which is taken as a primary crystallization age that approximates the age of formation of the ophiolite at a spreading ridge (Saleeby, 1976). The Calaveras Formation in this region is described by Saleeby (1977) as an eastward-dipping succession that consists of chert and argillite with limestone olistoliths, grading upward into massive white quartzite with limestone layers, in turn grading upward into silicic volcanic rocks. He states that the limestone olistoliths contain Permo-Carboniferous fossils and that the upper limestone layers contain Triassic and Early Jurassic fossils.

Nevada

Serpentinite crops out at widely-spaced localities along a narrow zone that trends north-south through north-central Nevada. The serpentinite bodies are few in number and are small in comparison with those of California and Oregon, the largest being no more than a kilometer or two in length. The most northerly reported occurrence of serpentinite in Nevada is at the north end of the Hot Springs Range (R. Willden, in Ross, 1961, p. 29). Southward, serpentinite bodies occur at several localities in the Toiyabe Range, as well as at the western edge of the Toquima Range, and at Monarch Ranch ruins (F. J. Kleinhampl, oral commun., 1974; F. G. Poole, written commun., 1975). The occurrence at the western edge of the Toquima Range may be the same as that reported by Ferguson (1924, p. 40)--perhaps the earliest reference to serpentinite in Nevada. The most southerly reported occurrence of serpentinite is in the Candelaria district. There the structural and distributional trends are east-west, having changed from the generally north-south trend of the preceeding localities. As described by Page (1959), the serpentinite of the Candelaria district occurs in a complex of rocks that includes metadolerite and metasedimentary rocks.

Although the serpentinite bodies are sparse and widely spaced, they follow the general lineal configuration of other regional lithofacies trends in north-central Nevada. Detailed descriptions of the individual occurrences are not available, but the serpentinites are described as small elongate, semi-concordant bodies that are tectonically interleaved with allochthonous lower and upper Paleozoic oceanic rocks in the upper plates of the Roberts Mountains and Golconda thrusts (Speed, 1971; Poole and Desborough, 1973), the oceanic plate presumably interacting with the continental margin during the Antler (Late Devonian and Early Mississippian) and Sonoma (Late Permian and Early Triassic) orogenies. The Paleozoic oceanic rocks are distributed in two parallel but somewhat overlapping belts, the lower Paleozoic generally to the east of the upper Paleozoic. According to Stewart and Carlson (1974) the lower Paleozoic oceanic rocks are mainly Ordovician and include the Vinini, Valmy, Palmetto, and Comus Formations that consist of chert, shale, quartzite, greenstone, and minor limestone. The upper Paleozoic oceanic rocks are Late Mississippian to Early Permian in age, and include the Havallah and Pumpernickel Formations of chert, argillite, shale, and greenstone. Both the Roberts Mountains and Golconda thrusts are commonly considered to be easterly-directed overthrusts, with western assemblages of oceanic rocks being carried over eastern assemblages of transitional and miogeosynclinal rocks (Roberts and others, 1958). The serpentinite associated with the Golconda plate may be derived either from remobilization of the Antler orogenic terrane during the Sonoma orogeny, or from newly sliced upper mantle (Poole and Desborough, 1973).

East-central Oregon

Three lithologically distinct terranes of pre-Tertiary rocks are present in east-central Oregon (Vallier and others, 1977)--an oceanic, a continental margin, and an island arc terrane. These terranes are discontinuously exposed through windows in a vast expanse of Cenozoic volcanic and sedimentary strata that separate them from exposures of somewhat similar rocks in the Klamath Mountains and north-central Nevada. The ophiolitic rocks are associated with both the oceanic and island arc terranes.

The oceanic terrane includes dismembered ultramafic-mafic ophiolitic rocks, radiolarian chert, argillite, tuff, minor limestone, and rarely epiclastic sedimentary rocks. Contacts generally are tectonic. In the John Day region the oceanic terrane includes the Canyon Mountain ophiolite in addition to a broad belt of ophiolitic mélange along the northern and western border of the complex. In the Baker region to the east, formations that are thought to be correlative with the mélange include the Elkhorn Ridge Argillite and Burnt River Schist, and similar rocks may extend into the Riggins area of western Idaho.

The oceanic terrane has been considered mainly late Paleozoic based on the age of the small limestone bodies which at several localities contain a Permian Tethyan fusulinid fauna. This fauna is different from the more typical North American fusulinid fauna found in the limestone of the continental margin terrane (Bostwick and Nestell, 1967). Recognition of the mélange character of the oceanic terrane has lessened the value of the limestone for dating the total assemblage of rocks. A Mesozoic age for the formation of the mélange in the John Day area is indicated by a block of radiolarian chert which was first reported to be Jurassic (Jones and others, 1976) but now is considered to be Triassic in age (E. A. Pessagno, Jr., oral commun., 1977).

The Canyon Mountain Complex, the principal occurrence of ophiolite in eastern Oregon, is located a few kilometers southeast of John Day and occupies a nearly rectangular mountainous area about 20 kilometers long (east-west) by 8 kilometers wide. It has been the subject of many detailed studies by T. P. Thayer, C. E. Brown, Hans Avé Lallemant, G. R. Himmelberg, and others, and has been a popular site for field excursions including the Penrose Ophiolite Field Conference of 1972. The complex consists mainly of olivine-rich peridotite and gabbro, and lesser amounts of pyroxene-rich peridotite and pyroxenite, quartz diorite, and albite granite (Thayer, 1963). The complex is truncated on the east and northeast along steep faults where it is in contact with Tertiary volcanic rocks, and is in contact with the rocks now considered mélange on the northwest, west, and south (Thayer, 1956; Brown and Thayer, 1966). The Canyon Mountain Complex probably is a large allochthonous block in the mélange. The mélange is overlain by relatively undeformed sedimentary and volcanogenic strata of the Upper Triassic and Lower Jurassic Aldrich Mountains Group (Brown and Thayer, 1966) but a depositional relation is uncertain. According to Hans Avé Lallemant (oral commun., 1977) hornblende pegmatite that intrudes the Canyon Mountain Complex has an $Ar^{40-39}$ isotopic age of 254 m.y., and an amphibolite associated with the complex has an $Ar^{40-39}$ isotopic age of 248 m.y. All pertinent data considered, the age of the ophiolite seems most likely to be late Paleozoic (Permian?).

In the Sparta area, gabbro is the most abundant ophiolitic rock, but peridotite, serpentinite, quartz diorite, albite granite, and diabase also are present (Gilluly, 1937; Prostka, 1962). The quartz diorite and the Sparta granite (plagiogranite) both have $Ar^{40-39}$ isotopic ages of 218 m.y. (Hans Avé Lallemant, oral commun., 1977). Ophiolitic rocks of the Sparta area are overlain by Upper Triassic rocks of a volcanic arc terrane (Vallier and others, 1977). The volcanic arc terrane is possibly a detached fragment of the Wrangellia terrane that is more fully exposed from Vancouver Island northward to south-central Alaska (Jones and others, in press).

UPPER TRIASSIC OPHIOLITIC ROCKS

Klamath Mountains

Ophiolitic rocks of probable Late Triassic age occur in the so-called western Paleozoic and Triassic belt of the Klamath Mountains (Fig. 3). The belt is 300 kilometers long from south to north and is generally between 40 and 80 kilometers wide. On the east the rocks of the belt are thrust beneath Devonian metamorphic rocks of the central Klamath Mountains, and on the west are underthrust by Upper Jurassic flysch and volcanic rocks (Galice Formation). The belt consists of both mélange and coherent slabs

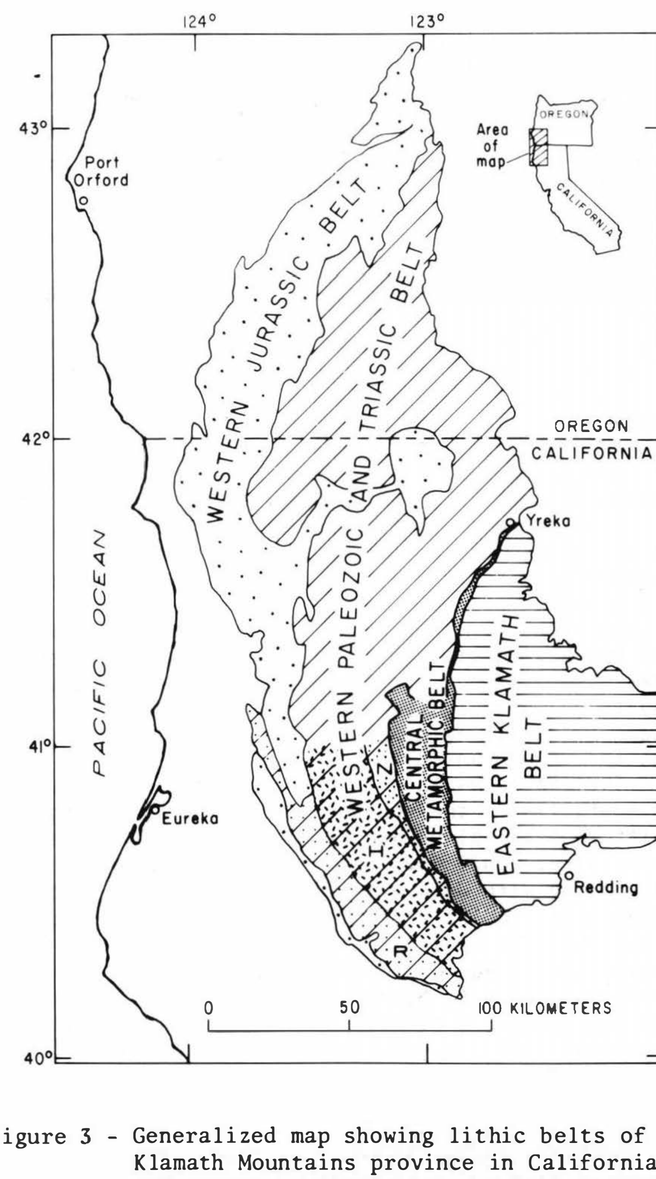



Figure 3 - Generalized map showing lithic belts of the Klamath Mountains province in California and Oregon. Subdivisions of the western Paleozoic and Triassic belt are indicated by letter symbol: N, North Fork terrane; H, Hayfork terrane; R, Rattlesnake Creek terrane.

of ophiolite and associated oceanic rocks as well as andesitic volcanic rocks. The southern part of the belt is subdivided into three parallel subunits, which from east to west are the North Fork, Hayfork, and Rattlesnake Creek terranes (Irwin, 1972). Northward, in the Salmon River area and beyond, the belt includes rocks that are correlative with the Stuart Fork Formation of Davis and Lipman (1962). In the Klamath Mountains of southwestern Oregon the belt consists of the Applegate Group. The Paleozoic age assigned to the rocks of much of the belt was based on fossils from some scattered occurrences of limestone, but now there is evidence that at least some of the limestone bodies are blocks floating in Jurassic mélange (Irwin, 1977; Irwin and others, in press). At three localities the presence of blocks of Permian limestone with fossils of Tethyan faunal affinity suggests the possibility of great tectonic dislocation.

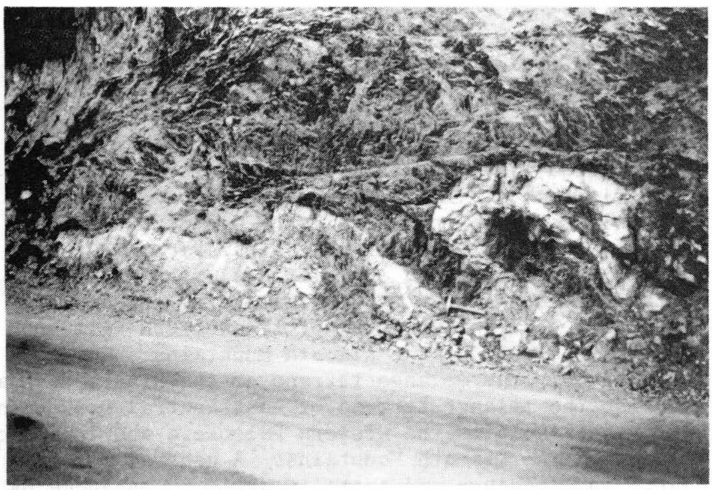


Figure 4 - A folded rodingitized mafic dike (light colored) in serpentinite of the Rattlesnake Creek terrane near Peanut, about 60 kilometers west-southwest of Redding. Note geologic hammer for scale.

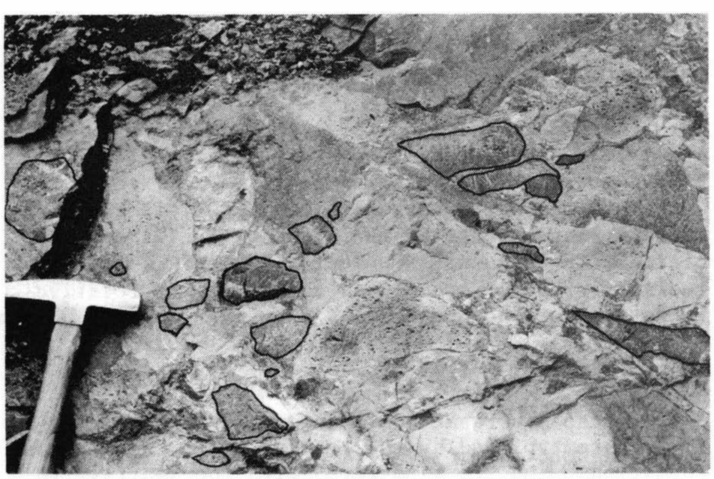


Figure 5 - Vesicular volcanic (pillow?) breccia containing fragments of red radiolarian chert (outlined). In ophiolite of Rattlesnake Creek terrane 5 kilometers west of Wildwood on Highway 36, about 65 kilometers west-southwest of Redding.

The ophiolitic rocks are the structurally lowest exposed part of the North Fork terrane, and are variously succeeded upward to the east by siliceous tuff, chert, mafic volcanic rock, small pods or lenses of limestone, phyllite, and locally pebble conglomerate. The North Fork ophiolite includes serpentinized peridotite, gabbro, diabase, pillow basalt, and red ribbon chert. Sparse sampling near the latitude of Hayfork shows that the red chert that overlies the pillow basalt contains Late Triassic radiolarians, while the cherts and siliceous tuff higher in the sequence yield Early or Middle Jurassic radiolarians (Irwin, 1977; Irwin and others, in press). Red chert from the Rattlesnake Creek terrane, dominantly an ophiolitic mélange, also yielded Late Triassic radiolarians. Although the radiolarian data are scanty, they nevertheless provide the most direct evidence of the age of the ophiolitic rock in the southern part of the western Paleozoic and Triassic belt.

Northward in the Salmon River area, the North Fork ophiolite is described briefly by Ando (1977) as the core of an antiform that is steeply overturned to

Figure 6 - Red radiolarian chert of the North Fork terrane in west-central Klamath Mountains. Exposed along Salmon River road just south of Jennings Gulch 9 kilometers northwest of Cecilville, about 55 kilometers north-northwest of Weaverville.

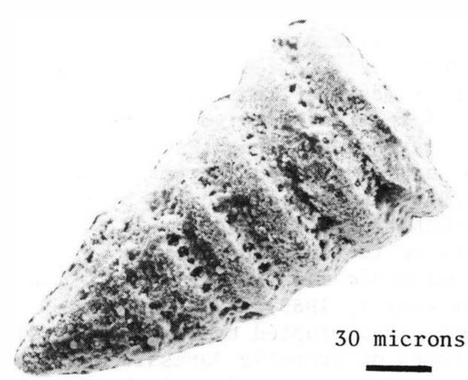



Figure 7 - SEM photo of an unnamed nassellariinid radiolarian of probable Triassic age from the chert locality (Fig. 6) near Jennings Gulch.

the west. Rocks of the North Fork and Stuart Fork terranes are on the east limb, and rocks of the Hayfork terrane on the west. A typical ophiolitic sequence of plutonic rock is not seen in the core. However, some cumulate gabbro that may be in the upper part grades upward a few meters into diabase. Dikes and sills are absent. Tectonitic harzburgite occurs structurally between the diabase and overlying pillow basalts. Ando states that these features create difficulty in comparing the North Fork rocks to "classical" ophiolites.

Glaucophane- and lawsonite-bearing schists are present (Davis, 1968) in the North Fork terrane on the east limb of the antiform south of the Salmon River, and also near the Trinity River (Cox, 1967), but isotopic dates of metamorphism are not available for them. In the Yreka-Fort Jones region to the north, and also in north-central Oregon, blueschists in presumably correlative rocks yield K-Ar isotopic ages of approximately 220 m.y. (Hotz and others, 1977). The blueschists are not thought to represent metamorphosed ophiolite. Their isotopic ages, if Middle Triassic, seem to indicate a metamorphic event somewhat older than the Late Triassic age of the upper part of the North Fork ophiolite indicated by the radiolarian chert.

Ophiolite in the Preston Peak area, just south of the Oregon border, forms the western lip of the thrust plate of so-called western Paleozoic and Triassic rocks. As described by Snoke (in press), the ophiolite consists of a basal sheet of serpentinized tectonitic peridotite and minor pyroxenite, overlain and intruded by a diabasic complex, with a metamorphic contrast between the two units. The diabasic complex is in turn overlain by metabasalt and metasedimentary rocks. In the ultramafic rock near the base of the sheet is a serpentinite mélange that contains blocks of amphibolite. An unusual aspect of the Preston Peak ophiolite is the presence of coarse jackstraw-textured talc-olivine rock in the peridotite. Snoke considers this spinifex-like texture to result from hydrothermal alteration of the peridotite rather than from a chilled ultramafic melt. The ophiolite is bordered on the west and underthrust by Upper Jurassic flysch (Galice Formation). It is overlain by volcanic and sedimentary rocks of the western Paleozoic and Triassic belt. Isotopic or paleontologic ages that are pertinent to the age of the Preston Peak ophiolite are not available, but the age probably is Triassic based on analogy with the age tentatively assigned to the ophiolite of the North Fork and Rattlesnake terranes.

### Sierra Nevada

In the northern Sierra Nevada, a broad belt of interbedded chert, phyllite, minor mafic volcanic rocks, and limestone trends generally along the western side of the Feather River ultramafic body and its southerly extensions along the Melones fault zone. These rocks are referred to the Calaveras Formation (Clark, 1976) and are considered by some to be a mélange (Schweickert and Cowan, 1975). They are thought to be tectonic correlatives of the so-called western Paleozoic and Triassic belt of the Klamath Mountains (Davis, 1969). Southward the continuity of these rocks is disrupted by large fault-slices of arc-related rocks of probable Jurassic age, and may be represented in the latitudes of the Consumnes and Mokelumne Rivers by a narrow sliver of mélange (Duffield and Sharp, 1975) consisting of rocks referred to by earlier workers as a western belt of Calaveras Formation (Clark, 1964). As with the western Paleozoic and Triassic belt, the Paleozoic age of the Calaveras west of the Feather River ultramafic body and the Melones fault zone is based on the presence of a few bodies of fossiliferous limestone. Three limestone bodies in the narrow sliver of "Calaveras" in the latitude of the Mokelumne River contain Permian fusulinids of Tethyan fauna affinity (Douglas, 1967). Similarly, limestones with fossils of Tethyan faunal affinity are present at three localities in the western Paleozoic and Triassic belt of the Klamath Mountains (Irwin and Galanis, 1976; Irwin, 1977). Another similarity to the western Paleozoic and Triassic belt is the presence of blueschist in the vicinity of the North Yuba River (Schweickert, 1976), but it apparently is in an anomalous patch of "Calaveras" rocks on the east of the main ultramafic belt rather than on the west.

The "Calaveras" west of the Feather River ultramafic body is divided into two units by Hietanen (1977, and oral commun., 1977)--the Calaveras Formation (restricted) on the east and the Horseshoe Bend Formation on the west, separated by island-arc metavolcanic rocks of the Franklin Canyon Formation. The ultramafic and other ophiolitic(?) rocks occur mainly with the Horseshoe Bend Formation. They include serpentinite, peridotite, metagabbro and hornblendite, metadiorite, and metatrondhjemite. The Horseshoe Bend Formation is interbedded volcanic, volcaniclastic, and sedimentary rocks, with lenses of recrystallized limestone that contain deformed and unidentified fossils. Although considered probably Paleozoic by Hietanen, there is no direct evidence of the age of these rocks in the Feather River region. However, the associated ophiolitic(?) rocks are here tentatively considered to be Triassic by analogy with presumably correlative rocks of the western Paleozoic and Triassic belt of the Klamath Mountains.

Not only the mélange terrane of the Sierra Nevada but also the mélange terrane of east-central Oregon in some ways resembles the western Paleozoic and Triassic terrane of the Klamath Mountains. A general resemblance in lithology and tectonic style is enhanced by the presence in Oregon of Permian limestone bodies with fossils of Tethyan faunal affinities (Bostwick and Nestell, 1967) and 220 m.y.-old blueschist (Hotz and others, in press). For the present discussion, the principal difference lies with the ophiolitic rocks for which, as previously noted, there is isotopic evidence of Late Paleozoic age in Oregon (Canyon Mountain Complex) and paleontologic evidence of Late Triassic age in the southern Klamath Mountains (ophiolites of North Fork and Rattlesnake Creek terranes).

## LOWER OR MIDDLE JURASSIC OPHIOLITIC ROCKS

Island arc volcanic rocks and flysch of Jurassic age lie in a belt that trends along the western side of the central and northern Sierra Nevada, and reappear to the northwest where they crop out along the arcuate western edge of the Klamath Mountains. Associated with these are abundant ultramafic and other rocks of probable ophiolitic affinity. Correlative rocks are not known to extend into east-central Oregon.

### Sierra Nevada

In the Sierra Nevada these volcanic rocks and flysch are typified by the Logtown Ridge and Mariposa Formations, and are associated with the Smartsville complex. The Logtown Ridge Formation ranges from fine-grained tuff to coarse breccia of pyroxene meta-andesite, and includes some pillow lava. It is about 1320 meters thick at the type locality (Clark, 1964) and is Middle or early Late Jurassic in age (Duffield and Sharp, 1975). The Mariposa Formation, as described by Clark (1964), consists mostly of slate, tuff, graywacke, and conglomerate, and locally includes a volcanic member that is lithologically similar to the Logtown Ridge Formation. Chert is rare. Thicknesses of partial sections measured at two localities are 830 and 1230 meters. The Mariposa locally overlies the Logtown Ridge Formation and is Late Jurassic (Oxfordian and Kimmeridgian) in age.

At the latitude of the Tuolumne River, Morgan (1973) describes an ophiolite that lies in a narrow zone between the Bear Mountain fault on the west and faulted Mariposa Formation on the east. The ophiolite is mainly dunite, wehrlite, gabbro, pillow lava and volcanic breccia. The gabbro intrudes both the ultramafic and volcanic rocks. The pillow lava and breccia are part of the Peñon Blanco Volcanics, which conformably underlie the Mariposa Formation elsewhere and which are probably Late Jurassic in age (Clark, 1964). Antigorite schists predominate near the base of the ophiolite. The presence of blocks of marble, slate, chert, greenstone, and amphibolite in the schists suggest a metamorphosed mélange. Morgan interprets the ophiolite as being thrust onto the continental margin.

The Smartsville complex is an ophiolite (Moores, 1975) that occupies a narrow wedge-shaped area about 70 kilometers long and 30 kilometers wide in the northern Sierra Nevada. It is separated from the alluviated floor of the Great Valley by a narrow band of pyroxene meta-andesite that resembles Logtown Ridge Formation, with which the ophiolite is in contact along a broad north-northwest-trending shear zone. On the east the Smartsville complex is in contact with mélange. The complex consists largely of pillow lava and sheeted dikes and is thought to be a slab of the upper level of pre-Late Jurassic oceanic crust that is surprisingly unaffected by younger tectonic events (Moores, 1975). Schweickert and Cowan (1975) also consider the complex a fragment of oceanic crust, but one that formed in an interarc or marginal basin during the Jurassic. The slab of crust is broadly antiformal along a north-northwest axis, as described by Buer (1977), and for a width of 2 to 5 kilometers along the core of the antiform nearly all of the rocks are diabase dikes that consistently strike north-northwest and dip steeply eastward. The dikes average 1 to 2 meters thick, have crosscutting relations, and are chilled against screens of plagiogranite, diorite, and gabbro near the base of the complex (Day, 1977). On the flanks of the dike complex is a sequence of pillow lavas about 1.5 kilometers thick that toward the lower contact is increasingly intruded by dikes, and as further described by Buer (1977), the pillow lavas are succeeded stratigraphically upward to the west by a 1.5 to 2 kilometer-thickness of pyroxene meta-andesitic and related volcanic rocks similar to those west of the Smartsville terrane.

### Klamath Mountains

In western Klamath Mountains, the flysch and volcanic arc deposits with which the ultramafic rocks are associated are the Galice and Rogue Formations. The rocks along the arcuate westerly-facing border of the province are underthrust from the west mainly by South Fork Mountain Schist (blueschist facies) in California and by tectonized Dothan Formation and related rocks in Oregon. On the east they are thrust beneath rocks of the western Paleozoic and Triassic belt. The Galice Formation is exposed along most of the 350 kilometer length of the western boundary of the province. It consists of slaty shale and graywacke, with some interlayers of volcanic rock, and is thousands of meters thick. Fossils indicate that the Galice is Late Jurassic (late Oxfordian to middle Kimmeridgian) in age, an equivalent of the Mariposa Formation of the western Sierra Nevada. At some places the volcanic rocks greatly predominate in the lower part of the section, where they are as much as several kilometers thick and are locally known as the Rogue Formation (Wells and Walker, 1953). The Rogue Formation is mainly pyroclastic meta-andesite but includes metabasalt and metarhyolite.

The largest exposure of ultramafic rock in the western Klamath Mountains is the Josephine Peridotite, which lies across the boundary between California and Oregon and which rivals the Trinity ultramafic sheet in size. It is exposed continuously for more than 150 kilometers along the western boundary of the province, and although the northern and southern parts are only a few kilometers wide, the width of the middle section is nearly 20 kilometers.

Exposures of cumulate gabbro, diabase, and spilite just north of the boundary between California and Oregon are described by Vail and Dasch (1977) as the mafic part of an ophiolite that includes the Josephine Peridotite. Major element compositions of the rocks are similar to those of other ophiolites, and the abundance patterns of titanium, zirconium, and rare earth elements show close affinities to oceanic ridge rocks. Slate and graywacke of the Galice Formation depositionally overlie the ophiolite, and although there is no bedded chert at the contact, minor amounts of radiolarian chert occur between pillows in the underlying volcanic rocks. Vail and Dasch interpret the ophiolite as oceanic crust that formed in a marginal basin, near a continent, behind an island arc that is represented by Rogue and Galice Formations. However, at some localities the Rogue is intruded by a multitude of diabase, gabbro, and diorite dikes, and these may represent a sheeted-dike part of an ophiolite (Ramp, 1975).

Rocks of the region along the northern part of the Josephine Peridotite are called Rogue River Island Arc Complex by Garcia (1976), who describes the complex as consisting of three units that are separated by steep northeast-trending faults. The western unit is essentially the Josephine Peridotite consisting of tectonized harzburgite and dunite, clinopyroxenite, lherzolite, troctolite, banded anorthosite and gabbro, and quartz diorite. These rocks were folded and metamorphosed prior to intrusion of 140 m.y.-old granodiorite. The middle unit, Briggs Creek amphibolite, is in fault contact with the ultramafic-mafic unit. It consists of a 3.5 kilometer width of folded and metamorphosed mafic igneous and sedimentary rocks that include bedded chert with manganese deposits. In fault contact on the east is a 5 kilometer thick section of predominantly intermediate to silicic volcanic rocks of the Rogue and Galice Formations. Garcia interprets the volcanic rocks as an island arc deposit based on their composition and dominantly pyroclastic nature, and he suggests that the ultramafic-mafic and amphibolite units represent fragments of oceanic lithosphere on which the volcanic arc was built.

## UPPER JURASSIC OPHIOLITIC ROCKS

### The Coast Range ophiolite

The youngest of the belts of ophiolitic rocks under discussion is the Coast Range ophiolite. It is exposed mainly along the Coast Ranges of California between the latitudes of the south end of the Klamath Mountains and the Transverse Ranges, but important exposures are present to the north in southwestern Oregon and as far to the south as the Vizcaino Peninsula of Baja California (Jones and others, 1976). The Coast Range ophiolite is the most thoroughly studied and perhaps the best understood in regard to structural position and regional tectonic development of any of the major divisions of ophiolites of the Pacific Coast region. It was described in broad outline and named in a now classic paper by Bailey, Blake, and Jones (1970)--a paper that sparked much additional interest in the study of ophiolites. The Coast Range ophiolite is now considered by most workers to be the basal part of a regional thrust plate of Great Valley sequence flysch, beneath which the Franciscan assemblage of the Coast Ranges is thrust (subducted) eastward.

Most of the ultramafic rock of the Coast Range province of California is in the northern part, where the general distribution was mapped in reconnaissance during the 1950's (Irwin, 1960). At that time the Franciscan "eugeosynclinal" rocks that made up most of the northern Coast Ranges were generally thought to be "basement" on which the Late Jurassic and Cretaceous flysch of the Great Valley sequence were deposited. However, the reconnaissance mapping showed that most of the ultramafic rock occurs as tabular masses that separate Franciscan rocks from flysch of the Great Valley sequence. The reconnaissance also resulted in the discovery of additional fossils, and owing to these and other factors it became apparent that the Franciscan rocks are the same general age as the Great Valley sequence rather than older (Irwin, 1957). To explain the distribution of the ultramafic rock, the two facies of coeval stratigraphic units, and other relations, it was postulated that the Great Valley sequence was thrust on top of Franciscan rocks along a great low-angle fault with a root zone to the east, and that the ultramafic rock intruded along the fault (Irwin, 1964). Along this great fault, later named the Coast Range thrust (Bailey and others, 1970), Great Valley sequence was considered to be thrust over Franciscan as much as 80 kilometers west of the main trace of the fault (Irwin, 1964). With the advent of the plate tectonic concept in the late 1960's, the oceanic aspect and blueschist content of the Franciscan quickly led to the interpretation of the Franciscan-Great Valley interface as a subduction zone along a convergent plate boundary. This raised the possibility that the Coast Range thrust may have resulted in much greater crustal shortening than the minimum 80 kilometers previously envisioned.

The Coast Range ophiolite, as described by Bailey and others (1970), is depositionally overlain by Upper Jurassic mudstone, sandstone, and conglomerate of the Great Valley sequence, and at the base is in fault contact with Franciscan rocks along the Coast Range thrust. They state that typically the lowest part of the ophiolite is pyroxene-bearing peridotite that locally includes dunite and minor pyroxenite and that generally is not more than 1500 meters thick. The peridotite gives way upward to a layered complex consisting of norite, gabbro, anorthosite, hornblendite, and minor trondjemite. These rocks are overlain by a 600- to 1500-meter thick section of volcanic rocks. The lower part of the volcanic section is diabase and basalt, some showing pillow structure. Keratophyre or quartz keratophyre locally predominate in the upper part of the volcanic section, which tends to be breccia and tuff. Chert commonly occurs at the top of the volcanic section, in some instances as silicified tuff, and in some as rhythmically bedded radiolarian chert. On the basis of 64 analyses of rocks from various localities of Coast Range ophiolite, Bailey and Blake (1974) concluded that the major element content is like that of rocks dredged from oceanic ridges, but that the data are not sufficient to distinguish among a spreading ridge, interarc basin, or root zone of an island arc, as the site of formation of the Coast Range ophiolite. According to R. G. Coleman (oral commun., 1977) the Coast Range ophiolite shows only oceanic hydrothermal alteration to zeolite or greenschist assemblages, and does not show evidence of blueschist metamorphism characteristic of some Franciscan rocks.

The foregoing generalizations regarding the Coast Range ophiolite are based mostly on relations at ten localities at various intervals along the northern and southern Coast Ranges, some localities that Bailey and others (1970) examined in the field, and others that they re-interpreted from published reports. As a result of quickened interest in ophiolites, much additional research has been done on the Coast Range ophiolite, and much of this has been at the localities described by Bailey and others (1970). Two of these localities in particular are at Point Sal and the Red Mountain-Del Puerto area, both of which are described in detail elsewhere in this volume.

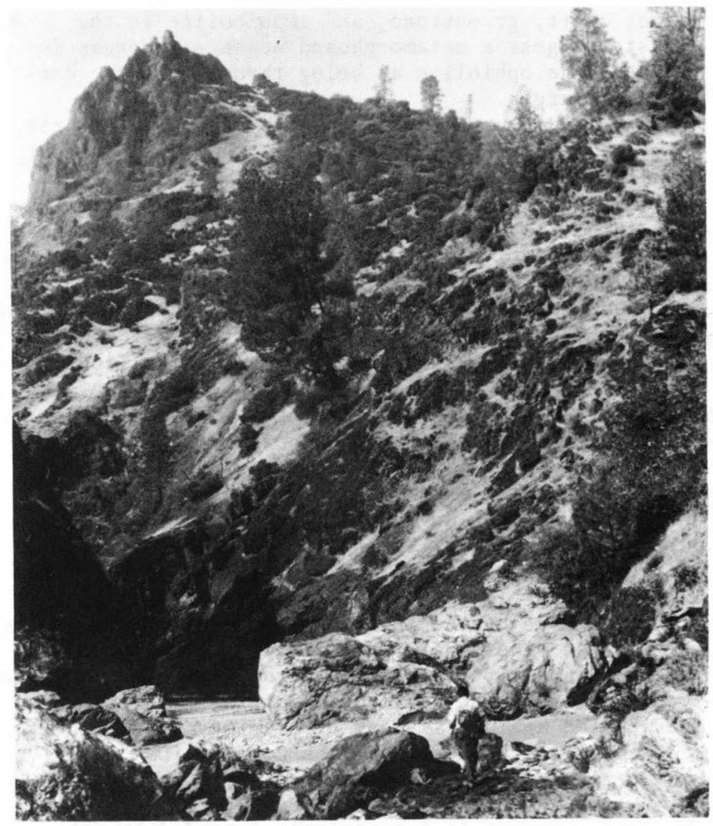


Figure 8 - Looking westward toward ophiolitic rocks at "The Gorge"--a narrow canyon where Thomes Creek cuts through the Coast Range ophiolite at the west edge of the Great Valley near Paskenta. Man in foreground is standing on serpentinite. Pillow lavas are exposed in the canyon walls.

A third locality mentioned by Bailey and others (1970) is an allochthon of Coast Range ophiolite near San Luis Obispo, west of the Salinian block. It is described in more detail by Page (1972) as a coherent synclinal remnant of mantle, oceanic crust, and overlying Mesozoic sedimentary rocks, resting on Franciscan mélange. The Mesozoic sedimentary rocks are marine shale and sandstone of the Toro Formation, a correlative of the Upper Jurassic (Tithonian) and Lower Cretaceous (Valanginian) rocks of the Great Valley sequence. Here the Toro is 600 meters thick, and at the base is 5 to 130 meters of Upper Jurassic radiolarian chert that Page considers to be clearly depositional on the pillow basalt at the top of the ophiolite sequence. The upper layer of the ophiolite

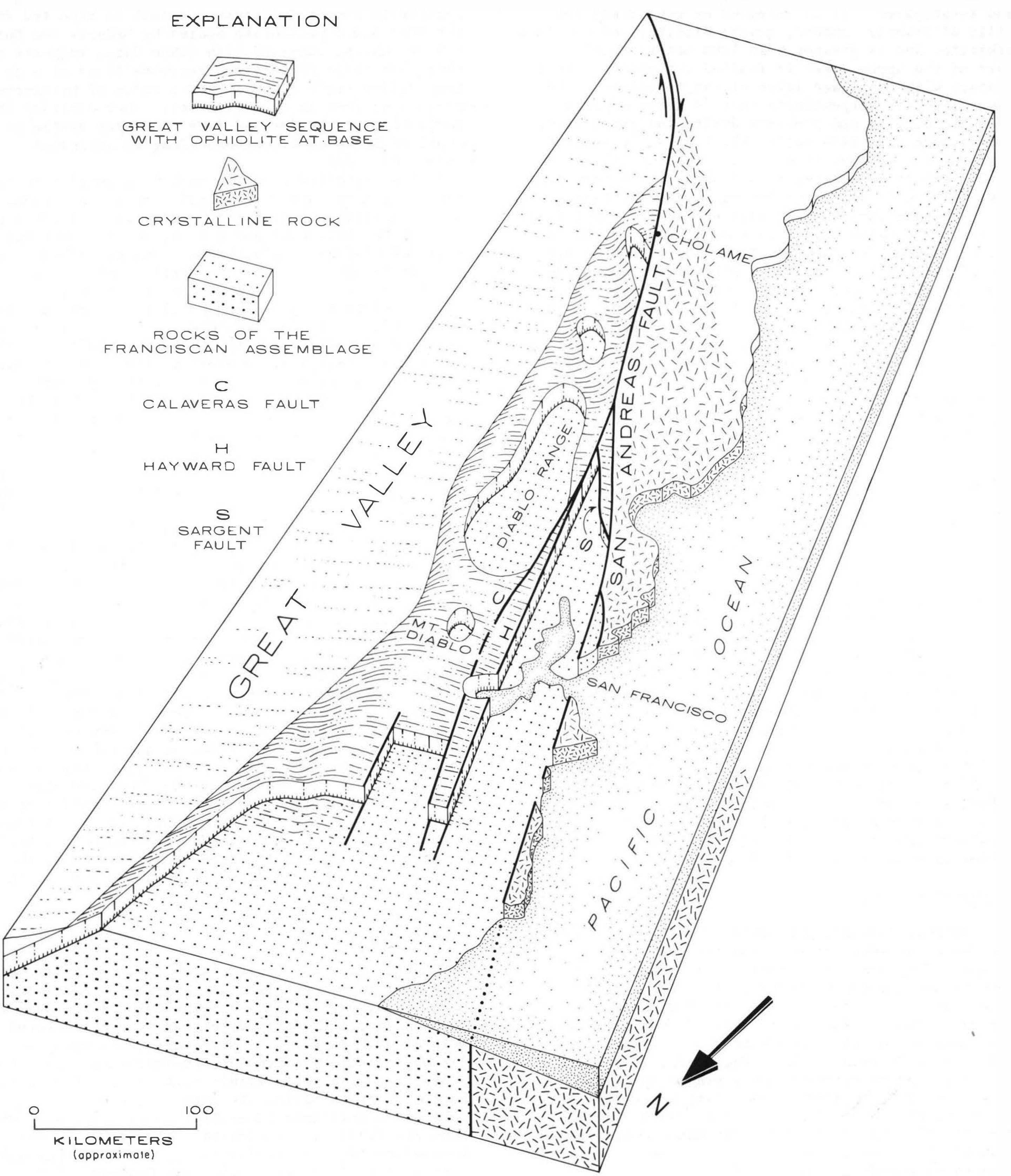



Figure 9 - Schematic block diagram of east-central California showing the gross structural relations between the Great Valley sequence, the Coast Range ophiolite, and the Franciscan assemblage. View is to the southeast along the San Andreas fault. The Coast Range ophiolite is at the base of the Great Valley sequence, and is in thrust fault contact with the underlying Franciscan rocks. Blue-schists that form a regional selvage in the Franciscan immediately beneath the upper plate of the thrust are not shown.

is not only pillow lava but includes basalt breccia and keratophyre. It is intruded by many dikes and sills of diabase, gabbro, quartz diorite, and quartz albitite, and is greater than 1200 meters thick. Part of the upper layer is faulted out where it is in contact with the lower layer of the ophiolite. The lower layer is serpentinite that is derived partly from harzburgite and probably dunite and pyroxenite, and is more than 1400 meters thick. Page visualizes this remnant of oceanic crust as floating like a raft while Franciscan mélange passed under it in the upper part of a subduction zone during plate convergence.

The Late Jurassic age assigned to the Coast Range ophiolite is based mainly on the relation of the Great Valley sequence to the ophiolite, the paleontologic age of the chert in the upper part of the ophiolite, the isotopic ages of gabbroic parts of the ophiolite, and on the tectonic relation of the ophiolite to the blueschist that regionally underlies the Coast Range thrust. The Great Valley sequence is depositional on the Coast Range ophiolite (Bailey and others, 1970), and the basal strata generally are considered to represent the middle or upper Tithonian Stage of the Upper Jurassic. However, at an ophiolite locality near Paskenta, the lowermost strata of overlying Great Valley sequence contain fossils, Buchia rugosa, that represent a middle to late Kimmeridgian Stage of the Upper Jurassic (Jones, 1975). Radiolarians from chert at a number of ophiolite localities, including Point Sal, are characteristic of Subzone 2A--the same age as the Buchia rugosa-bearing beds of the Great Valley sequence near Paskenta (Pessagno, 1977).

K-Ar isotopic ages were measured on Coast Range ophiolite at two localities by Lanphere (1971)--one locality at South Fork of Elder Creek near Paskenta, and the other in the Red Mountain-Del Puerto area. At the Elder Creek locality, gabbro dikes cut pyroxenite and are truncated at a contact with overlying mafic volcanic breccia. A sample of the gabbro about 2 meters below the breccia yielded an age of 151 m.y. Gabbro from the Red Mountain-Del Puerto ultramafic mass yielded an age of 160 m.y., and amphibole from an isolated block of cumulate peridotite yielded an age of 158 m.y. At Point Sal, Hopson and others (1975) report a concordant U-Pb isotopic age of 160 m.y. for zircon from an albite granite dike that cuts gabbro. They interpret the isotopic age to correspond to the Callovian Stage of the late Middle Jurassic.

## Southwestern Oregon

Several sizeable peridotite bodies lie west of the Josephine Peridotite in coastal southwestern Oregon. They are the Vondergreen Hill Peridotite of Koch (1966) and the Carpenterville, Signal Butte, and Snow Camp peridotites. The rocks of this area form a highly complex, dominantly upper Mesozoic terrane that consists of various thrust plates (Irwin, 1964; Dott, 1971; Coleman, 1972). The peridotite bodies probably are not fragments of a single ophiolite. None is clearly related to the Coast Range ophiolite, and although one of them, the Snow Camp peridotite, superficially resembles the Coast Range ophiolite, it probably is a dislocated thrust slice of an older Klamath Mountains ophiolite.

The peridotite bodies of southwestern Oregon are described by Coleman (1972) as ranging from massive serpentinized harzburgite, dunite, and minor orthopyroxenite, to completely sheared and flaky serpentinite. The serpentinite minerals are mainly lizardite and clinochrysotile. He considers much of the serpentinite to be a mélange that contains deformed bodies of altered diabase, gabbro, and diorite, as well as blocks of exotic rocks such as Colebrooke Schist, glaucophane schist, mafic volcanic rock, and amphibolite. The high-pressure mineral assemblage forsterite-enstatite-diopside-spinel is reported from the four named peridotite bodies by Medaris and Dott (1970), which, combined with other data, suggests to them that these peridotites recrystallized at high temperature (1100-1200°C) over a range of pressures decreasing from 19 to 5 kilobars. They consider the peridotites to be derived from the upper mantle at a depth of 50 to 60 kilometers along an ancestral spreading ridge.

The resemblance of the Snow Camp peridotite to the Coast Range ophiolite results from the presence of strata correlative with the Great Valley sequence that rest on the Snow Camp peridotite, and the presence of regional blueschist (Colebrooke Schist) in fault contact on the west of the peridotite. The strata correlative with the Great Valley sequence are depositional on the Snow Camp peridotite according to Dott (1971), but are considered to be allochthonous in relation to the peridotite by Coleman (1972, pl. 1). Unlike the Coast Range ophiolite, the Snow Camp peridotite is intruded by "Nevadan" diorite plutons--plutons of Coleman's "Pearse Peak type"--that commonly yield K-Ar isotopic ages of approximately 145 m.y. The oldest of these is the Collier Butte Diorite of Koch (1966), with an age of 151 m.y. (Dott, 1971). In addition to the "Nevadan" intrusives, dioritic to gabbroic rocks that are considered to be part of the mafic-ultramafic complex yield K-Ar isotopic ages that are considerably older than would be expected for the Coast Range ophiolite. Of these, the Saddle Mountain pluton in the Snow Camp peridotite gave an age of 285 m.y., and a small gabbro mass in the highly sheared Carpenterville peridotite gave 215 m.y. (Dott, 1971).

An allochthon of Coast Range ophiolite and associated strata correlative with Great Valley sequence occupy a wedge-shaped area of more than 700 square kilometers in the vicinity of Riddle at the north end of the Klamath Mountains. According to Jones (1973) the Upper Jurassic and Lower Cretaceous correlatives of the Great Valley sequence, in this region called Myrtle Group, are depositional on the ophiolite. He states that the ophiolite and associated Myrtle Group are thrust over the coeval Dothan Formation that borders the allochthon on the northwest, and that the fault probably is a northern equivalent of the Coast Range thrust of California. Just northwest of Riddle, a lateritic soil developed on the ophiolite is the source of the only significant production of nickel in the United States.

## Regional blueschist

Along much of its length of outcrop in California, the Coast Range ophiolite is separated from nonschistose Franciscan rocks by a relatively narrow band of blueschist-facies metasedimentary and interlayered metavolcanic rocks. The metasedimentary rock is lawsonite-bearing quartz-albite-muscovite-chlorite schist, and the metavolcanic rock is crossite-bearing albite-chlorite-actinolite-epidote gneiss. This regionally developed blueschist is called South Fork Mountain Schist in California, in areas where the metasedimentary and metavolcanic rocks are thoroughly recrystallized (Blake, Irwin, and Coleman, 1967, 1969), and is correlative with the Colebrooke Schist of Oregon (Coleman, 1972). The South Fork Mountain Schist is thought by some geologists to grade downward from the Coast Range thrust into the nonschistose Franciscan, and to have formed by metamorphism of Franciscan rocks during subduction beneath the Coast Range ophiolite. Isotopic ages on the blueschist indicate that the metamorphic event occurred during Early Cretaceous time, about 120 m.y. ago (Lanphere, Blake, and Irwin, in press). The relations between the Coast Range ophiolite, the South Fork Mountain Schist, and the Franciscan are highly controversial issues in California geology, and for some other views of these relations the reader is referred to Hsu (1971), Maxwell (1974), and Suppe (1972).

## The Franciscan assemblage

The Franciscan assemblage consists of northwest-trending areas of relatively coherent graywacke-shale sequences that are generally separated by linear zones of sheared and relatively chaotic mixed rocks. Some of the graywacke-shale sequences contain layers of radiolarian chert and mafic volcanic rocks. The areas of coherent rock are commonly tens of kilometers long and several kilometers wide. The rocks of the coherent sequences are clearly in fault contact with the rocks of the mixed zone at many places, but in some instances the contacts seem gradational. The mixed zone includes pillow lava and other mafic volcanic rock, keratophyre, radiolarian chert, graywacke, shale, conglomerate, blocks of schist, and serpentinite. The blocks of schist consist variously of high-grade blueschist, low-grade (regional) blueschist, and amphibolite. In contrast to the isotopic ages of about 120 m.y. for the low-grade (regional) blueschist, the blocks of high-grade blueschist and amphibolite generally yield ages of 140-150 m.y. (Coleman and Lanphere, 1971). The coherency and size of the components of the mixed zone range widely. Some components are more than a kilometer in length and can be traced discontinuously along trend for a much greater distance. Many occur as blocks only a few meters in greatest dimension in a matrix of sheared shale or graywacke, or less commonly in sheared serpentinite. The term mélange was introduced in the Pacific Coast region by Hsu (1968) in reference to the zones of mixed rocks in the Franciscan, and he emphasized the extremely chaotic and allochthonous nature of the Franciscan.

Figure 10 - View northward, looking edge-on at a west-dipping metasomatic zone of pectolite-prehnite rock at the west end of Black Lassic, in the Coast Ranges 70 kilometers southeast of Eureka. The zone of metasomatized rock (light colored) is nearly a meter thick and trends diagonally across the photo, separating serpentinite in the saddle on the west (left) from sandstone and shale of Black Lassic on the east (lower right).

Most of the megafossils found in the Franciscan are from the mélange and generally range from Late Jurassic to Early Cretaceous (Blake and Jones, 1974; Irwin and others, 1974). Most of the Franciscan radiolarian chert is Late Jurassic in age (Pessagno, 1977). The only older fossils known in the Franciscan are Early Jurassic radiolarians in chert from the San Rafael Mountains near Santa Barbara, and these are similar to the fauna in the cherty tuff of the North Fork terrane of the southern Klamath Mountains (Irwin and others, in press).

In many instances the affiliation of the serpentinite, pillow lavas and other ophiolitic rocks in the general region of Franciscan is not certain. Many of the larger occurrences of ophiolite west of the main Franciscan-Great Valley sequence interface are clearly fragments of Coast Range ophiolite associated with klippen of Great Valley sequence. Some probably are fault slices of Coast Range ophiolite that have been dragged for considerable distances along faults of the San Andreas system. Others may be fragments of ophiolite younger than the Coast Range ophiolite, or, as suggested by the presence of Lower Jurassic radiolarian chert in the Franciscan near Santa Barbara, perhaps even older. Blake and Jones (1974) propose that the matrix of the Franciscan mélange is the seaward portion of the *Buchia*-bearing basal sediments of the Great Valley sequence, and that the serpentinite, greenstone, and chert in the mélange were probably derived from the oceanic crust and upper mantle that immediately underlay the sediments.

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# THE CANYON MOUNTAIN COMPLEX, OREGON, AND SOME PROBLEMS OF OPHIOLITES

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CONTENTS



## ABSTRACT

The ophiolitic Canyon Mountain Complex in northeastern Oregon is of Early Permian age. About 80 percent of the complex, which has an area of 156 km$^2$, consists of a structural block of peridotite and gabbro that is bordered on the south by a sheeted dike unit. The peridotite and gabbro are believed to form a continuous cumulate sequence that grades from harzburgite (olivine and chromite cumulates) through wehrlite and clinopyroxenite (olivine, olivine-diopside, and diopside cumulates) to gabbro (olivine-plagioclase-diopside, plagioclase-diopside-hypersthene, and plagioclase cumulates). The ultramafic rocks and lower part of the gabbro in the eastern part of the complex are tightly folded and foliated. Absence of the earliest set of folds in gabbro in the western part of the complex is attributed to its high structural position in the westward-plunging complex and to deformation of cumulate crystal mush in the floor of the magma chamber during crystallization.

The sheeted dike unit consists mostly of intrusive sheets of basalt and diabase, and albite granite and keratophyre that intergrade texturally. Excepting two small quartz diorite plugs in gabbro, dikes in peridotite and gabbro and in the sheeted dike unit strike westward, parallel to the gabbro contact. Dikes of basalt and the plagiogranite-keratophyre sequence cut each other and overlapped in time. Screens of altered gabbro occur 1.8 km out in the dike swarm; no pillow basalts or other rocks definitely of surficial origin were found in the unit. The sheeted dike unit is interpreted as the substructure of a volcanic pile that formed part of the upper Paleozoic oceanic crust of the Blue Mountain region.

Rocks of the sheeted dike sequence followed fractures that cut across all structures in peridotite and gabbro. Gabbro was metamorphosed to epidiorite (hornblende-andesine rocks), and some pargasitic hornblende pegmatites were formed by solutions related to quartz diorite. Early basalt dikes were amphibolitized and followed by albite granite; late basalt dikes cut all the other rocks, and some probably were feeders for Triassic pillow basalt flows. Intrusion of plagiogranite preceded serpentinization of peridotite.

Field relations and geochemical disparities are cited as evidence against derivation of all the rocks in the complex from a single parental magma. Olivine and diopside normative tholeiites from the Mid-Atlantic Ridge that contain high-Al chromite might be related to the parent magma of peridotite and gabbro.

The Canyon Mountain Complex is believed to have originated along a spreading ridge in the ancestral Pacific Ocean, in a part isolated from terriginous sedimentation between Early Permian and Late Triassic time. In this interval, crustal rocks represented by the Elkhorn Ridge Argillite were metamorphosed to greenschist and amphibolite facies, fragmented, and then in places incorporated in serpentinite melange. The complex is bordered on the north and west by melange, whose formation preceded island-arc andesitic volcanism in the Late Triassic and Jurassic. The Canyon Mountain Complex shed debris into Upper Triassic sediments that were deposited unconformably on and then folded into the melange during island-arc volcanism. The complex is believed to have assumed its present form and structure by Lower Permian time, although it probably was raised 5,000 to 6,000 m by faulting during the Late Triassic and Early Jurassic.

## INTRODUCTION

The Canyon Mountain Complex forms the western half of the Strawberry Range in northeastern Oregon and is named after Canyon Mountain, at the western end of the range. The complex is 17-20 km long by 8-13 km wide, and covers an area of about 156 km$^2$ (pl. IV). The complex consists of four main kinds of rocks in the following areal proportions: olivine-rich peridotite* and serpentinite, 40-45 percent; pyroxene-rich wehrlitic peridotite and pyroxenite, about 5 percent; gabbro, 30-35 percent; basaltic dikes and silicic rocks that together form a sheeted dike complex or unit, about 20 percent. These rocks are distributed essentially in three east-west belts;

*For convenience, the ultramafic rocks, which include dunite, peridotite, pyroxenite, and serpentinite, as a group will be referred to as peridotite, in contrast to gabbro. Practically all the serpentinite has been derived from peridotite, and true pyroxenite (pyroxene >90 percent) probably makes up less than 1 percent of the ultramafic rocks.

the peridotite is on the north, the gabbro in the middle, and the basaltic dikes and silicic rocks on the south. The gabbro forms the highest part of the Canyon Mountain section of the range, which is 2000 to 2450 m in altitude and has local relief of 900-1400 m. Peridotite and gabbro are well exposed in the northern slopes, which were glaciated above about 1500 m. The sheeted dike unit and border of the gabbro are revealed in precipitous exposures on the southern side of the range. The complex is covered by Tertiary volcanic rocks on the south, and faulted on the other three sides. On the east it is faulted against Tertiary volcanic rocks, and on the west it is sheared against or over serpentinite melange.

### Previous investigations and present study

The first mapping of the Canyon Mountain Complex was done in 1939, when the U.S. Geological Survey examined the chromite deposits in the region and, in cooperation with the U.S. Bureau of Mines, appraised the chromite resources (Thayer, 1940). In 1946, the author undertook systematic geologic mapping of the complex in a search for guides to chromite ore bodies, a study that expanded ultimately to regional dimensions (Brown and Thayer, 1966a, 1966b; Thayer, 1956; Thayer and Brown, 1966). In 1975, the parts of the complex covered by the Strawberry Mountain Wilderness, mainly the gabbro and sheeted dike unit, were sampled geochemically, and certain areas, especially in the dike unit, were examined in detail (Thayer and Stotelmeyer, 1977). Studies were based primarily on field mapping with aerial photographs at scales of 1:20,000-1:25,000, thin section petrography, and rock analyses. Avé Lallemant (1976) contributed petrofabric data on the peridotite and gabbro.

### Aim of the paper and acknowledgments

The aim of this paper is to describe the field relations and some petrography of the rocks in the complex, and to present some chemical data. The information has led to interpretations that differ from some generally accepted petrologic and structural theories, but that are believed to be more consistent with features seen by the author or described by others in other ophiolitic complexes.

Interpretations regarding petrogenesis and emplacement of the various rocks in the Canyon Mountain Complex have changed drastically in 30 years as work progressed, and visitors came and went. The following members of the Geological Survey contributed to the field mapping: C. Ervin Brown, G.R. Himmelberg, R.A. Loney, P.W. Guild, A.B. Griggs, E.R. Force, and R.R. Carlson. As visitors, Ian Campbell, E.D. Jackson, A. Nicolas, and J.W. Dickey, Jr., provided critical ideas and observations. Discussions with B.A. Morgan, III and B.R. Lipin were invaluable in focusing and incorporating some of the ideas into the manuscript.

## PERIDOTITE-GABBRO SEQUENCE

### The rocks

The peridotite and gabbro form a continuous lithologic sequence, which probably was emplaced into its present structural setting as a solid tectonic unit. The sequence exemplifies the harzburgite subtype of alpine peridotite-gabbro complexes (Jackson and Thayer, 1972). The peridotite is dominantly harzburgite that probably averages about 80 percent olivine, in which enstatite ranges from 10-30 percent, diopside ranges from 4-10 percent (Avé Lallemant, 1976, p. 9) and podiform chromite deposits occur. The peridotite, however, ranges in composition from dunite to wehrlite, in a pattern that appears to characterize other alpine complexes. Dunite commonly forms dikes and small lenses in harzburgite, but the larger masses occur near or with wehrlite and pyroxenite (shown together as pyroxene-rich peridotite on the map) near the transition zone to gabbro. Unlike harzburgite, in wehrlite the olivine and pyroxene show a marked tendency to segregate into lenses of dunite and pyroxenite in which diopside crystals commonly are 5 cm but may be as large as 15 cm. Olivine in dunite also may be as much as 15 cm in size. Cumulus textures and layering are shown best in wehrlite composed about half-and-half of olivine and pyroxene that are 2-5 mm in grain size.

Podiform chromite deposits that range in size from a few kilograms to 125,000 tons are scattered through the peridotite (Thayer, 1940, 1956). The chromite ore ranges from massive chromitite, in which grains of chromite and occluded olivines are centimeters across, to fine-grained low-grade layered olivine chromitite (disseminated ore) that shows relict net texture (Thayer, 1969). Nodular texture is shown in several deposits. Undeformed solid nodules as much as 3 cm in diameter occur in the Haggard and New deposit (Thayer, 1940). Olivine or serpentine after olivine forms the matrix in most of the deposits, but diopside is also present in several deposits in the eastern part of the complex. At the Celebration mine, poikilitic diopside and a little enstatite form the matrix between undeformed nodules, and diopside is intergrown with chromite in the cores (Thayer, 1969, fig. 9); some specimens, however, show gradations to an olivine matrix between nodules. Cotectic diopside and chromite are shown in the Ajax deposit and are indicated in several others. In the Ajax, diopside is intergrown with high-Al chromite in the cores of lineated nodules that are enclosed in a matrix of serpentine after olivine (Thayer, 1940, fig. 13; 1969, fig. 12). Textures and distribution of diopside, chromite, and serpentine in higher grade and more strongly lineated ore at the Chambers and some other deposits indicate original cotectic intergrowths of chromite and diopside. Cleaned chromite from deposits in the Canyon Mountain Complex shows the following general range in composition, in weight percent (Bird, 1977): $Cr_2O_3$, 31-63; $Al_2O_3$, 8-39; total Fe as FeO, 14-22; MgO, 10-16. This compositional variation is similar to that found by Greenbaum (in press) in the Troodos Ophiolite Complex, Cyprus.

The gabbroic rocks of the complex range mostly between about 30 and 70 percent bytownite or labradorite; olivine, hypersthene, and augite in various proportions; and accessory magnetite or ilmenite. Primary brown hornblende is present only in some rocks as thin reaction rims. All gradations between feldspathic peridotite, pyroxenite, anorthosite, and gabbroic anorthosite containing as much as 30 percent quartz have been found. Intense shearing of quartz-rich layers to quartz-labradorite granulite shows that the quartz is primary. Anorthosite and quartz anorthosite occur only as layers a few centimeters thick. Layering is well developed over wide areas, and in places remarkable mineralogic contrasts are shown between adjacent layers (Thayer, 1963a, p. 56). Cumulus textures are common in unsheared gabbro and are implied by mineral-graded bands in gneissic facies. Layering is most prominent near peridotite, and much gabbro near the southern border shows little primary structure. The normal gabbro ranges from about 1 or 2 mm to about 5 mm in grain size, and shows allotromorphic or granulitic textures. No ophitic or diabasic textures have been seen in gabbro. In pegmatitic dikes and irregular masses, 10-15 cm crystals are common. In most gabbro pegmatites the plagioclase is $An_{75-90}$, both hypersthene and clinopyroxene are abundant, magnetite content is low, no amphibole or quartz is present. Although pegmatites cut both peridotite and gabbro, they seem to be most abundant in highly deformed gabbro.

Numerous isolated lenses of wehrlitic peridotite occur in the gabbro apparently as much as 2 km from the main peridotite contact. They range in size from about 2 by 10 m to 150 by 275 m, show poikilitic (apparently cumulus) or coarse granular texture, are rich in olivine, and grade into gabbro through a few centimeters of pyroxenite. They are interpreted as small channel deposits analogous to those described by Myers (1976) in the Fiskenaesset Anorthosite Complex in Greenland.

Petrologic continuity between peridotite and gabbro is manifest in outcrop, and is indicated by geochemistry. Complete transitions from harzburgite through wehrlite to gabbro by interlayering, by mineral grading across layering, by variation in composition of minerals, and by interfingering of layers along strike, are magnificently displayed along the divide between Pine and Indian Creeks south of Baldy Mountain, and in ridges to the west in Pine Creek basin. The cumulus and metacumulus textures in chromite deposits label their harzburgite host rocks as cumulates or metacumulates also. The succession harzburgite-wehrlite-olivine gabbro is explained most logically by enrichment of MgO-rich gabbroic magma in CaO, $Al_2O_3$, and $SiO_2$ as the result of fractional crystallization of olivine and enstatite. The appearance of diopside before plagioclase in the cumulate sequence indicates a high ratio of CaO to $Al_2O_3$ in the liquid, but the coprecipitation of high-Al chromite (30-36 percent $Al_2O_3$) with the diopside shows that the wehrlite formed from magma rich in $Al_2O_3$. Absence of an obvious iron-enrichment differentiation trend from harzburgite to wehrlite is consistent, on a larger scale, with the progressive decrease of total iron in chromite that Jackson (1963) found through 960 m of peridotite in the Stillwater Complex, and also indicated in the Great Dyke and Bushveld Complexes. Equilibration during intense deformation and recrystallization, furthermore, would reduce or eliminate original zoning.

