# THE ORIGIN OF GEORGIA DEPRESSION ;

# AND THE COAST PLUTONIC COMPLEX/

# INSULAR BELT PROVINCE

# BOUNDARY ON HARDWICKE AND

# WEST THURLOW ISLANDS, B.C.

by

JoAnne Lee Nelson

B.Sc., University of Washington 1973

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF

THE REQUIREMENTS FOR THE DEGREE OF

MASTER OF SCIENCE

in the Department

of

GEOLOGICAL SCIENCES

We accept this thesis as conforming to the required standard:

THE UNIVERSITY OF BRITISH COLUMBIA

July 1976

In presenting this thesis in partial fulfilment of the requirements for an advanced degree at the University of British Columbia, I agree that the Library shall make it freely available for reference and study. I further agree that permission for extensive copying of this thesis for scholarly purposes may be granted by the Head of my Department or by his representatives. It is understood that copying or publication of this thesis for financial gain shall not be allowed without my written permission.

Department of Geology

The University of British Columbia
2075 Wesbrook Place
Vancouver, Canada
V6T 1W5

Date 16 Aug 1976

## ABSTRACT

Georgia Depression is a major northwest-trending structural trough that lies between the Coast Mountains and Vancouver Island, along the western boundary of the Coast Plutonic Complex. Field work on a part of this boundary on Hardwicke and West Thurlow Islands in Johnstone Strait established it as an intrusive contact between an Upper Jurassic quartz diorite and stratified rocks of the Vancouver Group. The Vancouver Group in the vicinity of the contact is broken by two major west-northwest trending preplutonic faults. Contact metamorphic conditions determined from basic and carbonate assemblages are $P_{total} \leq 3$ kb. and $T_{maximum} = 650 - 720^{o}C$. Post-plutonic fractures and dikes trend northeast. The main result of the field study was to confirm for a particular area that the western margin of the Coast Plutonic Complex is essentially a magmatic front.

Georgia Depression began to subside in the Upper Cretaceous as the Coast Plutonic Complex rose. It is proposed that the initial subsidence of Georgia Depression was complementary to the uplift of the Coast Mountains. Georgia Depression is part of the Coastal Trough of western North America, an elongate depression that lies between a series of colinear magmatic arcs (the Coast Plutonic Complex, the Cascades, and the Sierra Nevada) and the western margin of the North American plate. The various sub-basins of the Coastal Trough formed at different times, from the Upper Jurassic to the Late Tertiary, that correspond to the times of uplift of the adjacent igneous complexes.

This coincidence suggests that the subsidence of the whole Coastal Trough, including Georgia Depression, was coupled to the uplift of the magmatic arcs of North America. The Coastal Trough may be analogous to the marginal synclines developed in Ramberg's (1967) diapir experiments.

# TABLE OF CONTENTS

# LIST OF TABLES

## LIST OF FIGURES

# LIST OF PLATES

## ACKNOWLEDGEMENTS

H. R. Wynne-Edwards suggested the project to me and provided moral and financial* support throughout. Stan Reamsbottom mentioned Hardwicke Island as an interesting exposure of the province boundary. Jenny Stark, Fang Bendickson and my parents, Anne and George Nelson, gave able and uncomplaining assistance in the field. Jenny and I owe our comfort (and perhaps survival) to Bruce, Edie and Harold Bendickson and Mac and Hug Carmichael of Hardwicke Island, fairy godparents indeed. Lee Pigage, H.R. Wynne-Edwards and Richard Armstrong visited Hardwicke Island and provided useful insights about field relations. Conversations with Jan Muller, Glenn Woodsworth, W.R. Danner, Frank Daneš, John Griffiths, Bill Coulbourn, Ian Duncan, and Lee Pigage contributed to the evolution of my ideas.

Joe Harakal, Richard Armstrong, Peter LeCouteur and Cathy Lamb carried out and/or coached me in analytical procedures for radiometric dating. Jan Muller and Glen Woodsworth improved the manuscript with their suggestions. Jeannette Williams did the final typing with efficiency and aplomb. I would like to thank the geology technicians, R.B. Cooper, and John Payne for kind assistance in the pursuit of my goals.

* National Research Council operating grant A8302.

## SECTION 1: INTRODUCTION AND REGIONAL SYNTHESIS

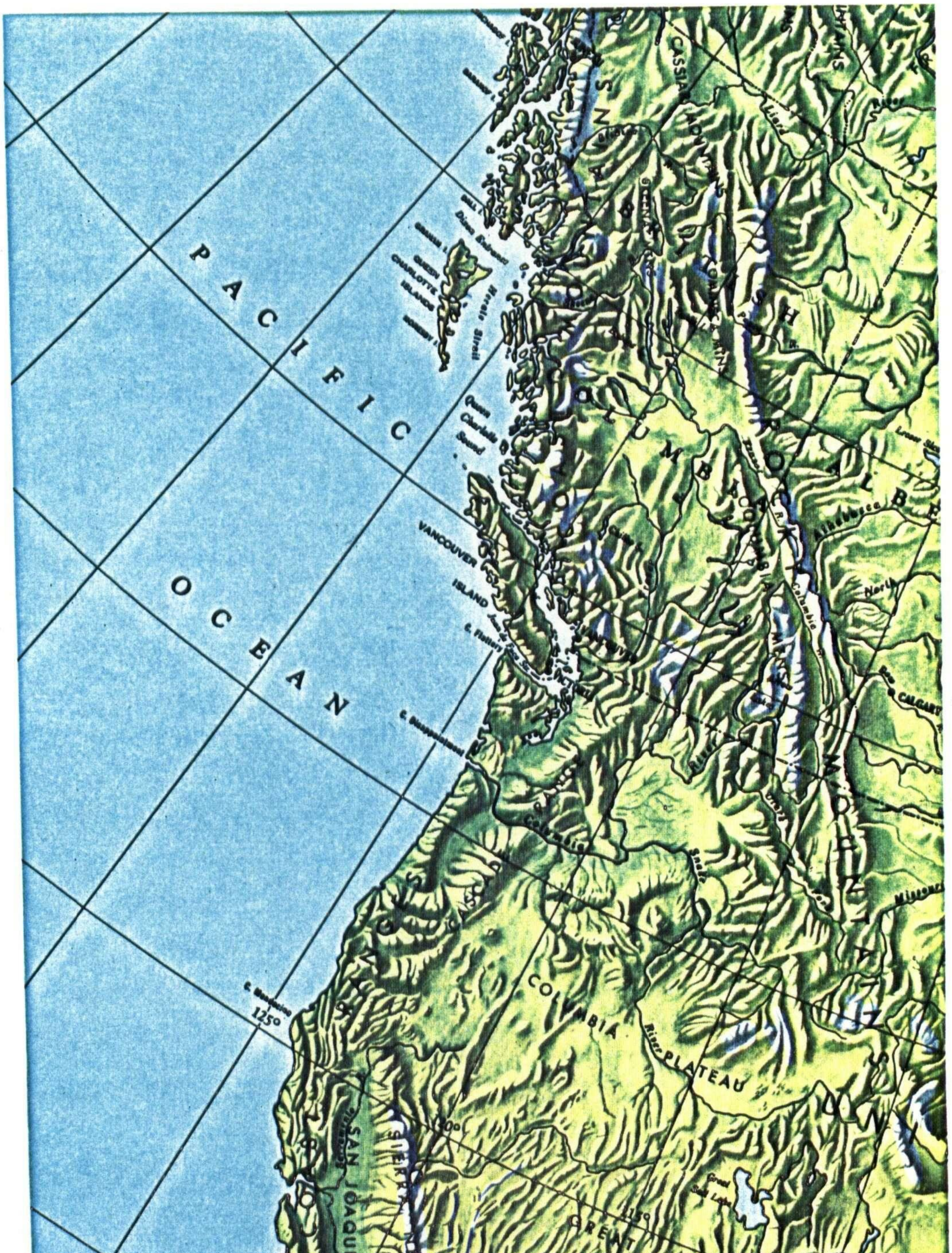
### Introduction

Georgia Depression is a topographic and structural low that lies between the mountainous regions of Vancouver Island and the Coast Mountains of British Columbia. It is a part of the Coastal Trough that extends from Hecate Strait in the north, through the Puget and Williamette Lowlands to the Central Valley of California (Figure 1). The Coastal Trough has been described in geomorphologic literature (Fenneman 1931; Holland 1964) but no attempt has been made so far to account for its origin. In contrast to previous studies which have emphasized local causes, this investigation, although principally concerned with the origin of Georgia Depression, has followed the hypothesis that since the Coastal Trough is a continuous feature then a common mode of origin should be sought.

Certain local features set Georgia Depression apart from the trough further south. It is deep: Tiffin(1969) estimates that half of the floor of Georgia Strait lies below continental shelf depth, 132 meters. It lies between the Coast Plutonic Complex (which in large part coincides with the physiographic Coast Mountains) and the Insular Belt (which includes Vancouver Island and the Queen Charlotte Islands). Several hypotheses for the origin of Georgia Depression have been proposed. They are:

1. Upper Cretaceous to early Tertiary downwarping and/or downfaulting (Jeletzky 1965; Sutherland Brown 1966; Muller and Jeletzky 1970).

Fig. 1


PACIFIC
OCEAN
QUEEN CHARLOTTE ISLANDS
Hecate Strait
Queen Charlotte Sound
VANCOUVER ISLAND
CASSIAR MOUNTAINS
BRITISH COLUMBIA
ROCKY MOUNTAINS
Columbia
Snake
Missouri
CALGARY
COLUMBIA PLATEAU
125°
120°
115°
110°
SAN JOAQUIN VALLEY
SIERRA NEVADA
Great Salt Lake
GREAT BASIN

2. Eastward tilting of the Vancouver Island block (Muller 1975).
3. Foundering of a zone of lithospheric weakness over a suture joining the two provinces. Monger et al. (1972) suggest that the Insular Belt is allochthonous, although they favor a suture along the Yalakom fault system.
4. Northeast-southwest extension associated with a Cretaceous or Cenozoic rift zone or rhombochasm (Bostrom 1968; Wynne-Edwards pers. comm. 1973).
5. Subsidence complementary to and accompanying the uplift of the Coast Mountains (this study).

These hypotheses are addressed to three issues: the timing, style, and ultimate causes of the Depression. In order to evaluate them this study utilizes both published literature and the results of a field investigation of the Insular Belt/Coast Plutonic Complex boundary carried out in the summer of 1974. It begins with a summary of present knowledge of the history of Georgia Depression and the two tectonic provinces that it spans. The field project provides further detail on the nature of the province boundary. The picture presented does not conform easily to either a suture or a rift; the position and timing of the Depression suggest that it formed as a complement to the rising Coast Mountains. Literature surveys of the rest of the trough show that the other segments likewise subsided synchronously with the uplift of the adjacent range. Possible mechanisms of coupling are discussed in the final chapter.

