

PETROLOGY AND PETROGENESIS OF BASALTS  
FROM THE OLYMPIC PENINSULA, WASHINGTON

by

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for the degree of Master of Science

at

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DALHOUSIE UNIVERSITY  
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ABSTRACT

The petrography and chemistry of basic volcanic rocks from the Olympic Peninsula, Washington are described, and 26 new chemical analyses are presented. Petrographically and chemically the majority of these rocks can be regarded as basalts or metamorphosed basalts, and not spilites as has been previously assumed. Changes in chemistry are discussed and it is shown that the major changes involve hydration, oxidation and carbonitisation. Normative calculations indicate that the large majority of rocks are tholeiitic, although some are alkali basalts.

Projections of rock compositions in relation to phase boundaries at different pressures within the normative and modified CMAS tetrahedra have enabled three petrogenetic models of the Olympic Peninsula basalts to be established, which involve partial melting of mantle peridotite at low, intermediate and high pressures. Partial melting of peridotite at intermediate or high pressures, followed by olivine fractionation from the magmas during their ascent to the surface, can account for the observed compositions of Olympic Peninsula basalts.

The Olympic Peninsula basalts, and basalts of similar age found in the Coast Ranges of western Washington and Oregon, are interpreted as the products of island arc volcanism associated with the subduction of oceanic lithosphere off Washington and Oregon during the early Cenozoic era. Problems associated with magma genesis in island arcs are discussed, and some consideration is given to the spatial and temporal aspects of magma genesis in island arcs.

Problems concerned with the nomenclature and origins of spilites are reviewed, and conclusions are made regarding the most likely



processes which cause spilitisation. Projections of basalts, spilites and Olympic Peninsula basalts onto the "spilite plane" diopside-clinocllore-feldspar within the CMAS tetrahedron are discussed, and conclusions are made regarding the nature of the Olympic Peninsula volcanic rocks, and on the relationship between spilites and basalts in general.

## INTRODUCTION

### I Purpose of Research

In 1969, several members of the Geology Department at the University of Alberta instituted a major programme of research on spilites. These efforts were directed towards determining the genesis of these rocks and the role which spilitisation plays, if any, in the formation of metalliferous ore deposits. An association of important metalliferous ore deposits with rocks of the spilite - keratophyre kindred has been recognised in many parts of the world, e.g. iron ore deposits at Kiruna, Sweden and Iron Mountain, Missouri; the iron and manganese deposits of eastern Switzerland (Amstutz 1958); and the manganese deposits of the Olympic Peninsula, Washington (Park 1946). Thus an investigation into the genesis of spilites may yield results that could be of value in predicting favourable metallogenic provinces.

In August 1969, rock samples were collected from the Olympic Peninsula, Washington, a region which has been regarded as a classic spilite locality (Turner and Verhoogen, 1960), although this reputation has been based upon extremely limited chemical data. As an integral part of the research on spilites, the major element chemistry and petrology of the Olympic Peninsula volcanic rocks have been studied, and this thesis deals with the results of that study.

### II Outline of Thesis

In order to develop a framework in which to discuss the petrology of

the Olympic Peninsula volcanic rocks, problems concerned with the nomenclature and origins of spilites have been discussed in Chapter 2. The petrology and chemistry of the volcanic rocks from the Olympic Peninsula are then described in Chapter 3. From these studies it became apparent that the majority of basic volcanic rocks were not true spilites, but rather basalts or altered and metamorphosed basalts. Spilitic basalts and albite diabases have been found among the Olympic Peninsula volcanic rocks which were studied, but appear to be in the minority compared to the volcanic sequence as a whole. Chemical changes in the volcanic rocks have been discussed, and it is shown that the major changes in chemistry involve hydration, oxidation and carbonitisation. Chapter 4 is concerned with the petrogenesis of the Olympic Peninsula basaltic rocks. The behaviour of these rocks in relation to phase boundaries at different pressures is examined by means of projections within the normative and modified CMAS (calcium-magnesium-aluminum-silicon) tetrahedra, and based on the results of these projections, petrogenetic models for the Olympic Peninsula basalts have been proposed, involving partial melting of mantle peridotite at low, intermediate and high pressures. Projections of Olympic Peninsula basalts, spilites and other basalts from various parts of the world onto the "spilite plane" diopside-clinocllore-feldspar are discussed, and conclusions are made regarding the nature of the Olympic Peninsula volcanic rocks, and on the relationship between basalts and spilites in general. In Chapter 5, the origin of the lower-middle Eocene basaltic rocks of the western Washington and Oregon Coast Ranges is discussed from a plate tectonic point of view. A new interpretation of the geological evolution of these Coast Ranges is presented, involving subduction of oceanic lithosphere off Washington

and Oregon during the early Cenozoic with resulting volcanic arc igneous activity. Inferred plate movements during the early Cenozoic are discussed with reference to this igneous activity. Chapter 6 deals directly with the problems of magma generation in island arcs. The spatial and temporal relations of tholeiitic and calc-alkaline lavas in island arcs are emphasised, and recent theories concerned with the genesis of these rocks are discussed. Conclusions are made regarding the most likely process of generation of calc-alkaline magmas, and a theory is proposed which accounts for the generation of tholeiitic magmas, their observed chemical characteristics, and their time of eruption in island arcs. Chapter 7 presents the list of conclusions drawn from the previous six chapters.

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## CHAPTER 1

### GEOLOGICAL HISTORY AND STRUCTURE OF THE OLYMPIC PENINSULA

#### I. Introduction

The Olympic Peninsula forms the northwestern part of the state of Washington, lying between latitudes 47°N and 49°N and longitudes 123°W and 125°W, and is bordered on the west by the Pacific Ocean, on the north by the Strait of Juan de Fuca, and on the east by Puget Sound and Hood Canal (Fig. 1). The Olympic Mountains occupy the central part of the Peninsula, comprising a complex system of glaciated valleys and canyons with intervening ridges and peaks that commonly attain altitudes of 6000 feet. Mount Olympus, the highest peak, has an elevation of 7954 feet. Relatively narrow, hilly lowlands border the mountains on the west, north and south, but the descent on the east to Hood Canal is abrupt. Thick surficial deposits and precipitation up to 140 inches per year over parts of the Peninsula result in heavy forestation, and consequently exposures of bedrock are poor and limited below the timber line. Exposures are generally good above the timber line but access to them is difficult.

#### II. The Olympic Peninsula: geological history

The basement geology of western Oregon and Washington is believed to be of the Nevadan complex which has evolved from the Paleozoic and early Mesozoic Cordilleran geosyncline as a result of folding, metamorphism and batholithic intrusion. In the Cordilleran geosyncline the cycles of

sedimentation and volcanism initiated in Precambrian time were brought to a close in a major orogeny that began in the middle of the Mesozoic and continued into the early Cenozoic. Two major mountain-building events occurred during this period: one at the end of the Jurassic period, called the Nevadan, during which the great granitic batholiths of the western Cordillera were intruded; and one at the end of the Cretaceous period, called the Laramide, when the Rocky Mountains were formed. The Klamath Mountains, Northern Cascade Mountains and Coast Ranges of British Columbia were largely formed during the Nevadan orogeny. Because these Nevadan complexes essentially encircle western Washington and Oregon (Fig. 2), it has been assumed that the basement beneath the overlying Tertiary formations in western Washington and Oregon is of Nevadan age and origin. The Olympic Mountains, Coast Ranges and extensive volcanic fields of western Washington and Oregon developed during the Tertiary, after the close of the Nevadan orogeny (Eardley 1962).

Snively and Wagner (1963) have discussed the Tertiary geological history of western Oregon and Washington. According to their interpretation a geosyncline extended from the southern end of Vancouver Island, through the Olympic Mountains, Coast Ranges and Puget-Willamette lowlands of western Washington and Oregon, to the northern end of the Klamath Mountains in southern Oregon at the beginning of Tertiary time (Fig. 3). This linear basin of deposition was over 400 miles long, its eastern margin extending beneath the present site of the Cascade Range, and its western margin extending some miles west of the present coastline.



Early in the Eocene a thick sequence of tholeiitic basaltic lavas and breccias was erupted from numerous centres onto the floor of the subsiding geosyncline. These volcanics are referred to as the Metchosin and Crescent formations in western Washington, and as the Umqua, Siletz River and Tillamook series in western Oregon. A lower Eocene age, based on stratigraphic considerations, was originally assigned to these volcanics by Weaver (1937). The base of the volcanics is exposed only on the Olympic Peninsula, where the Metchosin volcanics unconformably overly a metasedimentary sequence called the Solduc formation. This formation was originally believed to be Cretaceous in age (Weaver 1937), but later work by Cady and MacLeod (1963) has shown that the Solduc may be in part Palaeocene-Eocene in age. In a recent paper, Tabor (1972) has indicated that the stratigraphic age of all core rocks in the Olympic Mountains is probably Tertiary. An upper age limit for the Metchosin volcanics is given as the beginning of the middle Eocene by Weaver (1937), who records marine sediments containing a molluscan fauna of lower middle Eocene age overlying the volcanics at Crescent Bay on the south side of the Strait of Juan de Fuca. No radiometric age dates are available from the Eocene volcanic sequences.

The probable extent of this early Eocene volcanism in western Washington and Oregon is shown in Fig. 4, and a more detailed distribution of volcanics in the Olympic Peninsula is shown in Fig. 1. These basal Eocene volcanics are believed to have formed a vast lava field that extended from Vancouver Island southwards to the Klamath Mountains, and from a line west of the present coastline and extending some 150 miles inland. Weaver (1945) estimates that a minimum value for the average

thickness is 3000 feet and their volume to be greater than that of the Columbia plateau basalts. According to Snavely and Wagner (1963) the volcanic sequence totals more than 15000 feet in thickness on the Olympic Peninsula where the base is exposed.

Most of the lava was apparently erupted onto the sea floor, either as flows or extrusive breccia. Well-developed pillow structures are common, along with volcanic breccias and intercalated marine sediments (Park 1946), but Weaver (1945) also reports numerous small vents and dykes which he believes were the source of much of the volcanic material. Snavely and Wagner (1963) have also found interflow soil zones and interbedded, locally-derived mud-flow breccia and conglomerate, indicating that volcanic islands formed in places and that a part of the volcanic series was erupted subaerially (Fig. 3). Harvey (1959) notes that pillow lavas become less common towards the top of the Metchosin volcanics in western Washington, and that columnar jointing becomes more common.

Waters (1955) has described the volcanics as being largely aphanitic basalts composed of clinopyroxene and labradorite, set in a tachylitic base highly charged with magnetite dust. Phenocrysts of augite or plagioclase appear in some flows, but the series as a whole is characteristically non-porphyrific. Olivine is scarce or absent, and glass commonly accounts for 20 - 40 per cent of the rock. Park (1946) has discussed the spilitisation of these basalts in the Olympic Peninsula, and Snavely and Wagner (1963) indicate that zeolitisation rather than spilitisation is more common in the Coast Ranges of Oregon.

Throughout the Olympic Peninsula and Coast Ranges these volcanic units intertongue complexly with fossiliferous, tuffaceous siltstones that contain graded beds of volcanic, feldspathic and lithic wackes. Beds of chert and Globigerina-bearing limy siltstones occur locally (Snively and Wagner, 1963). In the Olympic Mountains, where the base of the Tertiary volcanics is exposed, a thick sequence of argillite and greywacke of Cretaceous (?) age, the Solduc formation, underlies the lower Eocene volcanics (Weaver 1937). The northeastern border of the geosyncline during the early Eocene was probably a broad, coastal plain across which meandering rivers carrying arkosic debris from the pre-Tertiary highlands in northern Washington. These continental deposits are represented in parts of the Puget and Chuckanut groups and the Swauk formation.

Between middle Eocene and middle Miocene times, sedimentation, intermittent volcanic activity and localised uplift continued in the geosyncline, while contemporaneous andesitic and dacitic volcanic activity occurred to the east of the geosynclinal margin (Snively and Wagner, 1963). In middle Eocene time, major uplift and erosion in the ancestral Klamath Mountains resulted in an influx into the southern part of the basin of great quantities of arkosic detritus, much of which was transported northwards into the deeper parts of the geosyncline by turbidity currents, to form the Tyee formation. In late Eocene time, local uplift and extrusion of alkalic basalt and breccia (largely in the Coast Ranges) divided the geosyncline into several separate basins of deposition. Tuffaceous sandstone and siltstone accumulated in these basins from the late Eocene to the Pliocene, and the general axis of sedimentation shifted westwards

with time. Folding and compression occurred during upper Miocene over western Washington and Oregon, resulting in northwest-southeast trending upwarps and intervening downwarps. Tabor (1972), on the basis of K-Ar dating of deformed sedimentary rocks in the core of the Olympic Mountains, reports the occurrence of two metamorphic events which affected the Olympic Peninsula: a regional one at 29 m.y., and a local one at 17 m.y. Erosion and further sedimentation took place around the margins of the Olympic Mountains during Pliocene time, and during late Pliocene or early Pleistocene further folding occurred. North-south structures were developed, emphasising the individuality of the Coast and Cascade Ranges and Puget-Willamette trough. These final phases of folding resulted in uplift and compression of the Solduc and overlying Tertiary formations into the prominent domal structure of the present Olympic Mountains (Weaver 1937, 1945; Eardley 1962).

### III. Structure of the Olympic Peninsula

Weaver (1937, 1945) indicated that in the interior of the Peninsula the Solduc formation has been greatly indurated, and in places metamorphosed, and is unconformably overlain by the later Tertiary volcanics. The structure of the Olympic Mountains has been broadly described as anticlinal, based on the following observations: the outcrop pattern, with younger sedimentary rocks on the flanks of the volcanics, suggests an anticline plunging to the east; the tops of the volcanics are everywhere towards the outside of the Peninsula; and the Solduc formation, stratigraphically below the volcanics, is more highly deformed and metamorphosed than the overlying Tertiary formations (Park 1950).

Stuart (1961), on the basis of a gravity study of crustal structures in western Washington, indicates that the pattern of gravity highs associated with the lower Eocene volcanics in the Olympic Mountains suggests the structure to be a southeast-plunging anticlinorium. Cady and Tabor (1964) have described large, northwesterly-plunging asymmetrical folds with limbs about a mile wide in the Solduc formation from the eastern Olympic Mountains. These folds lie stratigraphically below, and to the west of, nearly vertical volcanic rocks of the Crescent formation. Cady and Tabor (1964) believe that the volcanic rocks were not involved in the folding which affected the greywacke-slate sequence, thus indicating a strong tectonic break between the volcanic rocks and the underlying greywacke-slate section. However, they suggest that the confinement of the steeply plunging folds to the greywacke-slate sequence is analagous to salt dome structures, wherein the complexly folded salt may correspond to the folded, interbedded greywacke and slate, and the Crescent volcanics may correspond to the resistant cap rock.

Park (1950) cites facts that are not in accord with a simple anticlinal theory for the structure of the Olympic Mountains: fossils of Oligocene age were found by Warren (pers. comm. in Park 1950) in rocks of the Solduc (?) formation; and secondly, the tremendous apparent thickness of 120,000 feet in the section along the Dosewallips river, including the underlying Solduc formation, compared with an estimated average thickness of approximately 25,000 feet, is greater than can be explained by normal geological depositional processes. Park suggests that the great apparent thickness and complexity of structure has resulted from steeply dipping thrust faults and considerable buckling

distributed through the rocks of the Solduc formation. Additional evidence of the complexity of the structures in the Olympic Mountains, and of the controversy concerning the age of the Solduc formation, has been given by Cady and MacLeod (1963). They have found a persistent fossiliferous horizon within the Solduc formation which outcrops intermittently for about 21 miles along the strike, and varies between 500 and 1000 feet thick. This horizon lies approximately 15000 feet below the base of the lower Eocene Crescent formation, and has yielded megafossils of Palaeocene-Eocene age. This discovery may indicate that a considerable part of the Solduc formation is early Tertiary in age, rather than Cretaceous as was previously believed.

#### IV. Summary

The present Olympic Mountains and Coast Ranges evolved from the processes of sedimentation, volcanism, folding and metamorphism which all occurred within a geosyncline situated in western Washington and Oregon during the Tertiary. Voluminous volcanic material was erupted, largely under submarine conditions, during lower to middle Eocene times and is complexly interbedded with contemporaneous marine tuffaceous sediments. Sediment deposition, intermittent volcanic activity and localised uplift continued in the geosyncline until Miocene time, when folding and compression resulted in northwest-southeast trending upwarps and intervening downwarps. After erosion and further sedimentation, renewed folding in late Pliocene or early Pleistocene times resulted in the Olympic Mountains and Coast Ranges attaining their present elevations.

The structure of the Olympic Peninsula is poorly known. Originally believed to be an anticline plunging to the east, further evidence of thrusting has shown the structure to be more complex.

- Fig. 1 Location map of the Olympic Peninsula showing distribution of lower Tertiary volcanics.  
(from Tabor 1971, Fig. 1)



Lower Tertiary Volcanics

- Fig. 2 Geomorphological and geological divisions in western Washington and Oregon.  
(from Eardley 1962, Fig. 29.15)
- Fig. 3 Probable extent of the early Tertiary geosyncline in western Washington and Oregon.  
(from Snavely and Wagner 1963, Fig. 1)  
-----inferred eastern margin of geosyncline.
- Fig. 4 Distribution and extent of early Eocene volcanism in western Washington and Oregon  
(from Snavely and Wagner 1963, Fig. 2)
- Fig. 5 Geological map of the Olympic Peninsula and southern Vancouver Island.  
(after Hawkins 1967, Fig. 1)
- Fig. 5A Lower Tertiary stratigraphy, western Washington and Oregon.  
(from Weaver 1945 and Eardley 1962)



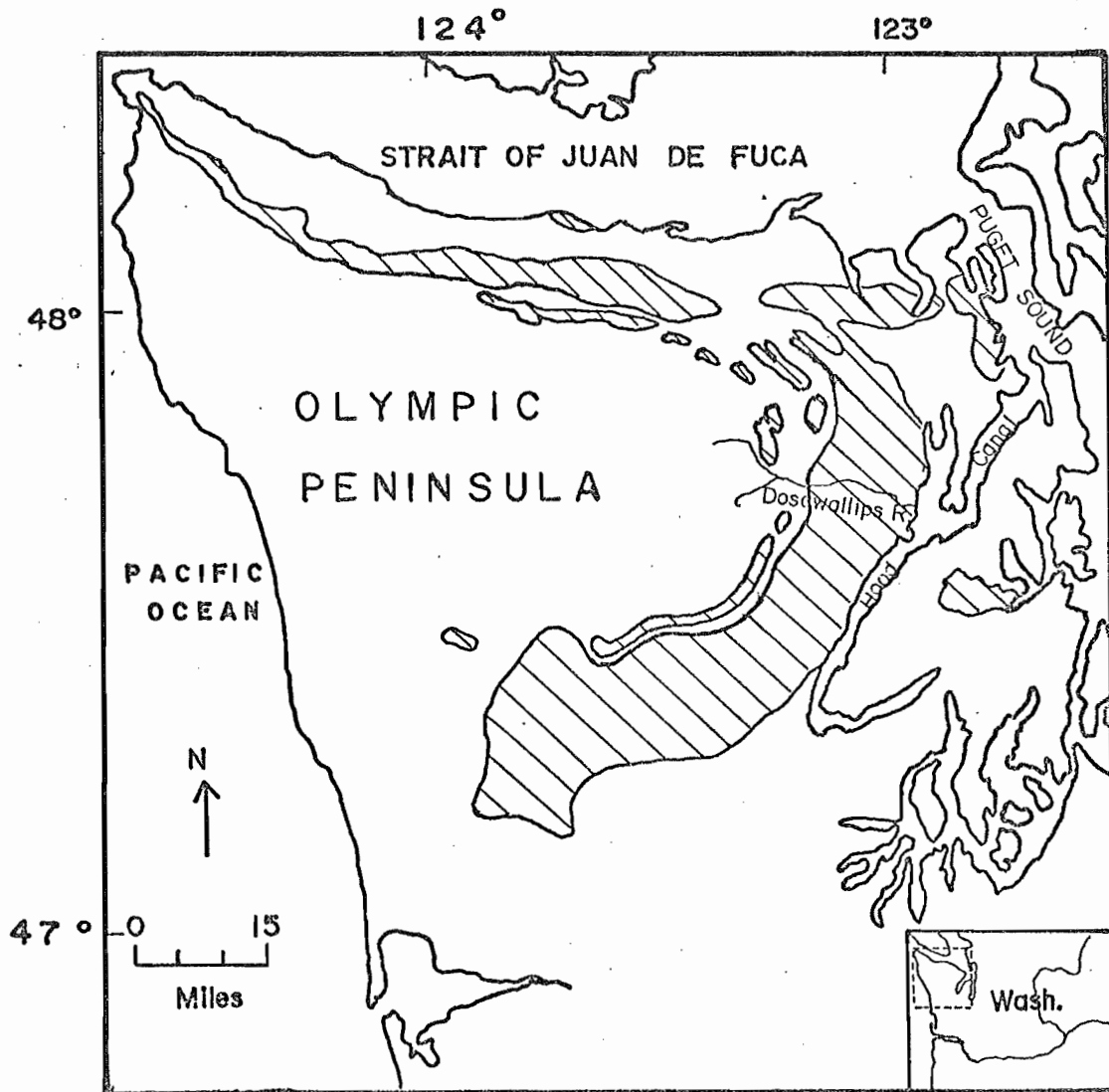


FIG. 1

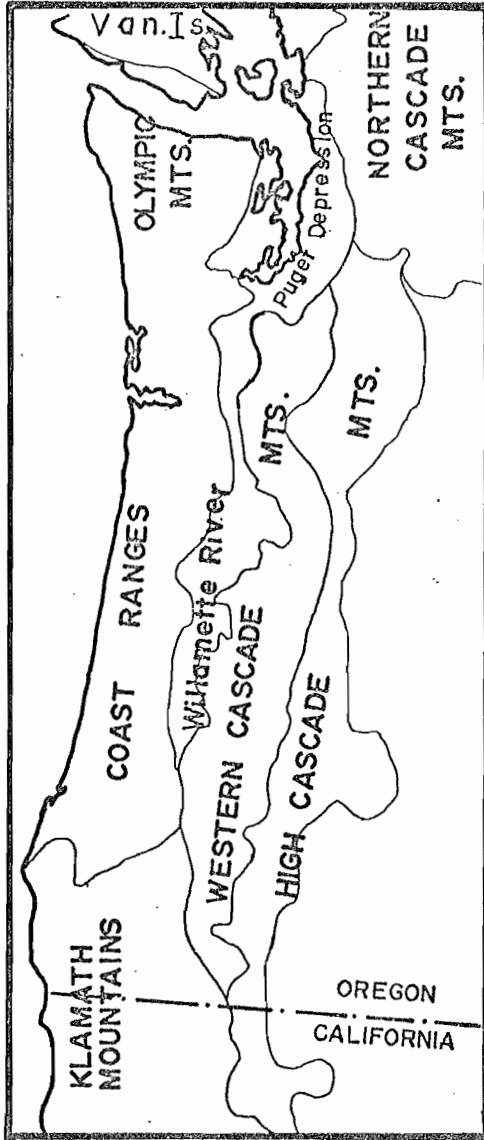


FIG. 2

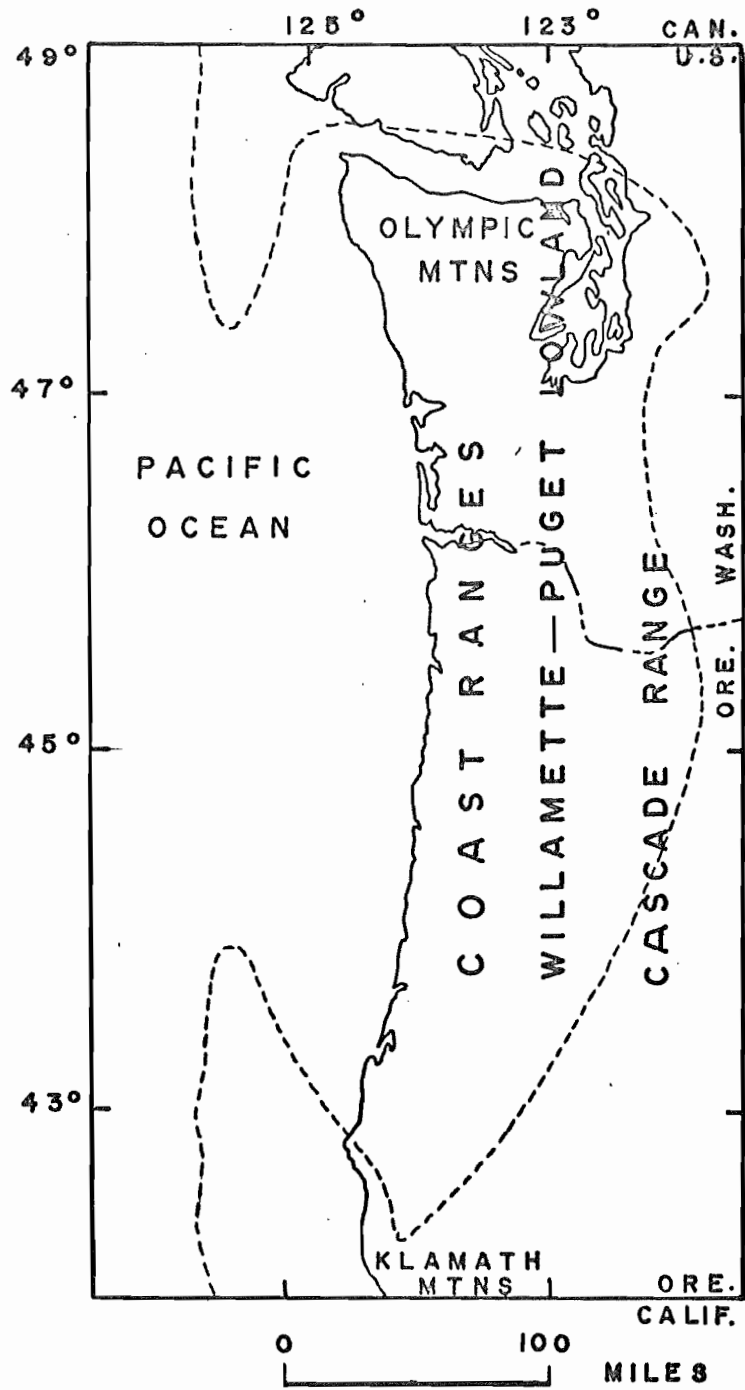


FIG. 3

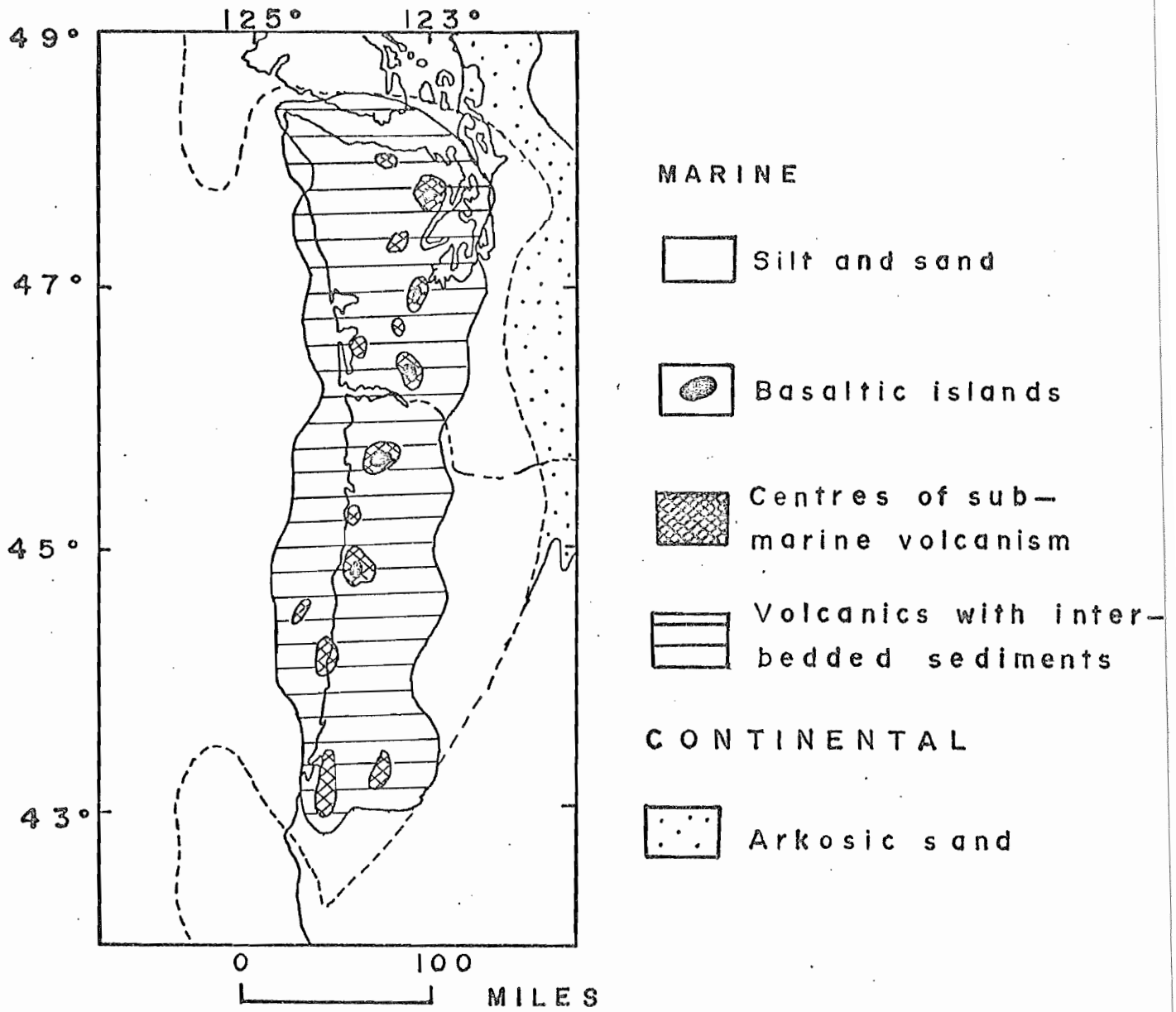
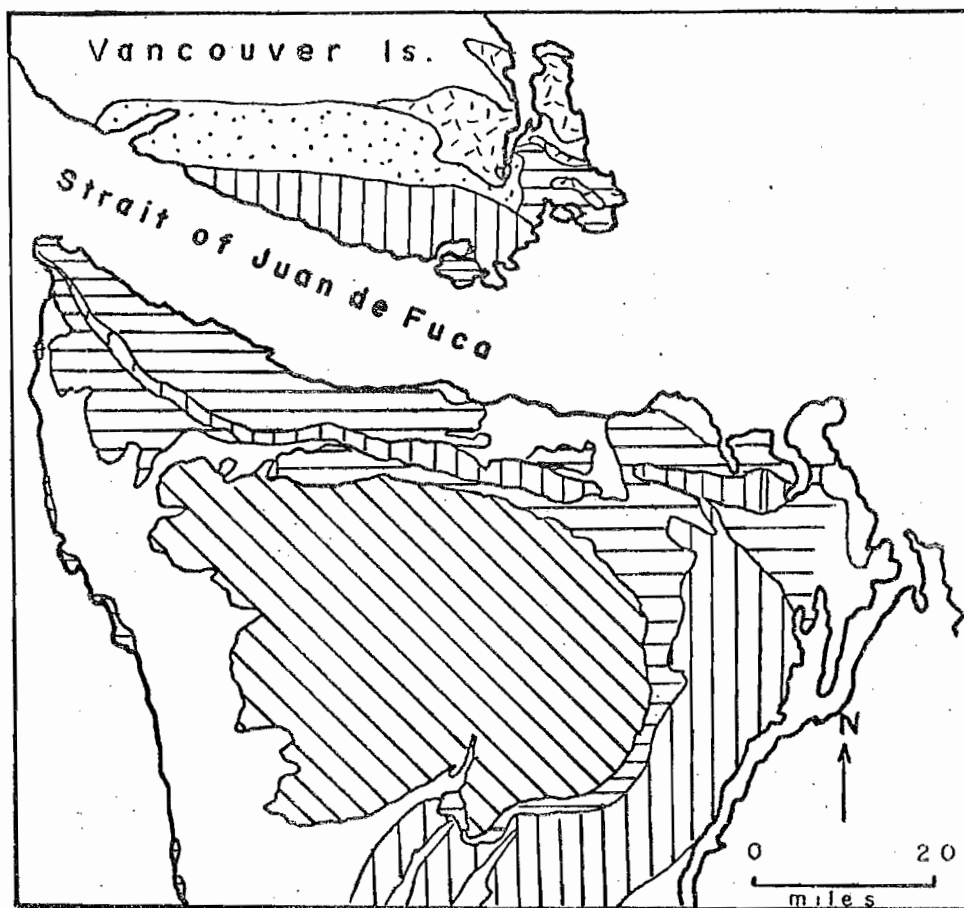


FIG. 4

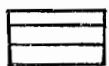
FIG. 5



**KEY**



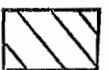
Quaternary



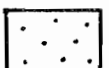
Tertiary sediments



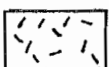
Eocene volcanics



Pre-Tertiary sediments



Devonian-Permian sediments  
and volcanics



Pre-Devonian plutonics

LOWER TERTIARY STRATIGRAPHY, W. WASHINGTON & OREGON

from Weaver 1945 and Eardley 1962

	W. OREGON	W. WASHINGTON		E. WASHINGTON
U. EOCENE	Coaledo Fm.	Cowlitz Fm.	Puget and Chuck-anut Gps.	Roslyn Fm.
M. EOCENE	Tyee Fm.	Crescent Fm.		
L. EOCENE	Umqua, Siletz River, Tillamook Series	Metchosin volcanics		Teannaway bas.
				Swauk Fm.
PALAEOCENE U. CRET.	P	Solduc Fm.		

FIG. 5 A

CHAPTER 2

SPIILITES: TERMINOLOGY AND ORIGINS

I. Introduction

The purpose of this chapter is to present a discussion on the origin of the term "spilite", and to present, in summary form, various arguments that have been proposed to account for the origin of spilites. Conclusions are made as to the most likely processes responsible for spilitisation, and to the meaning of the term spilite.

Within recent years the rock called spilite has received much attention, especially since the concepts of continental drift and sea-floor spreading have been synthesised under the general hypothesis of plate tectonics. Unfortunately, the increased usage of the term spilite has led to petrographic and petrogenetic confusion, because there is no common consensus of opinion on what a spilite actually is, and what its origins are. Most of these problems have arisen because various authors have tried to enforce a common process of origin for spilites from widely different geographical localities.

II. Historical Background

Amstutz (1968) and Vallance (1960, 1969) have recently given comprehensive reviews of the spilite problem, and the following brief historical account of the term spilite has been largely taken from Vallance (1960).

The term spilite was introduced by Brongniart (1827) as a loose, megascopic designation for a variety of extrusive rocks that had essentially one main feature in common - an aphanitic, greenish or purplish appearance. English-speaking geologists made little use of the term spilite during the 19th century, preferring the equally nebulous term of greenstone to categorise such rocks. Following the work of Teall (1894) and Flett (1907), Dewey and Flett (1911) published their review of British pillow lavas, reintroducing the term spilite. They noted the following features in the rocks they were calling spilites: the rocks very often showed signs of alteration and decomposition; the feldspars were invariably rich in soda; most of the feldspars occurred as microlites or fine laths, sometimes arranged in fluidal fashion or as radiating clusters; pyroxenes were generally relict and commonly replaced by secondary minerals such as chlorite, epidote and calcite; fresh olivine was never present, and many of their spilites contained devitrified glass.

The few chemical analyses available to Dewey and Flett (1911, p. 206) indicated markedly higher soda, carbonate and water compared with those of other known basic igneous rocks.

In calling these sodic rocks spilites, Dewey and Flett claimed that they were of the same type as the rocks called spilites by Brongniart. This is in fact true, but the mineralogical criteria, especially that of sodic feldspar, and the chemical criterion of high soda, now effected a considerable restriction upon the term spilite.

### III. A Working Definition of Spilite



There appear to be four main criteria used in the recognition of spilites, namely: texture, mineralogy, chemical composition and geological occurrence. Much of the confusion associated with the term spilite probably has derived from the fact that various workers have tended to emphasise one or another of these features occurring in their rocks, and have attempted to propose individual origins which may not necessarily be applicable to other spilites.

For the purposes of the following discussion the term spilite will be taken to mean a rock of approximately basaltic composition, characterised by the presence of sodic plagioclase, chlorite, a high soda content, and generally exhibiting textures typically found in basalts and diabases.

#### IV. Origins of Spilites

Spilitic rocks occur as bodies having broadly the characteristics of basic volcanics. Field evidence points to many being truly extrusive, whereas others form relatively shallow intrusions. There is considerable evidence from their mineralogy that most spilites are chemically adjusted to hydrous, low-temperature conditions. These conditions may prevail in any of several contrasted environments, such as late magmatic, hydrothermal and low-grade metamorphic. The important fact concerning the origin of spilites is whether the adjustment to hydrous, low-temperature conditions is achieved through magmatic or post-magmatic processes.

Amstutz (1968, Table 1) has conveniently tabulated the various proposals advanced to account for the origins of spilites. They can be simplified into two broad groups: 1) spilites are primary basic igneous

rocks resulting from direct crystallisation during magmatic cooling; and  
2) spilites owe their present characteristics to secondary, generally  
post-magmatic alteration of basic igneous rocks.

#### 1A. Direct Crystallisation of a Low-Temperature Spilitic Magma

Albite, clear and free from inclusions, has been reported in ophitic or sub-ophitic textural relations with clinopyroxene, and also as fibrous intergrowths with clinopyroxene from various spilitic bodies (Benson 1915a, b, Van Overeem 1948, Amstutz 1954, Scott 1951, Battey 1956). There appears little doubt that the clinopyroxene is a primary crystallisation product, and it has been argued that the textural relations indicate that albite must also be a primary crystallisation product. Thus, those spilites may have crystallised directly from a magma of spilitic composition.