## Structure of the Peridotite-gabbro Block

The peridotite and gabbro together form a structural unit that originated before intrusion of plagiogranite and basaltic dikes. Deformational features that characterize the peridotite and much of the gabbro have been outlined previously by Thayer (1963a, 1969) and described in considerable detail by Avé Lallemant (1976). Such features do not affect the sheeted dikes that cut them. The overall structure and evolution of this block still are obscure and controversial in some respects.

The distribution of peridotite and gabbro, and especially of the wehrlitic rocks, differs essentially in the eastern and western parts of the complex. At the west end, a southwestward-plunging anticline is indicated by symmetrical distribution, from the center outward, of olivine-rich harzburgite, dunite, and the wehrlite-pyroxenite facies (py on map) between gabbro in Canyon and Little Canyon Mountains. In the south limb of the anticline, gabbro has been dropped perhaps 500 m against harzburgite and wehrlitic rocks by an east-west fault between Dog and Canyon Creeks. Between Dog and Dean Creeks the harzburgite-wehrlite-gabbro sequence is intact, and except for small channel-type peridotite lenses, the gabbro forms a homogeneous, although complexly folded, mass. South and southwest of Baldy Mountain, peridotite and gabbro appear to be intercalated in complex lenticular masses that strike east to northeast. In this area, peridotite and gabbro show extremely complex field relations, and according to Avé Lallemant (1976, p. 13), the gabbro shows isoclinal folding that is not present in the western part.

The differences between the eastern and western parts of the peridotite-gabbro block may be explained by an overall westward plunge of the unit as a whole, and by folding and faulting. A recent aeromagnetic survey of the complex (Case and Thayer, 1977) shows steep magnetic gradients which indicate that peridotite extends southward under the gabbro in Canyon Mountain. South of Baldy Mountain, positive magnetic anomalies over peridotite lenses are disproportionately large, and broaden into a magnetic bench or terrace 900-1200 m in width. A magnetic trough about 1 km wide slopes southwestward across the west end of the magnetic bench from the northerly salient in the peridotite-gabbro contact west of Baldy Mountain. The magnetic trough is believed to indicate a steep westward plunge of the mixed peridotite-gabbro zone south of Baldy Mountain under the southeast limb of a synclinal trough filled with gabbro. The magnetic trough and the anticline at the west end of the complex appear to plunge at similar angles in the same direction. The distribution of peridotite and gabbro south and southwest of Baldy Mountain and the magnetic bench imply a series of anticlinal folds in the peridotite-gabbro transition zone. Large tight east-west folds and fault slices, which were cross-folded along northerly axes during development of Ave Lallemant's $F_1$ and $F_2$ folds could account for the present distribution of the rocks. Westward plunge of the complex would bury peridotite in the crests of anticlines under stratigraphically higher gabbro. Much more detailed mapping than has been done so far will be needed for a realistic estimate of the relative proportions or thicknesses of gabbro and peridotite in the complex.

Deformation of the magma chamber during crystallization of peridotite and gabbro, as George (1975) has postulated in Cyprus, would explain the westward decrease in deformation of gabbro that Avé Lallemant has described, and would be consistent with peridotite-gabbro relations in the ridge south of Baldy Mountain. There, as Frank Schairer observed in 1961, "Peridotite and gabbro sure have been stirred up with a long-handled platinum spoon." Unusual plasticity of feldspathic rocks in the presence of pyroxenite and gabbro magma is implied by foliated dikes in gabbro and peridotite (Thayer, 1963a, figs. 8, 9; Avé Lallemant, 1976, figs. 4c, 4d). Although Avé Lallemant (1976, table 1) postulated a definite sequence of igneous phases, others and I would compare the formation of foliated and undeformed gabbroic pegmatites (Thayer, 1963a, figs. 3, 10) to granite pegmatites in migmatite terrains. Whether the magma was introduced, as postulated by Avé Lallemant, or formed interstitially in mushy cumulate, deformation of peridotite and gabbro in the presence of interstial liquid seems indisputable. In accordance with George's hypothesis, the stratigraphically higher and younger gabbro in Canyon Mountain would not have crystallized sufficiently to transmit shear stresses by the time the basal gabbro and wehrlite were folded. In the western end of the complex, however, much of the gabbro is foliated and is more strongly folded than Avé Lallemant implies. Demonstration of gradually decreasing deformation in gabbro toward the west, therefore, might be very difficult.

## Petrogenesis of Peridotite and Gabbro

Although prevailing theory (Avé Lallemant, 1976; Bailey and Blake, 1974; Greenbaum, in press; Jackson, Green and Moores, 1975) postulates a major petrogenetic hiatus between tectonitic harzburgite as a refractory residuum and ultramafic and gabbroic cumulates as gravitational differentiates, comparison of the Canyon Mountain Complex with the Vourinos Ophiolite and Troodos Ophiolite Complex suggests a different interpretation. In the Vourinos, Jackson, Green, and Moores (1975) found a major "unconformity" between tectonitic harzburgite and a gently folded cumulate sequence from dunite through wehrlite and clinopyroxenite to hornblende gabbro. In the Troodos, George (1975) and Greenbaum (in press) have described a similar sequence. There, tectonitic harzburgite contains lenses of chromitite in dunite, principally near the border "where they grade into apophyses of large dunite bodies interdigitated with the harzburgite" (George, 1975, p. 21). Dunite forms masses 1 km or more in width (or thickness) and grades upward into wehrlite, clinopyroxenite, and gabbro. Intense penetrative deformation dies out in the dunite; chromite ore in different parts of the Chrome Mine ranges in texture from undeformed cumulus orbicular to tectonitic schlieren-banded. Cumulus textures in wehrlite and gabbro are undeformed. Chromite in the Troodos deposits ranges from about 35 to 63 percent $Cr_2O_3$, and shows the reciprocal Cr-Al variation typical of podiform deposits (Greenbaum, in press). The principal difference between the peridotite-gabbro sequences in the Canyon Mountain, the Vourinos, and the Troodos seems to be that part of the gabbro in Canyon Mountain shows the same deformational history as the harzburgite.

If the harzburgite in these complexes is not cogenetic with the associated wehrlitic and gabbroic rocks, repetition of the same suite of rocks in many alpine complexes must be explained. The three complexes show almost identical cumulus mineral successions in ascending order: olivine, olivine-chromite, olivine-diopside, olivine-diopside-plagioclase, and plagioclase-diopside-hypersthene. Podiform ore bodies of high-Cr chromite occur with dunite in the harzburgite in all three complexes. In the Canyon Mountain and in the ultramafic belt in the Coolac District, Australia, (Golding and Johnson, 1971) the unusual association chromite-diopside-olivine can be regarded as merely bringing together two minerals that form separate cumulate layers in the lower unit of the Vourinos stratiform complex (Jackson, Green, and Moores, 1975, p. 391). Furthermore, Smith (1958, p. 98) portrayed a very close association of high-Al chromite (63.6 percent $Cr_2O_3$+$Al_2O_3$, 22 percent $Al_2O_3$) with clinopyroxenite in dunite below gabbro and above harzburgite in the Bay of Island igneous complex in Newfoundland.

The petrologic continuity from harzburgite to gabbro and absence of any magmatic structural break in the Canyon Mountain sequence suggests to me that deformation during crystallization (George, 1975) has led to a fundamental misconception in relation of tectonitic harzburgite to associated gabbroic rocks. The high olivine content, lack of cryptic zoning, scarcity or absence of diopside and plagioclase, large volume, and deformed fabrics of alpine harzburgite seem easiest to explain by processes of partial melting. The high olivine content and associated dunite, however, can be explained just as well by slow deposition of cumulus olivine and extensive postcumulus overgrowth. Jackson, Green, and Moores (1975, p. 391) have emphasized this process in the Vourinos with the statement that "The general paucity of space-filling postcumulus phases indicates that cooling took a rather long time compared with large layered complexes of similar depth and thickness...". I believe that the Vourinos, Troodos, and Canyon Mountain complexes exhibit a harzburgite-wehrlite-clinopyroxenite-gabbro cumulus differentiation trend that characterizes ophiolites. The composition of segregated chromite and absence of prominent orthopyroxene cumulates are only two of several features that set this alpine trend apart from the Stillwater and Bushveld magmatic lines of descent. More careful study and consideration of the effects of deformation during crystallization of cumulates should clarify the problem.

## THE SHEETED DIKE SEQUENCE

The sheeted dike unit is a complex of silicic and basaltic rocks; it is exposed over a length of about 13 km and a width of 0.8-4 km. The unit consists mostly of sheets of basalt or diabase, keratophyre, and plagiogranite (fig. 1).

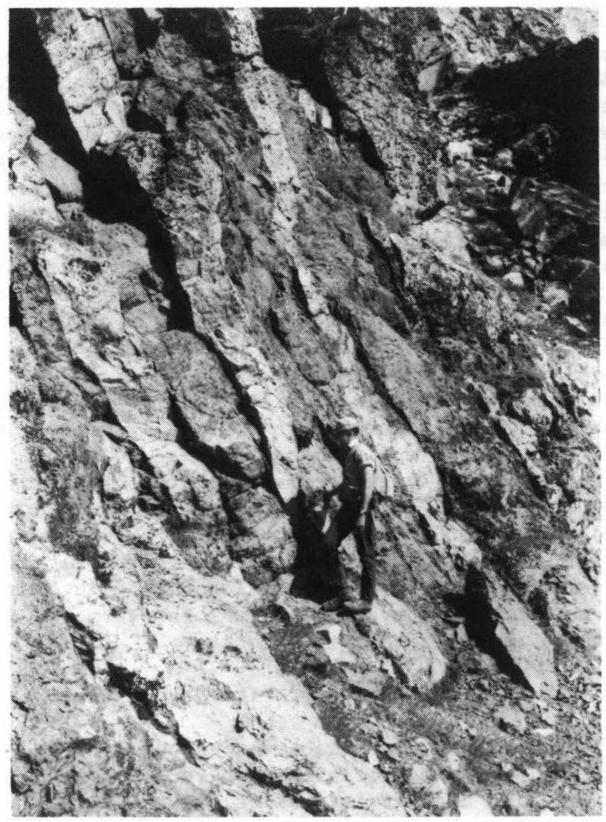

Figure 1. Dikes of albite granite (white) and basalt or diabase (gray) near the southern contact with gabbro, Canyon Mountain Complex, Oregon. At lower left, albite granite cuts across diabase, but other basaltic dikes appear to cut the albite granite.

Plagiogranite is a generic term for $K_2O$-deficient granitic rocks that range from quartz diorite to albite granite (Coleman and Peterman, 1975). The sheets range mostly from about 0.3 to 3 m in thickness, alternate singly or in groups in any order, dip 60° or more, and interfinger irregularly along strike. All three kinds of rocks are known to form dikes, but in many outcrops the order of emplacement of parallel dikes may be difficult to ascertain. Few individual dikes are traceable more than 100 m, although groups can be followed much farther. The proportions of the basaltic and silicic rocks vary widely from place to place and have not been determined. Plagiogranite, mostly albite granite, is concentrated as rather coherent units in a belt 500 to 1200 m wide next to the gabbro. The main zone of basaltic dikes appears to be about 2 km wide, but as the dikes die out to the north and south by decreasing in numbers, the limits of the zone are vague. Albite granite and keratophyre are mixed in various proportions over a distance of 2-2.5 km south of the albite granite belt. At the widest exposure of the dike unit, between Tamarack and Skin Shin Creeks, keratophyre predominates across the southern 1.5-1.8 km, and basaltic dikes are widely spaced. No pillow lavas, tuffaceous materials, or fragmental rocks definitely of surface origin were found in the dike unit.

The basaltic and silicic rocks related to the sheeted dike unit overlap chronologically. Basaltic dikes probably range in age from Early Permian to Early Jurassic. The oldest basaltic dikes were recrystallized by quartz diorite of the plagiogranite suite and converted to amphibolite. Dikes of intermediate age in gabbro were followed by albite granite, which had relatively little metamorphic effect. The youngest basaltic dikes cut albite granite, are altered only to uralite and prehnite or zeolites, and are regarded as probable feeders for Upper Triassic or Jurassic pillow basalt flows. Blocks of plagiogranite and lack of dikes in Upper Triassic graywacke show that all the plagiogranite is pre-Carnian in age. Foliation was found only in some basalt and plagiogranite dikes that cut peridotite and gabbro.

### Basaltic dikes

The basaltic dikes grade from very fine grained or aphanitic at the borders to medium grained (1-2 mm) in centers of dikes less than about 2 m wide, and to coarse grained (3-5 mm) in dikes more than about 6 m wide. Most dikes show even-grained ophitic texture, but some are porphyritic and carry equant or tabular plagioclase phenocrysts. None of the dikes now contain glass, and textures at contacts of most dikes indicate that they were finely crystalline rather than tachylitic. Some post-plagiogranite dikes, however, had glassy margins. The basaltic dikes are readily distinguished from gabbro by their chilled margins, diabasic and porphyritic textures, dark color, blocky fracture, and lack of layering and foliation.

### Plagiogranite and keratophyre

The silicic rocks probably constituted a volcanic pile in which all facies from subvolcanic intrusive rocks to near-surface dikes are represented. Some flows may be present in the southernmost part of the complex, but none were identified with certainty. The coarse-grained rocks (5-8 mm in grain size) are granitoid in appearance and range in composition from quartz diorite or tonalite to albite granite. The quartz diorite contains 30-40 percent of dark minerals, mainly hornblende, whereas the albite granite consists almost entirely of quartz and albite and only 5-10 percent of chlorite and epidote, and weathers white. All textural gradations from albite granite to porphyritic flow-banded keratophyre and quartz keratophyre are found and in many places sheets of albite granite cut keratophyre. Lacy myrmekitic intergrowths of albite and quartz characterize the fine-grained rocks; they form fringes on quartz phenocrysts and in some rocks form most of the groundmass. Patchy devitrification in some specimens shows an original high proportion of glass. Although quartz phenocrysts commonly are corroded and resorbed, many in flow-banded rocks and in holocrystalline granophyric facies show squarish outlines, indicating a high-temperature polymorph. Coarse, rude myrmekitic texture is present in some albite granite, but is not common. Various degrees of brecciation are widespread in albite granite and keratophyre. None of these rocks show textures similar to those in the average range of andesite and dacite.

## RELATIONSHIPS BETWEEN THE CUMULATE AND VOLCANIC ROCKS

### Intrusive relations and screens

Intrusion of the basaltic and silicic rocks apparently was controlled or guided by a major fracture system in the peridotite and gabbro. In general, the boundary between the sheeted dike unit and gabbro dips steeply and strikes westward parallel to the main peridotite-gabbro contact. Most dikes within the sheeted unit follow the same westward trends and dip 60° or more, either north or south. Basaltic dikes are widely scattered in peridotite and gabbro, but are most abundant near the southern border of the gabbro. Dikes that do not follow foliation or layering most commonly follow west-trending fractures or shear zones. The southern border of the gabbro is so splintered by basaltic dikes and larger plagiogranite intrusions that no definite contact can be plotted; that shown on the map is approximate, at best. East of Skin Shin Creek, screens or slivers of altered gabbro were found between dikes as much as 1.8 km south of the gabbro contact as mapped. Sheets of keratophyre between screens of gabbro and the main body of gabbro must have been dikes because of their structural position; they cannot have been surface flows.

Simple intrusive relations between albite granite, diabase, and gabbro are exemplified by the breccia dike shown in figure 2. The albite granite followed a diabase dike about 60 cm thick in layered gabbro, widened the original fracture by spalling slabs from the walls, and floated much of the diabase away. The albite granite has sharp contacts against both gabbro and diabase, and decreases in grain size from the usual 2 or 3 mm to 0.5-0.75 mm at the contacts. The diabase has been altered to blue-green amphibole and albite in which original calcic cores are indicated by distribution of amphibole and epidote. Ophitic texture is rather well preserved by the hornblende and albite.

### Metamorphism Related to Plagiogranite

Igneous and metamorphic relationships between the peridotite and gabbro as a petrologic unit and rocks related to the sheeted dike unit are complex. Two stocks of quartz diorite, both 250-300 m across, cut gabbro north and west of Canyon Mountain, and the West Fork of Pine Creek (Norton Creek) follows a gneissic migmatite zone in gabbro about 300 m wide and 2.5 km long. Hornblendic and sodic alteration caused by solutions accompanying intrusion of plagiogranite is widespread in the gabbro, and along the southern border much of the gabbro has been completely recrystallized pseudomorphically to epidiorite (hornblende-andesine rock: Thayer, 1972). In most epidiorite, the original granular gabbroic texture is preserved in hand specimen by development

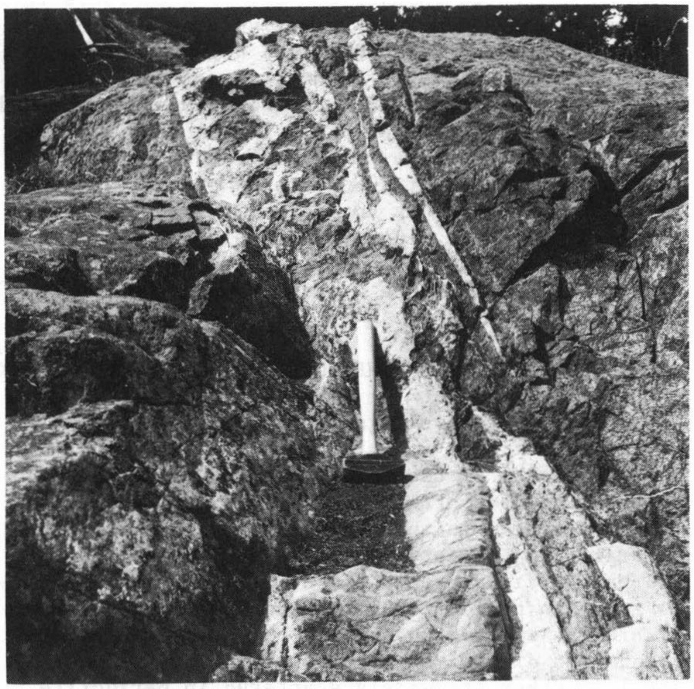

Figure 2. Breccia dike of diabase (light gray) and albite granite (white) in layered gabbro of Canyon Mountain Complex. The hammer lies on a block of diabase which extends from wall to wall. Above and below the hammer the dike has been widened by spalling of gabbro slabs from walls of the original dike.

of stubby black poikiloblastic hornblende crystals that replaced pyroxene and plagioclase. On weathered surfaces, the epidiorite has a black and chalky-white pepper and salt aspect that stands out against the warm buff, green, and cream tones of fresh gabbro.

Metamorphism by the dioritic facies of the plagiogranite is more intense than by albite granite; it has obscured some important relations of basalt against peridotite and gabbro, and revealed others. Near the plug in the cirque at the head of Dog Creek, coarsely recrystallized epidiorite is irregularly distributed through less altered hypersthene gabbro; in places, transition zones from one rock to the other are less than 10 mm wide. The feldspar in the epidiorite is andesine, and the hornblende is zoned from brownish green to green. Recrystallization of gabbro and dikes together can obscure distinctions between the two kinds of rocks. Much fresh gabbro near the southern border ranges from 1-2 mm in grain size, but, in the centers of thick diabase dikes, feldspars may be 5 mm long. Growth of poikiloblastic hornblende during recrystallization can form very similar textures in diabase and gabbro, and presence or absence of chilled contacts can be critical in distinguishing between the two rocks. My coworkers and I consistently have found chilled contacts between ophitic-textured rocks and gabbro, and have never found diabasic texture in fresh gabbro. We believe that Avé Lallemant's (1976, p. 13) description of ophitic textures in gabbro along the contact with the sheeted dike unit is mistaken.

Complete recrystallization of much gabbro destroyed the original granular textures and resulted in two kinds of new textures. In leucocratic and moderately mafic ($<40$ percent pyroxene) gabbro, hornblende forms stubby grains and slender prisms a few millimeters long, and random seriate texture imparts a dioritic appearance to the rock. In mafic gabbro, however, hornblende tends to form stubby subhedral oikocrysts several centimeters in size, and to be concentrated with feldspar in veinlets or dikelets. On a larger scale, irregular masses several meters across have been formed in which slender to stout hornblende prisms as much as 20 cm long occur in a feldspar matrix. In such masses random orientation of slender hornblendes resembles giant-grained diabase, although comb textures may occur along contacts. Some masses may have been formed by replacement of gabbroic fragments in intrusion breccias, but some form knots in apparently massive layered gabbro. Well-formed dikes, however, that range from 5 to 30 cm or more in thickness and cut across layered gabbro with sharp contacts (Avé Lallemant, 1976, fig. 3) appear to have been magmatic. The hornblende is pale brown to brownish green, weakly pleochroic, and compositionally is pargasite and tschermakitic hornblende. The feldspar apparently was andesine, but almost everywhere has been altered to a paste of more sodic feldspar, clinozoisite, prehnite, and similar minerals. These pegmatites are completely unlike the gabbro pegmatites in total absence of pyroxene or olivine and in freedom from penetrative deformation except where crossed by shear zones.

Many basaltic dikes in peridotite and gabbro have been altered to foliated amphibolite, marginally or completely. Foliation is parallel to the walls, and although mylonitization along contacts is common, chilled contacts usually are recognizable. Gabbro wall rock as a rule shows little shearing compared to the dikes, but where films or thin bands of plagiogranite are present along contacts, all three rocks tend to be sheared into banded amphibole-quartz-albite schist or mylonite. In the absence of basaltic dikes, plagiogranite and gabbro form similar mylonite along shear zones (Avé Lallemant, 1976, fig. 7). Amphibolitic basalt dikes have been found cutting hornblende pegmatite dikes.

The elongate mass that is mapped as quartz diorite along the Norton Fork of Pine Creek actually is a unique mixed zone of gabbro, plagiogranite, and basaltic dikes. There, angular to subrounded blocks of mafic gabbro and pyroxenite are enclosed in a matrix of gneissic leucogabbro and plagiogranite. The matrix shows strong flowage by bending of mesocratic and leucocratic schlieren around mafic blocks. The mafic blocks are partly hornblendized, and many schlieren of fine-grained pyroxene gabbro are rich in quartz. In places, basaltic dikes have been made almost unrecognizable by intrusion of plagiogranite along fractures and by recrystallization with the gabbro host rock. Along the strike of this zone in the southern slopes of Canyon Mountain, most structure in the gabbro has been obscured or obliterated by shearing. Presence of quartz gabbro schlieren free of hornblende suggests that plagiogranite followed a weak zone of quartz-rich mylonitic gabbro similar to quartz-rich gabbroic anorthosite in Indian Creek. The breccia dike shown in figure 2 is at the edge of this mixed zone. No gneissic structures like those that dominate this zone were seen in the sheeted dike unit.

## SERPENTINIZATION OF PERIDOTITE AND RODINGITE ALTERATION

Just as most peridotite in the Canyon Mountain Complex is more or less serpentinized, most gabbro is more or less altered to minerals of the rodingite suite: hydrogrossular, prehnite, clinozoisite, serpentine, chlorite, tremolite, and secondary diopside. This alteration was superimposed on the amphibolite and epidiorite-type metamorphism associated with plagiogranite. Much epidiorite is laced with veinlets of prehnite, and prehnite and clinozoisite replace both plagioclase and hornblende. At the Bald Eagle prospect (plate IV) serpentinized dunite matrix in chromite a few millimeters from the contact of a foliated amphibolitic basalt dike shows no sign of metamorphism. Prehnite replaces hornblende and feldspar in the dike, and serpentine pseudomorphs of plagioclase phenocrysts occur in the chilled border. Absence of tremolite or talc in the serpentine and rodingite alteration are interpreted to mean that the olivine in chromitite was fresh when the dike was amphibolitized. High temperature (600°-700°) during amphibolitization of basalt would account for stability of olivine in chromitite. Complete alteration of other amphibolitic and unsheared basaltic dikes to rodingite in serpentinite and peridotite provide widespread evidence that the main serpentinization occurred after intrusion of the plagiogranite and most of the basaltic dikes.

## GEOLOGIC SETTING AND AGE OF THE COMPLEX

The Canyon Mountain Complex is believed to represent the basement under a cover of Paleozoic and Mesozoic oceanic volcanic and sedimentary rocks that are exposed over a distance of 325-350 km in northeastern Oregon and western Idaho. The supracrustal rocks in the vicinity of the complex are of two contrasting assemblages separated by a profound regional unconformity. Rocks of Early Permian age, believed to be equivalent to the Elkhorn Ridge Argillite, consist of basaltic and keratophyric flows and volcaniclastic rocks, argillite, chert, and widely scattered limestone lenses (Vallier, Brooks, and Thayer, in press). These have been intensely folded, metamorphosed to green schist and amphibolite facies, and tectonically incorporated in places into serpentinite melange. From Late Triassic to Late Jurassic time about 20,000 m of graywacke, shale, conglomerate, and tuffs were deposited unconformably across the Permian rocks in a major basin southwest of Canyon Mountain (Brown and Thayer, in press). Thousands of meters of andesitic tuff and tuffaceous graywacke indicate a concentration of volcanic activity west or northwest of the present complex. Major angular unconformities within the sedimentary sequence reflect folding and faulting on a large scale. The deformation and volcanism together point to an island-arc environment that extended eastward to the vicinity of the present Idaho batholith.

These rocks are metamorphosed only to prehnite-pumpellyite metagraywacke facies (Brown and Thayer, 1963). The only satisfactory explanation of the regional unconformity between Lower Permian and Upper Triassic marine rocks is the formation of the two rock groups in completely different settings; a Late Triassic island arc was formed on older oceanic crust that had been isolated from terrigenous sedimentation for 40-50 million years. The only local evidence of sedimentation during the Early Permian to Late Triassic interval is a block of radiolarian chert of Middle Triassic age (Pessagno, written commun., 1977) in melange 5 km southwest of Miller Mountain. An area in the ancestral Pacific Ocean far from the present continental margin could have provided these conditions (Dickinson, in press).

The western half of the complex is surrounded on the west and north by a serpentinite melange which is exposed over a 13 by 19 km area west of Canyon Creek and north of the John Day River (Brown and Thayer, 1966a). The melange is exposed best and has been mapped in most detail in the vicinity of Miller Mountain (plate IV), and is named accordingly. The melange consists of rock units that range in size from a few meters to 3 km suspended in a matrix of serpentinite. Just west of the map area in the Mount Vernon quadrangle (Brown and Thayer, 1966b), jumbled blocks made up of Paleozoic argillite, chert, keratophyre, greenstone, and shale in various proportions occupy an arcuate belt 13 km long. Juxtaposition of pillow basalt, amphibolite and plagiogranite blocks shows that metamorphism preceded generation of the melange. The Triassic rocks in Miller Mountain are mostly unsheared pillow basalts, probably a few hundred meters thick, that overlie graywacke, mudstone, and breccia-conglomerates rich in clasts of albite granite. These rocks are unconformable on the melange, but they also have been folded steeply and faulted down into it. The melange contains remarkably few blocks of gabbro. Small pods of chromite are widely scattered in the serpentinite.

The contact at the west end of the complex seems to be a fairly well-defined shear zone that is followed by Canyon Creek for about 5.5 km. Pyroxenite and gabbro form steep slopes east of the creek, and landslides on melange come down to it on the west. West of Little Canyon Mountain where the creek crosses the gabbro, the contact of gabbro on melange dips eastward very irregularly at angles between about 25° and 45° over a vertical range of 300 m. Where the contact swings eastward across the mouth of the canyon just south of Canyon City, the gabbro dips 70-80° southward over serpentinite, in which there are prominent large blocks of amphibolite and phyllite. Between Canyon and Pine Creeks peridotite adjoins melange and the contact is indeterminate because of marginal serpentinization and shearing.

Radiometric dates on plagiogranite, hornblende pegmatite, and amphibolitic dikes in the Canyon Mountain Complex and on amphibolite inclusions in the melange are consistent with ages indicated by fusilinids in the sedimentary rocks. Hornblende from pegmatites has yielded K/Ar dates of 240-250 m.y., and K/Ar ages between about 186 and 234 m.y. were determined for hornblende from amphibolite blocks in melange, amphibolitic basalt cutting gabbro, and the quartz diorite in Dog Creek (Lanphere, written commun., 1973). Fusilinids from limestone blocks in the melange west of Dog Creek are of Early Permian age (Nestell, written commun., 1977). Slide breccias of albite granite and epidiorite in rocks of known Carnian Age in the Aldrich Mountains show that the Canyon Mountain Complex, and the Miller Mountain melange as well, were exposed to erosion by Late Triassic time (Brown and Thayer, in press).

## History of the Canyon Mountain Complex

The history of the Canyon Mountain Complex can be divided into three phases: 1) crystallization and penetrative deformation of the peridotite and gabbro; 2) intrusion of plagiogranite and basaltic dikes and related metamorphism; and 3) emplacement into oceanic crust, and deformation in an island-arc setting (Table 2). The first two phases probably took place during Permian time. The regional setting and geologic environment, according to the present consensus, point to origin of the complex at a spreading ridge in the Pacific Ocean. The peridotite and gabbro must have crystallized in a magma chamber at some depth less than 30 km, presumably near the crust-mantle boundary in an active tectonic environment. After crystallization and contemporaneous

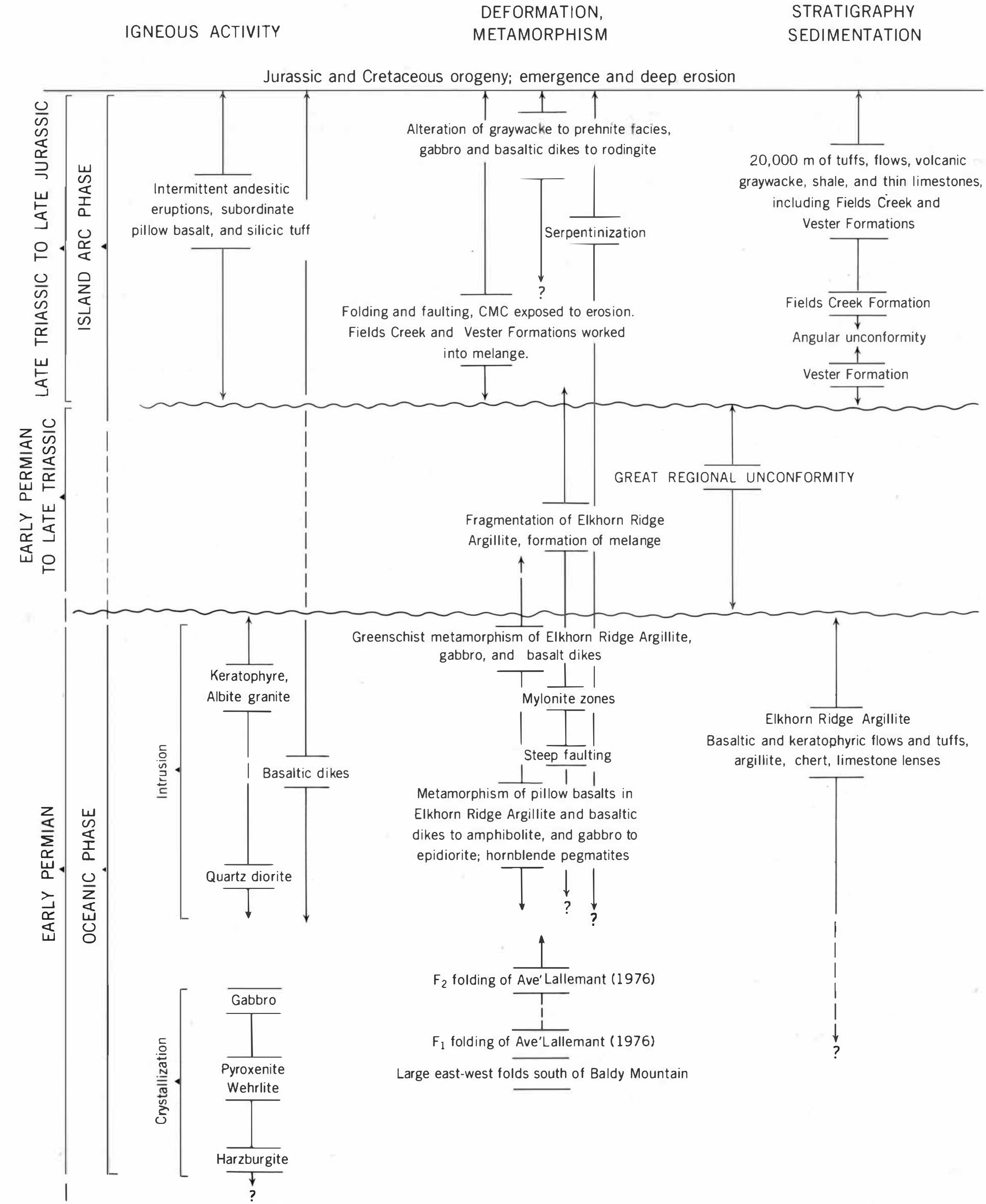


**Table 2.** Diagrammatic outline of the geologic history of the Canyon Mountain Complex and related rocks. Arrows indicate the approximate duration of processes or events described between brackets.

deformation as a cumulate body, the present peridotite-gabbro unit probably was raised as part of a fault block like those along the Mid-Atlantic Ridge. The southern border of the gabbro is interpreted as originally having been a steep fault that was obliterated by intrusion of plagiogranite and dikes. The ranges of textures in plagiogranite and metamorphism of basaltic dikes show that the block was incorporated into the subcrustal structure during active volcanism. The plagiogranite-keratophyre suite is believed to represent the subvolcanic structure of a silicic volcanic pile on the Permian sea floor, and the basaltic dikes are interpreted as feeders for basalt flows. Both kinds of rocks are abundant in the Elkhorn Ridge Argillite. The fractures that guided basaltic dikes into peridotite and gabbro were related to the boundary fault, and mylonite along them reflects adjustments during early stages in evolution of the oceanic floor. The length of the time interval between crystallization of peridotite and gabbro and intrusion of the sheeted dike sequence is one of the unsolved riddles of the ophiolite assemblage, but the consensus seems to be that it probably was rather short.

The breakup of the Permian crust and formation of melange probably continued over a long time. Boulders of gabbro and quartz diorite in Lower Permian conglomerate along the Snake River (Vallier, Brooks, and Thayer, in press) indicate that the oceanic crust was faulted on a large scale while it was being formed, just as oceanic crust now is faulted along slowly spreading ridges. A block of radiolarian chert believed to be of Middle Triassic age (Pessagno, written commun., 1977) suggests that the Miller Mountain melange was formed along an intraplate suture that developed as a precursor of Late Triassic and Jurassic island-arc andesitic volcanism. The larger structural relation of the complex to the melange, and the distance it may have moved as part of the melange is unknown. The rarity of gabbro in the melange is puzzling in view of its importance in the complex. Great abundance of large angular albite granite clasts in Triassic(?) conglomerate under pillow lavas in Miller Mountain, only 4 km from the sheeted dike unit, suggests uplift of the complex along the Canyon Creek fault. This uplift might have coincided with isoclinal folding of the Vester Formation in the Aldrich Mountains during Carnian time before deposition of the Fields Creek Formation (Brown and Thayer, in press). Movement presumably also occurred along the fault during folding of the conglomerate and pillow lavas in Miller Mountain, which have been mapped as part of the Fields Creek Formation (Brown and Thayer, 1966b). After major deformation of the Fields Creek, of Carnian Age, the Canyon Mountain Complex became part of the hanging wall block of the Aldrich Mountain fault, which has an estimated vertical displacement of 5,000-6,000 m. Serpentinization of peridotite probably has progressed intermittently since Permian time, but the main rodingite alteration in gabbro and basaltic dikes is believed to have taken place during the Jurassic, when graywackes of the Aldrich Mountains Group were metamorphosed to prehnite-pumpellyite facies (Brown and Thayer, 1963; Thayer and Himmelberg, 1968).

Although the Canyon Mountain Complex must have undergone substantial internal adjustment during two episodes of folding and faulting, in Oligocene(?) and Pliocene time, intrusion Upper Jurassic(?) quartz diorite nearby had the most obvious effects. Then, gabbro and serpentinite in Little Canyon Mountain were altered extensively to ankeritic carbonate, and gold-bearing quartz-calcite veins were formed. No large faults of Tertiary age, except the John Day and Indian Creek faults, are known to cut the complex. Its present topographic relief is believed to have resulted from lower and middle Pliocene folding and uplift along the John Day fault.

## THEORIES OF OPHIOLITE MAGMATIC PARENTAGE

### The One-Magma Theory

Analyses of representative rocks (table 1, fig. 3) show that chemically the Canyon Mountain

Table 1. Chemical Analyses of rocks from the Canyon Mountain Comolex, Oregon

| | Ultramafic rocks (11) | Gabbro (9) | Eoidiorite (4) | Basaltic Dike rocks (6) | Albite granite (3) | Kerato-phyre (2) |
|---|---|---|---|---|---|---|
| $SiO_2$ | 42.12 | 47.21 | 52.84 | 51.41 | 76.06 | 73.78 |
| $Al_2O_3$ | 1.70 | 17.96 | 14.86 | 15.86 | 12.87 | 12.68 |
| $Fe_2O_3$ | 2.49 | 1.21 | 1.77 | 1.37 | 1.12 | 1.53 |
| FeO | 4.79 | 4.45 | 6.77 | 8.76 | 1.62 | 2.50 |
| MgO | 35.59 | 10.70 | 8.49 | 5.92 | .36 | .70 |
| CaO | 5.58 | 14.51 | 9.08 | 8.97 | 1.62 | .95 |
| $Na_2O$ | .55 | 1.03 | 2.92 | 3.90 | 5.24 | 5.89 |
| $K_2O$ | .03 | .06 | .31 | .30 | .25 | .15 |
| $H_2O+$ / $H_2O-$ | 6.94 | 2.43 | 2.46 | 1.73 | .98 | 1.56 |
| $TiO_2$ | .04 | .25 | .52 | 1.36 | .19 | .26 |
| $P_2O_5$ | .02 | .03 | .06 | .14 | .04 | .07 |
| MnO | .13 | .12 | .15 | .20 | .04 | .07 |
| $Co_2$ | .12 | .03 | .03 | .13 | .14 | .01 |
| | 100.09 | 99.99 | 100.26 | 99.95 | 100.53 | 100.15 |

Analyses made by U.S. Geological Survey; (11) number of analyses averaged.

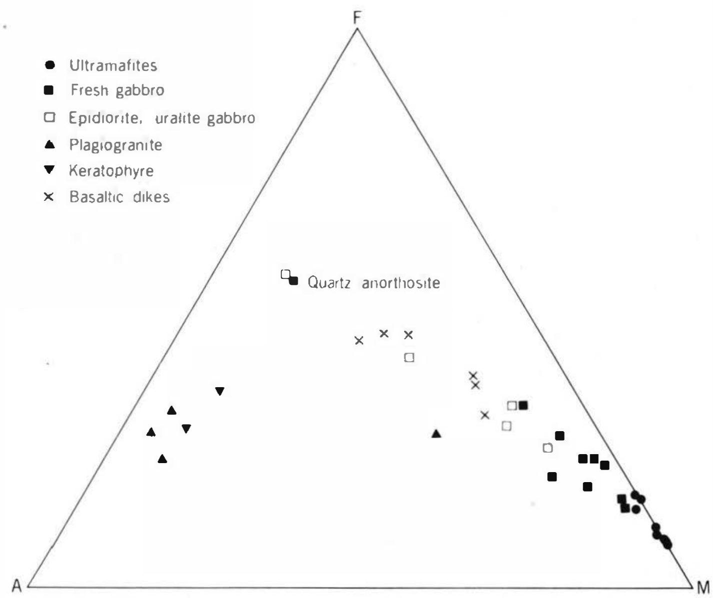


Figure 3. AFM diagram of rocks from the Canyon Mountain Complex, in mol percent. $A=Na_2O+K_2O$, F=total iron as FeO, and M=MgO.

Complex is remarkably like the Coast Range ophiolite in California and other ophiolite suites (Bailey and Blake, 1974; Coleman and Peterman, 1975; Thayer, 1972). Because many more chemical analyses are needed and mineralogical studies by others are in progress, detailed discussion of minor variations in compositions of rocks would not be fruitful. Previously (Thayer, 1963b, p. C85), I interpreted the gabbro and plagiogranite as belonging "to a distinctive line of magmatic descent, for which the name 'alpine mafic magma stem' is suggested." Then, I also believed that the gabbro had been emplaced into volcanic country rocks (now mapped as the sheeted dike unit) as crystal mush. Both ideas, I now believe, were erroneous, and I would like to reconsider some critical relations between the various rocks.

Most petrologists seem to agree that the cumulate rocks of the ophiolite assemblage are cogenetic differentiates with plagiogranite, keratophyre, and basaltic dikes from an oceanic basalt magma. Bailey and Blake (1974, p. 652) summarized the theory very well:

> "Various models suggested for the buildup of oceanic crust [ophiolite] at spreading axes postulate: (1) removal of basaltic magma from lherzolite or related mantle material to yield a residual, harzburgitic upper mantle, (2) coeval extrusion of the basaltic magma to form an overlying cover of pillow basalt, into which a related swarm of parallel dikes is intruded, and (3) intrusion of mafic magma at or near the base of the basalt layer to fill chambers in which the magma differentiates by crystal settling to yield layered cumulates of gabbroic to anorthositic rocks and a small amount of silicic magma that may be extruded to form keratophyre or consolidate in situ as trondhjemite."

Coleman and Peterman (1975, p. 1103) discussed differentiation of basalt according to this hypothesis in more detail:

> "Another chemographic method of relating rocks to differentiation processes is the MgO variation diagram. From this plot showing the parent liquid, the cumulate gabbros, and the oceanic leucocratic differentiates it is apparent that the selective removal of calcic plagioclase, pyroxene, and olivine from the parent magma to form the cumulate gabbros will lead to a residual magma that will crystallize abundant quartz and sodic plagioclase. The $SiO_2$, CaO, $Al_2O_3$ plots versus the MgO plot show that the removal of predominately calcic plagioclase with lesser amounts of clinopyroxene, orthopyroxene, and olivine could produce a residual melt whose composition is similar to the oceanic plagiogranites. However, the plots of FeO (total) and $TiO_2$ versus MgO show iron and titanium depletion in both the cumulate gabbros and the oceanic plagiogranites compared with the parent liquid."

## Problems of the One-Magma Theory

The two papers specified only gabbroic and anorthositic rocks as cumulates, but they omitted ultramafic rocks which should have been included. In the Vourinos, three-quarters of the 1565-m cumulate section described by Jackson, Green, and Moores (1975, p. 391) consists of dunite, wehrlite, and pyroxenite. In the Troodos, George (1975, p. 28) showed that ultramafic rocks make up about 65 percent of the cumulate section, compared with an estimate of 40-50 percent ultramafic by Greenbaum (1972). All three accounts agree that the rocks in question, from dunite to "high level" gabbro, form a continuous cumulate sequence that must be considered as a whole. Omitting the ultramafic cumulates simplifies, but also grossly skews, problems of petrogenesis. If cumulate processes formed hundreds of meters of dunite in the Vourinos and the Troodos, why not harzburgite, too?

The definitive field relations of peridotite and gabbro with plagiogranite and basaltic dikes in the Canyon Mountain present serious problems for the one-magma theory. Derivation of plagiogranite and gabbro from basalt that cuts the gabbro and is cut by the plagiogranite (figure 2) obviously is impossible. Advocates of a common parentage of all three rocks have proposed a variety of models in which basalt dikes are rooted in gabbro and plagiogranite is generated at the top of magma chambers above cumulates. Moores and Vine (1971, p. 462) have presented a model featured by a series of magma cells that are progressively older, one above the other. Basaltic magma and plagiogranite that were differentiated in the lower, younger cells form dikes in overlying cumulates and basalt flows that became part of the thickening roof.

Derivation of basalt dikes that cut an exposed peridotite-gabbro complex from a hypothetical younger mass at depth poses two questions. How many complexeslike the Canyon Mountain Complex can be superposed vertically within the narrow confines of a spreading center like the Mid-Atlantic Ridge (Moores and Vine, 1971, p. 462)? If basalt dikes were tapped from the top of a magma chamber during the time when peridotite and gabbro cumulates were crystallized, they should show marked differentiation trends. The composition of magma from which harzburgite or wehrlitic dunite and chromite crystallize at an early stage, as in the Vourinos, must be very different at the stage when hypersthene gabbro crystallizes from it.

Mutual relations of fresh gabbro, gabbroic pegmatites, epidiorite, and plagiogranite in Canyon Mountain cast doubt on the differentiation series from gabbro to plagiogranite proposed by Coleman and Peterman (1975, p. 1102):

> "Eight new analyses of the oceanic plagiogranites and associated gabbros from the Cyprus ophiolite are presented in Tables 1 and 2. High silica, low to moderate alumina, low total iron-magnesium, and extremely low $K_2O$ characterize the oceanic plagiogranites and their associates. Normative orthoclase is usually less than 4 mol %, and the normative An content of the plagioclase ranges from $An_{21}$ to $An_{61}$. The wide range in An content demonstrates differentiation from basaltic composition toward leucocratic types. Plotting normative Or, An, and Ab from these rocks on a triangular diagram reveals that they all fall within the low-pressure one-feldspar boundary."

Field and petrographic relations in the Canyon Mountain show beyond question that the plagiogranites intruded gabbro and metamorphosed it to epidiorite, after the gabbro had been intensely deformed. The two igneous suites are completely different. One can, however, show an overlapping range in composition of normative plagioclase from fresh gabbro to plagiogranite as follows: fresh gabbro, $An_{92-49}$; epidiorite, $An_{65-35}$; plagiogranite, $An_{45-4}$. Normative orthoclase has the following ranges: $Or_{0-2}$ in gabbro, $Or_{0.9-5.1}$ in epidiorite, and $Or_{1.4-4.7}$ in plagiogranite. It seems almost certain to me that Coleman and Peterman's (1975) analyses include epidiorite, which is abundant in Cyprus. Bailey and Blake (1974) expressed uncertainty about the significance of hornblende in gabbro and considered the possibility that it is metasomatic. Their descriptions (p. 640) of green hornblende as a primary mineral in gabbro and of pegmatitic gabbros which "in contrast to the layered gabbros, are non-foliated and commonly contain large randomly oriented crystals of hornblende and a little quartz" smack of epidiorite to me.

Efforts to calculate chemical balances between basalt as parent magma and alpine peridotite, gabbro, and plagiogranite as differentiates have not been notably successful (Thayer, 1976). Divorcing harzburgite from gabbro as a refractory residuum and overlooking ultramafic cumulates help, but do not solve the problem. Coleman and Peterman's (1975) plots of "FeO" and $TiO_2$ versus MgO in oceanic basalt, gabbro, and plagiogranite show excesses in the basalt comparable to those in my table 1. In my table, 6 basaltic dikes average 10 percent "FeO" and 1.4 percent $TiO_2$, compared to 5.5 percent "FeO" and 0.25 $TiO_2$ in gabbro and 3.1 "FeO" and 0.22 percent $TiO_2$ in albite granite and keratophyre. Alpine peridotite, gabbro, and plagiogranite just do not include Fe- and Ti-rich members needed for a chemical balance with common varieties of basalt as the parent magma. The close association of basaltic dikes with the plagiogranite--keratophyre suite and their low potash content suggest that they very likely are comagmatic.

Gabbroic pegmatites present strong evidence against derivation of tholeiitic magma from ophiolitic gabbro mush. In the Canyon Mountain Complex, these pegmatites have the composition of hypersthene gabbro rich in MgO and CaO, contain no hornblende or quartz, and have been deformed to various degrees (Thayer, 1963a). They clearly are formed locally, and are presumed to represent interstitial liquid from cumulate mush. Their composition is consistent with the lack of zoning in the cumulus minerals and is far from that of any plagiogranite.

The critical problem of the one-magma theory of ophiolite parentage is that no one yet has presented convincing evidence either of ophiolitic cumulate gabbro grading into diabase or of a differentiation sequence from gabbro to plagiogranite. Both relations are common in small stratiform gabbroic complexes. If the one-magma concept for ophiolites is valid, why has no one found "hard" field evidence to substantiate it?

## A Multi-Magma Theory of Ophiolite Parentage

The difficulties in trying to fit the various igneous components of the ophiolite assemblage into a structurally and geochemically coherent one-magma theory or model suggest that something is inherently wrong with the concept. The possibility of peridotitic lavas was rejected by most geologists until about 8 years ago when komatiites were discovered. Modern theories of plate tectonics are still in rapid flux, but the one-magma concept of ophiolites seems to have an air of petrologic sanctity. A more careful search through the many varieties of oceanic basalt should find a more suitable parent magma for cumulate peridotite and gabbro than the one Coleman and Peterman (1975, p. 1103) selected; magma that crystallized as an ordinary pillow lava obviously will not do.

Discovery of high-Al chromite by Sigurdsson and Schilling (1976) in some basalt from the Mid-Atlantic Ridge indicates progress in the search for a more compatible parent for alpine peridotite-gabbro cumulates. The flows are olivine and diopside normative tholeiite and $Al_2O_3$-rich picrite. Chromites from both kinds of rocks are typically alpine in affinity in that they are the high-Al variety ($Al_2O_3$>20 percent, $Cr_2O_3$ + $Al_2O_3$ >60 percent) and show a reciprocal relation between $Cr_2O_3$ and $Al_2O_3$. Chromite from the tholeiite flows ranges from 37.1 to 43.5 percent $Cr_2O_3$ and 23.7 to 30.8 percent $Al_2O_3$, whereas that from picrite ranges from 19.5 to 24 percent $Cr_2O_3$ and 43 to 47.3 percent $Al_2O_3$. Total iron as FeO and $TiO_2$ range from 13.2 to 17.8 percent and 0.15 and 0.42 percent, respectively. On a Cr:Al plot the samples show a strong bimodality, but a plot of Mg:Fe shows continuous variation; both trends characterize podiform or alpine chromites (Thayer, 1970). This suggests that the picrite, although it is described as noncumulate, may be a remobilized cumulate differentiate of the tholeiite.

Variations in the composition of chromites from different kinds of basalts reflect close magmatic control that should apply also to segregated chromitite. In addition to the high-Al chromite, Sigurdsson and Schilling found distinctly different chromite in a low-$SiO_2$ and high-$TiO_2$ basalt in the same segment of the Ridge. That chromite is 1.5-2 times as rich in iron (23-24 percent as FeO) and 4-10 times as rich in $TiO_2$ (about 1.7 percent). A Cr:Fe ratio of 1.45:1 in this chromite, in comparison with ratios of 2:1 to 2.35:1 in the high-Al chromites, places it squarely in the stratiform high-Fe trend (Thayer, 1970). Evans and Wright (1972) found liquidus chromite from Kilauea and Makaopuhi to be even richer in iron (25-33 percent as FeO) and $TiO_2$ (2.3-3.2 percent). The chemical "sensitivity" of chromite to magma composition would seem to severely restrict the choice of basalts that might represent parent magma for alpine peridotite, gabbro, and chromitite.

Gabbroic partial melts that chemically seem similar to the tholeiites containing high-Al chromite have been described in lherzolite in the Lanzo (Budier and Nicolas, 1972) and Serrania de la Ronda (Dickey, 1970) complexes. The "magmatic" gabbros in both complexes are "hyperaluminous" olivine and diopside normative tholeiites, contain more $Cr_2O_3$ than $TiO_2$, and have MgO:"FeO" ratios mostly between 3.2 and 4.6. Normative feldspar ranges between $An_{60-65}$ in gabbro from Lanzo, and $An_{77-83}$ in Serrania de la Ronda. Budier and Nicolas (1972, p. 50) calculated that the gabbro represents melting of 25-30 percent of the lherzolite at pressures between 5 and 8 kb.

A multi-magma hypothesis for genesis of ophiolite does not simplify conception of a plausible model of a spreading ridge; it probably complicates the problems. I believe, however, that it emphasizes the need for much more explicit information on the physical relations between sheeted dike swarms, plagiogranite, and alpine gabbro. Chemically and physically, the basalt dikes in the Canyon Mountain Complex seem cogenetically incompatible with the gabbro, but the plagiogranites might be related chemically to either. Their close structural and temporal ties, however, suggest that the plagiogranite and basaltic dikes probably are related.

The intimate association of the plutonic and volcanic rocks of the ophiolite assemblage is beyond argument and, in a broad sense, they must be related genetically through the processes of plate tectonics. The upper mantle, however, is being recognized as a complex affair in which many kinds of rocks originate. One might compare a spreading ridge to an old-fashioned barnyard. There, several kinds of animals live together. Some animals, such as chickens and goats, obviously have very different ancestry. Cows and horses have many features in common, but have evolved separately since the earliest Eocene. I would like to propose that the ophiolitic assemblage has many aspects of a petrologic and petrogenetic barnyard, in which basaltic magmas belonging to different genera and species have been brought together. This is the essence of the multi-magma theory.

## SUMMARY AND CONCLUSIONS

Excepting pillow lavas, the Canyon Mountain Complex comprises all the igneous components of the ophiolite assemblage: a plutonic suite of rocks that range from harzburgite to gabbro, and a volcanic suite made up of basaltic dikes and silicic rocks that range from quartz diorite to albite granite (plagiogranite) and keratophyre. The two suites of rocks are completely separate in outcrop, structurally and geochemically.

The plutonic rocks form a continuous sequence from harzburgite through wehrlite and clinopyroxenite to gabbro in which cumulus textures are preserved in chromitite if not in silicates. The petrologic sequence is remarkably similar to those in the Vourinos and the Troodos. In all three complexes deformation of the rocks decreases upward, but the upper limit of intense tectonism and sharpness of break between obvious cumulates and tectonites varies from complex to complex. These differences are attributed to variations in timing of accumulation and syntectonic deformation of cumulate crystal mush. In the Canyon Mountain, the peridotite and gabbro had formed a solid tectonic unit before intrusion of the volcanic suite.

The rocks of the volcanic suite form a composite sheeted dike unit that cuts gabbro off abruptly and encloses screens of it. Dikes of basalt and silicic rocks cut each other in the sheeted dike unit and where intruded along fracture systems are chilled against gabbro and peridotite. Gabbro has been extensively hornblendized to epidiorite near larger masses of plagiogranite, and pargasitic pegmatite occurs locally. Textural gradations from albite granite to flow-banded quartz keratophyre and classic intrusive relations of these rocks in basalt and gabbro imply that they represent the substructure of a silicic volcanic pile.

Although the complex is in a terrain that has been subjected to major folding repeatedly, from Permian to Pliocene time, the dikes in the sheeted unit are believed to be near their original attitude. The relations of basaltic dikes to gabbro in the Canyon Mountain are identical, except for details of gabbro structure, with those exposed along the coast of Hatay, Turkey (Parrot, 1973). Plagiogranite also cuts gabbro and basaltic dikes there, although in minor amounts. The original cover of volcanic and sedimentary rocks presumably has been eroded from the complex, as along the beach cliffs of Hatay. Interpretation of the sheets as sills that have been rotated 90° or more would involve a section of crust 11-12 km thick, and would require also that the now vertical diorite plugs in gabbro originally were intruded horizontally. Progressively lower grade metamorphism of gabbro and basalt by plagiogranite under conditions that formed pargasitic hornblende pegmatite and epidotic greenschist, respectively, indicates falling temperatures that would accompany elevation of the block. The mylonitic shear zones in gabbro involving plagiogranite probably are all pre-Late Triassic in age, although movements undoubtedly occurred along many of them much later. The complex has been compressed in the core of a late Tertiary anticline with 40° dips in the south limb, so it must have undergone much internal adjustment.

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# THE TRINITY OPHIOLITE
# KLAMATH MOUNTAINS, CALIFORNIA

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## ABSTRACT

The Ordovician (480-455 m.y.) Trinity ophiolite, located in the eastern Klamath Mountains of northern California, is a large, nearly horizontal sheet of mafic and ultramafic rocks with minor volcanic and diabasic rocks. Along its northwestern edge, the ultramafic rocks are predominantly harzburgite and feldspathis lherzolite, with minor dunite and pyroxenite. Three types of gabbro have been recognized: layered cumulate gabbro and associated diorite, amphibolitic gabbro with gneissic structures and a layer of recrystallized clinopyroxenite at its base, and younger (430 m.y.) pegmatitic gabbro which intrudes the rest of the ophiolitic sequence. Overlying the layered cumulate gabbro and diorite and gradational with them is a complex of diabasic dikes and sills which do not appear to exhibit sheeted structure. The diabase grades upward into mafic keratophyres and spilites. The ophiolite sequence is folded and faulted; some folding and faulting predates the intrusion of the 430 m.y. pegmatitic gabbros, and some is later.

Directly overlying the Trinity ophiolite is a melange consisting of several assemblages of distinct age and lithology. Conglomeratic lenses of probable middle Ordovician to early Devonian age within one melange unit contain clasts of diorite similar in composition and age (460 m.y.) to diorites of the ophiolite. This relationship suggests that part of the ophiolite was uplifted and being eroded before the melange formed. The melange also contains tectonic blocks of gabbro, pyroxenite, and peridote, as well as schist and amphibolite which may be metamorphosed ophiolitic rocks, and in a fenster 10 km to the west a similar melange contains blueschist blocks. Evidence for the mode of origin of the melange is contradictory; probably the melange formed by a combination of gravity sliding and tectonic processes.

Overlying both the melange and the ophiolite are sparse mafic volcanic rocks, locally pillowed. Dikes of similar composition cut melange and ophiolite, suggesting that these volcanic rocks were erupted after melange and ophiolite were juxtaposed. These volcanic rocks may be coeval with the mid-Devonian to Jurassic island arc sequences southeast of the Trinity ophiolite, and probably in late Paleozoic time the juxtaposed ophiolite and melange lay in the fore-arc region of an active volcanic arc.

Also overlying the ophiolite and juxtaposed subduction complex to the west and northwest is a stack of imbricate thrust slices, some of which probably represent parts of a dismembered turbidite fan. This thrust complex was probably emplaced after early Devonian time, but the original relationship of its component slices to the Trinity ophiolite is not clear.

## INTRODUCTION

The Trinity ophiolite is located in the Eastern Klamath Subprovince (Irwin, 1966) of the Klamath Mountains, northern California, just west of the Cenozoic volcanic rocks of the Cascade Range (Fig. 1). It is approximately 75 km long and 50 km wide, and consists predominantly of ultramafic and gabbroic rocks, with minor diabase and volcanic rocks. The area which I have studied in detail (Figs. 1, 3) includes about 250 square km of the northwestern edge of the ophiolite and the rocks overlying it.

Geophysical data (gravity and magnetic) suggests that the Trinity ophiolite is a relatively thin subhorizontal sheet overlying a less dense basement (Irwin and Bath, 1962; Irwin and Lipman, 1962; LaFehr, 1966). The geophysical data further suggest that rocks of this sheet probably extend westward under the Paleozoic sediments and metasediments as far as the Scott Valley (Fig. 1), and may extend some distance southeastward beneath younger rocks (Griscom, 1977).

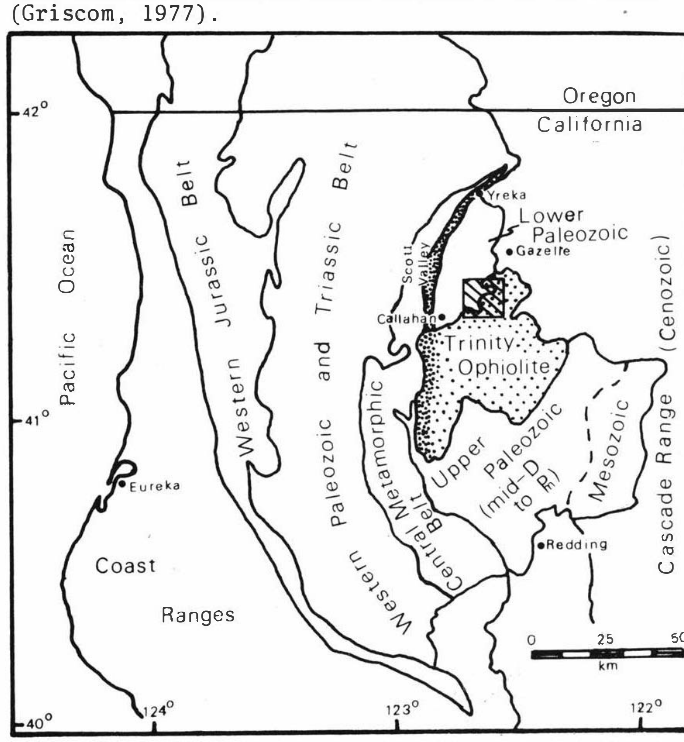


Figure 1. Location of the Trinity ophiolite (open stipple pattern). Dense stipple pattern: belt of severely deformed ultramafic and mafic rocks which have undergone a different tectonic history than the Trinity ophiolite and which may not be related to it. Area of diagonal bars: approximate area of Figure 3.