# CHAPTER I: THE HISTORY OF GEORGIA DEPRESSION AND VICINITY

The most obvious feature of the geologic setting of Georgia Depression is that it lies across the Coast Plutonic Complex/ Insular Belt province boundary. The history of these two provinces, so far as they are known, are summarized below. The origin of their present boundary is of particular interest. For clarity, the two provinces will first be discussed separately. (The Time-Space Plot, Plate 1, and the Fraser River 1 : 1,000,000 Map-Sheet (Roddick, Muller and Okulitch 1973) should be consulted during the following discussion).

## i. The Insular Belt

The oldest stratigraphic unit in Vancouver Island is the Sicker Group which contains strata from earliest Devonian or older (Muller 1976) to Pennsylvanian-Permian in age. Muller (1975) has divided the Sicker Group into three units. The lowest consists of andesite and dacite pyroclastics and subordinate flows. A turbidite sequence, restricted in occurrence, overlies the Sicker volcanics. The uppermost unit is a limestone containing fusilinids dated as Pennsylvanian or Permian. The species present are of Australian, not North American, affinity (Danner pers. comm. 1973). If this affinity has paleogeographic rather than paleoenvironmental significance the southern Insular Belt must be allochthonous. The uneven distribution of limestone and clastic sediments in Sicker Group exposures, as well as the abundance of andesite, has led Monger et al (1972) and Muller (1974) to suggest that it is the remains of a Paleozoic island arc.

The Sicker Group was metamorphosed to greenschist facies, and in places developed penetrative cleavage, before the Middle Triassic, as undeformed argillites of that age overlie it unconformably. The argillites are succeeded by the Karmutsen Formation of Upper Karnian age. The Karmutsen Formation is mainly a pile of basalt remarkable for its thickness (up to 8000 m.) and short eruption interval (part of a faunal zone). It embodies an emergent sequence (Carlisle and Suzuki 1974) ranging upward from pillow basalt to pillow breccia and breccia to amygdaloidal flows with sparse discontinuous sedimentary intercalations. Geochemically the Karmutsen basalts appear to resemble oceanic tholeiites or ocean island basalts more closely than island arc tholeiites (Kuniyoshi 1972). Kuniyoshi postulated that they were extruded at a mid-ocean ridge. Contact relations between the Karmutsen and the underlying Sicker Group are clearly depositional and not related to a thrust plane (Muller pers. comm. 1975), so Kuniyoshi's interpretation must be modified accordingly. Muller et al (1974) suggest that the basalts were formed at a ridge in an interarc basin. Symons (1971b) and Irving and Yole (1972) have determined paleomagnetic poles from the Karmutsen basalts. The poles lie about $50^{\circ}$ south of the mean Upper Triassic pole for North America. In order to reconcile the poles Vancouver Island would have had to lie in the southern hemisphere during the Upper Karnian. This is compatible with the Australian affinity of Pennsylvanian-Permian fusilinids in the Sicker Group.

The thin lensoid sedimentary intercalations in the Karmutsen Formation are probably lagoonal deposits. A typical lens contains an upward sequence of benthic limestone to pelagic limey shale that Carlisle and Suzuki (1974) interpret as submergent. Gentle subsidence, then, accompanied accumulation of the Karmutsen volcanic pile. The shallow-water rather than abyssal environment lessens the likelihood that the Karmutsen was formed at an oceanic ridge. It also precludes direct analogies with modern back-arc basins, which are deep (a necessary consequence of their thin crust).

The Karmutsen basalts are succeeded by upper Karnian to Pliensbachian sediments, which are divided into three formations with gradational contacts (also interpreted by Carlisle and Suzuki as a subergent sequence); the upper Karnian Quatsino Formation (limestone); the Norian Parson Bay Formation, a series of thin-bedded cherts, argillites, and carbonates; and the Lower Jurassic Harbledown Formation, distinguished from the Parson Bay Formation by absence of carbonate and presence of greywacke.

Volcanic clastic sediments in the Harbledown Formation mark the resurgence of volcanism. The Harbledown sediments are succeeded by and are in part coeval with the Sinemurian to Lower Pliensbachian Bonanza Volcanics. Dominantly andesitic, they are thought to represent a calc-alkaline island arc (Monger et al. 1972; Muller et al. 1974). Most of the plutons, and all of the large ones, on Vancouver Island are of Jurassic age (148 to 181 m.y.: Muller et al. 1974). The Jurassic Island Intrusions occupy about ten percent of the island's surface area and are roughly coeval with the Bonanza Volcanics. Mutual intrusive relationships have been observed: andesitic dikes in the plutons; and plutons intruding and metamorphosing Bonanza andesites. Northcote and Muller (1972) suggest that they are comagmatic. Confirmation awaits trace-element and isotopic comparisons and dating of possible Bonanza dikes in the plutons.

Symons (1971a) has determined a paleomagnetic pole for the Island Intrusions which, within the limits of error, coincides with the Middle Jurassic North American pole. The southern Insular Belt must have been in place by that time. The date of possible suturing is thus constrained between upper Karnian (approximately 210 m.y.) and 166 m.y., the oldest intrustion sampled by Symons.

Except for a few Lower Cretaceous bodies on Texada Island and the northeast shoulder of Vancouver Island, the youngest dated Island Intrustion is 148 m.y. old. Significant igneous activity in the southern Insular Belt ceased at that time. It did not resume until the Tertiary, and then in subdued form. This lack of Cretaceous igneous activity is one of the main differences between this province and the Coast Plutonic Complex, which contains an enormous volume of Cretaceous-Eocene intrusive material. During and after emplacement of the Island Intrusions the area now occupied by Vancouver Island and Georgia Strait was uplifted enough to shed appreciable clastic debris, including granitic material, westward onto the continental shelf (Jeletzky 1965; Muller et al. 1974). Additional evidence for regional uplift above wavebase includes: 1) the absence of Pliensbachian to Santonian strata on most of Vancouver Island and 2) the unconformity below the Upper Cretaceous Nanaimo Group which cuts across all older units and displays over 300 meters of relief (Ross and Barnes, unpub. ms.). Upper Jurassic and Valanginian to Barremian clastic marine sediments of the Longarm Formation accumulated on the northern and western edges of the uplift. A nearshore to offshore environment is inferred from the Longarm fauna. Conglomerates in the Longarm Formation contain few if any granitic clasts.

On northwestern Vancouver Island the Longarm Formation is overlain by the Aptian to Cenomanian Queen Charlotte Group. As in the type area in the Queen Charlotte Islands (Sutherland Brown 1968) it is partly non-marine and contains coarse granitic debris, which may reflect more vigourous uplift of the Insular Belt (Muller et al. 1974).

The Pacific Rim Complex, partly coeval with the Longarm Formation, occurs in fault-bounded blocks along the western margin of Vancouver Island. It consists mostly of highly deformed turbidites with subordinate chert, basalt and limestone. Page (1974) has described olistostromes in the Pacific Rim Complex. Both Page and Muller (1973) term it a melange. It may be a Franciscan equivalent, or at least an analogue, in which case the Late Jurassic - Early Cretaceous trench lay somewhat east of the present plate boundary.

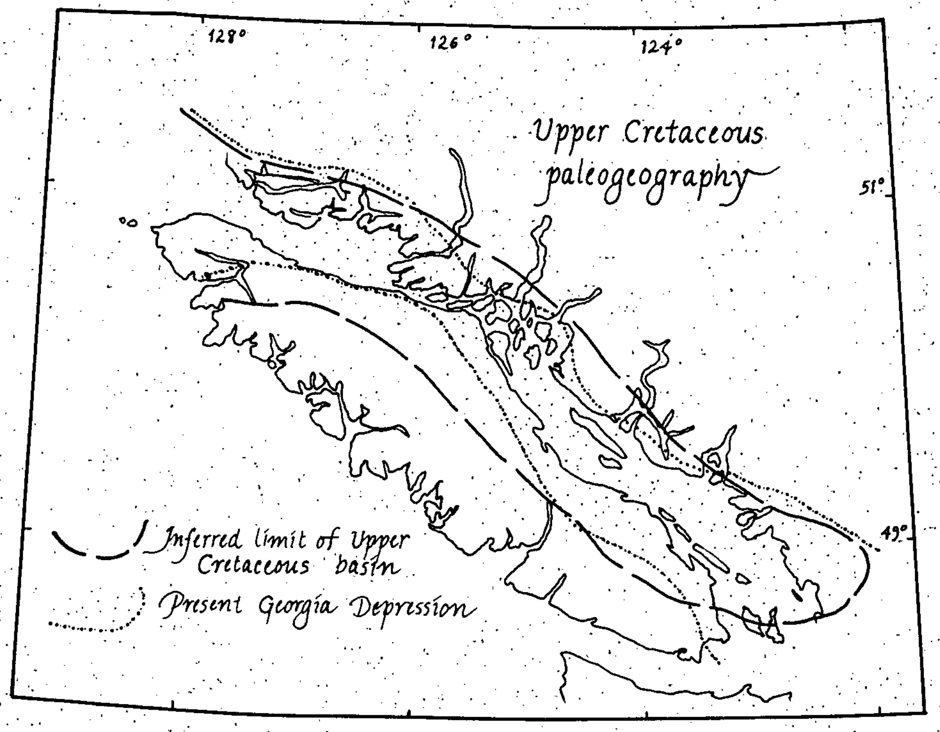
The Santonian to Maestrichtian Nanaimo Group outcrops in and around Georgia Strait. Its inferred original distribution (Figure 2) implies a basin configuration very similar to the modern Georgia Depression. The Nanaimo Group is a complex intertonguing sequence of greywacke-siltstone rhythmites, conglomerate lenses, shale, massive greywacke, shale and coal. It is currently interpreted as a complex delta/submarine fan in a subsiding basin (Ross and Barnes pers. comm. 1973; Rinne 1973; Simmons 1973; Hudson 1974; Sturdavant 1974). The source of sediment for the Nanaimo Group was the Insular Belt uplift: paleocurrents are easterly to northerly and cobbles in conglomerates suggest a Vancouver Island provenance. The Upper Cretaceous basin was the direct precursor of Georgia Depression, superimposed on a post-plutonic regional uplift of the southern Insular Belt. Its place in the history of the Coast Plutonic Complex will next be shown.

ii. The Coast Plutonic Complex

The Coast Mountains are dominated by plutonic and polydeformed metamorphic rocks (see Fraser River map-sheet, Roddick, Muller and Okulitch 1973) that have nearly obliterated the stratigraphic record.

The existence of Precambrian basement in the southern Coast Mountains is still unproven. The Custer Gneiss north of the Fraser River (Daly 1912) is considered correlative with the Skagit Gneiss in the North Cascades (Misch 1966) based on lithologic similarity. The Skagit Gneiss has given zircon ages

Fig. 2




In part after Muller and Jeletzky (1970).
note: fauna in Helm-Empetrum Fms. reidentified as Lower Cretaceous, thus not Nanaimo equivalent (Jeletzky to N. Green, 1974).