Several convincing arguments have been raised against such a spilite genesis. If spilites crystallised from a low-temperature magma, as their mineralogy would suggest, why do their textures mimic so closely the textures found in basalts, formed at approximately 1000°C? Pseudomorphs after olivine and relicts of calcic plagioclase in the cores of many albites both suggest that the original magma was of a high temperature origin, and that albites are of a replacement origin rather than a primary crystallisation phase.

Clinopyroxene compositions provide evidence against a low-temperature origin. Clinopyroxenes from spilitic rocks apparently have compositions

similar to common basaltic clinopyroxenes, and are very probably inherited from a high-temperature basaltic assemblage (Battey 1956, Donnelly 1966, Yoder 1967, Vallance 1969). Clinopyroxenes formed at lower temperatures with albite and chlorite would be expected to have different compositions from high-temperature basaltic augites. Smith (1970) has found evidence to this effect in pyroxenes from the albite diabase of the Sierra Ancha sill. The calcium-rich pyroxenes, which occur in mineral assemblages characteristic of spilites, apparently crystallised from a water-rich fluid at temperatures lower than the crystallisation temperatures of basaltic pyroxenes, and are richer in calcium and lower in aluminium and titanium than basaltic augites.

Experimental laboratory work on the melting of basalt under hydrous conditions gave little evidence for the formation of a spilitic magma. Yoder and Tilley (1962 p. 459) conclude "all major basalt compositions crystallised on cooling in the presence of water as hornblende gabbro". Yoder (1967) has also shown that even spilite plus water gives amphibole-bearing assemblages over the temperature range from 600°C to the beginning of melting at water pressures of 2-10kb.

#### 1B. Hydromagmatic Crystallisation of Basaltic Magma

According to this hypothesis, spilites can be derived from basaltic magmas rich in volatiles, chiefly water. The considerable water pressure at the depth of lava extrusion on the oceanic bottom has been suggested as the mechanism for causing volatile retention. This volatile retention, in turn, depresses the freezing range of the lavas allowing albite and

chlorite to crystallise as primary mineral phases (Van Overeem 1948; Amstutz 1954, 1958; Niggli 1952, Battey 1956).

Failure to account for the presence of pyroxenes, many of them fresh and unaltered, in spilites is one major problem this theory has to contend with, and the experimental laboratory results discussed in 1A certainly do not corroborate any aspect of this hydrous crystallisation theory. There are other objections to hydromagmatic crystallisation in lava bodies: retention and accumulation of volatiles are essential to the scheme, yet spilites have close textural affinities with basalts that have chilled rapidly; the occurrence of amygdales in many spilites strongly suggests escape of volatiles from the cooling magma. Relatively rapid loss of heat and magmatic volatiles are not conducive to hydromagmatic crystallisation.

#### 2A. Reaction of Basic Magma with Sea Water or Wet Sediments

Some workers, impressed by the frequent occurrence of spilites in former submarine environments, have attributed to sea water, or to sea water in sediments, an important chemical role in the development of spilites. Beskow (1929) believed that albitisation of calcic plagioclase and the conversion of ferromagnesian minerals to chlorite, as observed in spilitic rocks in Lapland, was due to the action of sea water heated during the extrusion of basaltic magma. Some of the soda involved in the albitisation was thus of marine origin. Park (1946) accepted the idea of sea water diffusion, involving an exchange of Na for Ca, to account for the Olympic Peninsula spilites, and Gilluly (1935) emphasised the

influence of wet sediments on magma. Rittman (1958) proposed a variation of this theory whereby chemically-active solutions escape from the magma to form hydrothermal solutions with the sea water immediately at the roof of the magma body. These solutions remained in contact with the magma bringing about the spilitisation of the original magma.

If the sea water reaction theory is a valid mechanism for causing spilitisation through a reaction between basic magma and sea water, then it is surprising that so many fresh and unaltered basaltic rocks have been dredged from the ocean floors. However, if the spilitisation process requires that sodium be derived other than from within the parent magma itself, then obvious sources exist where trapped pore waters in a basalt pile, or in sediments, are of marine origin. Hart (1970) has shown that hydration by sea water of tholeiitic, oceanic ridge basalts results in a gain of sodium in the altered portions of pillows.

## 2B. Autometamorphism of Basic Lavas

At least since the time of Dewey and Flett (1911), autometamorphism of basic magma has proved attractive to many workers to account for the origin of spilites. The fact that magmatic volatiles may be retained in the late fractions of a cooling magma has been established, as has the tendency for the late fractions to react with earlier-formed crystalline phases. In the differentiated Prospect intrusion (Wiltshire 1967), the action of late-magmatic fluids resulted in extensive autometamorphism or deuteritic alteration of the parental alkaline diabase. Accumulation of

volatiles in the upper half of this intrusion led to extensive deuteric alteration of olivine, pyroxene and plagioclase resulting in a secondary mineral assemblage consisting of sodium-rich zeolites, prehnite, clay minerals and carbonate.

Smith (1970) has described similar features from the Sierra Ancha diabase sill, in which accumulation of magmatic fluids has led to the development of a spilitic mineral assemblage. He traced a continuous sequence, in a vertical section, from fresh diabase with a primary mineral assemblage through to rocks which had completely recrystallised under deuteric conditions. The completely recrystallised rock type, albite diabase, contained albite, calcic pyroxene and chlorite, the mineral assemblage characteristic of spilites.

Fawcett (1965) believes that late-magmatic fluids were responsible for alteration features observed in flood basalts from the Isle of Mull, and Nicholls (1958) has postulated an autometamorphic origin for the Lower Spilite formation of the Builth Volcanic Series. Nicholls suggests that the spilitisation of the volcanics resulted from the development of related but immiscible liquids during the consolidation of the magma. The separation of a liquid phase relatively rich in Ca, Mg and Fe led to the enrichment of the host silicate liquid in Na and Si, which thus brought about the metasomatic alteration of the parent rock to produce a spilitic mineral assemblage.

Donnelly (1966) has also proposed an autometamorphic origin for the Water Island spilites in the U. S. Virgin Islands. He suggests that

eruption of a hydrous basic magma in considerable depth of sea water prevented or retarded the escape of magmatic volatiles, causing the original plagioclase to react with the magmatic aqueous phase to yield albite and oligoclase. Hekinian (1971), upon further research on the Water Island spilites, discloses two main features which he believes are indicative of an autometamorphic origin for these spilites, namely: the textural and mineralogical heterogeneity between "dark patches" and the green matrices of the spilites; and, the differences in alkali content between most "dark patches" and their green matrices. He suggests that this chemical and mineralogical heterogeneity resulted from autometasomatism caused by fluids enriched in  $K_2O$ ,  $Na_2O$  and  $H_2O$  derived from within the parent magma. However, Hekinian (1971) has also discussed the role of burial metamorphism in connection with the formation of the Water Island spilites, and believes that autometamorphism was not solely responsible for the formation of these spilites. He suggests that autometamorphism does not fully explain the presence and variation of certain minerals within the rock sequence, citing the presence of prehnite and pumpellyite in the upper 700 feet of the sequence and their eventual disappearance with depth below this level, concluding that burial pressures must be responsible. Accumulation and retention of magmatic volatiles are essential in the autometamorphic origin of spilites, but conditions existing during the eruption of basic lavas are generally not conducive for these features to occur, unless eruption occurs under considerable hydrostatic pressure or the lava pile is built up quickly by a rapid succession of flows. The likelihood of truly autometamorphic conditions being achieved in lava flows, other than in very thick ones, seems much less than in intrusive bodies.

Also, by definition, the acceptance of an autometamorphic origin for spilites denies the possibility of any external gain of sodium, thereby implying that the original parent magma must have had a spilitic composition. Thus, if the original magma was spilitic in composition, why should it have crystallised a basaltic assemblage as the primary crystallisation product, which then later reacted with the sodium-enriched residual magmatic fluid, which up until this stage had somehow been effectively isolated from the basaltic assemblage, rather than directly crystallise a spilite? It may be possible, however, for a fluid phase to separate from a magma body and become concentrated in the magma chamber, the fluids then later reacting with the magma, or with residual magma left from previous eruptions, to produce a spilitic assemblage. The fact that a magma of spilitic composition has yet to be found, and that Yoder's (1967) experimental work in the Spilite - H<sub>2</sub>O system failed to produce a spilitic assemblage upon recrystallisation, may indicate that a spilitic magma does not exist, and that autometamorphism, without gain of external sodium-bearing fluids, is not a viable process in the production of relatively thick sequences of spilites.

## 2C. Burial or Low-Grade Metamorphism of Basic Lavas

Thick accumulations of volcanic and sedimentary material may be exposed to increases in temperature and pressure during burial, sufficient for the formation of metamorphic mineral assemblages of the zeolite facies, or with increasing temperature and pressure, the prehnite-pumpellyite or the greenschist facies. Coombs et al. (1959) and Coombs (1960, 1961) have described such changes in rock sequences from New Zealand. Similar



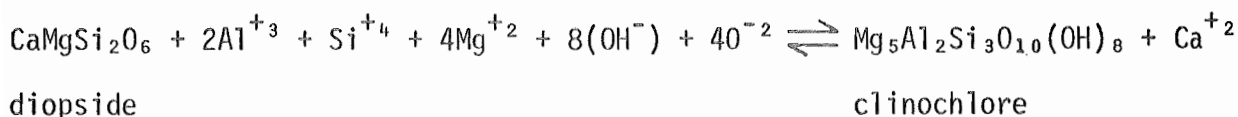
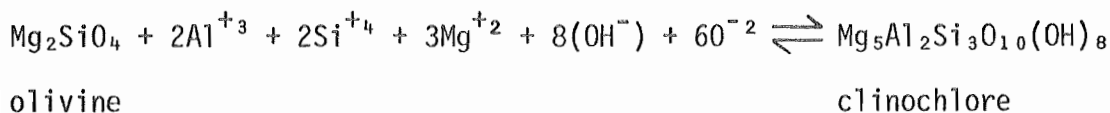
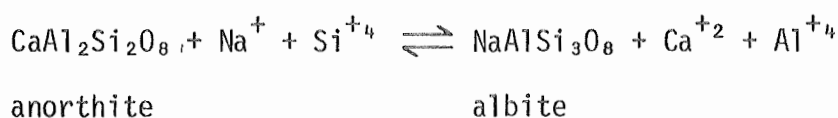
conditions may be expected to operate at active oceanic ridges where thick volcanic sequences are subject to conditions of elevated heat flow, intrusion by ultramafic bodies, and regional deformative movements. Under such conditions of metamorphism, it has been proposed that spilites are the low-grade metamorphic equivalents of basalts. Albite, which is the stable feldspar under these conditions, forms from the resulting instability of calcic plagioclase, and chlorite forms by the breakdown of clinopyroxene or olivine. Melson and Van Andel (1966) have described greenstones rich in albite and chlorite, and with high  $\text{Na}_2\text{O}$  contents, from dredge hauls on the Mid-Atlantic Ridge at  $22^\circ\text{N}$ . They attribute the origin of these rocks to the alteration of basaltic rocks by regional and cataclastic metamorphism associated with faulting and uplift in the area. Cann (1969) has described a similar suite of spilitic rocks from a dredge haul at  $5\ 1/2^\circ\text{N}$  on the Carlsberg Ridge. Texturally, these spilites are very similar to fresh basalts dredged from the same area but chemically and mineralogically they are strikingly different. The pillow cores, consisting of an albite-chlorite-augite-sphene-(actinolite-epidote) assemblage are strongly contrasted with the pillow margins, composed principally of chlorite. Because of the textural similarity between the basalts and the spilites, Cann believes that the spilites must be related to the basaltic pillow lavas, and suggests that a metamorphic origin is the one that best explains the origins of these spilites. He envisages an original basalt pillow pile being buried beneath later lava flows and heated sufficiently for an assemblage corresponding to the lower greenschist facies to be formed. Under these conditions, the texturally different components react differently. The glass, and perhaps the

cryptocrystalline groundmass, form chlorite. Olivine, and perhaps clinopyroxene, are replaced by chlorite; and calcic plagioclase is replaced, without change of shape, by albite. Thus, this process which consists of mineral replacements, can lead to a very contrasted development of minerals in the cores and margins of the pillows. The glassy margins will be transformed almost entirely to chlorite, while in the originally holocrystalline cores, chlorite will be much less abundant and albite much more so. Cann believes that such a metamorphic process, involving considerable transport of chemical components between cores and margins of pillows, can be achieved at oceanic ridge crests at depths of 1 1/2- 2 km, where heat flow values may be greater than six times the normal for oceanic basins.

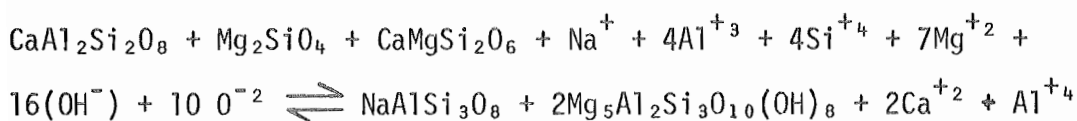
Smith (1968) has described an Ordovician geanticlinal marine sequence of basic lavas, labile sedimentary and pyroclastic rocks which have undergone extensive adjustment during regional burial metamorphism. The basic lavas developed a patchy coloration due to the formation of alteration minerals of the prehnite-pumpellyite facies, which formed during the burial metamorphism. The heterogeneous appearance of the basic lavas reflects a pronounced chemical and mineralogical departure from the original composition. Two diverging trends have resulted; one leading to a Ca-enriched lithology and the other, which dominates in area at each outcrop, to a spilite lithology. Smith believes that under the conditions of burial metamorphism, chemical components released by the primary igneous minerals were remobilised giving rise to the new mineralogical distributions. The replacement of calcic plagioclase by albite, part of the clinopyroxene

and most of the glass by chlorite released much of the calcium. At this facies of burial metamorphism, the Ca was then precipitated by the formation of the stable Ca-bearing phases, pumpellyite, epidote, prehnite, sphene or calcite. Thus, within the lavas, mineralogical patches enriched in albite, clinopyroxene and chlorite, and patches enriched in the secondary Ca-bearing phases are developed. This origin of spilites by burial metamorphism was proposed by Waters (1955) for the spilitic lavas of Washington and Oregon, by Battey (1955) for the keratophyre lavas of northern New Zealand, and by Vallance (1960, 1965) for geosynclinal spilites.

The chemical changes involved during the metamorphic transformation of basalt to spilite can be described by the following simplified equations:



Mass Balance:



It can be seen from the mass balance equation that the transformation

requires the addition of water, sodium, magnesium and iron (not shown) and silica to produce the spilitic mineral assemblage. Water, sodium, magnesium and iron may be present in trapped interstitial fluids in submarine lava piles, or in sea water, but it is unlikely that silica will be present in these fluids in the amounts required by the transformation process. The fact that water and the elements Na, Mg, Fe and Si are required in this transformation process indicates that, if spilites are derived by a metamorphic process from basalts, metamorphism did not behave as a closed system, but was open to the influence of fluids.

#### V. Sources of Sodium and Fluid Diffusion

A question that remains unanswered, and one that bears heavily on the origin of spilites, is whether or not any sodium is derived from an external source during the process of spilitisation. To convert a basalt or diabase containing 50%  $An_{50}$  into a spilite with plagioclase composition of  $An_0$  requires approximately double the amount of sodium assuming there has been no reduction in the modal content of plagioclase. If this additional sodium is not derived from the parent magma by the accumulation of alkali-enriched residual magmatic fluids, then it must be derived from an external source. When the frequency distributions of  $Na_2O$  in spilites and basalts are compared (Vallance 1969, Fig. 1), it is seen that the mean soda content of spilites is greater than that of basalts, suggesting that there has been a gain in sodium in spilites. However, in certain cases, apparent sodium enrichment may be the result of biased sampling of spilitic material, the dangers of which Amstutz (1958) has already indicated.

If sodium-enrichment does occur during the spilitisation process, two conditions must be satisfied: first, there must be a readily available supply of a sodium-bearing fluid; and secondly, this fluid phase must be capable of diffusing into either magmas during their cooling and crystallisation histories, or into solidified lava bodies which become reheated during metamorphism. In the former case, obvious supplies of sodium may be found in trapped sea water within submarine basalt piles, in the pore waters of marine sediments, in hydrothermal solutions which accompany volcanism or from the breakdown of volcanic glass and vitric tuffs under hydrous conditions.

In the latter case, evidence for the diffusion of fluids into magmas and lavas has been proven from the study of oxygen isotopes. Taylor (1968) and Taylor and Forester (1971) have shown on the basis of oxygen isotope ratios from Tertiary stocks, ring dykes, cone sheets and plateau basalts from Skye, Mull and Ardnamurchan, Western Scotland, that almost all the rocks in the vicinity of the central ring complexes are strongly depleted in  $O^{18}$  relative to "normal" igneous rocks from other areas. Their data indicate that very large hydrothermal convection systems involving heated low- $O^{18}$  meteoric ground waters were established in these areas at the time of igneous intrusion. The abnormally low  $O^{18}$  contents of these rocks resulted from diffusion of low- $O^{18}$  meteoric water into the intrusions and overlying plateau basalts at high temperatures (approx.  $500^{\circ}\text{C}$ ), causing isotopic exchange with the granites, gabbros and basalts, thus lowering their  $O^{18}/O^{16}$  ratios. This type of  $O^{18}$  depletion has also been shown to have occurred in a number of areas where Tertiary igneous intrusions have been emplaced into young flat-lying permeable volcanic

rocks e.g. in the margins and upper levels of the Skaergaard intrusion, east Greenland (Taylor 1968), the San Juan Mountains, Colorado (Taylor and Forester, 1971), and the Western Cascade Range, Oregon (Taylor 1970). In addition, hydrogen isotope data have been utilised further to demonstrate the importance of groundwater interaction with the Scottish Hebridean intrusions (Taylor and Epstein, 1968). Additional evidence of isotopic diffusion during metamorphism is given by Taylor and Coleman (1968). They found that six different metasediments and metavolcanics, collected from within 25 metres of one another, all showed almost uniform  $O^{18}/O^{16}$  ratios and identical mineral  $\delta$ -values. Because of the variety of rock types involved, Taylor and Coleman believe it highly unlikely that the original pre-metamorphic  $O^{18}/O^{16}$  ratios of the six rocks were just such that by coincidence after metamorphism all the minerals obtained essentially the same  $\delta$ -values. They conclude that the isotopic similarities must be the result of the metamorphic process itself, and suggest that a very reasonable interpretation is that all the rocks equilibrated with a widespread, ubiquitous oxygen-bearing fluid ( $H_2O$ -rich ?) that permeated the rocks at the time of metamorphism. That the partial pressure of water was high during metamorphism is attested to by the occurrence of abundant hydrous minerals, including muscovite, glaucophane, lawsonite, chlorite and pumpellyite.

Muehlenbachs and Clayton (1972) have recently reported on the oxygen isotope geochemistry of submarine greenstones. They found a range of 2.8 to 6.8‰ in  $\delta O^{18}$  from 14 greenstones, which implies that, at least in some cases, metamorphism in the oceanic crust was not closed with respect to oxygen, and that on the average it tended to lower the  $\delta O^{18}$  of the rocks. Three coarse-grained basic rocks also showed  $O^{18}$  depletion. The

measured oxygen isotope fractionations between coexisting metamorphic minerals in the greenstones gave temperatures of metamorphism in the range 200-300°C, which is in good agreement with independently inferred temperature ranges based on the petrographic work of others (Melson and Van Andel, 1966, Cann 1969, Aumento et al. 1971). Meuhlenbachs and Clayton estimate that water in equilibrium with the metamorphic minerals in this temperature range has a  $\delta O^{18}$  near 0‰. They suggest that this argues strongly that little or no "juvenile" water, or water equilibrated at very high temperatures with rocks of the mantle, which is expected to have a  $\delta O^{18}$  of 7-9‰, participated directly in the metamorphism of the greenstones, and conclude that the oxygen isotope data are consistent with the postulate of participation of sea water in the metamorphism of submarine greenstones.

The oxygen isotope evidence from a variety of igneous rocks indicates that water diffused into cooling magma bodies and into crystalline magma upon metamorphic reheating. The fact that water has diffused into magma does not in itself prove that sodium also diffused into that magma. The diffusion behaviour of sodium may be different from that of water, and it may not diffuse into magma as readily as water. Until evidence is shown to the contrary, it is reasonable to expect that sodium will diffuse into a magma along with water, and thus effect sodium enrichment within the igneous body.

## VI. Summary

Before concluding this chapter of the origins of spilites, it is appropriate and important to summarise the analogies which exist between

spilites and basalts, for the two appear to be genetically related.

In general, a marked similarity in texture exists between spilites and basalts, extending from microscopic to macroscopic features (Amstutz 1968). The similarities are closest for those basalts, such as sub-aqueous types, with heterogeneous textures related to locally variable cooling rates. Such basalts may be fragmented at their margins and tend to be richer in glass than internally. Many spilitic rocks, inferred from their mode of occurrence to have been originally erupted in sub-aqueous environments, also show increasing development of the analogues of crystallised basalt textures away from what appear to have been their initial cooling surfaces.

Despite the textural similarities, the mineral phases stand in marked contrast to those of basalts. The common mineral associations in spilites have been listed by Vallance (1960), the main ones being albite-clinopyroxene-chlorite, albite-chlorite, albite-chlorite-epidote-calcite. The main feature to be emphasised is that the textural positions of calcic plagioclase in basalt are occupied, as a rule, by low temperature albite in spilite; and those of glass by crystalline aggregates commonly rich in chlorite or chlorite plus a hydrous Ca-Al silicate. Olivine is never found in spilites, although apparently pseudomorphous patches of chlorite or carbonate suggest its former presence in some cases. The prior existence of pyroxene is likewise indicated by such textural criteria, however, a significant number of spilites contain fresh clinopyroxene in those sites which can be associated with the pyroxenes of basalts.

Despite the obvious contrasts in mineralogy between spilites and



basalts, comparison of the frequency distribution curves and the mean compositions for the major oxides between spilites and basalts (Vallance 1969, Fig. 1 and Table 1) indicates that the only major differences are found in  $\text{Na}_2\text{O}$ ,  $\text{CaO}$ ,  $\text{H}_2\text{O}^+$  and  $\text{CO}_2$ . Not only do  $\text{Na}_2\text{O}$  values tend to be higher and  $\text{CaO}$  values lower in spilites, but each oxide shows a wider spread of values with much lower percent frequency at their modes than they do in basalts. Arithmetic means for  $\text{SiO}_2$ ,  $\text{Al}_2\text{O}_3$ ,  $\text{TiO}_2$ ,  $\text{MnO}$  and  $\text{P}_2\text{O}_5$  are all close in both spilites and basalts, with  $\text{MgO}$  and  $\text{FeO}$  being slightly lower, and  $\text{Fe}_2\text{O}_3$  slightly higher, in spilites.

Apart from the pyroxenes, formation of all the characteristic phases found in spilites requires conditions lower in temperature and richer in fluids than those prevailing during the crystallisation of the normal phases of basalt. If reliance can be placed on the textural analogies, one is drawn either to expect a genetic relation between spilite and basalt, or to postulate a different parental material for spilite which can, with a different cooling history, produce basaltic textures during low temperature crystallisation and consolidation.

Probably no single hypothesis can account for the origins of all spilites. The existence of a primary spilitic magma, or the separation of a spilitic magma from basaltic melts, has yet to be proven; however, the possible effect of sea water diffusion into a cooling magma is an aspect that is well worth investigation. In some cases, available evidence suggests that volatile enrichment in basic magmas led to the formation of rocks with spilitic compositions. In other cases evidence suggests that post-consolidation changes arising from low grade metamorphism were responsible for the formation of spilites. The important

fact is that all these major processes are convergent in that the resulting mineral products tend to be similar. Whether the controls are magmatic, or local as in deuteric alteration, or more widespread as in metamorphism, all the rocks at a given horizon will be affected, and convergent mineralogies will result. On deciding the origin of any particular group of spilites the field, chemical, mineralogical and textural evidence obtained from these rocks should be carefully scrutinised to see which, if any, of the above processes appears to be the most likely one to have given rise to these spilites.

The definition of a spilite should now be reconsidered. The great diversity found in chemistry and mineralogy would indicate that a restrictive definition is not desirable. Because of the heterogeneity found in spilites, particularly evident in spilitic pillow lavas, it does not seem justified to restrict the term only to those portions of the rock which are characterised by albite, clinopyroxene and chlorite. Portions of spilites rich in albite, in albite and chlorite, in chlorite and epidote, and so on, exist in close proximity and are intimately related in origin to those "spilitic" portions. If spilite is to be retained as a petrographic term, it is better that it be used in a general sense to include all the rock variants which have been derived from a common basaltic parent, but which still retain their igneous textural characteristics. One could then talk of albite-clinopyroxene-chlorite spilites, albite-chlorite spilites, albite-epidote spilites, albite-rich spilites etc., rather than restrict the term to a part of the overall assemblage.

## CHAPTER 3

### GENERAL PETROGRAPHY AND CHEMISTRY

#### I. Introduction

This chapter is primarily divided into two parts, the first part dealing with the general petrography of the Olympic Peninsula volcanic rocks, and the second part dealing with the chemistry of these rocks. In the summary, the petrographic and chemical criteria are used to interpret the origin of the Olympic Peninsula volcanic rocks, and to present a picture of the type and degree of alteration and metamorphism which has affected these rocks.

#### II. General Petrography

The general petrography of the major volcanic rock types found on the Olympic Peninsula will be discussed as follows: basalts, pillow basalts, diabases and basaltic breccias and tuffs.

##### 1. Basalts

The basalts are largely aphanitic, dark green rocks rich in glass or devitrified glass, with textures ranging through variolitic, intersertal, hyalopilitic and intergranular. The majority of basalts contain glass or devitrified glass in their groundmass, although holocrystalline varieties also occur. Plagioclase, clinopyroxene and titanomagnetite are the common mineral phases present, with relict olivine present in

less than half of the observed rocks.

Plagioclase may occur as microlites in the glassy groundmass, as skeletal microphenocrysts, or occasionally as subhedral to euhedral phenocrysts. Determination of the chemical composition of the plagioclase by optical methods was made difficult by the extremely small grain size of the plagioclase crystals, the lack of suitable twins, the effects of alteration and the incomplete crystallisation of the plagioclase from the glassy groundmass. Where plagioclase compositions could be determined, they fell within the range expected for normal basalts. The colourless to pale brown clinopyroxene occurs as subhedral microphenocrysts and in feathery intergrowths with plagioclase, and is generally unaltered, although minor replacement by pale green actinolite does occur. Olivine is never abundant when present, occurring as both euhedral and skeletal microphenocrysts. It is invariably pseudomorphed by green bowlingite, chlorite or calcite, or mixtures of these phases. Small, euhedral red-brown spinels are often found poikilitically enclosed within the pseudomorphed olivines. Titanomagnetite (identified on the basis of 6 polished sections) is common throughout, occurring either as subhedral or skeletal grains or as dense specks and rods in the groundmass.

The groundmass may be represented by a dark brown, almost isotropic glass, or a lighter brown, partly spherulitic glass. Where devitrification has been more complete, the glass has crystallised into spherulitic or feathery intergrowths of plagioclase and pyroxene, the outlines of which are often rimmed by opaque minerals or haematite. Vesicles, when present, are generally filled with calcite or chlorite, or mixtures of

both. Quartz and prehnite are also found in some vesicles. Zeolitic material was found in only a few of the basalts in this group. Chlorite is invariably present, either in veinlets, vesicles or replacing glass, olivine, pyroxene and plagioclase. Calcite veinlets, and to a lesser extent quartz veinlets, are common, and prehnite, with rare pumpellyite, is found in association with the calcite veinlets. Interstitial quartz is present in several basalts, and may be related to quartz veinlets in some instances.

## 2. Pillow Basalts

Serial sections were cut across several small basaltic pillows ranging in diameter from approximately 10 to 30 cms. Generally two zones were apparent: an outer, largely glassy zone, and an inner more crystalline one. The outer glassy zones consist of dark brown glass, often spherulitic and charged with opaque minerals. Haematite staining is common. In the smaller pillows the inner zones consist of spherulitic or feathery intergrowths of plagioclase and pyroxene, while in the more crystalline inner zones of the larger pillows incipient variolitic textures with radial and sheaf-like clusters of plagioclase are more common.

Plagioclase, clinopyroxene and olivine microphenocrysts are found in the pillows, olivine being absent in only one of the examined pillows. Plagioclase microlites and microphenocrysts, often with a skeletal habit, usually have some degree of alteration and replacement. Replacement of the microlites, and to a lesser degree the microphenocrysts, by chlorite is found, along with kaolinitisation and sericitisation. Small patches

of chlorite within the plagioclase may represent replacement of former fluid inclusions. No identifiable saussuritisation of the plagioclase was observed. Complete replacement of some plagioclase by zeolitic material was observed in one pillow, with partial replacement by prehnite in all pillows being fairly common. In some pillows, plagioclase microphenocrysts are partially or wholly replaced by a mosaic of small, granular, dark grey and white grains, possibly indicating breakdown to zeolite and prehnite. Colourless to pale brown, subhedral clinopyroxene microphenocrysts are quite fresh in appearance, and exhibit a tendency to form in clusters. Euhedral and skeletal olivine microphenocrysts are present, with the invariable replacement as in the normal basalts.

Calcite veining is common, with prehnite often found in association, and both minerals are also present in vesicles. Chlorite is common in vesicles and veinlets, and also replacing glass. Quartz is found in vesicles, veinlets and interstitial patches but is never abundant. Zeolitic material is scarce, being found in only one pillow, where it occurs in vesicles and replacing plagioclase. Rare pumpellyite (?) may be present in one pillow.

### 3. Diabases

Because it was not always possible to tell from the field relations whether medium-grained rocks represent intrusive bodies or the coarser grained sections of thick lava flows, the term diabase, as used in the following discussions, refers to a basic igneous rock having an average grain size between 5mm and 3 cm. The diabases are largely medium-grained

rocks exhibiting subophitic textures. The two distinct types of diabases which have been found are a normal diabase containing calcic plagioclase, and an albite diabase containing sodic plagioclase. Both types of diabase have been altered in varying degrees, the albite diabases generally more severely. In normal diabase the plagioclase is generally quite fresh, although some kaolinitisation and sericitisation is invariably present. In albite diabase the plagioclase is commonly affected by kaolinitisation and sericitisation. Replacement by sericite, and in some cases prehnite and more rarely chlorite, is complete at times. Partial replacement by zeolitic material is occasionally found, with laumontite being recorded in one case. Inclusions in the plagioclase are numerous, consisting mostly of chlorite with an occasional patch of epidote. The clinopyroxenes, often pinkish in colour, are generally quite fresh in both types of diabase. They occur both in intergranular and sub-ophitic relationships with the plagioclase grains. Partial replacement along grain edges and cracks by green chlorite is common and occurs more frequently in the albite diabases. Occasionally, replacement at grain edges or in small internal patches is by a brown hornblende, or more rarely by pale green actinolite, probably the result of deuteric alteration. Fresh olivine was not found in any of the diabases, but chlorite pseudomorphs after olivine suggest its former presence in some diabases. Skeletal and subhedral titanomagnetites are present in all diabases, and exhibit varying degrees of alteration, probably to sphene. In general, the more severely altered titanomagnetites are found in the more altered diabases, but exceptions exist. Green chlorite is ubiquitous throughout both types of diabase, occurring most abundantly as interstitial

patches. It is also present in veinlets and replacing olivine, pyroxene and plagioclase to varying degrees. Prehnite is fairly common and is found associated with calcite and chlorite veins, and also replacing plagioclase. Epidote is quite rare, generally occurring as small veinlets or patches, and pumpellyite was recorded in one diabase, associated with calcite veinlets. Apatite is a common accessory mineral.

#### 4. Breccias and Tuffs

The volcanic breccias are largely represented by two types: hyaloclastite breccias, and mixed breccias. The hyaloclastite breccias are composed to a large degree of angular glass fragments which exhibit various stages of hydration and alteration. The freshest fragments consist of dark brown, almost isotropic glass, which may have devitrified into brown spherulitic aggregates. Hydration has resulted in the fragments being rimmed by yellow-brown palagonite, with complete palagonitisation in some cases. The most highly altered fragments have been completely replaced by a bluish-green chlorite, except for several fragments which appear to have been replaced by an intimate mixture of microgranular quartz and chlorite.

The mixed breccias consist of varying proportions of glassy basalt fragments, glass fragments and feldspar and pyroxene crystals. The glass fragments exhibit similar alteration features to those observed in the hyaloclastite breccias. The basaltic fragments consist of dark, glassy fragments, often showing fluidal orientation of plagioclase microlites. The glassy matrix may be fresh or totally replaced by a green, often



spherulitic, chlorite. The feldspar crystals are more abundant than the pyroxene crystals, and are generally kaolinitised and sericitised, with some replacement by prehnite. The pyroxenes are quite fresh in appearance, but somewhat shattered. Calcite and chlorite are common vein minerals in both types of breccias, and act as cementing agents in several cases. Prehnite and pumpellyite are associated with calcite in veinlets.

Three tuffaceous rocks have been identified in the sample collection. Two of these tuffs were collected from a locality where the strata is reported overturned, and petrographic evidence indicates that they have suffered shearing. One sample consists of a heterogeneous mixture of ragged, pale green actinolite and partially sericitised albite grains distributed throughout a streaky, kaolinitised groundmass. The other sample has been thoroughly epidotised, and now consists of a heterogeneous mixture of epidote and actinolite grains, with the occasional sericitised feldspar, set in a similar streaky groundmass. Prehnite is present in vein form in both rocks, and is found replacing feldspar. The third rock is a heterogeneous crystal tuff composed of shattered pyroxenes and sericitised feldspar, set in a chloritised groundmass. Prehnite is again present replacing feldspar, with pumpellyite present in vein form.

### III. Chemistry

In Table 1, 26 chemical analyses of volcanic rocks from the Olympic Peninsula are presented. Table 2 represents these analyses recalculated

to 100% on a total H<sub>2</sub>O and CO<sub>2</sub> - free basis. Table 3 and 4 represent the analyses recalculated to 100% on a total H<sub>2</sub>O and CO<sub>2</sub>, and total H<sub>2</sub>O and CaCO<sub>3</sub> - free basis, respectively, with iron oxide values recalculated by the method of Irvine and Baragar (1971). The major oxides and normative constituents are given in weight per cent. Analyses 32C, M and E and 10C, M and E represent serial analyses from the centres to the edges of small basaltic pillows. Each rock is designated by the field number allocated to it at the time of collection, except that the prefix "SP 69/S" has been dropped e.g. rock 5 corresponds to field number SP 69/S 5. Petrographic descriptions of analysed rocks and analytical techniques are given in Appendices A and B.

## 1. Possible Changes in Chemistry during Metamorphism

### A. The Basalt Screen

In view of the alteration and metamorphism which appears to have affected the Olympic Peninsula volcanic rocks, the question of addition or removal of chemical components, or their redistribution, in these rocks should not be overlooked. To obtain estimates of possible loss or gain of chemical components on these rocks, the chemical analyses were subjected to the basalt screen devised by Manson (1967), which is reproduced in Table 5. The basalt screen is a chemical screen test such that 33 individual requirements (16 based on chemical analyses, 11 on the corresponding norms, and the final 6 on parameters calculated from the norms) have to be satisfied before any rock may be considered to be chemically and normatively a basalt. The screen, therefore, defines limits for the chemical variation in basaltic rocks. By subjecting the chemical analyses

of the Olympic Peninsula volcanic rocks to this chemical screen, it should be possible to discover if, and where, any major variations from basalt chemistry occur.

By employing this chemical screen, the Olympic Peninsula volcanic rocks are treated as if they were originally normal basaltic rocks, without consideration being given to the possibility that some of these rocks may not have been originally basalts, or basalts which may have had unusual, but original, features of chemistry. The implications of the decision to use the basalt screen, despite the above possibilities, are fully realised. On the basis of the observed petrography of the Olympic Peninsula volcanic rocks, and on the fairly liberal limits of the basalt screen, particularly with regard to sodium, it is considered that the application of the basalt screen is still a valid procedure for determining possible changes in chemistry in the Olympic Peninsula volcanic rocks.

The results of the basalt screen applied to the analysed rocks (analyses "as is", Table 1) are tabulated in Table 6. Only 7 rocks (13, 1411, 16, 38, 43, 47, 69) managed to pass the screen. Of the remaining 19 rocks, 10 failed because of excess total  $H_2O$ ,  $H_2O^+$ ,  $CO_2$  or  $Fe_2O_3$ , or various combinations of these factors; and of the 9 remaining rocks, 3 failed because of low crystallisation Indices (Poldervaart and Parker, 1964) and/or Anorthite values\* (rocks 171, 23 and 80). In rocks 30 and 60 the relatively high  $K_2O$  values (1.0% and 1.2% respectively) indirectly account for the presence of the excess olivine. In rock 30, high  $K_2O$  has indirectly resulted in production of nepheline in the norm causing

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\* Crystallisation indices and Anorthite values are reported in Table 7.

conversion of provisional hypersthene into olivine, thus increasing olivine above the screen limits. Similarly with rock 60, although no nepheline appears in the norm, the relatively high alkalis have consumed much of the available silica, thus causing olivine to form from other provisional ferromagnesian minerals. Rocks 171, 23 and 80 all have Crystallisation Indices below the required minima of Table 5, and rocks 171 and 80 also have An values below the required minima. These failures are directly attributable to the high alkali contents found in these rocks (Table 1), which result in increased production of normative albite.

The remaining 4 rocks in the group of 19 (rocks 5, 8 48 and 57) provide more direct evidence of chemical changes, since the causes of their failing to pass the screen result directly from their unusual chemical compositions. Both rocks 5 and 48 have unusually low CaO values for basaltic rocks, even although petrographic evidence indicates that they were formerly basaltic in character. The fact that their other major oxides (with the exception of H<sub>2</sub>O and CO<sub>2</sub> in rock 5, and H<sub>2</sub>O in rock 48) fall within the range for normal basaltic rocks strongly suggests considerable loss in Ca from these particular rocks. Rocks 8 and 57 failed to pass the screen because of excess Na<sub>2</sub>O, with rock 8 also having an unusually low MgO content. The extremely fine-grained nature of rock 57 (Appendix A) makes the detection of soda addition difficult, but the high soda content is suggestive of gain. Rock 8 is an albite diabase (Appendix A), and on the assumption that the original plagioclase was approximately An<sub>50</sub>, the rock must now contain approximately double its original soda content, for petrographic evidence indicates no obvious reduction in the absolute amount of the original plagioclase. It is

impossible to say whether this increased soda resulted from late magmatic deuteric alteration, or by hydrothermal addition from external fluids. Whatever agent was responsible, it may also have been responsible for the removal of some MgO from this rock.