This large mafic-ultramafic complex was originally named the "Trinity Alps batholith" by Hinds (1935, p. 363). Since then, numerous hypotheses have been proposed to explain its origin; these have been summarized elsewhere (Lindsley-Griffin and Rohr, 1977). Many workers have described the Trinity complex as oceanic crust (Hamilton, 1969) or as an ophiolite (Mattinson and Hopson, 1972; Hopson and Mattinson, 1973; Lindsley-Griffin, 1973; D'Allura and others, 1974).

Along the southwestern edge of the Trinity ophiolite, and continuous with it (Fig. 1), lies a thin belt of highly deformed and metamorphosed ophiolitic rocks which has undergone a different tectonic history. This deformed belt extends northward from Callahan to Yreka (Fig. 1) where it is faulted against greenschist metavolcanic rocks containing blocks of glaucophane-lawsonite blueschist (Hotz, 1973, 1974). This belt of rocks, formerly described as being part of the Trinity ophiolite (Irwin, 1966), should probably be considered as a distinct and separate ophiolite fragment (D'Allura and others, 1974; Cashman, 1977; Lindsley-Griffin, 1977).

<u>Age</u>

Prior to 1968, rocks of the Trinity complex were considered to be late Jurassic or older in age. In 1968, Lanphere and others published K-Ar dates of 418 and 439 m.y. for gabbros associated with the Trinity ultramafics. The bodies sampled are part of the Trinity ophiolite although they lie southwest of the area of detailed mapping shown in Figure 3. The K-Ar ages are probably a little young; U-Pb dates for diorite and layered gabbro lie in the range 455 to 480 m.y., and around 430 m.y. for the pegmatitic gabbros, (Mattinson and Hopson, 1972; Hopson and Mattinson, 1973; Hopson, personal comm., 1976).

## STRATIGRAPHY

Although the Trinity ophiolite is folded and faulted, portions of the original stratigraphy are preserved within individual fault blocks. Careful mapping to define fault blocks, together with correlation between similar rock types, permits the development of a columnar section shown diagrammatically in Figure 2. Thicknesses of the units shown in this section are maximum thicknesses since parts of the section may have been repeated by unrecognized faults or folds.

<u>Ultramafic Rocks</u>

The lowest stratigraphic unit of the Trinity ophiolite is serpentinized ultramafic rocks of variable composition ranging from harzburgite (en + ol + sp) and feldspathic lherzolite (en + di + ol + sp + f) to dunite (ol + sp + en) interlayered with peridotite. Selected mineral analyses are shown in Table 1.

North of China Mountain (Fig. 3), the peridotites are predominantly massive serpentinized harzburgite, locally layered. Ultramafic rocks exposed southeast of China Mountain and South China Mountain (Fig. 3) consist of interlayered dunite and lherzolite with minor harzburgite. West of Cory Peak (Fig. 3), the ultramafic rocks are a mixture of harzburgite and lherzolite in which the lherzolite consists of rather irregular masses which grade over a distance of a few meters to a few tens of meters into harzburgite.

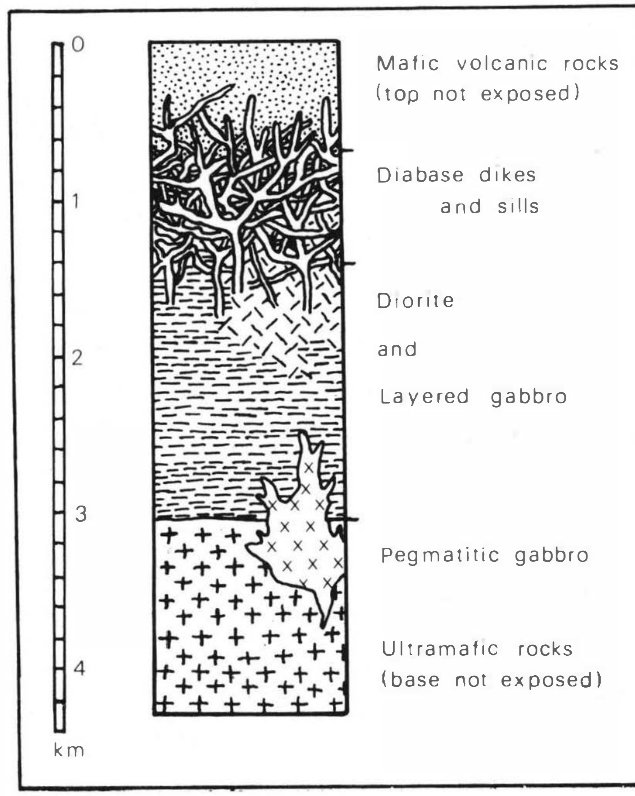


Figure 2. Diagrammatic columnar section of the Trinity ophiolite. Thicknesses of units are approximate. Mafic volcanic rocks (massive to brecciated spilites and keratophyres) grade downwards into diabase dikes and sills, which grade downwards into quartz diorite, hornblende diorite and gabbro. The gabbro and some of the diorite exhibit cumulate layering, and become more mafic downward. Lower contact of layered gabbro is sheared, and may be intruded by massive pegmatitic gabbro dikes and stocks. Ultramafic rocks are massive to layered serpentinized peridotite with minor dunite and pyroxenite.

Occurrences of feldspar and clinopyroxene are generally difficult to observe in the field because of the pervasive serpentinization and small grain size of the feldspar (<0.5 mm) and clinopyroxene (<1 mm). Thus, detailed relationships between harzburgite and lherzolite have not been mapped.

Layering in the peridotites, defined by a change in the relative proportions of pyroxene and serpentinized olivine, ranges in thickness from a few centimeters to about a meter. A foliation defined by flattened and elongated pyroxenes is superimposed on the layering but is often masked by the pervasive serpentinization. Where both are present, the foliation is parallel or nearly parallel to the layering. Both the foliation and the layering are folded, and are cut by a less deformed foliation defined by asbestos- or serpentine-filled veins which are present locally.

On a microscopic scale, the lherzolites typically exhibit a texture characterized by grains of diopside clustered around irregular blebs of saussuritized feldspar. The diopside and feldspar are elongated parallel to the foliation, and are associated with

TABLE 1. PRELIMINARY ELECTRON MICROPROBE ANALYSES OF MINERALS FROM ULTRAMAFIC ROCKS OF THE TRINITY OPHIOLITE

| MINERAL | Orthopyroxene | | | Clinopyroxene | | Olivine | Chrome-spinel | | |
|---|---|---|---|---|---|---|---|---|---|
| SAMPLE | a (large) | a (small) | b | a | c | a | a | b(1) | c |
| $SiO_2$ | 52.41 | 53.18 | 37.21 | 51.91 | 53.07 | 41.24 | --- | 9.85 | --- |
| $TiO_2$ | 0.16 | 0.24 | 0.00 | 0.37 | 0.15 | --- | 0.28 | 0.11 | 0.37 |
| $Al_2O_3$ | 2.77 | 2.42 | 2.11 | 3.18 | 2.18 | --- | 28.02 | 4.23 | 27.23 |
| $Cr_2O_3$ | 0.84 | 0.54 | 0.66 | 1.25 | 0.54 | --- | 36.45 | 28.71 | 29.19 |
| FeO* | 7.17 | 6.61 | 5.87 | 2.86 | 3.35 | 11.18 | 23.02 | 43.28 | 34.15 |
| MnO | 0.15 | 0.15 | 0.05 | 0.10 | 0.12 | 0.14 | 0.33 | 1.53 | 0.15 |
| MgO | 33.48 | 34.11 | 37.24 | 16.57 | 17.00 | 48.86 | 12.06 | 10.13 | 9.09 |
| CaO | 0.63 | 0.73 | 0.01 | 23.63 | 24.32 | 0.02 | --- | --- | --- |
| $Na_2O$ | 0.005 | 0.12 | 0.01 | 0.25 | 0.11 | --- | --- | --- | --- |
| | 97.62 | 98.08 | 83.16 | 100.12 | 100.84 | 101.44 | 100.16 | 97.84 | 100.18 |

Analyses performed at University of California, Davis, by the author.

Sample a: Feldspathic lherzolite, Scott Mountain. Small opx + ol + cpx + sp + saussurite interlayered with large opx + ol + sp.
Sample b: Serpentinized harzburgite, Rock Fence Creek. Opx + sp + serpentine; no olivine.
Sample c: Recrystallized clinopyroxenite, China Mountain. Cpx + sp.

(1) This chrome-spinel appears to be optically pure in both plane-polarized and reflected light. When the electron microprobe analyses all proved to be about 10% low, an elemental scan was performed in which the only detectable anomaly was silica. It is possible that the silica is due to finely disseminated serpentine not optically visible, as the harzburgites are all intensely altered.

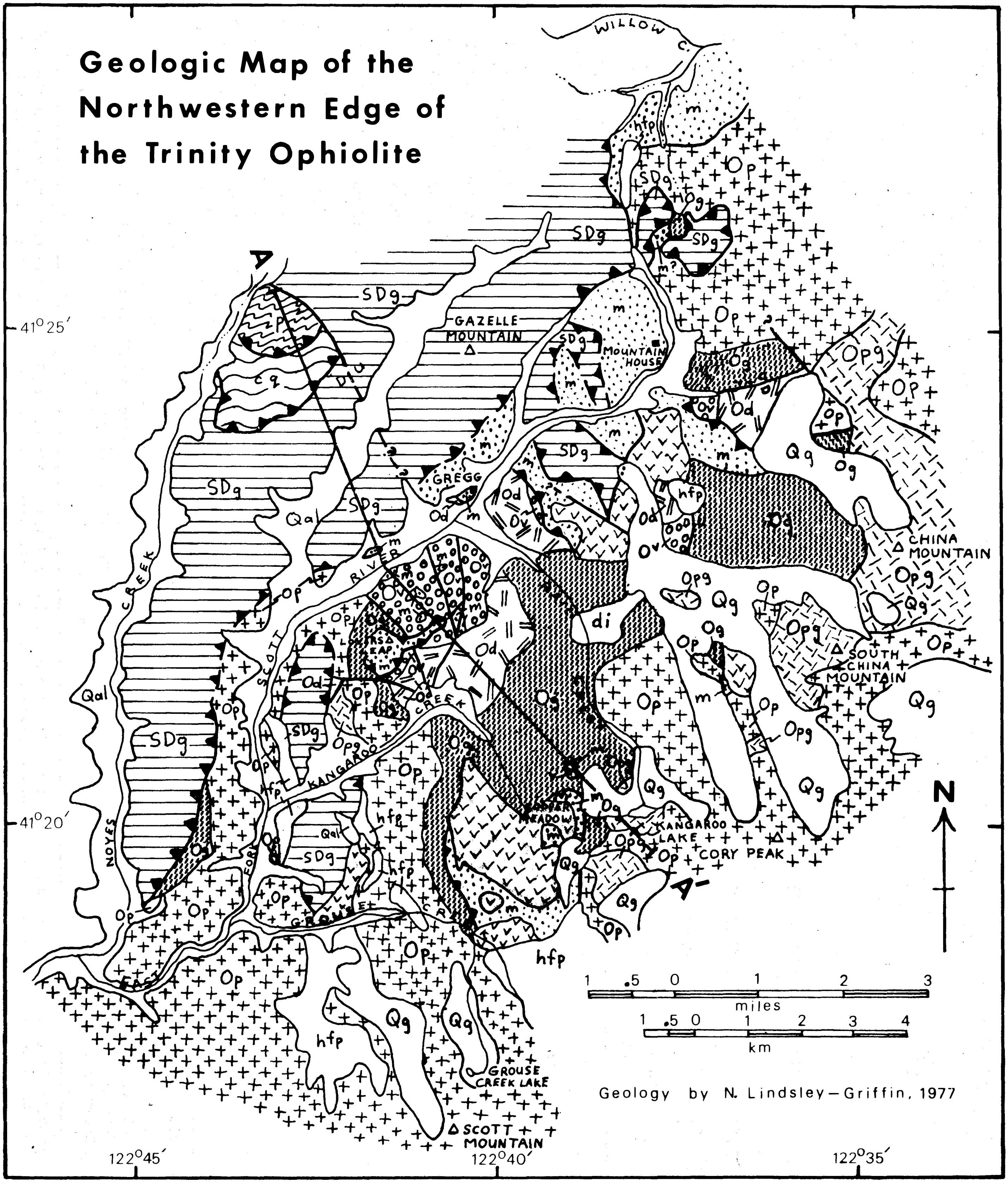
Geologic Map of the Northwestern Edge of the Trinity Ophiolite
WILLOW C.
GAZELLE MOUNTAIN
MOUNTAIN HOUSE
GREGG
SCOTT RIVER
CHINA MOUNTAIN
SOUTH CHINA MOUNTAIN
KANGAROO CREEK
KANGAROO LAKE
CORY PEAK
NOYES CREEK
EAST FORK
GROUSE CR.
GROUSE CREEK LAKE
SCOTT MOUNTAIN
A
A'
N
41°25′
41°20′
122°45′
122°40′
122°35′
miles
km
Geology by N. Lindsley-Griffin, 1977

Figure 3. Geologic map of the northwestern edge of the Trinity ophiolite, between Callahan and Gazelle, California.

LEGEND. Surficial deposits: Qal, Quaternary alluvium (no pattern); Qg, Glacial and fluvioglacial debris (no pattern).

Younger igneous rocks: hfp, hornblende-feldspar porphyry intrusions (no pattern); di, diorite and quartz diorite intrusions (no pattern); v, post-lower Devonian volcanic rocks: tuff, breccia, pillow lava ("v" pattern).

Rocks of the overthrust plates: p, phyllite (tight, asymmetric fold pattern); cq, calcareous quartzite (open fold pattern); SDg, Siluro-Devonian Gazelle Formation (horizontal bar pattern); m, melange, containing blocks of Ordovician, Silurian, and Devonian sedimentary rocks, and of igneous and metamorphic rocks derived in part from the ophiolite (stipple pattern).

Lower Ordovician Trinity ophiolite: Opg, pegmatitic gabbro (perpendicular dashes); Ov, mafic volcanic rocks (open circles); Od, diabase dikes and sills, diorite in lower part (randomly oriented sets of two parallel lines); Og, hornblende gabbro and diorite (closely spaced parallel dashes); Op, serpentinized peridotite with minor dunite and pyroxenite (crosses).

A-A': line of cross section (Fig. 4).

small enstatite grains, olivine, and chrome-spinel. Between the lenses or bands of lherzolite are zones consisting of harzburgite (large enstatite grains, olivine, and chrome-spinel) with little or no clinopyroxene or feldspar. Similar textures described at Papua (England and Davies, 1973) and Othris (Menzies, 1972; Menzies and Allen, 1974) have been attributed to incomplete extraction of partial fusion products in the upper mantle.

## Gabbroic Rocks

Stratigraphically above the ultramafic rocks lie gabbros and diorites exhibiting cumulate textures and layering. The cumulate rocks are characterized by the graded bedding, cross-bedding, cut-and-fill, and slump structures typical of igneous rocks formed by processes of sedimentation within a magma chamber. These features have been described in other ophiolites (Hopson, 1975; Ewing, 1976; Harkins and others, 1976).

In composition these are now hornblende gabbros and hornblendites, grading upward into hornblende diorites and quartz diorites. The hornblendes are commonly pseudomorphous after pyroxene crystals. Feldspars range in composition from anorthite to labradorite, with albite being common in the quartz diorite to diabase zone.

The contact between the gabbroic rocks and the ultramafic rocks is invariably sheared and serpentinized. Near the contact, structures in both the ultramafic and the gabbroic rocks may strike either parallel or perpendicular to the contact. The gabbroic rocks are typically severely deformed near this contact, and exhibit gneissic structures which become less common away from the contact.

The fault-bounded, synformally folded block of gabbro exposed on the west shoulder of China Mountain (Fig. 3) is characterized throughout by gneissic or amphibolitic structures, which are especially well-developed near its base. Along the base of the block (to the east and north) is exposed a band, 100 to 300 meters thick, of clinopyroxenite characterized by boudin structures. In thin section, this rock exhibits even grain size, straight or slightly curved grain boundaries, and equiangular triple points between grains, suggesting that it has been partially to completely recrystallized. This recrystallized clinopyroxenite is found only in association with amphibolitic gabbros, and northwest of China Mountain (Fig. 3) is faulted against serpentinized harzburgite. This fault is in turn cut off by an intrusive contact with the 430 m.y. pegmatitic gabbros which form the peak of China Mountain.

## Dike and Sill Complex

Stratigraphically above and gradational with the layered gabbro/diorite is a complex of mafic dikes and sills of variable composition and texture, which typically consist of pyroxene, hornblende, and/or feldspar phenocrysts in a very fine-grained to aphanitic groundmass. The minerals are commonly altered to a characteristic greenschist assemblage which includes epidote, chlorite, and albite.

It is not possible to demonstrate sheeted structure, partly due to the fact that the diabase weathers readily and is poorly exposed. However, the few good exposures that have been observed suggest a very roughly parallel set of dikes cut by numerous anastomosing sills and dikes at various other orientations.

As the contact between the gabbro/diorite zone and the dike/sill zone is approached, a few fine-grained diabase dikes may be observed, ordinarily cutting hornblende diorite or quartz diorite, but sometimes cutting hornblende gabbro. As one continues upsection, the amount of diabase rapidly increases relative to the diorite, until the rocks consist predominately of diabase or very fine-grained diorite. Blocks of layered gabbro and hornblendite may occur fairly high within the diabasic zone and presumably represent patches of screen.

## Volcanic Rocks

The mafic volcanic rocks associated with the Trinity ophiolite are spilites and keratophyres exhibiting typical prehnite-pumpellyite and low greenschist facies mineral assemblages. In the field, phenocrysts of feldspar and pyroxene may be observed, as well as small grains of red chert or jasper. Considerable amounts of secondary silica and calcite are present (the matrix consists locally of as much as 15% calcite) and rocks of this unit will "fizz" upon application of HCl. Contacts between diabase and volcanic rocks were drawn partly on this basis.

This unit is further characterized by extensive metallic mineralization. Sulfide minerals in small veins, or disseminated, are common. Numerous small copper prospects and pits dot the exposures of volcanic rocks, although none have developed into large-scale operations.

A chemical analysis of a sample of keratophyre published by Potter and Scheidegger (1973) shows the major element composition of these rocks to be fairly characteristic of other ophiolitic and oceanic volcanics which have undergone prehnite-pumpellyite to low-greenschist metamorphism (Smith, 1968; Hynes, 1975; Church and Coish, 1976).

Both massive lavas and agglomerates are present, and both may be strongly foliated due to the pervasive shearing characteristic of this unit. No undoubted pillow structures have been recognized, but many oblate structures or "pseudo-pillows" may be observed which could be sheared pillow structures.

Although the mafic volcanic rocks exposed at Lovers Leap (Fig. 3) have been variously described as thrust over the ophiolite (Rohr, 1972), or as part of an island arc (Potter and others, 1975), the mapped relationships do not support either hypothesis. The contact is not well-exposed but based on the small outcrops available, and on float, there appears to be a gradation upward from undoubted dike-and-sill complex into a massive fine-grained rock resembling the diabase, into a porphyritic rock resembling the volcanics, and finally into undoubted volcanic rock. Particularly compelling is the observation that in all localities where these rather distinctive volcanic rocks crop out, they exhibit the same stratigraphic relationship, grading downward into dike-and-sill complex, and then into diorite or gabbro (Figs. 2, 3).

## Pegmatitic Gabbros

Pegmatitic gabbros underlie South China Mountain, part of China Mountain, Kangaroo Lake, and occur as dikes and small stocks throughout the mapped area (Fig. 3). They typically consist of altered feldspar (usually saussuritized but sometimes altered to hydrogarnet) and hornblende-rimmed pyroxene crystals as large as 10 cm in diameter, and locally may exhibit zones of medium- or coarse-grained (.2-.5 cm) gabbro. The pegmatitic gabbros are commonly cut by numerous dikes of quartz diorite, hornblende pegmatite, and aplite.

Field relations demonstrate that the pegmatitic gabbros were intruded into the ophiolite complex after some folding and faulting occurred, but before any of the overlying sediments or volcanics were present. For example, the relationships on the west shoulder of China Mountain (Fig. 3) indicate that the amphibolitic gabbro and recrystallized clinopyroxenite were faulted against serpentinized harzburgite before the intrusion of the pegmatitic gabbro body which forms the peak of China Mountain. This observation agrees with U-Pb dates obtained by Mattinson and Hopson which suggest that the pegmatitic gabbros are about 30-40 m.y. younger than the layered gabbros (C. A. Hopson, pers. comm., 1975).

Because the pegmatitic gabbros are associated only with the ophiolitic rocks, they are included in the Trinity ophiolite complex. It is likely that they are the result of late-stage magmatic activity within oceanic crust.

## Sedimentary Rocks

The Trinity ophiolite has been deeply eroded. No pelagic sediments are preserved at its top, and probably most of the original section of volcanic rocks has been removed. Some very immature volcanogenic graywackes and mudstones are interbedded with the mafic volcanic rocks at Lovers Leap (Fig. 3), but they are of limited extent.

One other sedimentary deposit which may be related to the Trinity ophiolite has been observed. This deposit, at the north end of the mapped area, is marked "m?" on Figure 3. The deposit is an unbedded sedimentary breccia consisting of angular unsorted clasts up to 50 cm in diameter. Rock types present as clasts include layered gabbro, amphibolitic gabbro, diorite, quartz diorite (all similar to rocks of the Trinity ophiolite); diabase, siliceous argillite, and quartzite or recrystallized chert. About 80-90% of the clasts are ophiolitic rocks.

My field assistant, Catherine Anderson, undertook a petrographic study of clasts from this breccia in 1976 as a senior research project at the University of California, Davis. She found that many of the fine-grained clasts are actually mylonites, and that many of the coarser-grained clasts exhibit various stages of cataclasis. Electron microprobe analyses of gabbros and metagabbro clasts show that they could have been derived from gabbros of the Trinity ophiolite (C. Anderson, written comm., 1976). One gabbroic clast collected by Ms. Anderson and me is composed partly of diabase which was intruded after cataclasis occurred.

The breccia deposit overlies serpentinized harzburgite of the Trinity ophiolite, and on the east side it appears to be faulted against layered gabbro. Above it are the overthrust shales and siltstones of the Gazelle Formation. Similar breccias have not been observed elsewhere within the mapped area (Fig. 3).

It is likely that the clasts in this breccia originated in an oceanic fracture zone and that the breccia itself formed in or near such a fracture zone. The complete absence of ultramafic clasts in the breccia suggests that it was transported to its present location subsequent to its formation.

## Structure

Structure within the Trinity ophiolite is complex. The ultramafic tectonites exhibit at least three generations of structures; the mafic cumulates above them exhibit less deformation, but at least one generation of folds is present (Lindsley-Griffin, in prep.). The contact between the ultramafic and mafic rocks is sheared, and layering (defined in both lithologies by a change in the relative proportions of minerals) strikes into the contact at varying angles. Layering within either the peridotite or the gabbro may also be locally parallel or perpendicular to the contact.

Several generations of faulting have affected the Trinity ophiolite, the earliest having occurred before juxtaposition of the ophiolite and the overlying rocks, and before intrusion of the pegmatitic gabbros. Some of the early folding and faulting probably occurred within the oceanic crust, but some may have occurred during emplacement. The jumbling and local overturning of fault blocks along the northwestern edge of the ophiolite (Fig. 3) may have occurred any time before or during the juxtaposition of the overlying thrust sheets. Some of the minor high-angle reverse faults may be related to emplacement of the overlying thrust sheets, although not all of these faults cut the thrust sheets.

# OPHIOLITES AND MELANGE

Ophiolites are typically associated with belts of melange, which may include tectonic fragments of blueschist (Hsu, 1971; Coleman and Irwin, 1974). In 1968, Hsu (p. 1065) defined melanges as "mappable bodies of deformed rocks characterized by the inclusion of tectonically mixed fragments or blocks, which may range up to several miles long, in a pervasively sheared, fine-grained and commonly pelitic matrix".

Hsu distinguished between native blocks, "disrupted brittle layers which were once interbedded with the ductiley deformed matrix", and exotic blocks, "tectonic inclusions detached from some rock-stratigraphic units foreign to the main body of the melange" (Hsu, 1968, p. 1065). Every melange contains native and exotic blocks, as well as a matrix.

Since then, numerous workers have contributed to the continuing discussion of melanges. Berkland and others (1972, p. 2296) redefined the term as "...a mappable body of rock characterized by the inclusion of fragmented and generally sheared matrix of more tractable material." This definition permits the term melange to be applied to rock units of both tectonic and sedimentary (olistostromal) origin. However, Raymond (1975) followed Hsu (1968) in restricting the term melange to assemblages containing exotic blocks, whereas both Cowan (1974) and Beutner (1975) suggested that this restriction is too limiting.

Exotic blocks may consist of a variety of lithologies including sedimentary, metamorphic, and igneous rocks; fragments derived from ophiolites are common. Exotic block compositions reported by Maxwell (1974) for the Franciscan complex include chert, volcanic rocks, graywacke, serpentinite, blueschist, schist, phyllite, and greenstone. The diverse melange units may be intercalated with, or partially overlain by, relatively undisturbed sedimentary sequences; they may also include relatively large blocks in which the original stratigraphic succession is still preserved (Fox, 1976).

Maxwell's model (1974) for the origin of the Franciscan complex invokes a complex interaction between the processes of tectonism, diapirism, and normal sedimentation aided by gravity-sliding in a trench environment to explain the observed relationships. Variations in basement composition as well as details of local patterns of sedimentation and tectonism produce melange units characterized by distinct assemblages of exotic blocks.

As Hsu (1968) pointed out, in rock units of this type it cannot be assumed that faunal assemblages contain no exotic elements, for the principles of stratal continuity and superposition can not be rigidly applied. In my opinion, it is in lithologic assemblages of this type that careful and exhaustive paleontologic study becomes indispensable. Once the diverse melange units are recognized and mapped, consideration of faunas contained within each distinct unit may permit reconstruction of the environment and age of the source area for those rocks. Although it might not be possible to restore the individual melange units to their original positions relative to one another, such data can be of help in visualizing the paleogeography of the source area.

But just what is "melange" and how is it to be recognized? The many and often conflicting opinions in the literature suggest that "melange" is a concept which is still evolving. However, I personally believe that for the present, the more flexible definition of Berkland and others (1972) is more useful, that is, that melange may be both tectonic and sedimentary in origin. Thus, I have adopted their definition in discussing the eastern Klamath Mountains.

## Melange Related to the Trinity Ophiolite

Structurally overlying the Trinity ophiolite along its northwestern edge (Figs. 3, 4) is an assemblage of rocks of diverse age and origin. Stratal continuity can be demonstrated locally, but few lithologic units can be traced more than 100 meters, and most are much smaller. Careful mapping by the author in some of the better-exposed localities reveals that distinctive lithologies commonly occur as lensoidal or phacoidal-shaped bodies that are of limited extent, and which exhibit sub-parallel alignment. Internal bedding within the lenses or phacoids is rarely parallel to their external shape, thus arguing against simple interbedding of different facies.

Other significant characteristics include: (1) juxtaposition of lithologies of different age (based upon fossils collected in part by the author and identified by A.J. Boucot and his associates), (2) close juxtaposition of lithologies of different environment of deposition and provenance without obvious signs of faulting, and (3) presence of exotic blocks of ophiolitic fragments and metamorphic rocks (Table 2). Between the blocks of diverse lithology is a very poorly exposed matrix of mudstone or siltstone, which does not exhibit bedding. It is massive to poorly foliated, and is commonly seen only as inconspicuous fragments in the float. Significant sedimentary lithologies (summarized in Table 2) include reefal limestone of various ages, bedded chert, graywacke, shale, and a variety of distinctive conglomerate lenses containing different assemblages of clasts. Blocks of igneous and metamorphic rocks are rare to absent in some localities but are common in others (Table 2, Fig. 5).

These data indicate that this assemblage is a melange, but one in which exotic blocks are rare, metamorphism is low to non-existant, and intense deformation such as is seen in much of the Franciscan complex is not present. Such a melange might conceivably develop at very shallow depths within a trench, or perhaps on the edge of the trench-slope break (Karig, 1972, 1974). In the latter case, gravity-sliding interspersed with spillovers from an inner trench basin (Karig, 1974) would likely be more im-

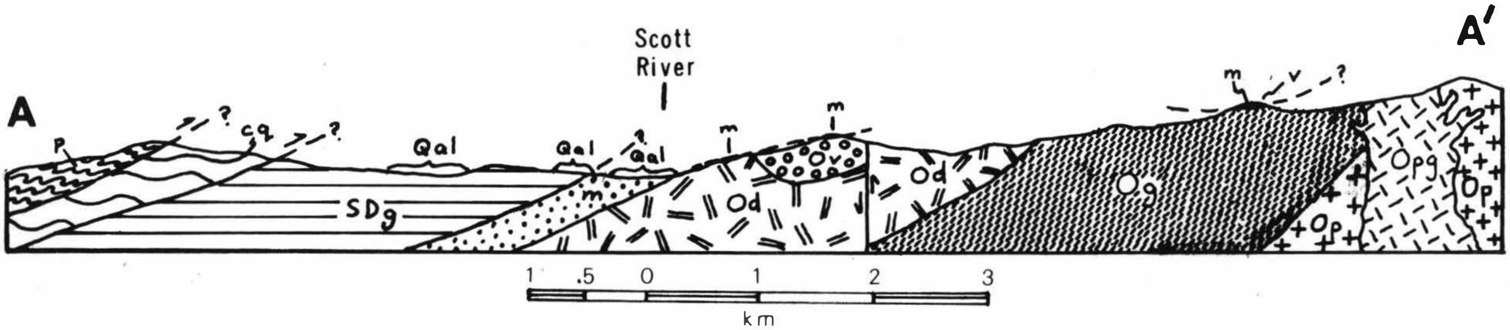


Figure 4. Structural cross section along line A-A' of Figure 3. Symbols and patterns same as in Figure 3.

TABLE 2. MELANGE UNITS AND THEIR CHARACTERISTIC LITHOLOGIES.

| UNIT | LITHOLOGIES | | |
|---|---|---|---|
| | Sedimentary (excluding conglomerate) | Conglomerate types distinguished by clasts (with graywacke matrix) | Igneous-metamorphic |
| Willow Creek | black chert<br>siltstone<br>graywacke | graywacke and siltstone | marble or recrystallized limestone |
| Mountain House | undeformed bedded chert<br>strongly folded bedded chert (may be soft-sediment deformation)<br>bedded shale<br>massive limestone (age unknown)<br>quartz wacke | graywacke with Devonian limestone<br>strongly foliated conglomerate with flattened clasts of gray limestone, black silty limestone, non-ophiolitic diorite, chert, mudstone and siltstone, phyllite, bedded shale, chloritic quartzite, chert breccia (1) | amphibolite<br>layered hornblende gabbro and hornblendite<br>peridotite<br>actinolite schist<br>chloritic quartzite and phyllite<br>marble or recrystallized limestone<br>slate or very siliceous shale |
| Crater Creek | bedded shale<br>Devonian (2) limestone<br>volcanogenic graywacke | limestone and graywacke<br>rounded mafic volcanics in angular volcanogenic matrix of same composition<br>angular ophiolitic volcanics different from the above, in matrix of same composition | not observed |
| Gregg Ranch | well-bedded Ordovician (2) shale<br>Devonian (2) algal limestone<br>massive limestone (age unknown)<br>very angular volcanogenic graywacke | siltstone and mudstone<br>predominantly volcanics with minor ophiolitic diorite and graywacke<br>both graywacke and limestone in varying proportions<br>entirely limestone, but different types in each lens<br>chert granules and pebbles (3) | massive micaceous siltstone (part of melange matrix?)<br>greenstone<br>ophiolitic diorite (exotic block? or basement showing through thin veneer?) |
| Lovers Leap | several large blocks .5-1 km in diameter of Ordovician (4) and possibly some interbedded graywacke and mudstone<br>Silurian (4) limestone<br>graywacke interbedded with shale | Ordovician (4) limestone clasts<br>clasts of Silurian (5) volcanics, rare limestone, siltstone and graywacke, and diorite which is the same age and composition as those of the Trinity complex (6) | not observed |
| Cabin Meadow | shale<br>massive graywacke (part of melange matrix?) | graywacke and siltstone<br>Devonian limestone; graywacke, siltstone | not observed |

TABLE 2--continued

| UNIT | LITHOLOGIES: Sedimentary (excluding conglomerate) | LITHOLOGIES: Conglomerate types distinguished by clasts (with graywacke matrix) | LITHOLOGIES: Igneous-metamorphic |
|---|---|---|---|
| Upper Grouse Creek | mudstone and shale<br>quartz wacke<br>massive limestone<br>interbedded siltstone and shale with slump structures<br>feldspathic sandstone | graywacke<br>rounded volcanics in volcanogenic matrix, interbedded with graywacke and minor shale | not observed |
| Lower Grouse Creek | Devonian (7) fossiliferous gray limestone<br>interbedded shale and siltstone<br>interbedded mudstone and graywacke | angular clasts of diorite, bull quartz, quartzite (metamorphic), quartzite conglomerate, quartz sandstone, keratophyre, phyllite (not same phyllite as the one in Mountain House conglomerate) | serpentinized peridotite<br>pyroxenite<br>pegmatitic gabbro |
| Bonnet Rock (Yreka Quad) | large block (about .5 km diameter)consisting of a sequence of graywackes and cherts capped by massive limestone<br>red shale and siltstone<br>green shale and siltstone interbedded with very silty limestone | chert, volcanic pebbles, quartz-feldspar sandstone | volcanics and pillow volcanics (local rare interbeds of pelitic sediments) |
| Horseshoe Gulch (Fort Jones Quad) | middle Ordovician (9) shale<br>several types of limestone of Ordovician, Silurian, and Devonian age (10)<br>arkose (10) | limestone, diorite (10)<br>limestone, graywacke<br>volcanics | serpentinite (11)<br>volcanic rocks (12)<br>blueschist (13) |
| Moffett Gulch (14) (Yreka Quad) | limestone<br>This assemblage was described by Hotz, 1974 as: "greenschist-albite-chlorite-epidote-actinolite schist and phyllite; limestone lenses; minor siliceous phyllite and semischist; a few small tectonic blocks of glaucophane-lawsonite blueschist." | | marble<br>greenstone<br>blueschist<br>trondjhemite (15)<br>matrix of greenschist-grade phyllite and schist. |

(1) Appearance is strikingly similar to that of tectonic melange described by Cowan (1974) and shown in his Figure 4b.
(2) Dates by A. J. Boucot and associates, in part based on fossils collected by the author.
(3) May belong to overthrust Gazelle Formation.
(4) Dates from Rohr, 1972.
(5) A. J. Boucot, personal communication, 1973.
(6) Mattinson and Hopson, 1973.
(7) Savage, 1976.
(8) Based in part on work by Michael Churkin, Jr. (Churkin and Langenheim, 1960); on reconnaissance by John R. Griffin, 1970-71; supplemented by reconnaissance by the author, 1970-75.
(9) Berry and others, 1973.
(10) Zdanowicz, 1971.
(11) John R. Griffin, personal communication, 1974.
(12) According to Potter and others (1975), this is quartz keratophyre and may belong to the Trinity ophiolite.
(13) Susan M. Cashman, written communication, January, 1977. (Blueschist was first recognized there by P. E. Hotz; later confirmed by Cashman.)
(14) Hotz, 1974; supplemented by reconnaissance in 1974-75 by the author and John R. Griffin.
(15) According to Hotz, 1974, similar to rocks at Lovers Leap which the author considers to be part of the Trinity ophiolite.

portant contributors of material than tectonic mixing of basement fragments. Tectonically controlled basins of limited extent which have been described at accreting plate margins (Schweller and Kulm, 1976) might permit the accumulation of relatively large "blocks" of turbidite sediment in which internal stratigraphy is preserved, or they might serve as catch-basins for submarine mudflows or debris flows carrying exotic blocks acquired upslope.

Although no blueschist has yet been found in the melange immediately adjacent to the ophiolite, blueschist has been reported in Horseshoe Gulch (S. M. Cashman, pers. comm., 1977), which lies to the west (Fig. 5; Table 2). Hotz (1974) reported blueschist blocks in Moffett Gulch (Fig. 5) to the northwest of the Trinity ophiolite. Associated with the blueschist in Moffett Gulch are a variety of rocks, including some which may be ophiolite fragments (Table 2). Reconnaissance by John R. Griffin and me in 1974 suggests that these rocks may be part of a melange which has experienced greater deformation and metamorphism than the melange adjacent to the Trinity ophiolite. However, it is not clear what the relationship between the two assemblages may be. The Moffett Gulch assemblage might be a continuation of the melange exposed along the edge of the Trinity ophiolite, or it might be completely unrelated.

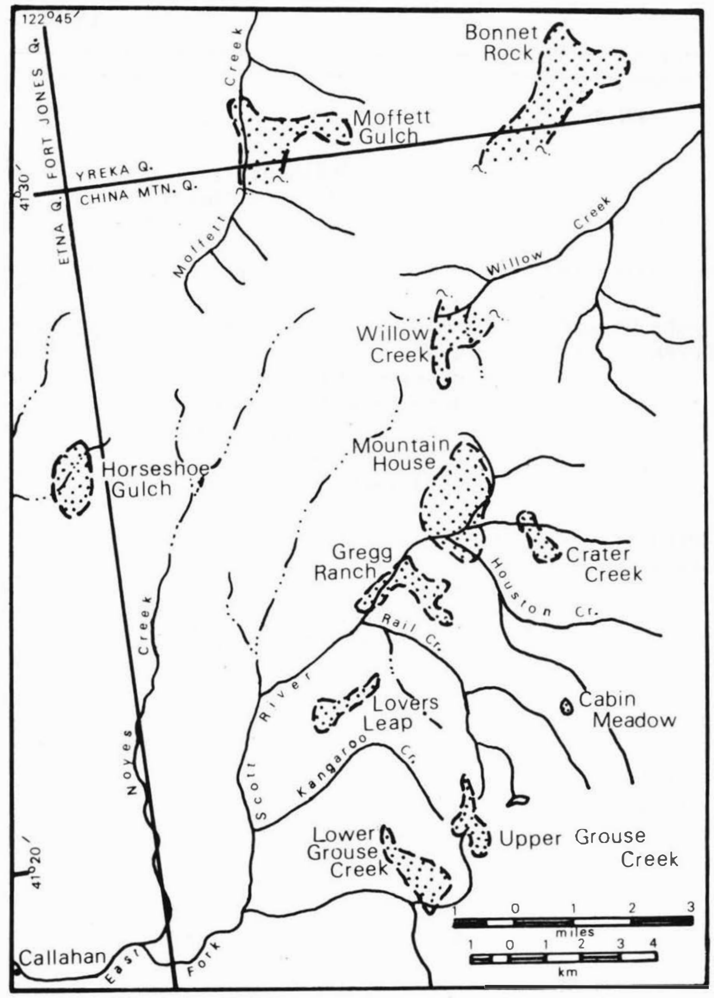


Figure 5. Locations of melange units described in Table 2.

## REGIONAL SETTING OF THE TRINITY OPHIOLITE

Before determining the time of accretion of an ophiolite onto the continental margin, its relationship to the surrounding rock units must be ascertained. Of considerable importance is the age and type of basement, if any, over which the ophiolite is emplaced.

In the case of the Trinity ophiolite, however, neither older nor structurally lower rocks have been discovered. It has been suggested that the Trinity complex was emplaced by thrusting or intrusion into rocks of the Central Metamorphic Belt (Lipman, 1964; Davis, 1968), but the area where these relationships have been observed is 50 km southwest of the main body of the Trinity ophiolite. A comparison between structural styles and metamorphism of the two areas (Lipman, 1964; Hotz, 1974; Lindsley-Griffin and Rohr, 1977) suggests that the rocks along the western edge of the Trinity sheet (Fig. 1) have experienced a different tectonic history than the main body, and may even be entirely unrelated to it. This strip of ultramafic and mafic rocks, as shown in Figure 1, is continuous with the Trinity sheet along its southwestern edge but near Callahan diverges from the main sheet and continues northward as a thin septum along Scott Valley to northeast of Yreka.

### Rocks of the Overthrust Sheets

To the west and northwest, the Trinity ophiolite and the juxtaposed melange are overlain by a complex of imbricated thrust sheets (Figs. 3, 4) which are composed of sedimentary and metasedimentary rocks (Hotz, 1974; Lindsley-Griffin and Rohr, 1977). As shown on Figure 3, these include the Gazelle Formation, calcareous quartzite and metasiltstone which may be correlative to the Moffett Creek Formation (Potter and others, 1977; Hotz, in press), and phyllites which were originally assigned to the Duzel Formation of Wells and others (1959) and more recently to the Duzel Phyllite (Hotz, in press).

The Gazelle Formation (Fig. 3) consists of a sequence of siliceous shales interbedded with tuffaceous siltstones, and cherty sandstones and granule to pebble conglomerates. Rare interbeds of a very sandy, organic-rich, rusty-weathering black limestone are also present. The shales are commonly thin bedded, whereas the coarser interbeds vary in thickness from a few centimeters to about .5 meters. Rocks of this thrust plate are probably Siluro-Devonian in age (A.J. Boucot, pers. comm., 1974; Savage, 1977) and formed in a quiet, fairly deep marine environment. They exhibit a single generation of folds and have undergone little or no metamorphism. The Gazelle Formation thrust plate directly overlies the Trinity ophiolite in some places, whereas in other places it overlies the melange sheet which in turn overlies the ophiolite (Figs. 3, 4).

Although I have mapped the lower contact of the Gazelle Formation as a low-angle fault, Potter (1977), in discussing a locality just north of my area, describes this contact as a "sheared depositional contact". Such apparently contradictory evidence as to the nature of contacts is common in melange terranes (Hsu, 1968; Cowan, 1974; Maxwell, 1974). According to Hsu (1968), a contact between a coherent rock-stratigraphic unit and a tectonic melange which seems to be allochthonous in one locality and autochthonous in another may be evidence that the two are coeval. Thus, the Gazelle Formation may have been deposited over the melange while the melange was forming.

Thrust over the Gazelle Formation are rocks which probably belong to the Moffett Creek Formation (Potter and others, 1977; Hotz, in press). This thrust sheet, at least in the area of Figure 3, is composed of interbedded phyllite and calcareous quartzite. The rather distinctive quartzite is an extremely hard, fine-grained, gray-brown, quartz-rich rock, often massive in appearance, which weathers to a light brown color. This quartzite-phyllite sequence exhibits graded bedding, convolute bedding and other features typical of turbidite sequences, and it probably represents a portion of a deep-sea turbidite fan. It is also characterized by zones of disrupted bedding, slump folds, and sedimentary breccia, and thus probably formed at least in part where relatively steep slopes were present. These rocks have been metamorphosed to very low greenschist facies and exhibit at least two generations of tectonic folding (Lindsley-Griffin, in prep.). Along Willow Creek, just north of the area of Figure 3, they are faulted against the Trinity ophiolite along high-angle faults.

The phyllites (Fig. 3) which are thrust over both the calcareous quartzite and the Gazelle Formation are green, chlorite-rich, thin-bedded phyllites with sparse interbeds of very fine metasiltstone. They probably should be placed in the Duzel Phyllite (Hotz, in press). They have been metamorphosed to low greenschist facies, perhaps a little higher than the calcareous quartzite, and exhibit at least two, possibly three, generations of folding (Lindsley-Griffin, in prep.). These phyllites probably represent distal turbidites, but their age is not known.

### Younger Igneous Rocks

The Trinity ophiolite is cut by numerous dikes and stocks, many of which also intrude the sedimentary and metasedimentary rocks of the overlying thrust sheets. Some of the dikes can be traced upward into volcanic rocks which overlie both the Trinity ophiolite and the melange in numerous localities (Fig. 3, Unit "v"). These younger volcanic rocks are spilites and keratophyres, commonly characterized by large green pyroxene phenocrysts. They consist of massive to pillowed lavas and breccias, with rare interbeds of tuff or volcanogenic graywacke. Field relations clearly demonstrate that these volcanic rocks are distinct from the ophiolitic volcanics and that they are younger than both the ophiolite and the melange. Field relations further suggest that these younger volcanic rocks were erupted through the ophiolite and the melange after their juxtaposition. Unfortunately, their relationship to the other thrust plates (Gazelle, calcareous quartzite, Duzel) is not clear, as the volcanics are faulted against rocks of the Gazelle plate and are not in contact with the other two.

Stocks and dikes of diorite, quartz diorite, and hornblende-feldspar porphyry intrude the Trinity ophiolite, the melange, and the Gazelle Formation, and are frequently localized along major shear zones (Fig. 3). Although these intrusions have not been dated radiometrically, they are clearly younger than the faulting which affected lower Devonian rocks of the Gazelle Formation thrust plate. Intrusions of similar composition into the Trinity ophiolite to the east and south of the area shown in Figure 3 have been dated as 224 m.y. (Castle Crags) and 133 m.y. (Craggy Peak) according to Lanphere and others (1968).

## TECTONIC HISTORY OF THE TRINITY OPHIOLITE

The Trinity ophiolite presumably formed at an oceanic spreading center during early Ordovician time (480 to 455 m.y.). This spreading center which may have been located along a mid-ocean ridge or in a marginal basin was not immediately adjacent to a continent during the early Paleozoic (Lindsley-Griffin, 1976; Churkin and McKee, 1974). Shortly after its formation the ophiolite experienced some folding and faulting (perhaps related to transform faulting or to its emplacement) and was then intruded by the pegmatitic gabbros at about the end of the Ordovician.

The occurrence within the melange of conglomerate which contains clasts of the same age and composition as the Trinity ophiolite (Hopson and Mattinson, 1973) is significant. Either a part of the Trinity ophiolite, or another ophiolite of similar age and composition, was situated in a location which permitted it to shed large clasts into conglomeratic deposits, or possibly into submarine debris flows. Geologic relationships suggest that this may have occurred as early as middle Ordovician, but no later than early Devonian. Thus, sometime during that interval, oceanic crust may have been emplaced on the leading edge of the overriding plate as uplifted obducted slabs, or perhaps serpentinization of the leading edge of the overriding plate occurred due to dewatering of subducted sediments, permitting it to be uplifted.

The sedimentary rocks now included as blocks within the melange apparently formed during the period from middle Ordovician to early Devonian; at least, rocks older or younger than that have not yet been observed within the melange. However, the age of formation of the melange in a subduction zone can not be determined solely from the ages of blocks contained within the melange (Hsu, 1968). It is possible that the melange was forming continuously during the time from middle Ordovician to early Devonian, or it may have formed only after the early Devonian, or it may have formed in several stages at various times.

If the contact between the melange and the Siluro-Devonian Gazelle Formation is indeed depositional in places (Potter, 1977), then the two are coeval, at least in part. Thus, a subduction zone could have been active in the early Devonian, and possibly during the late Silurian. Since there is no evidence for a nearby volcanic arc in the early Devonian, this subduction zone may have been too short-lived for an arc to develop. An alternate possibility is that the dip of the subduction zone was very shallow, causing the volcanic arc to develop at a great distance from the trench.

Since the melange is not significantly metamorphosed and exotic blocks are rare, a reasonable environment for its formation would be the upper portion of the trench slope, near the trench-slope break (Karig, 1974). The presence of some metamorphosed sedimentary and ophiolitic fragments suggests, however, that some in-shuffling of deeper basement and subducted sediments was occurring--probably along high-angle reverse faults dipping away from the trench. Portions of the sedimentary rocks within the melange may have been trapped in fault-controlled basins, supplied by density flows or by gravity-sliding of blocks formed elsewhere.

The fact that the pegmatitic gabbros intrude only the ophiolite and not the melange suggests that the thrust sheet consisting of melange arrived in its present position relative to the ophiolite after about 430 m.y., or sometime after the end of the Ordovician. However, the actual time of arrival of the melange and of the other thrust plates cannot be dated precisely. Nevertheless, it would not be unreasonable to conclude that the emplacement of the melange sheet, and possibly the other thrust sheets as well, was related to the emplacement of the strip of ultramafic and amphibolitic rocks along the western edge of the thrust complex (Fig. 1). K-Ar dates for the metamorphism of these amphibolites of 382 and 391 m.y. (Hotz, 1974) suggests emplacement of this ultramafic belt in the early Devonian, which is also the most likely time for the melange to have formed.

The absence of a well-developed, thick island arc sequence of early Paleozoic age in the Klamath Mountains is puzzling, since early Paleozoic island arcs have been mentioned frequently in the literature (Hamilton, 1969; Condie and Snansieng, 1971; Churkin and McKee, 1974; Burchfiel and Davis, 1975; to name only a few). The earliest indisputable preserved island arc sequence in the eastern Klamaths is middle to late Paleozoic (D'Allura and others, 1974). These rocks are middle Devonian (Eifelian, Boucot and others, 1974) through Permian (Strand, 1964) in age, and are overlain unconformably by Triassic and Jurassic volcanic and sedimentary rocks (Fig. 1). This sequence probably is correlative with the 7-km-thick sequence of the same age in the Sierra Nevada (D'Allura and others, 1974).

Although no thick island arc sequence is known to exist for the early Paleozoic in the Klamath Mountains, it is probable that there was some volcanic activity nearby, and island arcs may have existed at some distance. Many of the sediments which make up the various components of the melange overlying the Trinity ophiolite are at least partially volcanic in origin (Table 1). It might be hypothesized that any volcanic arc which developed could lie to the east of the present location of the Trinity ophiolite, conveniently hidden by the late Paleozoic, Mesozoic, and Cenozoic volcanic arcs now located there (Fig. 1). An alternate hypothesis might be that such an arc was later removed by transform faulting. However, the most reasonable conclusion is that no island arc developed in the near vicinity of the Trinity ophiolite during the early Paleozoic, perhaps because subduction during that time was too short-lived to produce more than minor volcanic activity.

The Devonian or younger volcanics which unconformably overlie both the melange and the Trinity ophiolite (Figs. 3, 4) are very minor in volume, and it is probable that they are related to the development of the post-lower Devonian island arc sequence (shown as upper Paleozoic on Fig. 1) lying to the east and southeast of the Trinity ophiolite. Thus, during the late Paleozoic, the Trinity ophiolite, the juxtaposed melange related to it, and possibly some or all of the overlying thrust plates, lay in the fore-arc region of an island arc and was the site of minor volcanic activity.

## CONCLUSION

The Trinity ophiolite meets the general stratigraphic and petrologic requirements for an ophiolite, but it has some peculiarities which either are not part of the definition of ophiolites (Anon., 1972) or which are different from most previously described ophiolites: 1. The ultramafic rocks include a relatively large proportion of lherzolite and feldspathic lherzolite, although both the feldspar and the clinopyroxene are very small and thus difficult to recognize in the field. Are there other ophiolites in which lherzolite is also common but difficult to recognize and map? 2. Sheeted structure does not appear to be present in the diabase. There is a distinct diabase zone, but internal structures are complex, and cross-cutting relationships suggest several sequential episodes of dike and sill intrusion. Is it possible that this is characteristic of most ophiolites or of ophiolites formed in a particular tectonic setting? 3. Early (pre-emplacement) folding and faulting appear to have been important in the Trinity ophiolite. Is this related to some of the other peculiarities of the Trinity ophiolite, or is it common in all ophiolites?

Like many other ophiolites, the Trinity has experienced a long and complex tectonic history since its formation. Knowledge of regional relationships is still incomplete, and there will always be some questions that can never be answered. Thus, it is not possible at this time to construct a specific model for the emplacement and subsequent history of the Trinity ophiolite.

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# THE GEOLOGY AND PETROLOGY OF THE DEL PUERTO OPHIOLITE, DIABLO RANGE, CENTRAL CALIFORNIA COAST RANGES

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## ABSTRACT

A dismembered remnant of the late Jurassic ophiolitic basement of the Great Valley sequence is exposed in the Del Puerto (Red Mountain)area of the northern Diablo Range. Rocks representing several levels within the original ophiolite section are preserved in four fault-bounded blocks along the major tectonic contact between the Great Valley sequence and the Franciscan Complex. The largest block is a mass of alpine-type (metamorphic tectonite) peridotite consisting of: 1) a unit of refractory harzburgite (depleted mantle), 2), a unit of dunite and lesser wehrlite containing podiform chromitites and mafic minerals somewhat less magnesian than those of the harzburgite (interpreted as a penetratively-deformed olivine+chromian spinel±clinopyroxene cumulate), and 3) various kinds of serpentinites, including antigorite schists. A fault marked by distinctive black serpentinite separates the tectonized peridotites from the overlying constructional parts of the ophiolite. Two small fault blocks contain plutonic igneous rocks from the middle levels of the original section. These consist of peridotite and gabbro cumulates, hornblende gabbros, and hornblende quartz diorites considered to be parts of a single differentiated mafic pluton. These rocks have been intensely fractured, faulted, and intruded by dikes and irregular bodies of diabase, microdiorite, and plagiogranite. The top of the ophiolite is preserved in a block of volcanic flows and breccias and hypabyssal intrusions, mainly sills. Earlier mafic to intermediate lavas may represent liquids periodically tapped from the differentiating pluton, whereas younger intermediate to silicic lavas are apparently related to the intrusive bodies cutting the pluton. All igneous rocks in the ophiolite are the products of differentiation of hydrous low-K subalkaline basaltic magma, in which an early trend of iron-enrichment was terminated by crystallization and settling of hornblende and lesser magnetite, the ultimate products being extremely silica-rich plagiogranites and quartz keratophyres. Deposited on the ophiolite volcanic rocks are tuffaceous cherts interbedded with volcaniclastic sandstones derived from an active calc-alkaline volcanic arc; ophiolitic debris is absent. Zeolites and other secondary minerals in the sediments are ascribed to burial metamorphism, but most of the alteration of the ophiolite rocks is attributed to low-P metamorphism contemporaneous with igneous activity.

The fragmental nature of the occurrence and the pervasive secondary alteration impede satisfactory evaluation of the tectonic setting of ophiolite formation, but a marginal basin origin is considered most probable. Formation within an island arc is also plausible, but generation at a normal mid-ocean ridge, as now understood, is unlikely.

## INTRODUCTION

Ultramafic rocks, mainly serpentinites, have long been recognized as distinctive and characteristic components of the late Mesozoic Franciscan terrane of the California Coast Ranges (Bailey and others, 1964). Such rocks are particularly prevalent in the vicinity of the fundamental geologic boundary separating the structurally complex and metamorphosed clastic and volcanic rocks of the Franciscan Complex (Berkland and others, 1972) to the west from coeval but essentially undeformed and unmetamorphosed clastic sediments to the east, referred to as the Great Valley sequence (Bailey and others, 1964). The nature of this contact, which is crucial to tectonic and paleogeographic interpretations of the region, is disputed (Taliaferro, 1943; Bailey and others, 1964, 1970; Bailey and Blake, 1969; Ernst, 1970; Maxwell, 1974). Along much of this boundary, the two great lithologic belts are physically separated by substantial exposures of serpentinized peridotite, raising the question of whether the ultramafic rocks are linked in some fashion with rocks of one of the other of the two belts, or are unrelated solid intrusions injected along a preexisting contact.

The recognition that in many localities these ultramafic bodies are intimately associated with various mafic to silicic plutonic and volcanic rocks, forming a typical ophiolite affiliation (Bailey and others, 1970), has had a profound effect on recent geologic models of the Coast Ranges. These authors proposed that the ophiolite occurrences represent fragments of oceanic lithosphere upon which the late Jurassic to latest Cretaceous Great Valley sequence was deposited, and that the Great Valley rocks plus basal ophiolite had formed the lip of an upper, North American continental plate that was thrust over the Franciscan assemblage and its underlying Pacific Ocean crust during late Cretaceous time. The contact between the Franciscan and ultramafic rocks was therefore considered to mark the surficial expression of this major fault, named the Coast Range Thrust.

Whole-rock chemical data from numerous scattered localities within the ophiolite belt have been collected by Bailey and Blake (1974), who stressed the chemical similarities with ophiolite occurrences elsewhere in the world and with oceanic crustal rocks. The data also indicate an impressive chemical diversity among the ophiolitic rocks, especially pronounced with respect to the volcanic lithologies. The highly silicic nature of some ophiolite volcanics from the southern Coast Ranges is singularly striking. It thus appears that the Great Valley ophiolite is far from a single homogenous petrologic entity, and it may in fact be composed of several discrete ophiolite "plates", created by somewhat different processes, in disparate tectonic environments, and perhaps at slightly but significantly different times. Detailed studies of individual occurrences using combined geologic, petrologic, and geochemical tools will be necessary to substantially improve our understanding of the role of the ophiolites in western North American Mesozoic geology. This paper is a brief presentation of some of the results of one such investigation.

## GEOLOGIC SETTING

The Del Puerto (or Red Mountain) ophiolite (locality 5 of Bailey and others, 1970) is located along the eastern margin of the Diablo Range in central California, approximately 100 km southeast of San Francisco. The Diablo Range (fig. 1) is a topotraphic and geologic entity composed of a central core of Franciscan rocks exhibiting a characteristic disrupted structure and complicated metamorphic history (Kerrick and Cotton, 1971; Ernst, 1971a; Cotton, 1972; Raymond, 1973; Cowan, 1974). The core is flanked on nearly all sides by outward-dipping strata of the Great Valley sequence. The two units are juxtaposed across a series of faults that together comprise the local segment of the Coast Range Thrust. The present structural configuration of the range is attributable to postthrusting diapiric uplift of the Franciscan core in late Cenozoic time, resulting in rotation of the original thrust surface along the borders of the uplifted block into its current, near vertical position. Erosion has completely stripped the upper plate rocks from most of the elevated area, revealing the subjacent Franciscan assemblage. Because of this late movement, original, undistrubed contacts along the Coast Range Thrust are rarely, if ever, preserved (Raymond, 1973).

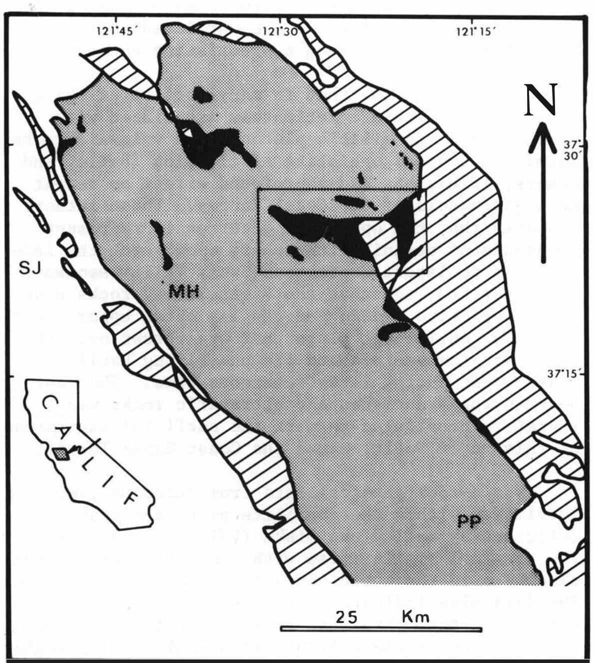



Figure 1. Geologic sketch map of the northern Diablo Range, California, showing distribution of major rock units: Franciscan Complex (gray), Great Valley sequence (diagonal rule), Cenozoic rocks (blank), ultramafic and associated igneous rocks of ophiolite affinity (black). SJ = San Jose, MH = Mount Hamilton, PP = Pacheco Pass. Del Puerto area outlined.

Highly sheared serpentinite commonly intervenes between the Franciscan and Great Valley units along the boundary faults, though it is much less abundant here than in the northern Coast Ranges. Unsheared and only partially serpentinized peridotites are restricted to a pair of large outcrops in the northernmost parts of the Diablo Range--in the Red Mountain area described in this paper, and the Cedar Mountain area to the northwest--both of which are associated with other typical ophiolitic igneous rocks (Bauder and Liou, 1976). The two occurrences may actually be parts of a single ophiolite fragment that have been offset by a north-northwest-trending right-lateral strike-slip fault (Ernst, 1971a).

A simplified geologic map of the Del Puerto area (fig. 2) shows the scattered distribution of the various members of the ophiolite. The immediately surrounding Franciscan terrane is a graywacke-chert melange containing slabs of coherent stratigraphic sections, up to several hundred meters long, in a pervasively sheared argillitic matrix which is easily eroded and thus rarely exposed. The graywacke is jadeite-bearing, and typically possesses a distinct foliation, being similar in this respect to the semischist described by Cowan (1974) from the Pacheco Pass area. Structural trends in the Diablo Range Franciscan are generally oriented north-northwest, but near the ophiolite peridotite they swing sharply east-west parallel to the contact. High-grade blueschists (Coleman and Lanphere, 1971) are distributed irregularly throughout the melange (see Maddock, 1964) and are not exceptionally abundant along the contacts with the ultramafic rocks.