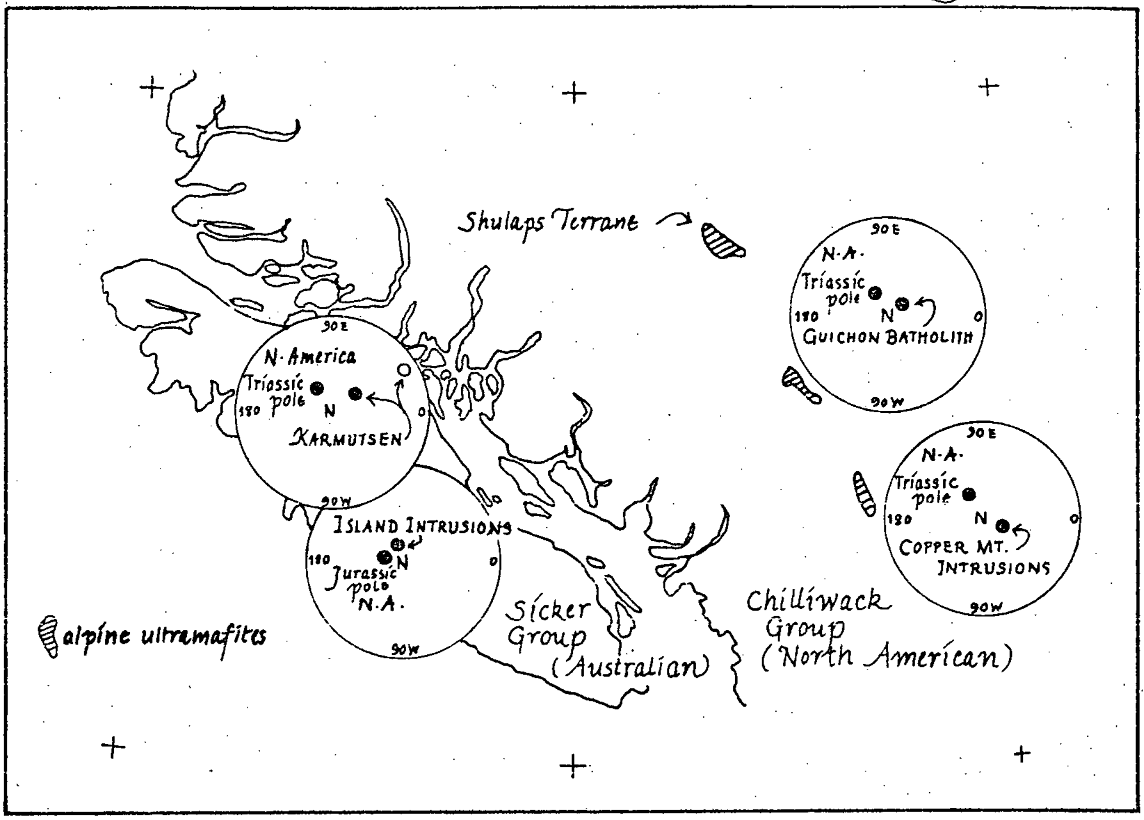
greater than 1650 m.y. (Mattinson 1973). Even assuming that the Custer Gneiss is autochthonous and not transported from the south by faulting, the extent of basement in the Coast Mountains cannot be guessed at.

Upper Paleozoic (Lower Pennsylvanian to Lower Permian) strata in the southern Coast Mountains are confined to partly fault-bounded septa near Harrison Lake. They are correlative with the Chilliwack Group in the North Cascades (Monger 1970). Lithologically the Chilliwack Group resembles the Sicker Group, but its stratigraphic order is different. It includes volcanic sandstone, pelite, chert, siltstone, limestone, metabasalt and/or meta-andesite, and volcaniclastics. As fusilinids in the Chilliwack limestones are of North American affinity (Monger and Ross 1971), the Sicker and Chilliwack Groups may have been deposited thousands of kilometers apart. If so a suture zone lies between them. Does it follow Georgia Depression? The Triassic record sheds some light on this question.

The Karmutsen Formation is truncated abruptly on the east by voluminous intrusions. There are no certain Karmutsen equivalents within the Coast Plutonic Complex, although a few have been suggested. Pillow basalts in Jervis Inlet bear a striking resemblance to the Karmutsen (Woodsworth pers. comm. 1974). Tipper (pers. comm.) also reports likely analogues. These may argue against a post-Triassic suture between the two provinces. Triassic strata in the southern Coast Mountains between Pemberton and Lillooet are included in the Fergusson and Cadwallader Groups, The Fergusson Group is dominated by chert, argillite and basalt with minor limestone. Part of it has been dated as Ladinian (Cameron and Monger 1971), coeval with the Middle Triassic argillites on Vancouver Island. Monger et al. (1972) interpret it as an oceanic assemblage. The Hozomeen Group, lying between the Yale and Hozameen Faults, has yielded no fossils, but is correlated on lithologic grounds with the Fergusson Group (Cameron and Monger 1971). The Cadwallader Group has yielded fossils of Norian age (Roddick and Hutchison 1973). It contains greenstone, clastic sediments including volcanic conglomerate with quartz monzonite boulders, argillite, pyroclastics, chert and limestone. Unlike the Karmutsen Formation all Triassic units within the Coast Mountains are highly deformed.

Figure 3 shows fusiline affinities and paleomagnetic poles in southern British Columbia to demonstrate possible positions for an Upper Triassic - Lower Jurassic suture. Of the possible locations, the Insular Belt/Coast Plutonic Complex boundary seems much less likely than the Shulaps Terrane, which fits well into the preconception of suture. In it are cherts - "ocean bottom sediments", greywackes - "trench sediments", volcanics, and ultramafites, all highly deformed and faulted. The Shulaps ultramafic body overlies Fergusson Group rocks along a shallowly east-dipping contact, possibly a shear zone (Nagel pers. comm. 1975). The ultramafics are thought to have been emplaced in Upper Triassic to Lower Jurassic time, as they are faulted against Middle Triassic and Norian rocks and detrital chromite appears in late Lower

Fig.3




*Paleomagnetic poles and faunal affinities southwestern British Columbia*

pole positions from:
Symons 1971a, 1971b, 1971c, 1973
Irving and Yole 1972

Jurassic sediments (Leech 1953). Although shear zones are described in the eastern Insular Belt (Kuniyoshi 1972; Muller pers. comm. 1975; Carlisle pers. comm. 1975) they are not of the same scale, nor are ultramafites present along them.

The most complete Jurassic-Cretaceous record pertaining to the Coast Plutonic Complex is preserved in a narrow fault-bounded band (including the Pasayten and Methow Grabens) south and east of the Coast Mountains. The sequence, ranging in age from Sinemurian to Cenomanian, consists of greywacke-siltstone rhythmites, conglomerate and phyllite. A few volcanic and volcanic clastic units occur east of Mt. Waddington (Tipper 1969). Until recently these sediments were thought to have been deposited in a fairly restricted trough, the Tyaughton Trough. However, several detailed facies and paleocurrent studies (Cole 1973; Coates 1974; Tennyson 1974) have shown that the "trough" was open to the west until Aptian time. Current directions were from the east or longitudinal, as was sediment transport deduced from facies changes. The paleoslope was westerly: there is no indication of a proto-Coast Mountains either as a positive topographic feature or as a sediment source until Albian time.

Jurassic-Lower Cretaceous strata within the Coast Mountains include 1) the Gambier Group and correlative strata, 2) the Fire Lake Group, and 3) the Harrison Lake Group. The sequence at Harrison Lake is Middle Jurassic to Lower Cretaceous, largely coeval with the Pasayten Graben sequence. The source for the marine clastic sediments, including granitic cobbles, at Harrison

Lake is undetermined. In contrast to the Pasayten strata the section is broken by angular unconformities and contains abundant andesite and pyroclastics. The Fire Lake Group north of Harrison Lake includes Upper Jurassic/Lower Cretaceous volcanics and marine sediments, including granitic debris (Roddick 1965). The Gambier Group and its equivalents outcrop intermittently for 150 km. along the southwestern side of the Coast Mountains. They rest unconformably on plutonic rocks (Roddick 1965; Caron 1974) but are intruded and metamorphosed by later plutons. Albian ammonites have been found in the Gambier Group (Jeletzky pers. comm. to Mathews 1972): these are the youngest known marine sediments in the western Coast Mountains. The sparse Jurassic and Lower Cretaceous sediments within the Coast Mountains, mostly marine, suggest that much of the Coast Plutonic Complex lay at or below sea level during that time. Sufficient erosion occurred to unroof plutons: but some of the granitic debris may have been derived from Insular Belt and Intermontane highlands. Certainly no extensive plutonic/metamorphic terrane was exposed. The earliest westerly-derived conglomerates in the Methow and Pasayten Grabens contain mainly chert and volcanic cobbles (Tennyson 1974). These are Aptian-Albian in age and provide a maximum age for the southern Coast Mountains as a high-standing range.

The Jurassic plutonic history of the Coast Mountains is undefined. Two Jurassic dates have been obtained so far in the southern part of the range (Mathews 1968; McKillop 1973). Granitic cobbles in Upper Jurassic-Lower Cretaceous conglomerates indicate early plutonism but nothing can be said yet about its intensity or timing. Was the Coast Plutonic Complex a locus of intense magmatism then as in the Cretaceous, or was its Jurassic intrusive history similar to that of the Insular Belt? This problem may be resolved by Rb/Sr and U/Pb dating.

In the Early Cretaceous the Coast Plutonic Complex became the site of a narrow zone of intense plutonism. It remained so until Eocene time. Potassium-argon dates in the southern Coast mountains generally decrease in age eastward from Early Cretaceous along the western margin to Eocene in the east. The ultimate origin of the plutonic rocks is debatable. Two sources have been proposed: generation within a subducted lithospheric plate, and crustal anatexis, possibly involving remelting of Benioff-zone-derived rocks (Griffiths pers. comm.). Crustal derivation is favored by Geological Survey workers who have mapped the northern Coast Mountains (Hutchison 1970; Roddick and Hutchison 1972). They have offered convincing arguments that plutonic bodies in the Prince Rupert area can be traced to origins in the Central Gneiss Complex. A gradational sequence extends from migmatitic, highly foliated autochthonous bodies, concordant with surrounding migmatites, to unfoliated, discordant plutons intruding low-grade rocks. Some, termed "para-autochthonous plutons", embody the entire sequence. Roddick and Hutchison envisage parts of the migmatitic terrane becoming mobilized and moving by plastic flow into the overlying rocks. This model may or may not apply to the southern Coast Mountains.