#### B. Additional Evidence of Changes in Chemistry

During low-grade metamorphism, such as that which appears to have affected the Olympic Peninsula volcanic rocks, calcium is released during the breakdown of calcic plagioclase and may be precipitated as  $\text{CaCO}_3$  under suitable physico-chemical conditions. If the calcium is precipitated immediately within the parent rock then no overall change in bulk chemical composition will occur. If, however, the calcium is removed in a fluid phase before precipitation, then depletion of calcium occurs in the parent rocks, with possible calcium enrichment in nearby rocks.

Many of the analysed rocks contain calcite in scattered patches, in vesicles and in veinlets, thus the possibility exists that calcium enrichment, from a source external to the analysed sample, may have occurred in these rocks. To test for this possible enrichment, the analyses were recalculated on a  $\text{CaCO}_3$  - free basis (Table 4). If an increase in calcium has occurred in a particular rock, as a result of calcite addition from an external source, then it should be indicated by a significant decrease in CaO in the analysis of that particular rock recalculated  $\text{CaCO}_3$  - free. Comparison of the analyses of Tables 3 and 4 indicates that for the majority of analysed rocks, the analyses recalculated on a  $\text{CaCO}_3$  - free basis differ only slightly from the analyses recalculated on a  $\text{CO}_2$ -free basis. This suggests that there has been little

or no gain in calcium in the majority of analysed rocks. In rocks such as 5, 77, 32C, M and E and 10C, M and E, all of which contain considerable levels of  $\text{CO}_2$ , significant decreases in CaO do occur in their analyses recalculated on a  $\text{CaCO}_3$ -free basis. These rocks all contain scattered calcite patches and veinlets, the veinlets being particularly well represented in rocks 32 and 10. These facts suggest that calcium enrichment, resulting from externally derived calcium, has occurred in these particular rocks.

## 2. Major Element Chemistry and Normative Compositions

When considering the chemical variation and normative compositions of rocks, it is extremely important that the chemical analyses of these rocks represent, as closely as possible, the original chemical composition when the rocks crystallised and solidified, since post-magmatic changes in chemistry will cause changes in the proportions of chemical components, and thus in the proportions of normative constituents. The results of the basalt screen (Table 6) have shown that excess  $\text{H}_2\text{O}^+$ , total  $\text{H}_2\text{O}$ ,  $\text{CO}_2$  or  $\text{Fe}_2\text{O}_3$ , or combinations of these factors, were the main reasons why Olympic Peninsula volcanic rocks failed to pass the screen. In view of the alteration and metamorphism which appears to have affected the Olympic Peninsula volcanic rocks, these departures from normal basaltic chemistry can probably be regarded as products of post-magmatic processes; thus, it is unlikely that the chemical analyses of Table 1 are totally representative of the original chemical composition of the Olympic Peninsula volcanic rocks. Therefore, it is desirable that the analyses of the Olympic Peninsula volcanic rocks be presented in such a manner that

these post-magmatic effects, notably increased  $H_2O$ ,  $CO_2$  and  $Fe_2O_3$ , be reduced or eliminated as far as possible.

In order that the chemical analyses of the Olympic Peninsula volcanic rocks represent the original chemical compositions as closely as possible, the analyses have been recalculated to 100% on a total  $H_2O$  and  $CaCO_3$ -free basis, and the iron oxide values have been adjusted according to the method of Irvine and Baragar (1971). The results of such recalculations are given in Table 4. Recalculation of the analyses on a  $CaCO_3$ -free basis, rather than a  $CO_2$ -free basis, is preferred since this calculation takes into account the likely addition of  $CaO$  to the rock when  $CO_2$  levels are high.

In basalts, the effect of oxidation on ferrous iron tends to increase normative hypersthene at the expense of normative olivine, thus, any rock which has suffered oxidation would be expected to portray a more oversaturated normative composition. The results of the basalt screen (Table 6) indicate that the Olympic Peninsula volcanic rocks have been oxidised; therefore, their iron oxide values have been adjusted to give a closer representation of the proportions of normative hypersthene and olivine. Coombs (1963) considers 1.5%  $Fe_2O_3$  to be a theoretically acceptable figure for basalts, any excess in  $Fe_2O_3$  being recalculated as  $FeO$ . Such an adjustment will give a slightly more undersaturated normative composition relative to the composition which would have been obtained if the unconverted iron oxide values had been used. The acceptable value for  $Fe_2O_3$  reported by Coombs was not used as the basis of recalculations in the analyses of Table 4, for the reason that adopting a uniform  $Fe_2O_3$

value throughout implies an unjustifiably variable oxidation state in all the analysed rocks. Instead, the method of Irvine and Baragar (1971) was used to recalculate the iron oxide values, in which an upper limit of  $\text{Fe}_2\text{O}_3$  is set according to the following equation:

$$\% \text{Fe}_2\text{O}_3 = \% \text{TiO}_2 + 1.5$$

Their equation is based on the observation that  $\text{Fe}_2\text{O}_3$  and  $\text{TiO}_2$  generally have similar trends of variation in unaltered volcanic rocks. If the analysed value for  $\text{Fe}_2\text{O}_3$  is less than the calculated value, no change in the iron oxide values are made, however, if it is greater, the "excess"  $\text{Fe}_2\text{O}_3$  is converted to  $\text{FeO}$ .

When the recalculated analyses of Table 4 are put through the basalt screen, 18 rocks now pass the screen compared with 7 rocks which originally passed the screen. It appears, therefore, that the recalculated analyses of Table 4 best represent the original chemical composition of the Olympic Peninsula volcanic rocks. In the remainder of this chapter and in following chapters, all chemical variation diagrams, normative compositions and various data projection schemes will be based on the analyses of Table 4, unless otherwise stated.

#### A. Major Element Chemistry

The major element variation in the Olympic Peninsula volcanic rocks is depicted on a plot of weight per cent oxides against weight per cent silica (Fig. 6), and on a total alkalis versus silica diagram (Fig. 7). All plots are based upon the analyses recalculated to 100% on a total



H<sub>2</sub>O and CaCO<sub>3</sub>-free basis, with adjusted iron oxide values (Table 4).

It is apparent from Fig. 6 that no strong trends are present in the plots of the major oxides versus silica, with considerable scatter present in the plots for MgO, FeO, CaO, Na<sub>2</sub>O and K<sub>2</sub>O. Similarly, considerable scatter occurs on the alkali-silica diagram (Fig. 7). The boundary line constructed by Macdonald and Katsura (1964) separates the fields between Hawaiian alkalic and tholeiitic basalts, and although its position varies somewhat for different volcanic provinces, Yoder (1967) considers that the boundary is diagnostic within narrow limits. On the basis of normative compositions (to be discussed), the great majority of Olympic Peninsula volcanic rocks have been classified as tholeiitic, yet over half of the analysed rocks plot on the alkalic side of the Macdonald and Katsura boundary line. A considerable shift in the position of the boundary line, greater than what can be regarded as "within narrow limits", is required before the majority of analysed rocks would fall within the tholeiitic field. The approximate position of a line, below which no Olympic Peninsula alkali basalts plot, has been constructed on Fig. 7. Rocks 34 and 10C, containing .52 and .44 per cent nepheline, respectively, do actually plot below this line, but the appearance of nepheline in the norms of these rocks at such low levels may be regarded as within the limits of analytical error.

#### B. Normative Compositions

Discussion of the normative compositions of the Olympic Peninsula volcanic rocks is based upon the analyses recalculated to 100% on a total

H<sub>2</sub>O and CaCO<sub>3</sub>- free basis, with adjusted iron oxide values (Table 4).

The majority of analysed rocks contain hypersthene in the norm and can therefore be classified as tholeiitic in the terminology of Yoder and Tilley (1962). Rocks 5, 8 and 91 contain quartz in the norm. Rock 91 is fresh and unaltered; the presence of quartz veinlets in rocks 5 and 8, and also the unusually low CaO content of rock 5, are probably responsible for the appearance of quartz as a normative constituent in these rocks. The remaining rocks - 30, 34, 57, 77 and 10C - are nepheline normative. In rocks, 30, 34 and 10C nepheline is less than one per cent in each case, and its appearance at these levels may be regarded as within the limits of analytical error. Rocks 30 and 34 also have high K<sub>2</sub>O values (Table 4) and consequently high normative orthoclase, the formation of which would tend to deplete available silica in the norm calculations. The relatively low SiO<sub>2</sub> accounts for the appearance of normative nepheline in rock 10C. Rocks 57 and 77 have nepheline values of 1.77 and 3.05 per cent respectively, and may represent true alkali basalts. It should be noted, however, that rock 57 has a very high Na<sub>2</sub>O value (6.3%), and that rock 77 has high FeO (14.1%) and CaO (10.5%) values.

#### IV. Summary

The presence of pillow basalts, intercalated marine sediments, abundant volcanic glass and hyaloclastite material suggest that submarine basalts, breccias and tuffs comprise at least part of the Olympic Peninsula volcanic sequence. The majority of basalts are aphanitic, with a primary

mineral assemblage of microphenocrysts of plagioclase and clinopyroxene, with titanomagnetite, set in a glassy and often chloritised groundmass. Pseudomorphed olivines are present in some of the basalts, but are never abundant. Plagioclase compositions are difficult to determine optically, but where grains are large enough and fresh enough, the compositions fall in the range for normal basaltic rocks. Diabasic rocks, probably intrusive, are present as two distinct types, namely, normal diabases with calcic plagioclase, and albite diabase with plagioclase compositions in the range  $An_0 - An_{10}$ .

Virtually all the volcanic rocks have suffered some degree of alteration and metamorphism. The most common form of change is the replacement of glass and groundmass by chlorite, and the sericitisation or kaolinitisation of plagioclase. In other cases, the presence of zeolites and partial, or complete replete replacement of plagioclase by zeolitic material, and the presence of prehnite and pumpellyite indicate that metamorphism has attained the zeolite facies and prehnite-pumpellyite facies (Coombs et al. 1959, Coombs 1960, 1961) over parts of the Olympic Peninsula. Under these conditions of metamorphism, chemical components, notably Na, K and Ca, may become mobile (Smith 1968; Hekinian 1971) and cause enrichment or depletion within the parent rocks or surrounding rocks. The observed scatter in the variation diagrams (Fig. 6) and the alkali-silica diagram (Fig. 7) may, in part, be the result of redistribution of chemical components under these metamorphic conditions.

Evidence of changes in the original chemistry of the Olympic Peninsula volcanic rocks has been obtained from the results of the basalt screen, where

high levels of  $H_2O$ ,  $CO_2$  and  $Fe_2O_3$  appear to be the most common changes involved. Evidence of greater departures from normal basalt chemistry is indicated in cases where there appears to have been addition of Na, K and Ca, and in some cases, loss of Ca and Mg. For the most part, large changes in chemistry appear to be the exception rather than the rule, but within the fairly liberal limits of the basalt screen, particularly with regard to sodium, loss or gain of chemical components can go undetected.

The basalt screen has indicated that the chemical analyses recalculated to 100% on a total  $H_2O$  and  $CaCO_3$ -free basis, with adjusted iron oxide values, best represent the original chemical compositions of the Olympic Peninsula volcanic rocks. On the basis of these recalculated analyses, chemical variation diagrams indicated that no strong trends were present in the analysed rocks, and that considerable scatter existed in several major oxides. Normative calculations, also on the basis of the recalculated analyses, indicates the majority of analysed rocks to be tholeiitic. The appearance of nepheline in the norms of certain rocks can be explained in terms of their unusual chemistry, although two of these rocks may represent true alkali basalts.

Tables 1, 2, 3 and 4.	Chemical Analyses and Norms.
Table 1.	Analyses "as is".
Table 2.	Analyses recalculated to 100% on a total H <sub>2</sub> O and CO <sub>2</sub> -free basis.
Table 3.	Analyses recalculated to 100% on a total H <sub>2</sub> O and CO <sub>2</sub> -free basis, with adjusted iron oxide values.
Table 4.	Analyses recalculated to 100% on a total H <sub>2</sub> O and CaCO <sub>3</sub> -free basis, with adjusted iron oxide values.
Table 5.	The basalt screen. From Manson (1967, Table II).
Table 6.	Basalt screen results for Olympic Peninsula volcanic rocks.
Table 7.	Crystallisation indices and Anorthite values for Olympic Peninsula volcanic rocks.









TABLE 1

	32M	32E	10C	10M	10E
SI02	46.5	48.6	44.2	49.8	50.9
AL203	13.6	14.8	12.2	13.0	13.7
FE203	5.2	4.1	6.1	4.9	4.4
FEO	3.7	4.4	6.2	7.2	7.8
CAO	13.0	10.4	14.0	10.4	9.7
MGO	5.0	6.9	6.9	7.4	7.5
NA2O	3.9	2.8	2.9	3.2	3.1
K2O	.62	1.5	.25	.30	.38
TI02	1.1	1.1	.69	.74	.75
P2O5	.27	.26	.09	.05	.04
MNO	.15	.16	.20	.22	.22
CO2	3.4	1.2	3.2	1.2	.54
H2O+	2.4	2.6	2.7	1.7	1.7
H2O-	.23	.40	.22	.24	.25
TOTAL	99.1	99.2	99.8	100.3	101.0
Q	.84	.68	0.00	1.42	1.33
OR	3.80	9.22	1.53	1.80	2.27
AB	34.21	24.62	25.31	27.51	26.48
AN	18.43	24.30	20.15	20.55	22.57
NE	0.00	0.00	0.00	0.00	0.00
DI	19.09	15.58	24.05	19.25	18.10
HY	4.98	13.78	7.68	17.92	20.04
OL	0.00	0.00	3.08	0.00	0.00
MT	7.82	6.18	9.12	7.22	6.44
IL	2.17	2.17	1.35	1.43	1.44
HM	0.00	0.00	0.00	0.00	0.00
AP	.65	.63	.21	.12	.09
C	0.00	0.00	0.00	0.00	0.00
CC	8.02	2.84	7.51	2.77	1.24
WO	0.00	0.00	0.00	0.00	0.00
TOTAL	100.00	100.00	100.00	100.00	100.00







TABLE 2

	32M	32E	10C	10M	10E
SI02	50.0	51.1	47.2	51.2	51.7
AL203	14.6	15.6	13.0	13.4	13.9
FE203	5.6	4.3	6.5	5.0	4.5
FEO	4.0	4.6	6.6	7.4	7.9
CAO	14.0	10.9	14.9	10.7	9.8
MGO	5.4	7.3	7.4	7.6	7.6
NA2O	4.2	2.9	3.1	3.3	3.1
K2O	.67	1.6	.27	.31	.39
TI02	1.2	1.2	.74	.76	.76
P2O5	.29	.27	.10	.05	.04
MNO	.16	.17	.21	.23	.22
TOTAL	100.1	99.9	100.0	99.9	99.9
Q	0.00	0.00	0.00	0.00	.94
OR	3.96	9.47	1.60	1.84	2.31
AB	26.32	24.55	15.96	27.93	26.25
AN	18.99	24.84	20.76	20.85	22.88
NE	4.97	0.00	5.56	0.00	0.00
DI	30.91	21.78	42.03	25.79	20.69
HY	0.00	7.98	0.00	13.91	18.86
OL	0.00	2.24	3.04	.87	0.00
MT	8.11	6.24	9.42	7.25	6.53
IL	2.28	2.28	1.41	1.44	1.45
HM	0.00	0.00	0.00	0.00	0.00
AP	.67	.63	.23	.12	.09
C	0.00	0.00	0.00	0.00	0.00
CC	0.00	0.00	0.00	0.00	0.00
WO	3.80	0.00	0.00	0.00	0.00
TOTAL	100.00	100.00	100.00	100.00	100.00









TABLE 3

	32M	32E	10C	10M	10E
SI02	50.1	51.3	47.4	51.4	51.8
AL203	14.7	15.6	13.1	13.4	13.9
FE203	2.8	2.7	2.4	2.3	2.3
FEO	6.5	6.0	10.4	9.9	9.9
CAO	14.0	11.0	15.0	10.7	9.9
MGO	5.4	7.3	7.4	7.6	7.6
NA2O	4.2	3.0	3.1	3.3	3.2
K2O	.67	1.6	.27	.31	.39
TIO2	1.2	1.2	.74	.76	.76
P2O5	.29	.27	.10	.05	.04
MNO	.16	.17	.21	.23	.22
TOTAL	100.0	100.1	100.1	99.9	100.0
Q	0.00	0.00	0.00	0.00	0.00
OR	3.96	9.45	1.59	1.83	2.31
AB	20.92	25.35	10.82	27.93	27.07
AN	19.28	24.34	21.01	20.85	22.41
NE	7.91	0.00	8.33	0.00	0.00
DI	39.27	22.84	43.40	26.25	21.73
HY	0.00	3.29	0.00	9.35	15.98
OL	1.65	7.92	9.74	8.89	5.63
MT	4.06	3.91	3.47	3.34	3.33
IL	2.28	2.28	1.40	1.44	1.44
HM	0.00	0.00	0.00	0.00	0.00
AP	.67	.63	.23	.12	.09
C	0.00	0.00	0.00	0.00	0.00
CC	0.00	0.00	0.00	0.00	0.00
WO	0.00	0.00	0.00	0.00	0.00
TOTAL	100.00	100.00	100.00	100.00	100.00







TABLE 4

	32M	32E	10C	10M	10E
SI02	52.6	52.1	49.5	52.2	52.2
AL203	15.4	15.9	13.7	13.6	14.0
FE203	2.9	2.8	2.5	2.3	2.4
FEO	6.8	6.1	10.9	10.1	9.9
CAO	9.8	9.5	11.1	9.3	9.2
MGO	5.7	7.4	7.7	7.8	7.7
NA2O	4.4	3.0	3.3	3.4	3.2
K2O	.70	1.6	.28	.31	.39
TI02	1.2	1.2	.77	.78	.77
P2O5	.30	.28	.10	.05	.04
MNO	.17	.17	.22	.23	.23
TOTAL	100.0	100.0	100.1	100.1	100.0
Q	0.00	0.00	0.00	0.00	0.00
OR	4.14	9.46	1.66	1.83	2.31
AB	37.24	25.37	27.09	28.75	27.07
AN	20.21	25.18	21.73	20.92	22.68
NE	0.00	0.00	.44	0.00	0.00
DI	21.53	16.17	26.91	20.43	18.63
HY	3.17	14.21	0.00	17.68	21.34
OL	6.53	2.64	16.86	5.46	2.94
MT	4.21	4.06	3.62	3.33	3.48
IL	2.28	2.28	1.46	1.48	1.46
HM	0.00	0.00	0.00	0.00	0.00
AP	.70	.65	.23	.12	.09
C	0.00	0.00	0.00	0.00	0.00
CC	0.00	0.00	0.00	0.00	0.00
WO	0.00	0.00	0.00	0.00	0.00
TOTAL	100.00	100.00	100.00	100.00	100.00

TABLE 5

## Chemical screen for basaltic rocks

	More than	Less than	
SiO <sub>2</sub>	*	56.00	
TiO <sub>2</sub>	*	5.50	
Al <sub>2</sub> O <sub>3</sub>	10.50	22.00	
Fe <sub>2</sub> O <sub>3</sub>	*	6.00	
FeO	2.50	15.00	
MnO	*	1.00	
MgO (FeO < 10%)	3.00	*	
MgO (FeO > 10%)	2.00	*	
CaO	5.00	15.00	
Na <sub>2</sub> O	*	5.50	
P <sub>2</sub> O <sub>5</sub>	*	1.50	
total H <sub>2</sub> O	*	4.00	
H <sub>2</sub> O <sup>+</sup>	*	3.00	
H <sub>2</sub> O <sup>-</sup>	*	1.00	
CO <sub>2</sub>	*	0.50	
Analysis total	99.00	101.00	
q	*	12.50	
or (ne = 0)	*	15.00	
or (ne < 10%)	*	20.00	
or (ne > 10%)	*	25.00	
No c			
No wo			
ol (ne = 0)	*	15.00	
ol (ne < 10%)	*	20.00	
ol (ne > 10%)	*	25.00	
ol (lc > 0)	*	25.00	
No hm			
CI (FeO < 10%) (Crystallization index, Poldervaart and Parker, 1964)	35.00	70.00	
CI (FeO > 10%)	25.00	70.00	
An [100 × an/(an + ab)] (FeO < 10%; or < 10%; lc = 0)	35.00	80.00	-An 1
An (FeO > 10%; or < 10%; lc = 0)	25.00	80.00	-An 2
An (FeO < 10%; or > 10%; lc = 0)	35.00	90.00	-An 3
An (lc > 0)	*	100.00	-An 4

\* No limit specified.



TABLE 7

## Crystallisation Indices and Anorthite Values

Rock No.	Crystallisation Index (CI)*	Anorthite Value (An)
5	30.0	43.8
8	27.4	25.1
13	48.0	52.9
1411	47.8	52.3
16	53.1	57.9
171	33.3	27.2
23	34.7	35.4
30	50.1	46.3
34	54.5	49.9
38	40.9	39.7
40	47.1	57.3
43	53.9	59.4
47	47.9	57.8
48	28.5	20.3
57	23.0	10.9
60	53.6	57.7
69	40.5	41.5
77	38.4	28.9
80	33.0	29.8
91	48.0	56.2
32C	42.9	36.2
32M	39.6	35.0
32E	45.9	49.7
10C	45.2	44.3
10M	43.0	42.8
10E	44.2	46.0

\* Crystallisation Index - CI (Poldervaart and Parker, 1964)

CI = (An + Di' + Fo' + Sp') where

An = normative anorthite

Di' = magnesian diopside,  $\text{CaMgSi}_2\text{O}_6$ , calculated from normative diopside

Fo' = normative forsterite plus normative enstatite converted to forsterite

Sp' = magnesian spinel,  $\text{MgAl}_2\text{O}_4$ , calculated from normative corundum in ultramafic rocks



Fig. 6. Major element variation diagrams. Weight per cent oxides versus weight per cent silica. Analyses from Table 4.

Fig. 7. Alkali-silica diagram. Total alkalis ( $\text{Na}_2\text{O} + \text{K}_2\text{O}$ ) versus silica, in weight per cent. Analyses from Table 4.

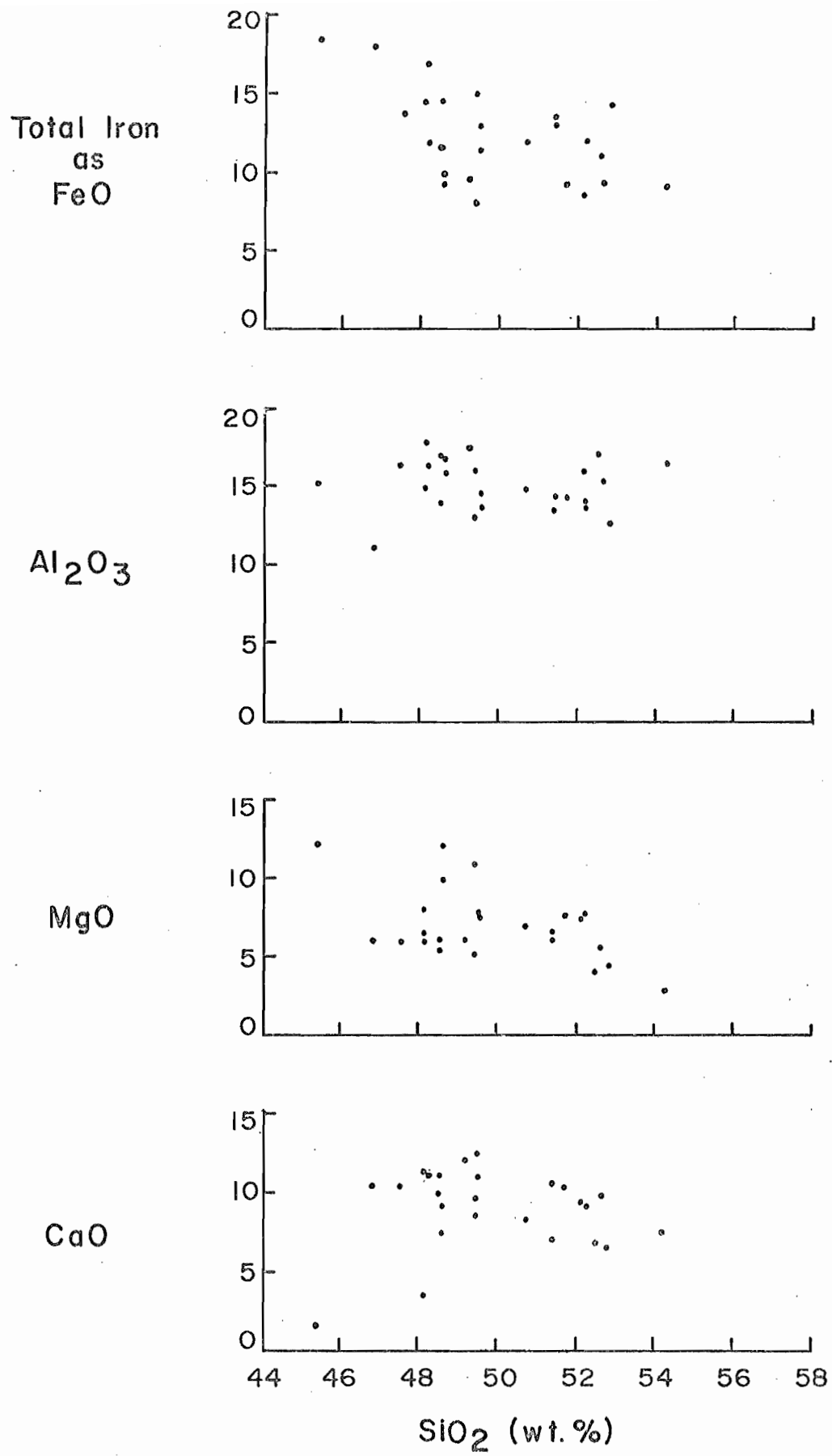


Fig. 6

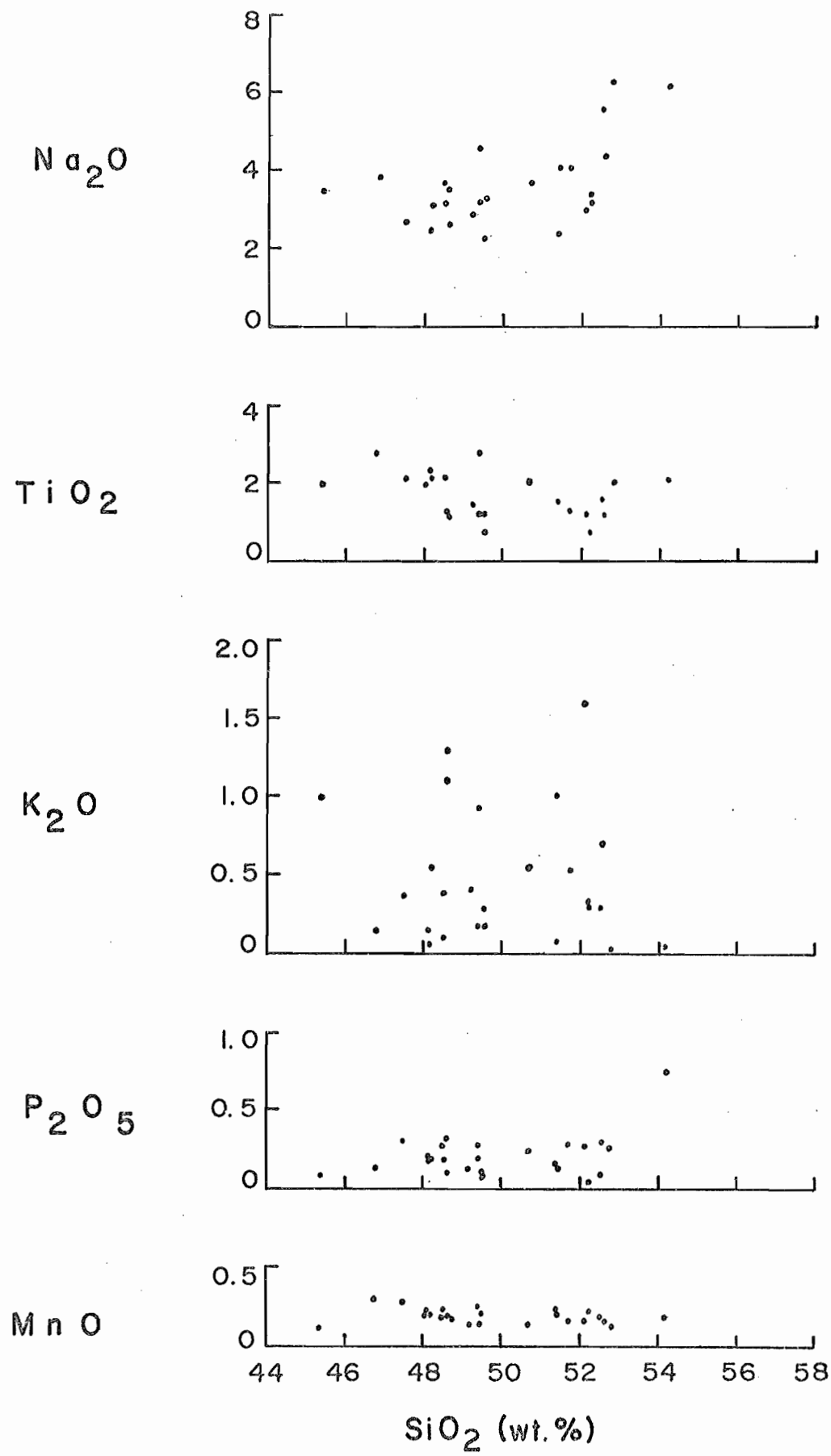


Fig. 6

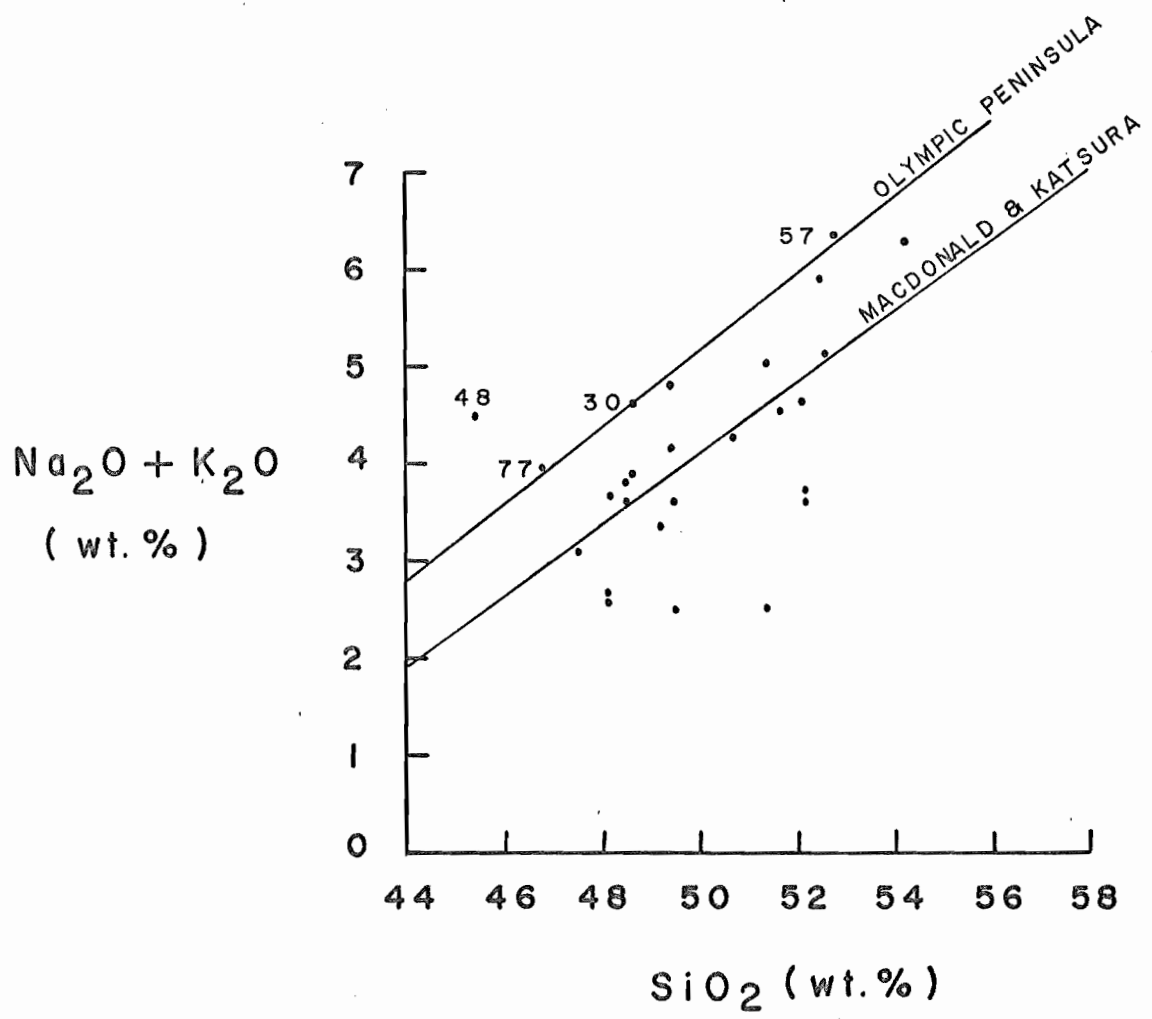


FIG. 7

CHAPTER 4

PETROGENESIS OF THE OLYMPIC PENINSULA VOLCANIC ROCKS

I. Introduction

This chapter deals with the Olympic Peninsula volcanic rock compositions as they relate to the simple normative basalt tetrahedron, and the modified lime-magnesia-alumina-silica (CMAS) tetrahedron. Based on the results of these projections, discussion is presented concerning the petrogenesis of the Olympic Peninsula volcanic rocks. Also, for comparative purposes the Olympic Peninsula volcanic rocks, spilites and published basalts have been projected onto the "spilite plane" within the CMAS tetrahedron, and general conclusions are drawn regarding the relationship between spilites and basalts.

II. Simple Normative Basalt Tetrahedron

The variation diagrams presented in Chapter 3 were mainly based on selected pairs of elements and thus are not representative of the bulk composition of the analysed rocks. The simple normative basalt tetrahedron is a diagram which involves all the major chemical components of rocks combined into minerals (Fig. 8). In this diagram, the CIPW normative constituents (weight per cent) are plotted at the four corners of the tetrahedron in the following manner:

Olivine = normative (forsterite + fayalite) + olivine component of

normative hypersthene

Clinopyroxene = normative clinopyroxene

Plagioclase = normative (albite + anorthite)

Quartz = normative quartz + quartz component of normative hypersthene

The projections are given from 4 corners of the tetrahedron (Figs. 9 - 12), and are based on the normative constituents obtained from the recalculated analyses of Table 4. The positions of the atmospheric pressure, natural cotectic boundaries in Figs. 9-12 have been taken from Clarke (1968), who obtained the positions by plotting, in each projection, published basalts with experimentally determined phase relations. These basalts are listed by Clarke (1968, p. 53).

Figures 9-12 illustrate the projections of the Olympic Peninsula volcanic rocks from the Olivine, Clinopyroxene, Plagioclase and Quartz corners respectively. In the quartz projection (Fig. 12), rocks 30, 34, 77 and 10C (open circles), which are all nepheline normative, have also been projected onto the Cpx-Oliv-Plag plane from the nepheline corner of the tetrahedron.

The main points to be noted from these projections are:

1. All rocks plot fairly close to the natural cotectic boundaries, except rocks 5 and 48, both of which have unusual chemistry.
2. All rocks plot remote from the quartz corner of the tetrahedron and close to the Cpx-Oliv-Plag plane.
3. No strong trend or clustering of points is evident on any of the projections.
4. On the quartz projection (Fig. 12), which involves the minimum distortion in the spread of the projected points by virtue of the fact that the quartz apex lies farthest away from the plotted points, a fairly restricted range of compositions is indicated for the Olympic Peninsula

volcanic rocks. This projection also indicates that the projected rocks can be grouped into 3 categories: a group which falls on or very close to the cotectic boundaries; a second group which falls within the Plagioclase field; and a third group which falls within the Olivine field. This grouping is better illustrated on Fig. 14, which is a Diopside projection within the CMAS system, and its significance will be discussed in the following section on CMAS projections.

5. There is a variation in the projected positions of several parts of individual basaltic pillows e.g. rocks 10 and 32, which is large relative to the variation in the province as a whole.

### Summary

The norm projections show the tendency for projected rock compositions to cluster around the natural cotectic boundaries, suggesting well-advanced control of rock compositions by fractional crystallisation. In the terminology of Yoder and Tilley (1962) and O'Hara (1968a), the Olympic Peninsula rocks can be regarded as fairly evolved basalts, having equilibrated at low pressures with plagioclase and clinopyroxene. This being true, then any evidence of former evolutionary trends has been lost.

### III. CMAS Projections

In order to aid in the graphical presentation of analyses of igneous rocks, a pseudo-quaternary system C-M-A-S has recently been used by O'Hara (1968a, Figs. 4-6). By considering chemical affinities between cations

in basic magmas, O'Hara has devised a scheme that allows 10 of the major and minor oxides of basic rocks to be expressed in terms of 4 components, namely: CaO, MgO, Al<sub>2</sub>O<sub>3</sub> and SiO<sub>2</sub> (CMAS). The object of this mode of projection is to present results in a form directly comparable with weight per cent plots and projections of phase equilibria in the system CaO-MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub> at various pressures. Basalt analyses, once calculated into the pseudo-quaternary system CMAS, can then be projected onto ternary planes within the pseudo-quaternary system. The plotted positions of basalts within the ternary system can then be obtained relative to phase boundaries derived from projecting experimentally studied natural or synthetic basalt compositions. O'Hara has written a computer program which calculates the coordinates of analyses within the main tetrahedron and in the various ternary projections. Jamieson (1970) has discussed the advantages of plotting within the CMAS system relative to other data projection schemes.