## DEL PUERTO OPHIOLITE

The ophiolite occurs as four separate pieces strewn on both sides of the Franciscan-Great Valley boundary fault, each of which is petrologically distinct and all of which have faulted contacts with adjacent units (see fig. 2). The reconstructed sequence (fig. 3) would presumably consist of, from the base upwards, an alpine peridotite member, a complicated plutonic member (found in two fragments, the eastern one representing a higher structural level), and a volcanic member. The volcanic rocks are overlain by tuffaceous cherts and volcaniclastic sandstones that form the lowest stratigraphic unit of the Great Valley sequence in this area. These strata grade up into clay shales and nontuffaceous, continent-derived sandstones that are the more typical Great Valley sequence rock types. Latest Jurassic (Tithonian) megafossils have been recovered from the shales (Maddock, 1964), placing a minimum age on the underlying ophiolite. The exposed thickness of the Great Valley sequence in the Del Puerto area totals nearly 10 km (Maddock, 1964; Bishop, 1970). The section immediately to the east of the ophiolite outcrops has been described by Bishop (1970).

### Alpine Peridotite Member

The largest single fragment of the Del Puerto ophiolite consists of an alpine-type peridotite of the harzburgite subtype (Jackson and Thayer, 1972) which is one of the best samples of relatively "fresh" ultramafic rock in the entire Coast Ranges. The peridotite is divisible into several distinct mappable units (fig. 4) which are discussed separately below.

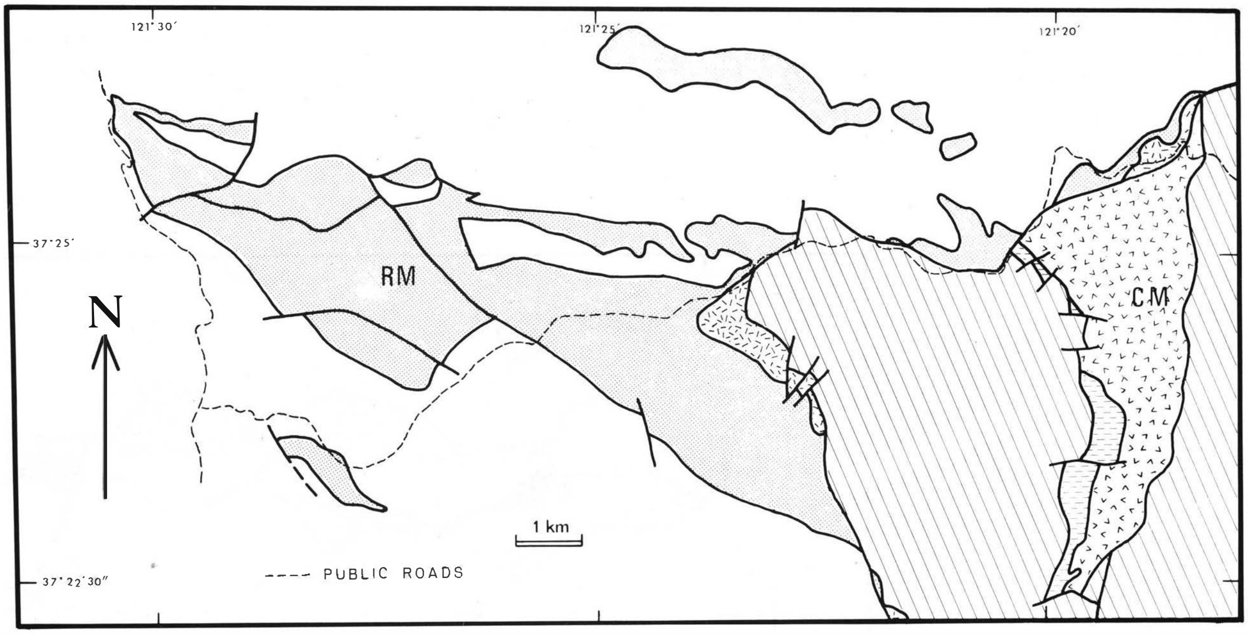



Figure 2. Simplified geologic map of the Del Puerto area. Franciscan Complex (blank), Great Valley sequence shales of late Jurassic-early Cretaceous age (diagonal rule), late Jurassic basal volcaniclastic sandstones and cherts of Great Valley sequence (horizontal dashes). Ophiolite components: alpine peridotite member and serpentinite outliers (dotted), plutonic member (random dash pattern), and volcanic member (V-pattern).

*Harzburgite Unit*.--The most abundant rock type in the alpine peridotite is a rather monotonous, massive harzburgite virtually identical in most respects to the well-studied occurrences at Burro Mountain, California (Burch, 1968; Loney and others, 1971) and Vulcan Peak, Oregon (Himmelberg and Loney, 1973; Loney and Himmelberg, 1976), the most obvious difference being the very poor development of foliation and compositional layering in the Del Puerto body. The weak layering that is discernible displays a consistent orientation (N. 70°-80° E., near-vertical dips) throughout the mass, which therefore cannot have been folded subsequent to emplacement into its current position (Maddock, 1964; Saad, 1969).

Contacts with the surrounding Franciscan metasediments are predominantly high-angle faults. The peridotite in a wide zone (up to 200 m) adjacent to the contacts is intensely sheared and serpentinized. Gravity data suggest that this sheared serpentinite extends beneath the central area of massive harzburgite at a shallow depth (Thompson and Robinson, 1975) so that the peridotite body, despite the steep contacts, has the general form of a horizontal sheet.

The harzburgite, like all alpine-type peridotites, is a metamorphic tectonite. Its microstructure is remarkably uniform throughout, being composed of an interlocking mosaic of anhedral, equant to amoeboid grains of olivine and orthopyroxene averaging 3-4 mm. Minor clinopyroxene is finer grained (<2 mm), and occupies interstitial positions relative to the two other silicates. Chromian spinel (0.5-2.0 mm) is a persistent accessory phase that varies in shape from equant and rounded to highly irregular. Lenticular dunite bodies are scattered through the harzburgite, and these often contain minor spinel concentrations which display apparently cumulus textures (Thayer, 1964, 1969, 1970). In addition to a uniform texture, the harzburgite also has a uniform modal composition (fig. 5), which is matched by a very limited variation in bulk rock and mineral chemistry. The ratio: 100Mg/Mg+Fe* in bulk rock analyses of harzburgites and associated dunites ranges from 90.3 to 91.4, CaO and $Al_2O_3$ are very low, and the alkalies are present in negligible amounts (Bodenlos, 1950; Himmelberg and Coleman, 1968). Electron microprobe analyses of the constituent minerals reveal the following ranges in 100Mg/Mg+Fe* (see fig. 6): olivine, 90.7-91.8; orthopyroxene, 90.6-91.8; clinopyroxene, 93.7-95.0, with 0.4-1.4 weight percent $Cr_2O_3$. Chromian spinels show much greater compositional variation (Cr/Cr+Al=0.33-0.73; $Mg/Mg+Fe^{2+}$=0.42-0.68), and display a negative correlation between $Mg/Fe^{2+}$ and Cr/Al that is a universal characteristic of alpine peridotites (Irvine, 1967; Irvine and Findlay, 1972). Olivine in the associated dunites is slightly more magnesian (100Mg/Mg+Fe*=91.5-93.2), and the coexisting spinel is richer in chromium (Cr/Cr+Al=0.74-0.82).

Olivine microfabric analyses (Evarts, unpub. data) of two typical harzburgites reveals the presence of strong X=[010] maxima oriented approximately normal to the plane of the compositional layering. Combined with the general absence of a dimensional preferred orientation of the olivine in these rocks, and the large percentage of optically strain-free crystals, the strong fabric suggests that syntectonic recrystallization under high-temperature, low-strain-rate conditions was the dominant deformation process Avé Lallemant and Carter, 1970; Loney and others, 1971; Avé Lallemant, 1975). Such conditions are likely to exist within the upper mantle, especially in the vicinity of a spreading center, hot spot, or in the roots of an island arc. The chemically refractory nature of the peridotite is most readily explained as due to extensive partial melting of a lherzolitic parent mantle and removal of most of the segregated melt, perhaps at the same time the rock was being subjected to deformation (for example, Irvine and Findlay, 1972; Ringwood, 1975).

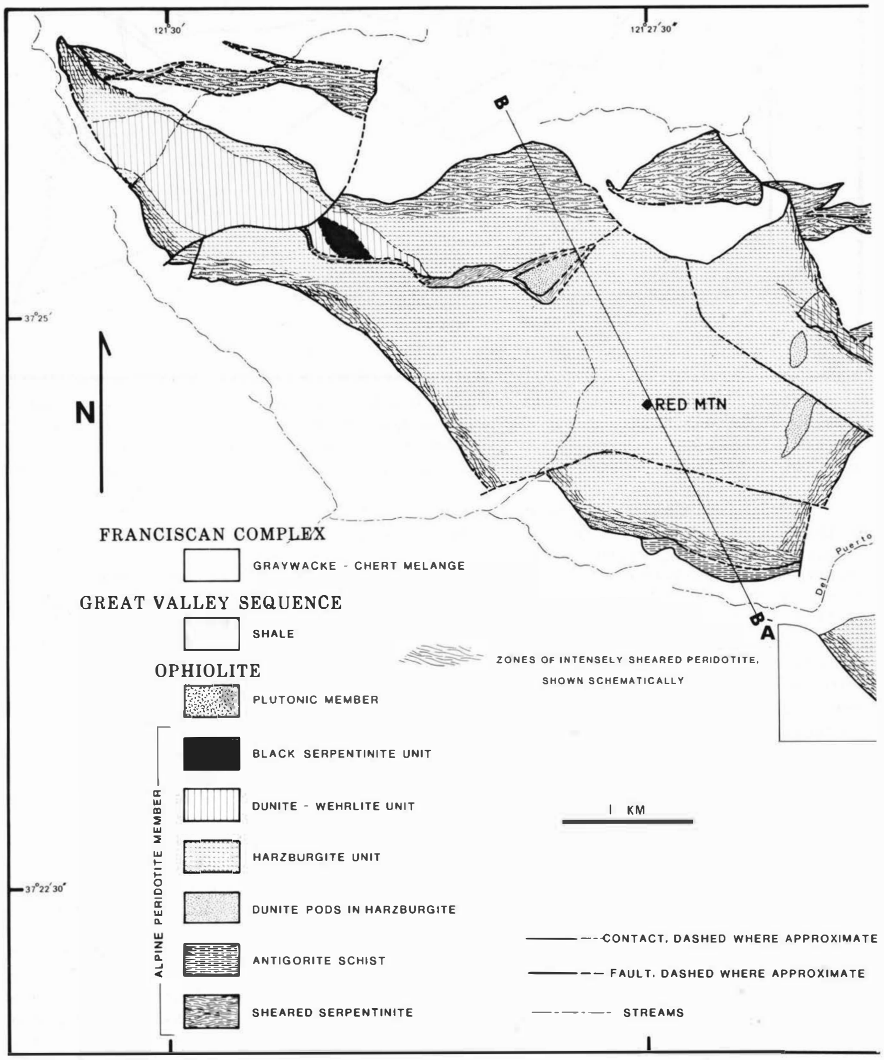
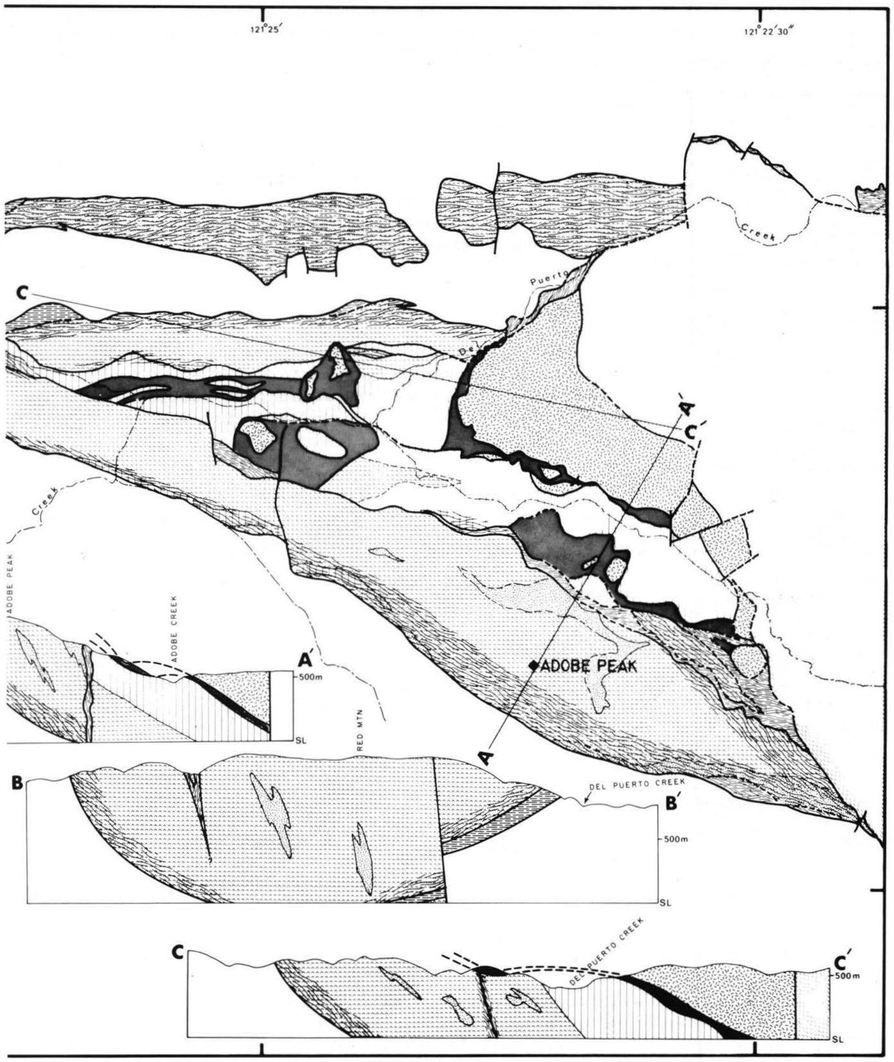



Figure 4. Geologic map and cross-sections of the




alpine peridotite member of the Del Puerto ophiolite.

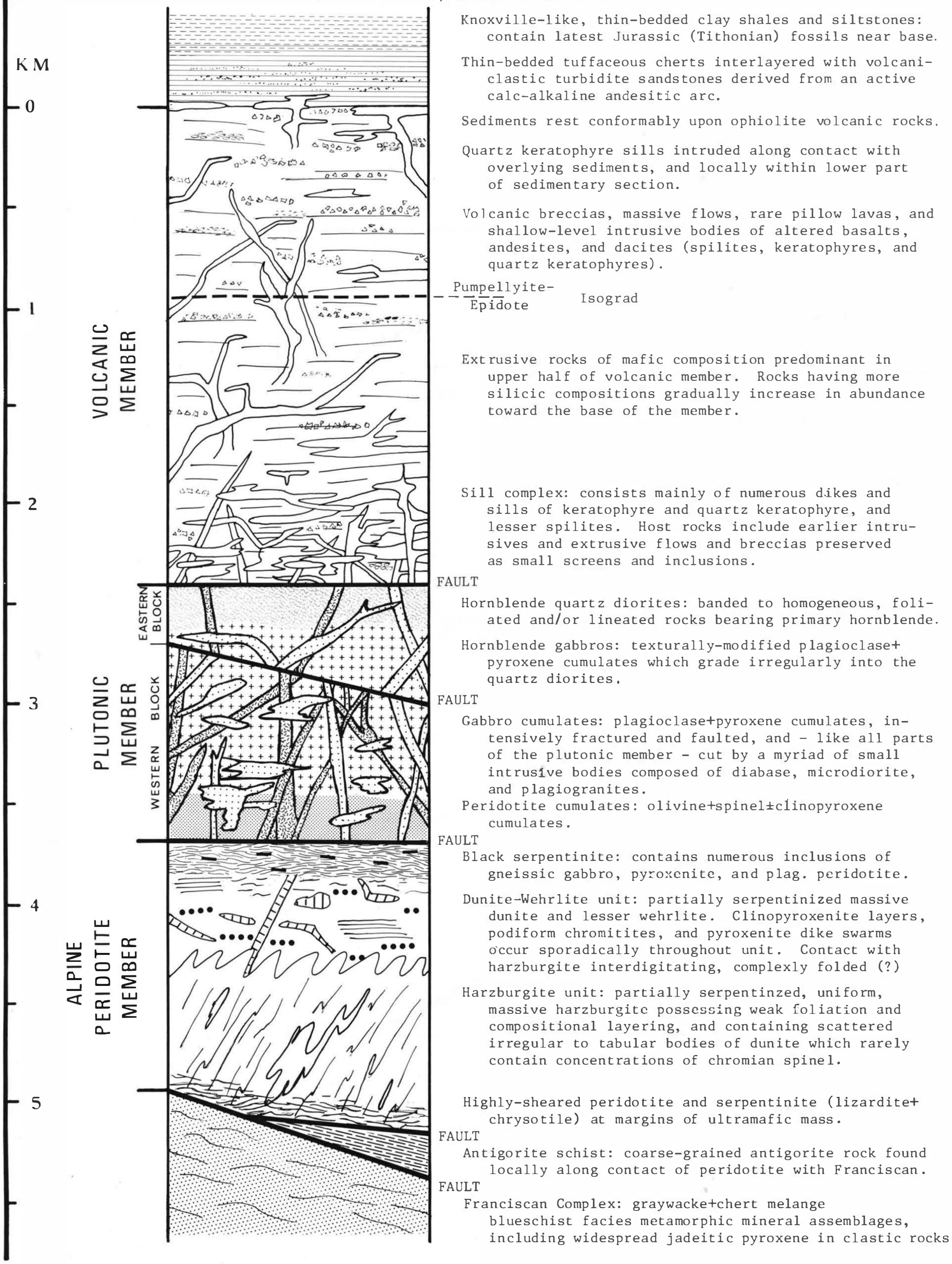



Figure 3. Schematic reconstruction of the Del Puerto ophiolite sequence.

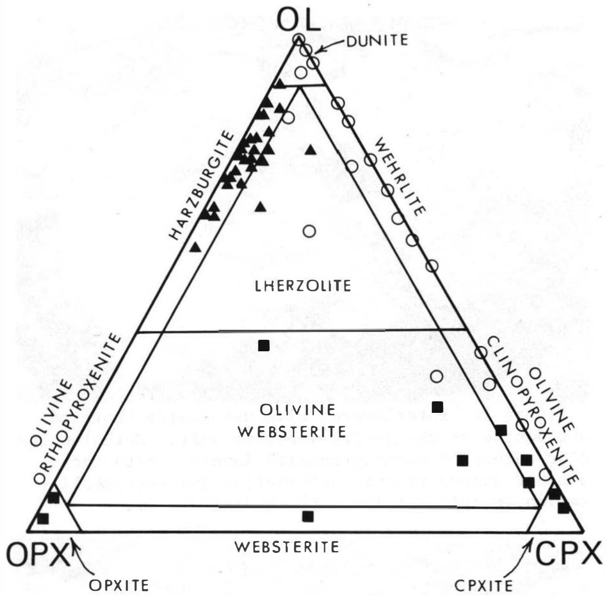



Figure 5. Olivine(OL)-orthopyroxene(OPX)-clinopyroxene(CPX) triangular diagram showing modal composition of rocks from harzburgite unit (triangles,) dunite-wehrlite unit (circles), and pyroxene-rich dikes cutting dunite-wehrlite unit (squares). Rock nomenclature from IUGS classification (Streckeisen, 1976); OPXITE = orthopyroxenite, CPXITE = clinopyroxenite.

*Dunite-Wehrlite Unit.*--A second major unit in the alpine peridotite, as shown on figure 4, is composed primarily of dunites that, unlike the smaller dunite bodies dispersed throughout the harzburgite, invariably contain at least traces of clinopyroxene, and locally grade into wehrlites. Sets of clinopyroxenite layers (fig. 7) occur sporadically, but harzburgites are uncommon, orthopyroxene rarely exceeding 5 percent. The range of modal variation in the major minerals is shown in figure 5. Chromian spinel in the dunites is ubiquitous in amounts of 1-2 percent; in addition, podiform chromitites, as described by Hawkes and others (1942), are scattered through the dunite-wehrlite unit, and in fact are restricted to it. As in the harzburgite unit, plagioclase is totally absent.

The contact between the dunite-wehrlite and harzburgite units is in most places a steeply dipping fault marked by intensely sheared serpentinite. Where not destroyed by faulting, the original contact is extremely difficult to trace in the field, because of extensive serpentinization and because it is exceedingly irregular. It appears, however, that the two units interdigitate in a complex fashion.

Microstructural evidence of plastic deformation of olivine is much more common in the dunites and wehrlites than in the harzburgites: nearly all olivine crystals contain numerous, closely spaced kink-band boundaries, and so-called porphyroclastic textures (Mercier and Nicolas, 1975; Pike and Schwartzman, 1977) are well developed in olivine clinopyroxenites. In addition, the average grain size is distinctly smaller (1-2 mm). Predictably, these deformed rocks possess rather strong olivine microfabrics, similar to those in the harzburgites (Evarts, unpub. data).

A distinctive petrographic feature of the dunites and wehrlites is the delicate interstitial, cuspate to holly-leaf habit of the clinopyroxene (fig. 8). The pyroxene shows no evidence of the extensive strain exhibited by the coexisting olivine, so must have formed after deformation ceased. It is suggested that, during the deformation event, the dunitic rocks contained minor amounts of interstitial silicate liquid, and that postdeformation crystallization of this liquid produced the strain-free, holly-leaf clinopyroxene grains.

Rocks of the dunite-wehrlite unit are readily distinguished from those of the harzburgite unit by bulk-rock and phase chemistry. 100Mg/Mg+Fe* of the bulk rocks can be as high as that in the harzburgites (91.8), but ranges to lower values (82.5). CaO is higher, due to the presence of clinopyroxene, but $Al_2O_3$ remains low (table 1, analysis 1x). The component silicates are of course also lower in magnesia and are more variable. 100Mg/Mg+Fe* in olivine, for example, ranges from 90.5 to about 84.0, and in clinopyroxene, 93.4-89.4. Orthopyroxene coexists only with the more iron-rich olivines, and is similarly iron-rich (88.0-85.0)(see fig. 6). Accessory chromian spinels exhibit a range in Cr/Al nearly identical to that found in the harzburgites, but are decidedly richer in iron and have higher $Fe^{3+}/Fe^{2+}$ ratios (fig. 9).

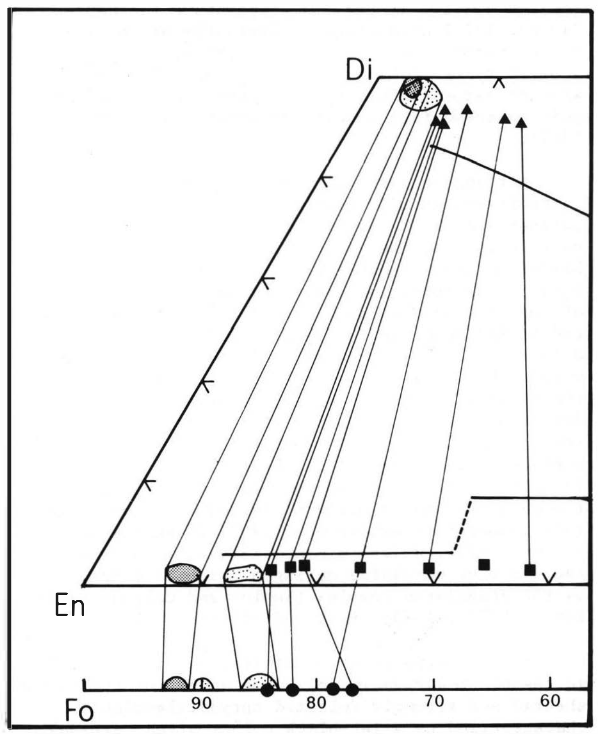



Figure 6. Major element compositions of coexisting clinopyroxene (triangles), orthopyroxene (squares), and olivine (circles) in peridotite and gabbro cumulates. Also shown are compositional fields of same minerals from the metamorphic or tectonite harzburgite (gray) and dunite-wehrlite (stippled) units. Pyroxene trend lines (heavy lines) for the Skaergaard and similar tholeiitic layered intrusions (Brown, 1967) shown for reference.

The dunite-wehrlite unit is provisionally interpreted as consisting of olivine+spinel±clinopyroxene cumulates that underwent considerable solid state penetrative flow prior to complete solidification of trapped intercumulus melt, now represented in the dunites by the holly-leaf clinopyroxenes. This suggestion is supported by the occurrence of cumulus textures in the podiform chromitites, as well as the ferrous compositions and coherent "cryptic" variation displayed by the coexisting silicates. This interpretation is essentially the same as that suggested by George (1975) to explain very similar structural relationships in ultramafic rocks of the Troodos ophiolite complex.

Yet another feature which sets the dunite-wehrlite unit apart from the harzburgite unit is the widespread occurrence in the former of irregular to tabular intrusive bodies of coarse-grained pyroxenites and gabbros. These rocks exhibit a bewildering variety of textures and compositions; some have been strongly deformed along with their host rocks whereas others appear to be unstrained and presumably postdate the deformation event. Counterparts to these intrusives have not been encountered in any other units of the ophiolite; little work has been done on them, and their origin and significance is an unsolved problem.

*Serpentinites*.--The other mappable units that are part of the alpine ultramafic body are serpentinites of various sorts. The most common serpentinites consist of the low-temperature (Wenner and Taylor, 1971) assemblage: lizardite+chrysotile+magnetite±brucite, and include both massive and sheared types (Coleman, 1971a). Formation of this kind of serpentinite may be continuing at present under near-surface conditions (Barnes and O'Neil, 1969).

Antigorite serpentinite that predates the lizardite-chrysotile assemblage is found in small patches scattered throughout the harzburgite, particularly in the western half of the mass. Larger bodies of coarse-grained antigorite rock occur at sporadic localities along the outer margins of the harzburgite unit, most notably at the southern end of Red Mountain, where antigorite schist is in direct contact with schistose jadeitized metagraywacke; the foliation in both rocks is parallel to their mutual contact, which dips beneath the peridotite at a moderate angle. This contact may represent a rare preserved segment of the original fault surface of the Coast Range Thrust in the Diablo Range. Antigorite from a sheared serpentinite lens east of the main ultramafic outcrop yielded an oxygen isotope temperature estimate of 225°C (Wenner and Taylor, 1973), which is compatible with inferred temperatures obtaining during blueschist metamorphism of the Franciscan Complex (Taylor and Coleman, 1968; Ernst, 1971b, 1973; Ernst and others, 1970).

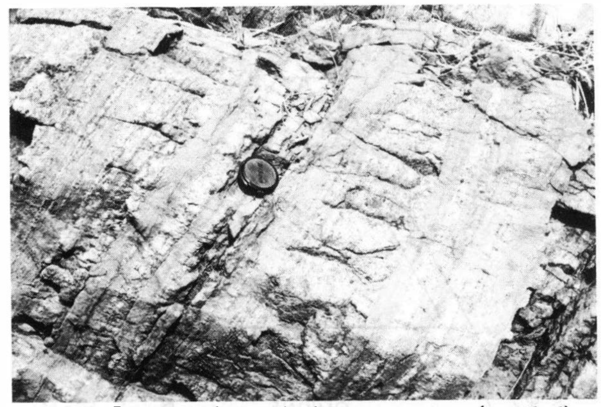


Figure 7. Interlayered clinopyroxenite (mottled) and dunite of the dunite-wehrlite unit. Note lenticular nature of clinopyroxenite layers. Both rocks exhibit strong plastic deformation and recrystallization in thin section. Black disk = 6 cm.

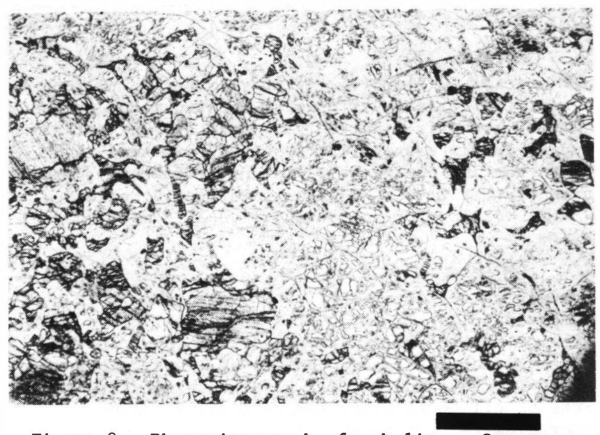


Figure 8. Photomicrograph of wehrlite. Cuspate clinopyroxene (dark gray) occupying interstices between olivine (white) largely altered to serpentine. Plane polarized light. Bar = 1 mm.

A third type of serpentinite, here referred to as the black serpentinite, forms a unit of thoroughly sheared and strongly foliated chrysotile-rich rock characterized by a jet-black color, often carrying overtones of deep blue to purple. This serpentinite is restricted to the fault zone that separates the tectonized ultramafic rocks of the alpine peridotite member from the overlying plutonic member of the ophiolite, in which the rocks are totally free of penetrative deformation features. The fault zone dips eastward beneath the block of plutonic rocks at the eastern margin of the peridotite (figs. 2 and 4) and flattens to a horizontal attitude toward the west. Several small klippen of the plutonic member lie west of the main outcrop, resting on black serpentinite which in turn grades downward into the dunite-wehrlite unit. The age and origin of the black serpentinite fault zone are uncertain, but possibly it was created by imbricate slivering of the ophiolite base of the upper plate during movement along the Coast Range Thrust. The black serpentinite contains numerous tectonic inclusions of mechanically resistant pyroxenites and gabbros obviously derived from the rock units above and below, and the serpentine matrix itself was formed largely from the dunitic parts of the underlying dunite-wehrlite unit.

Plutonic Member

Various lithologic units representative of the middle levels of the ophiolite sequence occur in two lenticular fault blocks alongside the Franciscan-Great Valley boundary fault (fig. 2). The western block rests structurally above the peridotite member as just described, whereas the eastern block abuts against the northern border of the volcanic member; the eastern block contains rock types that would have been found at a higher structural level than those in the western block prior to tectonic dismemberment of the ophiolite (fig. 3).

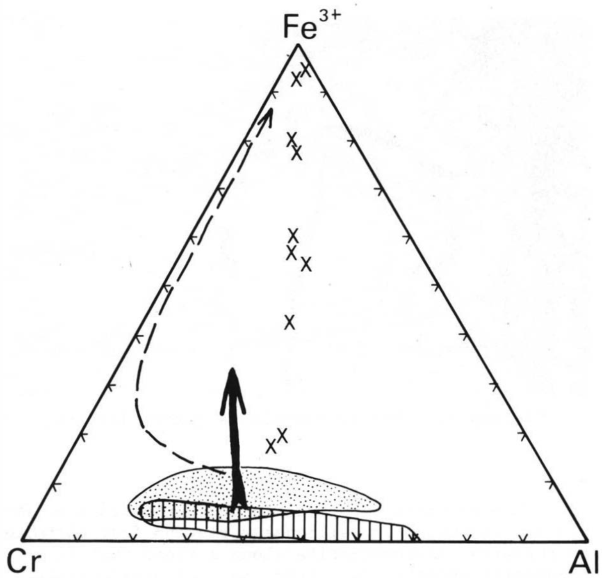



Figure 9. Compositions of cumulus chromian spinels (X's) in peridotites of plutonic member with respect to trivalent cations (ferric iron calculated from microprobe analyses assuming stoichiometry). Fields of spinels from the harzburgite (vertical rule) and dunite-wehrlite (stipple) units also shown. Heavy arrow indicates general compositional trend of chromian spinels from the Stillwater Complex (Jackson, 1969). Dashed line is compositional variation shown by spinels in metamorphosed ultramafic rocks (Bliss and MacLean, 1975; Evans and Frost, 1975).

*Cumulate Rocks*.--The western block consists of peridotite and gabbro cumulates intruded by a profusion of tabular to irregular masses of metadiabase, microdiorite, and coarse- to medium-grained plagiogranite. The entire section is intensely fractured and faulted, so that no large coherent section of cumulate rocks remains.

The ultramafic cumulates consist of cumulus olivine and chromian spinel, commonly accompanied by clinopyroxene, and rarely by orthopyroxene; the pyroxenes may occur as postcumulus phases as well, along with ubiquitous poikilitic plagioclase (fig. 10). All of the peridotite cumulates thus far examined possess heteradcumulus textures (Wager, 1968) in which pyroxene and plagioclase oikocrysts may attain 2-3 cm in diameter. Reaction textures (Jackson, 1961) involving replacement of olivine by either pyroxene are common, and textures suggesting replacement of orthopyroxene by clinopyroxene, and vice versa, are present in a few specimens. In addition, the pyroxenes often exhibit variable degrees of replacement by late-magmatic pale green hornblende. Serpentinization of olivine and concomitant calcium metasomatism of adjacent plagioclase grains are widespread features, resulting in textures and mineralogy identical to those described from the Stillwater Complex by Page (1976).

The appearance of plagioclase as a liquidus phase gives rise to plagioclase-two pyroxene cumulates (gabbronorites) having adcumulus textures (fig. 11). Rocks containing both olivine and plagioclase as cumulus minerals have not been encountered. Strong preferred orientations of tabular orthopyroxene and plagioclase impart a distinct igneous lamination (Wager, 1968) to all of the gabbroic cumulates, but layering due to fine-scale variations in grain size or modal mineralogy is exceedingly rare, nor is there visible lineation (Jackson and others, 1975). The absence of layering or any other sedimentary structure in the gabbros implies deposition under extremely quiescent conditions. As in the peridotite cumulates, postcumulus hornblende may occur as a replacement of the pyroxene crystals. Also notable is the presence of subpoikilitic titanomagnetite in many of the gabbros, in some cases reaching up to about 8 percent of the mode. It is not apparent whether this mineral is entirely a postcumulus space filler, or is a result of extensive overgrowths on settled magnetite crystals.

In the eastern block of plutonic rocks, only gabbroic cumulates are present, and these grade irregularly into hornblende gabbros formed by reaction of settled pyroxenes with intercumulus melt. Relict clinopyroxene cores within the hornblende are commonly present, but in some rocks replacement is total. In habit, the hornblende ranges from small interstitial patches to subhedral crystals up to 1 cm long

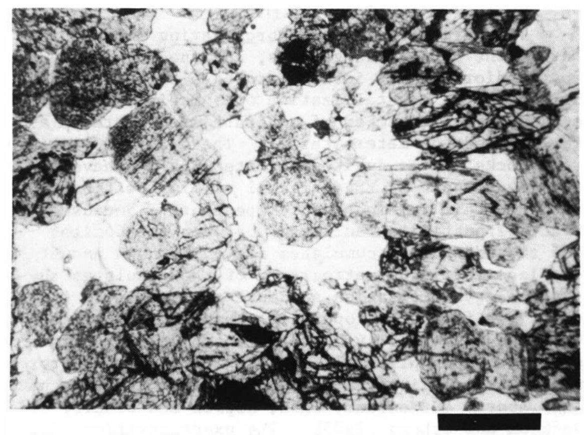


Figure 10. Photomicrograph of typical peridotite cumulate. Euhedral olivine and clinopyroxene (gray) poikilitically enclosed by plagioclase crystal (white). Plane polarized light. Bar = 1 mm.

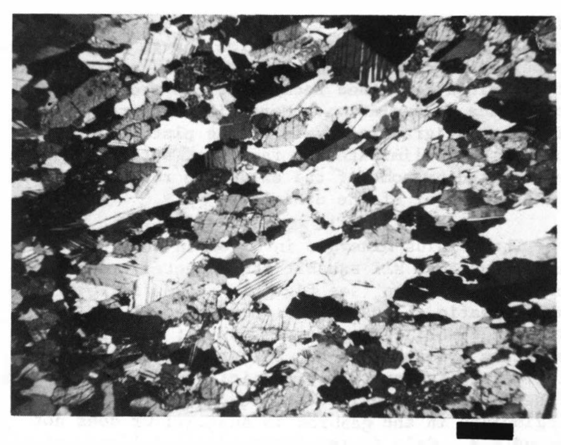


Figure 11. Photomicrograph of typical gabbro cumulate showing adcumulus texture and an igneous lamination formed by tabular plagioclase and orthopyroxene grains. Crossed polars. Bar = 1 mm.

poikilitically enclosing plagioclase tablets. This enclosed feldspar is unzoned, and exhibits a conspicuous preferred orientation inherited from the primary igneous lamination. In contrast, late-stage plagioclase outside of the hornblende crystals is markedly zoned (bytownite-andesine) and unoriented.

Closely associated with the hornblende gabbros are banded hornblende quartz diorites (fig. 12) that are not of apparent cumulus origin. Relationships are unclear due to faulting and poor exposures, but contacts between the hornblende gabbros and quartz diorites appear to be irregular and mutually gradational. The mineralogy of the quartz diorites consists of euhedral prisms of primary green hornblende, subhedral zoned plagioclase (labradorite-oligoclase), anhedral titanomagnetite, and interstitial quartz (see fig. 12). The streaky banding is similar to that observed in some large granitic plutons (for example, Wilshire, 1968), and likewise is possibly a result of segregation of suspended crystals during flow of a crystal-rich mush. Nonbanded quartz diorites are also present, and in these the hornblende crystals form a strong lineation, unaccompanied by a foliation, that is also attributed to magmatic flow.

The peridotites and gabbros having cumulus textures, the hornblende gabbros, and the hornblende quartz diorites are inferred to have been generated by fractional crystallization of a mafic melt within a single magma chamber of moderate size (perhaps a few hundreds of meters thick). The deformed olivine-rich rocks of the dunite-wehrlite unit may have formed the lowest part of the original cumulus section, although this remains highly speculative because the black serpentinite fault zone between the dunites and the undeformed cumulates has an unknown amount of displacement. The petrography of the cumulates suggests a crystallization sequence comparable to those described from other ophiolite complexes: olivine+chromian spinel, olivine+chromian spinel+clinopyroxene± orthopyroxene, plagioclase+clinopyroxene+orthopyroxene± magnetite, plagioclase+hornblende+magnetite±(noncumulus) quartz (Greenbaum, 1972; Hopson and others, 1975; Jackson and others, 1975). The exact position of orthopyroxene in this sequence is not entirely clear from the petrography; orthopyroxene appears to have occurred as a minor cumulus phase sporadically throughout the original section, but only in plagioclase cumulates does it occur in significant abundance (more than about 5 percent). The hornblende quartz diorites are the most differentiated rocks recognized as part of the inferred intrusion. Even so, some have rather mafic compositions (see table 1, analysis 5x, of a typical nonbanded rock) that are probably due to hornblende accumulation. More siliceous rocks from higher stratigraphic levels of the pluton have presumably been eliminated by faulting, but they may be represented by some of the extrusive rocks of the volcanic member of the ophiolite.

The cumulus minerals involved in this differentiation process all exhibit the expected trends of chemical variation (see fig. 6). 100Mg/Mg+Fe* ranges for the mafic silicates are: olivine, 84-76; orthopyroxene 85-62; clinopyroxene, 87-72; hornblende replacing pyroxene, 68-65; primary hornblende, 65-56. Plagioclase is remarkably calcic: intercumulus plagioclase in the peridotites is $An_{95-93}$, whereas cumulus plagioclase in the gabbros is $An_{93-85}$; Or does not exceed 0.3. Only in the interstitial feldspar of the hornblende gabbros and in the quartz diorites do less calcic compositions appear.

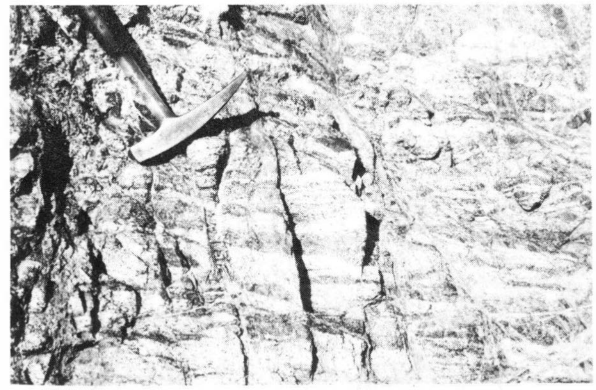


Figure 12. Banded hornblende quartz diorite.

Cumulus chromian spinel in the peridotites exhibits a continuous compositional variation from chromite to chromian titanomagnetite along a trend that is apparently magmatic in origin, and not attributable to subsolidus alteration or metamorphism (Evans and Frost, 1975; Bliss and MacLean, 1975)(fig. 9). In most large stratiform intrusions, such as the Muskox (Irvine and Smith, 1969), a major compositional and stratigraphic gap separates chromian spinel in the ultramafic cumulates from Cr-poor magnetite found at higher levels. The termination of chromite crystallization in these cases roughly coincides with the first appearance of cumulus pyroxene, a connection that can be explained as due to a peritectic reaction relationship involving Cr-bearing pyroxene, chromian spinel, and silicate melt (Irvine, 1967; Irvine and Smith, 1969). The data from the Del Puerto cumulates indicates that clinopyroxene and a spinel phase crystalized simultaneously with olivine, along a eutectic rather than a peritectic type of liquidus boundary curve, and that crystallization of an oxide mineral was more or less continuous all during differentiation. Higher oxygen fugacity in the ophiolite magma may have caused this different kind of behavior (Hill and Roeder, 1974).

*Mafic dike rocks*.--Minor intrusives of mafic to intermediate composition are found cutting through the cumulate sequence everywhere but do not form anything approaching a sheeted complex. Most of these rocks fall into one of two categories: plagioclase-clinopyroxene metadiabases (fig. 14), or microdiorites containing primary igneous hornblende and minor amounts of quartz (fig. 15). The metadiabases have suffered varying degrees of low-grade recrystallization and metasomatism, in extreme cases having been converted into epidosites; even in these rocks, however, relict ophitic to subophitic textures remain. In a few of the less altered samples, fresh clinopyroxene or calcic plagioclase may be found (fig. 14).

The microdiorites have less secondary alteration, the most widespread effect being partial conversion of the calcic cores of zoned plagioclase grains to clinozoisite. These intrusives are generally younger than the metadiabases, and it may be that they postdate the major period of metamorphism responsible for the alteration of the latter. Representative modes of microdiorites are included in figure 13, and a typical analysis is given in table 1 (analysis 7).

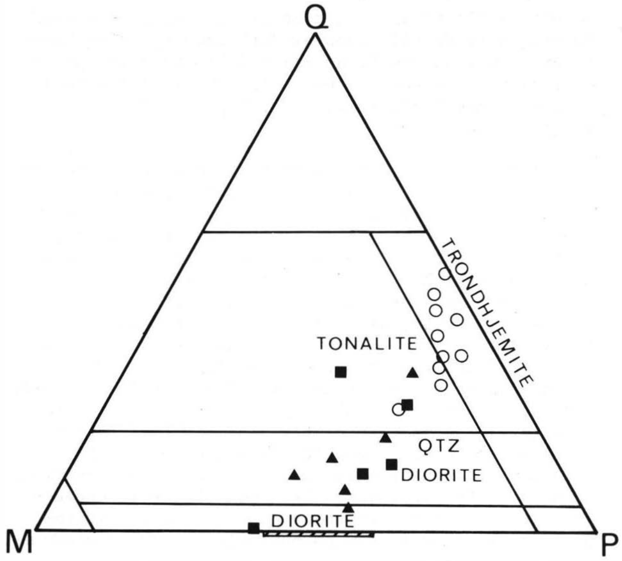



Figure 13. Quartz(Q)-plagioclase(P)-hornblende+ oxides+chlorite(M) diagram showing modal variations in hornblende quartz diorites at top of cumulus section (triangles), microdiorite dikes cutting cumulates (squares), and plagiogranite intrusives cutting cumulates (circles). Bar along base of triangle shows range of hornblende gabbros (texturally modified gabbro cumulates). Rock nomenclature from IUGS classification (Streckeisen, 1976).

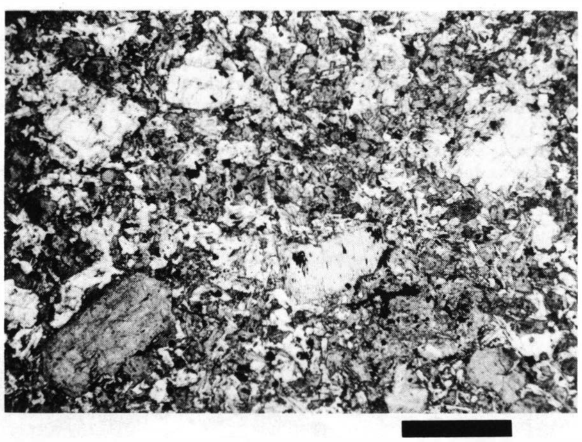


Figure 14. Photomicrograph of metadiabase dike rock from plutonic member. Thoroughly uralitized clinopyroxene (gray) and unaltered bytownite (light gray and white). Crossed polars. Bar = 1 mm.

*Plagiogranites.*--The cumulates are cut by an additional group of even younger, very leucocratic rocks equivalent to the oceanic plagiogranites of Coleman and Peterman (1975). These occur as thin dikes or larger irregular bodies measurable in tens of meters. Coarser rocks from the larger bodies are hypidiomorphic granular to aplitic (figs. 16, 17), and consist of zoned sodic plagioclase (andesine to albite) plus quartz, along with minor amounts of hornblende, magnetite, ilmenite, and chlorite (fig. 13). Granophyric textures exist in some of the smaller dikes (fig. 18). The plagiogranites are totally devoid of potassium feldspar or biotite, and are characterized chemically by extremely low $K_2O$ contents (see table 1, analyses 9x and 10). That the plagiogranites were hydrous and lost water upon crystallization is shown by the development of hornblendite reaction zones within the host rock cumulates at the intrusive contacts. Associated with the emplacement of the leucocratic magmas was injection of numerous late-stage veins of quartz.

Volcanic Member

The volcanic member of the Del Puerto ophiolite consists of a sequence of submarine flows, breccias, and hypabyssal intrusive rocks that crop out as a single fault-bounded block east of the alpine peridotite (fig. 2). The gross stratigraphy within this block strikes north-south and dips uniformly to the west at 60° to 70°. Along the west margin of the exposure, the ophiolite volcanic rocks are depositionally overlain by the basal sedimentary rocks of the Great Valley sequence, whereas the eastern margin is a steep normal fault along which the volcanic section has been uplifted several kilometers relative to the adjacent shales. The exposed thickness of the volcanic pile above the faulted base is approximately 2½ km, somewhat greater than in most other described ophiolites (Coleman, 1971b).

Massive flows, some of them pillowed, are volumetrically subordinate to monolithologic volcanic breccias (fig. 19) that appear to be hyaloclastites having a (now completely altered) tuffaceous matrix. In the lower portions of the section, sills become increasingly abundant. Epiclastic sediments are virtually absent.

Chemically, the rocks range from mafic to silicic, with the average composition of the entire section probably falling in the mafic andesite range. All have undergone low-grade alteration to albite-bearing assemblages and are petrographically representatives of the spilite-keratophyre suite. Chronologically, the mafic extrusive units precede the more silicic rocks; the latter tend to be massive nonvesicular types that are evidently sills emplaced within the older mafic section. Near the faulted base of the

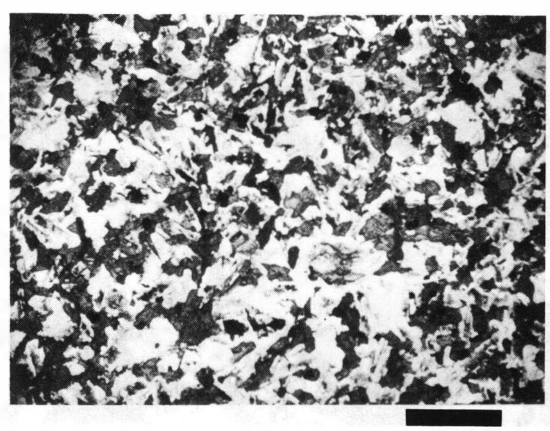


Figure 15. Photomicrograph of typical microdiorite dike from plutonic member. Zoned plagioclase (white), hornblende (dark gray), magnetite (black), and minor quartz (also white). Plane polarized light. Bar = 1 mm.

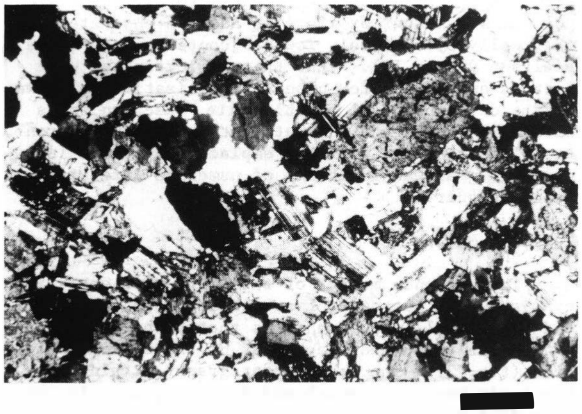


Figure 16. Photomicrograph of hornblende tonalite (plagiogranite). Blocky plagioclase and subhedral hornblende grains, anhedral quartz, and minor chlorite. Crossed polars. Bar = 1 mm.

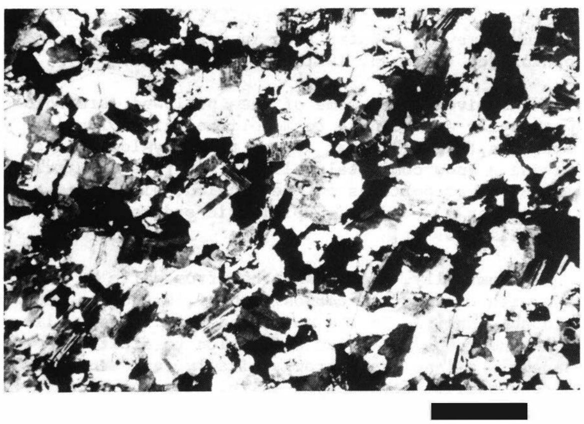


Figure 17. Photomicrograph of quartz albitite (plagiogranite) dike rock. Albite+quartz (95 percent) exhibit allotriomorphic texture; remainder is chlorite and epidote. Crossed polars. Bar = 1 mm.

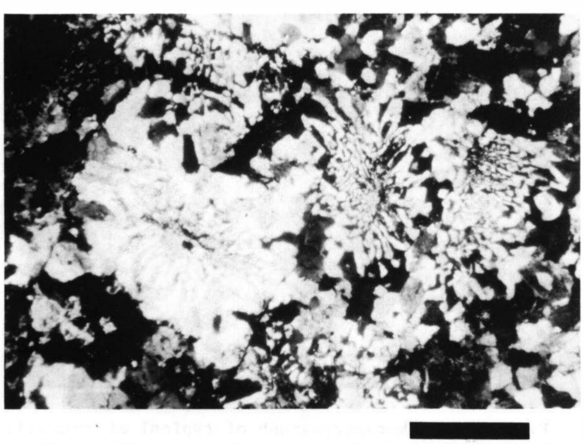


**Figure 18. Photomicrograph of plagiogranite dike rock displaying granophyric intergrowths of sodic plagioclase (gray) and quartz (white). Crossed polars. Bar = 1 mm.**

member, sills of silicic composition are predominant, forming a crude sill complex analogous to those known to exist within the Point Sal and San Luis Obispo ophiolite sequences of the southern California Coast Ranges (Page, 1972; Pike, 1974; Hopson and others, 1975).

The majority of the volcanic rocks are vesicular, sparsely porphyritic varieties bearing phenocrysts of albitized plagioclase. Unaltered igneous clinopyroxene phenocrysts are common in the spilitic samples (fig. 20), which may contain rare recognizable pseudomorphs after euhedral olivine phenocrysts as well. In a number of cases, these olivine pseudomorphs include tiny octahedral crystals of chromian spinel (fig. 21). Quartz joins plagioclase as a phenocryst in many of the highly silicic quartz keratophyres, exhibiting β-quartz morphology and resorption textures common in normal calc-alkaline dacites and rhyolites (fig. 22). As is true of the plutonic rocks, there is no evidence for the prior existence of primary potassium-bearing phases, such as biotite or sanadine, in any of the volcanic rocks. Hornblende is exceedingly rare, being observed as partly altered microphenocrysts in only two porphyritic quartz keratophyres.

A well defined pattern in secondary mineralogy is exhibited by the volcanic member. Iron-rich pumpellyite is widespread and often abundant in the upper 1 km or so of the section, giving way rather abruptly to abundant epidote in the deeper levels. Prehnite occurs in very minor amounts throughout. The pumpellyite-epidote isograd is parallel to the upper surface of the ophiolite volcanic pile. Metastable igneous clinopyroxene persists to levels below that of the pumpellyite-epidote transition, but the approximate horizon at which it is replaced by uralitic amphibole is difficult to pin down due to the paucity of rocks of appropriate (mafic) compositions in the lower half of the sequence. A similar metamorphic pattern occurring within the volcanic portion of the Point Sal ophiolite (Hopson and others, 1975) has been attributed to a combination of hydrothermal alteration and contact metamorphism ("ocean-floor metamorphism"), the fluid involved being circulating sea water, and the heat source being the subjacent magma chamber represented by cumulate and related rocks. A closely similar origin seems applicable to the metamorphism of the Del Puerto locality.

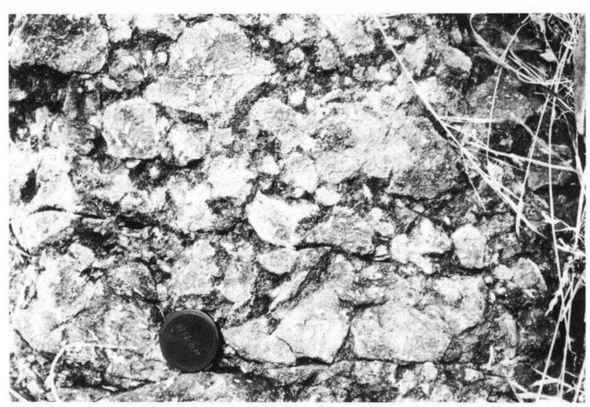


**Figure 19. Volcanic breccia (hyaloclastite). Note large irregular amygdules filled with white calcite. Black disk = 6 cm.**

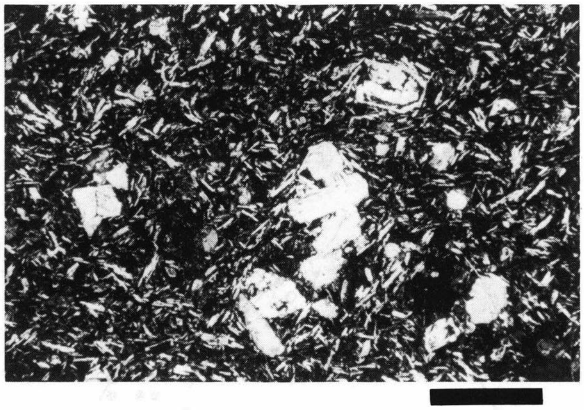


Figure 20. Photomicrograph of metabasalt (spilite) flow from volcanic member. Phenocrysts of albitized plagioclase and unaltered clinopyroxene in a groundmass of feldspar microlites set in mesotasis of chlorite, magnetite, and pumpellyite. Crossed polars. Bar = 1 mm.

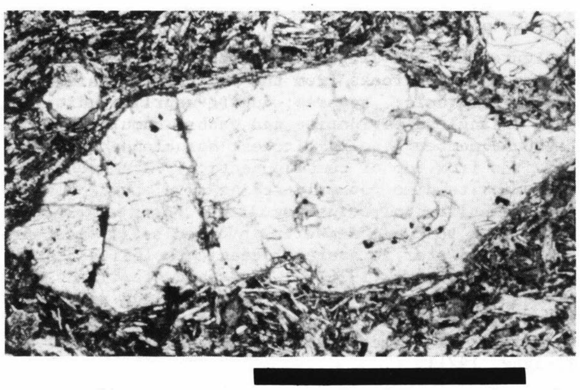


Figure 21. Photomicrograph of calcite pseudomorph of olivine phenocryst in metabaşalt (spilite) flow from volcanic member. Minute black grains within pseudomorph are unaltered chromian spinel octahedra. Plane polarized light. Bar = 1 mm.

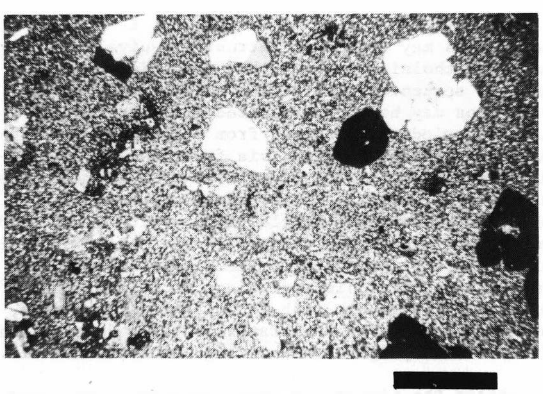


Figure 22. Photomicrograph of porphyritic quartz keratophyre. β- quartz phenocrysts (white, and black-at extinction) in aplitic groundmass composed of albite+quartz. Highly altered plagioclase phenocrysts are also present but difficult to distinguish from groundmass. Crossed polars. Bar = 1 mm.

## BASAL SEDIMENTS OF THE GREAT VALLEY SEQUENCE

Directly above the ophiolite volcanic member rests a distinctive sedimentary unit composed of interbedded tuffaceous radiolarian cherts (fig. 23) and volcaniclastic sandstones and conglomerates derived from a nearby active volcanic source. The beds of this unit, which totals roughly 400 m thick, were laid down conformably on top of the ophiolite, but the original contact has been modified in most places by faulting, extensive replacement of rocks on both sides by laumontite, or emplacement of quartz keratophyre sills along the contact horizon. Locally, sill-like bodies of keratophyre are found at somewhat higher levels within the sedimentary section, indicating overlap in time of silicic ophiolite volcanism and hemipelagic sedimentation. No evidence for subaerial or submarine erosion of the ophiolitic rocks has been recognized.

The coarse-grained clastic sediments of the unit are turbidites consisting of essentially 100 percent volcanic debris of nonophiolitic provenance. The most common lithic clast type (fig. 24) is a typical calc-alkaline andesite (Ewart, 1976) bearing copious phenocrysts of intermediate (labradorite-andesine) plagioclase, clinopyroxene, magnetite, and green hornblende. The ubiquitous hornblende, in particular, provides a striking contrast to the ophiolite volcanics, which lack that phase almost entirely. Stained slabs of volcanic lithic clasts show that the groundmass of these rocks contains considerably more potassium than is found in even the most silica-rich ophiolite samples, but neither sanadine nor biotite phenocrysts have been seen. Clasts of basaltic and dacitic volcanics also occur, but are much less abundant than the andesites.

Except where swamped by calcite cement, the volcaniclastic sandstones invariably contain at least traces of secondary zeolites, both in the rock matrix and replacing detrital feldspar. The most common zeolites are analcime and a member of the heulandite-clinoptilolite series. With the exception of the previously mentioned replacement deposits along the ophiolite-sediment contact, laumontite is restricted to sparse narrow veinlets. Other zeolites that have been identified from limited areas include stilbite, epistilbite, natrolite, and thompsonite. The original glassy material in all of the sandstones has been completely replaced by smectite clay minerals. Prehnite occurs as very fine grained mats in a number of samples, but pumpellyite is apparently absent.

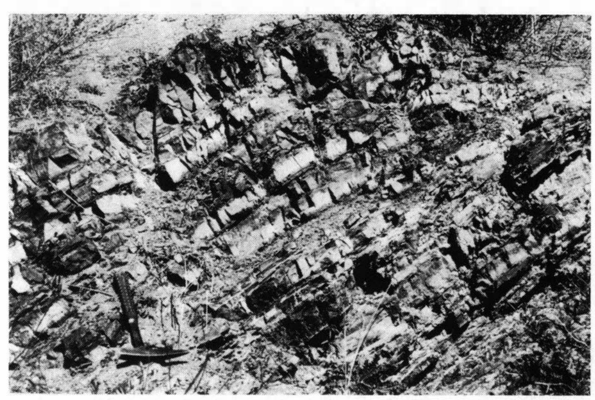


Figure 23. Thin-bedded tuffaceous cherts immediately above ophiolite volcanic rocks.

The very low-grade incomplete alteration represented by these secondary minerals is ascribed to burial metamorphism beneath an estimated 10 km of Cretaceous strata of the Great Valley sequence (compare Dickinson and others, 1969). The effects of this event on the previously altered ophiolite suite was minimal, apparently being limited to deposition of laumontite in scattered veins in the volcanic member and of prehnite in veins within the lower plutonic section. Because pumpellyite appears so abruptly at the top of the ophiolite, and shows the definite relationship to epidote in the volcanic section, it is believed to have formed during the earlier low-pressure metamorphism, although the development of some pumpellyite during burial cannot be entirely discounted. The same problem exists with respect to laumontite, but at least some laumontite must have been generated long after activity within the ophiolite ceased, as shown by the veins within the upper strata of the volcaniclastic sandstone-chert unit.