An important consequence of plate-tectonic theory is that calc-alkaline magmas may be derived by partial melting along Benioff zones. By analogy with the Andean arc, Hamilton (1969) reasons that long arcuate volcanic/plutonic/metamorphic terranes such as the Sierra Nevada are the deeply eroded remnants of volcanic arcs. Useful guidelines for recognizing fossil magmatic arcs include:

1. elongate arcuate shape and/or parallelism to an inferred plate boundary (such as a continental margin); reasonable proximity to the plate boundary, say 100 to 400 km.
2. predominance of andesites and their intrusive equivalents, granodiorite or quartz diorite.
3. high T/P metamorphism
4. a coeval low T/P metamorphic zone towards the plate boundary (Miyashiro 1961; Ernst 1971).
5. younging of intrusions away from the plate boundary.
6. increase in $K_2O$ likewise.
7. low Sr 87/86 and Pb 206/204 ratios particularily in volcanics, as an index of subcrustal origin.

The Cretaceous/Eocene Coast Plutonic Complex conforms to 1), 2), 3), 5), and 6); the Pacific Rim Complex as a possible Franciscan correlative suggests that it conforms to 4), although glaucophane and lawsonite have not as yet been found in it. Strontium isotopic ratios in the Coast Plutonic Complex tend to be low. However, they are not easily interpreted (see below).

The Coast Plutonic Complex as a magmatic arc is not incompatible with the derivation of its plutonic bodies by crustal melting. Greenwood (1975) favors a two-stage history for intrusive rocks: primary origin at a Benioff zone and remelting within the crust. This model is favoured by Presnall and Bateman (1973) for the Sierra Nevada.

Strontium isotopes are not useful for distinguishing between subcrustal and crustal origin for intrusions in the Coast Mountains. Although as yet inadequately sampled, they tend to be rudidium poor; their Sr 87/86 ratios increase very slowly through time (Culbert 1972; and this study). Magmas derived from them would have essentially the same Sr-isotopic ratios as Benioff zone melts.

Pre-Cretaceous metamorphic events have been deduced by various authors (Tipper 1969; Roddick and Hutchison 1972; Reamsbottom 1973). Like pre-Cretaceous plutonism, assessments of degree, extent or timing of these events cannot be made. Extensive regional metamorphism accompanied the Cretaceous-Eocene plutonism. The pattern of Cretaceous metamorphism is complex and not well known due to a lack of detailed pressure-temperature studies. Although presumably accompanying plutonism it is in many cases unrelated to specific plutons (Hollister 1969; Reamsbottom 1973; Woodsworth 1974). Metamorphic grade seems to increase towards the eastern Coast Mountains, possibly reflecting a deeper erosion level. The kyanite-sillimanite transition is observed in several places in the eastern Coast Mountains (Hollister 1969; Lowes 1972; Pigage 1973). This reaction requires a minimum of 3.8 kb pressure (Holdaway 1971). Woodsworth (pers. comm.) reports that there may be a pattern of metamorphic highs and lows in the southern part of the range.

Despite the sketchiness and obvious intricacy of the Cretaceous story three fairly safe and significant generalizations can be made:

1. The gross characteristics of the Coast Plutonic Complex - its preponderance of plutonic rock, its prevalent moderate to high grade of metamorphism, its narrowness and abrupt boundaries, and the great depth of erosion attained in places within it - were acquired or (in the case of erosion) initiated during the Cretaceous.
2. The present boundaries of the Coast Plutonic Complex are either intrusive contacts or faults (Baer 1973; Roddick and Hutchison 1974). The western contact is a line along which Insular Belt strata are truncated by intrusions mostly dated as Early Cretaceous. The western edge of the province thus was established in Lower Cretaceous time.
3. The creation of the Coast Plutonic Complex was linked to subduction.

The amount of uplift is variable: some septa are in lower greenschist facies; some have been metamorphosed at pressures above the aluminosilicate triple point. Generally the erosion level depens eastward, but this is complicated by warping, faulting and smaller-scale metamorphic culminations. Evidence for timing of uplift is sparse. Westerly-derived Aptian-Albian sediments in the Pasayten Graben indicate that the eastern Coast Mountains began to rise in the Middle Cretaceous. Much of the uplift was accomplished along the Fraser River Fault System. Movement on the Hozameen Fault, which juxtaposes Middle Triassic Hozameen Group and Jura-Cretaceous sediments, ceased before 84 m.y. ago, the date of a small pluton that cuts across it and is not displaced (Coates 1974). The Yale Fault has brought the Custer Gneiss in contact with Hozameen rocks. Tilted Eocene conglomerate along it attests to Tertiary activity. The Chilliwack Batholith, 26 m.y. old, (Fraser River map-sheet) cuts across its southern extension, the Straight Creek Fault.

It is suggested that the eastern Coast Mountains began to rise in Middle Cretaceous time and were the western source that Coates(1974) infers for the Pasayten/Jackass Mountain Formations. The western part of the range could not have been a high-standing feature until at least the Late Cretaceous, as there are Albian marine sediments in the Gambier Group. Futhermore, had the western Coast Mountains existed in Santonian time, they surely would have contributed sediment to the proto-Georgia Depression then forming immediately to the west. There is no evidence that they did. The Paleocene (?) Chuckanut Formation has a dominantly eastern source: the Cascade and Coast Mountains must have been elevated by that time.

The Lions Gate Member of Campanian age (Rouse et al 1975), the oldest Nanaimo correlative on the east side of Georgia Strait, rests unconformably on plutonic rocks in Capilano Canyon. It contains less than 10% granitic clasts (less even than the Gambier Group) and virtually no high-grade metamorphic or foliated plutonic clasts, reflecting a mostly volcanic source terrane without large amounts of exposed plutonic/crystalline rocks as are seen in the southern Coast Mountains today. This shallow erosion level as late as Upper Cretaceous is another indication that strong regional uplift had not occurred before that time.

In summary, the province boundary that underlies Georgia Depression is a (roughly) Early Cretaceous intrusive contact. The Depression itself, however, began to form in the Santonian: it well postdates the establishment of this boundary. What evidence there is suggests that it accompanied another significant event in Coast Plutonic Complex history, the Upper Cretaceous-Eocene uplift of the range.

iii Georgia Depression

The Upper Cretaceous structural depression was not a graben *sensu stricto*, as both downfaulting and downwarping were involved. The prevalence of faults on geologic maps is misleading: there are several generations of faults in the Depression, some of which have nothing to do with its origin. The massive volcanic units of Vancouver Island have tended to undergo brittle failure under compression, yielding a map pattern of fault blocks and open folds. Major faults on the east side of Georgia Strait have been proposed (Hopkins 1966; Berry et al. 1971) but not confirmed. It is hoped that future studies will separate the different stress systems.

The present Georgia Depression is a modified version of the original basin. Its Cenozoic history is difficult to reconstruct because, except for a few outcrops near Vancouver, no Tertiary rocks are exposed above sea level. The sedimentological record of the uplift of the Coast Mountains is likewise scanty (Eisabacher, pers. comm. 1975). Nanaimo Group deposition apparently ceased in the Maestrichtian. It may have continued into the Paleocene (Muller and Jeletzký 1970); it may have evolved in the south into continental-deltaic sedimentation of the Chuckanut Formation (P.Ward, pers. comm. 1974). Both the Chuckanut Formation and the Nanaimo Group were involved in pre-Middle Eocene deformation. Folds and faults in the Chuckanut Formation are overlain unconformably by the Middle Eocene Huntington Formation (Misch 1966). On Saltspring Island the Nanaimo Group is folded about northwesterly axes. The major folds verge westerly. Eastern limbs of the anticlines are generally steep; some are overturned. Faulting was more important than folding on Vancouver Island where a thin cover of Nanaimo sediments overlies volcanic/plutonic basement (Ross and Barnes,Ms. unpub.). Blocks bounded by steep faults dip gently east towards Georgia Strait (Muller 1974). It was probably during the Early Tertiary episode that the Nanaimo Group rocks in Mt. Washington and the Forbidden Plateau were raised to their present elevation almost 2000 meters above sea level.

At least part of Georgia Depression remained a sedimentary basin throughout Tertiary time. The Tertiary record in Whatcom

Basin near Vancouver is fairly complete. Dating is mainly based on spores and pollen (Hopkins 1966; Rouse et al 1975). Deposition in the Whatcom Basin was not interrupted by the Eocene deformation. The Lions Gate Member passes without significant break into Eocene deltaic sediments of the Kitsilano Formation (Hopkins 1966; Blunden 1971). The Kitsilano Formation is succeeded by an unnamed unit of Miocene age seen only in drillholes (Hopkins 1966). Total thickness of the Tertiary section exceeds 3000 meters. The Tertiary depositional environment was terrestrial deltaic. Tiffin (1969),using seismic reflection profiles, traced Tertiary units part of the way across Georgia Strait. Major Tertiary faults under the strait are possible. Tiffin suggested that the Tertiary sequence is truncated by a major fault at the base of the eastern slope of Vancouver Island.

The southern part of Georgia Depression was affected by the same Miocene event that produced west-northwest folds in the Columbian Embayment. The San Juan Islands are the backbone of a Miocene upwarp (Snavely and Wagner 1963) that cuts obliquely across the southern end of Georgia Strait. Fold orientations in the Nanaimo Group change from northwest to west-northwest between the Gulf Islands and the San Juans; this may partly be due to the Miocene tectonic event.

Late Tertiary movements have accentuated Georgia Depression; it has remained low while the flanking ranges rose. The lack of marine sediments except in the latest Pleistocene, however, indicates that it probably stood above sea level throughout the

Tertiary. Late Cenozoic uplift of the Coast Mountains amounted to about two thousand meters. (Roddick 1966). Recognizable Miocene and Pliocene erosion surfaces penetrate deep into the range. Miocene volcanics lie on deeply beveled surfaces (Baer 1973 and the Fraser River map-sheet). Downward movement of the Depression was probably somewhat less than the uplift (Blunden 1971).