Figure 13 shows the pseudo-quaternary system CMAS and various planes of projection within that system. Figures 14-16 illustrate ternary projections of Olympic Peninsula volcanic rocks in relation to the boundaries at different pressures between the primary phase volumes of crystallisation of olivine, orthopyroxene, clinopyroxene and an alumina-rich phase (plagioclase, spinel or garnet, according to pressure). Figures 14 and 15 show the projections of Olympic Peninsula volcanic rocks from Diopside (CMS<sub>2</sub>) onto the ternary plane C<sub>3</sub>A-M-S, and Fig. 16 shows the projection of these rocks from Olivine (M<sub>2</sub>S) onto the ternary plane CS-MS (Enstatite)-A. The positions of the various phase boundaries



and normative divides have been taken from O'Hara (1968a). When the liquid B is referred to, it will always represent the initial liquid formed on partial melting of a 4-phase peridotite at the specified pressure.

Using the results of these three projections, 3 petrogenetic models based upon the partial melting of mantle peridotite at low, intermediate and high pressures will now be discussed.

#### 1. Partial Melting at Low Pressure

Figure 14 shows that the restricted field of rock compositions would be on the boundary of or within the primary phase volume of olivine crystallisation at pressures around 5kb; all the rock compositions (with the exception of rocks 30 and 48) fall outside the primary phase volume of olivine at pressures greater than 7kb. Unmodified primary magmas produced by the partial melting of mantle peridotite must be in equilibrium with olivine at the pressures of their formation (Cohen et al. 1967, O'Hara 1968 a, b). This means that Olympic Peninsula rocks cannot represent primary magma produced by the partial melting of mantle peridotite except at pressures less than approximately 5kb, which according to O'Hara (1968 b), implies depths of less than 15km and geothermal gradients greater than 80°C per km. If these rocks do represent erupted, unmodified primary magma, then they should have compositions represented by the first or successive liquids to form at 5kb from the partial melting of plagioclase peridotite.

As can be seen from Fig. 14, the Olympic Peninsula volcanic rocks do not have compositions represented by the first or successive liquids to form at 5kb from the partial melting of plagioclase peridotite. Figures 14 and 15 also show, with the exceptions of rocks 5, 48 and 91, that the primary phase volume of orthopyroxene never touches the Olympic Peninsula rock compositions at any pressure, and it does not seem in the least likely that these rock compositions could be generated directly on the surface of liquids in equilibrium with olivine and orthopyroxene. in the pressure range 1-5kb. If true, this observation precludes the Olympic Peninsula volcanic rocks from being the erupted, primary partial melt product of any currently accepted mantle peridotite, since any liquid formed as the partial melt of peridotite must be in equilibrium with olivine and orthopyroxene. This observation also precludes the Olympic Peninsula volcanic rocks from being the modified or fractionated products of a primary magma formed by partial melting of peridotite at low pressures. Fractionation of olivine or orthopyroxene cannot move the composition of liquid B, or successive liquids formed by greater degrees of partial melting, into the compositional field of Olympic Peninsula rocks. In fact, the reaction relationship between olivine and liquid at low pressures ( Schairer and Yoder, 1964) causes liquid compositions to move from B down the path of liquids in equilibrium with orthopyroxene and plagioclase during fractional crystallisation, and away from Olympic Peninsula rock compositions.

## 2. Partial Melting at Intermediate Pressures

Figure 15 is identical to Figure 14 except that it shows the primary

phase volumes of crystallisation of olivine, orthopyroxene and spinel or garnet at pressures of 15kb, 20kb and 30kb. In the 12-25kb pressure range, the first liquid B to form on partial melting of 4-phase peridotite mantle lies below the olivine-gabbro divide and is therefore nepheline-normative.

During advanced degrees of partial melting of peridotite, the liquid B changes composition along the Opx-Oliv boundary, which is the locus of liquids in equilibrium with harzburgitic mantle residuum. Successive liquid compositions eventually cross the olivine-gabbro divide and enter the hypersthene-normative volume since this divide is only operative up to pressures of 8kb. (O'Hara 1968a). Thus erupted magmas obtained by the partial melting of spinel peridotite at intermediate pressures may be either nepheline-normative or hypersthene-normative, depending upon the degree of partial melting. The close approach to low pressure, cotectic behaviour found in the Olympic Peninsula volcanic rocks (Figs. 14 and 16) argues against these rocks being the erupted primary magmas formed by partial melting of spinel peridotite at intermediate pressures. However, Olympic Peninsula rock compositions can be derived from the liquids formed on partial melting at intermediate pressures if olivine fractionation occurred throughout the ascent of the primary liquids towards the surface. Olivine control lines can be drawn from the Olivine composition point ( $M_2S$ ), through points representing liquid B and successive liquids formed on greater degrees of partial melting, which effectively encompass the whole range of Olympic Peninsula rock compositions (Fig. 15). An added attraction of this mode of origin

is that it can account for the nepheline-normative compositions of rocks 30, 34, 57 and 77, which all fall below the olivine-gabbro divide in the nepheline-normative volume (Fig. 15). Nepheline-normative liquids form during the early partial melting of spinel peridotite in the pressure range 12-25kb (Fig. 15), and some of these liquids may move upwards towards the surface before their compositions cross the normative olivine-gabbro divide. Olivine fractionation during the ascent of these liquids to the surface would then give erupted compositions identical with those of rocks 30, 34, 57 and 77. As partial melting increases, successive liquids will cross into the hypersthene-normative volume, and subsequent olivine fractionation during their ascent to the surface would then give erupted compositions typical of the majority of Olympic Peninsula volcanic rocks. Thus, the Olympic Peninsula volcanic rocks may represent erupted, fractionated primary magmas formed by the partial melting of spinel peridotite mantle in the pressure range 12-25kb.

The apparent grouping of compositions observed in the quartz projection (Fig. 12) can readily be seen in Fig. 14, where different groups of compositions clearly fall within the plagioclase and olivine primary phase volumes, and on or very close to the 1 atmosphere natural cotectic boundaries. This grouping has resulted as follows: the group which falls on or very close to the cotectic boundaries represents liquid compositions which equilibrated at or very near atmospheric pressure in a high level magma chamber and were subsequently erupted and crystallised. The groups falling within the plagioclase and olivine primary phase volumes are approximately collinear with the group falling on the cotectic boundaries,

and thus may represent complementary compositions which have been enriched in plagioclase and olivine respectively, relative to the cotectic group. As magma equilibrates in a high level magma chamber and precipitates olivine and plagioclase, olivine will sink through the magma faster than plagioclase because of its greater density. Thus liquids erupted from the upper part of the magma chamber will be preferentially enriched in plagioclase, while liquids erupted from the lower part of the magma chamber will be preferentially enriched in olivine. This theory is substantiated by the fact that plagioclase is the dominant mineral phase in the group falling within the plagioclase primary phase volume. However, apart from rocks 30 and 34, which contain recognisable chlorite pseudomorphs after olivine, this argument does not seem applicable to the remaining rocks which fall within the olivine primary phase volume, since they do not contain any recognisable olivine or olivine pseudomorphs. It is important to note at this point that rocks 30 and 34, which do contain recognisable chlorite pseudomorphs after olivine, still plot within the olivine primary phase volume (Fig. 14) and project in the positions expected for unaltered olivine basalts. This indicates that despite the widespread chloritisation which has affected the Olympic Peninsula rocks little change in the overall bulk chemistry has occurred. A second, a perhaps more likely, explanation why the rocks of this group fall within the olivine primary phase volume can be given. If the magmas which gave rise to these rocks were erupted quickly towards the surface, without having had time to equilibrate completely in a high level magma chamber, the expansion of the primary phase volume of crystallisation

of olivine with falling pressure (O'Hara 1968a, b), may have allowed it to overrun the liquid compositions. Thus these magmas would arrive at the surface containing a higher content of potential olivine than they would have had if they had completely equilibrated at atmospheric pressure. If this is true, it might still be expected that some evidence of olivine formation be found in these rocks. None has been found, and it is suggested that any evidence of the former presence of olivine has been destroyed by the ubiquitous chloritisation which has affected these rocks..

The existence of the hypersthene-gabbro divide (Fig. 14) inhibits development of quartz-normative compositions from olivine-normative compositions at pressures greater than 5kb (O'Hara 1968a). The Olympic Peninsula rocks have apparently equilibrated at pressures less than 5kb but contain only one true quartz-normative composition (rock 91). The rarity of quartz-normative compositions among these rocks can be explained on the grounds that the magmas underwent insufficient fractionation at low pressures to drive the liquid compositions to the invariant point  $Ol+Plag+Opx+(Cpx)+Liquid$ , and thus into quartz-normative compositions. Rock 91, which is the only true quartz-normative composition among the analysed rocks and is also the only rock (with the exception of rocks 5 and 48) which plots in the Opx primary phase volume, can therefore be regarded as the crystallised product of a more extremely fractionated liquid which managed to pass through the invariant point.

### 3. Partial Melting at High Pressure

O'Hara (1968a) has cautioned against the general validity of the

conclusion that the close approach to low pressure cotectic character of erupted magmas is indicative of an origin by partial melting at low pressures or of advanced crystal fractionation at low pressure (Yoder and Tilly, 1962, O'Hara 1965). O'Hara notes that the close approach to low pressure cotectic behaviour may also be indicative of extensive olivine fractionation throughout the ascent of a high pressure liquid, sufficient to keep the residual liquid composition close to the edge of the primary phase volume of olivine.

An alternative origin for the Olympic Peninsula volcanic rocks might therefore be an origin from picritic magma formed by partial melting of garnet peridotite at pressures about 30kb (O'Hara 1968a), followed by olivine fractionation during the ascent of the high pressure liquid to give the resulting, erupted rock compositions. A picritic magma so produced should have a composition similar to those of the Olympic Peninsula volcanic rocks, except that it would have a higher content of potential olivine. Such an origin is suggested in Fig. 15 by the facts that: 1) there is a collinearity among some Olympic Peninsula basalt compositions, the liquid B formed on the initial partial melting of a 4-phase lherzolite (O'Hara 1968a, Fig. 2) at 30kb, and the olivine composition point ( $M_2S$ ); and 2) olivine control lines can be drawn from the olivine composition point, through points representing liquid B and successive liquids formed on greater degrees of partial melting, which essentially encompass almost all the range of Olympic Peninsula rock compositions. However, this mode of origin cannot account for the presence of the nepheline-normative rocks which lie below the olivine-

gabbro divide within the nepheline-normative volume. The liquid B formed on initial partial melting of garnet peridotite at 30kb is already a hypersthene-normative picrite (O'Hara 1968a), and with higher degrees of partial melting this liquid migrates along the line of liquids in equilibrium with harzburgitic mantle residuum, and therefore farther into the hypersthene-normative volume. Under no circumstances, can olivine fractionation during the ascent of liquid B or successive liquids caused by greater degrees of partial melting, produce nepheline-normative compositions.

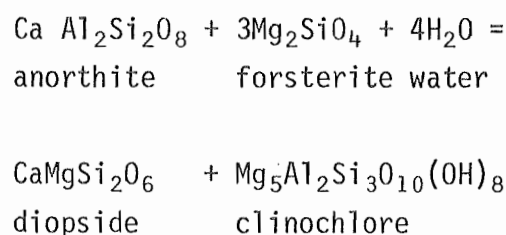
However, if the picritic magmas formed at 30kb were arrested during their ascent towards the surface at a depth of 60km (20kb), by virtue of the fact that they have moved into cooler levels of the mantle, they will crystallise and fractionate olivine and the chemical composition of the remaining liquid will change, moving towards the Opx-Oliv cotectic boundary (Fig. 15). Once the liquid composition arrives at this boundary orthopyroxene and olivine will crystallise simultaneously, and the liquid composition will migrate along the Opx-Oliv cotectic boundary towards the invariant point Opx+Oliv+Cpx+Sp+liquid. This movement of liquid down the isotherms towards the invariant point is exactly the reverse of the partial melting of peridotite at 20kb, when the liquid B initially forms at the invariant point and the liquid composition subsequently migrates up the Opx-Oliv cotectic boundary in response to greater degrees of partial melting. Thus, liquid compositions formed along the Opx-Oliv cotectic boundary at 20kb during the crystallisation of the partial melt of mantle peridotite produced at 30kb will be identical to, and indistinguishable from, liquids produced by the partial melting of peridotite at 20kb. Thus,



the Olympic magmas may have originated at 30kb, equilibrated for a period at 20kb, and then underwent olivine fractionation during their final ascent to the surface. It might be expected, according to this theory, that cognate accumulates of harzburgite be erupted with the Olympic Peninsula magmas, but none have been found.

#### IV. CMAS and the Origins of Spilites

Kushiro and Yoder (1966) added water to speed rates of reaction in the system An-Fo and discovered the following reaction, written schematically without regard for solid solution:



This reaction was recognised as crucial to the spilite problem since it yielded one of the principal coexisting mineral pairs (clinopyroxene-chlorite) characteristic of spilites. Observing that the critical mineral assemblage in regionally unmetamorphosed spilites is plagioclase + chlorite + clinopyroxene, Yoder (1967) conducted further experimental work in the system Anorthite-Forsterite-H<sub>2</sub>O, making the significant discovery that the assemblage clinopyroxene-chlorite develops in the solid state: that is, this mineral assemblage does not appear on or above the solidus (Yoder 1967, Fig. 49). The broad implications of this discovery, as stated by Yoder, is that spilites are not the direct products of crystallisation of a hydrous magma, but are autometamorphic products from a

pre-existing basalt or gabbro assemblage. An alternative view, also given by Yoder, is that a normal basalt flow could be subjected at depth to metamorphic conditions suitable for the production of the spilite mineralogy.

Yoder (1967, Fig. 45) has illustrated the relationship between basalt and a portion of the anhydrous Albite-Diopside-Clinocllore plane (the "spilite plane"), representative of the principal spilite mineral assemblage. This "spilite plane" can also be constructed within the CMAS tetrahedron. Because all albite and orthoclase molecules plot as the equivalent weight of anorthite molecule ( $CAS_2$ ), according to the equations derived by O'Hara (1968a, Fig. 4), the "spilite plane" can be represented by the compositions Diopside ( $CMS_2$ )-Clinocllore ( $M_5AS_3$ )-Anorthite ( $CAS_2$ ) within the CMAS tetrahedron (Fig. 17).

In recognition of the fact that spilites are not primary crystallisation products but are probably derived from basaltic rocks (Yoder 1967), it is instructive to calculate spilitic and basaltic compositions in terms of C, M, A and S and then project them onto the plane  $C_2S_3$ - $M_2S$ - $A_2S$  (Fig. 17), since this plane contains within its boundaries the plane of principal spilitic mineralogy i.e. the "spilite plane"  $CMS_2$ - $M_5AS_3$ - $CAS_2$ .

Figures 18 to 21 illustrate the projections, from or towards the S apex of the CMAS tetrahedron, of basaltic and spilitic rocks onto the plane  $C_2S_3$ - $M_2S$ - $A_2S$  (Fig. 21). Figure 18 shows the projections of published spilites, with  $Na_2O$  contents equal to or greater than 4 per cent by weight, onto this plane. Where possible to check given petrographic

reports, all projected spilites have essential mineralogies given by Albite + Chlorite + Clinopyroxene. The level of 4 per cent  $\text{Na}_2\text{O}$  is considered a convenient lower limit for spilites, since this figure effectively separates them from basalts, which generally do not have  $\text{Na}_2\text{O}$  contents much above 3 per cent (Manson 1967, Tables III-VI). Note that all but one of the projected spilites lie within the boundaries of the "spilite plane" and close to the  $\text{CaS}_2$  (feldspar) component.

Figure 19 and 20 illustrate the projections of published basalts onto the plane  $\text{C}_2\text{S}_3\text{-M}_2\text{S-A}_2\text{S}$  and their relation to the "spilite plane". Figure 19 represents the projections of all the published analyses of Eocene basalts from the Washington and Oregon Coast Ranges and Fig. 20 represents the projections of Carlsberg Ridge basalts and averages of individual basalt suites. The sources of these analyses are reported elsewhere in this thesis.

The projections of Olympic Peninsula volcanic rock compositions (analyses, Table 4) onto the plane  $\text{C}_2\text{S}_3\text{-M}_2\text{S-A}_2\text{S}$ , in relation to the natural cotectic boundaries at 1 atmosphere, are illustrated in Fig. 21. The positions of the cotectic boundaries were obtained by projecting published basalts with experimentally determined phase relations (Yoder et al. 1963, 1964, 1965, 1967) onto the plane  $\text{C}_2\text{S}_3\text{-M}_2\text{S-A}_2\text{S}$ . All the Olympic Peninsula rocks, with the exceptions of rocks 5 and 48, plot within the "spilite plane", and show a 1 atmosphere cotectic character.

For discussion of the projections of basalts and spilites onto the "spilite plane" to be truly meaningful, it has to be demonstrated that the "spilite plane" does adequately represent the compositions of basalts

and spilites. To demonstrate that this is the case, the means and standard deviations of residual S have been calculated for the projected rocks, and are reported in Table 8. If the mineralogical components of the "spilite plane" i.e.  $CaMgSi_2O_6$  (diopside) +  $M_5Al_3Si_3O_{10}(OH)_2$  (clinochlore) +  $CaAl_2Si_2O_8$  (feldspar) adequately represent the chemical compositions of rocks recalculated into the S- $Ca_2Si_2O_7$ - $M_2SiO_5$ - $Al_2SiO_5$  tetrahedron (Fig. 17), then these rocks should lie on, or very close to, the base of this tetrahedron remote from the S apex. In addition, these rocks should also plot within the boundaries of the "spilite plane" when projected from or towards S onto the plane  $Ca_2Si_2O_7$ - $M_2SiO_5$ - $Al_2SiO_5$  (Fig. 17). Therefore, a measure of how far above (+VeS) or below (-VeS) the "spilite plane" any particular rock plots can be given in terms of residual S. Thus the calculated means and standard deviations of residual S for any particular suite or group of rocks will give a measure of the ability of the "spilite plane" to represent their particular compositions.

The results of Table 8 indicate that the "spilite plane" can, and does, represent extremely well the compositions of basalts and spilites. Figures 18-21 also illustrate that all projected spilites and basalts (with the exceptions of rock 5 and 48) plot on or within the boundaries of the "spilite plane).

The boundaries for the respective fields of projected rock compositions from Figs. 18-21 have been drawn, and the relationship between these boundaries are illustrated in Fig. 22. The Olympic Peninsula boundary line has been drawn exclusive of rocks 5, 8, 30, 48 and 80 (Fig. 21), the justification for this being that these particular rocks do not

represent the typical chemical variation found within the province as a whole. Rocks 5 and 48 are rejected on grounds of unusual chemistry, as has been previously discussed in Chapter 3; rock 30 is unusually rich in olivine and consequently MgO, and thus plots closer to the  $M_2S$  (olivine) component; rocks 8 and 80 are both soda-rich, containing 6.2% and 5.6%  $Na_2O$  respectively.

It can be seen from Fig. 22 that strong overlap exists between the fields of Washington and Oregon basalts and Olympic Peninsula volcanic rocks ("altered basalts"), and also between these respective fields and the field of "normal" basalts (average basalts and Carlsberg Ridge basalts). The field of spilites extends largely out towards soda-rich compositions i.e. towards the  $CAS_2$  (feldspar) component, and only partly overlaps any of the above basalt fields at its soda-poor end.

The field of Olympic Peninsula volcanic rocks apparently has closer affinities with the fields of basalts than the field of spilites, a fact that has been previously suggested from their petrography and chemistry, and this may indicate that the Olympic Peninsula rocks are largely basaltic rather than spilitic. Although statistical data is required concerning the significance of the overlap between the fields of spilites and Olympic Peninsula volcanic rocks before definite conclusions can be made regarding the relationship of spilites to Olympic Peninsula rocks, the validity of the Olympic Peninsula being a classic spilite locality must be open to serious questioning.

Other interesting features also arise from Fig. 22. An important continuous spectrum in chemistry between basalts and spilites is suggested

from this figure, which could be interpreted as further proof that spilites are derived from rocks of basaltic composition. Although spilites overlap on the field of Average basalts and Carlsberg Ridge basalts, the results of Table 8 indicate that the "spilite plane" represents the compositions of Carlsberg Ridge basalts much more closely than it does the compositions of the Average basalts, which are largely continental in origin. This may indicate that, if spilites are derived from basalts, they may be derived from oceanic-type basalts rather than continental basalts. This suggestion would be consistent with the theories postulated in this thesis that spilites are largely derived from basalts which originate at oceanic ridges and in island arcs. However, both of these features are mere suggestions, based upon extremely limited data, and obviously larger representative samples and statistical treatment of the resulting data are required before these suggestions can be substantiated or disproved.

#### V. Summary

Ternary projections from Diopside ( $CaMgSi_2O_6$ ) and Olivine ( $Mg_2SiO_4$ ) within the CMAS system have further shown the restricted range of compositions and the low pressure equilibration of Olympic Peninsula magmas. In addition, these projections have also provided possible insight into the pressure and depth of generation of magmas which were parental to the Olympic Peninsula volcanic rocks, and three petrogenetic models have been discussed. Derivation of the Olympic Peninsula volcanic rocks by the eruption of primary magmas formed by low pressure partial melting of plagioclase peridotite, or by subsequent fractionation of a low pressure

partial melt, is rejected on the grounds that the compositions of these rocks cannot be derived from such magmas according to the known phase relations. Partial melting of spinel peridotite at intermediate pressures (12-25kb) or garnet peridotite at high pressures (25-30kb) can give rise to the Olympic Peninsula rock compositions if the primary liquids subsequently underwent olivine fractionation during their ascent to the surface. Of these two models, partial melting of spinel peridotite at intermediate pressures is preferable, since this model allows derivation of nepheline-normative rock compositions which cannot be obtained by any simple fractionation scheme from primary liquids formed during partial melting at high pressures. However, if picritic magmas which form at 30kb become arrested at 20kb during their ascent to the surface and fractionate olivine and orthopyroxene, liquid compositions will form which are identical to liquids which form by the partial melting of peridotite at 20kb. Under these circumstances, nepheline normative rock compositions may be produced by partial melting of peridotite at 30kb. It should also be remembered from discussions in Chapter 3 that the production of nepheline in the norms of rocks 30, 34, 57 and 77 may be the result of alkali addition, and that these rocks may not represent true alkali basalts.

Projections of spilites, basalts and Olympic Peninsula rocks onto the "spilite plane" within the CMAS system have suggested that the Olympic Peninsula volcanic rocks are basaltic rather than spilitic in character, and thus the validity of the Olympic Peninsula being a classic spilite locality must remain in question. However, relevant statistical data is required before this assertion can be substantiated. These projections have also suggested that there is an apparent continuous spectrum in

chemistry between spilites and basalts, and that spilites may be derived from oceanic-type basalts rather than continental basalts. Again, more representative samples and statistical data are required to prove or disprove these suggestions.



TABLE 8

## Mean and Standard Deviation of Residual S

Rock Type	No. of Samples	Mean (%S)	Standard Deviation
Spilites	23	+0.34	5.39
	22	-0.55	3.42
Olympic Peninsula volcanics	26	+0.58	3.69
	25	+0.13	2.94
Washington-Oregon basalts	12	+3.26	2.90
	11	+2.49	1.23
Average basalts	9	+5.05	2.98
Calrsberg Ridge basalts	6	+0.99	0.74

In each of the first three rock types, one projected analysis was anomalous because of a certain unusual feature of chemistry; consequently, unusually high and atypical residual S values were obtained from these analyses e.g. one spilite with a SiO<sub>2</sub> content of 63.6% gave a residual S value of +19.7. To illustrate this effect, and therefore give a more accurate representation of the means and standard deviations, the above values are reported for all the projected analyses in each group, and also the analyses minus the anomalous rock in each one of the first three groups.

Fig. 8 The Simple Normative Basalt Tetrahedron  
(from Yoder and Tilley, 1962)

Fig. 9 Olivine projection, simple normative basalt tetrahedron.

Fig. 10 Clinopyroxene projection, simple normative basalt tetrahedron.

Fig. 11 Plagioclase projection, simple normative basalt tetrahedron.

Fig. 12 Quartz projection, simple normative basalt tetrahedron  
o nepheline normative compositions (rocks 30, 34, 77 and 10C)

Rocks 8, 57 and 80 all have normative  $(An+Di+Hyp+Oliv) < 45$  and are therefore not considered as basalts for the purposes of projection within the normative basalt tetrahedron. Consequently, these rocks have not been projected in Figs. 9-12.

Rocks 30, 34, 77 and 10C are all nepheline normative, and consequently only appear in the Quartz projection (Fig. 12), where they have been projected from the Ne apex of the simple normative basalt tetrahedron onto the plane CPX-PLAG-OLIV.

Thus, only 19 rocks have been projected in Figs. 9-11, while 23 rocks have been projected in Fig. 12.