## DISCUSSION

Most available chemical analyses of rocks from the Del Puerto ophiolite have been plotted on the AFM diagram of figure 25. The trend shown by the data is strictly neither tholeiitic nor calc-alkaline in nature, but possesses elements of both. A simple petrogenetic interpretation of the trend is rendered impossible by two important facts: 1) the extrusive and hybabyssal rocks that define the trend are generally metamorphosed and have undergone metasomatic alteration to an extent that is difficult to evaluate, and 2) the field relations clearly demonstrate that the ophiolite was generated by a series of separate events, and the magmas intruded at one point during the evolution of the complex need not bear any petrologic relation to magmatic events occurring at other times. Nevertheless, it does seem to be true that all of the rocks share some fundamental characteristics, the most basic of which are the subalkaline low-potassium nature of the magmas, and the clear tendency to produce ultimate differentiates very high in silica and low in iron.

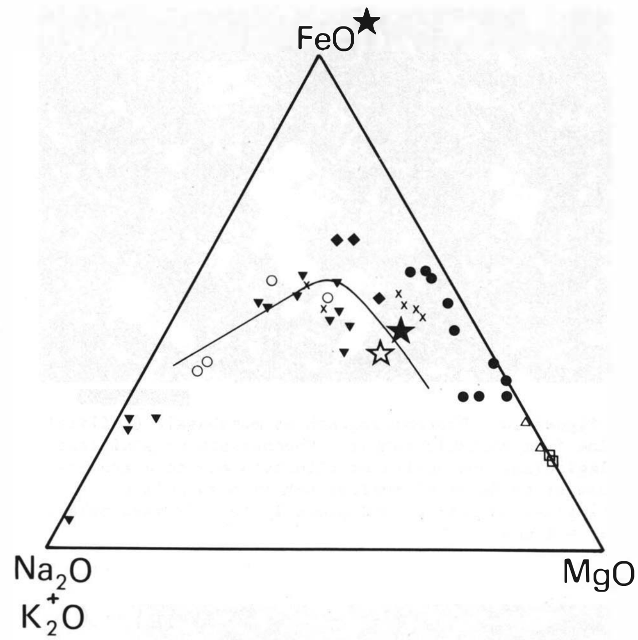



Figure 25. AFM diagram on which are plotted most of the analyzed rocks from the Del Puerto ophiolite. Harzburgite unit: squares; dunite-wehrlite unit: open triangles; peridotite and gabbro cumulates from plutonic member: solid circles; hornblende quartz diorites from top of cumulus section: diamonds; microdiorite dike rocks intruding cumulates: X's; plagiogranites intruding cumulates: open circles; rocks from volcanic member: inverted solid triangles. Large solid star is average oceanic tholeiite from Melson and others (1976). Large open star is average greenstone from Mid-Atlantic Ridge (Melson and van Andel, 1966). The line separates the fields of tholeiitic and calc-alkaline rocks (Irvine and Baragar, 1971).

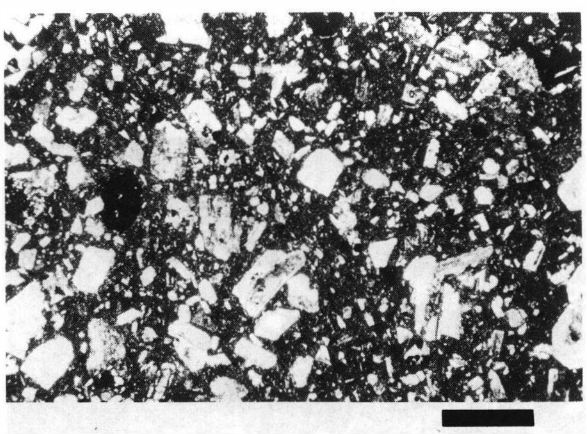


**Figure 24. Photomicrograph of hornblende-clinopyroxene andesite clast from volcaniclastic sandstone bed interlayered with tuffaceous cherts. Plagioclase phenocrysts (white), partly altered to heulandite, are much more abundant than in typical ophiolite volcanic rock (compare with Fig. 20). Crossed polars. Bar = 1 mm.**

The apparent sequence of events as now understood commences with the extrusion of mafic lavas, which, judging from their high vesicularity and the prevalence of breccia units, probably took place in relatively shallow water--possibly at depths less than 500 m. These lavas may represent extrusive equivalents of the low-K tholeiitic magma that gave rise to the cumulate sequence; some of the more silicic early extrusives may have crystallized from supernatant liquids periodically tapped from the top of the magma chamber as differentiation via fractional crystallization progressed. The lower most olivine-rich cumulates (dunite-wehrlite unit) were laid down on a substrate of depleted mantle (harzburgite unit), and penetratively deformed along with the subjacent peridotite shortly thereafter. The black serpentinite zone contains tectonic inclusions of feldspathic peridotite and gabbro with strong gneissic fabrics, indicating that high-temperature ductile deformation affected some plagioclase-bearing cumulates as well.

After the pluton had completely solidified and cooled, mafic to intermediate magmas, some of them hydrous, were injected into it, and crystallized as diabases and microdiorites. These were followed by the extremely leucocratic plagiogranites, which are chemically equivalent to the highly silicic quartz keratophyres that represent the last manifestation of

Table 1. Chemical analyses of selected rocks from the Del Puerto ophiolite, California

| | 1x | 2 | 3 | 4x | 5x | 6 | 7 | 8 | 9x | 10 | 11x | 12x |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 41.0 | 40.5 | 47.6 | 44.4 | 50.9 | 51.0 | 53.4 | 61.7 | 62.1 | 72.4 | 50.9 | 77.5 |
| $Al_2O_3$ | 1.0 | 4.4 | 17.8 | 18.3 | 16.9 | 17.8 | 14.2 | 15.7 | 15.8 | 13.8 | 16.1 | 11.2 |
| $Fe_2O_3$ | 8.4 | 6.2 | 2.9 | 12.4 | 13.1 | 2.2 | 3.7 | 2.9 | 7.3 | 1.0 | 10.4 | 2.0 |
| FeO | -- | 7.6 | 6.3 | -- | -- | 5.5 | 6.9 | 4.5 | -- | 2.5 | -- | -- |
| MgO | 33.4 | 29.4 | 9.0 | 8.0 | 4.8 | 6.6 | 8.4 | 3.5 | 3.1 | 1.1 | 5.8 | .26 |
| CaO | 5.8 | 5.2 | 14.6 | 11.3 | 8.4 | 12.1 | 9.2 | 6.8 | 6.2 | 2.4 | 8.0 | 1.1 |
| $Na_2O$ | .00 | .08 | .52 | 1.0 | 2.0 | 1.5 | 2.2 | 3.1 | 2.6 | 5.2 | 5.0 | 4.8 |
| $K_2O$ | .00 | .08 | .07 | .09 | .37 | .07 | .20 | .40 | .21 | .13 | .10 | 1.3 |
| $TiO_2$ | .04 | .10 | .37 | .49 | .63 | .32 | .45 | .44 | .42 | .40 | .42 | .17 |
| MnO | .13 | .17 | .15 | .20 | .18 | .12 | .18 | .13 | .13 | .11 | .11 | .04 |
| $P_2O_5$ | .02 | .05 | .05 | .08 | .08 | .08 | .08 | .13 | .09 | .10 | .07 | .03 |
| $H_2O^+$ | 10.0 | 5.5 | .70 | 1.8 | 1.6 | 1.5 | 1.3 | 1.5 | 1.6 | 1.3 | 3.5 | 1.34 |
| $H_2O^-$ | -- | .22 | .06 | -- | -- | .11 | .25 | .18 | -- | .19 | -- | -- |
| $CO_2$ | -- | .04 | .02 | -- | -- | .08 | .01 | .01 | -- | .01 | -- | -- |
| Total | 99.8 | 99.5 | 100.1 | 98.1 | 99.0 | 99.0 | 100.5 | 101.0 | 99.6 | 100.6 | 100.4 | 99.7 |
| M-value* | 88.6 | 79.7 | 63.9 | 56.1 | 41.8 | 60.8 | 59.0 | 46.3 | 45.8 | 35.8 | 52.4 | 30.6 |
| S.G.** | 2.85 | 3.10 | 3.04 | 2.97 | 2.91 | 2.87 | 2.87 | 2.80 | 2.80 | 2.69 | 2.87 | 2.65 |

1. Wehrlite from dunite-wehrlite unit. (65-39)
2. Olivine-chromite-clinopyroxene cumulate; plutonic member. (75-11)
3. Plagioclase-clinopyroxene-orthopyroxene cumulate; plutonic member. (64-157a)
4. Plagioclase-clinopyroxene-orthopyroxene cumulate having abundant postcumulus hornblende; plutonic member. (93-2a)
5. Hornblende quartz diorite from plutonic member. (93-16b)
6. Metadiabase dike cutting gabbro cumulates in plutonic member. (75-6)
7. Microdiorite dike cutting gabbro cumulates in plutonic member. (65-22a)
8. Micro quartz diorite dike cutting cumulates in plutonic member. (75-10)
9. Tonalite intruding gabbro cumulates in plutonic member. (65-38c)
10. Albite granite dike cutting peridotite cumulates in plutonic member. (65-33)
11. Pumpellyite-bearing metabasalt (spilite); volcanic member. (99-9)
12. Porphyritic quartz keratophyre sill injected along contact between top of ophiolite volcanic sequence and tuffaceous cherts. (220-14a)

Note: Samples bearing the "x" suffix are X-ray fluorescence analyses performed at the University of California at Santa Cruz using the methods of Norrish and Hutton (1969). For these data, $Fe_2O_3$=total iron as $Fe_2O_3$ and $H_2O^+$=loss on ignition. All others are rapid rock analyses performed by the U.S. Geological Survey using the methods of Shapiro and others (1975); Lowell Artis, analyst.

*M-value is atomic ratio 100Mg/Mg+Fe*.

**S.G. is specific gravity of bulk rock.

ophiolite volcanism. These younger intrusive and extrusive rocks have no obvious source in the parts of the ophiolite that are now exposed; there is absolutely no indication that these melts ever passed through the underlying harzburgite. A second differentiated pluton may be imagined to have been emplaced below the original one, in order to account for the later leucocratic magmas, but if such a body ever existed, it has been completely removed by more recent fault movements. Clearly, development of a comprehensive model for the ophiolite is not possible because several pieces of the puzzle are missing.

The evidence from the cumulus pluton suggests that late-stage differentiates rich in silica and poor in iron can be produced by normal fractional crystallization of a subalkaline magma. The role of hornblende (and lesser magnetite) appears to be crucial in this process. Fractionation of amphibole and magnetite along with plagioclase is believed to be responsible for terminating the trend of iron-enrichment typical of tholeiitic suites during the middle stages of crystallization, and replacing it with the silica-enrichment trend considered more characteristic of calc-alkaline suites (Ringwood, 1974; Cawthorn and O'Hara, 1976). This in turn implies that the magma was hydrous, although the water need not be "juvenile" mantle-derived water--it could just as well have been introduced (ocean?) water. If so, the kind of continuous compositional variation shown by the cumulus spinels and the extremely calcic compositions of the cumulus plagioclase (for example, Yoder, 1969) suggest that the magma acquired its high $pH_2O$ and high $f_{O_2}$ relatively early in its crystallization history.

The tectonic setting within which the Del Puerto ophiolite developed, and consequently its precise role in the late Jurassic paleogeographic and tectonic environment of central California, remains enigmatic. Both mid-ocean ridge (Page, 1972; Hopson and others, 1975) and interarc or marginal basin (Blake and Jones, 1974; Schweikert and Cowan, 1975) origins have been suggested for all or parts of the Great Valley ophiolite, and as emphasized earlier, the ophiolite belt probably contains a number of segments having somewhat different origins. The most definitive evidence on the setting of the Del Puerto occurrence is provided by the sedimentary petrology of the overlying sediments, which clearly place the site of origin near an active calc-alkaline volcanic arc. Such a situation is more likely to exist in a marginal basin than in the open ocean (Dickinson, 1974; Karig and Moore, 1975). Thick volcanic arc sequences of appropriate age (that is, equivalent in age to that of the ophiolite: c. 155-160 m.y., Lanphere, 1971) in the western foothill belt of the Sierra Nevada (Schweikert and Cowan, 1975) would form a logical source area for the volcaniclastic debris.

The petrology of the Del Puerto rocks does not provide definitive evidence concerning the tectonic setting in which the ophiolite originated. Low-K tholeiites such as those in the Del Puerto locality occur in island arcs as well as in oceanic and marginal basin crust (Jakes and White, 1972; Miyashiro, 1974; Hawkins, 1976; Bryan and others, 1976). The chemical differences among them are minor and readily blurred by secondary alteration. The plagioclase+clinopyroxene-phyric ophiolite lavas, however, are petrographically analogous to neither the typical olivine+plagioclase-phyric oceanic tholeiites nor to the phenocryst-rich lavas characteristic of volcanic arcs (Miyashiro and others, 1970; Bryan and others, 1976; Ewart, 1976).

The relative abundance of intermediate to silicic compositions among the ophiolite volcanic rocks is a problem that the Del Puerto occurrence shares with many other ophiolites (Moores, 1969; Moores and Vine, 1971; Coleman and Peterman, 1975; Coombs and others, 1976). Such rocks are extremely rare among all the samples of oceanic crust thus far obtained. Plagiogranite and equivalent low-K volcanic rocks are known, however, from several island arcs, including Japan (Ishiyaka and Yanagi, 1975; 1977), the Caribbean (for example, Longshore, 1965; Bowin, 1966; Mattinson and others, 1973; Gunn and Roobol, 1976), and Fiji (Gill, 1970), and Tonga (Ewart and Bryan, 1972), suggesting affinities of ophiolites with an arc environment, as argued by Miyashiro (1973, 1975). It seems, nevertheless, that low-K tholeiite, given the opportunity to fractionate under appropriate conditions, would produce low-K differentiates regardless of tectonic environment. The apparent abundance of such differentiates in island arcs compared to oceanic crust may be largely a result of more complete sampling of eroded and subaerially-exposed arc rocks compared to submerged oceanic crust. Alternatively, the conditions required to produce significant quantities of leucocratic rocks may be more likely to prevail in arc environments. The absence of a sheeted dike complex at Del Puerto, and the uniform nature of the cumulates, imply a relatively stable, nontensional regime very different from that obtaining under normal spreading centers, but such a regime might be expected within an arc or in a marginal basin where spreading is more diffusely distributed. Thus, all things considered, a marginal basin origin is preferred for the ophiolite, although a primitive island arc environment is also plausible. There are, however, strong arguments against the Del Puerto suite being directly analogous to oceanic crust as now perceived (Bottinga and Allegre, 1976).

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# FRACTURE ZONE TECTONICS, CONTINENTAL MARGIN FRAGMENTATION, AND EMPLACEMENT OF THE KINGS-KAWEAH OPHIOLITE BELT, SOUTHWEST SIERRA NEVADA, CALIFORNIA

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## INTRODUCTION

The Sierra Nevada foothill metamorphic belt is a 450 km long assemblage of remnant continent-derived epiclastics, arc volcanics, pelagic-hemipelagic sediments, and ophiolite slices. The various lithologic units range in age from Ordovician to Jurassic. Lithologic units are lenticular at scales ranging up to 150 km and strike about N. 30°W. parallel to the trend of the metamorphic belt (Fig. 1). Many units are penetratively deformed with a variety of near vertical foliation surfaces. The lithologic units are generally bounded by steep dipping fault and melange zones, but locally depositional contacts can be recognized. From at least latitude 38°30'N southward, latest Paleozoic to possibly early Mesozoic disrupted ophiolite occurs as remnant oceanic basement beneath Triassic to Jurassic arc volcanics and interstratified continent-derived epiclastics. Along the northern part of this segment of the metamorphic belt the ophiolitic rocks occur as scattered basement exposures surrounded by the younger volcanic and epiclastic rocks (Morgan and Stern, 1977; Behrman, 1978; Saleeby, unpub. field data). Further south in the Kings-Kaweah terrane deeper structural levels of the foothill metamorphic belt are exposed. Here a nearly continuous 125 km long ophiolite belt occurs with scattered remnants of early Mesozoic arc volcanic and epiclastic rocks depositionally above it. The ophiolite belt is named informally the Kings-Kaweah ophiolite belt after the Kings and Kaweah Rivers which transect it. This ophiolite belt constitutes part of the same oceanic basement terrane that is locally exposed further north amidst the arc volcanics and epiclastics.

The Kings-Kaweah ophiolite belt constitutes a significant segment of the foothill metamorphic belt. Within it exist the only remnants of a complete ophiolite succession to be found throughout the entire Sierran terrane. In addition, since it represents the deepest exposure of foothill metamorphic rocks, it affords the best opportunity to study the tectonic and petrogenetic history of the oceanic basement terrane upon which Mesozoic continental margin rocks were deposited. The purpose of this paper is to: 1) give a general description of the ophiolite belt; 2) expand upon critical relationships within the ophiolite belt which bear on the tectonics of ophiolite genesis, deformation and emplacement, and 3) discuss briefly the tectonics of the ophiolite belt with respect to the regional tectonics of the southwest United States. Detailed structural and petrologic data are presented in Saleeby (in press a, in press b). The paleogeographic implications of the ophiolite belt and adjacent metasedimentary and metavolcanic rocks are discussed in Saleeby and others (in prep.). The geochronological evolution of the ophiolite belt is discussed in Saleeby (in prep. a).

Significant conclusions drawn in this paper are: 1) the Kings-Kaweah ophiolite belt originated at a distant oceanic spreading center where cut by a transverse fracture zone; 2) deformation of the ophiolite was progressive and occurred primarily by fracture zone tectonics while in route to the ancient continental margin; 3) the ancient continental margin was fragmented and tectonically eroded along an extension of the fracture zone; 4) the disrupted ophiolite was juxtaposed against the raw edge of the continent as the continental fragments were displaced; 5) the disrupted ophiolite was accreted to the continental margin as the hanging wall of a subduction zone as a result of a change in plate motions; 6) the tectonically accreted ophiolite belt subsequently served as frontal arc basement during subduction related arc activity; and 7) the suture between oceanic and continental basement terranes remained tectonically active as a longitudinal intra-arc deformation zone during arc activity. This model of ophiolite genesis, deformation, emplacement and subsequent tectonic history is believed to be applicable along the entire length of the Sierra Nevada foothill metamorphic belt (Saleeby, in prep. b).

## GENERAL DESCRIPTION OF THE OPHIOLITE BELT AND RELATED ROCKS

### The Ophiolite Belt

Plate V is a general geologic map of the Kings-Kaweah ophiolite belt. The gross structure of the belt is that of a huge tectonic megabreccia with a schistose serpentinite matrix. At the north end of the belt clasts range up to 20 km in length, and are referred to as tectonic slabs (after Hsu, 1968) since they contain internal mappable stratigraphic units. The slabs are collectively named the Kings River ophiolite after the Kings River which transects the slab cluster. The slabs are separated by narrow serpentinite melange zones and by cross-cutting plutons of the Sierra Nevada batholith. Southward from the Kings River area the slabs decrease in size to monolithologic blocks. In doing so the ophiolite belt grades into serpentinite matrix melange. The greater part of the melange is named the Kaweah serpentinite melange after the Kaweah River which transects it.

The entire ophiolite belt has been metamorphosed in the albite-epidote to mainly hornblende hornfels facies (after Turner, 1968) by the Cretaceous Sierra Nevada batholith. Metamorphic recrystallization of the ophiolite belt is in many places incomplete. Thus some insight into original mineralogy of the ophiolite protoliths is available. In addition, even where metamorphic recrystallization is complete, earlier textures and structures are commonly well-preserved. Thus protoliths of the metamorphosed ophiolite belt have been readily deduced from field, petrographic, mineralogical and chemical data. Detailed treatment of this data is not the intent of this paper. For sake of brevity the ophiolite will

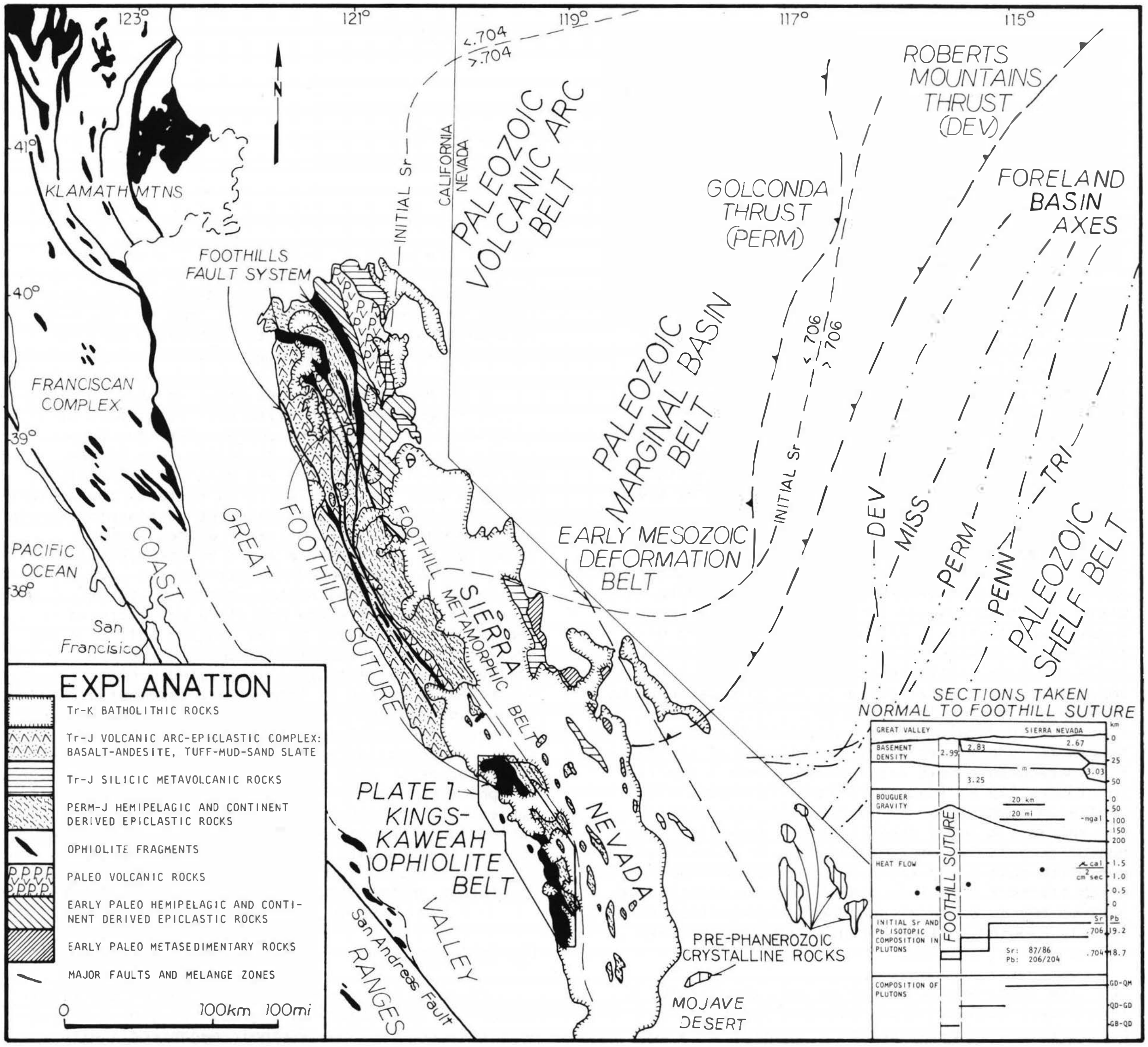



Figure 1. Map showing location of Kings-Kaweah ophiolite belt and significant regional geological features discussed in text. General geology of Sierra Nevada after Clark (1964, 1976), Jennings and Strand (1966), D'Allura and others (1977), Schweickert and others (1977), Saleeby and others (in prep.). Paleozoic regional thrust faults after Burchfiel and Davis (1972), Speed (1977). Paleozoic paleographic belts after Churkin (1974), Stevens (1977). Paleozoic foreland basin axes after Poole and others (1977), Speed (1977), Stevesn (1977). Early Mesozoic deformation belt after Stewart and others (1966), Burchfiel and others (1970), Stevens and Olson (1972), Kelley and Stevens (1975). Initial strontium contours for autocthonous post-Paleozoic igneous rocks after Kistler and Peterman (1973). Major faults and ophiolite fragments of Klamath Mountains and Coast Ranges after Jennings and Strand (1966). Inset at lower right shows plots of several geological parameters taken along a section perpendicular to southern part of foothill suture. Bouguer gravity after Oliver and Robbins (1975); rock density after Cady (1975), Saleeby (1975); heat flow after Lachenbruch (1968); isotopic composition on autocthonous igneous rocks after Kistler and Peterman (1973), Dow and Delevaux (1973), Saleeby (1975), Chen (1977); composition of volumetrically important Mesozoic plutons after Moore (1959), Saleeby (1975), Chen (1977) with gb=gabbro, qd=quartz diorite, gd=granordiorite, qm=quartz monorite.

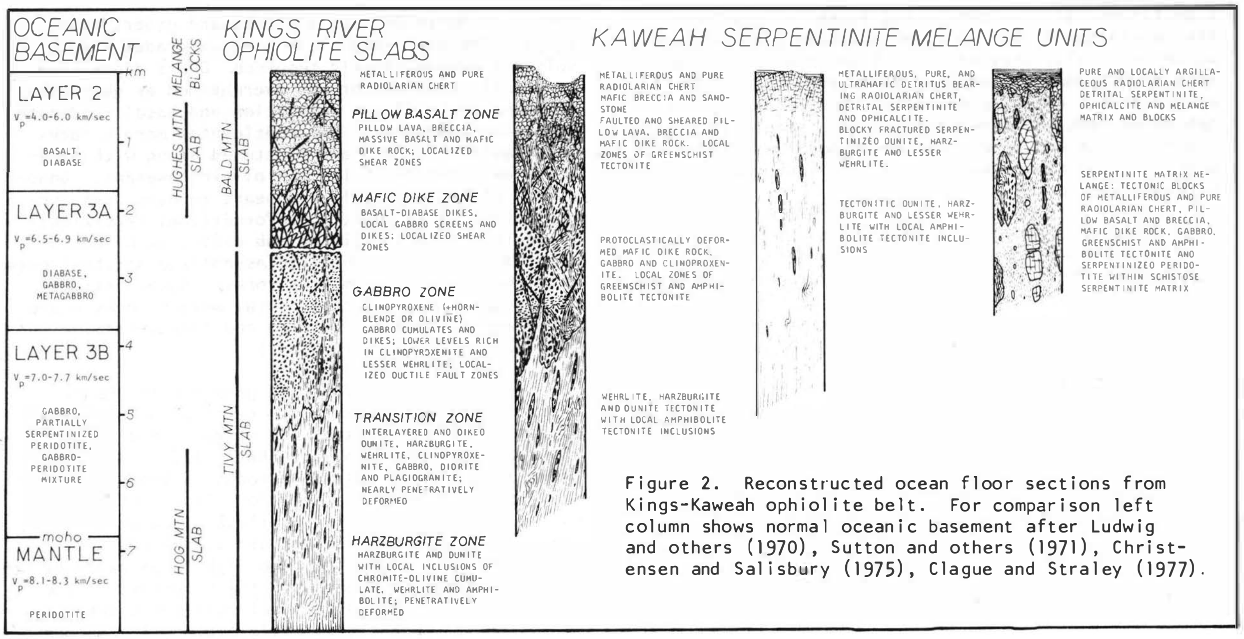



Figure 2. Reconstructed ocean floor sections from Kings-Kaweah ophiolite belt. For comparison left column shows normal oceanic basement after Ludwig and others (1970), Sutton and others (1971), Christensen and Salisbury (1975), Clague and Straley (1977).

be discussed in terms of pre-batholith protoliths. For further information on batholith related metamorphism see Saleeby (1975, 1977, in press a and b, in prep. a).

The Kings River ophiolitic slabs are named after the highest encompassed peak. The slabs are bounded by serpentinite melange zones in most instances. Where not present, the melange zones can be inferred to have existed prior to batholith emplacement. The Hog and Tivy Mountain slabs are predominantly peridotite. These slabs grade into the neighboring melange matrix. Peridotite foliations grade into or are cut by domains of schistose serpentinite. Towards the melange zones the schistose serpentinite domains become dominant until the original peridotite fabric and mineralogy is obliterated producing the melange matrix. Within short distances exotic blocks of gabbro, basalt and chert occur within the serpentinite matrix. The strike of the matrix schistosity is parallel to the long axes of the slabs and to the regional trend of the ophiolite belt.

Within the slabs structures which pre-date melange mixing also occur. Peridotite and gabbro within the Hog Mountain and Tivy Mountain slabs contain a mylonitic foliation. The trend of this foliation is mainly parallel to the trend of the ophiolite belt. However, domains in which foliations have been folded and rotated are common. The folded and rotated domains are truncated by mylonite foliation surfaces which are identical to the folded surfaces. Structural analysis of the mylonites shows that mylonitization proceeded in repeated pulses with early-stage foliation surfaces being truncated, rotated, folded and refolded during succeeding stages of mylonitization. This complex family of foliation surfaces is referred to as $S_1$. Within $S_1$ there is a steep plunging lineation ($L_1$) defined by elongated dimensional markers such as pyroxene porphyroclasts and deformed mafic inclusions. Folding of $S_1$ and $L_1$ was predominately about steep plunging axes. Folds in $S_1$ and $L_1$ are referred to as $F_1$. $F_1$ geometry is variable and complex. Asymmetries suggestive of a dextral sense of motion are not uncommon.

The schistosity of the melange matrix and similar schistosities of the Hog Mountain and Tivy Mountain slabs are referred to collectively as $S_2$. Along both margins of the Hog Mountain slab, and along the western margin of the Tivy Mountain slab $S_1$ grades into $S_2$. This is manifested by a progressive increase of schistose serpentine relative to flattened and streaked out olivine and pyroxene. Along the eastern margin of the Tivy Mountain slab $S_2$ cuts sharply across $S_1$. Local zones of both $S_2$ cutting $S_1$ and $S_1$ grading into $S_2$ occur within the ultramafic slabs. $S_1$ in northwest orientations grades into $S_2$, whereas $S_1$ in other orientations is cut by $S_2$.

Mafic slabs of Bald Mountain and Hughes Mountain contain a relict shear fabric which is only locally developed in each of them except for the northwest part of the Hughes Mountain slab where it is penetrative. The shear fabric is also steeply dipping and parallel to the regional trend of the ophiolite belt. It is thought to be equivalent primarily to $S_1$ of the Hog Mountain and Tivy Mountain slabs. Similar shear fabrics in mafic melange blocks appear to be surfaces along which the blocks were rifted apart during melange mixing. Thus the shear fabric may in part also be equivalent to $S_2$ of the ultramafic slabs and melange zones. As will be discussed later, development of $S_1$ and $S_2$ are thought to have partly overlapped in time.

The slabs contain various segments of the original ophiolite stratigraphy. A reconstructed stratigraphic section is shown in Figure 2. Next to the graphic section the intervals spanned by the Kings River slabs are shown. The reconstructed stratal thicknesses are taken from the Tivy Mountain and the Bald Mountain slabs which fit immediately adjacent to one another when the ophiolite is palinspastically restored to its pre-batholith configuration (Fig. 3). The reconstructued ophiolite section consists from the base up of: 1) greater than 4 km harzburgite-dunite with traces of chromitite, wehrlite, clinopyroxenite and gabbro; 2) 2.5 km mafic-ultramafic transition zone composed of the same rocks except wehrlite, clinopyroxenite and gabbro are more significant; 3) 2 km gabbro and lesser clinpyroxenite cumulates; 4) 0.7 km basalt-diabase dike complex which is locally sheeted; 5) 1.8 km basaltic pillow lava and pillow breccia; and 6) greater than 20 m metalliferous radiolarian chert. The reconstructed ophiolite section is interpreted as a sample of oceanic crust and upper mantle. A more detailed discussion of the ophiolite section is presented in Saleeby (in press a).

Southward from the Kings River area the ophiolite fragments decrease in size to form tectonic blocks in serpentinite melange. The large gabbro block at the north end of Smith Mountain is intermediate in size between the Kings River slabs and common melange blocks which range between 1 km and several meters in diameter. Geophysical data (Saleeby, 1975) indicates that the Smith Mountain block continues in the subsurface for at least 7 km north of Smith Mountain. A significant feature of the serpentinite melange is that it consists only of ophiolite assemblage blocks. Blocks of dunite, harzburgite, wehrlite, clinopyroxenite, gabbro, mafic dike rock, pillow basalt,ophicalcite and radiolarian chert are suspended in schistose serpentinite. Ultramafic blocks usually grade outward into the matrix in a fashion similar to that described for the ultramafic slabs of the Kings River area. In contrast, contacts between matrix and mafic and chert melange blocks are usually sharp.

Melange blocks are invariably elongate parallel to the matrix schistosity and the regional trend of the ophiolite belt. Internal structures of the blocks such as mylonite or metamorphic foliation and shear surfaces are usually oriented parallel to the blocks long axes. Chert blocks are usually tabular in shape with bedding also oriented parallel to long axes. Many melange blocks have transverse extension fractures which are occasionally injected with schistose serpentinite. In many instances blocks have been pulled apart along the tension fractures like large boudins. Local kinks in blocks and small-scale folds in the matrix schistosity occur; these are invariably about near vertical axes with many of them having asymmetries indicating a dextral sense of motion.

Outcrop mapping of the melange revealed a clustering of blocks of similar lithology or lithologies. The clusters are shown as melange units on Plate 1. The melange units appear to be the vestiges of once larger blocks or slabs that have been distended into a myriad of smaller blocks by faulting and injection of the more mobile matrix. Within the melange units there are vestiges of primary igneous and sedimentary contacts between different members of the ophiolite assemblage. As discussed below, some primary contacts formed during melange development. The melange units are interpreted as the mixed remnants of ocean floor stratigraphic successions. Stratigraphic successions reconstructed from the units are also shown in Figure 2. The implications of the reconstructions will be discussed below.

Continental Margin Rocks

Depositional remnants of continental margin rocks occur above the Kaweah serpentinite melange. The oldest of these rocks is a chert-argillite olistostrome complex containing olistoliths of shallow water limestone and interbeds of chert and quartzose to subarkosic sandstone. The shallow water limestone blocks contain late Permian fauna believed to be exotic to North America (Saleeby and others, in prep). The chert-argillite complex grades into a volcanic arc-epiclastic sequence. Chert deposition apparently subsided or was overwhelmed as quartzose to sub-arkosic flysch deposition and basalt-andesite volcanism commenced. The continental margin rocks were faulted, folded and flattened along with late-stage deformation of its ophiolitic basement. Deposition of this assemblage appears to have been syntectonic with abundant intraformational reworking. In addition, local uplifts and exposures of ophiolite basement shed ophiolite assemblage olistostromes into the continental margin rocks. Age constraints on the deposition of continental margin rocks place it between the latest Permian and late Jurassic (Saleeby and others, in prep.).

Middle and late Jurassic gabbroic to quartz dioritic plutonic rocks which cut the ophiolite belt appear to be the roots of the volcanic arc rocks (Saleeby, 1975; Saleeby and Sharp, 1977). An important feature of the plutonic rocks is that they were emplaced late in the deformation history of the ophiolite belt following significant tectonic mixing. Thus the plutons cut melange structures and are structurally in tact, but have high temperature deformation features on trend with the structure of the ophiolite belt. The structural relation between the ophiolite belt and the Jurassic plutons is analogous to the structural relation between the depositional remnants of continental margin rocks and the ophiolite belt. The petrogenesis of each was late-stage syntectonic along the pre-existing structural trends of the ophiolite basement. Foliation surfaces of the Jurassic plutons and the continental margin rocks are designated $S_3$ on Plate V.

The ophiolite belt is in tectonic contact along its eastern margin with an additional assemblage of continental margin rocks. This assemblage consists of quartzite-argillite olistostromes, quartzose to sub-arkosic massive sandstone and flysch, carbonate turbidites and slide blocks and an upper section of shallow marine and silicic volcanic rocks. It is thought to be equivalent to the upper intervals of the Calaveras Complex exposed further north along the foothill metamorphic belt (Saleeby and Goodin, 1977; Schweickert and others, 1977). Late Triassic to early Jurassic fossils have been recovered from the upper part of this assemblage (Christensen, 1963; Jones and Moore, 1973; Saleeby and others, in prep.). Recent mapping and petrographic work suggests that part of this assemblage is a proximal facies of the epiclastic rocks deposited on top of the ophiolite (Saleeby and others, in prep.).

Geochronology

Geochronological work in conjunction with structural and petrologic work has revealed a prolonged history of igneous and metamorphic events along the ophiolite belt. Gabbro of the Kings River ophiolite transition zone contains rare pods and dikelets of diorite and plagiogranite which appear to be autocthonous magmatic differentiates. Zircon separates from these rocks and similar rocks from three widely spaced gabbro-peridotite blocks from the Kaweah serpentinite melange yield a suite of discordant U/Pb ages whose minimum ages range between 205 m.y. and 270 m.y., and whose upper intercept ages cluster around 300 m.y. Zircon discordance is attributed to Cretaceous thermal metamorphism related to emplacement of the batholith. Intercept ages on young zircon populations are difficult to interpret. The tentative interpretation is that initial crystallization could have ranged back to 300 m.y.; however, it could have occurred as late as about 250 m.y. Since the sparce leucocratic rocks are an integral part of the ophiolite belt's igneous assemblage, their initial crystallization age constraints are taken as ophiolite genesis age constraints.

Regional thermal metamorphism related to the Sierra Nevada batholith is of hornblende hornfels facies. A significant exception is where local zones of albite-epidote hornfels facies rocks occur at significant distances from contacts with batholithic rocks. From these zones mafic metamorphic tectonites of the ophiolite have been dated by K/Ar techniques. The minimum age of these tectonites is 190 m.y. Where similar tectonites have been collected from zones of hornblende hornfels facies rocks the K/Ar system has been completely reset to batholith ages. As discussed below, mesoscopic field relations suggest that the true metamorphic age of these tectonites is the same as the ophiolite genesis age.

The timing of ophiolite deformation is bounded on the upper end by voluminous plutons of the batholith which cut $S_1$, $S_2$ and $S_3$ and lack a tectonic fabric. These plutons have yielded numerous early Cretaceous concordant zircon ages. The intact plutonic rocks which contain $S_3$ but cut melange structures have yielded concordant middle and late Jurassic zircon ages. Thus tectonic mixing of the ophiolite belt ceased by middle Jurassic time and subsequent deformation of the ophiolite belt and Jurassic plutons ceased by early Cretaceous time. The tectonics of ophiolite genesis, transport, deformation, melange development and emplacement are the subjects of the remainder of this paper.

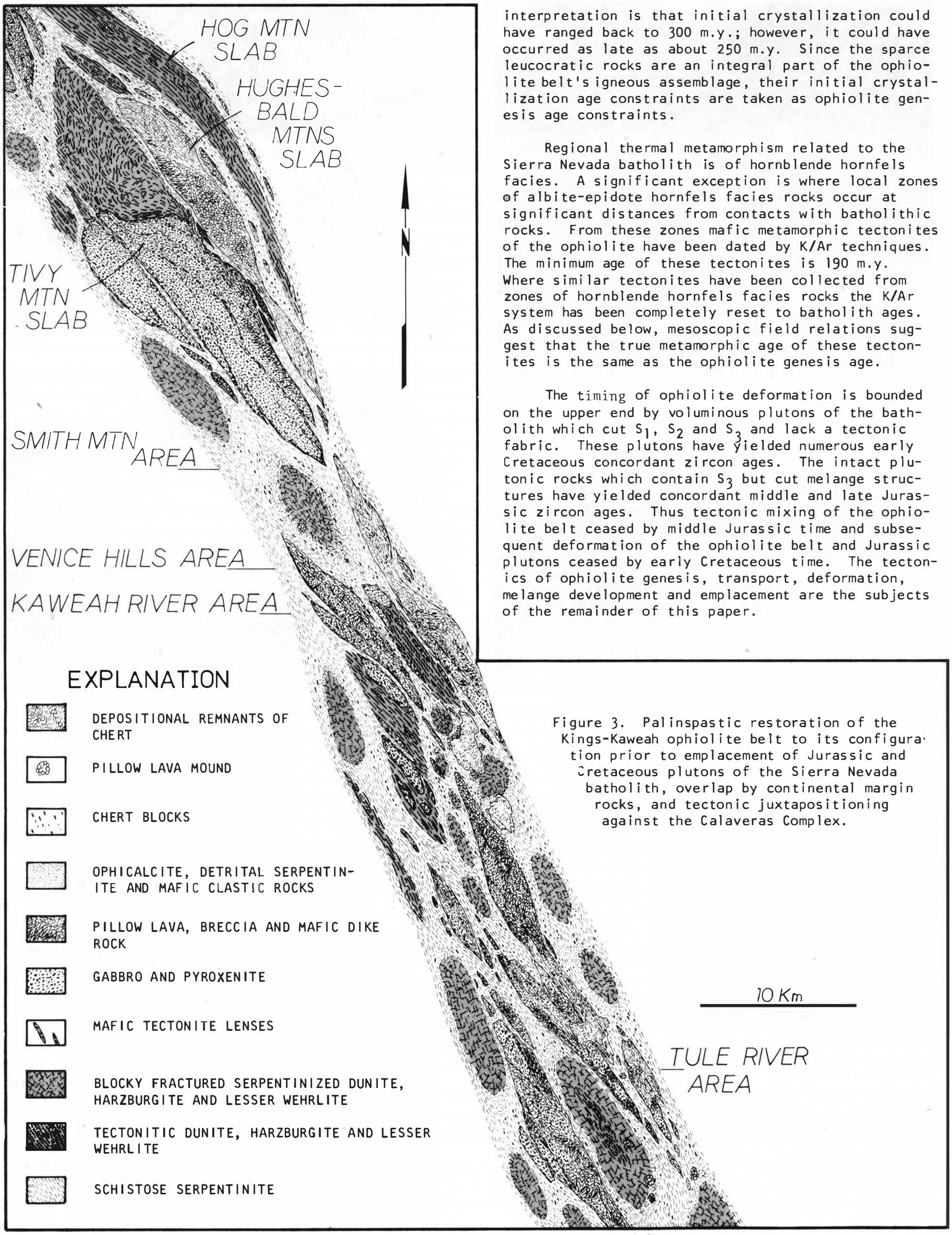
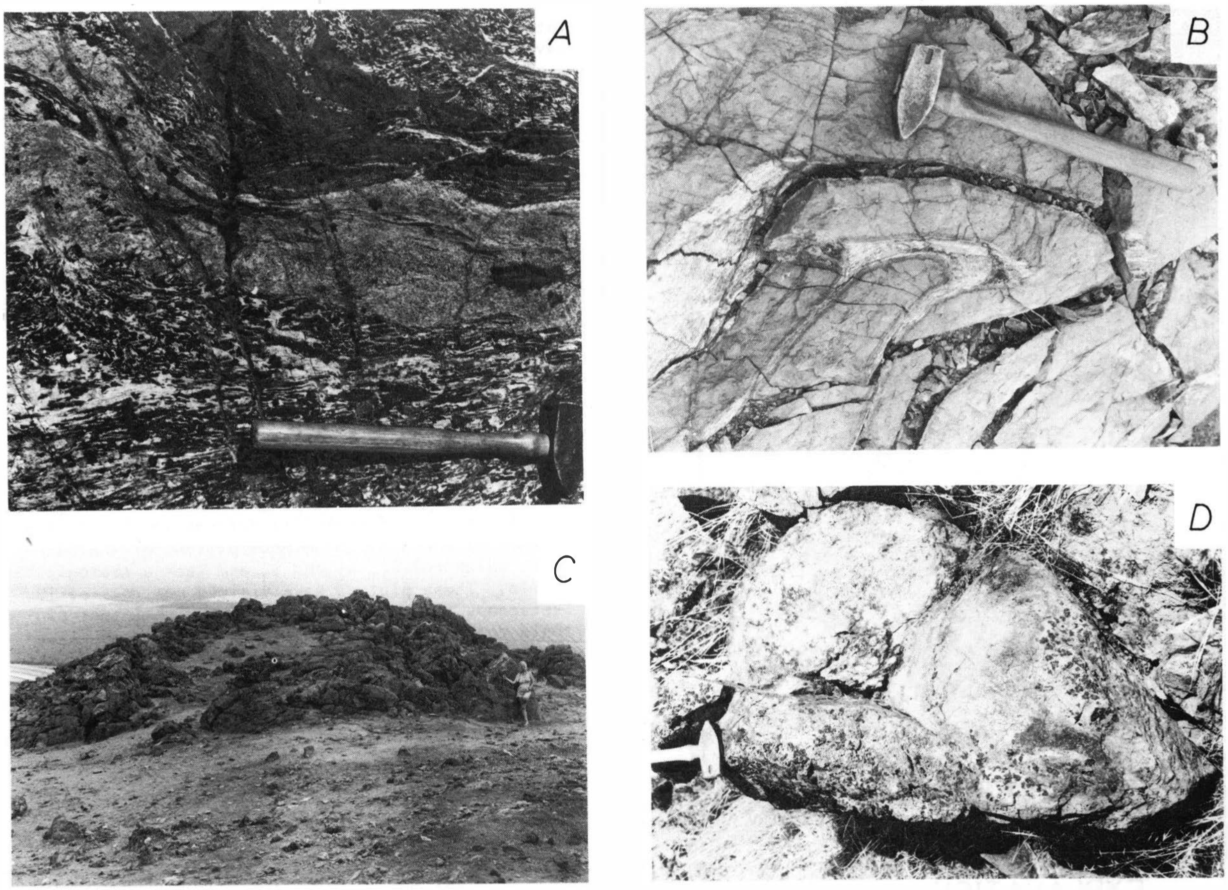



Figure 3. Palinspastic restoration of the Kings-Kaweah ophiolite belt to its configuration prior to emplacement of Jurassic and Cretaceous plutons of the Sierra Nevada batholith, overlap by continental margin rocks, and tectonic juxtapositioning against the Calaveras Complex.




THE FRACTURE ZONE MODEL

A palinspastic restoration of the ophiolite belt to its configuration prior to emplacement of Jurassic to Cretaceous plutons and deposition of the continental margin assemblage is shown in Figure 3. The reconstruction shows the ophiolite belt as a tectonic megabreccia with a penetrative vertical planar fabric. This configuration is significantly different from many other ophiolites which occur as moderately dipping sheets (Moores, 1969; Coleman, 1971; Davies, 1971; Dewey and Bird, 1971; Moores and Vine, 1971; Church, 1972; Gealey, 1977). For this reason an obduction or overthrust model of emplacement is not adopted for the Kings-Kaweah ophiolite belt. A continental margin subduction model of emplacement does not seem applicable either. The tectonic melange of the Kings-Kaweah ophiolite belt is oceanic in origin. Continental margin rocks were deposited across ophiolite melange late in its deformational history subsequent to significant tectonic mixing. In addition, the sedimentary record of the ophiolite belt records transport from the oceanic regime into a continental margin regime which was characterized by non-volcanic hemi-pelagic and terrigenous sedimentation. A volcanic arc was not approached by the sea floor spreading transport of the ophiolite as would be the case in a subduction emplacement model. Finally, metamorphic tectonites of the ophiolite belt are greenschist to amphibolite facies. Blueschist and eclogite facies rocks, which are generally considered characteristic of subduction complexes, are not present along the Kings-Kaweah belt.

An alternative to an obduction or subduction model of ophiolite deformation and emplacement is developed below. Specific structural, petrologic and stratigraphic relations along the ophiolite belt are used in conjunction with recent discoveries in marine tectonics to develop an oceanic fracture zone model. The role of some of the relationships used to develop the model is summarized in Figure 5. Relationships between petrogenesis and deformation of the ophiolite are of primary interest. Critical relationships in igneous, pre-batholith metamorphic, and oceanic sedimentary rocks are covered respectively. Emphasis is placed on the fact that deformation and disruption of the ophiolite was oceanic and progressive, and furthermore, ophiolite genesis was syntectonic. Specific relationships discussed below may have noteworthy alternative interpretations. However, in each instance the fracture zone interpretation appears to be as good or better than alternative interpretations. Furthermore, when all of the relationships are considered together, the fracture zone model seems to be the only model that cannot be dismissed. For sake of brevity the alternative interpretations will not be given equal treatment.

<u>Igneous Deformation</u>

Temporal relationships between the igneous generation of the ophiolite belt and commencement of its deformation history support the fracture zone model. Development of $S_1$ commenced during the igneous generation of the ophiolite at the oceanic spreading

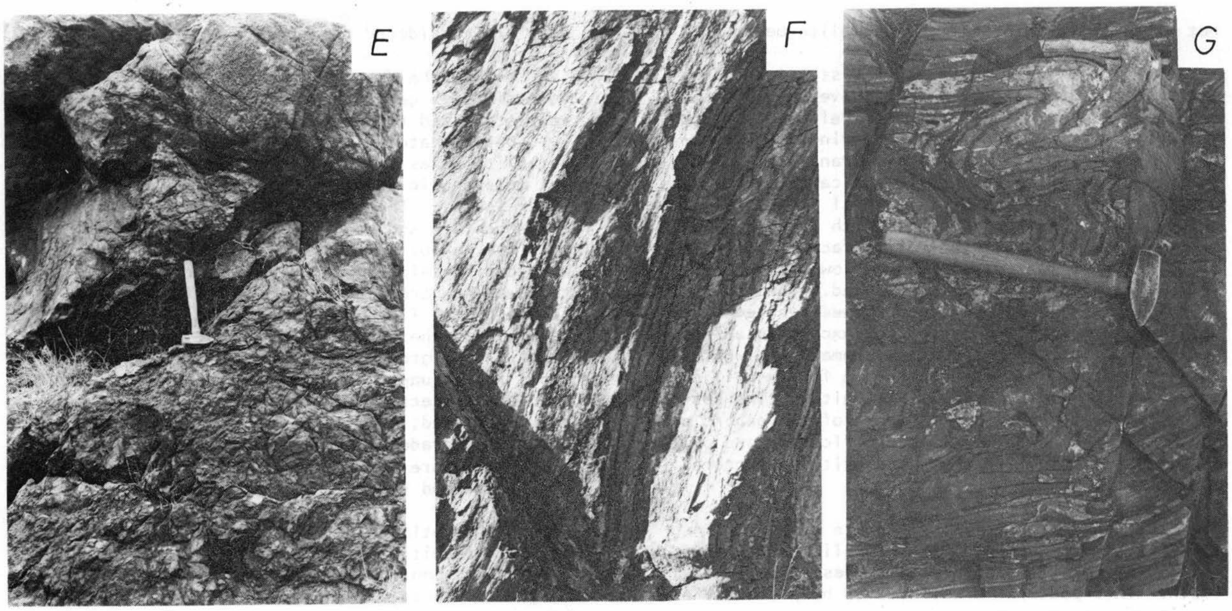



Figure 4. Photographs of some important features along the Kings-Kaweah ophiolite belt. A: Protoclastic deformation in diorite and basalt dikes cutting clinopyroxene gabbro. B: Deformed hydrothermal veins in harzburgite tectonite; veins are flattened into $S_1$ which is tightly folded around vertical axis with homoaxial open refold. C: Blocky fracturing in serpentinized harzburgite. D: Large clast of ophicalcite composed of smaller ophicalcite clasts (areas rich in dark ultramafic detritus) in micritic matrix; the large clast occurs with other clasts of peridotite and ophicalcite within a micritic matrix. E: Crude bedding in sedimentary serpentinite; upper bed contains up to boulder size clasts, lower bed contains up to cobble size clasts. F: Steep-plunging elongation lineation in harzburgite; $L_1$ is accentuated by transposed hydrothermal veins. G: Soft sediment folding and brecciation in metalliferous radiolarian chert; dark bands are nearly pure oxide minerals; note how disrupted interval is bounded by intact intervals.

center. Intrusive and extrusive pulses overlapped in time with pulses of mylonitization. Some intrusive masses cut $S_1$ sharply along part of their length but are in turn cut or transposed into $S_1$ further along their length. Some merely show chaotic protoclastic-type structures which in most cases converge into $S_1$ of the surrounding rocks (Fig. 4a).

The position and amount of highly differentiated igneous rocks raises an important point. Pillow lavas of the ophiolite belt are basaltic. Keratophyre and quartz keratophyre are apparently absent. The mafic dike and cumulate gabbro zones of the Kings River ophiolite and equivalent melange blocks lack diorite and plagiogranite. This paucity of highly differentiated rocks contrasts with many other ophiolites which contain significant amounts of intermediate to silicic intrusive and extrusive rocks (Moores, 1969; Dewey and Bird, 1971; Bailey and Blake, 1974; Coleman and Peterman, 1975). Diorite and plagiogranite do occur in trace amounts in the Kings River transition zone and in equivalent melange blocks. It thus appears that the only environment suitable for stagnation and extreme differentiation of magma bodies was in the deeper levels of the ophiolite beneath the main plutonic part of the section. The pockets and dikes of magma which stagnated in the transition zone also concentrated magmatic water during differentiation. This is shown by the presence of primary brown hornblende and by hydrothermal aureoles and veins that formed in the ultramafic host rock. In the hydrothermal zones dunite, harzburgite and wehrlite have been altered to various combinations of serpentine, Cr-chlorite and talc (Fig. 4b). An important feature of these alteration zones is that they also show developmental pulses which overlapped in time with mylonitization pulses in that the zones cut and are cut or transposed by $S_1$ to various degrees.

Structural analysis of the $S_1$ tectonites reveals high amounts of flattening and constrictional strain. In addition, persistent pulses of translational movements with folding and rotation about steep axes accompanied the flattening and constrictional strain. It is difficult to envision such complex tectonites forming at a normal oceanic spreading center. However, tectonites similar to those of the Kings-Kaweah ophiolite belt have been recovered from transverse fracture zones (Aumento and others, 1971; Bonatti and others, 1971; Melson and others, 1972; Thompson and Melson, 1972; Bonatti and Honnorez, 1976; Fox and others, 1976). It is proposed that the $S_1$ tectonites of the Kings-Kaweah ophiolite belt developed along a transverse fracture zone. The $S_1$ deformation began at the intersection of the fracture zone with the ridge axis and continued for some unknown distance off the ridge axis. Thus, plutonic masses and their contact metamorphic derivatives were protoclastically deformed by $S_1$ and subsequently folded, rotated and redeformed with development of later-stage $S_1$. It is important to emphasize that the penetrative tectonite fabric that is present throughout the harzburgite is also present in dunite, wehrlite, pyroxenite, gabbro, diabase, diorite and plagiogranite. A distinctive tectonite-cumulate contact or contact zone, as would be expected with a normal ridge derived ophiolite, does not exist in the Kings-Kaweah ophiolite belt.

Development of $S_1$ varies with stratigraphic depth in the reconstructed Kings River ophiolite section (Fig. 2). This variation reflects a change in material behavior with depth during $S_1$ development. The harzburgite and lower transition zones behaved by penetrative ductile and cataclastic flow. Notable exceptions to this are small isolated mafic bodies in the harzburgite zone which syntectonically recrystallized in the amphibolite facies and intrusive masses in the transition and lower gabbro zones which were protoclastically deformed. The upper transition and lowermost gabbro zones behaved similar to rocks lower in the section except in a less penetrative fashion. Thus local domains in which igneous textures and structures are fairly-well preserved occur within these $S_1$ tectonites. The tectonites extend through the main part of the gabbro zone as ductile fault zones. In the mafic dike and pillow basalt zones localized shear and brittle fracture zones occur.

The general deformation pattern displayed in the reconstructed Kings River ophiolite section is increasing ductility and pervasiveness with stratal depth. This pattern is believed to be primarily a result of a steep ocean ridge thermal gradient with higher temperatures favoring greater ductility and pervasiveness of deformation. This deformation pattern is believed to have been masked by intense protoclasis at the intersection of the fracture zone axis with the spreading axis. Mafic melange units commonly contain blocks with extremely complex internal structures in which chaotic mixtures of pillow lava, mafic dike rock, gabbro and mafic metamorphic tectonites have contradictory relationships with $S_1$. The chaotic mafic melange blocks are interpreted as remnants of the intersection zone. The spatial relationships envisioned between the melange units which contain the chaotic mafic blocks and the large slabs which fit into a more conventional ophiolite stratigraphy will be discussed below.

## Metamorphic Tectonites

Recent studies of oceanic ridges and fracture zones have shown that these zones are characterized by a distinctive steep vertical metamorphic gradient which passes through zeolite, greenschist and amphibolite facies (Miyashiro and others, 1971; Miyashiro, 1972; Spooner and Fyfe, 1973; Fox and others, 1976). This vertically compressed facies series apparently results from a steep ocean ridge thermal gradient which is related to heat liberated during ocean floor genesis. The effects of a steep ocean ridge thermal gradient are evident in the metamorphic grade of mafic tectonites present along the ophiolite belt (Fig. 2). Data which pertain to this subject comes from within the Tivy Mountain slab and from zones along the serpentinite melange where contact metamorphism by the batholith is at its lowest grade. As stated earlier K/Ar data on the mafic metamorphic tectonites reveals a minimum metamorphic age of 190 m.y. Contact metamorphism by the batholith has severely altered both U/Pb and K/Ar systems of the ophiolite belt, so the true metamorphic age of the tectonites is probably significantly greater. Since protoclastic deformation of diorite-plagiogranite dikes and metamorphic recrystallization of the mafic tectonites are both $S_1$ features, the true metamorphic age of the tectonites is probably close to the igneous age of the diorite-plagiogranite dikes. Thus metamorphic heat and ophiolite genesis heat are considered the same.

Within the deeper stratal levels of the Tivy Mountain slab gabbroic masses were syntectonically recrystallized to amphibolites during development of $S_1$. Unfortunately, contact metamorphism by the batholith makes it impossible to resolve the original metamorphic grade of the Tivy Mountain slab's upper levels. The same problem exists with the Bald Mountain and Hughes Mountain slabs which contain the ophiolite's uppermost stratal levels. What can be said is that metamorphic recrystallization was nowhere near as pervasive in the upper levels of the Tivy Mountain slab as in its lower levels, and that ductile, cataclastic and protoclastic flow greatly predominated as deformation modes in its upper levels. Where the protoliths of amphibolite tectonite blocks in serpentinite melange can be deduced, they are usually gabbro. In contrast, low grade amphibolite and greenschist tectonite blocks are most commonly derivatives of mafic hypabyssal and volcanic rock.

The relationships outlined above are interpreted as a result of a steep ocean ridge thermal gradient which controlled metamorphic mineral assemblages developed along the fracture zone where metamorphic recrystallization was the preferred mode of deformation. Pervasive amphibolite facies metamorphism is present between stratal depths of 7 and 11 km in the reconstructed Kings River section. With a temperature range of about 450°C to 650°C for the amphibolite facies (Turner, 1968), this depth-temperature relation corresponds with calculated ocean ridge geotherms (Oxburgh and Turcotte, 1968; Sclater and Franchetean, 1970). The lower grade conditions which existed higher in the section are manifested by localized zones of syntectonic metamorphic recrystallization now preserved only within mafic melange blocks. This localization of metamorphic tectonites at higher stratal levels is thought to be a result of three variables which worked together to produce them: 1) zones of concentrated deformation; 2) a rapidly declining high thermal gradient; and 3) migration of water. As discussed below, the zones of concentrated deformation are thought to have widened with time, and as a result the influx of water into the deforming ocean floor increased with time. However, the rapidly declining thermal gradient put tight time-space constraints on the interval over which metamorphic recrystallization could operate as a significant deformation mode at upper crustal levels.

Progressive serpentinization of the ophiolite belt's ultramafic rocks is thought to have been an important fracture zone process. Serpentinization is known to be an important process along modern oceanic fracture zones (Bonatti and others, 1971; Melson and Thompson, 1971; Bonatti, 1976; Bonatti and Honnarez, 1976). Serpentinization of the Kings-Kaweah belt began with transition zone hydrothermal metamorphism during ophiolite genesis and initial deformation. As stated earlier the hydrothermal zones cut and are cut by or transposed into $S_1$ to various degrees. The hydrothermal serpentinites do not appear to be directly related to $S_2$-bearing schistose serpentinites. However, $S_2$ serpentinization is also thought to have overlapped in time with development of $S_1$. This is suggested by the gradational relations between $S_1$ and $S_2$. $S_1$ represents the initial deformation and disruption of the newly created ocean floor. As stated earlier $S_1$ development was progressive. As $S_1$ developed migration of ocean water into the deforming ocean floors deeper stratal levels was facilitated. As water migrated into the ultramafic rocks syntectonic serpentine growth progressively replaced ductile and cataclastic flow of olivine and pyroxene. Slabs and blocks of $S_1$-bearing peridotites are the incompletely digested remnants of the young ocean floor's ultramafic zones. It is important to note that steep plunging folds which are so common in the $S_1$-bearing slabs also occur locally in $S_2$ of the ultramafic slabs and the melange matrix.

Progressive serpentinization is believed to have led to greater mobility in the young ocean floor. As $S_2$ domains developed differential tectonic movements were preferentially concentrated along them. This accelerated both tectonic mixing and further serpentinization which led to serpentinite melange formation. As will be discussed below protrusions and surficial debris flows of ultramafic rock appear to have played an integral part in this stage of the ophiolites disruption history.