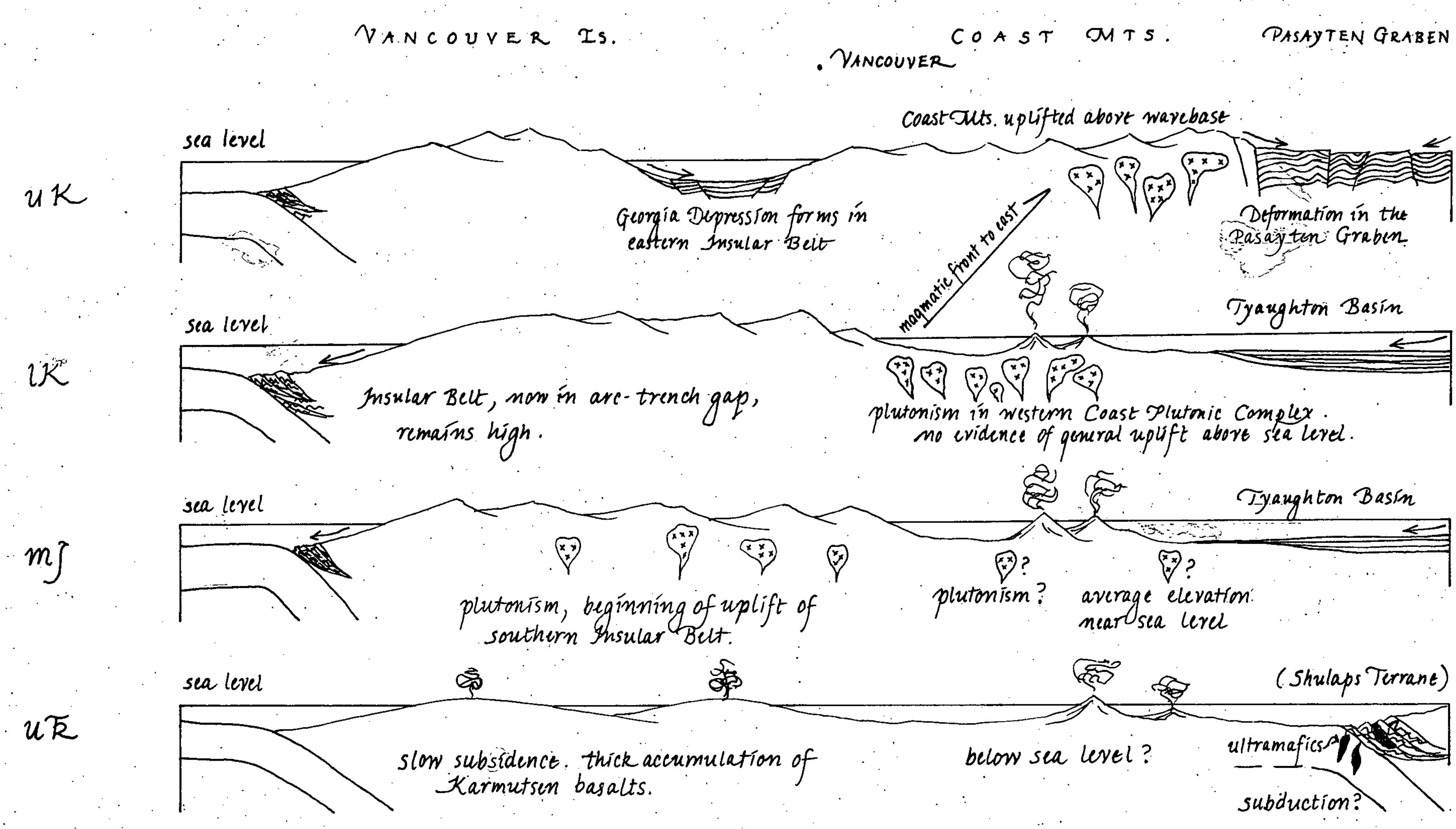
The present strait owes much of its depth and narrowness to glacial erosion, first because glaciers were channelled between the adjacent ranges and second because of differential erosion of soft Nanaimo Group sediments (Mathews 1968). The southeastern part of the Depression is currently subsiding. The pattern seems unrelated to the geometry of the Fraser Delta (Mathews et. al. 1970) and thus may be partly tectonic.

iv. Summary and Evaluation of Regional Synthesis

Figure 4 summarizes the tectonic and paleogeographic evolution of the Insular Belt and the Coast Plutonic Complex up to the origin of Georgia Depression. This historical view bears on the hypotheses listed on Page 1. The timing of origin of the Depression is certain. The Santonian-Maestrichtian Nanaimo Group sediments were deposited in a basin whose configuration mimics the present one. Therefore the Depression was created in the Upper Cretaceous: Cenozoic rifting, faulting and erosion can only have modified a pre-existing trough. The structural nature of the Depression is poorly known. It is referred to as a combination downwarp-downfault in regional studies, but definitive evidence as to the importance of fault control has yet to be

# Paleogeographic evolution of southwestern British Columbia

Fig 4



presented. Hopefully some of the studies now in progress will fill this gap.

The ultimate cause of the subsidence of Georgia Depression has been linked to a suture, a rift zone, or a complement to the uplift of the Coast Mountains. It is worthwhile before evaluating these hypotheses to enumerate the evidence that pertains to each.

Coupled uplifts and subsidences occur of course simultaneously, making field evidence as to timing clearly valuable. On the other hand, as no particular structural style is required, field evidence pertaining to structure is not diagnostic. Great faults such as the Teton Fault may be involved; or the movement may consist of gentle tilting.

The current method of identifying sutures and rift zones is by comparison with accepted examples. Examples of sutures which have received general acceptance are those in the Klamath Mountains (Irwin 1973; Cashman 1974); northern New Guinea (Dewey and Bird 1970); and Newfoundland (Williams et al. 1971). Dewey and Bird (1970) and Le Pichon et al (1973) treat suturing and suture zones in general. Characteristics of suture zones are:

1. juxtaposition of unrelated geologic provinces.
2. a line of ophiolites or alpine ultramafites, "dismembered ophiolites" marking the suture.
3. blueschist and/or melange terrane between the two provinces, indicating a fossil subduction zone.

4. intense deformation - isoclinal folding, thrusting, obduction.

Since there is no recognizable Upper Cretaceous oceanic crust in Georgia Depression, any rifting there must have involved a small amount of extension. There are several well-known modern continental rifts of this sort, the Baikal and Rhine Grabens (Beloussov 1969; Milanovsky 1972), and the East African rift system (Girdler et al 1969; Baker et al 1972). Important characteristics of continental rift zones include:

1. They are narrow linear complex grabens. Faults either dip steeply inwards or are vertical. Dominant faults parallel the sides of the graben; cross faults are also present.
2. Characteristic total vertical displacement is 2 to 5 km.
3. High heat flow and vulcanism (tholeiitic and alkalic) accompany rifting. High temperature metamorphic assemblages may provide evidence for high heat flow in deeply eroded fossil rift zones.
4. The graben will probably fill with sediment, producing a large negative gravity anomaly. Fault control should be evidenced by coarse, unsorted border facies.
5. In east Africa a regional negative Bouguer anomaly has been ascribed to low-density material in the upper mantle (Girdler et al 1969).

The evidence for a suture zone between the Insular Belt and the Coast Plutonic Complex is that Vancouver Island lay in the southern hemisphere in the Karnian but was in place by the Middle Jurassic (166 m.y.). Paleomagnetic data (Symons 1971b;

Irving and Yole 1972) requires the suture to lie between Vancouver Island and the southern Intermontane Belt: it need not coincide with the Insular Belt/Coast Plutonic Complex boundary. Monger et al.'s (1972) suggestion that the Upper Triassic - Lower Jurassic suture lies along the Yalakom Fault System is probably the correct one, as the Triassic rocks there are highly deformed and faulted and pierced by Upper Triassic-Lower Jurassic ultramafic bodies. The improbability of a suture between the Insular Belt and the Coast Plutonic Complex is further demonstrated by the field study described in the next section. A rift zone, if it engendered the Depression, must have been active in the Upper Cretaceous. Arguments based on structural style are futile at this point. There are no volcanics in the Nanaimo Group. Heat flow at present in the Georgia Strait area is low (Hyndman 1976). Past heat flow can only be guessed at from metamorphic assemblages. Metamorphic phases in the Nanaimo Group include heulandite, laumontite and unidentified phyllosilicates (Stewart and Page 1974). The absence of high-temperature zeolites such as wairakite might be taken as evidence against high heat flow (and consequently rifting) in the past.

At the time of its formation Georgia Depression lay between an evolving magmatic arc, the Coast Plutonic Complex, and its trench. Because it lay within an arc-trench gap, the Cretaceous depression was by definition a forearc basin (Dickinson 1974). It is important that this term be used without genetic significance. The forearc basins off the coast of Chile (Ross and Shor 1965; Coulbourn and Moberly 1975) lie immediately above the trench-slopebreak. They are bounded oceanward by a highly deformed and thrusted welt of sediments. Georgia Depression, by contrast, lies a hundred kilometers from the Pacific/North American plate boundary and is basement-controlled. Its location immediately west of the western limit of cretaceous plutonism and its coincidence with the major uplift of the Coast Plutonic Complex suggest that its subsidence was linked to the rise of the range, rather than to some subduction - related process.

## SECTION II: FIELD STUDY

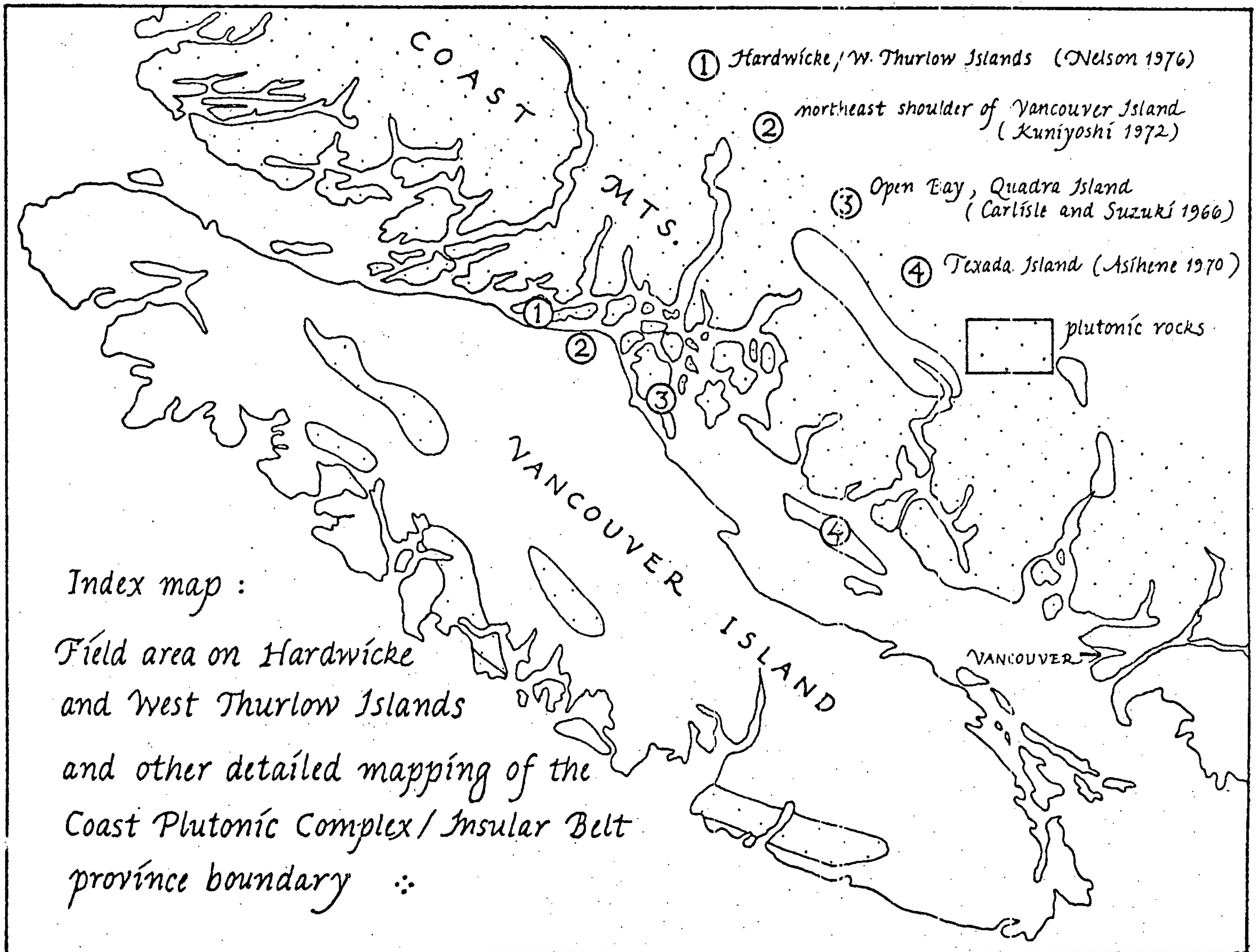
### Introduction - General Geology of the Map Area

The position of the Coastal Trough in British Columbia over the Coast Plutonic Complex/Insular Belt boundary ought to have genetic significance, whatever the cause of the depression. I undertook a field study of a section of the boundary in order to investigate its nature and possibly its role in engendering a structural trough. The map-area chosen for this purpose includes Hardwicke, West Thurlow and neighboring smaller islands, and the coast of Vancouver Island near Kelsey Bay (Figure 5). The islands were previously mapped at 4 miles to the inch during the ten-year reconnaissance of the Coast Mountains conducted by the Geological Survey of Canada (Roddick et al. 1971 and in preparation). Kuniyoshi (1972) and Muller (1972) mapped the Vancouver Island shore at ½ mile and 4 miles to the inch respectively.