SIMPLE NORMATIVE BASALT  
TETRAHEDRON

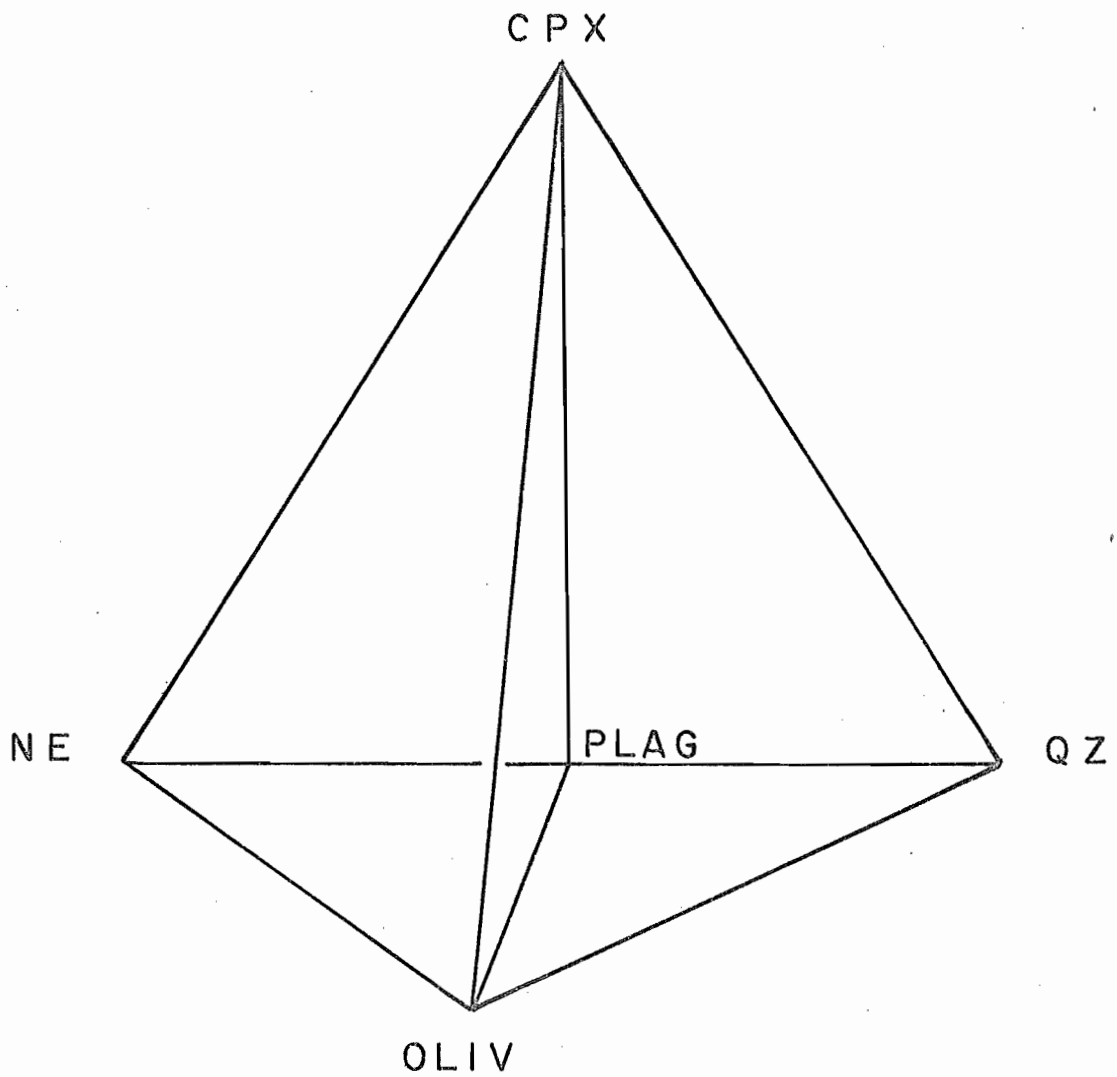


FIG. 8

OLIVINE PROJECTION

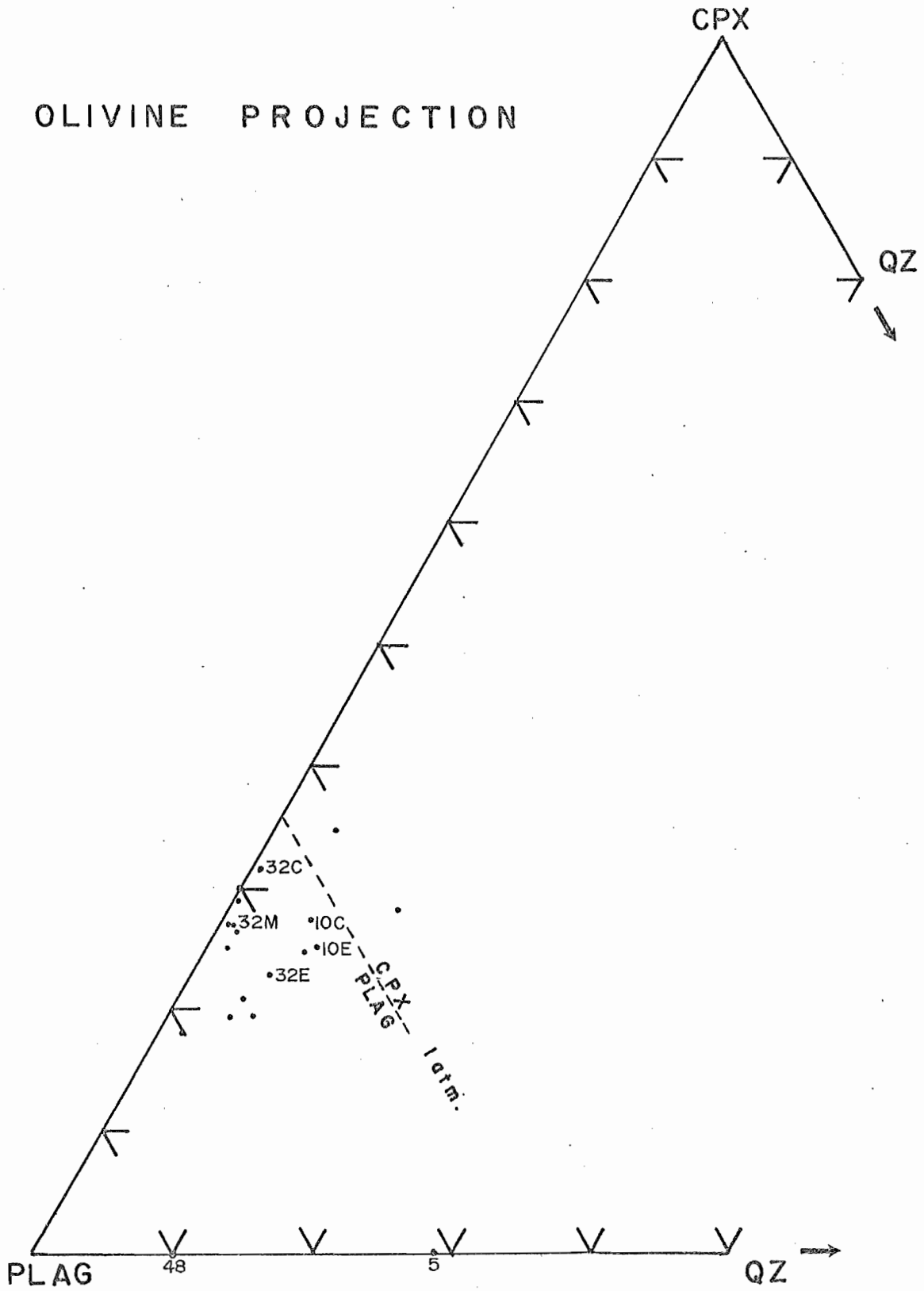


FIG. 9

CLINOPYROXENE  
PROJECTION

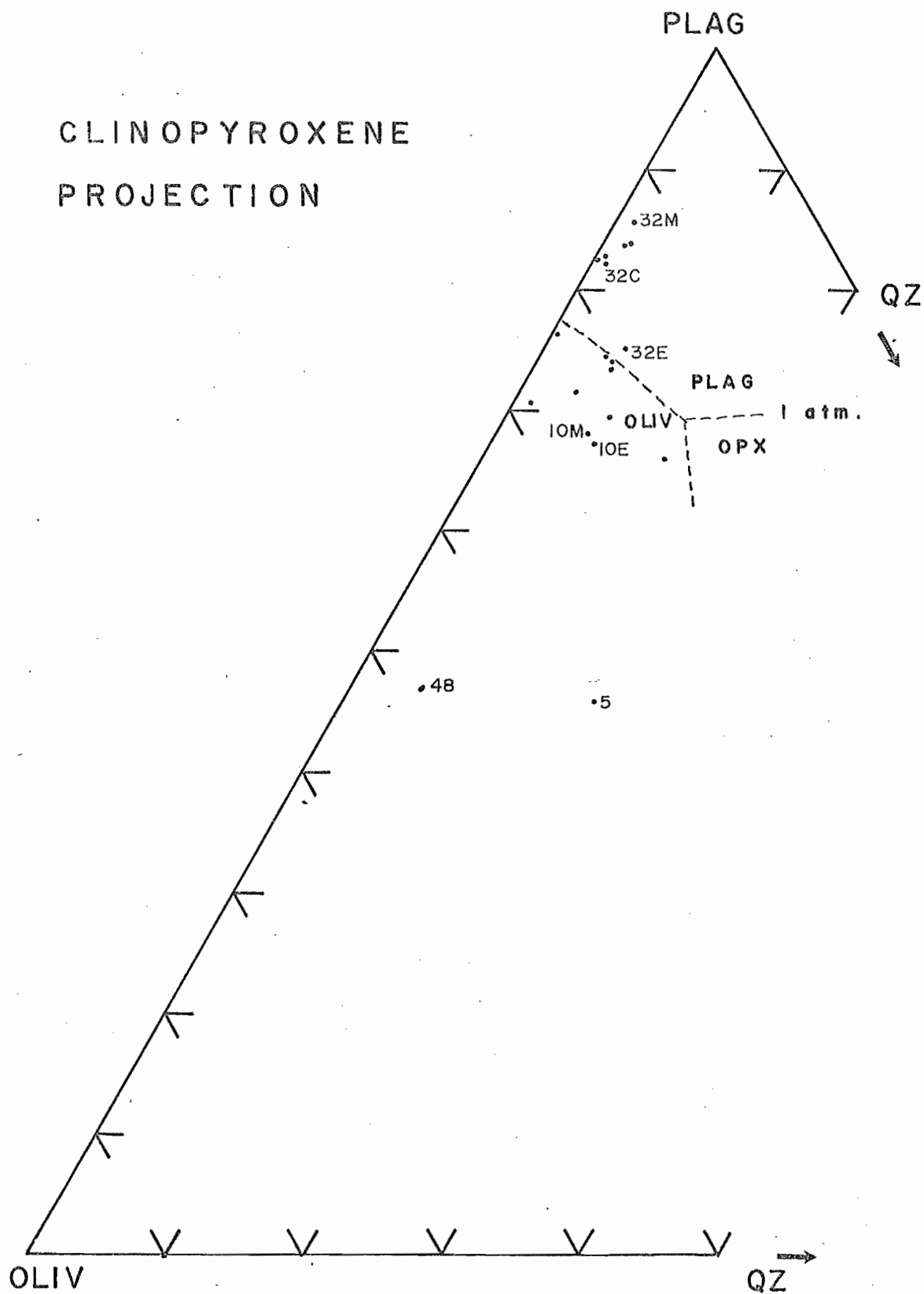


FIG. 10

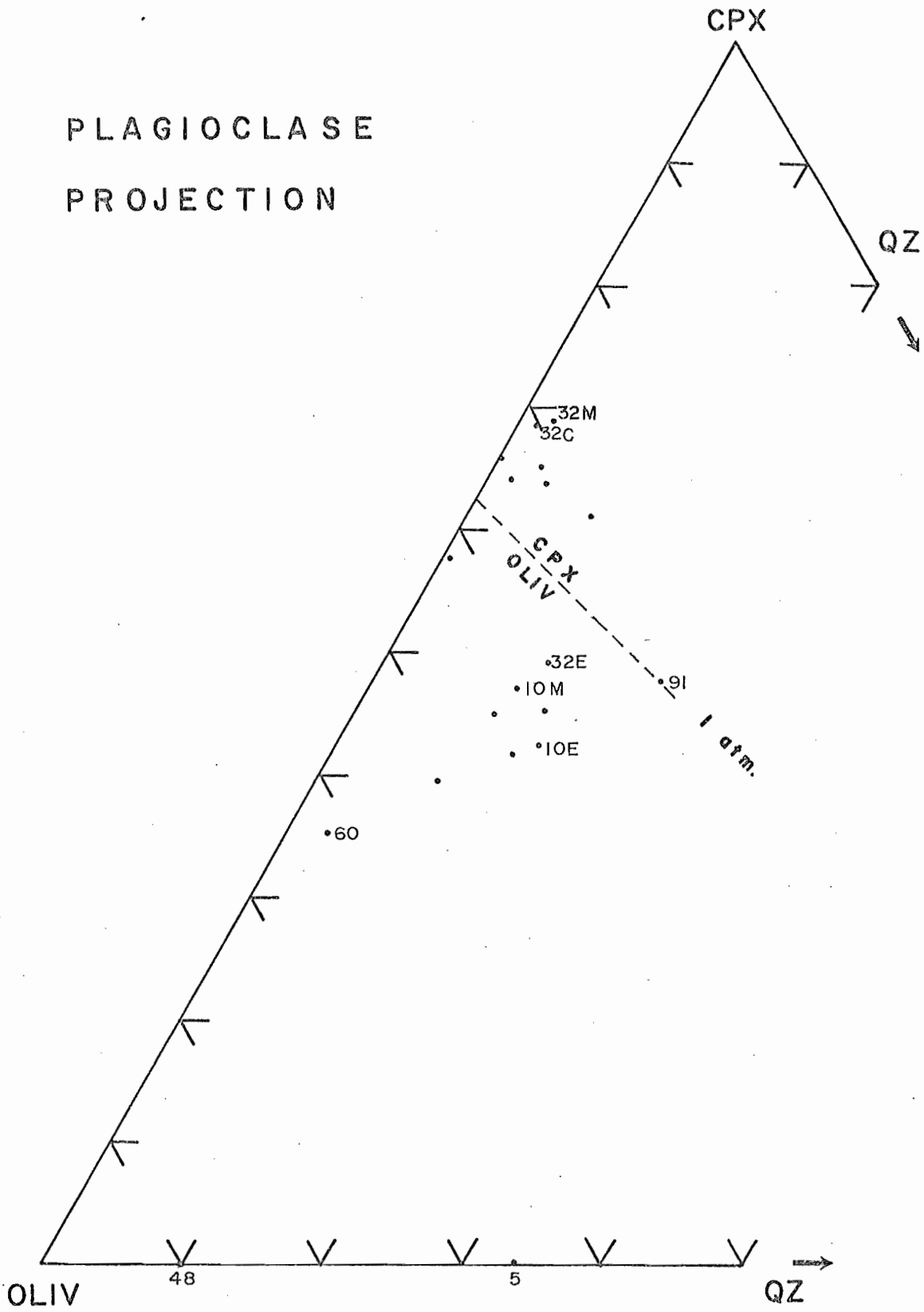


FIG. II

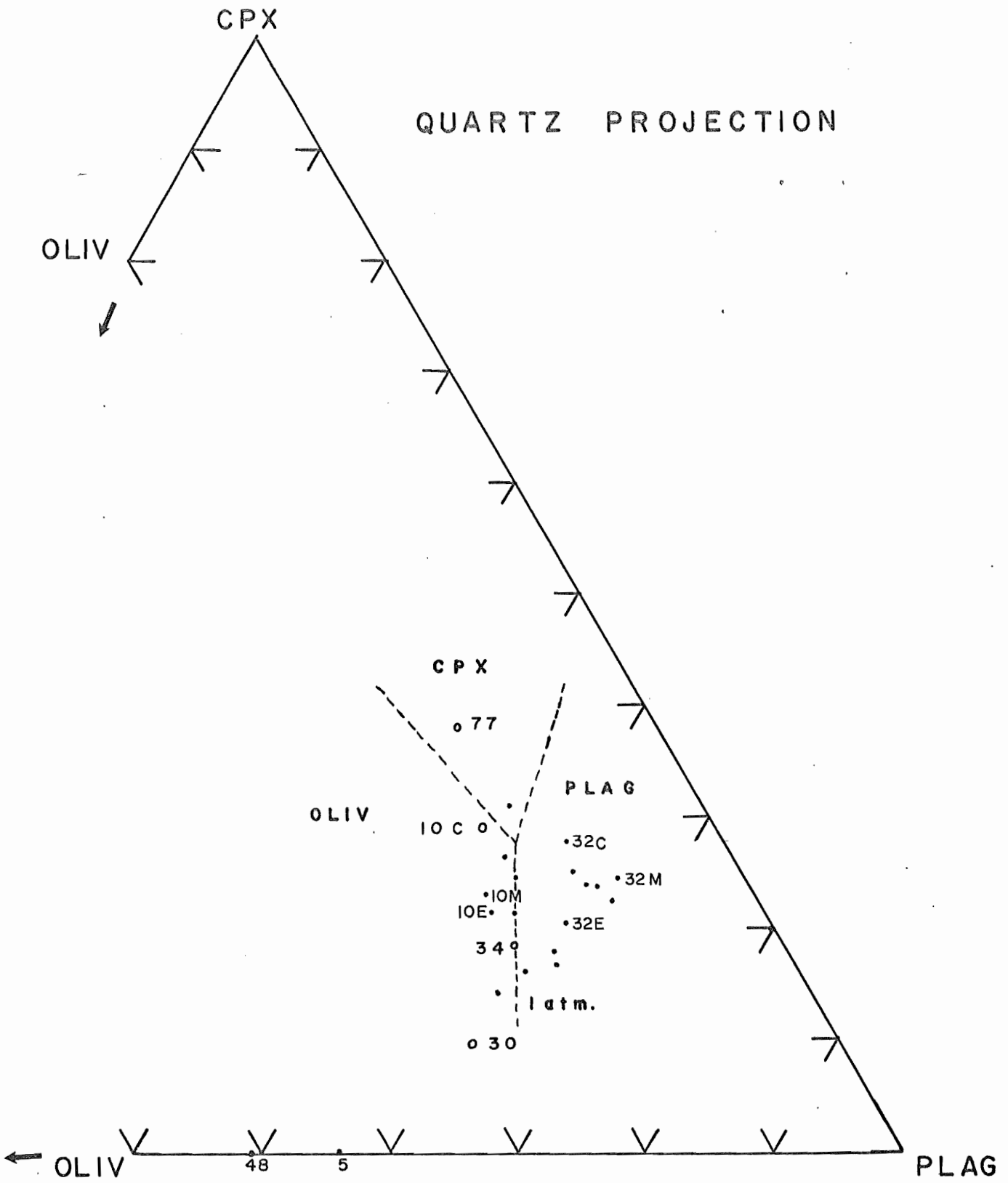


FIG. 12

- Fig. 13 CMAS Tetrahedron, showing ternary planes of projection
- Fig. 14 Diopside ( $\text{CMS}_2$ ) projection, CMAS tetrahedron
- Fig. 15 Diopside ( $\text{CMS}_2$ ) projection, CMAS tetrahedron
- Fig. 16 Olivine ( $\text{M}_2\text{S}$ ) projection, CMAS tetrahedron
- Fig. 17 CMAS tetrahedron, showing position of the "spilite plane"



C M A S T E T R A H E D R O N

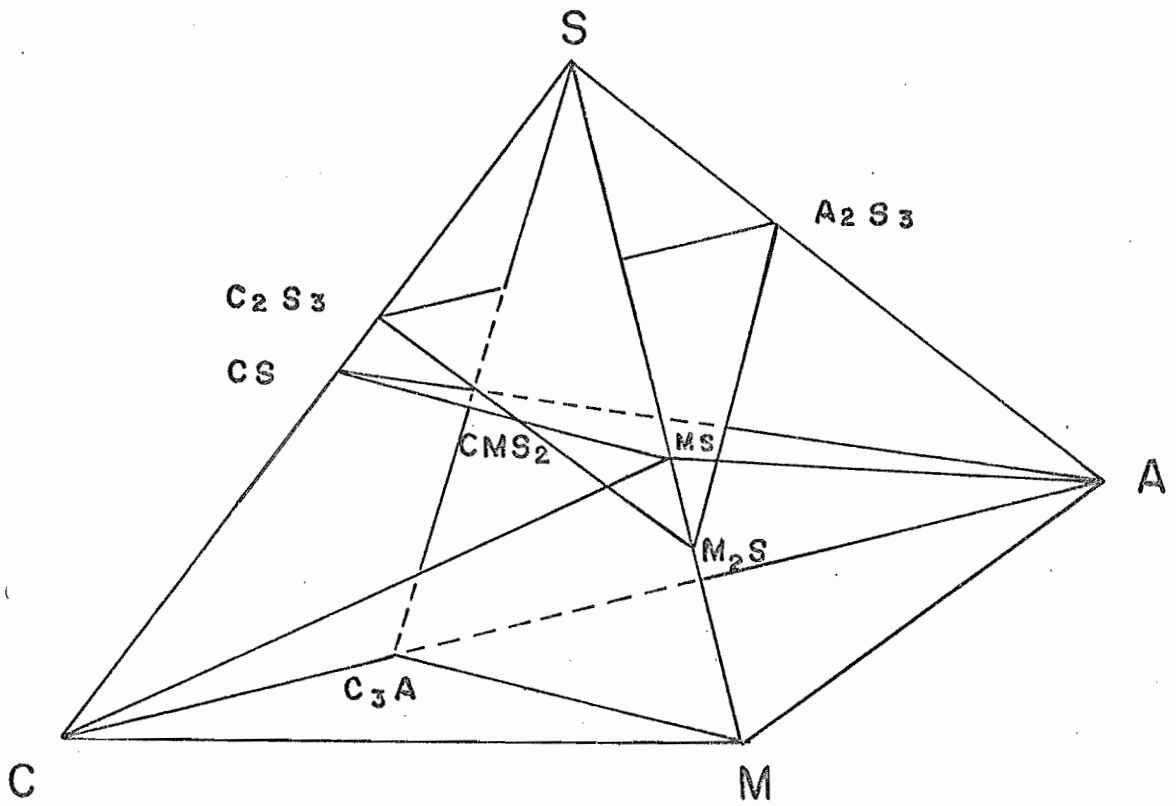


FIG. 13

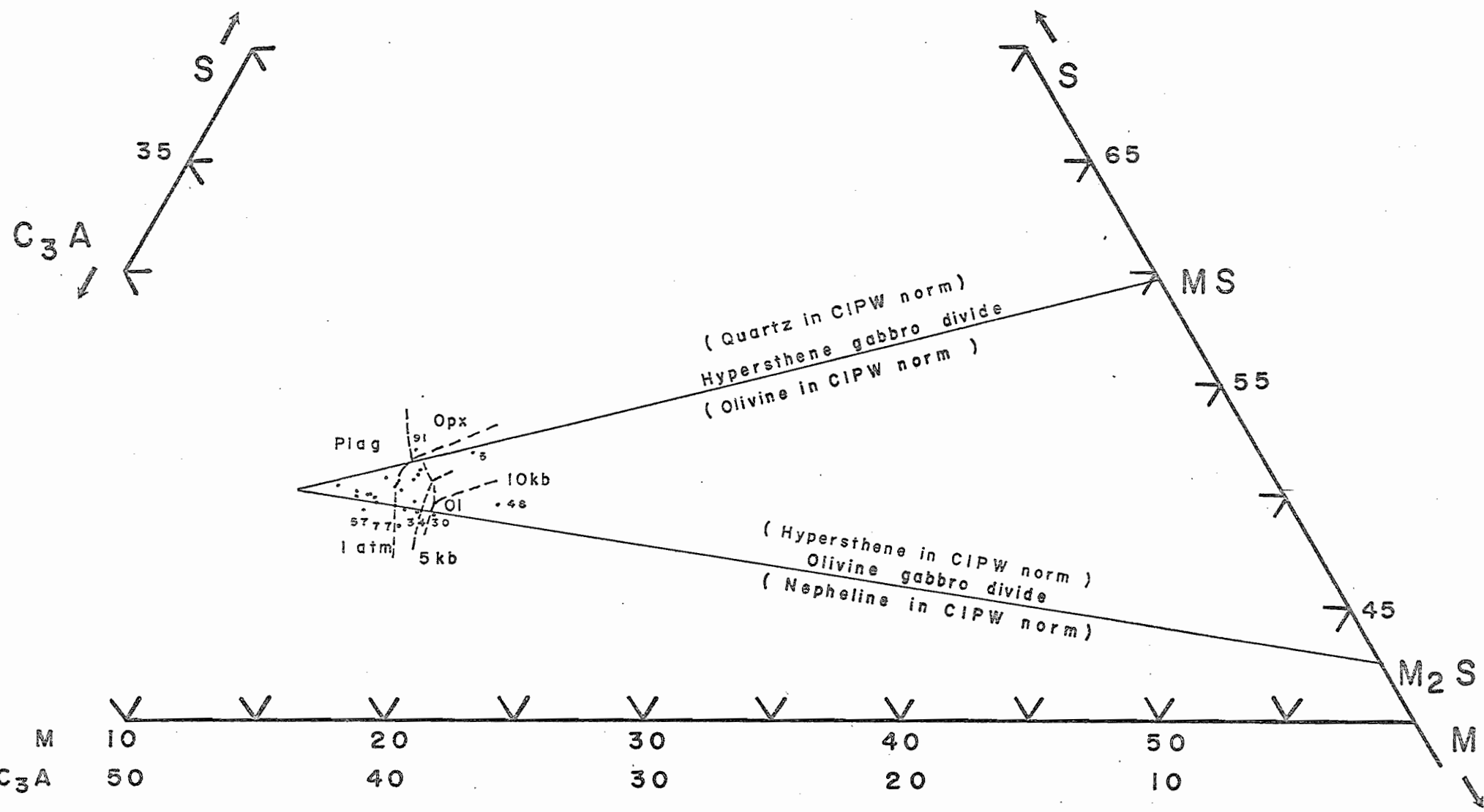


FIG. 14

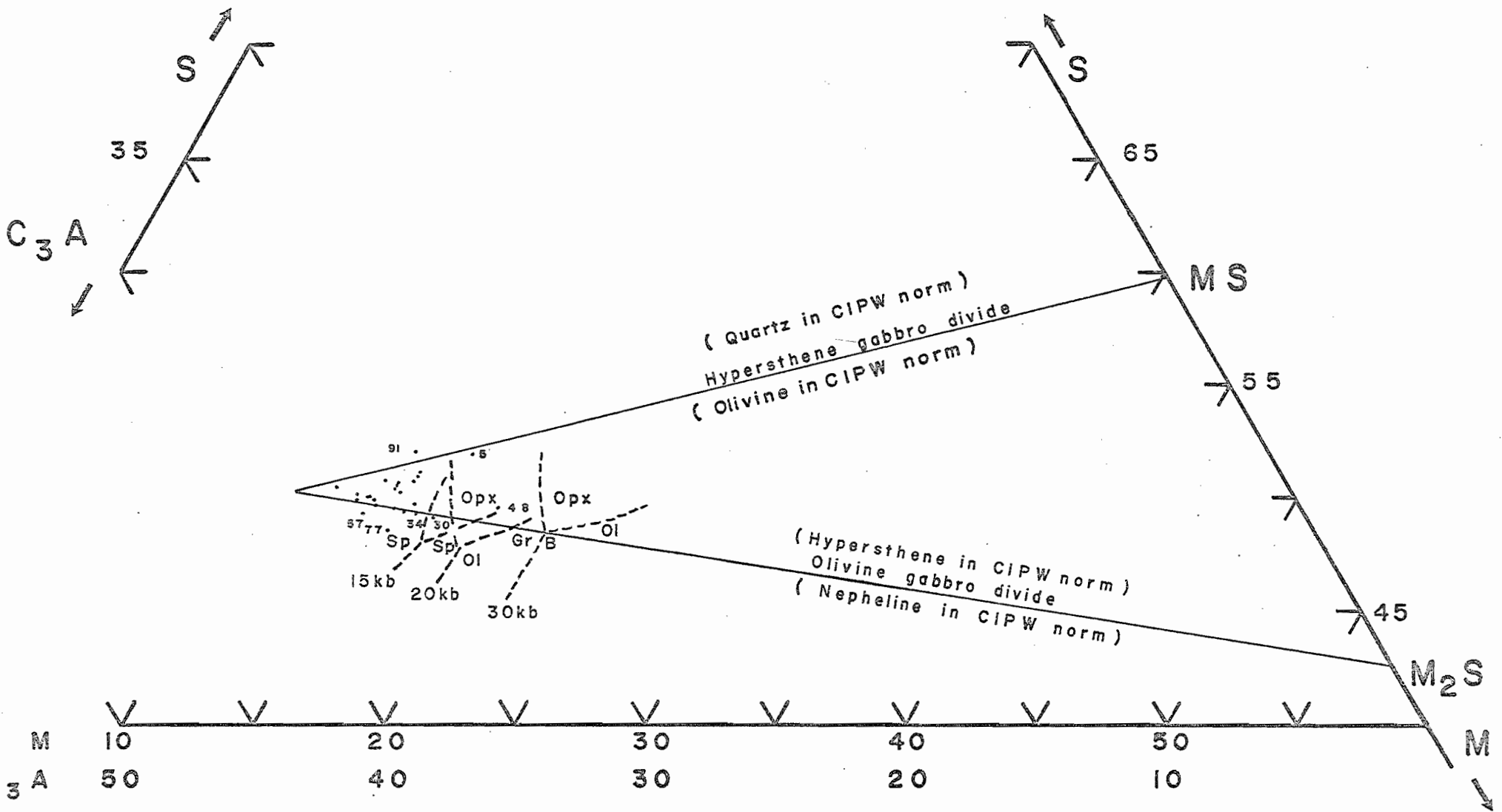


FIG. 15

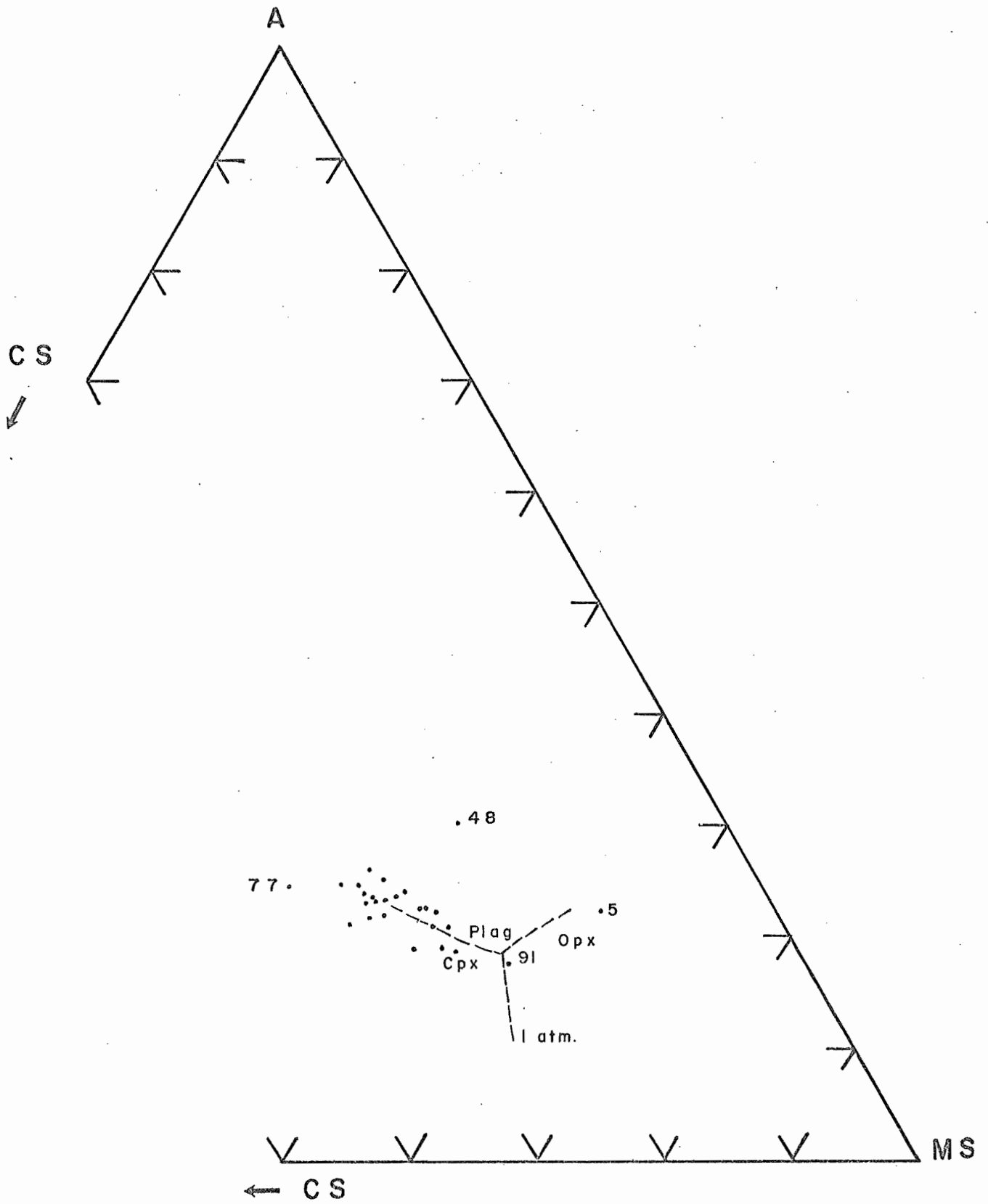


FIG. 16

CMAS with the " SPILITE PLANE "

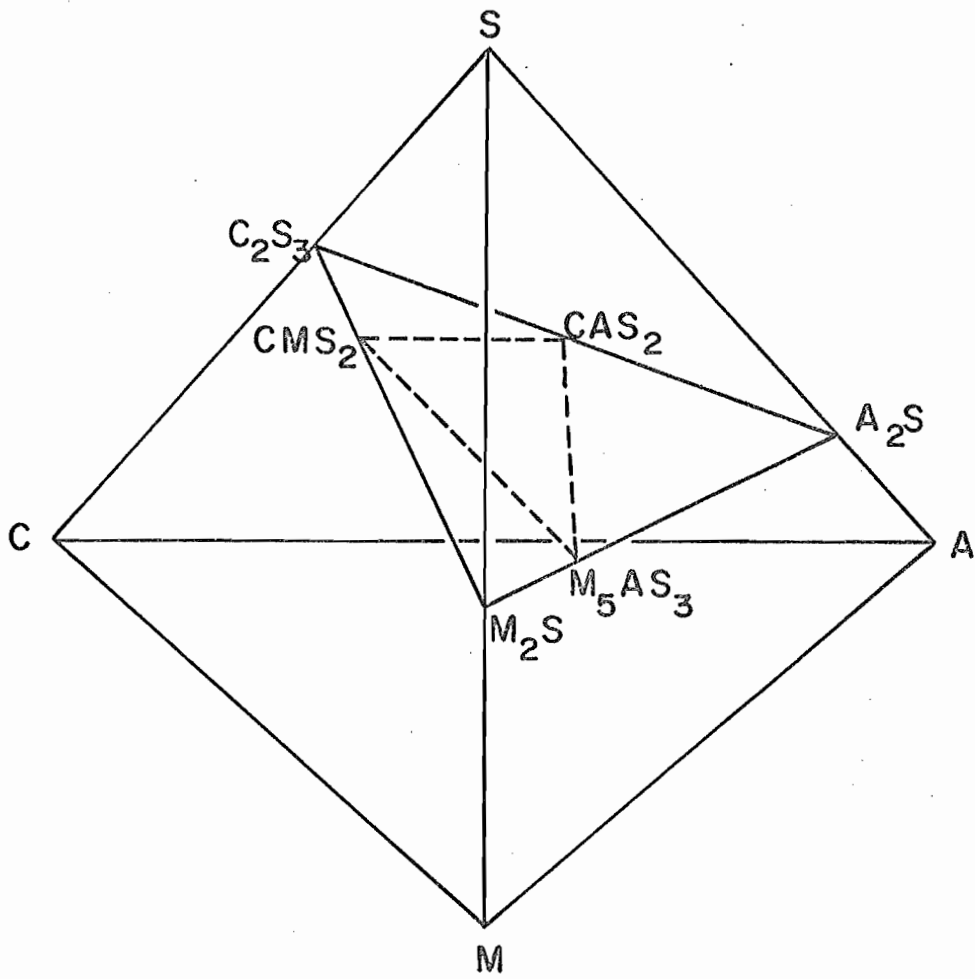


FIG. 17

Fig. 18 Projection of spilites onto the plane  $C_2S_3 - M_2S - A_2S$ ,  
CMAS tetrahedron

Sources of Analyses:

- o Vallance 1969, Table 1, col. 2. Average of 225 spilites.
- \* Vallance 1960, Table 2, col. 93. Average of 92 spilites.
- + Sundius 1930, Table II, col. 1. Average spilite.
- Wells 1923, page 69, col. 1. Average of 7 spilites.
- Miscellaneous spilites from -
  - Cann 1969, Table 1, cols. 1 and 2
  - Aumento and Loncarevic 1969, Table 1, col. 7
  - Melson and Van Andel 1966, Table VI, cols. 4 and 5
  - Smith 1970, Table 1, cols. 5, 7 and 8
  - Bartrum 1936, Table 1, cols. 2 and 3
  - Gilluly 1935, Table 1, col. 8
  - Nicholls, 1958, Table I, cols. 1 and 2; Table II, cols. 5 and 6;  
Table III, cols. 7, 10 and 12
  - Batthey 1956, Table 3, col. 3

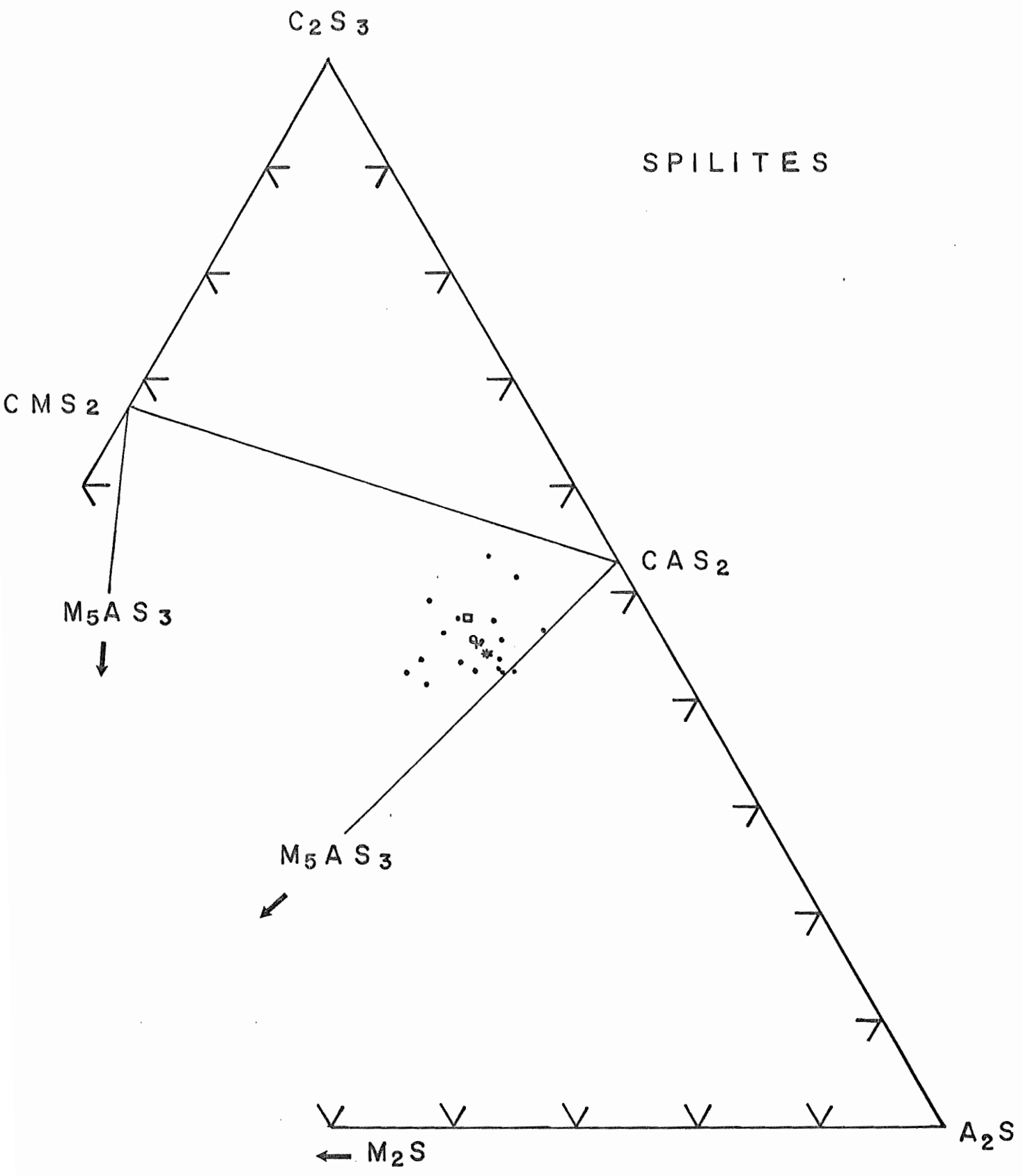


FIG. 18

Fig. 19      Projections of Washington and Oregon Eocene basalts  
                 onto the plane  $C_2S_3$ - $M_2S$ - $A_2S$ , CMAS tetrahedron

Sources of Analyses:

- x    Park 1946, Table II, cols. 1 and 2. Olympic Peninsula basalts.
- o    Waters 1955, Table 1, col. 1. Average of 4 Eocene basalts.
- Snaveley et al. 1968, Table 2, cols. 1-8. Siletz River basalts.
- \*    Snaveley et al. 1968, Table 3, col. 3. Average of 10 Crescent basalts.



WASHINGTON & OREGON  
BASALTS

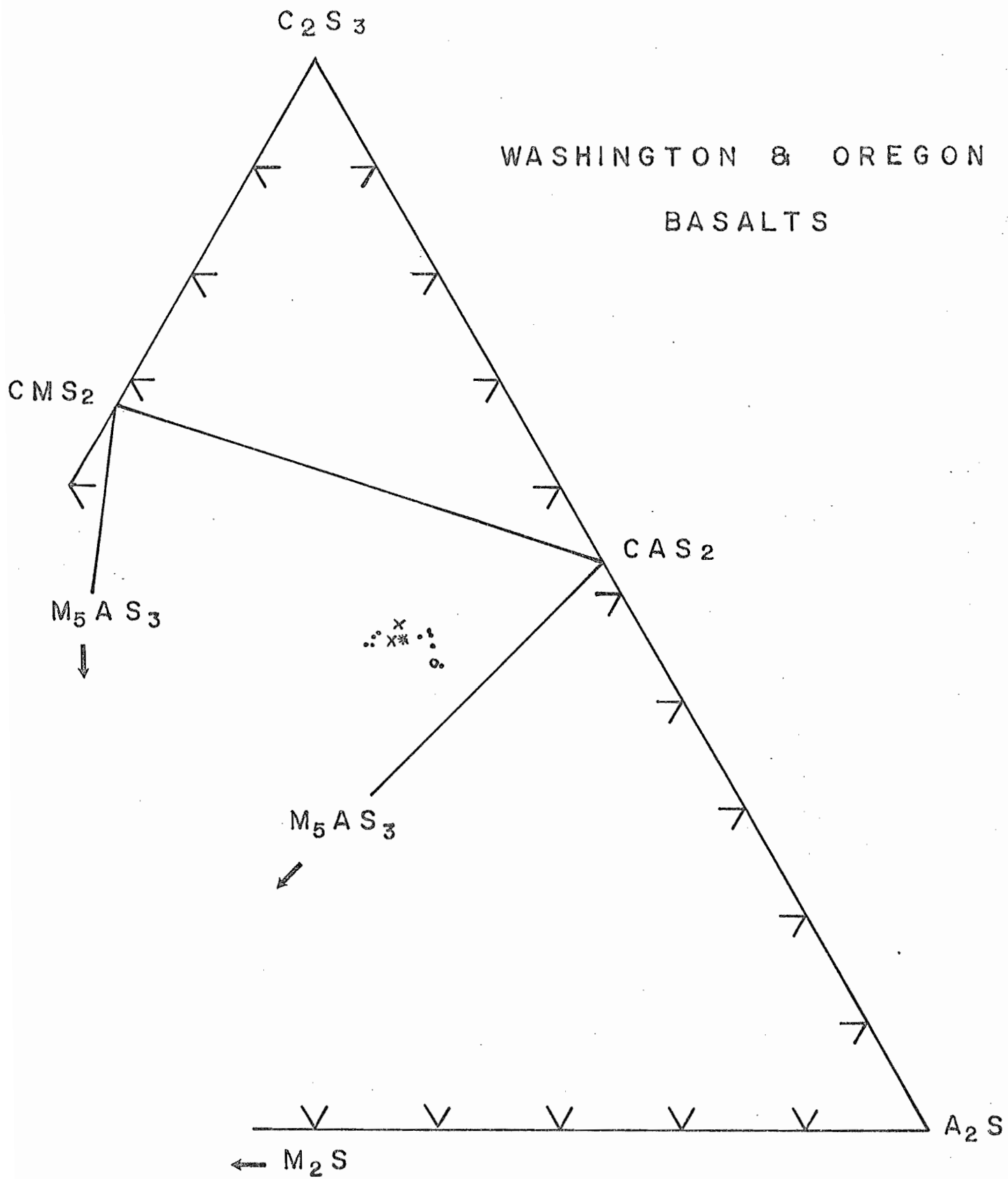


FIG. 19

Fig. 20 Projections of Carlsberg Ridge and Average basalt suites onto the plane  $C_2S_3$ - $M_2S$ - $A_2S$ , CMAS tetrahedron

Sources of Analysis:

- o Carlsberg Ridge basalts - Cann 1969, Table 3, cols. 1-6.
- Average basalt suites -
  - Engel et al. 1965, Table 3, col. 1. Average of 10 deep oceanic tholeiites.
  - Manson 1967, Table VI, col. 8. Average of 182 olivine tholeiites.
  - Manson 1967, Table VI, col. 17. Average of 897 tholeiitic basalts.
  - Cox and Hornung 1966, Tables 2 and 3. Average of 21 Lesotho basalts.
  - Cox et al. 1967, Table 3, nos. A, B, C and D<sub>1</sub>. Average of 20 low-Mg Rhodesian Karroo basalts.