## Syntectonic Petrogenesis

Structural analysis of the ophiolite belt's igneous and pre-batholith metamorphic rocks indicates a syntectonic petrogenesis of the ophiolite belt's crustal segments which can be best explained with a fracture zone model. Progressive deformation along the fracture zone during transport away from the spreading axis led to the formation of ocean floor melange. The fact that the ophiolite belt's melange is oceanic in origin is best displayed in the record of oceanic sedimentation. As with the ophiolite belt's igneous and metamorphic rocks, petrogenesis of its sedimentary rocks was syntectonic. Thus the earliest formed sedimentary rocks are thoroughly mixed into serpentinite melange, with later deposits being mixed to a lesser extent. In the following discussion the oceanic sedimentation history of the ophiolite belt is treated in two sections, clastic and biogenic. It must be emphasized, however, that these sedimentation modes operated simultaneously.

## Clastic Sedimentation

Sedimentary breccia and coarse angular sandstone composed of basalt, diabase, gabbro, and rare chert and amphibolite detritus occurs as melange blocks in several localities along the ophiolite belt. Relict bedding is preserved in some blocks. Sedimentary fabrics suggest both talus slope accumulation and debris flow deposition modes. In a couple of blocks deformation makes it impossible to decipher if the breccia is a deformed sedimentary rock or if it originated in a fault zone. Angular clast fault breccias along with rare sedimentary breccias have also been observed in the pillow section of the Bald Mountain slab. Since the sedimentary breccias occur most commonly as melange blocks the deeper stratal levels of the ophiolite were at least locally exposed and eroded prior to melange mixing. Deep level exposures of the ocean floor are only known to occur along fault scarps of transverse fracture zones (Bonatti and others, 1971; Melson and Thompson, 1971; Melson and others, 1972; Thompson and Melson, 1972; Fox and others, 1976; Bonatti and Honnorez, 1976). The mafic sedimentary breccias are interpreted as having been shed from fault scarps formed during the early stages of ophiolite disruption. The breccias were subsequently engulfed into serpentinite melange as the fracture zone evolved to a more chaotic state.

Ultramafic detrital rocks also occur along the ophiolite belt. These consist of detrital serpentinites and ophicalcites. Nearly identical rocks have been recovered from modern fracture zones (Bonatti and others, 1973, 1974). Rarely fine detrital serpentinite will occur as sedimentary matrix for mafic clast breccias, and occasionally basalt and gabbro clasts occur in ophicalcite. The ultrmafic breccias have complex developmental histories which are directly related to deeper level tectonics and also involve abundant surficial reworking.

A significant number of ultramafic melange blocks have structural features which differ significantly from the $S_1$-$S_2$ relationships discussed earlier. Superimposed over the peridotite foliation ($S_1$) is a rounded blocky fracture system with arcuate schistose serpentinite zones woven through the peridotite autoclasts (Fig. 4c). These features grade into several different features. In some instances the schistose serpentinite zones become more pervasive, less accurate and ultimately converge into $S_2$ of the melange matrix leaving small clasts of serpentinized peridotite dispersed in matrix adjacent to the parent block. In other instances, the serpentinite becomes less or non-schistose with the clasts dispersed through it in a chaotic fashion. In a significant number of instances the ultramafic melange blocks go through similar gradations as mentioned immediately above except calcite and dolomite occur in different amounts through the sequence. First the carbonate occurs interstitial to ultramafic fragments and then it progressively increases in concentration until the ultramafic material is dispersed in a carbonate matrix (Fig. 4d).

The brecciation sequence outlined above is interpreted as having two intimately associated stages. The first stage is autobrecciation as the ultramafic material moved up diapirically into the fracture zone. The second stage is dispersal of the brecciated ultramafic rock as debris flows and turbidities upon surfacing of the ultramafic protrusion. In many instances it is impossible to distinguish between protrusive breccias and sedimentary breccias. Breccias interpreted as definitely protrusive are parts of semi-intact peridotite blocks. Breccias interpreted as definitely sedimentary contain sedimentary structures and clasts or interbeds of chert (Fig. 4e). A similar intimate relationship between protrusion and sedimentary breccias can be observed in ultramafic flows of the California Coast Ranges (Eckel and Myers, 1946; Dickinson, 1966; Cowan and Mansfield, 1970; Lockwood, 1972), and are apparent in modern fracture zones (Bonatti and others, 1974; Bonatti and Honnorez, 1976).

Vertical protrusion of ultramafic rock is known to be an important process along modern fracture zones (Melson and others, 1967, 1972; Thompson and Melson, 1972; Bonatti and Honnorez, 1976; Fox and others, 1976). The steeply plunging elongation lineation ($L_1$) within $S_1$ indicates a dominant component of upper mantle - lower crustal vertical flow during and immediately following ophiolite genesis (Fig. 4f). In a fracture zone environment the accent of the hot ultramafic rock would not be confined by lateral spreading about the ridge axis. Thus $S_1$-$L_1$ development not only reflects wrench tectonics, but also vertical protrusion tectonics. Protrusion was probably accelerated as water migrated into the deforming ocean floor and serpentinization of the hot ultramafic rock commenced. The blocky fracture pattern discussed above suggests a volume increase during serpentinization. Similar patterns are present around the Burro Mountain ultramafic body of the California Coast Ranges where expansion has been documented (Coleman and Keith, 1971). In addition, serpentinite under high temperature conditions exists in a thermally weakened state (Raleigh and Patterson, 1965). Thus, the upward ductile and cataclastic flow of peridotite is envisioned as having accelerated due to the expansion and weakening of serpentinization. As vertical flow and serpentinization progressed, the protrusive rock continued to weaken, increasing its mobility. The importance of strain history with respect to progressive weakening in these type of bodies has been demonstrated by Cowan and Mansfield (1970). Surfacing of the fracture zone protrusions resulted in monolithologic sedimentary breccias of ultramafic rock.

Talus piles of protruded ultramafic rock are believed to be the main environment of ophicalcite formation. Interaction with percolating ocean water and/or hydrothermal fluids is believed to have been the main cause of ophicalcite formation. A biogenic origin is not considered important here since biogenic limestones are rare along the entire ophiolite belt. A subaerial pedogenic origin (Folk and McBride, 1976) is not considered since radiolarion chert is locally interbedded with and overlies ophicalcite.

The detrital ultramafic rocks show a complex sedimentation history with abundant reworking. Clasts of ophicalcite containing abundant ultramafic detritus occur in later generation ophicalcites and in detrital serpentinites. In addition, interbeds of ophicalcite occur within detrital serpentinite and interbeds of both ophicalcite and detrital serpentinite occur within radiolarian chert. The ophicalcite interbeds appear to have been accumulations of carbonate mud with pebble to sand size ultramafic fragments. In several instances these "diamictites" compose the matrix of chert-clast breccia. As will be discussed in the section on biogenic sedimentation, soft sediment deformation and reworking was an important process along the fracture zone. The protruded accumulations of ultramafic detritus and related ophicalcites were probably disrupted and reworked by further protrusion and wrench tectonics. Local disruption may have also occurred when small mounds of pillow lava were built on the detrital ultramafic rocks.

The detrital ultramafic rocks were readily incorporated into serpentinite melange as both blocks and matrix. In numerous instances the friable detrital serpentinites can be observed in intermediate to advanced stages of disintegration into melange matrix by development of $S_2$. The fact that the ultramafic clastic sedimentary rocks occur as depositional remnants above melange, as melange blocks and as a local protolith of the melange matrix indicates their syntectonic genesis.

## Biogenic Sedimentation

Deposition, soft sediment deformation, lithification and hard rock deformation of radiolarian chert proceeded throughout the disruption history of the ophiolite belt. The earliest formed cherts are mixed as tectonic blocks throughout serpentinite melange, and occur locally within pillow lava slabs and melange blocks. Later-stage cherts rest as highly deformed depositional remnants above detrital ultramafic rocks, late-stage pillow lava mounds and serpentinite melange. Chert melange blocks commonly have stratigraphic thicknesses of about 20 m. Thicknesses between 100 and 200 m occur in the depositional remnants, but these are gross thicknesses due to intense deformation. Since chert deposition was syntectonic a coherent chert section probably never existed. The thicknesses of both the melange blocks and depositional remnants suggest that at least 200 m of chert was deposited on the ophiolite belt prior to deposition of the continental margin assemblage. However, the earlier-formed chert intervals were tectonically mixed into melange prior to and during deposition of the later-formed intervals. Contact metamorphic recrystallization of radiolaria tests prohibits paleontological dating of the cherts.

A significant relationship exists between the composition of the cherts and their structural setting. Cherts occurring as tectonic blocks throughout the melange commonly contain black to dark purple interbeds and disseminations of oxide minerals. These impurities are primarily iron oxide with trace manganese oxide. The metalliferous cherts are notably lacking in argillaceous or volcanic impurities. Cherts occuring as highly deformed depositional remnants above serpentinite melange locally contain thin interbeds and disseminations of argillaceous material. The argillaceous cherts lack significant amounts of oxide minerals, and lack volcanic impurities. Volcanic impurities occur only rarely in cherts that occur with pillow lava. Cherts lacking any significant impurities occur both as dispersed melange blocks, with or without metalliferous chert, and as bedded intervals in depositional remnants which contain the argillaceous cherts.

The relationships presented above are interpreted to be a result of: 1) deposition of radiolarian ooze commencing during ophiolite genesis and continuing throughout ophiolite disruption along the fracture zone; 2) early to middle-stage deposition of basal metalliferous sediments from hydrothermal solutions (Bostrom and Peterson, 1969; Bostrom and others, 1976) emanating from depth at the spreading axis and possibly along the fracture zone for some distance off the spreading axis; 3) later-stage sporadic influx of fine terrigenous material shed from a distant source that was being approached by sea floor spreading transport of the fracture zone complex. Following deposition of the argillaceous cherts the next rocks that appear in the sedimentary record are chert-argillite olistostromes which contain shallow water limestone olistoliths, and interbeds of both continent derived sandstone and argillaceous chert. Thus the ophiolite was approaching a landmass that was not contributing volcanic detritus to the sedimentary record. This important relationship will be discussed further in conjunction with continental margin tectonics.

Stratigraphic settings of the various cherts indicate progressive disruption of the oceanic basement during biogenic sedimentation. The early to middle-stage cherts occur in association with pillow lavas or as dispersed blocks in serpentinite melange. These cherts appear to have had two depositional settings: 1) mafic oceanic basement as shown by their presence in both pillow lava slabs and in pillow lava-bearing melange units; and 2) protruded ultramafic basement as shown by their presence in peridotite melange units where they are associated with ophicalcite and detrital serpentinite. The later-stage cherts were deposited on basement consisting of serpentinite melange. detrital ultramafic rocks and mounds of late-stage pillow lava.

Structural features of the chert assemblage reflect continuous deformational activity along the fracture zone. Structures that are best interpreted as soft sediment in origin occur both in tectonic blocks and in the depositional remnants. These consist of stratigraphic intervals of chert-cemented chert -clast breccias and associated chaotic folds (Fig. 4h). The chaotic intervals are bounded by bedded intervals both of which are commonly cut by hard rock tectonic structures. In several instances chert clast debris flow deposits occur with an ophicalcite matrix. The common occurence of soft sediment breccias and folds is taken as an indicator of an unstable depositional environment. Since the younger cherts overlie serpentinite melange, which contains blocks of older cherts, the instability of the pelagic depositional environment is shown to have been persistent and tectonic in origin.

Early to middle-stage cherts, which occur primarily as melange blocks, were lithified in most cases prior to melange mixing. This is shown by the presence of brittle shear and tension fractures which are the surfaces along which the melange blocks initially broke apart. Ductile deformation features such as pinching and swelling or streaking-out of bedding in many cases cannot be distinguished from soft sediment features. In some instances chert blocks can be shown to have been ductily deformed along with $S_2$ deformation of the matrix. Thus early to middle stage cherts underwent localized chaotic folding and brecciation prior to lithification, and following lithification they underwent brittle fragmentation to form melange blocks some of which underwent subsequent ductile deformation along with the melange matrix.

Late-stage cherts which occur as depositional remnants above melange are in some ways structurally more complex than the earlier-stage cherts. This relationship is interpreted as a result of deposition and lithification of the earlier cherts occuring on semi-intact basement slabs undergoing localized deformation with deposition of the later cherts occuring on serpentinite melange basement undergoing penetrative deofrmation. Thus, once lithified the earlier cherts were able to escape much of the deformation that the later cherts experienced in a soft sediment to semi-lithified state while sitting on active melange. Chaotic folds and breccias of probable soft sediment origin are locally important in the late-stage cherts. Another important feature of these cherts, which appears to have been inherited from the soft sediment to semi-lithified state, is the presence of large massive domains with local clasts and rootless folds. The massive domains grade into or sharply abut against bedded domains. The massive domains are interpreted as ponds of reworked radiolaria ooze which slid across and ripped up beds of compacted ooze. In addition some chert beds are graded with respect to radiolaria test size which suggests reworking of radiolaria ooze by turbidity current mechanisms. Both bedded and chaotic domains of the depositional remnants are commonly cut by a spaced cleavage which locally grades into a penetrative cleavage. Where bedding is preserved the cleavage is occasionally axial planar to steep-plunging folds. In several instances the cleavage cuts across limbs of chaotic soft sediment folds. The cleavage is coplanar to $S_2$ of the underlying melange matrix. The outcrop pattern of the depositional remnants is highly suggestive of an infold relationship with the underlying melange (Plate V).

It must be emphasized that a distinct line cannot be drawn between early, middle and late-stage cherts. Early and late-stage cherts can be distinguished by the structure-composition relations outlined above. Middle-stage cherts appear to represent a gradition both in structure and composition between early and late-stage cherts. The fact that soft sediment and progressive hard rock deformation occur throughout the chert assemblage coupled with the compositional variation outlined above indicates that the ocean floor was progressively disrupted in an oceanic environment en-route to the continental environment. A survey of present day marine tectonic environments reveals that a large fracture zone will serve as the only suitable analogue.

## Fracture Zone Tectonics and the Structure of Oceanic Crust

The syntectonic history of igneous, metamorphic and sedimentary petrogenesis displayed in the Kings-Kaweah ophiolite belt is diagramatically summarized using a fracture zone model in Figure 5. Mafic magma and harzburgite residue are shown ascending beneath the ridge axis in accord with sea floor spreading theory (Green and Ringwood, 1967; Kay and others, 1970; Green, 1970, 1971; Dewey and Bird, 1971). The model shows anomalous oceanic crust being created at the intersection of the spreading center and the axis of the fracture zone. The anomalous crust is shown as having two main components. 1) Mafic pillow lava, hypabyssal and deeper plutonic rock characterized by chaotic protoclastic deformation and mixing. These rocks now exist as complex melange blocks and melange units. 2) Ultramafic protrusions which ascended into the fracture zone from the mantle while hot, and which upon reaching crustal levels underwent serpentinization and subsequently surfaced, shedding ultramafic detritus. The large proportion of ultramafic rock along the ophiolite belt, the common occurence of peridotite-chert melange units, and the monolithologic nature of most detrital ultramafic rocks indicate that a significant amount of anomalous oceanic crust was created by ultramafic protrusion. It follows that a significant amount of spreading probably occurred along the axis of the fracture zone due to protrusion tectonics. This has been suggested for large modern fracture zones (Van Andel and others, 1969; Thompson and Melson, 1972; Bonatti and Honnorez, 1976). In addition, protrusion tectonics probably played a significant role in serpentinite melange formation. The anomalous oceanic crust was generated in a semi-tectonically mixed state with chaotically deformed mafic crustal blocks interwoven with ultramafic protrusions and their sedimentary derivatives. This configuration was subjected to prolonged wrench tectonics which resulted in serpentinite melange. The oceanic sedimentation record above the serpentinite melange indicates that the melange too represents anomalous oceanic crust. It is suggested that serpentinite melange is a significant anomalous crustal component in present-day fracture zones with large ridge offsets.

The model also shows a more conventional-type of oceanic crust exposed by faulting along the margin of the fracture zone. This is presently represented by the Kings River ophiolite which has the remnants of normal ocean floor stratigraphy (Fig. 2). The relationship portrayed above is best displayed today along the Vema Fracture Zone of the equatorial Atlantic where normal oceanic crust is apparently exposed along the fracture zone's northern wall while disrupted and protruded crust is exposed along its axes and southern wall (Bonatti and Honnnorez, 1976).

The effects of fracture zone tectonics (protrusive and wrench) are evident in the Kings River ophiolite; however, its stratal succession indicates that normal ridge crustal generation processes were permitted to operate. This pattern is intuitively pleasing since anomalous fracture zone crust must at some interval grade laterally into normal ridge created crust. The Kings River ophiolite is interpreted as having originated in such a gradation interval.

The complexity of the ophiolite belts $S_1$ tectonites can be conceived of as a result of both wrench and protrusion tectonics. As the hot upper mantle and lower crust ascended into the fracture zone it was polydeformed. The deformations consist of: 1) vertical extension by upward flow; 2) flattening in the plane of the fracture zone by forcing its crustal levels apart during protrusion; 3) shear and translation in the plane of the fracture zone by wrench faulting; 4) rotation and folding about steep axes in the plane of the fracture zone due to wrench movements; and 5) a probable complex system of dip-slip faults, antithetic strike-slip faults and shallow plunging folds that are ubiquitous in continental wrench zones (Moody and Hill, 1956; Lillie, 1964; Reed, 1964; Dickinson, 1966; Harding, 1973, 1974; Wilcox and others, 1973; Sylvester and Smith, 1976).

The fact that fracture zone deformation of the ophiolite belt was progressive is well-displayed by the contradictory cross-cutting relations between steep to shallow plunging folds, different stage $S_1$ surfaces, and different stage igneous pulses. The inclusion of chert blocks in serpentinite melange and the deposition of later-stage cherts across the melange with their subsequent deformation demonstrates the longevity of progressive deformation. As outlined in the next section, this deformation continuum is believed to have extended from the oceanic realm into the ancient continental margin as the ophiolite belt was transported and emplaced into its present position.

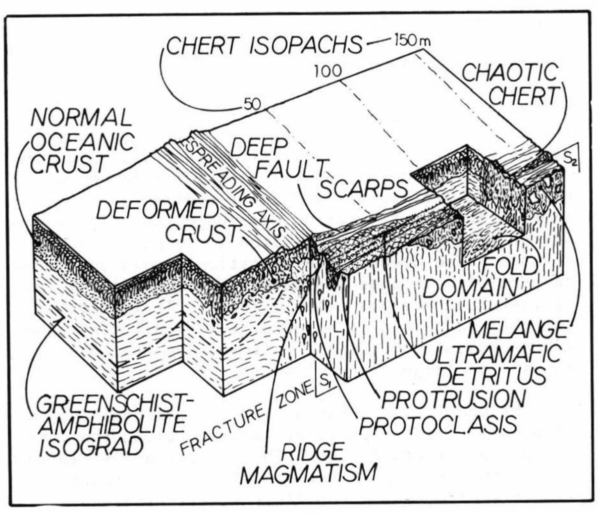



Figure 5. Schematic block diagram showing how critical features of the Kings-Kaweah ophiolite belt fit into an oceanic fracture zone tectonic model. Regional relationships suggest a north-south trend for the fracture zone making this view of the block diagram towards the northwest.

## THE ANCIENT CONTINENTAL MARGIN

The fracture zone model for the origin, deformation and sea-floor spreading transport of the Kings-Kaweah ophiolite belt has been developed above using data solely from the ophiolite belt itself. The fracture zone history of the ophiolite belt apparently began in the latest Paleozoic and probably extended into the Triassic. Recent workers have cited regional structural and stratigraphic evidence for early Mesozoic tectonic truncation and transcurrent faulting of the ancient southwest continental margin (Hamilton and Myers, 1966; Jones and others, 1972; Jones and Moore, 1973; Burchfiel and Davis, 1972; Silver and Anderson, 1974; Schweickert, 1976). This tectonic regime is believed to have been directly related to fracture zone tectonics as discussed above. Thus the fracture zone is believed to have extended from well within the oceanic realm into the ancient continental margin. Outstanding modern examples of complex wrench systems which involve both oceanic and continental domains include the Macquarie Ridge-Alpine Fault system of the southwest Pacific (Griffiths, 1971; Griffiths and Varne, 1972), the San Andreas-Queen Charlotte system which rims western North America (Wilson, 1965) and the Spitsbergen Fracture zone of the Arctic Ocean (Lowell, 1972). Considering the complex histories of these sytems one is forced to conclude that significant complexities that could have involved triple junctions and microplates are probably irresolvable in the ancient system. With this in mind, a simplistic tectonic model is outlined below for the continental margin deformation and emplacement of the Kings-Kaweah ophiolite belt. A fuller treatment of this model is given in Saleeby and others (in prep.) and Saleeby (in prep. b).

### Foothill Suture

The Sierra Nevada foothill metamorphic belt coincides with a tectonic suture in the earth's crust (Fig. 1). Suture is used here to mean a zone of joining. As discussed below the foothill suture joins fossil late Paleozoic to early Mesozoic oceanic lithosphere to older continental lithosphere. In the south the suture is defined by the Kings-Kaweah ophiolite belt. In the north it is defined by the foothill fault system (Clark, 1960; Schweickert and others, 1977). The significance of the foothill suture is shown by several points: 1) all Sierra Nevada ophiolite remnants occur along it; 2) highly deformed and tectonically mixed rocks which occur along it (Clark, 1960, 1964; Morgan, 1973; Duffield and Sharp, 1975; Ehrenberg, 1975; Behrman, 1978; Saleeby, in press a, b) indicate that it was a zone of major translation; 3) the distinct changes in the gross structure of the crust and upper mantle which coincide with the suture (Fig. 1, inset) can be best explained as a result of a fossil contact between oceanic and continental lithosphere; 4) Jurassic and Cretaceous batholithic rocks emplaced into and to the east of the suture have systematic petrochemical variations (Fig. 1, inset) which can be best explained as a result of the batholith having been emplaced across a contact between oceanic and continental lithosphere; and 5) as discussed below highly contrasting lithologic and structural terranes are juxtaposed along it (Fig. 1).

The suture in the south is exposed as a penetratively deformed ophiolite terrane, whereas in the north it's exposed as a fault system in primarily younger epiclastic and arc volcanic rocks. This is believed to be a result of deeper levels of exposure occurring towards the southern end of the metamorphic belt as discussed earlier. Many of the penetrative deformational features present in the metamorphic belts ophiolitic basement rocks pre-date overlap by the younger epiclastic and arc volcanic rocks. Deformational features in the younger rocks and the foothill fault system are the last expressions of deformation along the suture. It is significant that these late-stage deformations followed the older trends established in the ophiolitic basement rocks.

The foothill suture represents the locus of significant tectonic juxtapositioning. Late Paleozoic ophiolite remnants and overlying early Mesozoic epiclastic and arc volcanic rocks are juxtaposed against a complex of Paleozoic to early Mesozoic continental margin rocks which lie east of the suture. These rocks appear to be remnants of the fragmented continental margin. This fragmentation is believed to be linked to fracture zone tectonics of the Kings-Kaweah ophiolite belt, and to emplacement of the ophiolite belt against the continent's edge.

Continental Margin Fragmentation

Paleozoic rocks east of the Sierra Nevada constitute the southern end of a system of paleogeographic belts that can be traced as far north as central Alaska (Churkin, 1974). The paleogeographic belts consist of volcanic arc, marginal basin and shelf terranes (Fig. 1). Through eastern California and Nevada the paleogeographic belts have northeast trends which are exemplified by facies patterns and Paleozoic thrust belts (Fig. 1). The shelf rocks appear to overlie pre-Phanerozoic crystalline basement, whereas marginal basin and volcanic arc rocks were apparently deposited on transitional or oceanic basement.

Remnants of the Paleozoic belts are present in roof pendants of the eastern Sierra Nevada and possibly in the Shoo Fly complex of the northern Sierra Nevada (Speed and Kistler, 1977; J.N. Moore, personal communication, 1977). These exposures mark the western limit of the Paleozoic belts, and thus a zone of pre-batholith tectonic truncation must have passed longitudinally through the Sierra Nevada. Tectonic truncation of the Paleozoic belts is believed to have been a direct result of wrench movements along the foothill suture. The foothill suture is envisaged as a segment of a transform plate juncture which extended from the fracture zone and cut obliquely across the ancient continental margin. Fragments of the Paleozoic belts were displaced by oceanic lithosphere during truncation.

Truncation of the Paleozoic belts resulted in a major change in the structural grain of the southwest continental margin. Northeast structural and stratigraphic trends, which prevailed throughout the Paleozoic, were terminated and replaced by Mesozoic northwest trends (Fig. 1). The northwest trends have persisted through the Cenozoic and are now manifested by the San Andreas fault system.

This change in structural grain is evident within Paleozoic strata adjacent to the truncation zone. In Paleozoic strata exposed immediately east of the Sierra Nevada northwest trending fold axes, cleavages, thrust faults and strike-slip faults of Mesozoic age are superposed over earlier northeast trending structures (Stewart and others, 1966; Burchfiel and others, 1970; Stevens and Olson, 1972; Kelley and Stevens, 1975; Sylvester and Babcock, 1975; Dunne and Gulliver, 1976; J.N. Moore, personal communication, 1977; Saleeby, unpub. data). A similar pattern of superposed structures exists in Paleozoic rocks present in roof pendants of the eastern Sierra Nevada (Kistler, 1966; Brook, 1977; Russel and Nokelberg, 1977). It must be emphasized that the northwest trending structures of the Sierra Nevada do not represent a single deformational event. Instead, deformation along northwest trends occurred continuously, or in numerous pulses, throughout the Mesozoic (Nokelberg and Kistler, 1977; Saleeby, in prep. b).

Roof pendants east of the Kings-Kaweah ophiolite belt record the history of early Mesozoic sedimentation and tectonics along the fragmented edge. These rocks are treated in-depth in Saleeby and others (in prep.). As discussed earlier they consist of continent derived massive sandstone, flysch, olistostromes and an upper section of shallow marine and silicic volcanic rocks. This assemblage was probably deposited on continental crust as shown by isotopic studies on their enclosing batholithic rocks (Kistler and Peterman, 1973, 1975; Doe and Delevaux, 1973; Chen, 1977). However, this assemblage may not be in its original position relative to similar age rocks resting above Paleozoic strata immediately east of the Sierra Nevada (Jones and Moore, 1973; Saleeby and others, in prep). The clastic rocks were reworked from the truncated Paleozoic shelf belt. They were apparently shed as submarine aprons and fans across fragmented continental basement. The basement was probably undergoing longitudinal wrench movements along the new Mesozoic trends during clastic sedimentation. This tectonically active depositional environment is believed to have given rise to the chaotic deposits of this assemblage.

Structural data on Paleozoic and Mesozoic continental margin rocks east of the Kings-Kaweah ophiolite belt suggest that a longitudinal dextral wrench system worked in conjunction with transverse shortening during the early Mesozoic (Saleeby, in prep. b). This pattern is also evident along the Kings-Kaweah ophiolite belt. These structural patterns suggest that the fracture zone complex was transported from the south, and that the continental margin fragments were displaced northward.

Plate tectonic transport of the ophiolite belt and the displaced continental fragments to the north is also implied by regional considerations. 1) Lower Paleozoic rocks of southeastern Alaska constitute part of an anomalous continental fragment which may have been transported by right-slip faulting from the California region (Mongar and Ross, 1971; Jones and others, 1972). 2) Another anomalous terrane of Triassic age, which extends from south-central Alaska through British Columbia, has yielded equatorial paleolatitudes (Jones and others, in press). 3) Plate tectonic reconstructions of the Mesozoic western Pacific yield mainly east-west trending spreading axes with large north-south trending fracture zones (Larson and Chase, 1972; Hilde and others, 1977). In addition, there is known to have been 4,500 km of northward drift of the Pacific ocean floor since the middle Mesozoic (Larson and Chase, 1972). The time intervals for which these plate tectonic relations are applicable post-date the fracture zone history of the ophiolite belt. However, the consistency between these relations, and the structural configuration of the ophiolite belt and the ancient continental margin suggest that all of these tectonic processes are related to the same kinematic regime.

Continental margin rocks lying above the ophiolite belt record its transport history into proximity of North America. The chert-argillite olistostrome complex was acquired at some unknown distance from the continental margin during transport from the South Pacific (Saleeby and others, in prep.). Large submarine sliding covering thousands of square kilometers of ocean floor is a significant modern process adjacent to both stable and mobile continental margins (T.C. Moore and others, 1970; Embley, 1976; D.G. Moore and others, 1976). The source for the ancient olistostrome complex is unknown. The exotic nature of the fauna within limestone olistoliths indicates that the source was not the North American continent. The olistostrome complex and its exotic fauna may have been derived from outboard borderland and/or orogenic terranes which rimmed the western and southern margins of North America in the latest Paleozoic (Saleeby and others, in prep.).

It is significant that the chert-argillite olistostrome complex grades into distal quartoze to subarkosic flysch. Furthermore, this flysch sequence appears to be the distal equivalent of clastic rocks shed directly off the fragmented North American Paleozoic shelf. The extremities of a large submarine fan system derived from the fragmented shelf are envisaged as lapping across the cite of final chert-argillite deposition. The possible spatial and temporal complexities of this relationship are discussed in Saleeby and others (in prep.).

Shortly after clastic sedimentation began subduction tectonics commenced along the Mesozoic trends. This is shown by the remnants of the arc rocks along the ophiolite belt and the early Mesozoic silicic volcanic rocks east of the ophiolite belt. Regional age data on volcanic and plutonic rocks suggest that this transition occurred during the Triassic (Evernden and Kistler, 1970; Crowder and others, 1973; Schweickert, 1976b; Morgan and Stern, 1977; Saleeby and others, in prep.; P.C. Bateman and O.T. Tobisch, oral communication, 1977).

Subduction and Ophiolite Emplacement

Transcurrent (wrench) faulting has recently been cited as an important mechanism for ophiolite emplacement along continental margins (Dewey and Karson, 1976; Brookfield, 1977). In this view initial juxtaposition of oceanic lithosphere against continental lithosphere occurs by wrench faulting, and actual ophiolite emplacement occurs during a change in plate motions which results in a convergent component between the juxtaposed plates. A similar mechanism is envisaged for the Kings-Kaweah ophiolite belt (Fig. 6). It is not unreasonable to assume that the ancient fracture zone complex was tens of kilometers wide during the later stages of its evolution - considering the width of modern fracture zones with large offsets (Thompson and Melson, 1972; Sclater and Fisher, 1974; Bonatti and Honnorez, 1976). A widely accepted corollary to plate tectonic theory is that continental lithosphere cannot be subducted beneath oceanic lithosphere due to their relative densities (McKenzie, 1969). Similar logic is used in deciphering the fate of the fracture zone complex during the onset of subduction. Taking into account that much of the fracture zone complex was serpentinite, and that serpentinite is significantly less dense than continental crust, the consuming break is believed to have formed on the oceanic side of the fracture zone complex. Thus the change in plate motions accreted the fracture zone complex to the "raw edge" of the continental margin. As the transform juncture evolved into an oblique subducting juncture the fracture zone complex was stranded as the subduction zone's hanging wall. Evolution of large fracture zones into subduction zones during changes in plate motions has been postulated for several present day Pacific subduction zones (Uyeda and Miyashiro, 1974; Falvey, 1975; Hilde and others, 1977). At least one of these instances (Tonga-Kermedac) has yielded ophiolite assemblage dredge hauls from its inner-trench walls (Fisher and Engel, 1969).

Following the change in plate motions the accreted fracture zone complex served as frontal arc basement. However, the arc rocks and their ophiolitic basement are not considered to have been in their final position along the foothill suture until the end of the Jurassic when tectonic deformation along the ophiolite belt ceased.

As stated earlier, the arc plutonic and volcanic rocks of the Kings-Kaweah region were syntectonically generated. Studies in the foothill metamorphic belt further north and in roof pendants to the east reveal similar relations (Parkison, 1976; Nokelberg and Kistler, 1977; Behrman, 1978; Saleeby, in prep. b; Saleeby and others, in prep.). In the Kings-Kaweah region the arc plutons were protoclastically deformed while the volcanic sequence was faulted and in some instances penetratively deformed. There were also uplifts and exposures of ophiolite basement which shed olistostromes into the arc sequence. The structural trends of the arc deformation followed pre-existing trends in the ophiolitic basement. It must be emphasized that Triassic and Jurassic arc rocks throughout California represent only small fragments of the original arc terrane. The original position of these fragments relative to one another may not be easily resolved.

Longitudinal wrench disruption and dispersion of active arc and inner trench wall terranes is known to be an important process along the modern circum-Pacific in zones of oblique convergence (Allen, 1962, 1965; Allen and others, 1970; Wilson, 1965; Fitch, 1972; Karig, 1974; Karig and others, 1975, 1977; Brookfield, 1977; Curray and others, in press). A significant northward component in Mesozoic oblique subduction is believed to have been dissipated by intra-arc wrench movements along the foothill suture and within the fragmented edge of the continent. Transverse shortening worked in conjunction with longitudinal wrench movements. This transpressive (after Harland, 1971) tectonic regime is believed to have been facilitated by the pre-weakened state of the arc basement which consisted of the fragmented continental edge and the tectonically accreted fracture zone complex.

## CONCLUSIONS

The Kings-Kaweah ophiolite belt was generated during the latest Paleozoic at a distant east-west trending oceanic spreading center where cut by a major north-northwest trending transverse fracture zone. The fracture zone extended from the oceanic realm into the ancient southwest continental margin where it truncated earlier northeast trending structures and facies patterns.

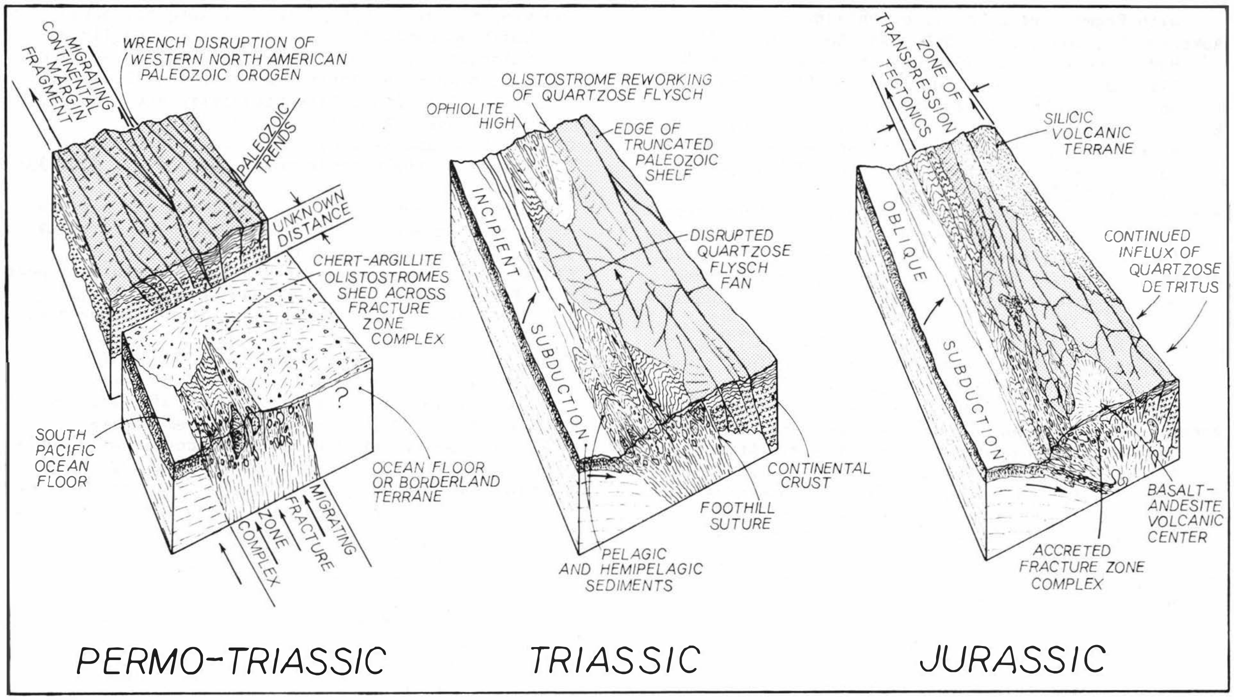



Figure 6. Series of block diagrams which show continental margin emplacement history of Kings-Kaweah ophiolite belt. View is northward along the foothill suture.

Along the axis of the fracture zone anomalous oceanic crust was created. This consisted of proto-clastically deformed mafic igneous rock and protruded ultramafic rock. Away from the axial region of the fracture zone normal oceanic crust was created. Remnants of the normal crust also show the effects of fracture zone deformation however. Metamorphic-tectonites of greenschist and amphibolite facies were created during fracture zone tectonics. The heat which drove the metamorphic reactions was the ambient heat of ophiolite genesis. Protrusion and wrench tectonics worked together to progressively disrupt the newly created ocean floor. Progressive disruption and serpentinization led to the development of serpentinite-matrix melange.

Oceanic sedimentation proceeded throughout the genesis and disruption history of the ophiolite belt. Mafic and ultramafic detrital rocks were shed off of upfaulted and protrusive highs. Radiolarian chert was also deposited during genesis and disruption. The earlier formed oceanic sedimentary rocks were thoroughly mixed into serpentinite melange. Later deposits were mixed to a lesser extent. Several of the latest deposits remain as highly deformed depositional remnants above melange. The latest formed cherts have local interbeds of argillaceous material which record encroachment of the fracture zone complex into the continental margin environment.

As the ophiolite belt was transported northward along the fracture zone into proximity of the continental margin fragments of the continental margin were displaced by wrench movements. Chert-argillite olistostromes with blocks of shallow water late Permian limestone containing exotic fauna were shed across the fracture zone complex enroute to the continental margin. By this time the ophiolite belt had already been chaotically mixed by fracture zone processes. As the ophiolite belt moved into closer proximity of the truncated margin terrigenous sedimentation overwhelmed hemi-pelagic sedimentation.

During the Triassic a significant convergent component had developed along the fracture zone. Disrupted ocean floor of the fracture zone was accreted to the truncated edge of the continent as the hanging wall of an oblique subduction zone. A magmatic arc developed along the fragmented edge of the continent in response to subduction. The arc lapped across the suture between the accreted fracture zone complex and the truncated edge of the continent. As the arc evolved it was deformed and disrupted by both transverse shortening and continued wrench movements. Arc deformation was facilitated by basement mobility. The basement consisted of a wide zone of fragmented ocean floor and continental margin rocks.

The tectonic regime outlined above led directly to the Franciscan regime and the related emplacement of the major part of the Sierra Nevada batholith. The Cretaceous batholith further disrupted and metamorphosed the Kings-Kaweah ophiolite belt leaving it as a healed tectonic suture in the earth's crust.

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# IGNEOUS HISTORY OF THE POINT SAL OPHIOLITE, SOUTHERN CALIFORNIA

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## ABSTRACT

A Jurassic ophiolite ~3 km thick, together with its cover of Upper Jurassic sedimentary rocks, is exposed along the southern California Coast near Point Sal. The ophiolite is partly dismembered by faulting but the original stratigraphy is nearly complete above the base of the ultramafic cumulate sequence. Cenozoic deposits mantle the ophiolite and isolate it from other exposures of the Jurassic ophiolite and the Franciscan Complex in nearby parts of the southern Coast Ranges.

The ophiolite sequence, going upward from the fault at its base, consists of serpentinite, layered ultramafic rocks (ol cumulates and ol-cpx cumulates) and olivine gabbro (ol-cpx-pl cumulates), non-cumulus uralite gabbro, diorite, and quartz diorite, a sill complex and associated dikes, and submarine lavas. Swarms of mafic dikes younger than the main dike and sill complex cut the lower (cumulate) part of the plutonic section. Tuffaceous radiolarian chert, overlain by mudstone with thin sandstone interbeds (flysch), rest on top of the volcanic rocks.

Pillowed and massive lavas are more than 1.3 km thick. The upper lava sequence is chiefly olivine basalt, with strong smectite alteration due partly to seafloor weathering. The much thicker lower lava sequence consists of strongly fractionated cpx-pl microphyric and aphyric basalts, now altered to spilite and keratophyre. Evidence from the chilled pillow rims and the pelagic interpillow sediments suggest that both groups of lavas were erupted into deep water in an open ocean environment, distant from any source of terrigenous detritus or airborne tephra. Petrographic features, low $TiO_2$ contents, and light REE depleted patterns (Menzies and others, 1977a,b) suggest that both groups of lavas were Group I mid-ocean ridge tholeiites.

The plutonic sequence crystallized within a magma chamber ~1.5 km thick, beneath a cover of volcanic rocks and sheeted sill complex 1.3-1.5 km thick. Fractionation by crystal settling produced a lower sequence of ultramafic and olivine gabbro cumulates, and an upper sequence of non-cumulus amphibole-rich gabbros, diorite, and quartz diorite that developed along the liquid line of descent. Unconformities and extensive internal disruption within the cumulus sequence, and the intrusion of cumulate mushes as dikes and diapirs, point to repeated tectonic disturbance within the magma chamber during the period of crystal accumulation and post-cumulus crystallization. Retrograde boiling, fluid migration, and extensive uralitization were important during the later stages of crystallization and dike intrusion. This evidence of exceptionally hydrous conditions within the magma chamber supports the suggestion, based on Sr-isotopic evidence (Davis and Lass, 1976), that seawater entered the magma chamber. The inference that the plutonic sequence crystallized beneath an actively rifting spreading center is consistent with the evidence of repeated tectonic disturbance and seawater penetration.

The Upper Dike Series includes (1) swarms of dikes that penetrate the lower lavas and upper plutonic rocks and (2) a unique sheeted sill complex that separates the volcanic and plutonic zones. Detailed analysis of dike and sill attitudes, compositions, and age relationships distinguishes four stages of dike intrusion that span the entire period of time over which the ophiolite developed. The significance of the sheeted sill complex differs from that of vertical sheeted dikes in other ophiolites: it records the rise of a new body of magma and the initial stages of its penetration into earlier oceanic crust.

The Lower Dike Series which cuts the cumulates marks the late stage of intrusion of basic magma that crystallized abundant orthopyroxene, which is unique at Point Sal. This represents a separate batch of magma, perhaps intruded from the side.

The ophiolite apparently represents composite oceanic crust formed in two stages from separate magmas, both of MORB type. Stage I was the initial formation of oceanic crust, represented by the strongly fractionated lower lavas (>1.1 km) and their NNE-trending feeder dikes. This early crust was cut off, intruded, and metamorphosed from below by the Stage II magma, which formed the sheeted sill complex, the stratiform plutonic sequence, and the olivine-phyric upper lavas. The Stage II magma was first emplaced as unfractionated tholeiitic melt with liquidus olivine and chromite. This magma then fractionated in place, and the upper plutonic rocks and later members of the Upper Dike Series are its late-stage fractionation products.

Stage I volcanism occurred at water depths below the CCD (much shallower in the Jurassic) where siliceous radiolarian ooze was accumulating. Stage II volcanism occurred slightly above the CCD where mainly calcareous pelagic ooze was accumulating. Extrusion at different geographic sites is inferred; yet a spreading ocean ridge setting is deduced for both stages. Possibly this two-stage ophiolite records the jumping of a spreading ocean ridge from its initial location to a new, distant site amid older crust. Such spreading center jumps are well documented in Cenozoic oceanic crust of the equatorial eastern Pacific.

## INTRODUCTION

The existence of widespread remnants of Mesozoic oceanic crust at the base of the Great Valley sequence in the California Coast Ranges was first documented by Bailey, Blake, and Jones (1970). They concluded that the western (distal) edge of the terrigenous Great Valley sequence was deposited upon oceanic crust now represented by the ophiolitic igneous sequence and its capping of cherts and marine tuffaceous sediments, and that subsequently the combined oceanic and terrigenous sequences were underthrust from the west by the Franciscan eugeosynclinal assemblage. Lanphere (1971), Page (1972), Pessagno (1973, 1977), and Bailey and Blake (1974) have further elucidated the age, structural relations, and petrochemistry of these oceanic igneous rocks and their pelagic cover, and have generally strengthened the 1970 conclusions. Conflicting interpretations have arisen regarding the tectonic setting in which the ophiolite developed, its mode of tectonic emplacement, and its structural relations with the Franciscan Complex (Maxwell, 1974; Blake and Jones, 1974; Hopson and others, 1975b; Jones and others, 1976), but no one yet has seriously challenged an oceanic crustal origin for the ophiolite itself.

Detailed reconstruction of the California Coast Range ophiolite (i.e., the collective ophiolitic remnants at the base of the Great Valley sequence) is hampered by the tectonic dismembering of the sequence and by generally poor exposures. These difficulties are perhaps least severe at Point Sal - one of the ophiolite localities recognized by Bailey, Blake, and Jones in 1970. Here the ophiolitic sequence comes nearest to being complete, and it is also relatively well exposed in seacliffs and wave-cut platforms along the coast.

The present report describes the field relations and petrology of the ophiolite at Point Sal, and its tectonic setting at the time of formation. Its subsequent tectonic history will be described elsewhere. The 1975 guidebook description (Hopson and others, 1975a) is updated and substantially revised, and special attention is given to unique features of this ophiolite: its sheeted sill complex, its strongly disturbed cumulates, and the complex age relationships between the volcanic, plutonic, and dike sequences. It is concluded that the Point Sal ophiolite is a remnant of composite oceanic crust that developed in two separate stages in ocean-ridge settings, perhaps by jumping of spreading centers.

## Other Studies

Igneous rocks of the Point Sal area were first recognized and described by Fairbanks (1896), and the general geology of the district is mapped and described by Woodring and Bramlette (1950). Bailey and others (1970) first recognized the ophiolitic sequence at Point Sal. Special studies of this ophiolite have dealt with its field relations and petrology (Hopson and Frano, 1973; Hopson and others, 1973, 1975a; Hopson, 1975, 1976), tectonics (Hopson and others, 1975b), radiometric age (Mattinson, in Hopson and others, 1975a), major element chemistry (Bailey and Blake, 1974), trace element and rare earth element geochemistry (Menzies and others, 1977a,b), Sr-isotope geochemistry (Davis and Lass, 1976), paleomagnetism and magnetic stratigraphy (Kempner, 1976, 1977; Kempner, Luyendyk, and Cockerham, 1976), seismic structure (Nichols, 1977). Paleontological studies of Jurassic strata that rest on the ophiolite include the mollusca (Woodring and Bramlette, 1950; D. L. Jones, in Hopson and others, 1975a), the radiolaria (Pessagno, 1973, 1977), and the dinoflagellates (W. R. Evitt, pers. commun., 1975).

## Stratigraphy and Structure

A composite stratigraphic section through the ophiolite at Point Sal, combining the three partial sections in Figure 2, consists of the following units (designated hereafter as zones): (1) submarine volcanic rocks, consisting of (a) an upper group of mainly olivine basalt with pervasive low-grade alteration, and (b) a lower group of spilitic pyroxene basalt and keratophyre, with low to high-grade greenschist-facies alteration; (2) dioritic and gabbroic rocks, grading progressively downward from (a) hornblende quartz diorite and diorite to (b) hornblende gabbro and uralitic clinopyroxene gabbro, (c) olivine-clinopyroxene gabbro, and (d) olivine clinopyroxene gabbro with troctolite and minor anorthosite; (3) ultramafic rocks, grading downward from (a) chiefly olivine clinopyroxenite to (b) serpentinized dunite with subordinate ol-clinopyroxenite and wehrlite, and (c) serpentinized dunite. Swarms of basaltic, diabasic, microdioritic, and epidositic sills and dikes form a sheeted sill complex (zone 1c) between the volcanic and upper plutonic rocks; also, swarms of similar dikes cut the overlying volcanics (zone 1b) and the underlying diorite and gabbro (zone 2a-b). A different swarm of low-angle dikes, mainly noritic microgabbros and feldspathic ol-clinopyroxenite, cut plutonic zones 2c-3d.

Above the ophiolite a thin sequence of tuffaceous radiolarian chert with rare limestone nodules lies in depositional contact on the upper basalt (ophiolite zone 1a). Overlying the chert is a mudstone-sandstone flysch, representing the base of the Late Jurassic and Cretaceous Great Valley sequence. These strata represent pelagic and then distal terrigenous sedimentation on top of the igneous sequence.

The ophiolite is exposed mainly along the coastline, in a strip extending six miles southeast from Point Sal (Fig. 1). The upper part of the ophiolite and overlying Upper Jurassic sedimentary strata dip 35-60° north at Point Sal and Point Sal Ridge, and are offset by northeast-trending vertical faults with left-lateral displacements that total at least 1 kilometer (Figs. 1, 3). The lower part of the ophiolite, two to five miles southeast of Point Sal (Figs. 1, 6), dips more steeply to the north and northeast and is locally overturned. The Lions Head fault with 4,000 feet of minimum vertical separation truncates the base of the ophiolite, throwing serpentinized dunite (zone 3c-d) against Miocene Monterey Formation. A steep ENE-trending fault with up to 2,500 feet of apparent left-lateral displacement separates gabbro zones 2b and 2c (Figs. 1, 3), and the upper part of zone 2c is hidden by Cenozoic formations (Figs. 1, 2, 6).

The faults described above are post-middle Miocene. Pre-Oligocene deformation is recorded by the unconformity at the base of the Lospe Formation. This unconformity cuts across progressively higher parts of the Jurassic sequence from south to north (and northwest), i.e., it truncates ultramafic rocks (ophiolite zone 3b) at Lions Head, gabbro (zone 2c) at Point Lospe, and Tithonian flysch (Great Valley sequence) near Point Sal and at Corralitos Canyon (Figs. 1, 6). Thus, the early Tertiary (pre-Lospe) erosion surface cut across Jurassic rocks that dipped gently northward. Plio-Pleistocene folding has brought these rocks to their present steep attitudes (Woodring and Bramlette, 1950, p. 12).

Tertiary strata and Quaternary deposits blanket the surrounding area for many miles to the north, east, and south, and the ocean is on the west. The ophiolite is thus isolated from other remnants of

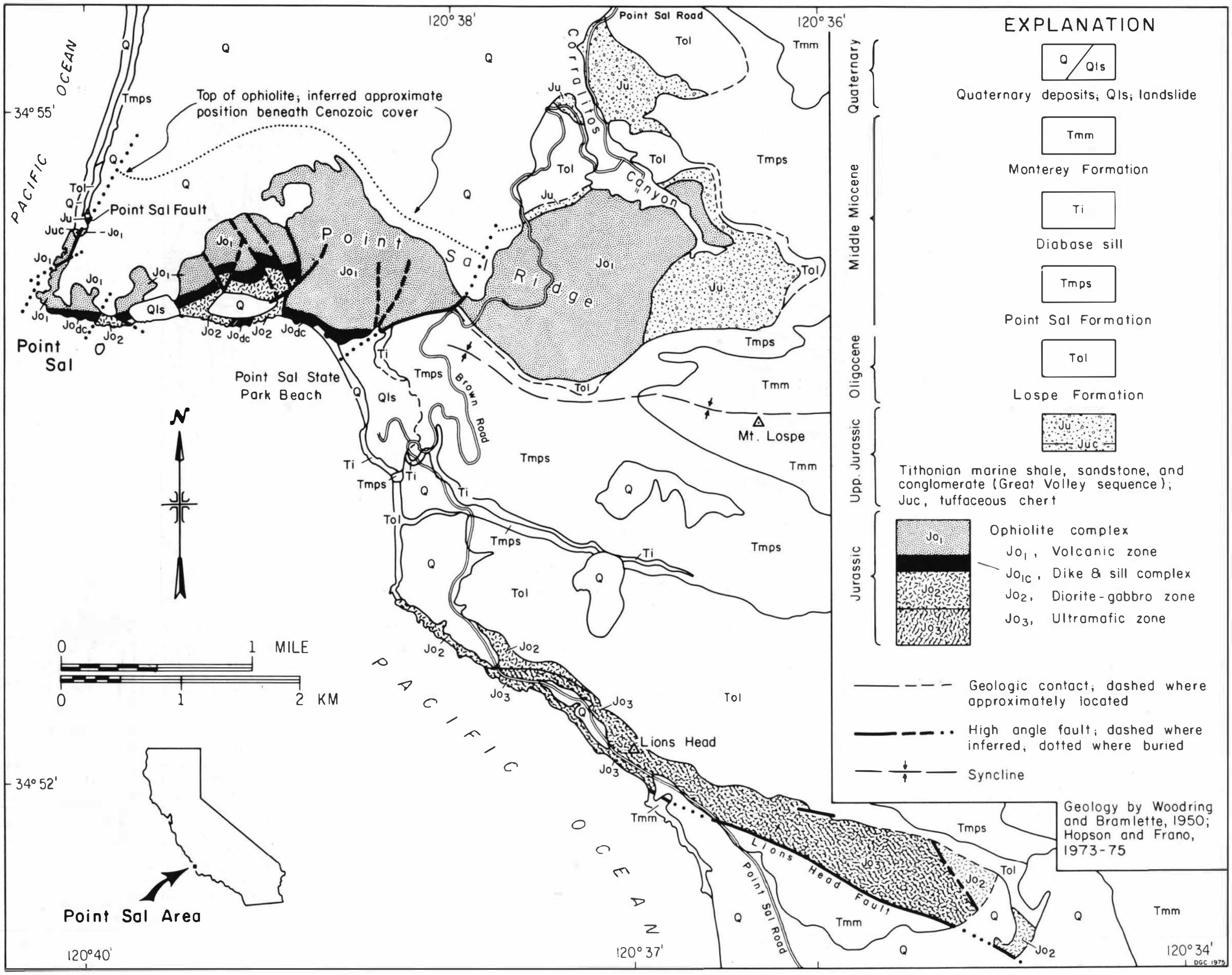


Figure 1. Geologic map of the Point Sal area, Santa Barbara County, California.

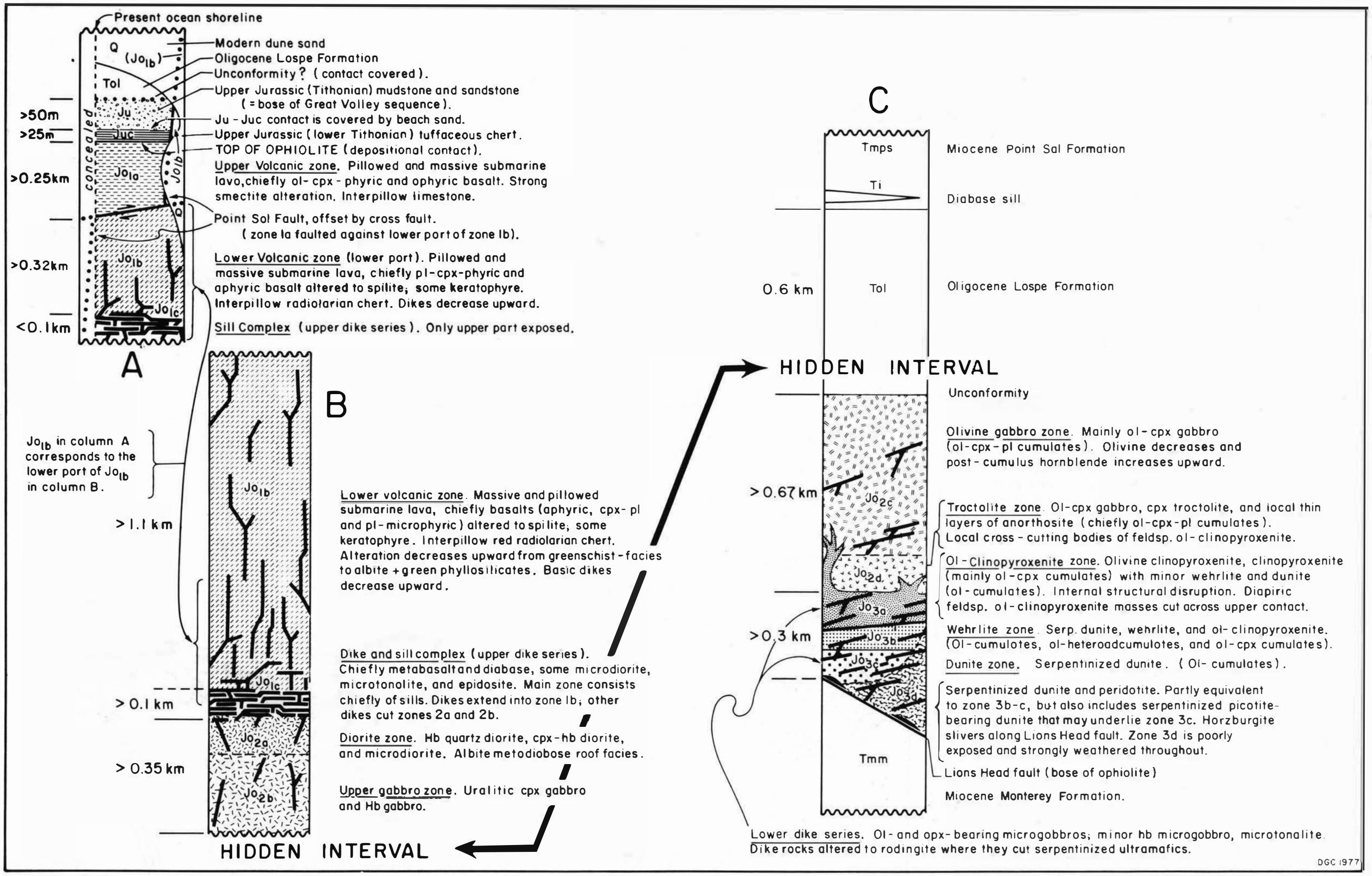


Figure 2. Columnar sections: A, coastal strip extending north from Point Sal (Fig. 3 inset); B, Point Sal Ridge, north and northwest of Point Sal State Park Beach (Fig. 3); C, Lions Head-Point Lospe area (Fig. 6).

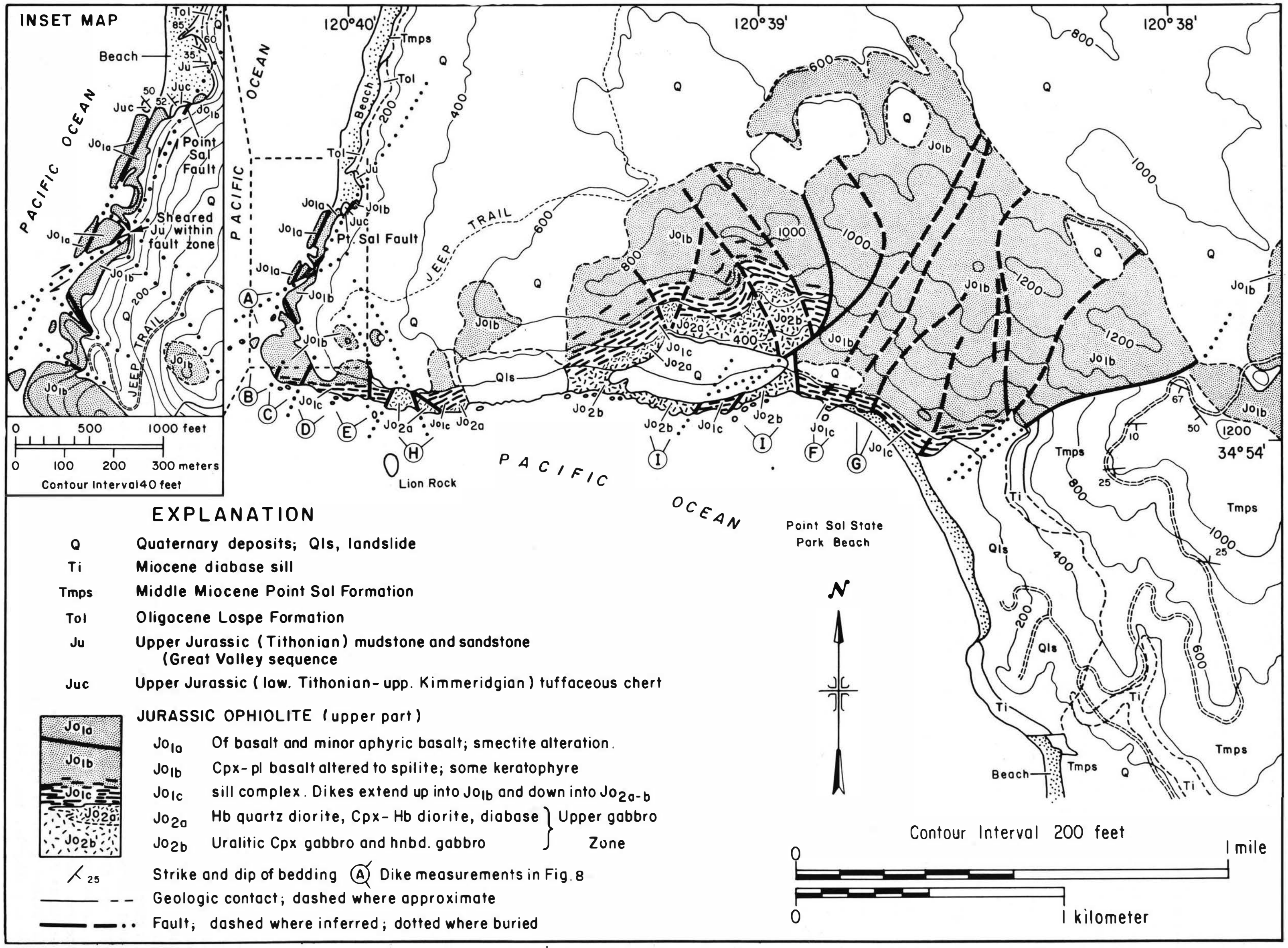


Figure 3. Geologic map of the Point Sal Ridge-State Park Beach area, showing ophiolite zones 1a-2b.

the Coast Range ophiolite and the underlying Franciscan Complex.