The province boundary has been mapped elsewhere, although not with the object of studying the relationship of the two provinces. The work of Carlisle and Suzuki (1965) on Quadra Island, Asihene (1970) on Texada Island, and Kuniyoshi (1972) on the northeast shoulder of Vancouver Island, will be discussed in the summary chapter for comparison.

The boundary between the Insular Belt and the Coast Plutonic Complex exposed on Hardwicke and West Thurlow Islands is an intrusive contact (Geologic Map, Plate 2). To the north

Fig. 5




Index map :

Field area on Hardwicke and West Thurlow Islands and other detailed mapping of the Coast Plutonic Complex / Insular Belt province boundary ∴

and east of it are plutonic rocks, dominated by a Late Jurassic quartz diorite. Stratified units of the Mesozoic Vancouver Group lie to the south and west. The southwestern ends of Hardwicke and West Thurlow Islands and the northern coast of Vancouver Island are underlain by basalts of the Karmutsen Formation. The Quatsino Formation outcrops on the southern coast of West Thurlow Island. It passes upwards into the Parson Bay Formation, which also occurs as septa within plutonic rocks near Eden Point. The Harbledown Formation outcrops east of Carmichael Point on Hardwicke Island. It is succeeded inland by the Bonanza Volcanics. Two major west-northwest-trending faults, the Johnstone Strait and the Telegraph Hill faults, have broken the Vancouver Group into homoclinal blocks. Prior to intrusion the Vancouver Group underwent prehnite-pumpellyite burial metamorphism. Contact-metamorphic greenschist assemblages are developed in Vancouver Group rocks less than 4 kilometers from the quartz diorite, and hornblende hornfels assemblages less than 500 meters from it. Synplutonic ductile deformation affected the Vancouver Group within the inner aureole.

Northeast-trending dikes, pervasively altered to chlorite-epidote-albite, cut both the quartz diorite and the older rocks. They mark the close of igneous activity and are the last recorded event in the map area except for Pleistocene glaciation.

# CHAPTER II: STRATIGRAPHY OF THE VANCOUVER GROUP

Every formation in the early Mesozoic Vancouver Group is represented in the map-area (Figure 6, column 3; also see the Geologic Map, Plate 2), although no complete section is seen. Most units are bounded by faults, intrusive contacts, or water. Probable fault repetition and, near the intrusions, tectonic thinning, render thickness estimates approximate only. A reasonable sequence may nevertheless be reconstructed by taking account of lithology and structure.

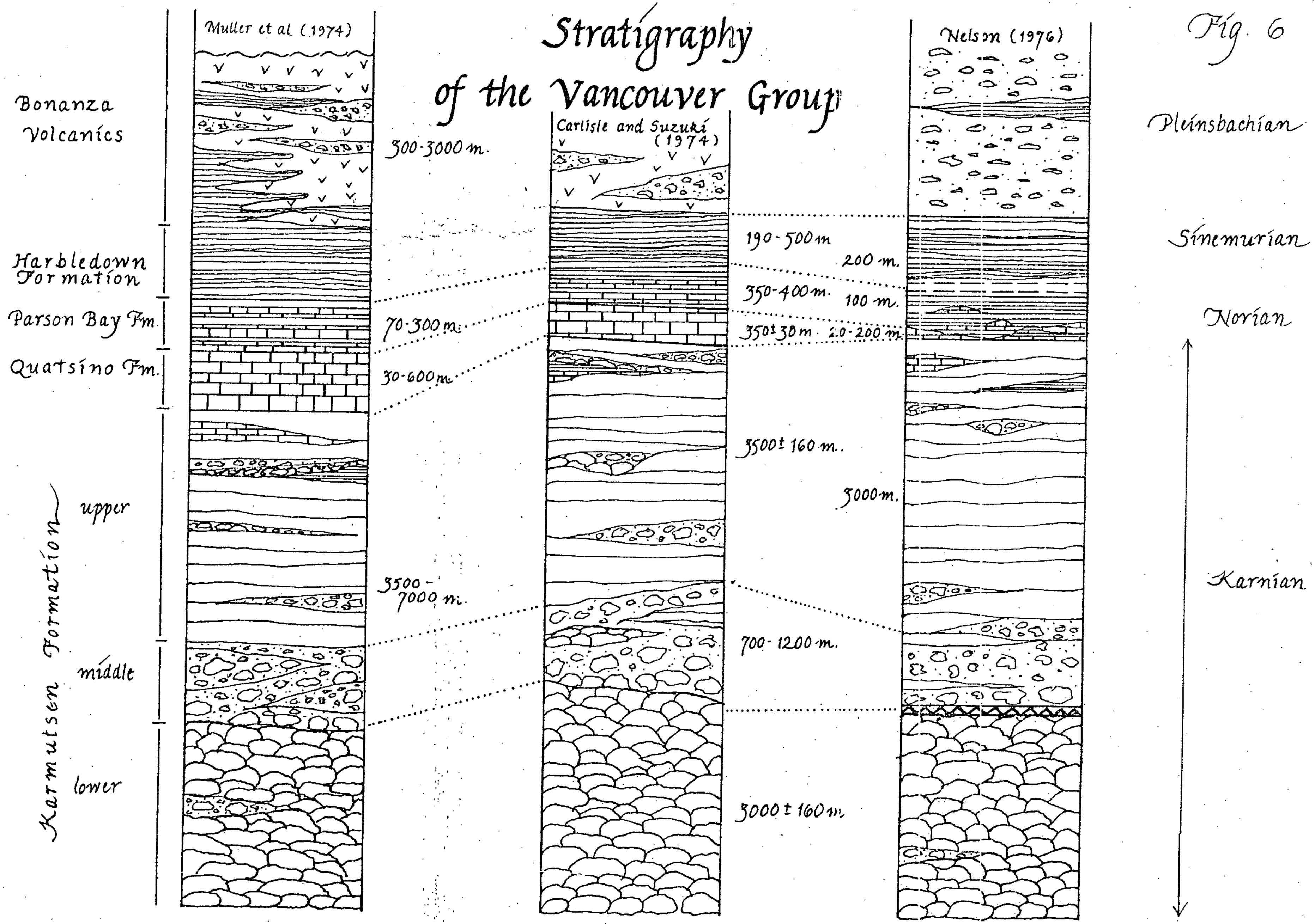
## i. Karmutsen Formation

The Karmutsen Formation is readily divided into three parts based on lithology. Kuniyoshi (1972) assigned formational names to them. However, to avoid ambiguity, Kuniyoshi's names will not be used here.

The lower Karmutsen Formation (Kuniyoshi's Kelsey Bay Formation) is well exposed as shore cliffs near Kelsey Bay. It is a monotonous pile of pillow basalt relieved by minor breccias and tongues and flows of massive basalt. The interiors of the pillows are hyalopilitic with small phenocrysts and amygdules (less than 1 mm. across). Their original glassy selvages are altered to fine-grained epidote-pumpellyite-chlorite mixtures. Pillow interstices are filled with quartz, calcite, epidote and clinozoisite. Only approximate bedding attitudes can be obtained from the pillows. These attitudes are somewhat erratic, perhaps reflecting initial sea-bottom topography.