CARLSBERG RIDGE  
&  
AVERAGE BASALTS

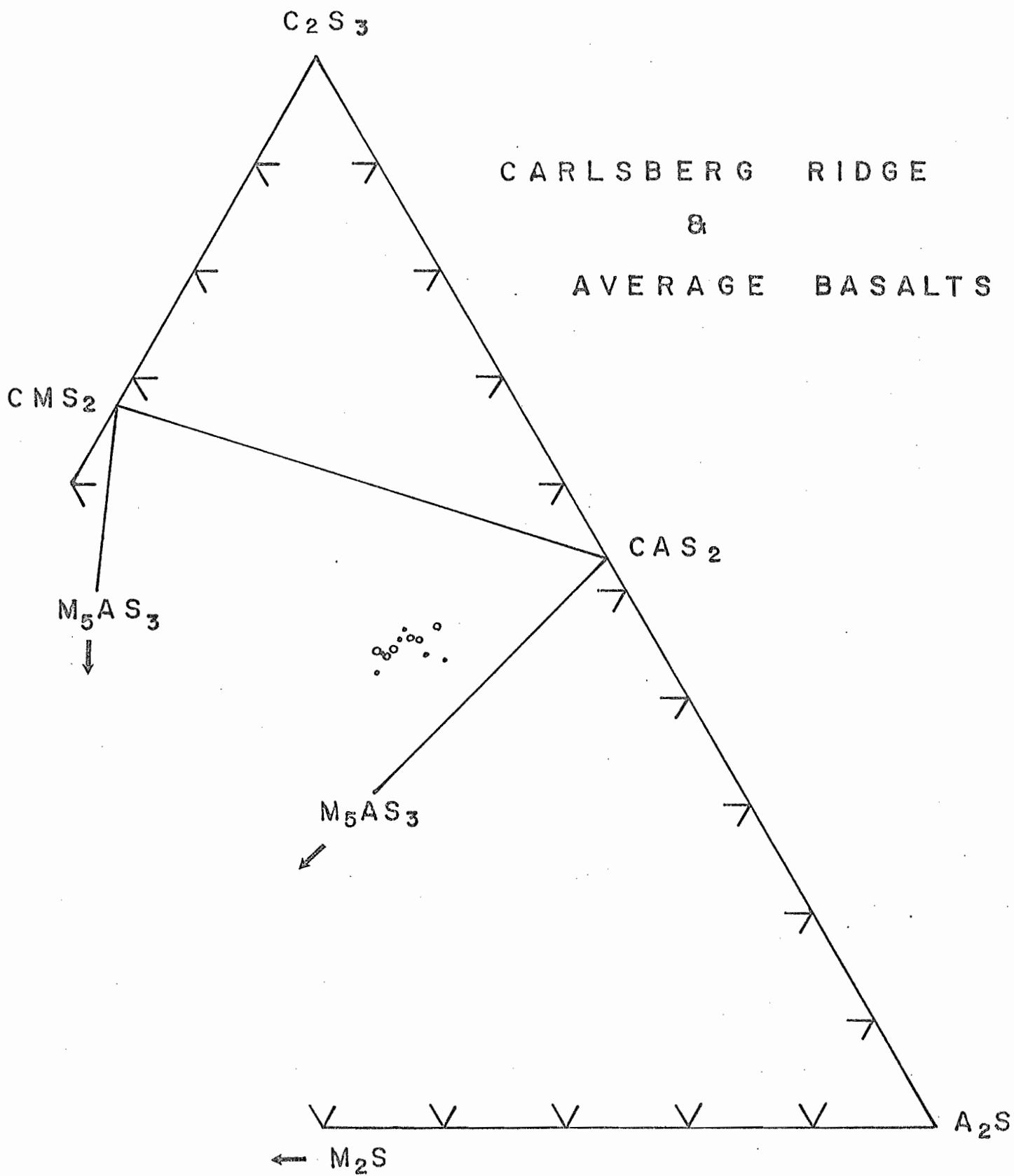


FIG. 20

Fig. 21      Projections of Olympic Peninsula volcanic rocks onto the  
plane  $C_2S_3$ - $M_2S$ - $A_2S$ , CMAS tetrahedron

Analyses, Table 4

OLYMPIC PENINSULA  
ROCKS

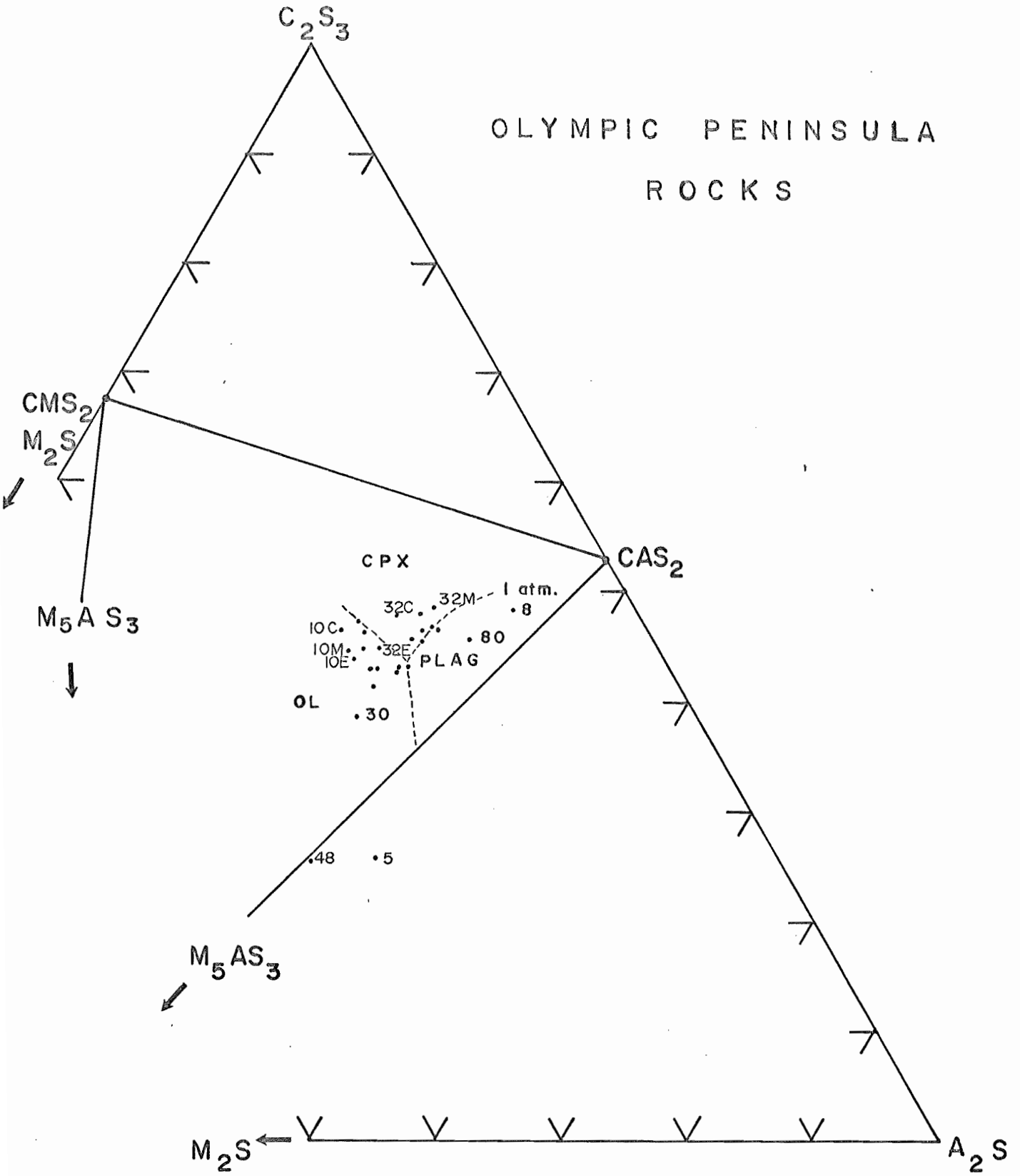


FIG. 21

Fig. 22      Boundaries of the fields for basalts, spilites and Olympic Peninsula volcanic rocks (Figs. 22-25) in relation to the "spilite plane"  $\text{CaS}_2\text{-M}_5\text{AS}_3\text{-CaS}_2$

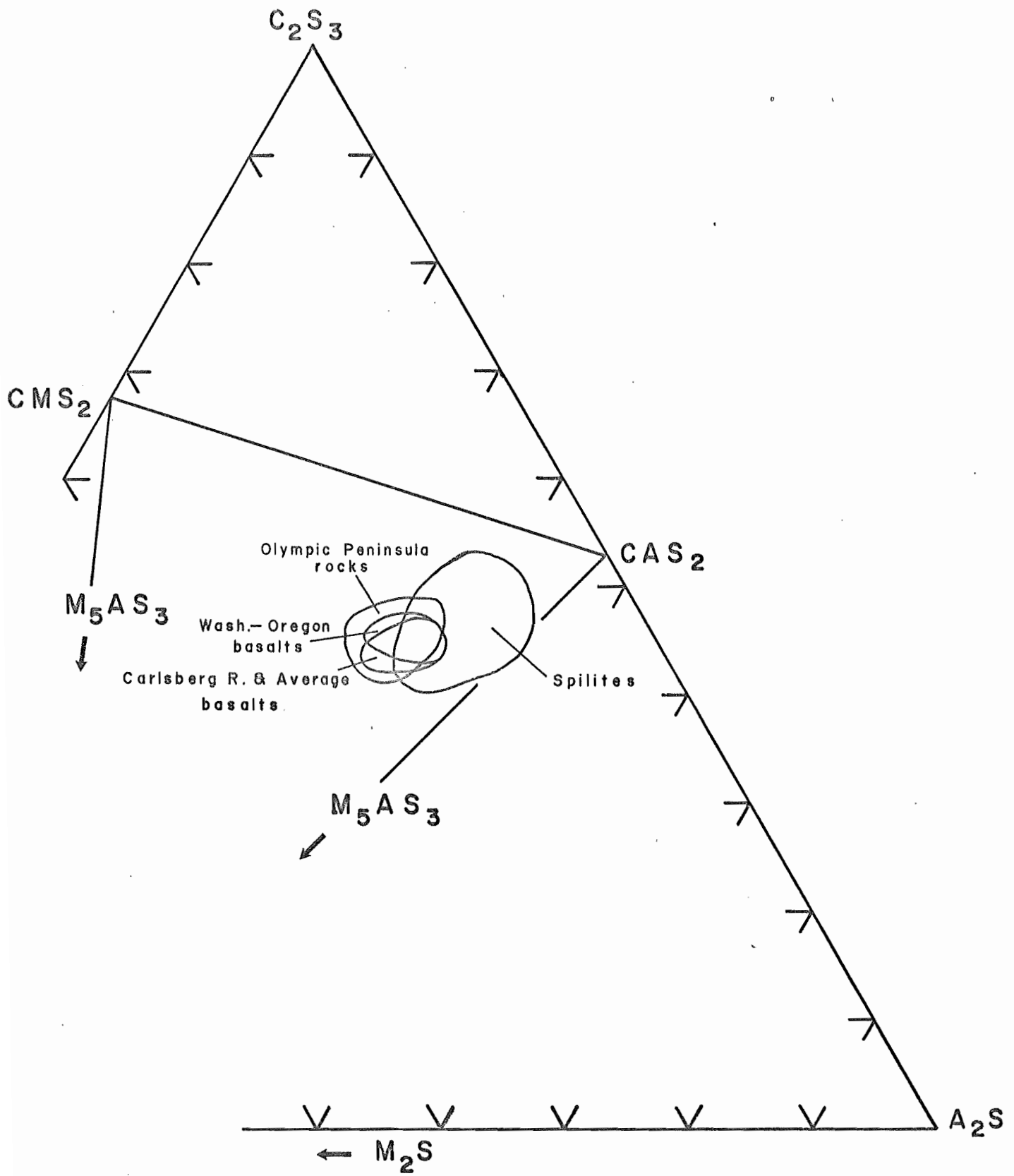


FIG. 2 2

CHAPTER 5

STRUCTURE AND TECTONIC EVOLUTION OF WESTERN WASHINGTON AND OREGON

I. Introduction

Marine geological and geophysical investigations have now shown the world-wide distribution of oceanic ridges and trenches. Also, the resulting theories of sea-floor spreading and plate tectonics (Vine 1966, Morgan 1968, Isacks et al. 1968) have led to the realisation that much of present-day oceanic and continental volcanism is probably directly related to spreading at oceanic ridges and subduction of oceanic lithosphere in off-shore trenches (Dickinson 1968).

Traditionally, interbedded sequences of marine sedimentary rocks, pillow lavas and breccias, and volcanoclastic deposits have been believed to originate in eugeosynclines, which formed by the downbuckling of continental margins. Orogeny and uplift then elevated these geosynclinal deposits above sea level, forming continental landmasses and mountain chains. Such an origin has been postulated for the Coast Ranges of western Washington and Oregon (Snively and Wagner, 1963).

Since at least the Mesozoic, the western margin of North America has been characterised by the interactions between oceanic and continental lithospheric plates, and numerous attempts have been made to explain the origins of on-shore and off-shore geological features as the results of these plate interactions e.g. Hamilton (1969) and Atwater (1970). While



considerable attention has been paid to explaining geological features found in the California Coast Ranges (Hamilton 1969; Page 1969, 1970; Ernst 1970; Bailey et al. 1970) and the southwestern Oregon Coast Range (Medaris and Dott, 1970, Medaris 1972) in these terms, so far no such explanation of the origin of the Coast Ranges in western Washington and Oregon has been attempted.

The voluminous lower - middle Eocene basaltic rocks found in these Coast Ranges have been recognised as largely marine in origin (Snively and Wagner, 1963, Snively et al. 1968, Lipman et al. 1972), and their origin as submarine volcanics seems fairly certain. It should be worthwhile, therefore, to consider their development in terms of early Cenozoic plate interactions and resulting volcanic activity. This consideration may then provide information or restraints upon inferred plate movements off the western United States during the early Cenozoic, and may also provide further information concerning continental accretion in an important area that has been almost totally ignored in theories concerned with plate movements off western North America (Lipman et al. 1972).

## II. Nature of the Crust in western Washington and Oregon

The traditional concept that the Coast Ranges of western Washington and Oregon developed from a marginal eugeosyncline implies that continental crust underlies these Coast Ranges (Eardley 1962, Snively and Wagner, 1963). If, according to plate theory, the volcanic rocks within these Coast Ranges originated from a spreading oceanic ridge or by island arc volcanism, oceanic crust or modified oceanic crust should underlie

these Coast Ranges. Therefore, information concerning the nature and thickness of the crust may provide some insight into the geological evolution of the Coast Ranges.

The available evidence indicates that the crust in western Washington and Oregon is abnormally thin for continental crust. Tatel and Tuve (1955) obtained a depth to mantle of 19 km and a mantle velocity of 8 km/sec from the eastern margin of the Olympic Peninsula, west of Seattle, and Shor et al. (1968) obtained a depth to mantle of 18.4 km with a mantle velocity of 8.1 km/sec (station G7) on the continental shelf off western Washington, adjacent to the southern border of the Olympic Peninsula. Berg et al. (1966) report a crustal thickness of 16km from the central Oregon Coast Range, consisting of a 10 km upper part with velocities of 4.4-5.5 km/sec overlying a 6 km lower part with velocities of 6.6-7.4 km/sec. This result is consistent with the results of Shor et al. (1968) who obtained a depth to mantle of 16.5 km with mantle velocity of 8.1 km/sec (station G8) on the continental shelf south of the Columbia River, and a mantle depth of 11.9 km with a mantle velocity of 8.0 km/sec from the continental shelf approximately 100 km further south (station G9). All these results are consistent with the crustal and sub-crustal profiles illustrated by Dehlinger et al. (1970, Figs. 21 and 22) for central western Oregon and Washington, where their profiles have been constructed on the basis of both seismic refraction results and free-air gravity measurements. Figure 23 is a schematic illustration of crustal thicknesses in western Washington and Oregon based upon the seismic refraction results just discussed.

Computations based on reported stratigraphic thicknesses (Snively and

Wagner, 1963, 1964; Weaver 1945) give maximum true thicknesses of approximately 13 km of crustal rocks on the Olympic Peninsula, and in excess of 10 km in the Oregon Coast Ranges. The 13 km estimate for the Olympic Peninsula includes both the Solduc formation (3km) and the overlying Tertiary lavas and sedimentary rock (10 km). Compared with a crustal thickness of 18-19km obtained from seismic refraction studies (Tatel and Tuve, 1955, Shor et al. 1968), this leaves approximately 6km of crust to be accounted for on the Olympic Peninsula before the mantle is reached. It should be noted, though, that the true stratigraphic thickness of crustal rock is not strictly comparable with the thickness obtained by refraction studies since possible thickening of the crust by folding will be included in the thickness obtained by the refraction method. If the basal Solduc formation was deposited directly on oceanic crust, then the apparent 6km of crust before the mantle is reached may very well represent the combined thickness of the "Transition layer" (layer 2) and the "Oceanic layer" (layer 3) of the oceanic crust, which is reported as  $6.11 \pm 2.28$  km Shor et al. (1970).

On the basis of the thin crust in the central Oregon Coast Range, the chemical similarities between the Siletz River volcanics and Pacific Ocean abyssal tholeiites (Snively et al. 1968), and the geographic location of this area adjacent to the Pacific basin, Snively et al. (1968) have suggested that the crust underlying the central part of the Oregon Coast Range may be oceanic rather than continental. The base of the volcanic sequence is not exposed in the Oregon Coast Ranges, unlike the Olympic Peninsula, and it is not known whether a sedimentary sequence equivalent to the Solduc formation underlies the volcanics in western

Oregon. Snavely et al. (1968) consider the upper 10 km of crust in Berg et al's (1966) (Fig. 23) profile to be probably entirely represented by the Siletz River volcanic sequence, and that older rocks represent the entire lower 6 km, thereby implying that an unknown (meta-sedimentary ?) sequence underlies the volcanics in western Oregon. However, if an older metasedimentary sequence does underlie the volcanics, and comprises the entire lower 6 km, then their interpretation of the crustal structure is difficult to reconcile with their view that oceanic crust underlies the central Oregon Coast Range, and with the known structure of the oceanic crust (Shor et al. 1970). If the entire lower 6 km represents an older metasedimentary sequence, then this sequence was apparently deposited directly on the mantle, since mantle velocities were obtained by Berg et al. (1966) directly below the base of the 6 km layer (Fig. 23). A more reasonable interpretation is that this lower 6 km represents the oceanic crust and the seismic velocities of 6.6-7.4 km/sec found in this layer are consistent with this interpretation. If true, metasediments comprising layer 1 or part of layer 2 could still underlie the volcanics in western Oregon, and this would be consistent with geological evolution in western Washington and Oregon. If the Coast Ranges in western Washington and Oregon have originated by common geological processes, a southern equivalent to the Solduc formation of the Olympic Peninsula should be present in western Oregon.

### III. The Environment of Eruption of the lower-middle Eocene Basalts

#### 1. Ocean Floor Origin

If the lower-middle Eocene basalts of the western Washington and Oregon Coast Ranges are oceanic in origin, then they may have either been formed at a spreading oceanic ridge, or by island arc volcanism associated with subduction of oceanic lithosphere. If these basalts originated at or near the crest of an actively spreading oceanic ridge, they would need to be transported as part of an eastward moving oceanic plate and then be obducted onto the continental margin to arrive at their present location. If these processes did occur, it might reasonably be expected that major thrust zones, melange deposits, blueschist belts, and ultramafic mantle material, or at least some of these features, be found within the Coast Ranges of western Washington and Oregon. Such features have been described from other regions in the world where oceanic crust has been obducted e.g. the Franciscan, California Coast Ranges (Bailey et al. 1970; Ernst 1970; Page 1970), Papua (Davies 1968), New Caledonia (Coleman 1967; Brothers 1970), the Macquarie Islands (Varne et al. 1969), but so far none of these features have been reported from the Coast Ranges of western Washington and Oregon. Medaris and Dott (1970) and Medaris (1972) have described mantle-derived peridotites from southwestern Oregon, but these are associated with the episode of late Mesozoic continental underthrusting which deformed and emplaced the Franciscan rocks of the California Coast Ranges.

## 2. Island Arc Origin

Tholeiitic volcanism associated with the development of volcanic arcs has been described from Japan (Kuno 1966, Sugimura 1968), the South

Sandwich Islands (Baker 1968) and Melanesia (Jakeš<sup>V</sup> and White, 1969), and Jakes and Gill (1970) and Jakeš<sup>V</sup> and White (1972) have recently emphasised the importance of tholeiitic volcanism in the formation of volcanic arcs.

A volcanic arc model can account for the deformation of the Solduc formation (Weaver 1937, 1945) prior to the eruption of the lower Eocene basalts. Silver (1969, 1971) has described folded and faulted sediments near the base of the continental slope off northernmost California which he has attributed to underthrusting of the continental margin during the late Cenozoic. The deformation of the Solduc formation may therefore have been caused by early Cenozoic underthrusting off Washington.

The descriptions of the pattern of lower-middle Eocene volcanism in western Washington and Oregon given by Snavely and Wagner (1963) are also compatible with a volcanic arc model. The great local accumulations of basalt, the formation of basaltic islands with subaerial flows and inter-flow soil zones, and interbedded, locally-derived mud-flow breccia and conglomerate are strongly suggestive of the development of large volcanic edifices scattered along a volcanic arc.

Calc-alkaline magmatic activity associated with plate convergence and subduction is characteristic of igneous rocks of active oceanic and continental arcs (Dickinson and Hatherton, 1967; Dickinson 1968). By analogy, the existence of similar calc-alkaline magmatic activity during the early Cenozoic in Washington and Oregon would further corroborate the presence of a subduction zone off Washington and Oregon during that period,

and would lend weight to the model of volcanic arc development of the lower Eocene basaltic rocks of the western Washington and Oregon Coast Ranges.

Lipman et al. (1972) have summarised the characteristics of Cenozoic volcanism in the western United States, describing a shifting pattern of volcanism in relation to regional tectonic features, and relating these changing volcanic-tectonic systems to concepts of the plate-tectonic evolution of the northeastern part of the Pacific Ocean and western North America. They have described extensive calc-alkaline volcanism, consisting largely of andesite activity, from the western United States. Of particular interest, they have described Eocene calc-alkalic volcanism from Washington and Oregon, which consisted largely of andesitic, rhyodacitic and dacitic activity with related monzonite, diorite, granite, syenite and shonkinite plutonic intrusions (Fig. 24). On the basis of similarities in petrology and chemical variation between these and circum-Pacific andesites and related rocks and other volcanic arcs, and on inferred plate movements during the Cenozoic, Lipman et al. have concluded that the volcanic rocks of the western United States originated as the products of continental arc volcanism generated by the subduction of oceanic lithosphere along the western margin of the United States during and up to late Cenozoic time.

### 3. Conclusion

It would therefore appear that calc-alkalic magmatic activity, which is typically found in presently active continental and island arcs associated

with subduction of lithosphere, was also manifest during the early Cenozoic in Washington and Oregon, and thus corroborates the existence of a subduction zone off Washington and Oregon during that period. This, combined with the absence of geologic and tectonic features expected during obduction of oceanic crust, and the described pattern of lower Eocene volcanic activity, leads to the conclusion that a volcanic arc type of volcanism best accounts for the development of the western Washington and Oregon Coast Ranges.

#### IV. Inferred Early Cenozoic Plate Movements

The sea-floor magnetic anomaly patterns of the northeast Pacific (Atwater and Menard, 1970) indicate that only the western half of the expected symmetrical pattern is represented. This geometry indicates that there was once another plate lying to the east of the East Pacific Rise, the Farallon plate of McKenzie and Morgan (1969), which contained the missing anomalies. Since only fragments of this plate are now preserved, as the Cocos plate off Mexico and the Juan de Fuca plate off Oregon and Washington, the intervening portion must have been consumed in a subduction zone at its margin with the North American plate (Fig. 25). The presence of magnetic anomaly 9 (29 m.y.) (Atwater 1970, Fig. 3) off the coast of northern California indicates that the East Pacific Rise was still a viable spreading system until late Oligocene, and that subduction occurred along the western margin of the United States up until that time.

North of the Mendocino fracture zone the older magnetic anomalies have been destroyed by later spreading and subduction associated with the



Gorda and Juan de Fuca Ridges, and hence early Cenozoic plate interactions along the Washington-Oregon margin can only be inferred by extrapolation from the known plate configurations and rates of sea-floor spreading of the Pacific and North American plates in middle and late Cenozoic times.

Atwater (1970, Fig. 18) has attempted an extrapolation of a constant-motion plate tectonic model back into the Mesozoic and early Cenozoic, and illustrates a transform fault plate boundary along the western edge of North America (Fig. 25). Lipman et al. (1972) have commented on the fact that it is difficult to reconcile the presence of subduction-type calc-alkalic igneous rocks of late Cretaceous and early Cenozoic age in western North America with Atwater's (1970) interpretation that transform motion was taking place at this time along the western margin of the United States. Atwater's extrapolation relies upon a constant relative motion between plates, and she has stated that because the Kula-Pacific plate motion (Grow and Atwater, 1970) since 60 m.y. age is unknown, and since no compelling reason exists to suggest that it was constant, the extrapolation must be regarded as extremely tenuous. Atwater (written communication, 1972) has suggested that the Kula-American relative plate motion (Atwater 1970, Fig. 18) could easily have been in a more northeasterly direction. If true, this would have resulted in oblique subduction all along the western margin of the United States and Canada, and thus plate configurations would then be compatible with the presence of early Cenozoic subduction-type calc-alkaline igneous rocks in Washington and Oregon. However, there appears to be no evidence that subduction-type igneous activity occurred north of the southern tip of Vancouver Island

during the early Cenozoic (Souther 1972), which this change in plate motions demands. It seems more reasonable that the triple junction between the Kula, North American and Farallon plates had reached a position at the southern end of Vancouver Island by at least Eocene time (Fig. 25B); alternatively, a transform fault may have once intersected the ridge between the Kula and Farallon plates, such that the eastern end of the ridge lay further north and off southern Vancouver Island by Eocene time. Both configurations would allow subduction along Washington and Oregon during the early Cenozoic, and would still maintain strike-slip motion along western Canada.

#### V. Summary

Computed crustal thicknesses are in agreement with results obtained from on-shore and off-shore refraction studies, and the seismic velocities and depths to mantle obtained from the seismic studies indicate that the western Washington and Oregon Coast Ranges comprise a sequence of crustal rocks 10-13 km in thickness overlying oceanic crust, or modified oceanic crust, approximately 6 km in thickness. The existence of a subduction system off western Washington and Oregon during the early Cenozoic is suggested by the presence of early Cenozoic subduction-type calc-alkalic igneous rocks in Washington and Oregon. Although inferred plate movements indicate a transform boundary along Washington and Oregon during the early Cenozoic (Atwater 1970), it is felt that the tenuous nature of the extrapolations involved allow the interpretation of subduction to be equally, if not more, valid. The inferred oceanic crust underlying the Coast

Ranges in western Washington and Oregon and the probable presence of an off-shore subduction system during the early Cenozoic, suggest that the lower Eocene basaltic rocks in these Coast Ranges may have originated either from a spreading oceanic ridge or by island-arc volcanism. The absence of geological and tectonic features associated with the obduction of oceanic crust, suggest that an island arc origin is more appropriate. The patterns of volcanism found within the lower Eocene basalts are also more suggestive of volcanic eruptions along the length of an island arc.

Fig. 23 Crustal thicknesses obtained by seismic refraction methods in western Washington and Oregon.

Numbers inside columns refer to seismic velocities; numbers outside columns refer to crustal thicknesses.

1 - from Tate1 and Tuve (1955). Dot indicates position of receiving station.

2 - from Berg et al. (1966). Dot indicates position of shot point.

G7-G10 - from Shor et al. (1968). Dots indicate positions of receiving stations.

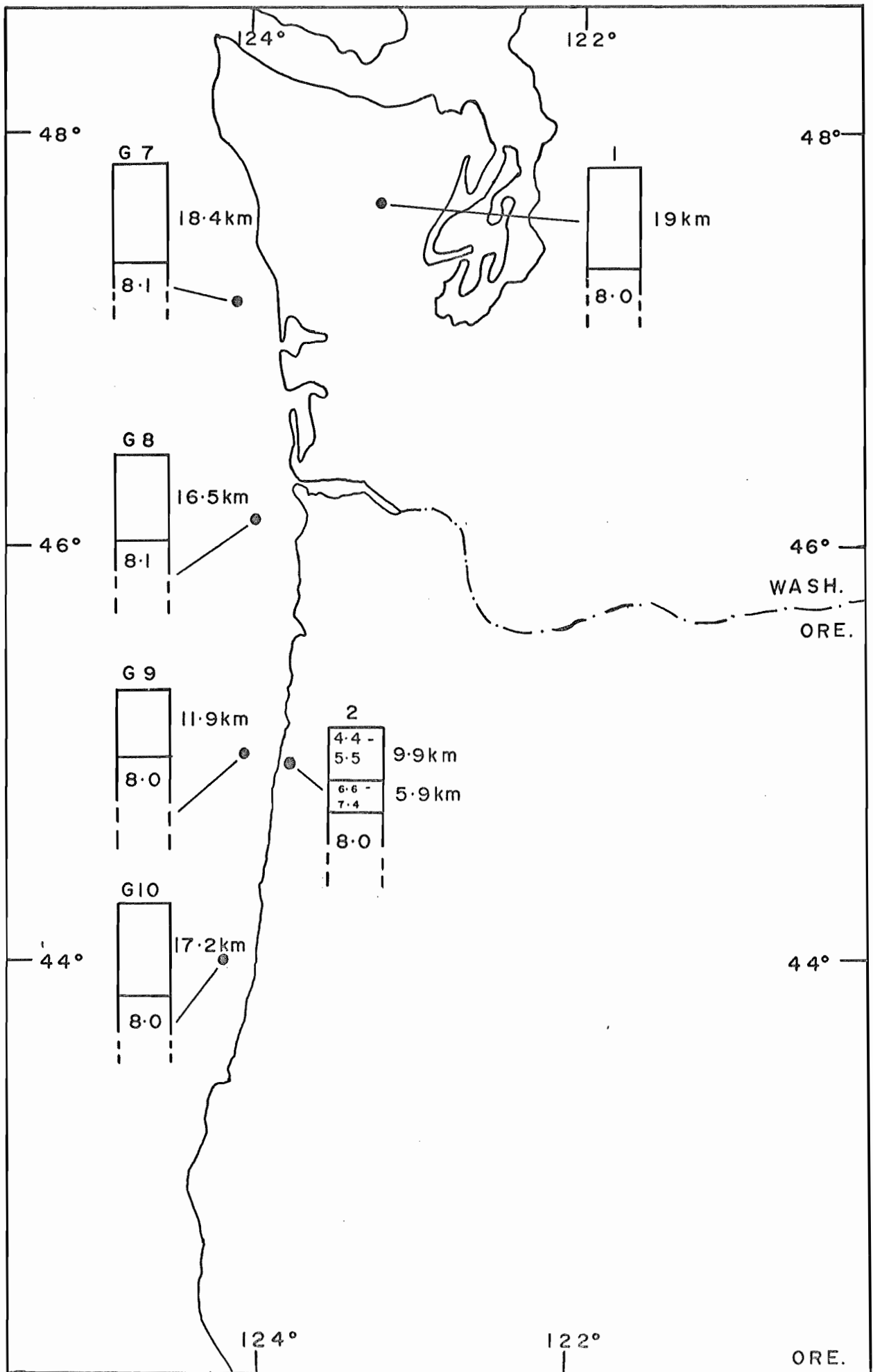
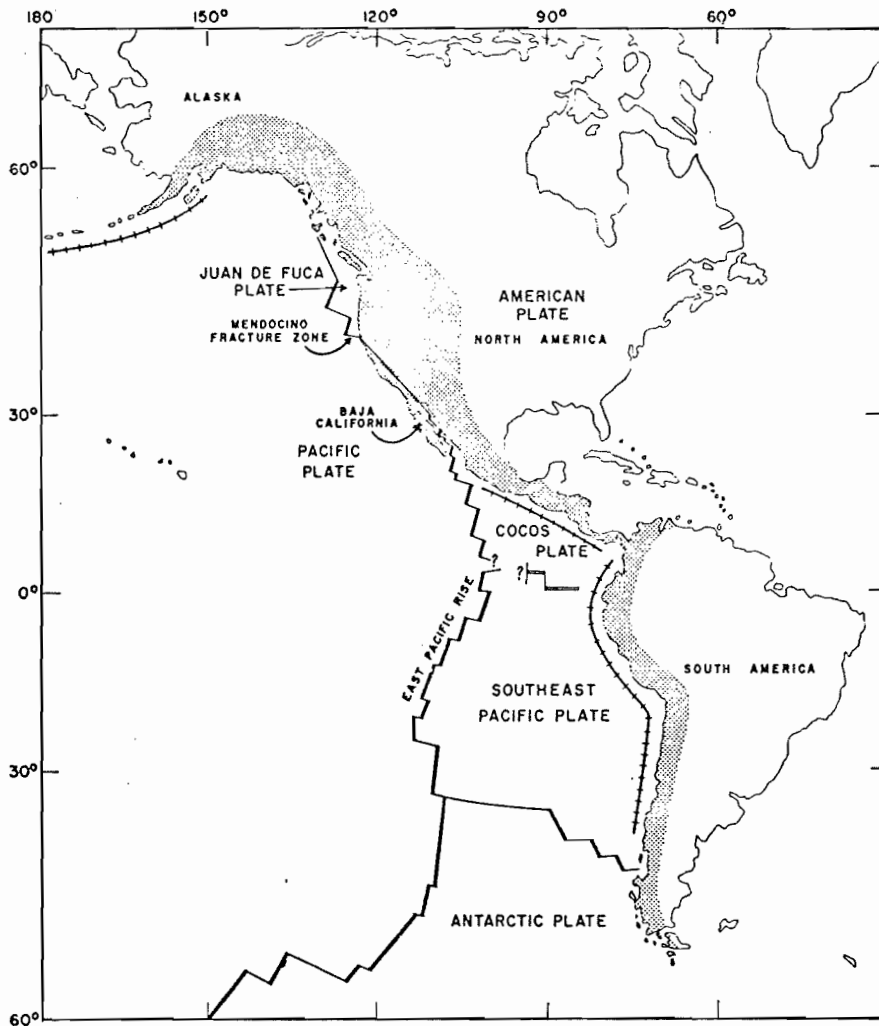


FIG. 23

ORE.  
CALIF.

Fig. 24 Present boundaries of major tectonic plates of the eastern Pacific region, and the approximate extent of late Mesozoic and Cenozoic calc-alkalic igneous belt (stippled) in western North America.

(from Lipman et al., 1972)



Present boundaries of major tectonic plates of the eastern Pacific region (modified from Bullard (1969) and Morgan (1968)), and approximate extent of late Mesozoic and Cenozoic calc-alkalic igneous belt (stippled) in western North and South America. Belt is about 500 km wide, except in the Western United States, where it is as much as 1500 km wide (width of belt in Alaskan area is exaggerated by Mercator projection). Double lines indicate spreading ridges; single lines, transform faults; 'railway tracks', oceanic trenches.

FIG. 24

Fig. 25 Plate configurations off western North America during the late Mesozoic, the Eocene and the present day. Single lines are transform faults; double lines are spreading centres; and hatched lines are zones of subduction. Large arrows show motions of plates; small arrows show relative motions at points along plate boundaries.

SF = San Francisco            S = Seattle

Fig. 25A Late Mesozoic (60 m.y.) plate configurations. From Atwater 1970, Fig. 18C.

Fig. 25B Eocene (50 m.y.) plate configurations. Modified from Atwater 1970, Fig. 18 to show subduction off western Washington and Oregon.

Fig. 25C Present day plate configurations off western North America. From Atwater 1970, Fig. 4.



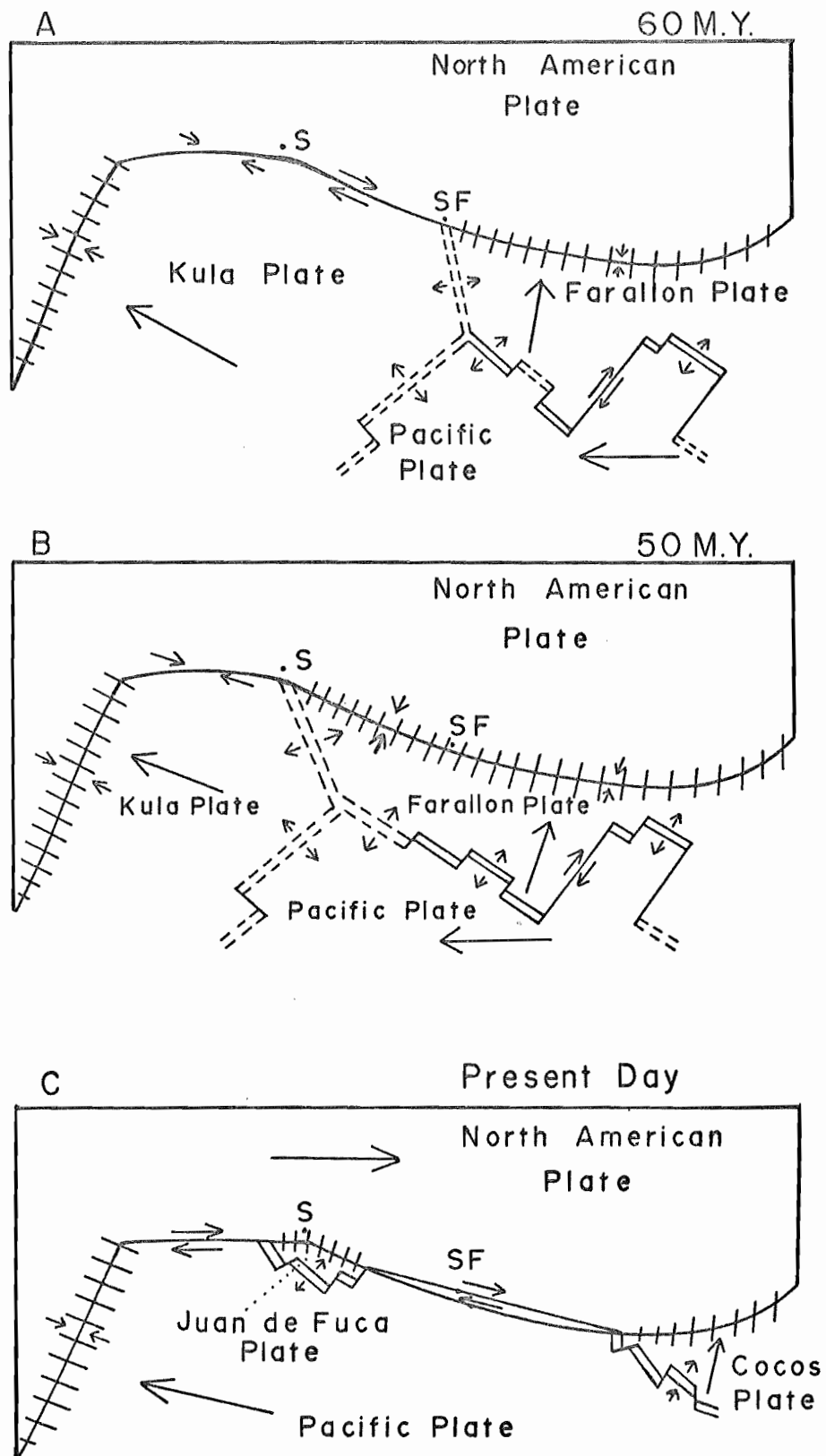


FIG. 25

## CHAPTER 6

### MAGMA GENESIS IN ISLAND ARCS

#### I. Introduction

The spatial and temporal aspects of magma genesis in island arcs have been discussed by Jakeš and Gill (1970) and Jakeš and White (1972). It appears that tholeiitic volcanism is the earliest manifestation of island arc evolution, followed either stratigraphically or laterally by calc-alkaline volcanics and finally, in old island arcs which have clearly defineable seismic planes dipping towards the continent, by shoshonitic (K-mafic lavas) activity. Assuming that the magmas erupted in island arcs originated on or close to the Benioff zone within the upper mantle, the spatial relations of lavas in island arcs dictate that tholeiitic magmas must have originated from shallower depths within the mantle than the calc-alkaline magmas. This is contrary to what may be expected if calc-alkaline magmas originate by the partial melting of the upper oceanic lithosphere, and the implications of this fundamental observation will be considered.

The petrogenesis of calc-alkaline magmas in volcanic arcs has been the subject of numerous studies e.g. Taylor and White (1966), Taylor (1969), Green and Ringwood (1968, 1969), Yoder (1969), Green (1972), but apart from the work of Jakeš and Gill (1970), Jakeš and White (1972) and Fitton (1971) relatively little study has been devoted to the origin of tholeiites in island arcs. It has been postulated that the Olympic

Peninsula basalts are the products of island arc volcanism, and petrogenetic models based upon phase equilibria studies have been proposed which account for the genesis of these basalts. Thus, opportunities exist to test the compatibility of these petrogenetic models with proposed petrogenetic schemes for island arc tholeiites, with the available chemical data on island arc tholeiites, and with the described thermal structures in island arcs.

## II. Origin of Island Arc Tholeiites

1. Projections of Olympic Peninsula basalts in relation to phase boundaries within the CMAS tetrahedron demonstrated that dry partial melting of peridotite at intermediate or high pressures, followed by olivine fractionation from the magmas during their ascent to the surface could give rise to erupted basalts with compositions identical to Olympic Peninsula basalts.
2. Fitton (1971) has proposed that the breakdown and melting of amphibole in a hydrous, amphibolitic upper lithosphere produced the island arc tholeiitic series. He has developed a theoretical model based on the melting relations of the upper 5 km of oceanic crust, consisting of amphibolite in the upper layer grading downwards into gabbro, allied with the thermal model of a downgoing slab of oceanic lithosphere given by Oxburgh and Turcotte (1970). Fitton (1971, Fig. 1) shows that liquids are produced by the breakdown and partial melting of amphibole in the depth range 40-60km. These liquids, Fitton claims, are tholeiitic and give rise to the island arc tholeiites when erupted. On general principles, it is extremely unlikely that basaltic liquids can form from the

partial melting of amphibole, and Fitton fails to substantiate his argument by providing the relevant chemical data on these liquid compositions. While amphibole compositions may approach those of basalt, the first liquids formed on partial melting of basalt are definitely not basaltic, but are composed of the low-melting fraction of basalt. Similarly, the liquids produced by the partial melting of amphibole would be unlikely to have the composition of basalt. Even the partial melting of amphibolite, itself derived from basalt, does not produce basaltic liquids but liquids which are andesitic to dacitic in composition (Green and Ringwood, 1968). Another difficulty with this proposal, as Green (1972) has pointed out, is that it would involve near complete melting of the upper part of the lithosphere to produce the island arc tholeiitic series, where the most abundant rock type has a silica value of 53% (Jakeš and Gill, 1970).