## DESCRIPTION OF THE OPHIOLITE

### The Volcanic Sequence

Field Relationships. Volcanic rocks more than 1 km thick overlie the upper plutonic rocks and sill complex at Point Sal Ridge (Fig. 2, section B), but their top is not exposed. Farther west, in the coastal strip extending north from the tip of Point Sal, nearly 600 meters of lava are continuously exposed beneath Upper Jurassic pelagic sedimentary rocks, which lie upon them in depositional contact (Fig. 2, section A). This section, however, comprises only the uppermost 250 m (zone 1a) and lowest 320 m (lower part of zone 1b) of the volcanic sequence; the middle of the sequence is cut out here by a major strike-slip fault (Point Sal fault, Figs. 1, 3). The volcanic section at Point Sal Ridge (section B) does not appear to extend high enough stratigraphically to include the distinctive upper lavas (zone 1a) of the coastal section (section A); therefore a composite section would include the full thickness of the former (approximately 1.1 km) plus at least zone 1a (250 m) of the latter. Thus, the total thickness of the composite volcanic section is not less than 1.3 km.

The volcanics consist mainly of lava, both pillowed and massive. A variety of pillow forms, including tubular pillows, are observed. Vesicles (now amygdules) are common within the interior of pillows but they become sparse and microscopic in the chilled outer rinds, characteristic of lavas erupted into very deep water (Moore, 1965, 1970; Moore and Schilling, 1973; Bryan, 1975). Breccia occurs locally within the volcanics but there is no bedded tuff. Small bits of pelagic sediment are widespread within the volcanic sequence, occurring as small pockets between the pillows and injected as dikelets and irregular gobs into brecciated lavas. This sediment is red radiolarian chert within the lower lavas (zone 1b) but chiefly pale gray coccolithic limestone (plus minor red mudstone and chert) in the upper lavas (zone 1a). Both of these sediments were originally biogenic oozes, and they are not contaminated by tuffaceous or terrigenous clastic components. Thus, the volcanic rocks at Point Sal are oceanic lavas that were probably erupted into deep water, in a setting that lay beyond the range of terrigenous sedimentation or airborne tephra from active arc-type volcanoes. An open-ocean environment is inferred.

Petrography. Lavas forming the upper 250 meters of the volcanic sections (zone 1a, coastal section A), are sparsely porphyritic olivine basalts and subordinate aphyric cpx basalt. Olivine, wholly pseudomorphed by smectite or calcite, originally formed sparse phenocrysts, while clinopyroxene and/or plagioclase occurs as much smaller microphenocrysts (Fig. 4; Table 1). Surviving original groundmass phases are plagioclase, clinopyroxene, and Fe-Ti oxide. Groundmass textures are most commonly intersertal, with the glass wholly altered to smectite and other products. Quench textures with skeletal crystal forms like some of those described by Bryan (1972) are also present.

Lavas forming the rest of the volcanic sequence (zone 1b, sections A, B), contrast strongly with the upper lavas in primary characteristics as well as their higher grade of alteration. Many of these lavas are aphyric; the rest are microporphyritic

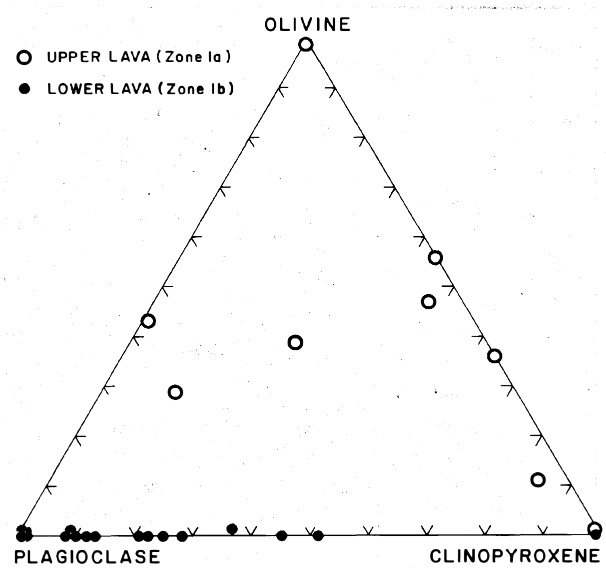


Figure 4. Ternary diagram showing modal abundance plot of phenocrystic and microphenocrystic olivine, clinopyroxene, and plagioclase in the Point Sal lavas (Table 1).

Table 1. Modal Analyses* of Phyric and Microphyric Lavas.

| UPPER LAVA (zone 1a) | OLIV | PLAG | CPX | GRDM | AMYG |
|---|---|---|---|---|---|
| PS69-1 | 5.5 | ... | 4.4 | 85.7 | 4.4 |
| PS69-2 | .5 | .3 | 4.0 | 91.4 | 3.8 |
| PS69-4 | ... | ... | 3.4 | 82.4 | 14.2 |
| PS75-64B | 10.4 | | 18.8 | 70.8 | |
| PS75-65B | 2.3 | ... | ... | 93.3 | 4.4 |
| PS153-1 | 7.0 | 9.2 | .2 | 83.6 | ... |
| PS154-1 | 2.2 | 1.9 | 1.6 | 93.0 | 1.3 |
| PS154-2 | 1.2 | 2.4 | .5 | 93.3 | 2.6 |
| PS155-1 | 1.6 | .2 | 1.6 | 94.7 | 1.9 |
| LOWER LAVA (zone 1b) | | | | | |
| PS19-1 | ... | 4.2 | ... | 84.8 | 11.0 |
| PS29-2 | ... | 5.2 | ... | 86.0 | 8.8 |
| PS31-1 | ... | 9.4 | 3.2 | 77.4 | 10.0 |
| PS42-2 | ... | 3.2 | 2.6 | 91.8 | 2.4 |
| PS65-1 | ... | ... | 1.8 | 97.8 | .4 |
| PS66-1 | .1 | 9.2 | 5.2 | 74.7 | 10.8 |
| PS67-4 | ... | 8.2 | 8.6 | 74.6 | 8.6 |
| PS109-1 | ... | 9.4 | 2.6 | 75.0 | 13.0 |
| PS114-1 | ... | 6.6 | 2.6 | 81.8 | 9.0 |
| PS115-1 | ... | 8.4 | ... | 70.0 | 21.6 |
| PS116-1 | ... | 13.0 | 1.8 | 67.0 | 18.2 |
| PS117-1 | ... | 11.4 | 3.0 | 65.4 | 20.2 |
| PS118-1 | ... | 4.2 | .6 | 87.2 | 8.0 |
| PS119-1 | ... | 10.2 | ... | 64.4 | 25.4 |
| PS121-1 | ... | 3.8 | .4 | 90.4 | 5.4 |
| PS122-1 | ... | 16.2 | 1.6 | 79.2 | 3.0 |
| PS123-1 | ... | 12.2 | 1.2 | 81.2 | 5.4 |

*Volume percent of phenocrysts (olivine, plagioclase, clinopyroxene), amygdules, and groundmass. Five hundred to 1100 points counted per specimen.

with plagioclase (now albitized) or clinopyroxene + plagioclase (albite) as the microphenocrystic phases (Fig. 4; Table 1). Pseudomorphs after olivine are very rare; olivine was evidently not a liquidus phase in most of the zone 1b lavas. The groundmass of these lavas is altered almost entirely to secondary minerals although original groundmass textures, especially lath-like feldspar microlites (now albite) are well preserved. Albite in the zone 1b lavas is evidently secondary, since relics of unreplaced plagioclase are still found within a few microphenocrysts.

Alteration. Pervasive alteration has strongly modified the mineralogy and chemical composition of lavas throughout the volcanic pile. The grade of alteration increases with stratigraphic depth, although precise details must await a fuller mineralogical study.

Smectite alteration characterizes the upper lava zone. All groundmass glass is replaced by green smectite; additional minor phases include calcite, celadonite, and brownish clay. Phenocrystic olivine is pseudomorphed by smectite or calcite, but clinopyroxene remains fresh. Groundmass plagioclase is clouded by clay alteration but is generally not albitized, except near the base of the zone. Zeolites appear to be lacking or quite rare, and it is probably misleading (Menzies and others, 1977a) to apply the term "zeolite facies" here. The strong smectite alteration in volcanic zone 1a, extending to the upper contact, appears to be comparable to seafloor weathering (halmyrolysis) described by various authors (Matthews, 1971; Miyashiro and others, 1971; Hart, 1973; Bass and others, 1973; Bass, 1975). In this connection it has been noted (Hopson and others, 1975a) that the top of the volcanic sequence experienced prolonged exposure to seawater prior to burial by sediments. The evidence is the large age gap (>10 m.y.) between the time the ophiolite formed at 160 m.y. B.P. (concordant zircon ages by J. M. Mattinson) and first burial of the upper lavas in the late Kimmeridgian-early Tithonian (Pessagno, 1977). Thus, there was ample opportunity for seafloor weathering to occur. Menzies and others (1977a,b) also reach this conclusion, based on negative Ce anomalies in the upper lavas.

The lower lavas (zone 1b) are pervasively albitized, but with good preservation of original textures. Typical secondary minerals in the upper parts of zone 1b include abundant albite and light-green phyllosilicate (chiefly iron-rich chlorite), plus sphene, calcite, quartz, small traces of epidote, rare pumpellyite (but no prehnite), and minor brown clay. The common amygduloidal minerals here are quartz, calcite, smectite, chlorite(?), and pumpellyite. The effects of more than one stage of alteration may be superimposed here, as described elsewhere by Bass (1975). We recommend discontinuing use of the term "prehnite-pumpellyite facies" to describe this low-pressure alteration assemblage (Hopson and others, 1975a; Menzies and others, 1977a,b), for the two key phases do not coexist here.

Still deeper, in the lower 200-300 m of the volcanic sequence, the common groundmass alteration products and amygdule-filling phases are albite, epidote, chlorite, rare actinolite, magnetite, sphene, quartz, and calcite - clearly a greenschist-facies assemblage. Also, strong epidote metasomatism is evident in the lower 100 meters of the volcanic sequence, where epidote pervasively replaces the matrix of basaltic breccias and pillow lavas. The alteration and metasomatism of the lower part of the volcanic sequence is doubtless a form of low-grade metamorphism, induced by heat and fluids rising from the magma body (plutonic sequence) that crystallized just beneath. Higher in volcanic zone 1b it is difficult to separate the superimposed effects of: (1) metamorphism which decreases upward, (2) halmyrolysis which dies out downward, and (3) pervasive alteration which may have resulted from the circulation of heated sea water shortly after the lavas accumulated.

Chemistry. Chemical analyses of representative Point Sal volcanic rocks are given in Table 2. Nearly all the Point Sal lavas are so highly altered that it seems fruitless to attempt close comparisons of their chemistry with fresh lavas and glasses from mid-ocean ridges or with moderately fresh DSDP lavas. Such comparisons show that the Point Sal lavas are significantly higher in $Na_2O$, $K_2O$, $SiO_2$, $FeO^*/MgO$, $Fe_2O_3/FeO+Fe_2O_3$, $H_2O$, and $CO_2$, slightly lower in $TiO_2$, and strongly depleted in CaO and MgO. Some of these differences are readily correlated with mineral alterations. For example, the complete alteration of glass and olivine to smectite caused gains in alkalis (especially $K_2O$), $H_2O$, and losses in CaO and MgO,

**Table 2. Chemical Analyses and CIPW Norms of Point Sal Volcanic Rocks.**

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| $SiO_2$ | 50.80 | 52.65 | 55.5 | 57.5 |
| $TiO_2$ | .57 | .64 | .55 | .61 |
| $Al_2O_3$ | 15.54 | 15.45 | 15.1 | 14.5 |
| $Fe_2O_3$ | 4.28 | 5.92 | 1.2 | 1.7 |
| FeO | 5.89 | 3.98 | 6.1 | 5.3 |
| MnO | .41 | .13 | .11 | .08 |
| MgO | 7.06 | 4.96 | 6.4 | 3.3 |
| CaO | 3.94 | 5.66 | 6.4 | 5.5 |
| $Na_2O$ | 4.99 | 5.10 | 4.0 | 5.2 |
| $K_2O$ | .98 | .78 | .34 | .20 |
| $H_2O^+$ | 3.38 | 2.53 | 2.4 | 2.6 |
| $H_2O^-$ | 1.42 | 1.42 | .51 | .28 |
| $P_2O_5$ | .08 | .09 | .05 | .06 |
| $CO_2$ | .70 | .69 | .08 | 3.5 |
| Total | 100.04 | 100.00 | 98.7 | 100.3 |
| FeO*/MgO | 1.38 | 1.88 | 1.12 | 2.07 |
| Q | 0.0 | 3.08 | 5.34 | 15.08 |
| Or | 5.97 | 4.76 | 2.07 | 1.19 |
| Ab | 46.19 | 47.27 | 37.09 | 47.20 |
| An | 15.05 | 17.51 | 22.98 | 4.82 |
| Cor | 1.04 | .0 | .0 | 4.39 |
| Di | .0 | 5.09 | 7.17 | .0 |
| Hy | 19.59 | 12.99 | 22.95 | 15.58 |
| Ol | 5.84 | .0 | .0 | .0 |
| Mt | 4.61 | 6.39 | 1.30 | 1.80 |
| Ilm | .82 | .92 | .79 | .86 |
| Ap | .17 | .19 | .11 | .13 |
| Ne | .0 | .0 | .0 | .0 |
| CC | 1.83 | 1.80 | .21 | 8.95 |

1. Olivine basalt (PS69-2; smectite alteration); ophiolite zone 1a. Analysis by H. N. Elsheimer, USGS.
2. Aphyric cpx basalt (PS68-1; smectite alteration); ophiolite zone 1a. Analysis by H. N. Elsheimer, USGS.
3. Spilite (70-B-302; albitized cpx basalt); ophiolite zone 1b. (Bailey and Blake, 1974, p. 643.)
4. Keratophyre (71-EB-105); ophiolite zone 1b. (Bailey and Blake, 1974, p. 644.)

especially in the smectite-rich upper lavas. Nearly complete albitization of plagioclase in the lower lavas (zone 1b) caused large gains in $Na_2O$, $SiO_2$, and depletion in CaO. Also significantly affecting the whole-rock analyses of Point Sal lavas are ubiquitous tiny micro-amygdules, filled with quartz, smectite, celadonite, albite, and other secondary minerals. Thus, because of the high degree of alteration (20-60% for the upper lavas; 70-100% for the lower lavas) and the difficulty of avoiding micro-amygdules, an accurate characterization of the Point Sal volcanic rocks in terms of major-element chemistry is probably not possible. Relict petrographic features therefore probably provide a more reliable, though limited, guide to the original character of these lavas than many aspects of their bulk chemistry.

The so-called keratophyres at Point Sal (Bailey and Blake, 1974; Hopson and others, 1975a) illustrate this point. These lavas (Table 2, nos. 3, 4) are classed as keratophyres following the usual criteria: high $SiO_2$, high $Na_2O$, low CaO, and albite as the principal feldspar (Gilluly, 1935; Hamilton, 1963, Fig. 67). But, petrographic evidence indicates that rocks with this chemistry at Point Sal are chiefly derived from the metasomatic alteration (particularly albitization) of aphyric and pl-microphyric basalts. The high $Na_2O$:CaO ratio of analyzed samples results from albitization of the original calcic plagioclase, and the high $SiO_2$ values result from the albitization and also partly from micro-amygdules filled with secondary quartz. If these lavas had originally been significantly quartz normative then quartz should be conspicuous in the groundmass alteration assemblage, but that is not commonly the case. Thus, lavas originally of intermediate ("andesitic") composition are now deemed to be only minor components of the Point Sal volcanic sequence.

Mobility of the major elements suggests that caution must also be exercised in the application of trace element and rare earth element data to the Point Sal lavas (Blake and Bailey, 1975; Menzies and others, 1977a,b). Certain trace elements (e.g., Ti, Y, Zr, Nb) are known to be resistant to moderate degrees of alteration and metamorphism (Bryan and others, 1976), but it should be borne in mind that most Point Sal lavas show 40-95 percent alteration to secondary minerals and that the major element bulk composition has changed significantly. Menzies and others (1977a) discuss these effects at Point Sal, particularly for the rare earth elements. They show that: (1) greenschist facies alteration of the lower lavas has not significantly effected the REE profiles, and (2) that low-temperature alteration of the upper lavas has induced some mobility of the light rare earths, resulting in negative Ce (or positive La) anomalies. It is important to note that both the upper and lower lavas still retain light REE-depleted patterns (Menzies and others, 1977a,b), which tend to be eliminated by reactions that involve sea water (Frey and others, 1974). The LREE-depleted patterns are significant here, for among oceanic lavas they are known only from basalts of MORB type (Bryan and others, 1976).

Ultramafic lava is reported from the Point Sal volcanic sequence by Blanchard and others (1977) and Menzies and others (1977b), on the basis of chemistry. This identification is from two drill core samples collected by Kempner (1977) from coastal section A, GSA guidebook locality 19 (Hopson and others, 1975a), in outcrops now placed in lowermost zone 1a (Fig. 3). I have examined Kempner's specimens and re-examined the outcrops and I disagree that the lavas (which are highly altered) are ultramafic. Kempner's specimens both contain phenocrystic olivine (wholly pseudomorphed by smectite and calcite) and microphenocrystic clinopyroxene (mostly fresh) in a groundmass with skeletal quench plagioclase (replaced by albite in PS75-65B), smectite, and Fe-oxides. Both rocks contain micro-amygdules, one filled with smectite and the other with smectite + celadonite + calcite. These two rocks are indeed somewhat richer in phenocrystic olivine and microphenocrystic pyroxene than other zone 1a lavas (Table 1), but this is due to crystal accumulation. Thus, they are not ultramafic nor similar to komatiites petrologically. The misnaming of these rocks, however, does not necessarily invalidate the conclusion of Menzies and others (1977a,b) that they represent the most primitive lava exposed at Point Sal; this seems supported by the high Cr, Ni, and Sc values and very low REE abundances.

Conclusions. The following conclusions are reached concerning the original character of the Point Sal lavas prior to their alteration.

(1) The upper and lower lavas were chiefly subalkaline (tholeiitic) basalts, including some rocks now altered to "keratophyre."

(2) These basalts correspond most closely to the Group I oceanic basalts of Bryan and others (1976), which is the kind typically erupted at spreading ocean ridges. Criteria that link them more closely with Group I than Group II basalts (which include the basalts from aseismic ridges, oceanic islands and other off-ridge centers) are their low $TiO_2$ and low $P_2O_5$ contents (Fig. 5), non-titaniferous pyroxenes, and their LREE-depleted character reported by Menzies and others (1977a,b).

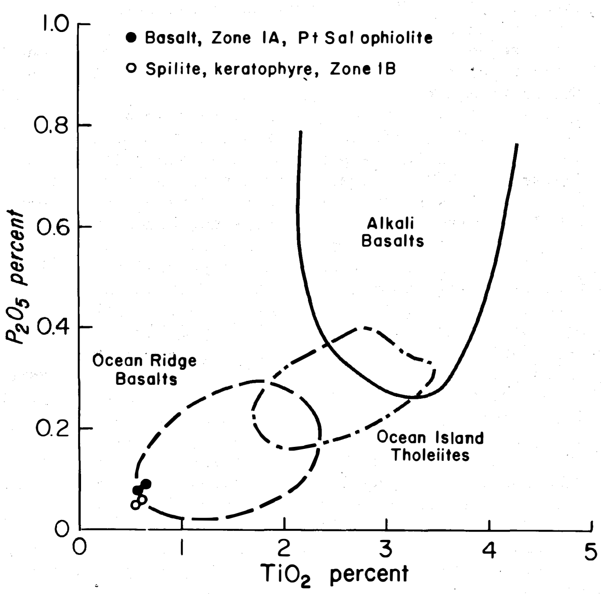


Figure 5. $TiO_2$-$P_2O_5$ plot for the Point Sal volcanic rocks. Fields for oceanic basalts from Bass and others, 1973.

(3) The upper lavas (zone 1a) are chiefly olivine-phyric, ol-cpx and ol-cpx-pl microphyric basalts, plus minor aphyric pl-cpx basalt. The apparent paragenesis is ol, ol+cpx±pl, pl+cpx±mt, comparable to the paragenesis of the underlying plutonic sequence. Some of the ol-phyric basalts are probably not extensively fractionated, a conclusion

supported also by the trace element and REE data (Menzies and others, 1977a,b). The scarcity of phenocrystic plagioclase is a unique feature in some of these lavas, which might be explained by the floating off of plagioclase in the underlying magma chamber (Bryan and Moore, 1977).

(4) The highly altered lower lavas, comprising most of the volcanic sequence (zone 1b), were chiefly derived from aphyric, pl-microphyric, and pl-cpx microphyric lavas (Fig. 4). A few examples of pl-cpx-ol phyric basalt are found but olivine-bearing lavas are rare in zone 1b. Thus, it appears that the lower lavas represent rather strongly fractionated basalts, probably corresponding to some of the highly fractionated (ol-free) MOR tholeiites found locally on the Mid-Atlantic Ridge (Hekinian and others, 1977), the western Atlantic (DSDP sites 100, 386; see esp. Ayuso and Bence, 1976), the south Atlantic (DSDP site 15; Frey and others, 1974), the Caribbean (DSDP sites 146, 150, 152, 153; Bence and others, 1975), the western Pacific (DSDP sites 58, 61, 63, 66, 285, 303, 304, 307), the central Pacific (DSDP site 317A), the northwestern Pacific (DSDP site 191), the northeastern Pacific (DSDP site 179), and the southeastern Pacific (Nazca plate, DSDP site 321, see esp. Mazzullo and Bence, 1976).

(5) Keratophyres that were truly oversaturated (i.e., have groundmass quartz) are scarce at Point Sal. They were the product of extreme fractionation and correspond to differentiated quartz diorites found in the plutonic sequence. This conclusion is also supported by the REE data (Menzies and others, 1977a).

(6) Lava reported to be ultramafic is actually altered basalt somewhat richer than average in phenocrystic olivine and microphenocrystic clinopyroxene.

Finally it is noted that the lower lavas were erupted in a deep ocean environment where red radiolarian ooze was the only sediment accumulating, whereas the upper lavas were erupted under conditions in which light-gray calcareous pelagic ooze was accumulating. It is suggested in a later section that the two lava groups are the product of two different magmas, erupted at geographically different sites.

## The Plutonic Sequence

A differentiated sequence of plutonic rocks more than 1 km thick lies beneath the volcanic rocks and the sheeted sill complex at Point Sal (Figs. 1, 2). It consists of an upper section of non-layered uralitic gabbro, diorite, and quartz diorite (zone 2a-b; Fig. 2, section B) separated by a fault and a covered interval from the lower section of layered olivine gabbro and ultramafic rocks (zones 2c-d, 3; Fig. 2, section C). The plutonic sequence is cut off below zone 3c by the Lions Head fault (Figs. 1, 6).

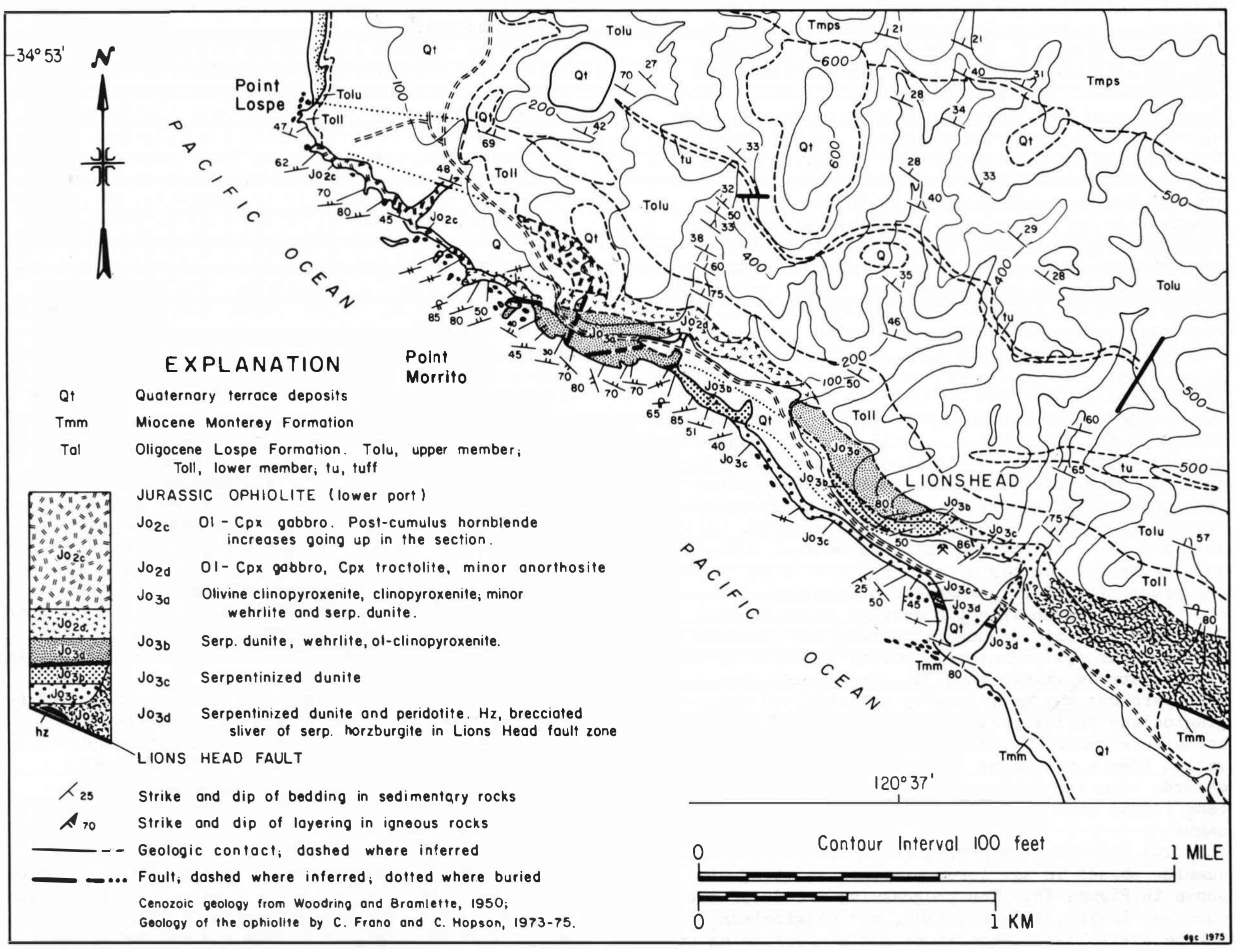


Figure 6. Geologic map of the Lions Head-Point Lospe area, showing ophiolite zones 2c-3d.

Lower Plutonic Sequence. Proceeding upward from the base, the sequence consists of more than 100 m of serpentinized dunite (zone 3c), 80 m of serpentinized dunite with some wehrlite and olivine clinopyroxenite (zone 3b), more than 120 m of olivine clinopyroxenite with subordinate clinopyroxenite, wehrlite, and dunite (zone 3a), 120 m of ol-cpx gabbro with minor troctolite, anorthosite, and cross-cutting intrusive masses of feldspathic ol-clinopyroxenite (zone 2d), and more than 500 m of cpx-ol gabbro with minor hornblende-bearing cpx gabbro (zone 2c). These are minimum thickness estimates for zones 2c, 3a, 3b, 3c, due to the unconformity at the top of 3c and faults between 3a-3b and at the base of 3c.

These rocks are all igneous cumulates, much disturbed and disrupted by penecontemporaneous deformation. Features indicative of cumulus origin are cumulus textures, planar orientation of elongate crystals (igneous lamination), cm- to m-scale ratio layering and size layering, and a regular progression of km-scale phase (mineral-graded) layering and cryptic (chemical-graded) layering going upward through the section (Fig. 7).

Zonal subdivisions within the lower plutonic section are based on the cumulus phase layering and related features. Zone 3c consists of olivine-chromite adcumulates and mesocumulates, with clinopyroxene and calcic plagioclase (altered) as minor post-cumulus phases. The dunite is more than 90 percent serpentinized but preserves its original texture. The dunite is massive except for local thin zones (layers?) with higher chromite content. Zone 3b continues as mainly olivine cumulates but clinopyroxene first appears here as a local cumulus phase (ol-cpx adcumulates) and also as post-cumulus oikacrysts (ol-heteroadcumulates). Zone 3a marks the incoming of clinopyroxene as a major cumulus phase. The main layers, much disrupted, are cpx-ol adcumulates and mesocumulates (with post-cumulus plagioclase and hornblende), ol-heteroadcumulates (wehrlite with cpx oikacrysts), and local thin ol-adcumulates (dunite). Zone 2d marks the first appearance of cumulus plagioclase. These rocks are mainly pl-cpx-ol adcumulates and some mesocumulates, with pale brown post-cumulus hornblende. The zone 2d gabbros are rich in olivine and grade locally into troctolite. Cm-scale anorthosite layers with reverse grading (pl-rich at the base) are a unique minor feature. Zone 2c continues as pl-cpx-ol adcumulates and mesocumulates but cumulus olivine decreases and disappears entirely near the top of the section. Post-cumulus hornblende increases in abundance upward. These mineralogical trends continue without major interruption into the upper plutonic section, which suggests that the covered interval between zones 2c and 2b consists mainly of cpx gabbro with an increasing proportion of late magmatic hornblende.

Orthopyroxene is not a cumulus phase in the Point Sal plutonic sequence. A very few remnants of post-cumulus opx (bronzite), mostly replaced by hornblende, appear within zone 3a. It appears that orthopyroxene may have locally crystallized following plagioclase in the paragenetic sequence, but that hornblende generally substituted instead. Crystallization from a wet magma is thus indicated. This accords with the Sr-isotopic evidence of Davis and Lass (1975) that sea water mixed with the Point Sal magma.

The results of a preliminary microprobe study of cumulus phases in the lower plutonic section, are shown in Figure 7B. The progressive trends of the cumulus olivine, clinopyroxene, and plagioclase toward their lower temperature end members going up section indicates crystallization and settling from a progressively more fractionated magma. The local reversals and fluctuations in this trend, especially within zone 2d, may reflect pulses of new magma; however, a more detailed study will doubtless modify and perhaps change some of the details shown here.

Numerous features within the lower plutonic section suggest that layers accumulating on the floor of the magma chamber were extensively deformed and disrupted before some of them were fully solidified. Most conspicuous is the pinching and swelling, boudinaging, and chaotic disruption of cumulus layering. The disruption is so extensive that individual primary layers can rarely be traced for more than a few meters. The resemblance to soft-sediment deformation and disruption of bedding is striking. Yet, igneous textures are well preserved, even within the most highly disrupted layers, and evidence of subsolidus recrystallization or low-temperature cataclasis is rare. Thus, deformation occurred at temperatures above solidus, while some layers are still mushes with interstitial melt. Another distinctive feature is intrusive dikes of ultramafic cumulate material that resemble sandstone dikes in disturbed sedimentary sequences. These intrusive ultramafics are common as dikes within pyroxenite zone 3a and they penetrate as dikes and diapiric masses up into the lower 250 m of the overlying gabbro (zones 2c-d). The textures and internal structure of these ultramafic intrusives suggest that they are emplaced as crystal mushes.

These deformations and auto-intrusions within the cumulus sequence can be explained as resulting from the strong tectonic disturbance of interlayered solids (adcumulates) and mushes (mesocumulates) at temperatures still above the solidus (Hopson, 1975).

Other features within the lower plutonic sequence point to repeated tectonic disturbance at the upper surface of the growing pile of cumulates, while crystal sedimentation was still actively in progress. These features include highly contorted small-scale folds along local horizons, due to slumping, and unconformities (mainly of buttress-type) between packets of cumulus layers. Both these features could have resulted from tilting of the magma chamber floor, if the upper layers had acquired some rigidity due to adcumulus crystallization. Slump folds were produced where thin cohesive layers (adcumulates) resting on mush layers (mesocumulates) were tilted and slid downhill. Continued crystal accumulation then built new horizontal layers above the deformed layers, creating unconformities. Still another feature perhaps related to large-scale sliding and mixing of crystal mushes (mesocumulates) are massive gabbros that have only local vestiges of contorted layering but are permeated by diffuse patches and dikelets of "pegmatitic" gabbro. These small gabbroic neosomes perhaps represent intercumulus melts that were filter pressed from mesocumulate crystal mush during sliding. This type of "veined" gabbro is abundant within the central part of zone 2c.

A final feature stemming from ocean-crust deformation before the close of plutonic activity are microgabbroic dikes with gneissic textures and schlieren banding, which cross-cut the deformed cumulates of zones 2c-3c. The internal gneissic texture and structure generally parallels the dike walls and cuts across the igneous structures outside the dikes. This results from protoclastic flowage and smearing of crystal mush within partly solidified dikes, when the cumulate host rocks are already solid and rigid. This records occasional pulses of deformation when the lower dike sequence was being emplaced into cumulates that were cooling below the

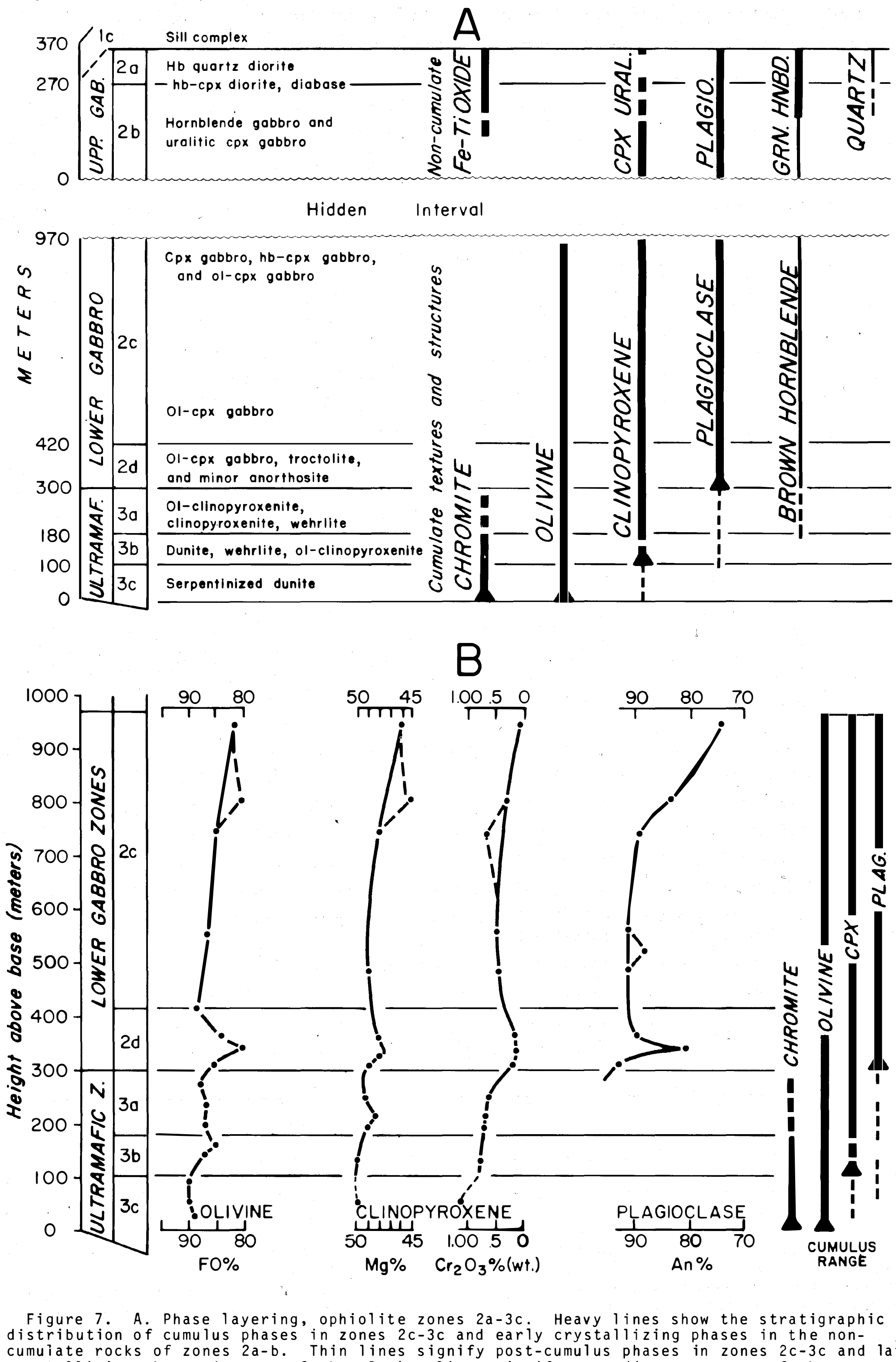


Figure 7. A. Phase layering, ophiolite zones 2a-3c. Heavy lines show the stratigraphic distribution of cumulus phases in zones 2c-3c and early crystallizing phases in the non-cumulate rocks of zones 2a-b. Thin lines signify post-cumulus phases in zones 2c-3c and late-crystallizing phases in zones 2a-b. Broken lines signify sporadic occurrence of phases.
B. Cryptic variation (chemically graded layering) in cumulate rocks of ophiolite zones 2c-3c (lower plutonic section). Stratigraphic range of cumulus phases shown on the right.

Table 3. Chemical Analyses and CIPW Norms for Ultramafic and Gabbro Cumulates, Lower Plutonic Section.

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 |
|---|---|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 33.07 | 39.9 | 40.45 | 43.99 | 44.6 | 50.63 | 48.65 | 43.32 | 46.9 | 47.74 | 48.45 |
| $TiO_2$ | .04 | .0 | .07 | .10 | .04 | .19 | .17 | .06 | .06 | .19 | .26 |
| $Al_2O_3$ | .65 | .70 | .68 | 1.44 | 2.0 | 4.01 | 3.75 | 21.55 | 20.6 | 18.15 | 18.19 |
| $Fe_2O_3$ | 5.34 | 6.5 | 6.23 | 2.29 | 3.7 | .51 | .92 | 1.13 | .40 | .56 | .73 |
| FeO | 3.58 | 2.0 | 2.88 | 4.28 | 3.6 | 2.95 | 4.18 | 3.98 | 4.1 | 3.06 | 2.84 |
| MnO | .16 | .08 | .12 | .13 | .09 | .10 | .11 | .07 | .04 | .07 | .07 |
| MgO | 38.18 | 34.6 | 32.44 | 28.54 | 27.5 | 17.50 | 20.12 | 11.40 | 9.0 | 11.11 | 9.76 |
| CaO | .30 | 2.3 | 4.02 | 10.01 | 11.5 | 20.66 | 17.57 | 11.79 | 14.2 | 16.98 | 15.09 |
| $Na_2O$ | .34 | .0 | .29 | .28 | .12 | .48 | .60 | 1.64 | 1.4 | .99 | 1.77 |
| $K_2O$ | .02 | .10 | .02 | .02 | .08 | .03 | .02 | .02 | .14 | .06 | .47 |
| $H_2O^+$ | 15.18 | 11.5 | 10.98 | 7.51 | 5.6 | 1.26 | 2.15 | 3.32 | 1.6 | 1.27 | 2.20 |
| $H_2O^-$ | .64 | 1.1 | .81 | .18 | .36 | .08 | .23 | .22 | .13 | .13 | .16 |
| $P_2O_5$ | .61 | – | .03 | .04 | – | .06 | .05 | .05 | – | .05 | .05 |
| $CO_2$ | .30 | .23 | .24 | <.05 | .04 | <.05 | <.05 | <.05 | .04 | <.05 | <.05 |
| $Cr_2O_3$ | .73 | | .22 | .44 | – | .73 | .44 | ... | ... | ... | ... |
| NiO | .22 | | .15 | .15 | – | .05 | .08 | ... | ... | ... | ... |
| Total | 99.36 | 99.0 | 99.63 | 99.40 | 99.23 | 99.24 | 99.04 | 98.55 | 98.6 | 100.36 | 100.04 |
| FeO*/MgO | 0.23 | .25 | .28 | .23 | .27 | .20 | .25 | .49 | .50 | .33 | .37 |
| Q | ... | ... | ... | ... | ... | ... | ... | ... | ... | ... | ... |
| Or | .13 | .63 | .12 | .12 | .48 | .18 | .12 | .12 | .84 | .35 | 2.78 |
| Ab | 3.30 | .0 | 2.75 | 2.58 | 1.09 | 3.29 | 4.46 | 13.50 | 12.74 | 8.80 | 14.99 |
| An | -4.75 | 1.71 | .52 | 2.69 | 4.74 | 8.72 | 7.51 | 52.41 | 50.20 | 44.44 | 40.35 |
| Ne | ... | ... | ... | ... | ... | .62 | .57 | .91 | ... | ... | .55 |
| Cor | 1.98 | ... | ... | ... | ... | ... | ... | ... | ... | ... | ... |
| Di | ... | 7.08 | 15.01 | 38.48 | 42.25 | 74.94 | 63.76 | 5.53 | 16.98 | 30.89 | 27.42 |
| Hy | 15.68 | 40.58 | 33.27 | 12.77 | 12.62 | ... | ... | ... | 6.43 | 1.13 | ... |
| Ol | 75.37 | 42.88 | 40.62 | 40.67 | 34.73 | 11.33 | 22.27 | 26.13 | 12.30 | 13.45 | 12.68 |
| Mt | 6.04 | 5.13 | 6.89 | 2.46 | 3.92 | .53 | .97 | 1.20 | .42 | .58 | .76 |
| Ilm | .06 | ... | .10 | .14 | .06 | .27 | .24 | .09 | .08 | .26 | .36 |
| Ap | 1.38 | ... | .07 | .09 | ... | .13 | .10 | .11 | ... | .10 | .10 |
| CC | .82 | .62 | .64 | ... | .10 | ... | ... | ... | ... | ... | ... |
| Ht | ... | 1.39 | ... | ... | ... | ... | ... | ... | ... | ... | ... |

1. Serpentinized dunite (PSF-151); ophiolite zone 3c near Lions Head. Analysis by H. N. Elsheimer, USGS.
2. Serpentinized wehrlite (71-EB-101); ophiolite zone 3b. (Bailey and Blake, 1974, p. 639.)
3. Wehrlite (PS20-1); ophiolite zone 3b at Lions Head. Analysis by H. N. Elsheimer, USGS.
4. Olivine clinopyroxenite (PS20-2); ophiolite zone 3a at Lions Head. Analysis by H. N. Elsheimer, USGS.
5. Olivine clinopyroxenite (71-EB-103); ophiolite zone 3a. (Bailey and Blake, 1974, p. 641.)
6. Clinopyroxenite (PSF-511); ophiolite zone 3a near Point Morrito. Analysis by H. N. Elsheimer, USGS.
7. Feldspathic olivine clinopyroxenite (PS12-2). Intrusive mass cutting gabbro, base of ophiolite zone 2d near Point Morrito. Analysis by H. N. Elsheimer, USGS.
8. Clinopyroxene troctolite (PM-31); ophiolite zone 2d near Point Morrito. Analysis by H. N. Elsheimer, USGS.
9. Ol-cpx gabbro (71-EB-104); ophiolite zone 2d. (Bailey and Blake, 1974, p. 642.)
10. Ol-cpx gabbro (PS15-1); base of ophiolite zone 2c near Point Morrito. Analysis by H. N. Elsheimer, USGS.
11. Hornblende-bearing cpx gabbro (PS16-1); highest exposure in ophiolite zone 2c near Point Lospe. Analysis by H. N. Elsheimer, USGS.

solidus. Strain at this stage was transmitted by movement through the incompetent mushy dikes.

To summarize: deformation and sliding occurred repeatedly while igneous cumulates were building up within the magma chamber. Further deformation then continued to internally disrupt the layered cumulate sequence as it cooled to solidus temperatures. Finally, still later deformations were transmitted through crystallizing cross-cutting dikes after the cumulates had fully solidified. The most likely setting where repeated strong tectonism would accompany the formation and cooling of plutonic oceanic crust is a spreading ocean ridge (Hopson, 1975).

Upper Plutonic Section. Gabbroic and dioritic rocks more than 350 m thick crop out along the beach and south side of Point Sal Ridge, 0.4 to 2.0 km east of the tip of Point Sal. These rocks are overlain by the dike and sill complex and the volcanic sequence on the north, and are faulted against those two units on the east. Slices of the dike and sill complex have locally been faulted down into the plutonic sequence (Fig. 3).

There is a gradation from uralitic cpx gabbro (zone 2b) through diorite to hornblende quartz diorite (zone 1a) going up the mountainside from the beach and also from east to west along the shoreline (Fig. 3). The mountainside is poorly exposed and the shoreline section is broken by faults and locally covered by a landslide; nevertheless, a reasonably complete stratigraphic section through the upper 350 m of the plutonic section can be studied here.

The zone 2b gabbros are structurally and texturally isotropic. There is no sign of layering or igneous lamination, nor do the rocks have cumulus textures. Rather, random mineral orientation, subophitic to hypidiomorphic textures, and small-scale variation in grain size from diabasic to pegmatitic are typical. Plagioclase (zoned $An_{70-30}$) and augite were the early liquidus phases. Green hornblende has partly replaced cpx and also grown as a primary phase, during the later stages of magmatic crystallization. Magnetite abruptly appears as an abundant (4-8 percent) late magmatic phase in the upper part of zone 2b. Interstitial spaces in the gabbro are filled by masses of very fine fibrous green amphibole, probably the product of vapor-phase mineralization.

Large, locally abundant miarolytic cavities are a distinctive feature of the zone 2b gabbro. Retrograde boiling is thought to be responsible. Locally these cavities are crudely aligned into "trains" that plunge SW; these linear strings of cavities would be subvertical if the ophiolite were restored to a horizontal position. Gas streaming is a probable explanation. The cavities are filled with finely fibrous vapor-phase amphibole.

Continuing up-section into zone 2a the plagioclase becomes more sodic, clinopyroxene becomes more extensively uralitized and then green hornblende takes its place as the early mafic phase, and finally interstitial quartz appears. Quartz diorite, consisting of zoned plagioclase ($An_{45-15}$), green hornblende, magnetite, and 5-15 percent of interstitial quartz or quartz-albite intergrowths forms a discontinuous zone 0-100 feet thick just below the top of the plutonic sequence. Small, diffuse segregations of aplitic albite granite occur locally within the quartz diorite. Continuing up section the quartz diorite appears to grade back into quartz-bearing augite-hornblende diorite and then into a thin roof facies of clinopyroxene diabase, which has been extensively recrystallized and metasomatically altered (albitized). Poor exposures and resemblance in the field of this roof-facies meta-diabase to rocks of the overlying sill complex make the uppermost part of zone 1a very difficult to study (see Hopson and others, 1975a, p. 31 for a fuller description).

The upper plutonic section (zones 2a-b) is cut by abundant dikes of meta-basalt, diabase, microdiorite, epidosite, and albitite. These are discussed in the section on Dike Sequences.

Discussion. The plutonic sequence (zones 2a-3c) crystallized and differentiated from a single magma body, within a chamber beneath the volcanic sequence. Crystal fractionation by gravity settling resulted in a lower section of igneous cumulates and a complementary upper section of non-cumulates that crystallized from progressively more fractionated residual magma. Solidification proceeded upward from the base and downward from the top to a sandwich horizon of quartz-bearing differentiates which lie just beneath the top of the sequence (zone 2a). Chief evidence that this sequence is the product of a single magma is the nearly continuous progression of phase layering and cryptic variation throughout the composite plutonic section (Fig. 7). This does not rule out the possibility that this magma was periodically replenished by small draughts of new magma as it differentiated, but such replenishment will only be detected by a more detailed study of the cryptic variation of the cumulus phases. The hidden interval between zones 2b and 2c represents a loss of part of the section, including the transition from cumulate to non-cumulate gabbros. Mineralogically, however, not much appears to be missing (Fig. 7A).

There is no place within the plutonic sequence where rock having the composition of the primitive magma may be sampled. Zones 2c-3c are crystal cumulates from magma that had liquidus olivine, and zone 2a-2b crystallized from magma fractionated sufficiently that olivine was no longer a liquidus phase. Diabase at the top of the roof facies may have solidified from primitive magma but it is now badly altered as well as difficult to recognize within the plexus of sills at the base zone 1c.

The mineral paragenesis tells something about

Table 4. Chemical Analyses and CIPW Norms for the Upper Plutonic Rocks and Dike Series.

| | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| $SiO_2$ | 47.86 | 55.02 | 58.47 | 54.14 | 55.69 | 48.96 |
| $TiO_2$ | .19 | 1.25 | 1.07 | .71 | .81 | .25 |
| $Al_2O_3$ | 18.22 | 16.62 | 15.55 | 15.06 | 13.86 | 11.18 |
| $Fe_2O_3$ | 1.27 | 4.07 | 4.02 | 4.09 | 4.88 | .93 |
| FeO | 3.20 | 6.35 | 5.89 | 6.12 | 5.96 | 6.18 |
| MnO | .09 | .14 | .19 | .18 | .18 | .15 |
| MgO | 9.49 | 2.66 | 2.51 | 5.53 | 4.28 | 13.00 |
| CaO | 15.45 | 5.35 | 5.37 | 6.33 | 10.03 | 16.22 |
| $Na_2O$ | 1.13 | 5.56 | 4.09 | 3.82 | .30 | 1.00 |
| $K_2O$ | .64 | .71 | .53 | 1.18 | .04 | .03 |
| $H_2O^+$ | 2.20 | 1.69 | 1.45 | 2.26 | 3.60 | 1.53 |
| $H_2O^-$ | .20 | .44 | .38 | .42 | .22 | .18 |
| $P_2O_5$ | .06 | .14 | .12 | .11 | .12 | .05 |
| $CO_2$ | <.05 | <.05 | <.05 | <.05 | <.05 | <.05 |
| $Cr_2$ 3 | ... | ... | ... | ... | ... | .14 |
| Total | 100.00 | 100.00 | 99.64 | 99.95 | 99.97 | 99.80 |
| FeO*/MgO | .47 | 3.92 | 3.95 | 1.85 | 2.53 | .55 |
| Q | ... | 3.37 | 15.11 | 4.69 | 25.20 | ... |
| Or | 3.82 | 4.29 | 3.25 | 7.18 | 2.56 | .18 |
| Ab | 10.24 | 51.07 | 38.16 | 35.34 | 2.92 | 9.05 |
| An | 43.17 | 18.72 | 23.39 | 21.08 | 38.20 | 26.14 |
| Ne | ... | ... | ... | ... | ... | ... |
| Cor | ... | ... | ... | ... | ... | ... |
| Di | 27.06 | 6.00 | 2.79 | 8.43 | 11.85 | 43.73 |
| Hy | 5.16 | 10.12 | 11.14 | 17.62 | 12.26 | 8.22 |
| Ol | 8.81 | ... | ... | ... | ... | 11.24 |
| Mt | 1.34 | 4.35 | 4.37 | 4.40 | 5.52 | .98 |
| Ilm | .27 | 1.78 | 1.55 | 1.02 | 1.22 | .35 |
| Ap | .13 | .30 | .26 | .24 | .27 | .11 |

1. Subophitic uralitic cpx gabbro (PS56-1). Lowest exposure of ophiolite zone 2b, northwest of Point Sal State Park beach. Analysis by H.N. Elsheimer, USGS.
2. Hornblende diorite (PS59-1). Ophiolite zone 2a, 800 m east of tip of Point Sal. Analysis by H. N. Elsheimer, USGS.
3. Hornblende quartz diorite (PS61-6). Ophiolite zone 2a, 400 m east of tip of Point Sal. Analysis by H. N. Elsheimer, USGS.
4. Augite diabase (PS17-1). Upper Dike Series stage 3, from sill complex (zone 1c) at north end of Point Sal State Park beach. Analysis by H. N. Elsheimer, USGS.
5. Epidosite (PS62-2). Upper Dike Series stage 3, from sill complex (zone 1c) near tip of Point Sal. Analysis by H. N. Elsheimer, USGS.
6. Hb-opx-cpx microgabbro (PS20-4). Lower Dike Series in zone 3b, Lions Head Quarry. Analysis by H. N. Elsheimer, USGS.

composition of the original magma in terms of its position within the basalt tetrahedron (Irvine, 1970). The apparent paragenesis, based on the cumulate sequence in zones 2c-3c and the crystallization sequence in zones 2a-b, is: ol+chr, cpx+ol±chr, pl+cpx+ol, hb(opx)+pl+cpx±ol, mt+hb+pl+cpx, qz+mt+hb+pl, qz+ab. Mid-ocean ridge basalt (MORB) magmas normally have a paragenesis in which cpx appears fourth, after ol, chr, pl. At Point Sal, however, cpx apparently precedes pl, which would indicate a slight but significant difference in composition from modern primitive MORB magmas. An alternative explanation that we favor, however, is that plagioclase preceded clinopyroxene in the crystallization sequence but followed it in the cumulate sequence, due to the similar densities of calcic plagioclase and basaltic melt (Bottinga and Weill, 1970). For example, Bryan and Moore (1977) demonstrate that plagioclase floated for a time in primitive MORB in the FAMOUS area. If this occurred at Point Sal then the settling (cumulate) sequence would accord with the crystal settling sequence predicted for the primitive MORB magma (Bryan and Moore, 1977, Fig. 16).

It has been suggested (Hodges, 1976) that the Point Sal magma had alkaline affinities, based on the absence of hypersthene from the paragenetic sequence. We disagree. The subalkaline character of this magma is indicated by its fractionation to quartz-rich derivatives, and by the low $TiO_2$ content and hypersthene-normative character of its differentiation products (Tables 3, 4. Note: four rocks have <1 percent normative Ne due to late hornblende). It fractionated along a tholeiitic trend (Hopson and others, 1975a, Fig. 9), and its LREE-depleted pattern (Menzies and others, 1977b) places it with the Group 1 mid-ocean ridge tholeiites (Bryan and others, 1976). In fact, orthopyroxene locally does appear as the fifth phase in the paragenetic sequence, but its place is generally taken instead by pale brown hornblende. This simply indicates that the magma had become quite hydrous by this stage of crystallization, probably from the entry and mixing in of sea water. This inference is supported by the Sr-isotopic results of Davis and Lass (1975).

In summary, the Point Sal plutonic sequence crystallized and differentiated in place from olivine-tholeiite magma, probably quite similar to modern Type 1 MORB magma. Repeated deformation accompanied the accumulation, post-cumulus crystallization, and subsolidus cooling of the cumulate sequence, indicating a tectonically active environment. An ocean-ridge setting is inferred.

### The Dike Sequences

Dikes and sills within the Point Sal ophiolite are readily assigned to two groups that differ from one another in stratigraphic position, petrography, and relative age. They are designated the upper and lower dike series. The upper dike series, found within zones 1b-2b, is a composite set of intrusives that spans a long time range. The lower dike series, cutting zones 2c-3c, are the youngest igneous rocks exposed.

Upper Dike Series. These rocks form an intrusive complex of sills and dikes that cluster between the volcanic and plutonic sequences (zone 1c, Figs. 2, 3). Dikes also penetrate up into the volcanic pile (zone 1b), and others cut the underlying quartz diorite and uralitic gabbro (zones 2a-b). Thus, the upper dike series occupies a stratigraphic position similar to the sheeted dike complexes of the Troodos, Semail, and other ophiolites. There are critical differences, however, as described below.

Description of the complex is facilitated by reference to Figure 8, which shows the distribution and orientation of these intrusives. Points plotted on the stereograms represent the poles of individual dikes and sills, and the points in each plot have been rotated by amounts necessary to bring the upper surface of the ophiolite back approximately to its original horizontal position. The following description begins with the dikes cutting the lavas and proceeds down-section to those cutting the upper plutonic rocks.

Dikes cutting the lower volcanics (locations A, B, C, Fig. 3; stereograms A, B, C, Fig. 8) were mainly pyroxene basalt and diabase, now extensively altered to greenschist-facies mineralogy. One group of these dikes are subvertical and trend mainly NNE; others branch off at lower angles. These dikes are crosscut by northeast-dipping basic dikes, representing a second group described below. The steep, NNE-trending dike set is older than the underlying sill complex (zone 1c), which cuts it off. The relative age of the NNE-trending dikes and their compositional similarity to the lavas they cut (aphyric and cpx-pl microphyric basalts) suggests that they are feeders for the lower volcanic sequence (zone 1b).

Separating the volcanic and plutonic rocks is a remarkable concentration of sills and low-angle dikes (zone 1c, Fig. 3, 8). The best exposures are at the north end of Point Sal State Park beach (locations F, G, Fig. 3). Here the repeated intrusion of sills and low-angle dikes alongside and across one another has so densely clustered hundreds of these bodies within a zone up to 100 meters wide that no septa of volcanic or gabbroic host rock remain between them. Similar concentrations of these sheeted sills are poorly exposed along the south side of Point Sal Ridge, where they angle downslope to meet the shoreline again near the tip of Point Sal (locations D, E, Fig. 3). These sheeted intrusives resemble the vertical sheeted dikes of other ophiolites, but the sheeting is parallel to the volcanic-plutonic contact instead of perpendicular to it. Moreover, when the stereoplots of dike attitudes are rotated to bring the upper surface of the ophiolite (i.e., the Jurassic sediment-basalt contact) back to a horizontal position, the dikes are subhorizontal (Fig. 8, plots D, E, F, G). Thus, these sheeted intrusives (zone 1c) are a sill complex.

The steep NNE-trending volcanic feeder dikes within zone 1b do not cut across the sill complex; rather, they terminate at its upper surface and form isolated, truncated remnants within it (best seen near location D, Figs. 3, 8). This demonstrates the younger age of the sill complex.

Rocks comprising the sill complex (zone 1c) are of three general types: (1) metabasaltic and diabasic rocks, extensively altered to greenschist-facies mineral assemblages; phenocrystic phases were plagioclase (now albitized or saussuritized), clinopyroxene (fresh to uralitic), and some olivine (pseudomorphed by chlorite); (2) hornblende microdiorite and microtonalite, variously altered; and (3) light green dikes of epidosite (chiefly chlorite-epidote-quartz), commonly grading to dark gray microdiorite at the margins. Where relative ages within the sill complex can be determined by local cross-cutting relations the type 1 basic dikes tend to be earlier than the type 2 and 3 dikes of intermediate composition. Youngest of all, however, are a few very thin metabasaltic dikes with olivine pseudomorphed by chlorite (type 4).

The diorites and gabbros beneath the sill complex are also cut by dikes, which are locally abundant. Those cutting the diorites (zone 2a, location

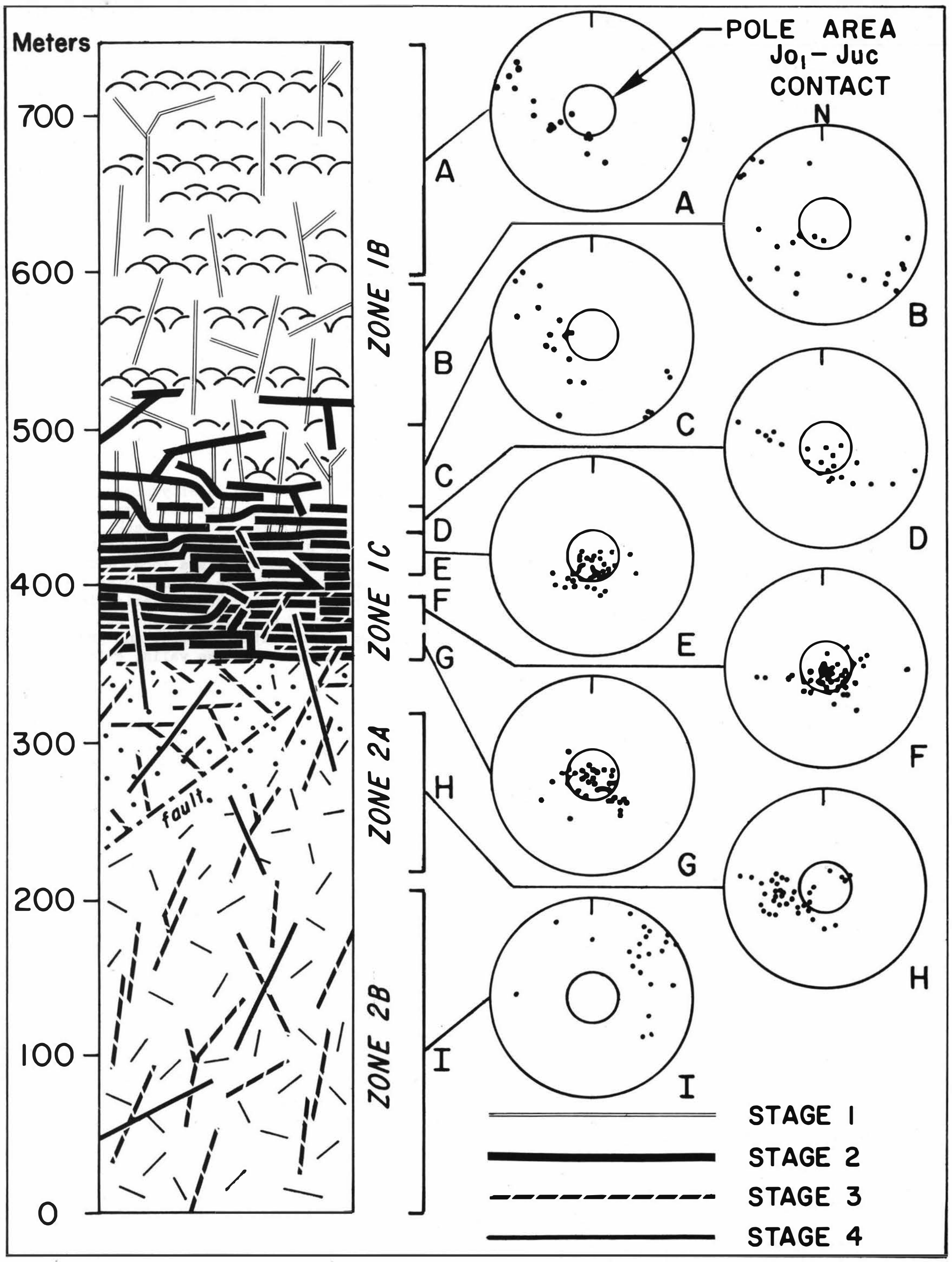


Figure 8. Columnar diagram and stereographic plots, showing dike and sill attitudes and age relations for the Upper Dike Series. Letters relate each stereoplot to the part of the column and to the geographic location (Fig. 3, lettered areas) where the measurements were made. Plotted points are the poles to dikes and sills. The stereoplots have been rotated to bring the upper surface of the ophiolite ($Jo_1$-Juc contact) back to a horizontal position. The small circle within each plot denotes a $\pm 20^{\circ}$ uncertainty in the $Jo_1$-Juc pole position. Stages of dike intrusion (1, oldest; 4, youngest) determined from field relations.