Fig. 6

# Stratigraphy of the Vancouver Group

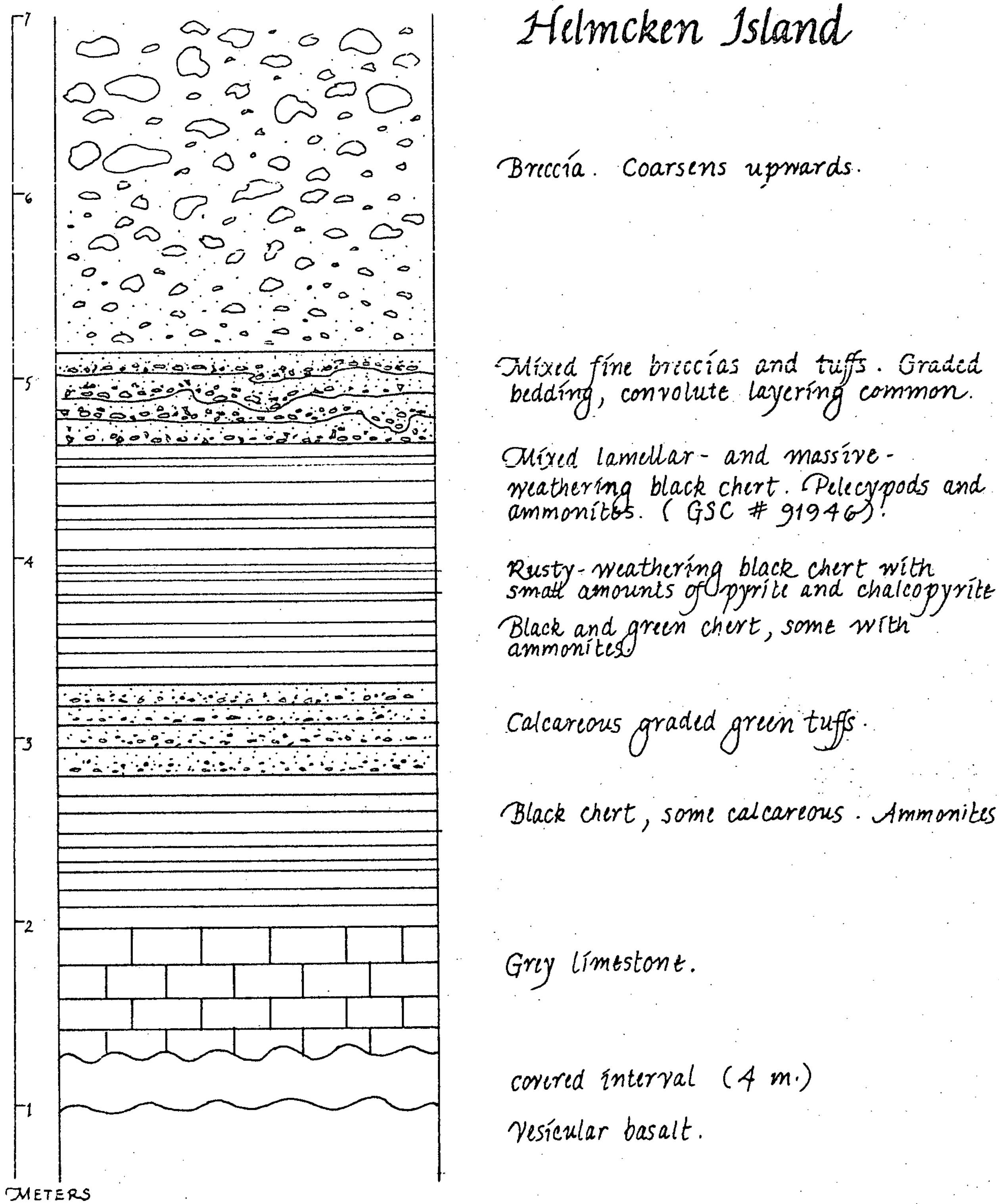


The middle Karmutsen Formation (Kuniyoshi's Hkusam Formation) is characterized by a predominance of pillow breccias with subsidiary flows, aquagene tuffs and true breccias. It is exposed in a homocline along the northeast shoulder of Vancouver Island, east of Camp Point. The basal contact is a fault (see Kuniyoshi 1972; Muller 1972). The contact with the upper Karmutsen Formation is a gradational zone in which amygdaloidal flows replace breccia upwards.

Several partial sections of the upper Karmutsen Formation (Kuniyoshi's Menzies Formation) are present in the northwesterly dipping homocline between the Johnstone Strait and Telegraph Hill Faults (see Chapter IV). All of the sections are predominantly composed of amygdaloidal flow basalt with varying precentages of breccia, pillow basalt, and interlava sediments. On the northeast shoulder of Vancouver island approximately 2000 meters of upper Karmutsen Formation overlie the middle Karmutsen. Pillow basalt, pillow breccia and breccia form about 10% of the section, while sediments are lacking. This assemblage is typical of the lower part of the upper Karmutsen. On Helmcken Island, by contrast, only one breccia-pillow breccia unit is present, and is associated with an interlava sedimentary cycle (Figure 7). The presence of sediments is typical of the top 1000 meters of the Karmutsen Formation (Kuniyoshi 1972). On southwestern Hardwicke Island the upper Karmutsen Formation consists entirely of flow basalt. Stratigraphically above it on Yorke Island are breccias, pillow basalts and interlava sediments. All of these

Fig 7

# Stratigraphy of intervolcanic sedimentary cycle, Helmcken Island



sections are in seeming structural continuity. If they are combined, however, the aggregate thickness, 10,000 meters, greatly exceeds all estimates for the upper Karmutsen Formation. A northeasterly crossfault is hypothesised between Helmcken Island and Bendickson Harbor. It divides the homocline into two packages of reasonable thickness, and places Helmcken Island within the uppermost Karmutsen Formation.

The upper Karmutsen Formation also occurs on southern West Thurlow Island where it is partly overlain by the Quatsino Formation, and partly truncated by intrusive rocks. The section contains five separate sedimentary intercalations and several thick pillow and/or pillow breccia units.

Flows in the upper Karmutsen Formation are typically coarsely amygdaloidal at their bases, sparsely amygdaloidal in their centers, and contain a densely amygdaloidal horizon near their tops. Pipe vesicles occur near flow bases. The amygdules tend to irregular amoeboid shapes. They are up to a centimeter in diameter, indicating deposition in water much shallower than that in which the lower Karmutsen Formation was deposited. The interlava sediment lenses within the upper Karmutsen Formation range from one to seven meters thick. They are never traceable for more than a hundred meters or so. Carlisle and Suzuki (1974) attributed their lack of continuity to deposition in small basins and lagoons. Three basic lithologies are present in varying amounts: limestone, thin-bedded carbonaceous chert, and fine-grained green greywacke interpreted by Carlisle and Suzuki as local basalt-weathering products. The interlava sediments on Helmcken Island follow a typical sequence (Figure 7). An ammonoid collected at this locality was examined by E.T. Tozer (see Appendix). He sets its age as Upper Triassic, a reassuring but unrevealing determination.

It has been observed before (Carlisle and Suzuki 1974) that these sedimentary intervals are scattered throughout the uppermost Karmutsen Formation. The only interval which seems to represent a consistent time horizon is the Intervolcanic Limestone (Muller et al. 1974) which lies near the top of the Karmutsen Formation. This limestone is seen in two places on West Thurlow Island, on Yorke Island, and possibly on Fanny Island. It consists of a series of lensoid bodies 3 to 7 meters thick. Lithologically it is identical to the Quatsino Formation.

ii. The Vancouver Group Sediments: Quatsino, Parson Bay and Harbledown Formations

The Quatsino Formation succeeds the upper Karmutsen Formation on southern West Thurlow Island. Large thickness variations suggest a lensoid shape. Above Miner's Bay approximately 150 meters of massive marble is truncated by the intrusion. East of Vansittart Point the Quatsino includes 5 meters of pure marble overlain by 15 meters of marble with chaotically folded chert interbeds. The lower contact is obscured by a post-plutonic dike. It may be a minor shear zone. The primary features of

Quatsino limestone described by other workers - stylotitic bedding, cross-bedding, algal structures - are absent due to metamorphic recrystallization. Near Miner's Bay the Quatsino Formation is a massive light grey marble with very sparse 2 to 5 centimeter cherty interlayers. East of Vansittart Point the percentage of chert layers increases upwards to pass into the Parson Bay Formation.

The lower boundary of the Parson Bay Formation is the point of which black limestone, shale, and siltstone predominate over pure limestone (Muller et al. 1974). This transition is observed east of Vansittart Point. Above it fine black siliceous siltstone predominates. Marble units in the Parson Bay Formation are lenticular or small pillow-like lenses generally less than a meter thick. In the upper part of the section there are a few thoroughly recrystallized greywackes which preserve primary features such as load casts, graded bedding, cross-bedding, and clastic dikes. Near Eden Point on western West Thurlow Island the Parson Bay Formation is 75% siltstone and 25% marble. The siltstone layers are black to yellow. Some, containing chalcopyrite and pyrite, weather to a rust color. One package of thin-bedded black siliceousits contains pelecypods identified as Monotis subcircularis Gabb of upper Norian age (E.T. Tozer pers. comm. see Appendix). This age agrees with that of the Parson Bay Formation elsewhere.

The sediments between Carmichael Point and Patterson Bay on Hardwicke Island yielded no fossils; however lithologically they resemble the Harbledown Formation which is distinguished from the Parson Bay Formation by its lack of limestone. The Hardwicke Island exposure consists of fine black siliceous siltstone, some of it slightly calcareous. It is overlain by volcanic breccia and massive siltstone of the Bonanze Volcanics.

iii. <u>Bonanza Volcanics</u>

The Bonanza Volcanics are most extensive inland on Hardwicke Island where exposures are poor. They also occur in septa east of Patterson Bay bounded to the north by quartz diorite and to the south by Johnstone Strait. Identification of the Bonanza is based on lithologic and mineralogic criteria. The interbedding of volcanic breccia and massive siltstone fits the Bonanza far better than the Karmutsen Formation. The breccias contain small flattened clasts 1 to 10 cm. in diameter. either primarily or tectonically flattened (or both) parallel to bedding. They are true breccias, not pillow breccias. Mineralogically they are andesites, not basalts. They contain abundant metamorphic quartz and biotite; plagioclase anorthite contents vary but in several cases are much less than in Karmutsen rocks at equivalent grade.

Figure 6 shows that the composite stratigraphic section in this map area closely resembles the Vancouver Group elsewhere on Vancouver Island, including the three subdivisions of the Karmutsen Formation, the lower part of the Bonanza Volcanics, and the intervening sedimentary formations. The regional history for Vancouver Island in the early Mesozoic outlined in Chapter I applies equally here. Although in this field study stratigraphy has mainly been used to define structure it also has interesting implications for depositional environments. Several workers have noted that the Vancouver Group sediments offer no evidence for strong currents, steep slopes, or nearby sediment sources (Asihene 1970; Kuniyoshi 1972; Carlisle and Suzuki 1974). This is even true in the easternmost exposures such as the present study area. An Upper Triassic-Lower Jurassic suturing event ought to generate a high-energy depositional environment. Yet the Upper Triassic-Lower Jurassic Vancouver Group shows no evidence for it. This argues against a suture near the province boundary.