3. Jakeš and Gill (1970) suggest that the island arc tholeiitic series may have originated by the mixing of partial melts from the upper part of the descending lithosphere and from the mantle of the overlying plate. They envisage the andesitic to dacitic low-melting fraction of the upper lithosphere (Green and Ringwood, 1968) rising above the Benioff zone, causing the overlying mantle to undergo partial fusion to yield a tholeiitic magma. They suggest that unspecified mixing of the tholeiitic and andesitic magmas would then yield magmas which would show the continuum between the calc-alkaline and island arc tholeiitic series. By its very nature, such a process is difficult to prove or disprove, and the resulting magma composition will obviously depend upon the volume relations of the two magma types. However, it may be possible by such a process to

produce low-silica tholeiites, involving little or no mixing of magmas, and high-silica tholeiites involving greater degrees of magmas mixing.

4. The rare earth element (REE) abundances in island arc tholeiites have also been discussed by Jakeš and Gill (1970), who have noted the following features:

- i) Enriched chondritic REE abundance patterns are characteristic of rocks of the island arc tholeiitic series.
- ii) No significant Eu anomalies occur in rocks of the island arc tholeiitic series.
- iii) REE abundance patterns in oceanic (abyssal) and island arc tholeiites are almost identical.

The above data suggests that island arc tholeiites have originated from a relatively primitive, (unfractionated?) material, namely, mantle peridotite. Fractional crystallisation of olivine and minor pyroxene could have occurred from the partial melt of peridotite without significantly changing the overall primitive REE pattern. Also, the absence of Eu anomalies suggests that olivine fractionation rather than plagioclase fractionation has dominated the early fractionation histories of island arc tholeiites. The low MgO, Ni and Cr contents found in island arc tholeiites (Jakeš and Gill, 1970) are also compatible with extensive olivine fractionation during the evolution of island arc tholeiitic magmas.

### III. Origin of the Calc-Alkaline Suite in Island Arcs

#### 1. Partial Melting of the Oceanic Crust under Hydrous Conditions

The eruption of calc-alkaline magmas in volcanic arcs over relatively

long periods and within relatively restricted zones e.g. the Cascades, strongly suggests that new source material for the production of calc-alkaline magmas must be made continually available at the requisite depth down the Benioff zone. Such a source material may well be the upper basaltic crust of the subducting oceanic lithosphere. Partial melting of basalt or amphibolite under hydrous conditions ( $P_{H_2O} < P_{load}$ ) in the pressure range 5-10kb and temperature range 900°C-1050°C (Green and Ringwood, 1968; Holloway and Burnham, 1972) has yielded liquids which range in composition from andesite to dacite, and which if erupted, or allowed to fractionate prior to eruption, would yield lavas representative of the calc-alkaline suite. Eruption of the liquids obtained at 10kb and higher temperatures (1000°-1050°C) would also be compatible with the andesite liquidus temperature of 1000-1050°C obtained at 2kb water pressure by Brown and Schairer (1967).

A process can be envisaged then, in which basaltic or amphibolitic oceanic crust undergoes partial melting at pressure of 10kb and temperatures in the range 1000°-1050°C, with  $P_{H_2O} < P_{load}$ , to produce calc-alkaline magmas. As the oceanic crust becomes depleted in its low melting fraction and continues down the subduction zone, "new" oceanic crust will enter the region of magma genesis and calc-alkaline magmas will continually form and be erupted at the surface. However, while this theory can explain the genesis of calc-alkaline magmas in island arcs, it is difficult to produce magma at a depth of 30-40 km (10kb) and have it erupted on the continental side of the island arc tholeiites. This apparent anomaly may simply mean that the temperature-pressure conditions

required for partial melting of oceanic crust inferred from laboratory experiments are not available at the corresponding depths down the Benioff zone. It is also possible, if partial melting does begin at 30-40 km, that insufficient liquid will be generated to produce a viable magma. Consequently, partial melting should increase with depth down the Benioff zone, thus making it easier to erupt calc-alkaline magmas at greater distances away from the trench. However, if partial melting does take place at greater depths down the Benioff zone, the magmas so formed will have more time to modify their compositions by fractionation before being erupted at the surface. The fact that andesite is the most voluminous member of the calc-alkaline suite may suggest that the magmas formed at greater depths are not in fact andesitic, but more basic in composition.

## 2. Partial Melting of the Mantle under Hydrous Conditions

Scar et al. (1967) and Kushiro et al. (1968) have established that enstatite melts incongruently in the presence of water at upper mantle pressures, yielding a silica-enriched liquid. It is thus conceivable that when peridotite melts in the presence of water, the initial liquid produced is silica-rich. Yoder (1969) has discussed the implications of this melting behaviour in the context of the diopside-forsterite-silica system studied by Kushiro (1969). At 20kb, and under hydrous conditions, the assemblage Fo + Opx + Cpx begins to melt at 1220°C, producing a liquid which if separated would crystallise as a two-pyroxene quartz assemblage. Thus hydrous melting of peridotite

may yield a magma capable of crystallising as a quartz-normative calc-alkalic andesite. If it can be demonstrated that considerable amounts of water are available in the upper mantle, perhaps by water being carried down the Benioff zone, by dehydration reactions involving amphibole or by outgassing from deeper levels within the mantle, then this mechanism of producing calc-alkaline magmas would have to be seriously considered. However, it is difficult to imagine this process operating over long periods, unless it is taking place at various levels within the mantle, since the peridotite should eventually become depleted in its low melting fraction.

### 3. Partial Melting of Eclogite

Green and Ringwood (1968, 1969) have conducted experimental work on the melting of synthetic quartz eclogite under high pressure, and suggest that calc-alkaline magmas can be obtained by the partial melting of quartz eclogite at depths of 80-150 km in the mantle. According to their experimental data (Green and Ringwood, 1968, Table 5), partial melting of quartz eclogite may begin at temperatures as low as 1370°C at 27kb. O'Hara and Yoder (1967) have demonstrated experimentally that eclogite apparently has a much higher temperature of beginning of melting. They found that partial melting at 30kb of 2 whole-rock natural eclogites (kyanite eclogite and bimineralic eclogite) yielded temperatures of beginning of melting at 1525°C and 1515°C, respectively, and have concluded (p. 115) "Where both (garnet peridotite and eclogite) rock types are available to be melted under normal circumstances, liquid would be produced preferentially from the garnet peridotite". This means, if



O'Hara and Yoder are correct, that it is impossible to partially melt eclogite without prior melting of the overlying mantle. If this assertion is correct, and calc-alkaline magmas are in fact derived by partial melting of quartz eclogite, then tholeiitic and calc-alkaline lavas should be erupted contemporaneously, and in comparable volumes, in island arcs. There is no evidence for this type of contemporaneous volcanic activity in island arcs, which may suggest that calc-alkaline magmas are not derived by the partial melting of quartz eclogite. In addition, the low melting temperature of eclogite which is required in the model of Green and Ringwood (1968) can only be obtained from a quartz eclogite, since bimineralec eclogite will have a higher melting temperature. It is debatable whether such quartz eclogite really does exist at depths in a descending lithospheric slab, which according to Green (1972), has already undergone at least one episode of partial melting to produce calc-alkaline liquids.

#### 4. Fractional Crystallisation of Basaltic Magma under Constant Oxygen Pressure

Osborn (1959, 1962, 1969 a, b) and Roeder and Osborn (1966a, b), on the basis of experimental studies in iron-bearing silicate systems under controlled oxidation conditions, have argued that calc-alkaline magmas may be derived by the fractional crystallisation of basaltic magma under constant, or near constant, oxygen pressures. The constant oxygen pressure is supposed to be obtained by diffusion of water into basaltic magmas which have been intruded into, and consequently heated, geosynclinal

sediments. Under such conditions of high temperature and high  $P$   $H_2O$ , the water dissociates resulting in high  $pO_2$  and magnetite is supposed to crystallise out early from the basaltic magmas. Fractionation of magnetite, along with olivine and pyroxene, is alleged to prevent iron enrichment in the magmas and leads to the production of silica-enriched magmas which give rise to the calc-alkaline lavas when erupted.

Green and Ringwood (1966) have estimated that over 65 per cent fractional crystallisation of basaltic magma must occur before andesitic liquids are produced. The volume of calc-alkaline rocks in island arcs implies a fractionation process which is rather too efficient to be credible. Such a fractionation process also demands that large quantities of mafic cumulates be formed along with the development of the calc-alkaline magmas, for which Green and Ringwood (1968) claim there is little or no field evidence. Even if such a fractionation process is feasible, the process demands the formation of prodigious volumes of basalt magma from a region of the mantle which is unlikely to reach melting temperatures in the vicinity of the descending lithospheric slab (to be discussed).

Taylor (1965, 1969) has discussed the trace element geochemistry of andesites and associated calc-alkaline rocks, and has indicated that the concentrations and ratios of certain elements are not what would be expected if calc-alkaline magmas were derived by the fractional crystallisation of basalt. Elements, such as Cs, Rb, Tl, Ba, the rare earths, and Th, U, Zr and Hf, which are normally concentrated by processes of fractional crystallisation or differentiation, are low in absolute

abundance in calc-alkaline rocks. In addition, K/Rb, K/Cs, Ba/Rb and Eu/Gd ratios are all high, and Rb/Sr and Ba/Sr ratios are low in andesites. Operation of processes of fractional crystallisation produce the opposite effect (low K/Rb, K/Cs, Ba/Rb, Eu/Gd, and high Rb/Sr and Ba/Sr).

The removal of magnetite from basaltic magmas will deplete the residual liquid in the ferromagnesian elements which enter magnetite. Duncan and Taylor (1968) have analysed 5 magnetites from andesites and dacites from New Zealand and found that they contained high contents of vanadium (0.5-0.8%). They have argued that the loss of magnetites having these compositions from basaltic magma will strongly deplete the magma in vanadium. By considering the differences in vanadium contents between basalts and andesites, which are small, they calculate that the maximum amount of magnetite which could have been fractionated from basalt to give this difference is only 2%. Holloway and Burnham (1972) have recently demonstrated that subtraction of amphibole from tholeiitic magmas may actually deplete the residual liquid in iron, indicating that no addition of oxygen to the magma is necessary to present iron enrichment.

#### IV Thermal Structure in Island Arcs and Tholeiitic Magma Genesis

Oxburgh and Turcotte (1970) and Minear and Toksöz (1970) have devised mathematical models which attempt to portray the thermal structure in island arcs. A basic difference in the thermal structure between the two models is that in the model of Minear and Toksöz higher temperatures lie closer to the upper surface of both the subducting plate and the overlying plate, and that hotter isotherms lie immediately above the Benioff

zone and have a marked upward inflection. An assumption made by Oxburgh and Turcotte may account for this difference between the two models. Oxburgh and Turcotte assume that the mantle within the subducting oceanic lithosphere was completely depleted in its low-melting basaltic fraction when it upwelled beneath an oceanic ridge. By making this assumption, they argue that heat production in the descending lithosphere from radioactivity and phase changes will then be negligible since most of the radioactive elements will have entered the low-melting basaltic fraction, and the residual mantle, having been depleted in most of its  $Al_2O_3$  and now composed largely of olivine and orthopyroxene, should be stable to depths of at least several hundred kilometres. It seems most unlikely that almost 100 km of mantle can have been totally depleted of its basaltic fraction by partial melting beneath an oceanic ridge. A more reasonable interpretation is that a narrow, vertical zone of the upwelling mantle underwent partial melting and only this part became depleted in its low-melting basaltic fraction. Minear and Toksöz have taken into account the effects of heating due to radioactivity and phase changes in the descending lithospheric slab, and consequently their thermal model will be used in preference to the model of Oxburgh and Turcotte in the following discussion.

According to Minear and Toksöz (1970, Fig. 10), for a slab 160 km thick subducting at the rate of 1 cm/year, the temperature in the mantle of the overlying plate immediately above the Benioff zone is approximately 1400°C at a depth of 100 km (30kb), and 1200°C at a depth of 60 km (20kb). In a descending slab 80 km thick (Minear and Toksöz, 1970, Fig. 12),

subducting at 1 cm/year, the temperature distribution above 100 km is practically identical to that in the 160 km slab. According to the data given by Ito and Kennedy (1967), these temperatures are similar to the temperatures of beginning of melting of a natural garnet peridotite at the corresponding pressures. However, if partial melting can occur, at these depths, it is questionable whether sufficient magma can be generated at these temperatures to produce a viable magma. According to the isotherm distribution (Minear and Toksöz, 1970, Fig. 10) substantial melting of mantle in the overlying plate can only occur in the depth range 100-300 km, where the temperature lies between 1400°C and 2000°C.

However, when the 1200°C, 1400°C and 1600°C isotherms (Minear and Toksöz, 1970, Fig. 10) and their corresponding pressures within the mantle (obtained from the highest inflection point of the isotherms just on or above the Benioff zone) are plotted in relation to the peridotite solidus (Fig. 26), it can be seen that these T-P points all fall below the solidus. Thus within the depth range of 60-100 km (20-50kb) it appears impossible to partially melt the mantle overlying the Benioff zone by simply appealing to increasing temperature with depth. It is emphasised, though, that this deduction is entirely dependent upon the accuracy of the inferred thermal structure in island arcs. If the solidus curve is extrapolated linearly to a pressure of 100kb (not shown), the only temperature point from the thermal model which falls above the solidus is the point T 2000°C, P 80kb (240 km). For a plate subducting at 45° at 1 cm/year (as in Fig. 10, Minear and Toksöz, 1970), 24 m.y. will be required before the plate reaches a depth of 240 km, and therefore

before the tholeiitic magmas can be generated. This is contrary to the observed spatial and temporal relations of tholeiites in island arcs where tholeiites lie closest to the trench, are the first products of volcanism, and are therefore presumably related to the initiation or early stages of subduction. Appealing to higher rates of subduction (Minear and Toksöz, 1970, Fig. 11) does not solve this problem either, since the isotherms are depressed further into the mantle thus making partial melting of the mantle even more difficult or impossible at the appropriate pressures.

#### V. The Relationship of the Subduction Zone to Magma Production

Thus the problem still exists: how can tholeiites form by partial melting of mantle peridotite if they are causally related to the initial or early stages of subduction? The following explanation is tentatively offered.

The oceanic geotherm lies closest to the peridotite solidus within the low velocity zone (Fig. 26) i.e. at depths in the vicinity of 100-150 km (30-50kb); consequently, the mantle within the low velocity zone is nearer its temperature of beginning of melting than the overlying or underlying mantle. When subduction within the oceanic lithosphere is initiated, stress relief associated with the development of the Benioff fault zone may be propagated through the mantle into the low velocity zone, presumably in the direction of the incipient fault plane. This stress relief may trigger the formation of mantle diapirs (Green and Ringwood, 1967) in the metastable low velocity zone, which is already

less dense than the overlying mantle. If the rising diapir cools adiabatically, without losing significant heat by conduction to the surrounding mantle, partial melting will occur in the diapir when its temperature intersects the peridotite solidus, to produce picritic magma. Olivine fractionation from these primary picritic magmas during their ascent to the surface would then produce tholeiitic lavas, as was described for the Olympic Peninsula basalts.

This theory has several advantages to it:

- i) it explains the origin and early eruption of tholeiitic lavas in island arcs. Other theories on the origin of island arc tholeiites, such as the mixing of magmas (Jakeš<sup>V</sup> and Gill, 1970) and the melting of amphibole (Fitton 1971) are unconvincing, and fail to answer why tholeiitic lavas, which should be generated after calc-alkaline lavas if partially melted from the same type of material, since they require higher melting temperatures, are erupted prior to calc-alkaline lavas;
- ii) it can account for the observed concentrations of REE, MgO, Ni and Cr (Jakeš<sup>V</sup> and Gill, 1970) in island arc tholeiites by the fractionation of olivine from the picritic magmas during their ascent to the surface;
- iii) it can account for the range of silica values found in island arc tholeiites and the apparent continuum between tholeiites and more intermediate compositions (Jakeš<sup>V</sup> and Gill, 1970; Jakeš<sup>V</sup> and White, 1972) by appealing to fractionation of picritic magmas during their ascent to the surface. The rate of magma ascent is regarded as the important factor controlling the erupted lava compositions. Magmas which ascend relatively slowly will have time to fractionate substantial amounts of olivine, and possibly pyroxene, thus enriching the residual magma in silica. Further

fractionation of olivine and pyroxene at low pressures will enhance the trend towards silica enrichment, and will produce the more intermediate members of the island arc tholeiitic series. Magmas which ascend at a relatively rapid rate will not fractionate as much olivine or pyroxene, and hence will have erupted compositions which will tend to be more silica deficient. The latter case would apply to the Olympic Peninsula basalts, which largely have silica contents less than 50 per cent (Table 1).

This model for the origin of island arc tholeiites is compatible with the petrogenetic scheme developed for the Olympic Peninsula basalts in Chapter 4 and briefly restated in Section B I of this chapter. The fact that the site of magma generation in the new model may be at greater depths than that proposed for the Olympic Peninsula basalts is not regarded as a serious problem: liquids generated at great depths within the mantle which subsequently equilibrate at shallower levels within the mantle can produce magmas identical to, and indistinguishable from, magmas produced by partial melting at the shallower levels. Thus, the Olympic Peninsula magmas may have been generated within the low velocity zone, or just above it, became arrested during their movement towards the surface at depths between 60 and 100 km and allowed to equilibrate, before their final eruption towards the surface.

Wyllie (1971) has also proposed a scheme for the generation of basaltic magmas in island arcs which invokes the rise of mantle diapirs from the low velocity zone. Wyllie suggests that migration of water, released by the dehydration of the descending slab, into the overlying low velocity zone caused incipient melting within the mantle above the Benioff zone.



The effects of water influx and incipient melting lower the density and viscosity of the peridotite above the Benioff zone, thus facilitating diapiric uprise from within the low velocity zone. These diapirs rise adiabatically and undergo partial melting to produce picritic magmas when they reach the level of the dry peridotite solidus, and tholeiitic magmas are eventually formed by the fractionation of these picritic magmas, as has been previously described.

This theory also has the advantages which were described for the stress-relief theory. However, according to Wyllie's theory, tholeiitic magmas cannot be generated until the descending slab has intersected the low velocity zone i.e. at depths greater than at least 100 km, whereas by the stress-relief theory, it is possible to produce tholeiites before the slab has descended to such depths. Until information is gained concerning the time of actual eruption of tholeiitic lavas in island arcs in relation to the initiation of subduction, both theories remain valid. The theory of Wyllie also demands that calc-alkaline magmas be generated at greater depths than has been suggested from experimental investigations upon the genesis of calc-alkaline magmas. Until more detailed information concerning physical processes within the mantle is available, it would appear that the problems concerned with magma genesis in island arcs will remain unsolved.

#### VI. Proposed Scheme for Magma Genesis in Island Arcs

A scheme for the processes of magma genesis in island arcs can now be summarised. Stress relief associated with the initiation of subduction

and the formation of the Benioff fault plane causes the rise of mantle diapirs from the metastable low velocity zone, which rise adiabatically and subsequently undergo partial melting to produce picritic magmas. Alternatively, the rise of diapirs from the low velocity zone may be initiated by the influx of water, derived from the dehydration of the descending slab, into the mantle above the Benioff zone. Fractionation of these picritic magmas, which may be arrested and allowed to equilibrate at shallower levels within the mantle, during their ascent to the surface then produces tholeiitic magmas. These magmas, when erupted, constitute the island arc tholeiitic series, the compositions of which will be governed primarily by the rate of magma ascent. As subduction develops, and the oceanic lithosphere reaches the requisite depth down the Benioff zone, the basaltic or amphibolitic crust will begin to partially melt under hydrous conditions to produce sufficient liquid to form viable calc-alkaline magmas, and also to satisfy the spatial aspects of calc-alkaline rocks in island arcs. Continued subduction provides "new" oceanic crust to be partially melted, and thus eruption of calc-alkaline magmas occurs over long time periods in island arcs.

## VII. Summary

The spatial and temporal aspects of magma genesis in island arcs demand that tholeiites be the first products of volcanism and originate from shallower depths within the mantle than the calc-alkaline magmas. Partial melting of the mantle above a descending lithospheric slab by simply appealing to increasing temperature with depth down the Benioff zone is not feasible, as has been demonstrated by plotting T-P points

obtained from on or close to the upper surface of the descending slab in relation to the peridotite solidus. The adiabatic rise of mantle diapirs from the low velocity zone, initiated either by stress relief associated with the beginning of subduction or by influx of water into the mantle above the Benioff zone, allows partial melting of peridotite to occur, producing picritic magmas, when the diapirs reach the level of the dry peridotite solidus. Fractionation of olivine and possibly pyroxene from these picritic magmas as they ascent to the surface can account for the range in silica contents found in island arc tholeiites, depending upon the rate of magma ascent, and can also account for the low contents of MgO, Ni and Cr and the primitive REE patterns found in island arc tholeiites.

No satisfactory explanation yet exists which can both account for the genesis of calc-alkaline magmas in island arcs, and the observed spatial and temporal relations. Partial melting of oceanic crust under hydrous conditions at pressures around 10kb provides a satisfactory explanation for the genesis of calc-alkaline magmas, their erupted temperatures and the eruption of these magmas over long time periods. However, it is still necessary to invoke greater degrees of partial melting at greater depths down the Benioff zone to allow calc-alkaline magmas to be erupted on the continental side of the island arc tholeiitic series. Partial melting of peridotite under hydrous conditions may be important in the genesis of calc-alkaline magmas, but it is not yet known whether sufficient water is available in the mantle for this mechanism to be effective. Partial melting of eclogite at high pressures is regarded

as an unlikely process to produce calc-alkaline magmas since the melting temperatures of eclogite are greater than those of the surrounding mantle peridotite. If calc-alkaline magmas are produced by the partial melting of eclogite, so also should picritic magmas be produced at the same time, and in comparable volumes, by the partial melting of the mantle. Evidence of this volume of picritic magmas generation is completely lacking in island arcs. Fractional crystallisation of basaltic magma under conditions of constant or near constant  $pO_2$  cannot produce the large volumes of calc-alkaline lavas found in island arcs without prior formation of prodigious volumes of basaltic magma and the operation of an extremely efficient fractionation process. In addition, the trace element chemistry of calc-alkaline rocks in island arcs indicate that they have not formed by a fractionation or differentiation process.

Fig. 26 T-P diagram showing positions of the peridotite solidus and the oceanic geotherm.

Peridotite solidus constructed from data given by Ito and Kennedy (1967) on the melting history of natural garnet peridotite KA 64-16.

Oceanic geotherm from Ringwood et al., 1964. Dots indicate T-P points obtained from Minear and Toksöz, 1970, Fig. 10.

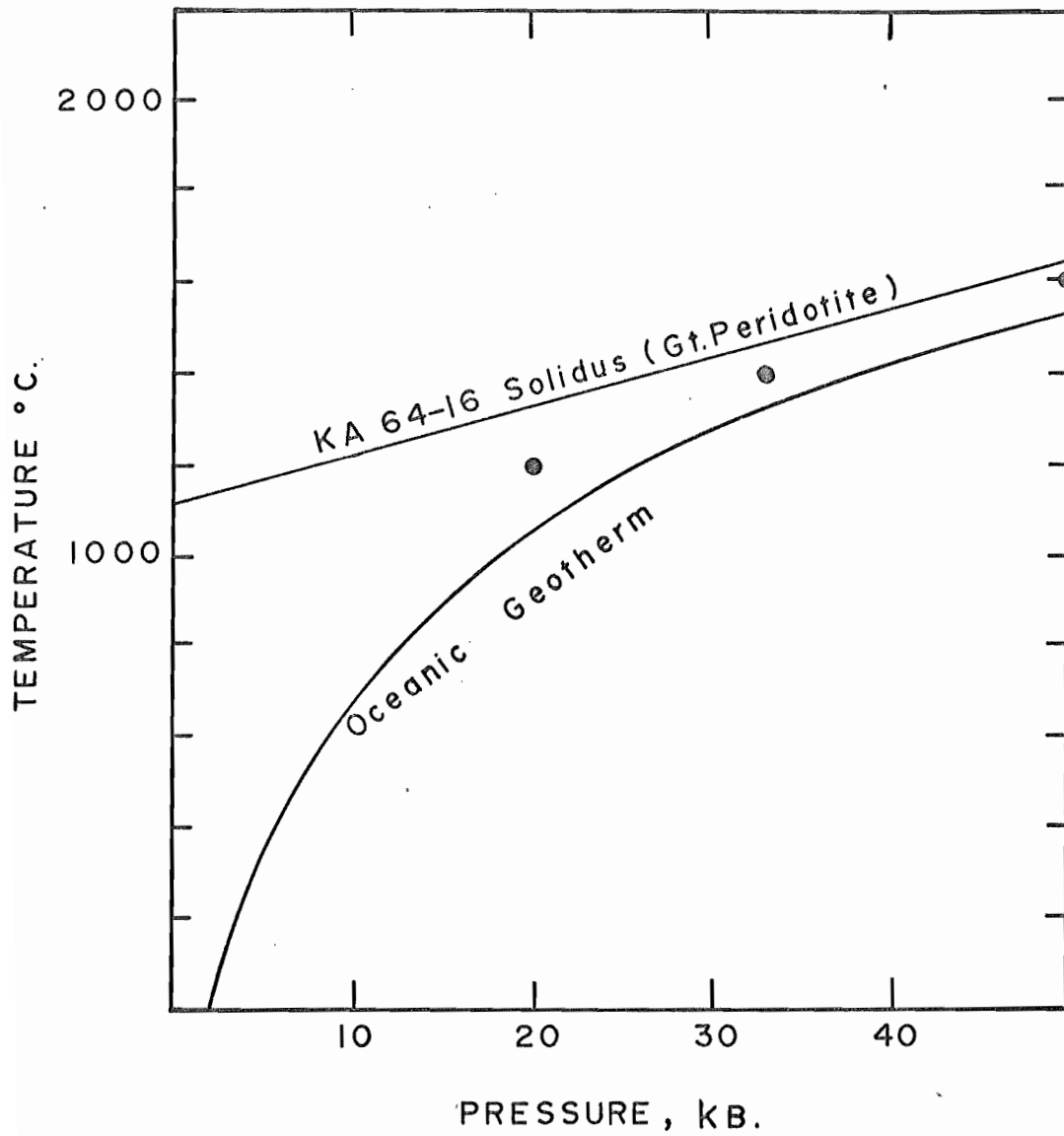


FIG. 26

CHAPTER 7

CONCLUSIONS

I. Concerning Spilites

First, the definition of spilite will be reiterated. Ideally, spilite should refer to a rock containing albite, clinopyroxene and chlorite, exhibiting relict igneous textures, and containing high soda. However, because of the diversity in mineralogy and chemistry found in spilites, it is not desirable to give a restrictive definition of the term spilite. If the term spilite is to be retained as a petrographic term, it is better that it be used in a general sense to include all the spilitic rock variants which contain albite, which have been derived from a common basaltic parent, but which still retain their igneous textural characteristics.

1. Spilitic magmas have yet to be observed, and the presence of high temperature igneous minerals and textures in spilites argue against the existence of a low temperature, primary spilitic magma.
2. Experimental work on the melting of basalt under hydrous conditions has given little evidence for the formation of a spilitic magma, nor does it corroborate any aspect of the hydrous crystallisation origin for spilites.
3. Autometamorphism of basaltic magmas can produce spilite on a small scale within relatively thick sills, but by itself cannot produce thick sequences of spilites, particularly in spilitic pillow sequences, unless a sodium-enriched fluid became concentrated in the magma chamber

and later reacted with the magma to produce a spilite.

4. Burial metamorphism or low grade metamorphism of basaltic rocks can create suitable T-P conditions for the development of spilitic mineralogy, but it is unlikely that thick sequences of spilites can form by this method alone if metamorphism operated strictly as a closed system. However, biased sampling of pillow lava sequences where spilites are present may have led to the erroneous conclusion that these sequences are entirely spilitic in composition.

5. In thick sequences of spilitic pillow lavas and in sills of albite diabase, sodium addition from fluids external to these magmas must be invoked to account for the overall increased sodium contents of these sequences. The most likely sources of sodium are in sea water, in sea water trapped within submarine basalt piles, and in the interstitial fluids of marine sediments.

6. Diffusion of sea water or trapped interstitial waters into cooling basaltic magmas, or into crystallised basaltic magma bodies during metamorphic reheating, is an important process in the genesis of spilites.

7. It is concluded that spilitic bodies may form by one of two processes, both of which involve the diffusion of a sodium-bearing fluid into either a cooling basaltic magma, or into crystallised basaltic magma during metamorphic reheating. Albite diabases most likely have formed by the diffusion of interstitial marine pore waters into a cooling basaltic sill. A magmatic intrusion emplaced in wet sediments will locally heat the interstitial fluids rapidly, increasing the  $P_{H_2O}$ . If the intrusion is emplaced in impermeable or semi-permeable sediments such as muds or



silts, the  $P_{H_2O}$  may become greater than the lithostatic pressure at that particular depth in the crust. Under such conditions, water can be expected to diffuse into the cooling magma and subsequently effect the spilitisation of the original mineralogical components. Even under low  $P_{H_2O}$  conditions, there can be some diffusion of  $H_2O$  into the magma if the magma was initially relatively "dry". Diffusion will continue until gradients in  $H_2O$  fugacity between the magma and its surroundings are eliminated.

Thick sequences of spilitic pillow lavas most likely have formed by the diffusion of sea water into basaltic lavas during reheating under low grade metamorphic conditions. The effects of burial metamorphism, tectonic movement, conditions of high heat flow, and possible thermal metamorphism from intrusion of ultramafic material, such as found at oceanic ridges and in island arcs, cause reheating of basaltic lava piles and allow sea water to diffuse into the basalts, which thus brings about spilitisation.

8. Oxygen isotope evidence from a variety of igneous rocks has proven that water has diffused into cooling magma bodies and into basaltic lavas upon metamorphic reheating. In addition, these studies have also shown that only sea water and not fluids derived by outgassing of the mantle, took part in the metamorphism which produced greenstones dredged from oceanic ridges.

9. As the result of items 7 and 8 above, it is generally concluded that spilites are derived from magmas or rocks of basaltic composition involving soda metasomatism through the agency of sea water or sea water trapped in submarine basalt piles or marine sediments. However, this general

conclusion does not rule out the formation of certain occurrences of spilites which may originate in different ways.

10. Projections of basalts and spilites onto the "spilite plane"

$\text{CMS}_2\text{-M}_5\text{AS}_3\text{-CAS}_2$  within the CMAS tetrahedron have suggested that there is a continuous spectrum in chemistry between spilites and basalts, and that spilites may be derived from oceanic-type basalts rather than continental basalts. However, more representative samples and statistical data is required before definite conclusions can be made from these suggestions.

11. The variation in mineralogical components, and hence in chemical components, found in spilites simply reflect the mobility and redistribution of elements under conditions of low grade metamorphism in the presence of a fluid phase. Depending on the partial pressures of the components of this fluid phase e.g.  $p\text{CO}_2$ ,  $p\text{O}_2$ ,  $p\text{H}_2\text{O}$  etc., various low grade metamorphic minerals such as calcite, epidote, prehnite, pumpellyite etc. may form.

## II. Olympic Peninsula Volcanic Rocks

1. Basalts, pillow basalts, diabases, breccias and tuffs comprise the volcanic rocks on the Olympic Peninsula.

2. The basalts are largely aphanitic with plagioclase and augitic pyroxene as the common microphenocryst phases, and often contain devitrified or chloritised glass in the groundmass. Olivine is present in some basalts but is rarely abundant.

3. Diabases are present as two distinct groups: normal diabases with calcic plagioclase, and albite diabases with sodic plagioclase.

4. Opaque minerals, largely titanomagnetite, are common in all rock types, and are frequently partially or completely altered to sphene.

5. On the basis of the rocks studied, spilites do not appear to be common in the volcanic sequence. The majority of rocks appear to be merely altered or metamorphosed basalts and diabases.
6. The presence of pillow basalts, hyaloclastite material and intercalated marine sediments indicate that part of the volcanic sequence, at least, was erupted under submarine conditions. The presence of skeletal plagioclase and titanomagnetite, quenched intergrowths of plagioclase and pyroxene, and strong zoning in some plagioclase grains testify to rapid cooling of magma.
7. Nearly all the volcanic rocks have suffered some degree of alteration or metamorphism.
8. Metamorphism varies from diagenetic changes involving only devitrification of volcanic glass through zeolite facies into prehnite-pumpellyite facies. Lack of stratigraphic control in sampling prohibits giving any comprehensive report on the metamorphism of the volcanic sequence.

### III. Chemistry

1. The major changes in the original chemistry of the Olympic Peninsula volcanic rocks have resulted from hydration, carbonitisation and oxidation. These changes have affected the majority of rocks.
2. Serious departures from basalt chemistry have occurred where there has been a gain in Na, K and Ca, and a loss in Mg and Ca from certain rocks. However, these changes are the exception rather than the rule.
3. Recalculation of the chemical analyses to 100% on a total H<sub>2</sub>O and CaCO<sub>3</sub>-free basis, with adjusted iron oxides values, restores the analyses

as closely as possible to the original chemistry. This recalculation is justified on the basis of the results of re-applying the basalt screen to the recalculated analyses, and also on the basis of the behaviour of certain olivine-bearing rocks (rocks 30 and 34) when projected into the CMAS tetrahedron.

4. Normative calculations on the basis of the recalculated analyses indicate that the majority of analysed rocks are tholeiitic. The appearance of nepheline in the norms of certain rocks may be the result of analytical error or mild alkali addition, but in the case of rock 77 this cannot be conclusively proven, and this rock may represent a true alkali basalt.

5. No strong trends are apparent on any of the major element variation diagrams.

#### IV. Petrogenesis

1. Olympic Peninsula basalts exhibit a marked low pressure cotectic behaviour. This behaviour can be seen in both the normative and CMAS projections.

2. A restricted range in chemical composition in the Olympic Peninsula volcanic rocks is indicated from both the normative and CMAS projections.

3. Olympic Peninsula basalts can be regarded as fairly evolved, having equilibrated at low pressures with plagioclase and clinopyroxene, and this accounts for the absence of strong chemical trends.

4. Derivation of the Olympic Peninsula basalts by the eruption of primary magmas formed during low pressure partial melting of four-phase peridotite, or by subsequent fractionation of a low pressure partial melt, is rejected

on the grounds that the compositions of these rocks cannot be derived from such magmas according to the known phase relations.

6. Partial melting of four-phase peridotite at intermediate pressures (12-25kb) or high pressure (25-30kb) can give rise to the Olympic Peninsula basalt compositions if the primary liquids subsequently underwent at least olivine fractionation during their ascent to the surface, and in the case of the 30kb partial melting, some harzburgitic fractionation as well.

7. The feasibility of olivine fractionation to account for the observed rock compositions is demonstrated by the fact that olivine control lines can be drawn from the olivine composition point ( $M_2S$ ), through points representing the initial and successive liquids formed on greater degrees of partial melting of four-phase peridotite at intermediate and high pressures, which effectively encompass the whole range of Olympic Peninsula basalt compositions.

8. Orthopyroxene fractionation alone cannot account for the observed Olympic Peninsula compositions.

9. Projections of spilites, basalts and Olympic Peninsula volcanic rocks onto the "spilite plane" within the CMAS tetrahedron have shown that the Olympic Peninsula volcanic rocks are dominantly basaltic in composition.

10. In general, Olympic Peninsula volcanic rocks should be regarded as altered or metamorphosed basalts and diabases rather than spilites.

11. Spilitic rocks are found within the Olympic Peninsula volcanic sequence, but are in the minority.

12. This survey has indicated that albite diabases occur more frequently than spilitic lavas.

13. The validity of the Olympic Peninsula being a classic spilite locality must remain open to question.

V. Geological Evolution of the western Washington and Oregon Coast Ranges

1. Oceanic crust, or modified oceanic crust, and not continental crust underlies the Coast Ranges in western Washington and Oregon.
2. The presence of Eocene, subduction-type, calc-alkaline igneous rocks in Washington and Oregon indicate that subduction occurred off Washington and Oregon during the early Cenozoic.
3. The lower-middle Eocene basalts of the western Washington and Oregon Coast Ranges originated as products of island arc volcanism in response to early Cenozoic subduction, and do not represent obducted oceanic crust.
4. The origin of these basalts by island arc volcanism explains certain geological features found in the Coast Ranges of western Washington and Oregon which are difficult to explain adequately if obduction of oceanic crust is invoked.

VI. Early Cenozoic Plate Movements

1. Inferred early Cenozoic plate configurations (Atwater 1970) are not compatible with observed geological features in Washington and Oregon; therefore changes in the inferred relative plate motions must be made to account for these geological features.
2. Migration of the Kula-American-Farallon triple junction to the southern end of Vancouver Island by early Cenozoic time best explains

the development of subduction along Washington and Oregon during the early Cenozoic.

3: A more northeasterly change in the Kula-American relative plate movement would also have produced subduction along Washington and Oregon during the early Cenozoic. However, this change would also have produced early Cenozoic subduction along the western margin of Canada, for which there is little geological evidence.

#### VII. Magma Genesis in Island Arcs

1. The spatial and temporal aspects of tholeiitic and calc-alkaline rocks in island arcs are fundamental features which must be explained in theories of magma genesis. Existing theories on the genesis of magmas in island arcs either fail to give convincing answers on the genesis of these magmas, fail to explain the spatial and temporal aspects of magma generation, or fail to explain both.

2. Tholeiitic magma generation in island arcs must be associated with the initial or early stages of subduction.

3. Partial melting of the mantle at depths down the Benioff zone has been shown to be unlikely to occur, although it is emphasised that this conclusion is entirely dependent upon the proposed thermal structure in island arcs.