Table 5. Intrusion Stages, Upper Dike Series.

| Stage | Rock Types | Occurrence |
|---|---|---|
| 1 | Cpx-pl basalt and diabase | Zone 1b |
| | Keratophyre | Zone 1b |
| 2 | (Ol)-cpx-pl basalt and diabase | Zone 1c, 1b |
| 3 | Cpx-pl diabase | Zone 2a, 2b, 1c |
| | Hb microdiorite | Zone 2a, 2b, 1c |
| | Hb microtonalite | Zone 2a, 2b, 1c |
| | Epidosite | Zone 2a, 2b, 1c |
| | Albitite | Zone 2b |
| 4 | (Ol)-cpx-pl basalt | Zone 2b, 1c |

Tabulation of dikes and sills within the Upper Dike Series, relating rock types to stage of intrusion and to spatial occurrence within the ophiolite. Most of the rock types listed show partial to complete alteration to deuteric or hydrothermal mineral assemblages, commonly of greenschist facies.

H, Fig. 3; plot H, Fig. 8) dip mainly to the east and northeast. Those cutting the upper gabbros (zone 2b, location I, Fig. 3; plot I, Fig. 8) dip mainly to the southwest and west. The dikes cutting both these zones are similar to types 2, 3, and 4 in the overlying sill complex; in addition, there are a few thin albitite dikes.

Table 5 summarizes all the dike types within the upper dike series, their age relations, and the zones in which they are found. The following events and correlations are inferred.

Stage I: Intrusion of subvertical NNE-trending dikes (with low-angle branches) that fed the lower lava sequence (zone 1b). These dikes of relatively fractionated (i.e., non-olivinic) basalt and diabase predate the sill complex and the underlying plutonic sequence. Their source is not exposed.

Stage 2: Repeated injection of thin sheets of basaltic magma along a horizontal zone of weakness between the lower volcanics (zone 1b) and its original substratum, which is not now exposed. These basaltic and diabasic sheets built up the main part of the sill complex (zone 1c) and also form low-angle dikes cutting up into volcanic zone 1b. At least some of these basic intrusives had liquidus olivine (now chlorite pseudomorphs), i.e., they represent relatively unfractionated magma. They are correlated with the olivine gabbro magma that subsequently differentiated to form the underlying plutonic sequence (zones 2-3). Thus, emplacement of the sill complex represents an early stage in the emplacement of the underlying pluton. This is discussed in a later section.

Stage 3: Dike intrusion of pyroxene diabase, hornblende microdiorite, microtonalite, and aplitic plagiogranite. Some of these dike rocks were altered metasomatically (e.g., microdiorite to epidosite; aplite to albitite) by gas streaming through the interiors of dikes as they solidified inward from their walls. Stage 3 dike emplacement occurred late, after the underlying plutonic sequence had solidified, for these dikes cut across its differentiated upper zones. Some of these dikes also rose up into zone 1c and spread out within it as sills, adding to the thickness of the sill complex. The compositions of the stage 3 dikes correspond to the highly differentiated upper parts of the plutonic sequence (zones 2a-b). Therefore, they are thought to have intruded from the side, from an adjacent part of the upper magma chamber while it was still crystallizing and differentiating through the pyroxene gabbro, diorite, and tonalite stages. The strong deuteric and metasomatic alteration within these dikes reflects the high water content of this residual magma.

Stage 4: Intrusion of a few very thin basaltic dikes that cut everything else. These dikes contained olivine (pseudomorphed now by chlorite). Perhaps they represent the extreme distal ends of dikes carrying relatively unfractionated magma from a distant source.

Lower Dike Series. Swarms of basic dikes cut the lower plutonic section (Fig. 2). These dikes are large and abundant within the dunite (zone 3c), smaller and fewer within the pyroxenites (zone 3a), and rare in the lower gabbro (zone 2c-d). The dikes cut across the cumulus layering at low angles and follow it as sills; also, the basic dike rock forms net veins in brecciated dunite. The dike rocks are altered to rodingite where the host rocks are strongly serpentinized.

The dike rocks are chiefly microgabbros that exhibit a considerable range of composition, from olivine-rich varieties to those lacking olivine but including two pyroxenes plus abundant brown hornblende. Dark melanorite is a local mafic variant. Hornblende micro-tonalite and one dike of albite granite are minor siliceous varieties in zone 2c. Isotropic fabrics are most common, but some dike rocks have protoclastic textures and gneissic banding that parallels the dike walls, evidencing mush flowage.

The presence of abundant orthopyroxene (bronzite) distinguishes basic rocks of the lower dike series from rocks of the upper dike series and the main plutonic sequence, where diopside or augite is generally the only pyroxene. Thus, the lower dike series evidently represents a different batch of magma that was injected into the cumulate sequence, either from the side or from below, after the cumulates had solidified. This magma differed by having a more hypersthene-normative composition, or by having a lower activity of water so that amphibole did not displace orthopyroxene from the paragenetic sequence.

Another, genetically different type of dike occurs within lower plutonic zones 2c-3a. These are feldspathic ol-clinopyroxenite and picrite dikes that originate within the ultramafic cumulates and cut up into the gabbros, where they locally form spectacular intrusion breccias. Petrographically these feldspathic ultramafic dikes are difficult to distinguish from mesocumulus ol-clinopyroxenites of the layered sequence itself. They appear to have been emplaced as hot crystal mushes, derived from the mobilization of cpx-ol cumulates that contained feldspathic intercumulus liquid. Local deformation of the stratiform sequence, at a stage when the adcumulus layers had completely solidified but the mesocumulus layers still retained interstitial melt, evidently resulted in the mobilization and intrusion of the mesocumulate liquid-crystal mushes.

## CORRELATION AND TIMING OF IGNEOUS EVENTS

The Point Sal igneous sequence apparently records two periods of volcanism, one prolonged period of plutonic crystallization and differentiation, and several periods of dike emplacement which overlap the volcanic and plutonic stages (Fig. 9). Here we

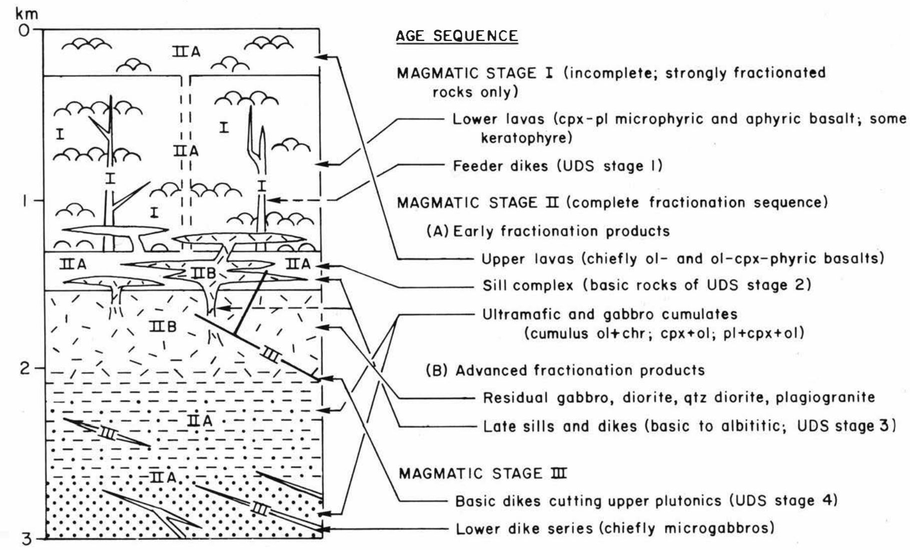


Figure 9. Diagram showing age relationships within the Point Sal ophiolite. Magmatic stages are those listed in Table 6. Upper Dike Series (UDS) stages are those listed in Table 5.

TABLE 6. CORRELATION OF IGNEOUS EVENTS

| VOLCANIC SEQUENCES | DIKE SEQUENCES | PLUTONIC SEQUENCES |
|---|---|---|
| I. Lower basalts<br>cpx+pl<br>pl | I. UDS stage 1 dikes<br>cpx+pl<br>pl | I. Not exposed |
| II. Upper basalts<br>ol<br>ol+cpx<br>ol+cpx+pl<br>cpx+pl | IIA. UDS stage 2 sills<br>ol+cpx±pl<br>cpx+pl | IIA. Dunite accumulation stage<br>ol+chr (primitive magma)<br>Pyroxenite accumulation stage<br>cpx+ol (+ floating pl?)<br>Ol-gabbro accumulation stage<br>pl+cpx+ol<br>Cpx-gabbro accumulation stage<br>pl+cpx |
| | IIB. UDS stage 3 dikes<br>pl+cpx±hb<br>pl+hb<br>pl+hb+qz<br>ab+qz | IIB. Uralite gabbro xtlzn. stage<br>cpx+pl+hb<br>Diorite xtlzn. stage<br>pl+hb+mt<br>Quartz diorite xtlzn. stage<br>pl+hb+mt+qz<br>Albite granite xtlzn. stage<br>ab+qz |
| | IIC. UDS stage 4 dikes (=III?)<br>ol+pl+cpx | III. Lower dike series<br>ol+pl+cpx±opx<br>pl+cpx+opx±hb(brn)<br>pl+cpx+hb(brn)<br>pl+hb+qz<br>ab+qz |

Volcanic, hypabyssal, and plutonic events arranged from top to bottom in order of decreasing age; correlative events are aligned left to right. Main stages are shown by Roman numerals; substages by letters. UDS (Upper Dike Series) stages are those shown in Figure 8. Mineral abbreviations refer to magma liquidus phases.

attempt to correlate these events and to develop an integrated igneous history. This will prove critical to understanding the sheeted sill complex and also the tectonic development of the ophiolite, taken up in the next two sections.

Discussion is facilitated by reference to Table 6, which arranges the volcanic, hypabyssal, and plutonic sequences each in order of decreasing age, and places them opposite one another to show inferred time equivalence. Two volcanic stages are distinguished on the basis of stratigraphic relations, distinctive lava compositions (Fig. 4), and the contrasting interpillow sediments. The upper dike series (labeled UDS in Table 6) is divided into stages and substages that correspond to successive periods of dike and sill emplacement described in the previous section (Table 5). Only one plutonic stage is recognized but it forms a continuous sequence of differentiation that can be subdivided into successive steps. Each step has a different set of liquidus phases that were crystallizing, as the melt became progressively more fractionated (see Fig. 7, and discussion of the plutonic paragenetic sequence). For comparison, the early liquidus (chiefly phenocrystic) phases are also shown for the volcanic and dike sequences. Thus, correlations between successive volcanic and diking events, and differentiation steps within the magma chamber (plutonic sequence), are based partly on cross-cutting relations in the field and partly from matching the early liquidus phase assemblages, which provide a rough index of fractionation.

Stage I. The oldest rocks exposed are the lower lavas (zone 1b) and their NNE-trending feeder dikes (UDS stage 1). Both these groups are cut off by the underlying sill complex, which therefore is younger (Figs. 8, 9). The lower lavas also predate the main plutonic sequence, for they formed the solid roof beneath which the latter crystallized. The stage I lavas and dikes crystallized from basic magma that had already reached an advanced stage of fractionation: olivine was not a liquidus phase in these melts, and some keratophyric flows were moderately quartz-normative. Thus, the lower lavas and their feeder dikes cannot be correlated with the exposed plutonic sequence, where olivine was a liquidus phase for the first two-thirds of its crystallization history (Fig. 7). The fractionated stage I lavas and dikes have no plutonic equivalents exposed.

Stage IIA. Next within the dike sequences are the UDS stage 2 metabasaltic and diabasic rocks of the sill complex (Table 5; Figs. 8, 9). Phenocrystic olivine (chlorite pseudomorphs), clinopyroxene, and plagioclase indicate that these minerals were liquidus phases in the magma at the time when early stages of sill emplacement occurred. This permits correlation of these sills with early stages in the differentiation of the plutonic complex (liquidus ol, cpx+ol, pl+cpx+ol) and also with the upper basalts of zone 1a (phenocrystic ol; microphenocrystic cpx, pl). No feeder dikes directly connecting with the upper olivine basalts have been observed, however.

Stage IIB. Dikes belonging to UDS stage 3 (Table 5; Figs. 8, 9) cut across the differentiated upper levels of the plutonic sequence (zones 2a-b) and spread out within the sill complex (zone 1c). Although these dikes and sills post-date solidification of the upper part of the exposed plutonic sequence they crystallized from highly fractionated melts that correspond to the zone 2a-b residual magma. Liquidus phases in these dikes and in successive late plutonic magma batches were cpx+pl+green hb, pl+green hb+mt, pl+hb+mt+qz, and ab+qz (Table 6). Thus, these dikes were probably intruded sideways from a lateral extension of the upper magma chamber, where solidification was a step behind that of the exposed sequence. This assumes a spreading ocean ridge-type magma chamber that is fed from the middle and solidifies continuously at the sides.

Stage IIC. Intrusion of a few very thin olivine basalt dikes which cut across all other rocks within zones 1c, 2a, 2b. This event may correspond to Stage III.

Stage III. The lower dike series, chiefly olivine- and orthopyroxene-bearing microgabbros, are later than Stage IIA, for they cut sharply across the olivine cumulates (zones 2c-3c). Furthermore, they probably are not contemporaneous with the upper plutonic sequence, Stage IIB, because their composition is distinctly more basic (compare analyses 4, 5, 6, Table 4). Thus, the lower dike series represents a still younger intrusive event, possibly correlative with Stage IIC (Table 6). Yet, these dikes are probably not much younger than Stages IIA-B, for the development of protoclastic structure within some of them is associated with the high temperature internal disruption of the cumulate sequence.

In summary, three main igneous events are discerned within the Point Sal ophiolite (Fig. 9). Stage I is the extrusion of strongly fractionated tholeiitic magma onto the sea floor in an environment where red radiolarian ooze (chert) accumulated between eruptions. Stage II marks the extrusion of relatively unfractionated olivine tholeiite magma onto the sea floor in an environment where gray calcareous ooze (coccolithic limestone) is the main interpillow sediment. The sharp contrast both in sediment type and magma composition suggests that events I and II were separated in time or in geographic location or both. Stage II also marks the intrusion of olivine tholeiite magma, forming sheeted sills beneath the Stage I volcanic crust and emplacing a magma body more than 1.4 km thick beneath the sills. Substage IIA includes growth of the main sill complex, and early differentiation within the magma chamber and the buildup of cumulates at its lower and middle levels. Substage IIB includes the late stages of differentiation, non-cumulus crystallization and retrograde boiling within the upper magma chamber. Sills of highly fractionated, water-rich melt were injected up into the sill complex during this period, and similar fractionated dikes were then intruded into the top of the newly solidified plutonic sequence from the side. Substage IIC marks the final intrusion of thin basaltic dikes through the top of the plutonic sequence. Stage III (= IIC?) marks the intrusion of swarms of chiefly basic low-angle dikes and sills across the cumulus sequence, probably while the latter was still cooling. These evidently came from a separate batch of orthopyroxene-rich magma, which fractionated from olivine-normative to quartz-normative compositions during the span of dike intrusion.

Igneous Stages I and II may be widely separated in time, for there is a striking break here both in the fractionation sequence and in the type of sedimentation that accompanied the volcanic eruptions. Igneous Stages II and III, however, were apparently close together and may be genetically related.

## EMPLACEMENT OF THE SILL COMPLEX

Vertical sheeted dike complexes separate the volcanic and plutonic zones of ophiolites in Cyprus, Oman, Newfoundland and elsewhere, but horizontal sheeted sill complexes are less well known. The Point Sal sill complex is not unique, however; similar sill complexes occur locally within other sections of the Jurassic Coast Range ophiolite in California; for example, near San Luis Obispo (Page, 1972; Pike, 1975), Del Puerto Canyon (Evarts, 1976), Mount St. Helena, and the South Fork of Elder Creek near Paskenta (personal observation). Sill complexes are also reported from ophiolites at Cedros Island, Baja California (Jones, Blake, and Rangin, 1976) and the Ergani district, Turkey (Bamba, 1974). Nor are they restricted only to oceanic rocks: sill complexes several hundred feet thick cap Tertiary epizonal plutons in the Cascade Mountains, particularly the Tatoosh pluton near Mount Rainier, Washington (Fiske and others, 1963).

Emplacement of the Tatoosh sill complex is instructive with respect to Point Sal. An early phase of the Tatoosh magma rose and spread out in repeated pulses along a stratigraphic zone of weakness between two volcanic formations, forming a plexus of sills and cross-cutting dikes concentrated along a zone hundreds of feet thick and extending laterally for several miles. Subsequently the main body of Tatoosh magma rose and spread out along the base of the sill complex, lifting the roof and swelling into a magma chamber several thousand feet thick (Hopson and others, 1970; Mattinson, 1977). The uplifted sill complex, reinjected and altered by the pluton, forms the immediate roof of the pluton in many places.

The Point Sal sill complex is comparable in many respects. Stage II basic magma invaded the base of an earlier (Stage I) volcanic crust (zone 1b), sending out sills along a horizontal zone of weakness. Subsequently a large body of Stage II basic magma crystallized and differentiated beneath the sills. The horizontal zone of weakness initially exploited by the sills must have had a floor as well as a roof at the time of emplacement; however, where the older floor should be we now find younger rocks (i.e., the differentiated plutonic rocks of Stage IIB). Thus, following initial emplacement of the sill complex more magma intruded between the sills and their floor, forming the large magma chamber filled by the present plutonic sequence (zones 2-3). The original floor to the sill complex should now be found beneath plutonic zone 3c; unfortunately, this level is not exposed.

Events associated with emplacement of the sill complex are reconstructed as follows. Oceanic crust formed during magmatic Stage I (Table 6; Figs. 9, 10a) consisted of a volcanic sequence (zone 1b) and a plutonic substratum (not now exposed). This crust subsequently hosted intrusions of magmatic Stage II. Tholeiitic olivine basalt magma rose from the mantle into the earlier oceanic crust, initially sending out a vanguard of sills between the volcanic and plutonic zones (Fig. 10b) but then continued to intrude at this level, splitting the volcanic layer and sills (zones 1b-c) from the floor and inflating the space between them into a floored magma chamber nearly 1.5 km thick (Fig. 10c). It seems likely that some of this magma also reached the surface, erupting onto the ocean floor (Fig. 10b, c). The Stage II lavas (volcanic zone 1a) have the right composition and stratigraphic position to correspond to such flows. Subsequently the magma chamber crystallized and differentiated (stages IIA-B), sending out later dikes and sills into the roof and sides of the chamber. A final, rather minor magmatic episode (Stage III) followed closely in time.

The formation of a sill complex in place of a vertical sheeted dike complex is not incompatible with development at an oceanic spreading center, but it seems to require a situation in which tensional fissuring of the upper crust cannot keep pace with the rapid upwelling of magma. Ascending magma rising into a relatively static roof will tend to form shallow intrusions that follow existing planes of weakness, especially stratification, analogous to intrusions on land. Experiments by Ramberg (1963) show that intrusions with low viscosity, rising to shallow levels, tend to spread out in horizontal sheets beneath a light cover. The initiation of a new spreading center within pre-existing oceanic crust, described in the next section, may thus be favorable for the development of sill complexes.

## TECTONIC SETTING

Jurassic oceanic crust represented by the Point Sal ophiolite formed in a spreading ocean ridge setting, far from land and well removed from any active volcanic arc. Evidence supporting these conclusions is summarized below.

(1) Pelagic interpillow sediments within the ophiolite are radiolarian chert (lower lava) and coccolithic limestone with traces of chert (upper lava). These were relatively pure biogenic oozes, uncontaminated by land-derived terrigenous clastic sediment or arc-derived tephra. They accumulated between submarine eruptions at water depths ranging from below to slightly above the calcium carbonate compensation depth (CCD), estimated to be near 2.7 km during much of the Jurassic (Bosellini and Winterer, 1975).

(2) Vesicles in the lava pillows become very sparse and tiny within the chilled outer pillow rims, characteristic of lava erupted into very deep water (Moore, 1965, 1970; Bryan, 1975).

(3) The lavas were mainly tholeiitic basalts with petrographic features comparable to mid-ocean ridge lavas. They range from ol-phyric to strongly fractionated cpx-pl microphyric and aphyric lavas like those recovered from many DSDP sites.

(4) The low $TiO_2$ and $P_2O_5$ contents are indicative of ocean ridge-type rather than oceanic island-type lavas (Fig. 5).

(5) The LREE-depleted character of the Point Sal lavas and plutonic rocks (Menzies and others, 1977a,b) identify them as Group I oceanic tholeiites of Bryan and others (1976). Group I tholeiites are characteristic of mid-ocean ridge basalts but are not known from aseismic ridges or oceanic islands.

(6) The plutonic rocks crystallized from exceptionally hydrous magma. This is manifested by the post-cumulus crystallization of hornblende in place of orthopyroxene in the cumulus sequence, by the extensive uralitization of clinopyroxene and primary crystallization of hornblende in the upper (non-cumulus) sequence, and by the retrograde boiling and subsequent filling of miarolytic cavities by vapor-phase amphibole near the top of the plutonic sequence. Strontium isotopic evidence by Davis and Lass (1976) indicates that sea water mixed with the magma, down even to the level of the cumulates. A rifting mid-ocean ridge is the most likely setting where large quantities of sea water might penetrate through the volcanic upper crust and pervade the magma chamber.

(7) The strong internal deformation, disruption, and mobilization of the layered cumulates, during post-cumulus crystallization at temperatures still above the solidus, point to magma chamber solidification in an environment of very active tectonism. Such effects might be expected during the buildup of magmatic cumulates beneath an actively rifting ridge at a diverging plate boundary (Hopson, 1975). They would not be expected in an aseismic oceanic setting, just as they are minor or lacking from most stratiform plutons on land.

Origin of the Point Sal ophiolite at a spreading ridge setting in the open ocean is thus indicated by evidence from the sedimentary, volcanic, and plutonic rocks. However, this ophiolite also appears to record the formation of oceanic crust in two separate stages, probably at different geographic locations. This is reasoned from the following evidence: (1) the igneous sequence records more than one cycle of magmatic differentiation; (2) the sheeted sill complex, which opened a new magmatic cycle, formed by the lateral spreading of rising magma beneath a cover of pre-existing oceanic crust; and (3) pelagic inter-pillow sediments within the volcanic sequence record two different environments of deposition. The lower (zone 1b) lavas, belonging to the first magmatic cycle, erupted below the calcium carbonate compensation depth (CCD) where only siliceous ooze was accumulating. Later, these early lavas formed the cover beneath which the sill complex developed at the start of the second magmatic cycle. During the second cycle the upper (zone 1a) lavas were erupted onto sea floor above the CCD, where calcareous ooze was accumulating.

Formation of the volcanic layer at a spreading ridge occurs within a brief span of time. Even at a slow spreading center such as the Mid-Atlantic Ridge the volcanic layer is chiefly formed within a few hundred thousand years (MacDonald, 1977). Thus, if the Point Sal lavas were erupted at a single spreading site it is difficult to understand how this volcanism could have occurred first below the CCD and then above it. It would seem necessary to postulate a rapid uplift of the ridge crest from below the CCD to above it while volcanism was still in progress. Such an event seems unlikely, for the crestal elevation of spreading ocean ridges is fairly constant, averaging near 2.7 km (Sclater, Anderson, and Bell, 1971).

The more plausible explanation is that the Point Sal volcanics were erupted first at one site, which lay below the CCD, and then at some other distant site which lay above it. It is well known that the CCD varies laterally. In the modern oceans the calcium carbonate compensation surface (CCS) has about 2 km of relief (Berger and Winterer, 1974). The CCS depth range is from about 3.5 to 5.5 km in the modern oceans but it was much shallower in the Jurassic (Bosellini and Winterer, 1975), when the Point Sal ophiolite formed. Thus, the Jurassic CCS could have bracketed the crestal depth (2.7 km) of spreading ocean ridges, rising above this depth at some locations and dropping beneath it at others. We conclude that the Point Sal ophiolite records an initial growth stage (Igneous Stage I) where volcanism occurred at water depths below the CCD, and a later growth stage (Igneous Stage II) at a different geographic location, where the CCS dipped beneath the level of the vents. Both sites, however, were at spreading ocean ridges, from the evidence previously cited.

Composite oceanic crust formed at two widely separated spreading-ridge sites might develop where an oceanic spreading center jumps from one location to another. This phenomenon is well known, especially in the eastern Pacific during the Cenozoic. For example, the Mathematician and Clipperton seamount chains between 0°-20°N are the old crest of the East Pacific Rise (EPR). This segment of the rise crest terminated at 5 mybp when the spreading center jumped 4° to the east, the present location of the EPR spreading axis (Sclater and others, 1971). Another example occurs in the disturbed magnetic zone between the Murray and Molokai fracture zones, where the spreading axis abruptly jumped 530 km eastwards at 51 mybp (Menard and Atwater, 1969) or at 40 mybp (Harrison and Sclater, 1972). Still another example is found between 5°S and 15°S, where spreading along the East Pacific Rise began only about 6.5 mybp. Here the ridge crest jumped from the Galapagos Rise (former EPR) approximately 900 km to the west to begin spreading in crust that was 13 m.y. old (Anderson and Sclater, 1972). Evidence of jumping ridge crests is also found in the South Pacific (Molnar and others, 1975).

In each example, a new spreading center developed within oceanic crust that had formed several million years earlier. In these cases the older igneous crust was invaded by younger magmatic rocks where these jumps took place. Two-stage igneous histories, perhaps similar to Point Sal, doubtless mark these crustal segments. Moreover, the pelagic sedimentation which accompanied the pre-jump volcanism probably differs also, since several million years of time and hundreds of kilometers in space separated the two volcanic stages in each case.

Anderson and Sclater (1972) deduce that spreading center jumps in the eastern Pacific (20°N-20°S) were due to changes in spreading directions caused by realignment of the Nazca (Farallon) and Pacific plates about a new pole of rotation. They suggest that such realignment of active spreading centers may have been caused by the breakup of the Farallon plate into smaller plates and/or the interaction of the East Pacific Rise with the North American continent. Similar conditions doubtless prevailed locally in the Mesozoic also, and the record of such events may be contained within ophiolites.

We conclude that the two-stage igneous history of the Point Sal ophiolite, and the contrasting pelagic depositional environments of sediments within the upper and lower lavas mark this as a remnant of composite oceanic crust formed at two geographically different ocean-ridge sites (Fig. 10). Jumping spreading centers may provide a logical explanation.

## CONCLUSIONS

(1) The Point Sal ophiolite is a remnant of Jurassic oceanic crust, nearly complete down to 3 km but truncated above the mantle boundary by faulting.

(2) The volcanic rocks are chiefly tholeiitic basalts of mid-ocean ridge type (MORB), whose chemistry and mineralogy are strongly modified by high-temperature alteration and by seafloor weathering.

(3) Two stages of volcanism are represented: an early stage of strongly fractionated cpx-pl basalts, erupted where radiolarian ooze was accumulating, and a later stage of olivine basalt, erupted where calcareous pelagic ooze was accumulating.

(4) The stratiform plutonic sequence records a single cycle of magmatic crystallization and differentiation of olivine tholeiite magma. This magma fed the olivine basalt of volcanic stage II before differentiation was far advanced.

(5) The ultramafic and gabbro cumulates show extensive internal disruption, and also remobilization as dikes. This records repeated tectonic disturbance within the magma chamber during the period of accumulation and post-cumulus crystallization.

(6) Extensive uralitization, retrograde boiling, and deposition of vapor-phase amphibole record exceptionally hydrous conditions in the upper part of the magma chamber during the final stages of crystallization. Leakage of seawater into the chamber is inferred.

(7) The upper dike series was emplaced in four widely spaced stages, overlapping the volcanic and plutonic events. The lower dike series was intruded only at a late stage, after the cumulates had solidified.

(8) The sheeted sill complex, between the volcanic and plutonic zones, formed where a rising body of new magma was first emplaced into older oceanic crust. Continued influx of new magma and its lateral spreading along this horizon split the sills and overlying

volcanic layer from their substratum, forming the magma chamber now occupied by the differentiated plutonic sequence.

(9) The ophiolite represents composite oceanic crust that formed in two stages from separate magmas. Stage I was the initial formation of oceanic crust, represented by the fractionated lower lava sequence and its NNE-trending feeder dikes. Stage II began with the intrusion of unfractionated olivine tholeiite magma into the older oceanic crust, forming the sill complex and then the pluton beneath. Extrusion of some of this magma formed the upper olivine basalts. Differentiation of this magma and the later stages of dike and sill intrusion closed out stage II.

(10) The two stages of ocean crust formation both occurred in open ocean environments, beyond the reach of terrigenous sedimentation or arc-derived tephra. Two different geographic locations are inferred from the contrasting character of the inter-pillow pelagic sediments within the two lava groups. Spreading ocean ridge settings are inferred from the petrology and geochemistry of the lavas, and from evidence that the plutonic rocks crystallized in an environment of strong tectonic disturbance.

(11) The two-stage development of this composite oceanic crust may have resulted from the jumping of a spreading center from its initial location to a new site within older oceanic crust.

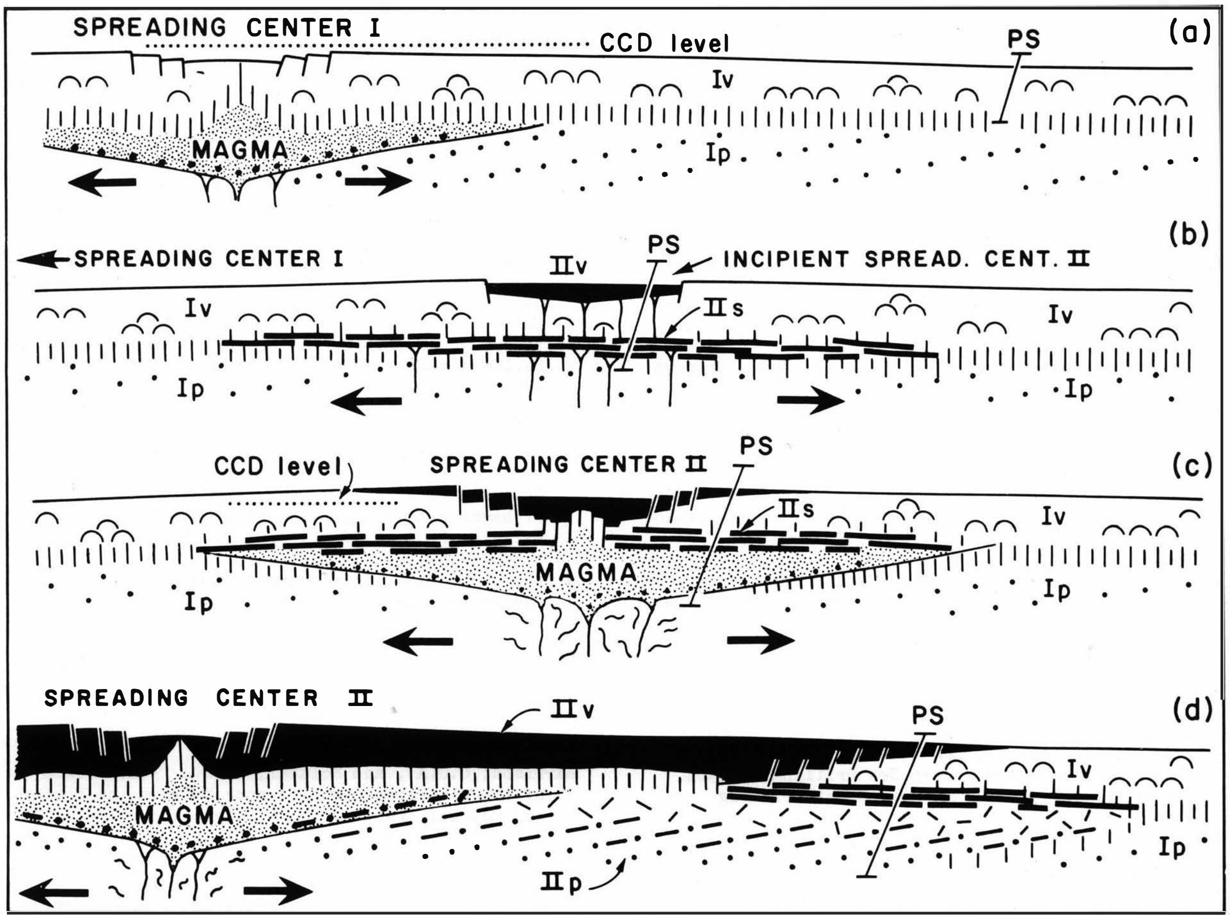


Figure 10. Diagram illustrating two stages of ocean crust formation and emplacement of the sheeted sill complex at Point Sal.

a. Stage I oceanic crust, formed at an initial spreading center. Iv, Ip: stage I volcanic and plutonic zones. PS signifies position of Point Sal section. Magma chamber shape adopted from Greenbaum, 1972.

b. Stage II (early): incipient development of new spreading center amid older (stage I) oceanic crust, by ridge-crest jumping. New, relatively unfractionated magma rises through fissures in rifting oceanic mantle and crust. Some magma intrudes laterally to form sheeted sills at base of volcanic layer Iv, and some is extruded at the surface. IIv, IIs: stage II volcanic rocks and sheeted sills.

c. Stage II (later): new magma continues to rise and to spread laterally along weak horizon beneath the sill complex, lifting the roof to form a floored magma chamber between zones Iv + IIs and zone Ip. Gravitative differentiation begins within the chamber. Magma continues to be extruded along the new rift zone, thickening the upper lava (IIv) as spreading proceeds.

d. Continued spreading and growth of oceanic crust at the new spreading center (II). Point Sal section (PS) occurs within the interval of overlapping old (I) and new (II) oceanic crust.

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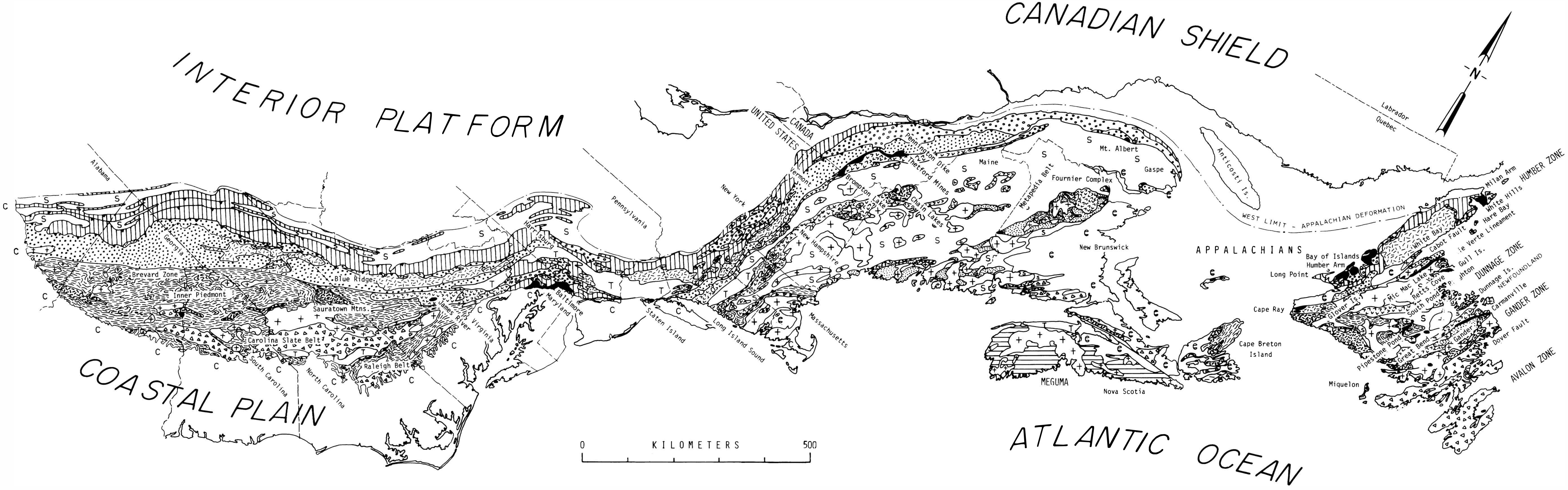



LEGEND (Figure 1)

- Carboniferous cover rocks, relatively undeformed.
- T — Triassic sedimentary and volcanic rocks with associated mafic intrusions.
- C — Mesozoic and younger cover rocks.
- S — Middle Ordovician to Devonian, mainly clastic sedimentary rocks deposited upon Ordovician and older rocks (includes Carboniferous rocks in Southern Appalachians).
- Mainly granitic intrusions

HUMBER ZONE

- Transported complexes, mainly ophiolitic.
- Taconic allochthons, mainly Cambrian to Middle Ordovician sedimentary rocks.
- Cambrian to Middle Ordovician carbonate sequence and overlying southeasterly-derived clastic sedimentary rocks.
- Late Precambrian to Cambrian mainly clastic sedimentary rocks and associated rift-facies volcanic rocks, mostly in amphibolite facies.
- Grenvillian inliers, commonly retrograded and deformed by Paleozoic orogenesis.

DUNNAGE ZONE

- Cambrian to Middle Ordovician marine facies volcanic rocks, slates, cherts, greywackes and mélange.
- Mainly ophiolite complexes.
- Amphibolite facies metamorphic rocks, mainly of unknown age and affinity.

GANDER ZONE

- Lower Ordovician and older clastic sedimentary rocks.
- Gneisses and migmatites, locally dated at ~ 1000 m.y. includes overlying marbles and quartzites.

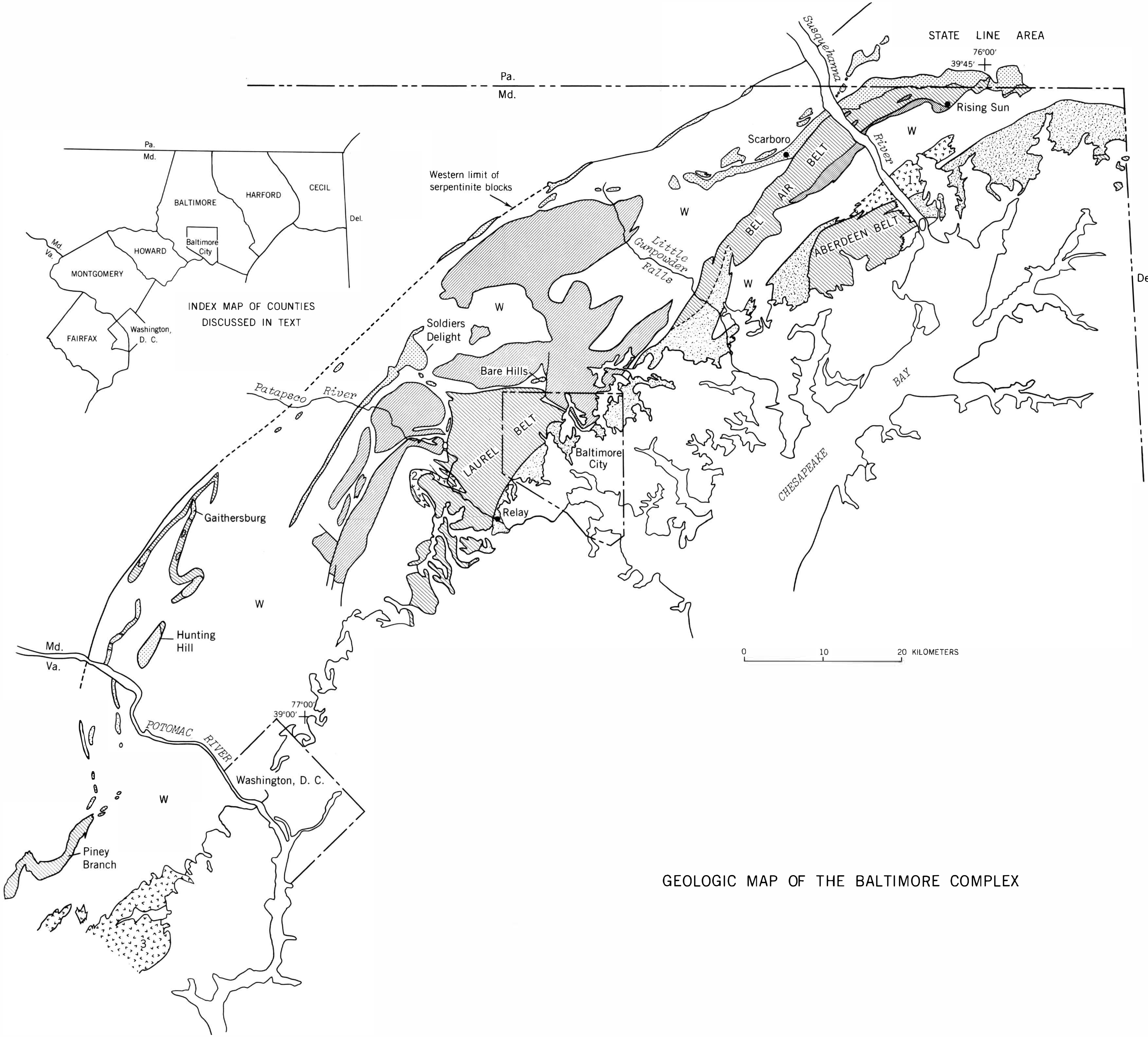
AVALON ZONE

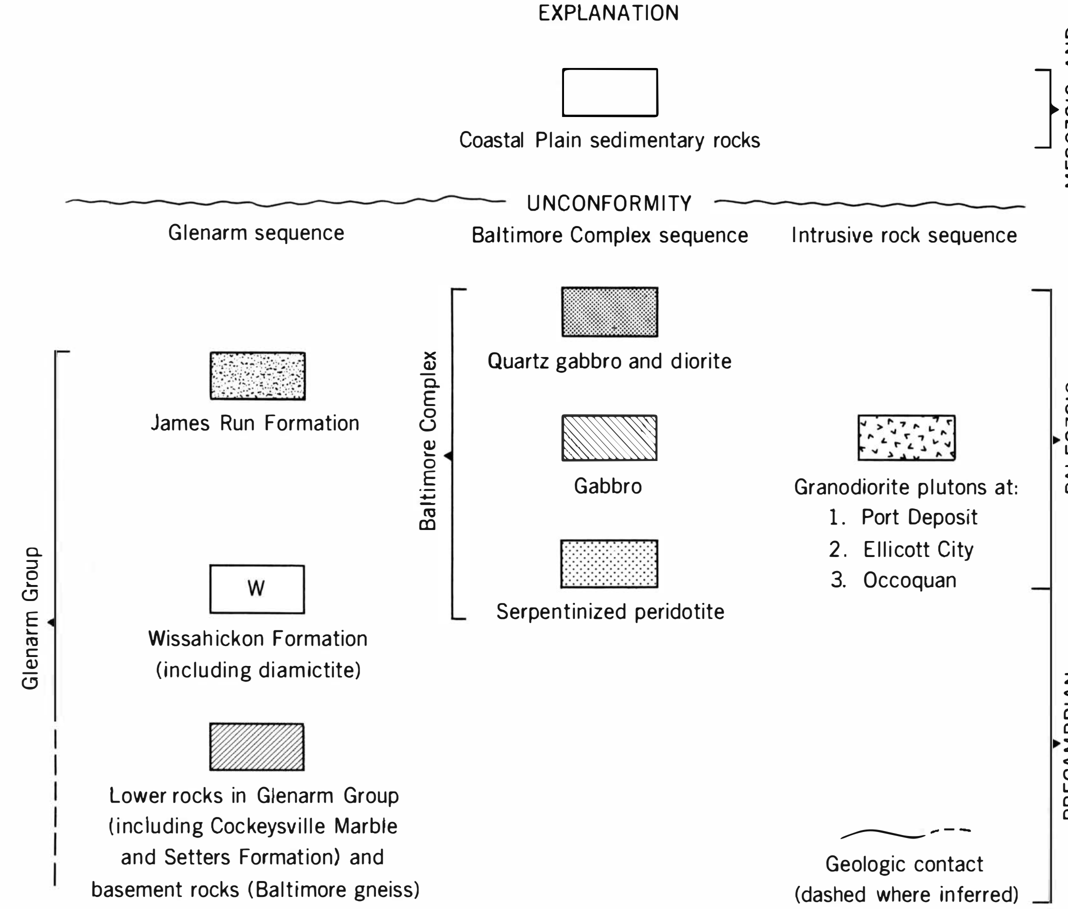
- Cambrian to Ordovician shales with Atlantic trilobite faunas
- Mainly Late Precambrian sedimentary and volcanic rocks.

MEGUMA ZONE

- Lower Ordovician and older greywackes and slates

PLATE I




GEOLOGIC MAP OF THE BALTIMORE COMPLEX

Compiled by B. A. Morgan, U.S. Geological Survey, 1977

**PLATE II**

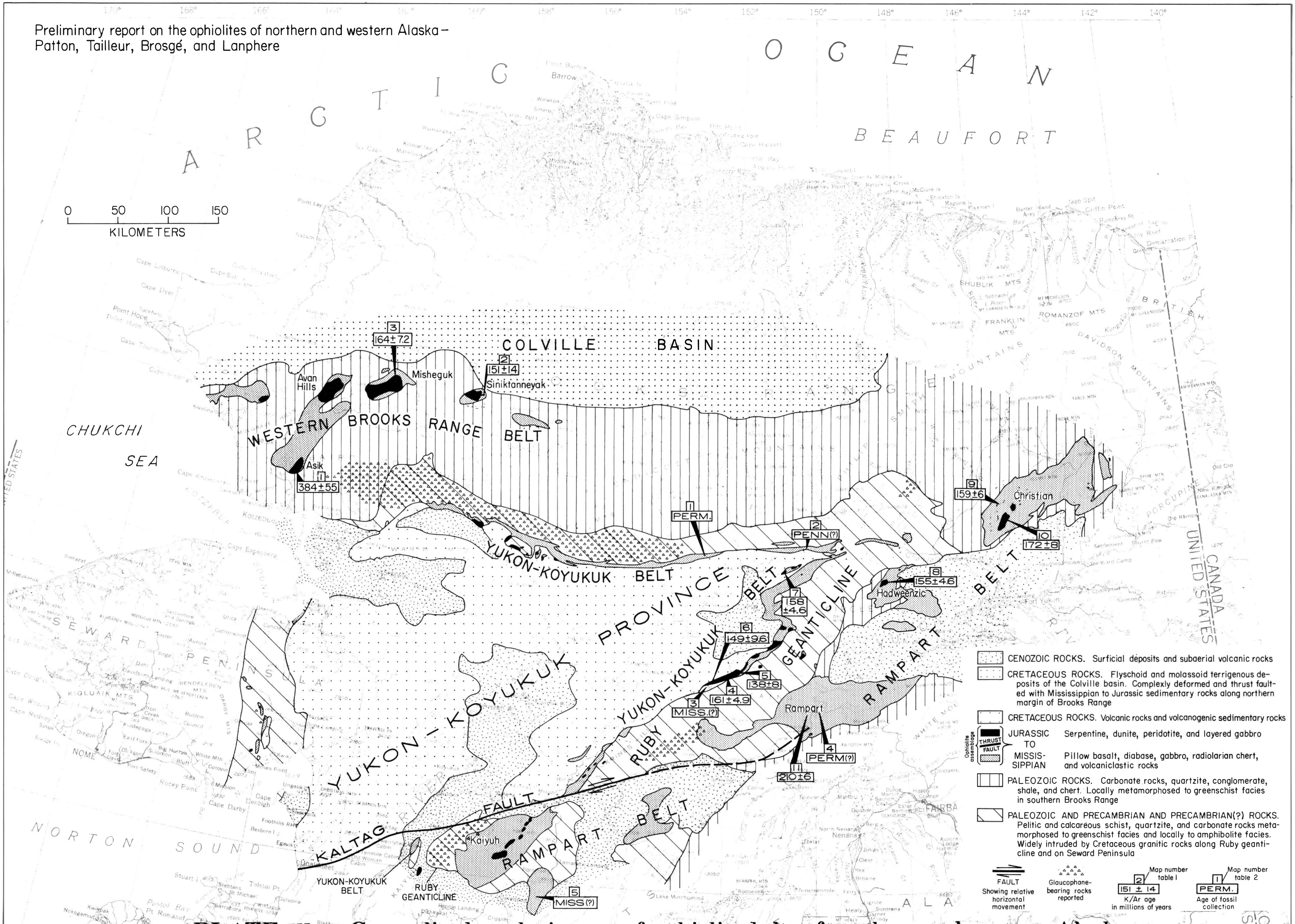



**PLATE III Generalized geologic map of ophiolite belts of northern and western Alaska**


GEOLOGIC MAP OF THE CANYON MOUNTAIN COMPLEX AND RELATED ROCKS, GRANT COUNTY, OREGON

Geology by T. P. Thayer, C. E. Brown, P. W. Guild, and A. B. Griggs, U.S. Geological Survey, 1977

**PLATE IV**

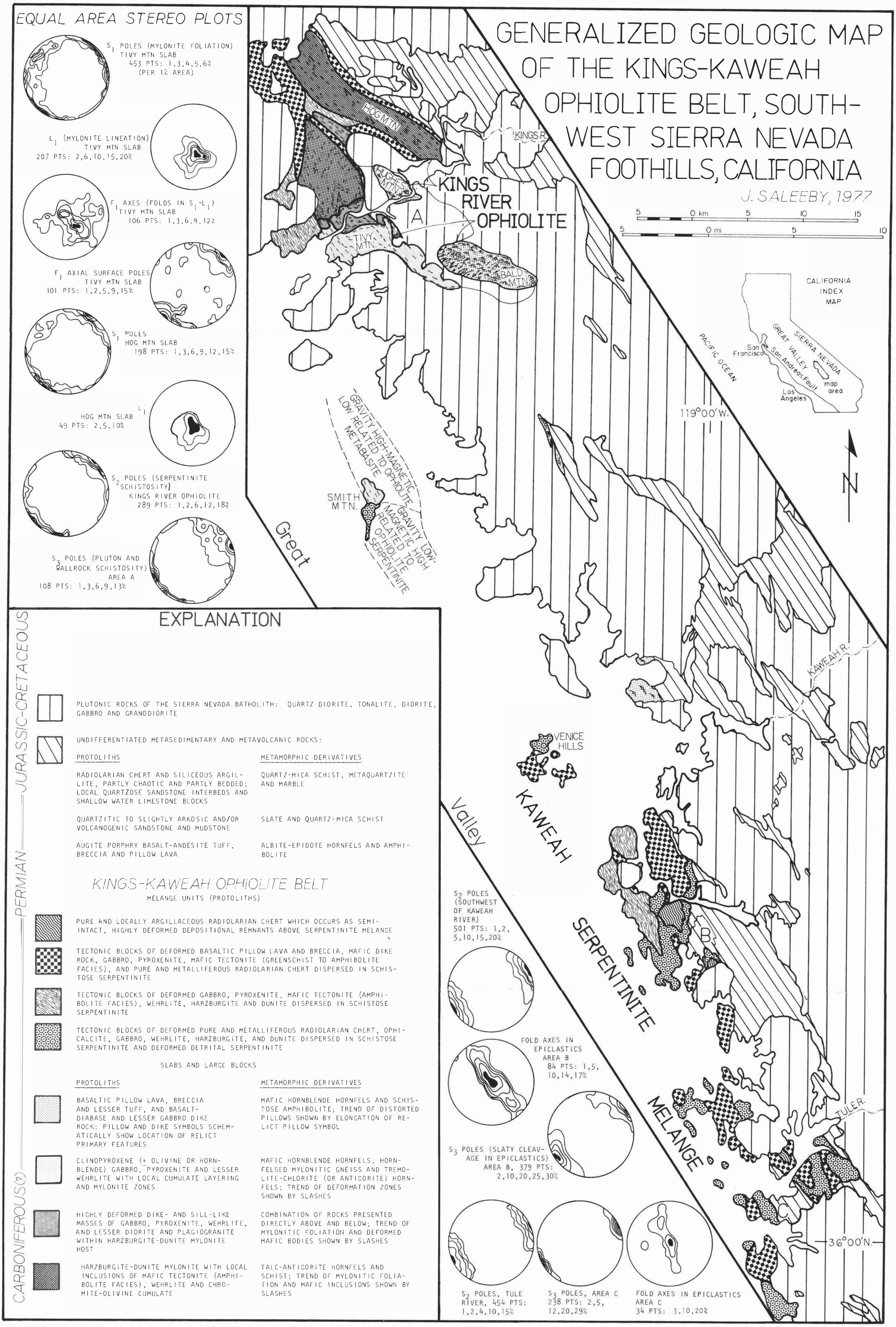

GENERALIZED GEOLOGIC MAP OF THE KINGS-KAWEAH OPHIOLITE BELT, SOUTHWEST SIERRA NEVADA FOOTHILLS, CALIFORNIA
J. SALEEBY, 1977
5 0 km 5 10 15
5 0 mi 5 10
CALIFORNIA INDEX MAP
SIERRA NEVADA
GREAT VALLEY
San Francisco
San Andreas Fault
map area
Los Angeles
PACIFIC OCEAN
119°00'W.
N
KINGS R.
HOG MTN.
HUGHES MTN.
KINGS RIVER OPHIOLITE
A
TIVY MTN.
BALD MTN.
GRAVITY HIGH-MAGNETIC LOW RELATED TO OPHIOLITE METABASITE
GRAVITY LOW-MAGNETIC HIGH RELATED TO OPHIOLITE SERPENTINITE
SMITH MTN.
Great
Valley
KAWEAH R.
VENICE HILLS
KAWEAH
SERPENTINITE
MELANGE
B
C
TULE R.
36°00'N.
EQUAL AREA STEREO PLOTS
S1 POLES (MYLONITE FOLIATION) TIVY MTN SLAB 453 PTS: 1,3,4,5,6% (PER 1% AREA)
L1 (MYLONITE LINEATION) TIVY MTN SLAB 207 PTS: 2,6,10,15,20%
F1 AXES (FOLDS IN S1-L1) TIVY MTN SLAB 106 PTS: 1,3,6,9,12%
F1 AXIAL SURFACE POLES TIVY MTN SLAB 101 PTS: 1,2,5,9,15%
S1 POLES HOG MTN SLAB 198 PTS: 1,3,6,9,12,15%
L1 HOG MTN SLAB 49 PTS: 2,5,10%
S2 POLES (SERPENTINITE SCHISTOSITY) KINGS RIVER OPHIOLITE 289 PTS: 1,2,6,12,18%
S3 POLES (PLUTON AND WALLROCK SCHISTOSITY) AREA A 108 PTS: 1,3,6,9,13%
S2 POLES (SOUTHWEST OF KAWEAH RIVER) 501 PTS: 1,2,5,10,15,20%
FOLD AXES IN EPICLASTICS AREA B 84 PTS: 1,5,10,14,17%
S3 POLES (SLATY CLEAVAGE IN EPICLASTICS) AREA B, 379 PTS: 2,10,20,25,30%
S2 POLES, TULE RIVER, 454 PTS: 1,2,4,10,15%
S3 POLES, AREA C 238 PTS: 2,5,12,20,29%
FOLD AXES IN EPICLASTICS AREA C 34 PTS: 3,10,20%
EXPLANATION
JURASSIC-CRETACEOUS
PLUTONIC ROCKS OF THE SIERRA NEVADA BATHOLITH: QUARTZ DIORITE, TONALITE, DIORITE, GABBRO AND GRANODIORITE
UNDIFFERENTIATED METASEDIMENTARY AND METAVOLCANIC ROCKS:
PROTOLITHS
METAMORPHIC DERIVATIVES
RADIOLARIAN CHERT AND SILICEOUS ARGILLITE, PARTLY CHAOTIC AND PARTLY BEDDED; LOCAL QUARTZOSE SANDSTONE INTERBEDS AND SHALLOW WATER LIMESTONE BLOCKS
QUARTZ-MICA SCHIST, METAQUARTZITE AND MARBLE
QUARTZITIC TO SLIGHTLY ARKOSIC AND/OR VOLCANOGENIC SANDSTONE AND MUDSTONE
SLATE AND QUARTZ-MICA SCHIST
AUGITE PORPHRY BASALT-ANDESITE TUFF, BRECCIA AND PILLOW LAVA
ALBITE-EPIDOTE HORNFELS AND AMPHIBOLITE
PERMIAN
KINGS-KAWEAH OPHIOLITE BELT
MELANGE UNITS (PROTOLITHS)
PURE AND LOCALLY ARGILLACEOUS RADIOLARIAN CHERT WHICH OCCURS AS SEMI-INTACT, HIGHLY DEFORMED DEPOSITIONAL REMNANTS ABOVE SERPENTINITE MELANGE
TECTONIC BLOCKS OF DEFORMED BASALTIC PILLOW LAVA AND BRECCIA, MAFIC DIKE ROCK, GABBRO, PYROXENITE, MAFIC TECTONITE (GREENSCHIST TO AMPHIBOLITE FACIES), AND PURE AND METALLIFEROUS RADIOLARIAN CHERT DISPERSED IN SCHISTOSE SERPENTINITE
TECTONIC BLOCKS OF DEFORMED GABBRO, PYROXENITE, MAFIC TECTONITE (AMPHIBOLITE FACIES), WEHRLITE, HARZBURGITE AND DUNITE DISPERSED IN SCHISTOSE SERPENTINITE
TECTONIC BLOCKS OF DEFORMED PURE AND METALLIFEROUS RADIOLARIAN CHERT, OPHICALCITE, GABBRO, WEHRLITE, HARZBURGITE, AND DUNITE DISPERSED IN SCHISTOSE SERPENTINITE AND DEFORMED DETRITAL SERPENTINITE
SLABS AND LARGE BLOCKS
PROTOLITHS
METAMORPHIC DERIVATIVES
BASALTIC PILLOW LAVA, BRECCIA AND LESSER TUFF, AND BASALT-DIABASE AND LESSER GABBRO DIKE ROCK: PILLOW AND DIKE SYMBOLS SCHEMATICALLY SHOW LOCATION OF RELICT PRIMARY FEATURES
MAFIC HORNBLENDE HORNFELS AND SCHISTOSE AMPHIBOLITE; TREND OF DISTORTED PILLOWS SHOWN BY ELONGATION OF RELICT PILLOW SYMBOL
CLINOPYROXENE (± OLIVINE OR HORNBLENDE) GABBRO, PYROXENITE AND LESSER WEHRLITE WITH LOCAL CUMULATE LAYERING AND MYLONITE ZONES
MAFIC HORNBLENDE HORNFELS, HORNFELSED MYLONITIC GNEISS AND TREMOLITE-CHLORITE (OR ANTIGORITE) HORNFELS; TREND OF DEFORMATION ZONES SHOWN BY SLASHES
HIGHLY DEFORMED DIKE- AND SILL-LIKE MASSES OF GABBRO, PYROXENITE, WEHRLITE, AND LESSER DIORITE AND PLAGIOGRANITE WITHIN HARZBURGITE-DUNITE MYLONITE HOST
COMBINATION OF ROCKS PRESENTED DIRECTLY ABOVE AND BELOW; TREND OF MYLONITIC FOLIATION AND DEFORMED MAFIC BODIES SHOWN BY SLASHES
HARZBURGITE-DUNITE MYLONITE WITH LOCAL INCLUSIONS OF MAFIC TECTONITE (AMPHIBOLITE FACIES), WEHRLITE AND CHROMITE-OLIVINE CUMULATE
TALC-ANTIGORITE HORNFELS AND SCHIST; TREND OF MYLONITIC FOLIATION AND MAFIC INCLUSIONS SHOWN BY SLASHES
CARBONIFEROUS(?)

PLATE V