4. The genesis of island arc tholeiites, their erupted compositions and the observed spatial and temporal relations of tholeiites in island arcs can be explained by invoking the adiabatic rise of mantle diapirs from within the low velocity zone. The initiation of diapiric uprise may be caused by stress relief associated with the development of subduction, or

by influx of water into the mantle above the Benioff zone, although this latter theory demands that subduction must be fairly advanced before tholeiitic magma can be generated.

5. The rate of magma ascent is regarded as the important factor controlling the composition of the erupted tholeiites. Magmas which ascent at a relatively rapid rate will not fractionate as much olivine and pyroxene as magmas which ascend at a relatively slower rate, and hence will have erupted compositions which tend to be lower in silica than those magmas erupted at relatively slower rates.

6. The continued eruption of calc-alkaline magmas over long time periods in island arcs strongly suggests that new source material for the production of calc-alkaline magmas must be made continually available at the requisite depth down the Benioff zone. Such a source may be the upper basaltic crust of the subduction lithosphere.

7. Partial melting of basalt or amphibolite under hydrous conditions ( $P_{H_2O} < P_{load}$ ) in the pressure range 5-10kb at temperatures of 900-1100°C is regarded as the most suitable mechanism for producing calc-alkaline magmas in island arcs, but this mechanism does not entirely satisfy the spatial relations found in island arcs.

8. Partial melting of the mantle under hydrous conditions may produce calc-alkaline magmas, which if erupted, would satisfy the spatial relations of magmas in island arcs. However, it has yet to be demonstrated that sufficient water is, or can be made available at the necessary depths within the mantle.

9. Fractional crystallisation of basaltic magma under conditions of constant or near constant  $pO_2$  cannot produce the large volumes of calc-alkaline lavas which are found in island arcs without appealing to an



extremely efficient fractionation process and the prior formation of prodigious volumes of basaltic magma. In addition, trace element chemistry of calc-alkaline rocks indicates that these rocks have not formed as the result of a fractionation or differentiation process.

10. Partial melting of quartz eclogite at great depths within a descending lithospheric slab to produce calc-alkaline magmas must be regarded as unproven in light of the experimental results of O'Hara and Yoder (1967), and by the fact that it is by no means certain that the descending slab is composed of quartz eclogite at the requisite depth.

APPENDIX A

PETROGRAPHIC DESCRIPTIONS OF ANALYSED ROCKS

Rock No.	Description
5.	Chloritised basalt. Ophitic, sericitised plagioclase laths, altered pyroxene and titanomagnetite set in chloritised groundmass heavily charged with altered opaque minerals. Pyroxene completely altered to chlorite-sphene mixtures. Chlorite ubiquitous as interstitial patches replacing glass, in veinlets and vesicles, and replacing plagioclase. Scattered calcite patches common, and zeolite present as small scattered patches in the groundmass. Vesicles infilled with chlorite-calcite mixtures. (Plate I)
8.	Albite diabase, possibly sheared. Dusty, zoned plagioclase laths ( $An_5$ ), some of which are bent, augitic clinopyroxene and skeletal titanomagnetite, largely altered to sphene. Quartz and epidote veinlets present, with interstitial chlorite. Green-brown hornblende and stilpnomelane (?) present in some interstitial chlorite patches. Apatite present.
13.	Diabase, with (sub)-ophitic texture. Strongly zoned plagioclase laths ( $An_{50}$ ), badly sericitised in places, pinkish clinopyroxene and skeletal chlorite. Prehnite found in association with chlorite, and in places replacing plagioclase. Rare interstitial quartz.

1411. Diabase, with (sub)-ophitic texture. Zoned plagioclase laths ( $An_{50}$ ), showing varying degrees of sericitisation, pinkish clinopyroxene and skeletal titanomagnetite, with interstitial patches of green chlorite. Prehnite found in association with chlorite, and in places completely replacing plagioclase. (Plate II)
16. Porphyritic diabase. Zoned plagioclase laths ( $An_{54}$ ), clinopyroxene and interstitial, partially altered titanomagnetite. Occasional clinopyroxene and plagioclase phenocryst, usually sericitised. Green chlorite found in veinlets, patches and replacing plagioclase. Rare epidote.
171. Albite diabase. Zoned plagioclase ( $An_4$ ), clinopyroxene and skeletal titanomagnetite. Minor sericitisation of plagioclase, with partial replacement by fibrous prehnite in some grains. Prehnite also present interstitially. Green chlorite found interstitially and also pseudomorphing olivine. Apatite present.
23. Chloritised basalt. Sheafs of plagioclase intergrown with granular pyroxene, with interstitial opaque minerals and green chlorite. Plagioclase has extremely mottled appearance and is completely replaced by calcite in patches. Calcite, chlorite and quartz veinlets presents, with minor interstitial quartz.
30. Chloritised basalt. Sericitised plagioclase laths, clinopyroxene, olivine microphenocrysts and opaque minerals, largely altered to

sphene, with interstitial green chlorite. Olivine completely pseudomorphed by green chlorite. Irregular blebs of sphene scattered throughout, with chlorite veinlets present. (Plate III)

34. Chloritised basalt. Sericitised plagioclase laths, clinopyroxene and olivine microphenocrysts with interstitial green chlorite patches. Olivine completely pseudomorphed by green chlorite, and opaque minerals completely altered to sphene. Calcite patches and irregular blebs of sphene scattered throughout.
38. Vesicular basalt. Skeletal plagioclase microphenocrysts set in a devitrified and partly chloritised groundmass containing plagioclase microlites and small birefringent (? pyroxene) grains. Vesicles largely infilled with calcite or green chlorite, although several are infilled by prehnite, quartz and pyrite.
40. Lithic tuff, probably sheared. Heterogeneous mixture of partially kaolinitised albite crystals and pale green, ragged actinolite crystals set in a streaky, partially chloritised and kaolinitised matrix. Prehnite scattered throughout and also replacing plagioclase.
43. Basalt. Plagioclase microphenocrysts ( $An_{54}$ ) and clinopyroxene microphenocrysts, showing partial or complete replacement by pale green actinolite, set in a devitrified groundmass composed of intergrowths of skeletal plagioclase and feathery clinopyroxene. Groundmass partially chloritised with minor chlorite veinlets and rare pumpellyite. (Plate IV)

47. Analysed by H. Baadsgaard, Geology Department, University of Alberta. Hand specimen not available for study.
48. Basaltic breccia (hyaloclastite). Angular to subrounded fragments of basaltic glass, some of which contain small plagioclase laths exhibiting flow orientation. Glass shows various stages of replacement by green, sometimes spherulitic chlorite and more rarely by analcite. Some glass fragments are fresh, dark and almost isotropic; others are completely replaced by chlorite. (Plate V)
57. Extremely fine-grained, altered basalt, containing the occasional plagioclase and pyroxene crystal. Rock largely altered to a mosaic of very small, granular, first order mineral. Extensive, very fine chlorite veinlets give the rock a fragmental appearance in thin section. Scattered calcite patches common.
60. Porphyritic, albite diabase. Sericitised albite laths, poikilitic clinopyroxene and skeletal titanomagnetite, with the occasional albite phenocryst. Interstitial green chlorite, prehnite present in veinlets and replacing plagioclase, and rare zeolite.
69. Albite diabase with (sub)-ophitic texture. Plagioclase laths ( $An_{10}$ ), clinopyroxene and subhedral titanomagnetite, partially altered to sphene. Green chlorite present interstitially and pseudomorphing olivine. Interstitial calcite, prehnite and laumontite also present.

77. Albite diabase. Plagioclase laths ( $An_4$ ), clinopyroxene and skeletal to subhedral titanomagnetite, with considerable calcite present in veinlets and interstitially. Green chlorite present interstitially, and possibly pseudomorphing olivine, with prehnite also found interstitially and largely associated with calcite. Rare interstitial epidote. Apatite present.
80. Gabbro. Cumulus plagioclase, strongly zoned with sodic borders (oligoclase), large intercumulus clinopyroxene and titanomagnetite. Plagioclase largely replaced by zeolite, with interstitial analcite also present. Clinopyroxene completely replaced by green-brown chlorophaeite in places.
91. Hypocrystalline basalt, fresh and unaltered. Plagioclase laths ( $An_{54}$ ), granular hourglass clinopyroxene and titanomagnetite, with interstitial devitrified glass. (Plate VI)
32. Pillow basalt, approximately 25 cm in diameter. Outer part of pillow composed of brown, devitrified glass containing euhedral microphenocrysts of olivine and feathery quench intergrowths of platioclase and clinopyroxene. Inner parts of pillow more crystalline with intergrowths of plagioclase and granular clinopyroxene, and olivine microphenocrysts. Vesicles are common in both zones and are infilled by calcite and chlorite. Olivine is completely pseudomorphed in every case by either bowlingite, chlorite, calcite, or intimate mixtures of these phases. Small,

equant, red-brown spinels are commonly found poikilitically enclosed in the pseudomorphed olivines. Calcite veinlets are common, and minor quartz veinlets and interstitial quartz is present, and is also found rimming several vesicles. Haematite staining is common. (Plates VII-X)

10. Pillow basalt, approximately 20 cms in diameter. Outer part of pillow composed of brown glass and spherulitic, feathery quench intergrowths of plagioclase and clinopyroxene, with skeletal plagioclase and clinopyroxene microphenocrysts. Inner zone more crystalline, composed of feathery aggregates of plagioclase and clinopyroxene, plagioclase and clinopyroxene microphenocrysts and interstitial opaque mineral. Rare olivine was present in either zones. Calcite veinlets common, and calcite and chlorite are common vesicle infillings. Minor quartz veinlets present with quartz rimming several vesicles. Haematite staining is present, and pumpellyite is associated with calcite in veinlets.  
(Plates XI-XIV)

Plate I. Rock 5 (x30, crossed polars). Chloritised basalt. Sericitised plagioclase, chloritised pyroxene and interstitial glass (dark, mottled patches) and altered titanomagnetite.

Plate II. Rock 14 11 (x30, plane polarised light). Diabase. Ophitic plagioclase, clinopyroxene, skeletal and subhedral titanomagnetite with interstitial chlorite (light grey and dark, mottled patches).





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5750  
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Plate III. Rock 30 (x30, plane polarised light). Olivine basalt. Olivine microphenocrysts (pseudomorphed by chlorite) set in groundmass composed of granular clinopyroxene and sericitised plagioclase laths. Irregular blebs of sphene within olivine grains.

Plate IV. Rock 43 (x30, plane polarised light). Basalt. Skeletal plagioclase and clinopyroxene microphenocrysts set in a devitrified and partly chloritised groundmass.

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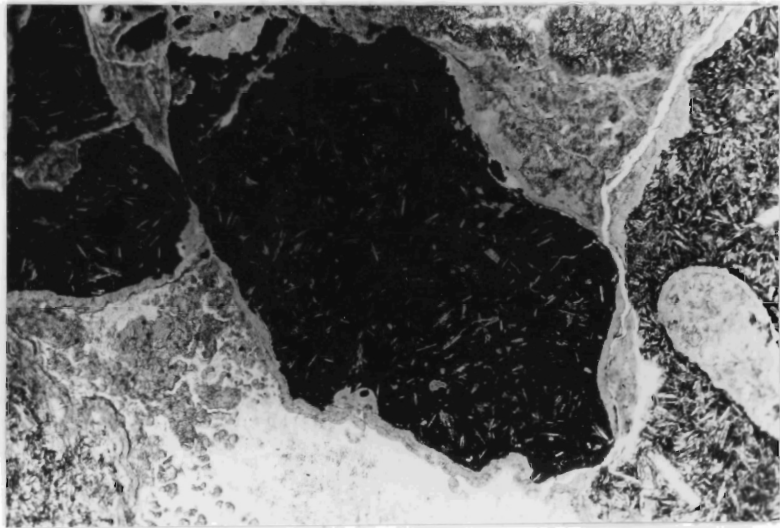


0170



Plate V. Rock 48 (x30, plane polarised light). Basalt breccia. Fragments of glassy basalt showing replacement by chlorite (grey) and zeolite (white).

Plate VI. Rock 91 (x30, plane polarised light). Hypocrystalline basalt. Granular clinopyroxene, plagioclase laths, titanomagnetite with interstitial patches of devitrified glass (large, dark patches).



1925

1925



1925

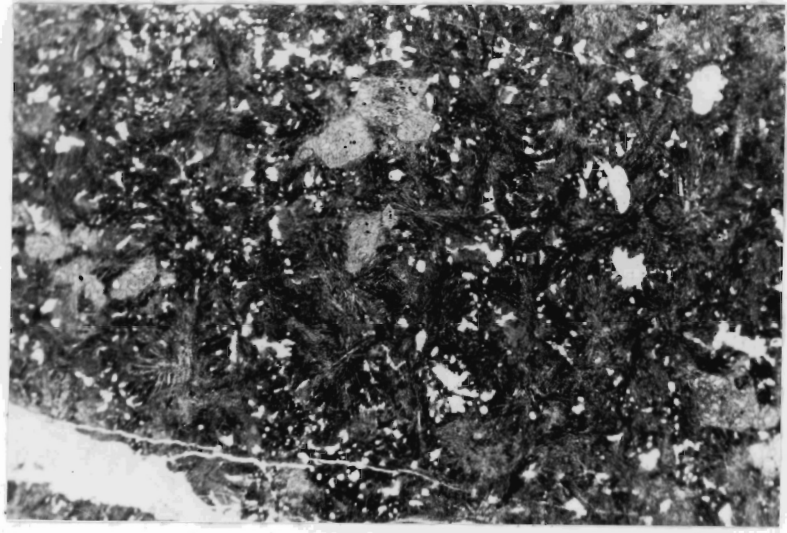
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Plate VII. Rock 32 (x30, plane polarised light). Chilled margin of pillow basalt. Olivine microphenocrysts set in quenched intergrowth of plagioclase and pyroxene, with interstitial chlorite (white) and calcite veinlet.

Plate VIII. Rock 32 (x105, plane polarised light). Chilled margin of pillow basalt in Plate VII, but at greater magnification, showing detail of the quenched groundmass. Olivine microphenocrysts poikilitically enclosing spinels.



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22  
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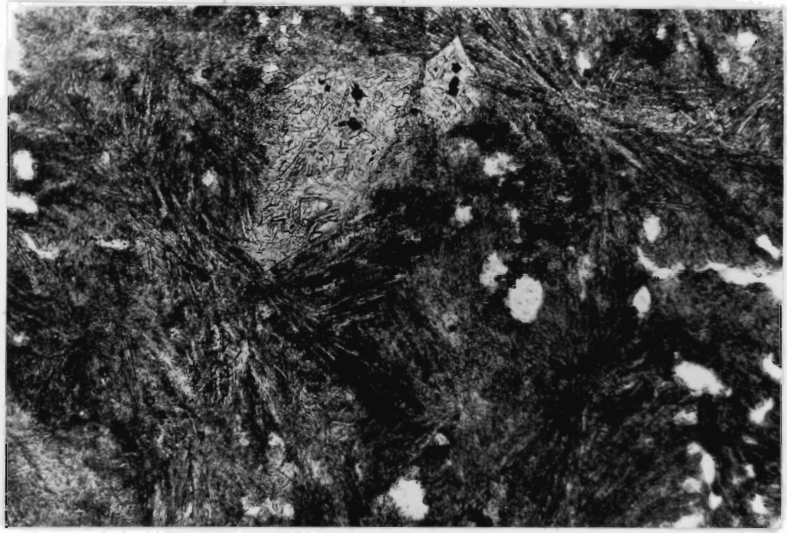


Plate IX. Rock 32 (x30, plane polarised light). Crystalline interior of pillow basalt. Olivine microphenocrysts, intergrown plagioclase and clinopyroxene (white), with interstitial opaque mineral (black). Clacite-infilled vesicle in lower right.

Plate X. Rock 32 (x105, plane polarised light). Crystalline interior of pillow basalt in Plate IX, but at greater magnification, showing greater detail of the intergrown plagioclase-pyroxene groundmass.





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-1425  
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03297



03297  
-1425

Plate XI. Rock 10 (x30, plane polarised light). Chilled margin of pillow basalt. Skeletal plagioclase and clinopyroxene microphenocrysts set in spherulitic, quenched intergrowths of plagioclase and pyroxene.

Plate XII. Rock 10 (x105, plane polarised light). Chilled margin of pillow basalt in Plate XI, but at greater magnification, showing detail of quenched plagioclase-pyroxene intergrowths with rimming by opaque mineral.

1967  
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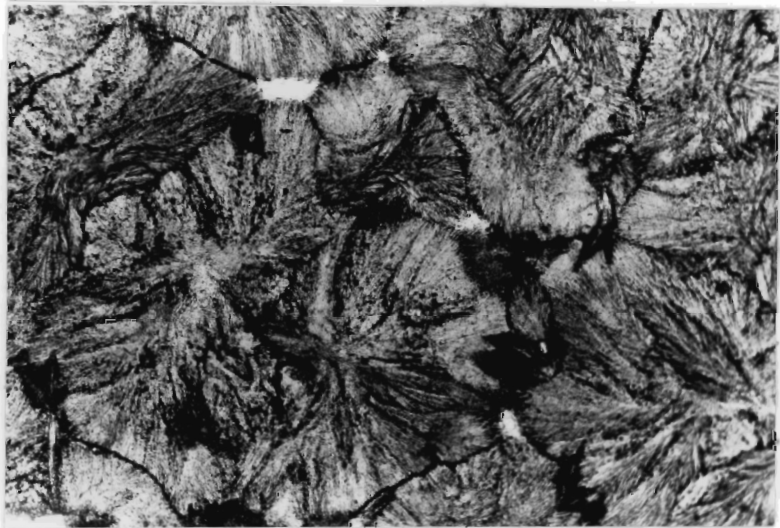
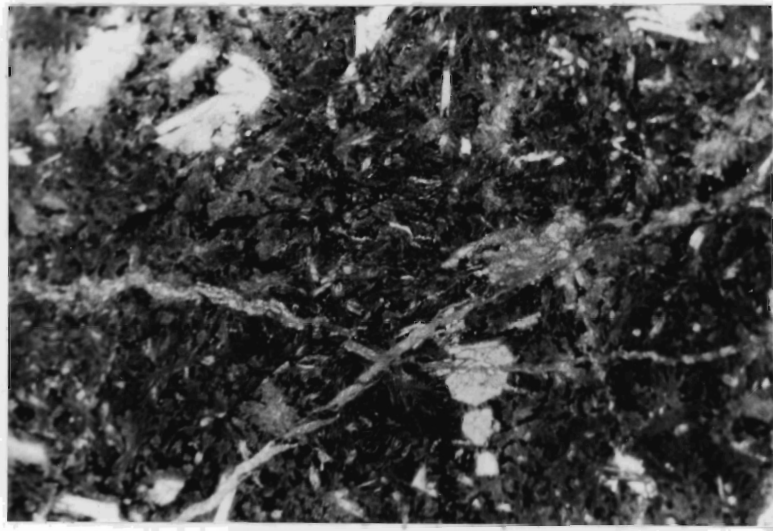


Plate XIII. Rock 10 (x30, plane polarised light). Interior of pillow basalt. Clinopyroxene and plagioclase microphenocrysts set in feathery intergrowth of plagioclase and pyroxene, with interstitial opaque mineral. Pumpellyite veinlets present.

Plate XIV. Rock 10 (x105, plane polarised light). Interior of pillow basalt in Plate XIII, but at greater magnification, showing detail of feathery intergrowth of plagioclase and pyroxene with interstitial opaque mineral.



1965  
MTC 1000000  
1965



1965

APPENDIX B

ANALYTICAL METHODS

I. Sample Preparation

Rock samples were first scrubbed with a nylon brush under hot, running water to remove any adhering foreign particles and loose, weathered surfaces, and then rinsed in distilled water. Any remaining weathered surfaces were removed by trimming with a diamond-edged sawblade, followed by further scrubblings and rinsings in distilled water. The remaining fresh rock samples were cut into thin slices with the diamond-edged sawblade, rinsed in distilled water, and broken manually between paper into small fragments. Approximately 50 gram aliquots of the rock fragments were then ground for 3 minutes in the hardened steel vessel of a Spex Industries Inc. #8500 Shatterbox, and the coarse powder sieved through a nylon mesh sieve to remove the less than 100 mesh fraction. The remaining coarse powder was further ground for 3 minute periods in a Fritsch Pulverisette 2 automatic grinder with agate mortar and pestel, until all the powder passed the 100 mesh sieve. The powders were then homogenised for 15 minutes in a P-K Twin Shell Dry Blender with plexiglas shells, and stored in tightly capped glass bottles with plastic tops.

Quantitative tests to determine whether contamination of the rock powders occurred during crushing and grinding were not carried out, but it is believed that the following precautions reduced the risk of contamination to a minimum.

Cross contamination from previously ground samples was demonstrated to be a major source of contamination by Thompson and Banksten (1970). Adopting their procedures, the steel vessel and pucks of the shatterbox were first scrubbed under hot, running water and then dried in acetone. Crushed pegmatite quartz was then ground for 3 minutes, followed by washing as before. A small aliquot of the rock sample was then ground in the shatterbox, brushed out and the vessel and pucks blown clean by compressed air, before grinding the remaining rock sample. This procedure was repeated for every sample ground.

A similar procedure was carried out with the automatic agate grinder, the grinder being blown clean of dust by compressed air before and after each run.

Nylon sieves were used to further reduce the risk of contamination.

## II. Analytical Procedure

### 1. Silica

Determined by the classical gravimetric technique (Volborth, 1969) on rock samples 13, 16, 171, 38, 48, 49, 80, 32M, 10M and 10E, and by fast neutron activation analysis on rock samples 5, 8, 13, 1411, 23, 30, 34, 40, 43, 57, 60, 77, 91, 32C and 32E. The method of fast neutron activation analysis of silicon is similar to that described by Volborth and Vincent (1967) for oxygen, except that the 1.78 meV peak of aluminium-28 was counted, irradiation and counting times were 30 seconds and 60 seconds respectively, and corrections made for phosphorus.

## 2. Manganese, Titanium, Calcium, Potassium

Determined by X-ray fluorescence on an ARL model 2500 X-ray Quantometer. Analyses were performed in duplicate utilising the combined fusion, dilution and heavy-absorber addition sample-preparation technique described by Rose, Adler and Flanagan (1963). The resulting fusion beads were ground and homogenised in a Buehler mill for 4 minutes and the powders pressed into discs on a boric acid backing at 70,000 psi for 30 seconds. Standards were prepared in exactly the same manner. Linear calibration curves constructed from these standards enabled the concentration of a particular element to be read directly.

## 3. Phosphorus, Magnesium, Sodium

Decomposition Procedure - to a 0.100 gram sample (approx. 150 mesh) in a teflon dish add 1 ml. concentrated  $\text{HNO}_3$ . Evaporate to dryness on a medium temperature sand bath (about  $250^\circ\text{C}$ ). Moisten with a few drops of water. Add 5 ml. concentrated  $\text{HCl}$ , and 5 ml.  $\text{HF}$  (48%) and break up residue with Teflon stirring rod. Evaporate to dryness. Moisten with a few drops of water. Add 5 ml.  $\text{HCl}$ , break up residue as before and again evaporate to dryness. Add 6 drops of concentrated  $\text{HCl}$  to the residue, break it up with the rod, and add 20 ml. water; heat until solution is complete, adding more water if necessary. Transfer the solution to a 100 ml. volumetric flask and dilute to volume - this is called the original solution.

### A. Phosphorus

Determined colorometrically on a 10 ml. aliquot of the original solution as the heteropoly blue complex, by the method of Boltz and Lueck (1958).



Absorbance was measured at 830 m $\mu$  on a Beckman Model B spectrophotometer in a 1 cm. cell using a reagent blank as the reference blank solution.

B. Magnesium

Determined on a 10 ml. aliquot of the original solution by atomic absorption, on a Techtron AA-3 atomic absorption spectrophotometer, using an air-acetylene flame. To the 10 ml. aliquot 10 ml. of 15,000 ppm. strontium solution was added and the volume made up to 100 ml. The resulting solution was then aspirated directly into the spectrophotometer and read using a reagent blank and high and low bracketing standards.

C. Sodium

Determined on the remaining original solution by flame photometry, on a Perkin-Elmer Model 146 flame photometer. To the remaining original solution, 4 ml. of radiation buffer was added, and the volume made up to 100 ml. The resulting solution was then read on the flame photometer using a reagent blank and high and low bracketing standards.

4. Alumina, and Total Iron as Fe<sub>2</sub>O<sub>3</sub>

Determined colorometrically on a Beckman Model B spectrophotometer from Solution B by the methods recommended by Riley (1958).

5. Ferrous Iron

Determined volumetrically using the method recommended by Volborth

(1969), utilising the inverted funnel into which CO<sub>2</sub> gas is led to prevent oxidation by the air. Analyses performed in duplicate.

6. Carbon Dioxide

Determined by the acid-attack train method of Volborth (1969).

7. Total Water

Determined by the Penfield tube method described by Volborth (1969), using anhydrous sodium tungstate as the flux.

8. Minus Water

Determined by the loss in weight of a previously weighed rock powder sample after drying in an oven for 2 hours at 105-110°C.

9. Plus Water

Obtained by subtraction of minus water from total water.

Using the same solution B on which Alumina and total Iron were determined colorimetrically (procedure 4), Ca, Mg, Na, K, Al and total Iron (as Fe<sub>2</sub>O<sub>3</sub>) were also determined by atomic absorption on a Perkin-Elmer Model 303 spectrophotometer at Memorial University through the courtesy of Dr. David Strong of the Geology Department. A nitrous oxide-acetylene flame was used in the determination of Aluminium, and an air-acetylene flame for the other elements. Thus duplicate analyses are available for the following elements: Mg, Ca, Na and K; and 2 analyses from the same original Solution B are available for Al and total Iron (as Fe<sub>2</sub>O<sub>3</sub>).

Table 9 lists the absolute differences between the duplicate analyses for  $\text{Na}_2\text{O}$ ,  $\text{K}_2\text{O}$ ,  $\text{CaO}$ ,  $\text{MgO}$  and  $\text{FeO}$ , and also the differences between the two analyses from the same Solution B for  $\text{Al}_2\text{O}_3$  and total Iron (as  $\text{Fe}_2\text{O}_3$ ). The average of the two sets of results was taken in every case, except those marked by an asterisk, where differences were considered too great. In these cases, when the total of all oxides fell outside acceptable limits by using the average value of the oxide in question, the oxide value which gave an acceptable total was chosen.

TABLE 9

## Absolute Differences Between Chemical Determinations

Rock No.	Na <sub>2</sub> O	K <sub>2</sub> O	CaO	MgO	FeO	Total Fe as Fe <sub>2</sub> O <sub>3</sub>	Al <sub>2</sub> O <sub>3</sub>
5	.1	.02	.1	.1	0	.4*	0
8	.4	.02	.2	.1	.2	.1	.2
13	.1	.03	.4	0	.2	.2	.1
1411	.1	.04	.3	.1	.1	.2	0
16	.2	.01	.4	.2	.1	.2	.3
171	.1	.03	.4	.1	.2	.4	.9*
23	0	.06	.1	0	.1	.5*	.3
30	.1	.02	.1	.1	.2	.3	.1
34	.1	.07	.1	.2	.1	.3	.1
38	.2	.02	.3	0	.1	.5	.2
40	.2	.02	.2	.1	.2	.2	.1
43	.1	0	.6	0	.2	.1	0
47+							
48	.1	.05	.3	.1	.2	.5	.3
57	.3	.04	.1	0	.1	.1	.9*
60	.2	.13	.1	.1	.2	.3	.1
69	.2	.08	.1	.3	.2	.3	.4
77	0	0	.2	.1	0	0	1.3*
80	.2	.07	.1	0	.1	.2	.2
91	.1	.01	.4	.2	.1	.4*	.2
32C	.2	.01	.3	.2	.2	.3	.7*
32M	0	.07	.5	.1	.1	.6*	1.3*
32E	.2	.05	0	.3	.1	.3	.4
10C	.2	.01	.4	.6	.2	0	.1
10M	.2	.04	.4	.1	.1	0	0
10E	.1	.04	.2	0	.1	0	.1

+ Analyst: H. Baadsgaard, Geology Department  
University of Alberta

\* Average value not used. For explanation, see text, Appendix B.

APPENDIX C

FIELD RELATIONS AND COLLECTION LOCALITIES OF ANALYSED ROCKS

For each rock, refer to Fig. 27 for the map location of the sample.

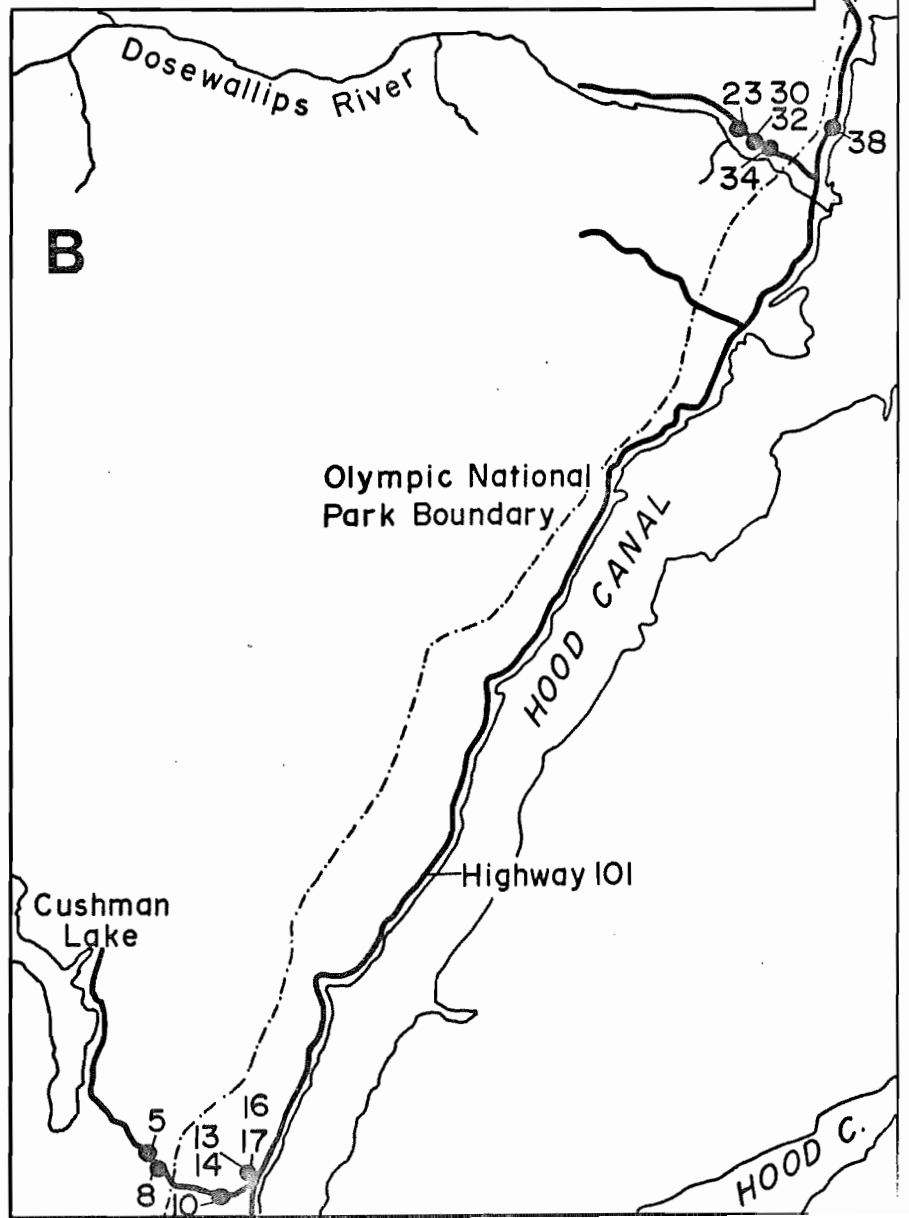
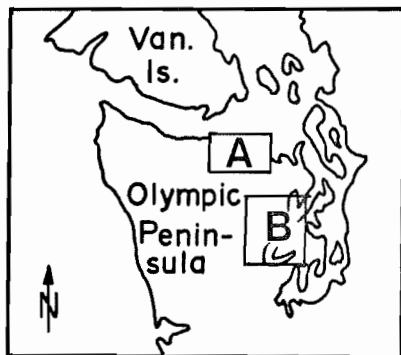
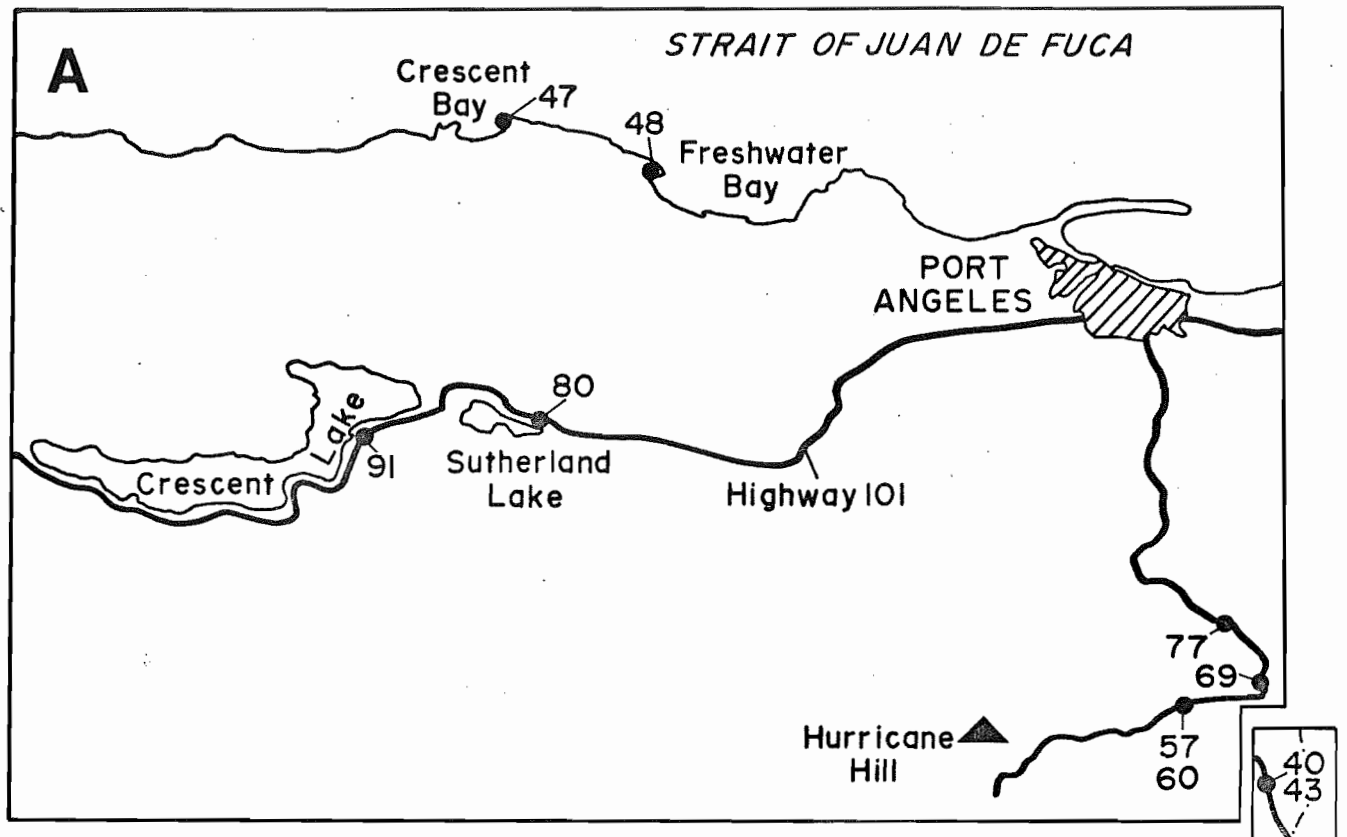
Rock No.	Field Relations and Collection Localities
5.	Extrusive, 50 feet from upper sediment contact. Cushman Lake Road, 0.5 miles west of National Park boundary.
8.	Probable intrusive. Sample from centre of sill. Cushman Lake Road, 0.4 miles west of National Park boundary.
10.	Extrusive, pillow basalt. Cushman Lake Road, 1.3 miles east of National Park boundary.
13 and 14	Intrusives (?) from massive unit of volcanics. Cushman Lake Road, 2.2 miles east of National Park boundary.
16 and 17	Intrusives (?), both from massive volcanic unit. Cushman Lake Road, 2.4 miles east of National Park boundary.
23.	Extrusive, from cliff section of thick sequence of extrusive volcanics. Dosewallips River Road, 1.5 miles west of National Park boundary.

- 30 and Extrusives, from thick section of basalts and pillow basalts.
32. Dosewallips River Road, 1.0 miles west of National Park boundary.
34. Extrusive. Dosewallips River Road, 0.7 miles west of National Park boundary.
38. Extrusive. 1.1 miles north of intersection of Dosewallips River Road with Highway 101.
- 40 and Extrusives from exposure of volcanics. 4.7 miles north of  
43. intersection of Dosewallips River Road with Highway 101.
47. Extrusive from headland composed of pillow lavas, lavas and hyaloclastite material. Tongue Point, east side of Crescent Bay, approximately 10 miles west of Port Angeles.
48. Hyaloclastite, from exposure of pillow lavas and hyaloclastite. West side of Freshwater Bay, west of Port Angeles.
57. Extrusive from exposure of volcanics. 4.7 miles from the southern end of Hurricane Hill Road, south of Port Angeles.
60. Intrusive (?) from exposure of volcanics. 4.7 miles from the southern end of Hurricane Hill Road, south of Port Angeles.

69. Intrusive (?), from massive unit of volcanics. 6.7 miles from the southern end of Hurricane Hill Road, south of Port Angeles.
  
77. Intrusive (?), from thick sequence of volcanics. 8.0 miles from the southern end of Hurricane Hill Road, south of Port Angeles.
  
80. Intrusive in sequence of volcanics. 0.2 miles west of the eastern end of Lake Sutherland, west of Port Angeles.
  
91. Extrusive from large sequence of pillow lavas and lavas. 4.5 miles west of the eastern end of Lake Sutherland, on the shore of Crescent Lake, near Sledgehammer Point.

Fig. 27      Locations of analysed rocks





**FIG. 27**  
**SAMPLE LOCATIONS**

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