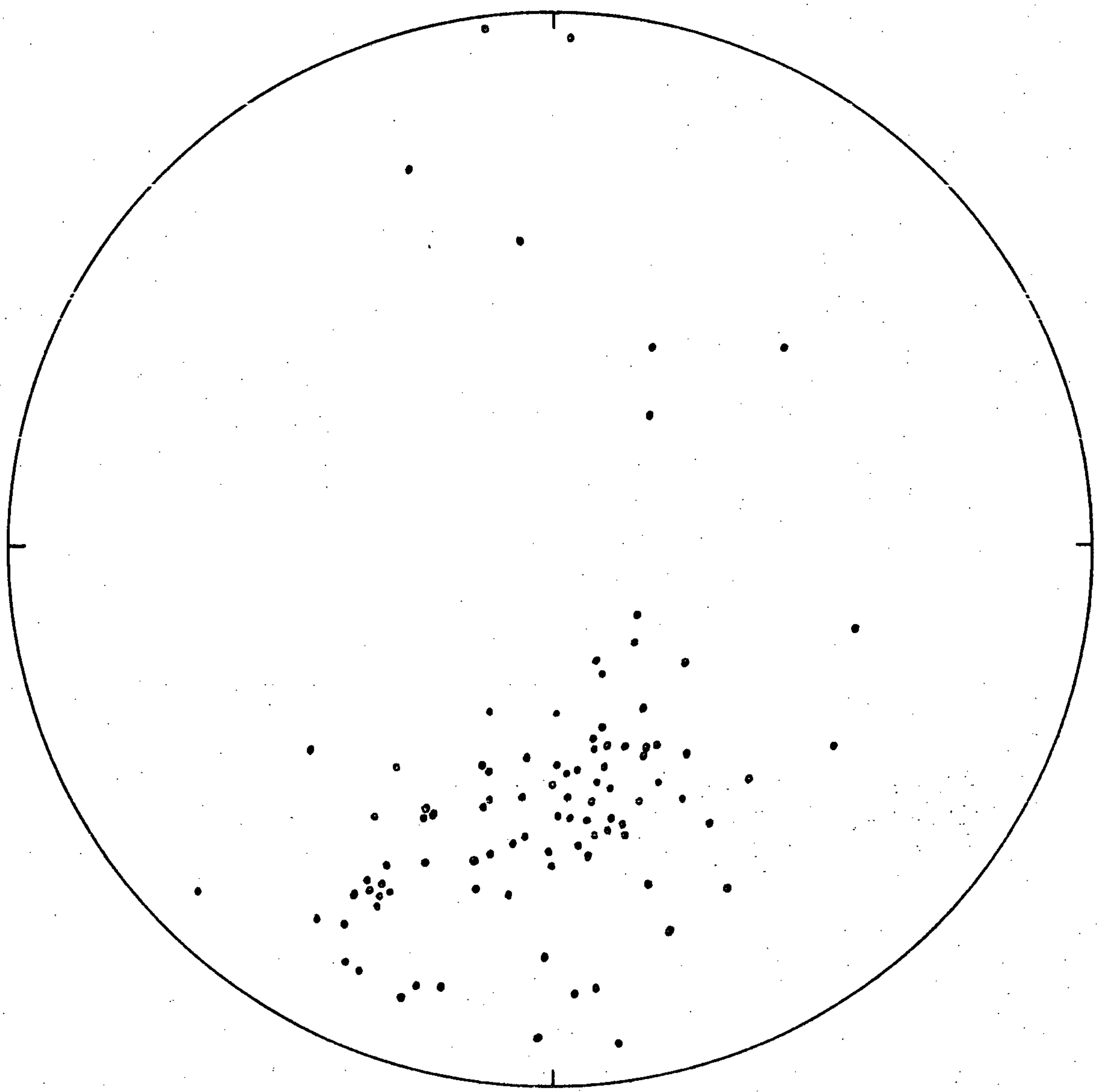
iv. Stratigraphy and Major Structure

Three homoclinal domains, based on bedding attitudes and stratigraphic successions, are inferred. Domain I includes West Thurlow Island and Hardwicke Island northeast of Telegraph Hill. The average bedding attitude in it is 85/40 N. (Figure 8). The lowest part of the section seen in this domain is the upper Karmutsen Formation on southern West Thurlow Island succeeded by the Quatsino Formation at Miner's Bay and east of Vansittart Point; and the Parson Bay Formation east of Vansittart Point and at Eden Point (where a tongue of the intrusion has obliterated the Quatsino limestone). In the same domain, and in proper stratigraphic order, are the Harbledown and Bonanza Formations on Hardwicke Island. Domain II is the narrow northwest - trending homocline extending from Yorke Island to the northeast shoulder of Vancouver Island. It is separated from Domain I by the

Fig. 8

# Poles to bedding, Domain I

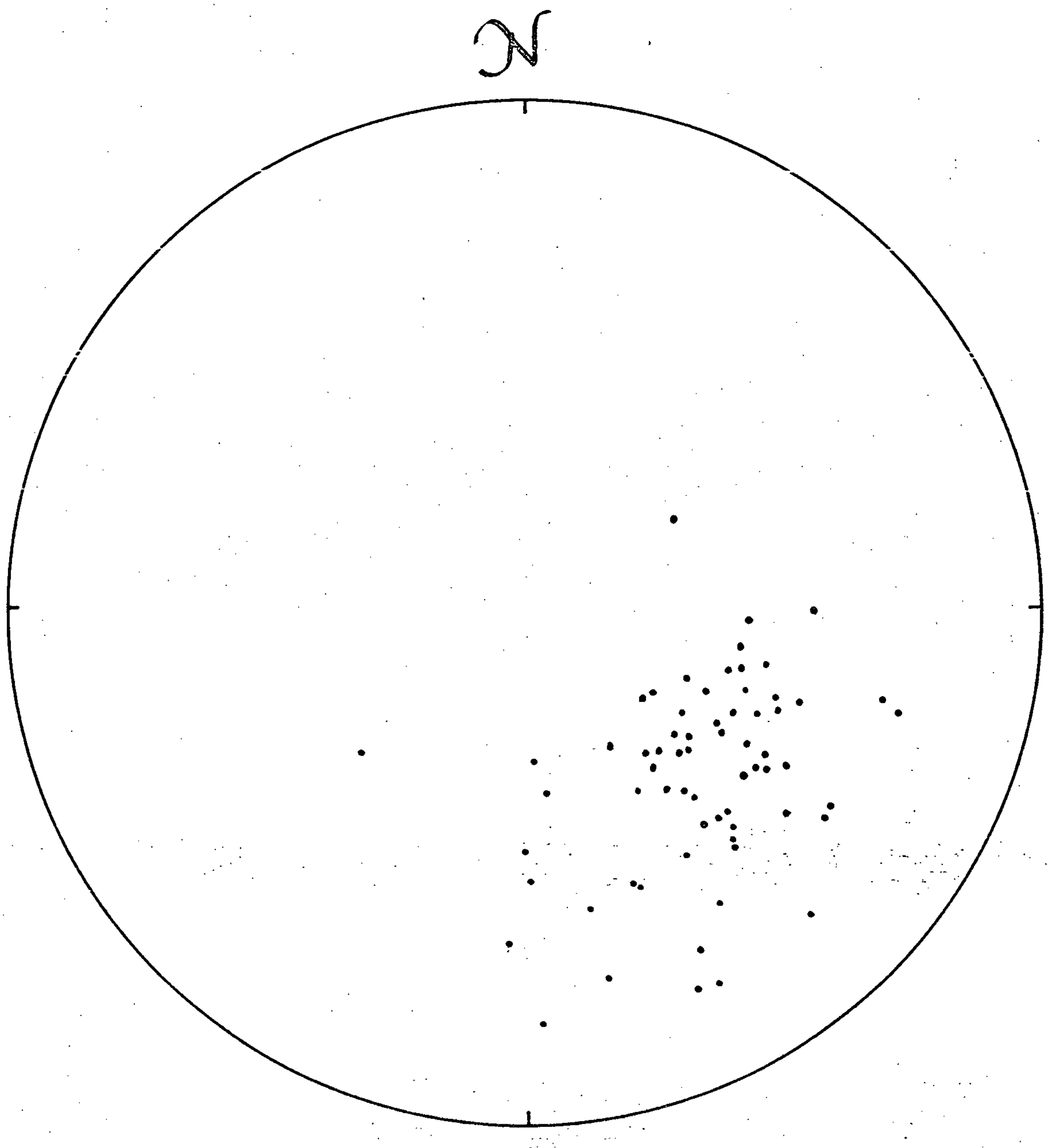
N



Telegraph Hill Fault. Average bedding in it is 44/32 N.W. (Figure 9). A cross-fault is inferred south of Bendickson Harbor to give reasonable thicknesses of the upper Karmutsen formation in Domain II. Domain III encompasses the Vancouver Island coast west of Camp Point, southwest of the Johnstone Strait Fault. It consists of near-horizontal lower Karmutsen Formation. The relationship between these domains and the major faults that separate them are discussed in Chapter IV.

Fig. 9

# Poles to bedding, Domain II



## CHAPTER III: PLUTONIC ROCKS

One prominent feature of the Insular Belt/Coast Plutonic Complex boundary is the virtual disappearance of the Vancouver Group east of it. In some instances the Insular Belt strata are sharply truncated by Coast Mountains intrusions; in others there seems to be a continuum between metavolcanics and plutonic rocks (Roddick and Hutchison, 1974,and Carlisle, pers. comm. 1975). This chapter documents the nature of the plutonic bodies on Hardwicke and West Thurlow Islands and their relation to the Vancouver Group.

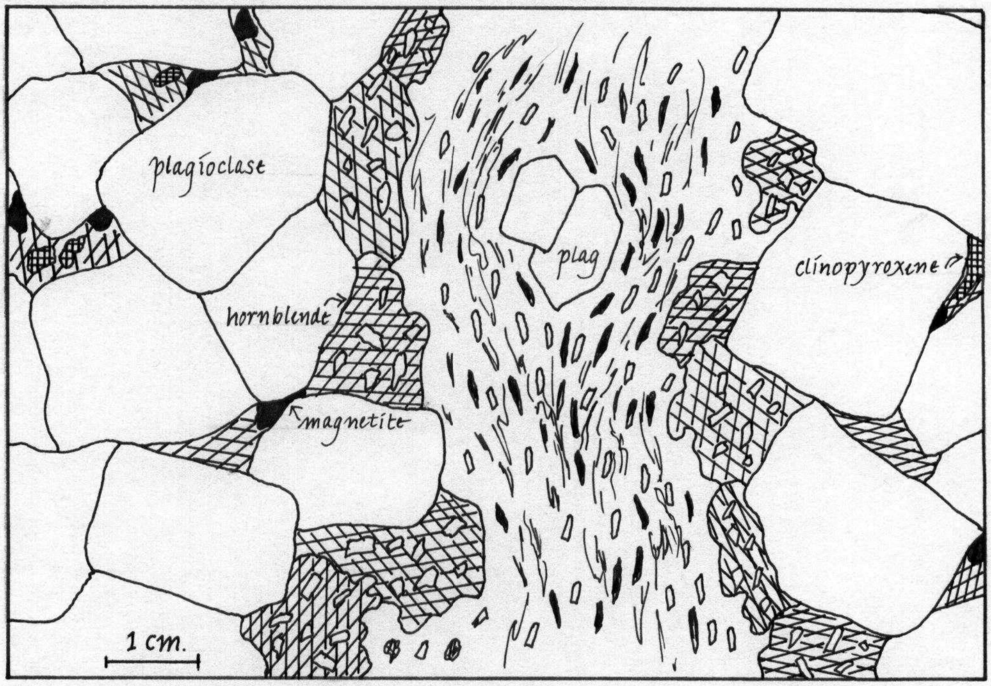
There are at least three distinct plutonic phases in the map area (see Geologic Map, Plate 2). The youngest is quartz diorite with a hornblende gabbro border phase that occupies most of the surface area of the islands. It is responsible for the second-phase structures described in Chapter IV and the contact metamorphism described in Chapter V. By Rb/Sr isochrons it is approximately 154 m.y. old; by K/Ar it is between 147 and 158 m.y. old. The older plutons are truncated and metamorphosed by it. Because resetting was expected they were not radiometrically dated. Their contact aureoles are not separable from that of the main intrusion.

### i. Older Bodies

One of the older bodies is a gabbro which intrudes the upper Karmutsen Formation west of Miner's Bay. Its texture is distinctive (Figure 10a), consisting of large squarish euhedral plagioclase surrounded by interstitial clinopyroxene, hornblende and

Fig. 10

## A. MINER'S BAY GABBRO WITH BASALTIC DIKE

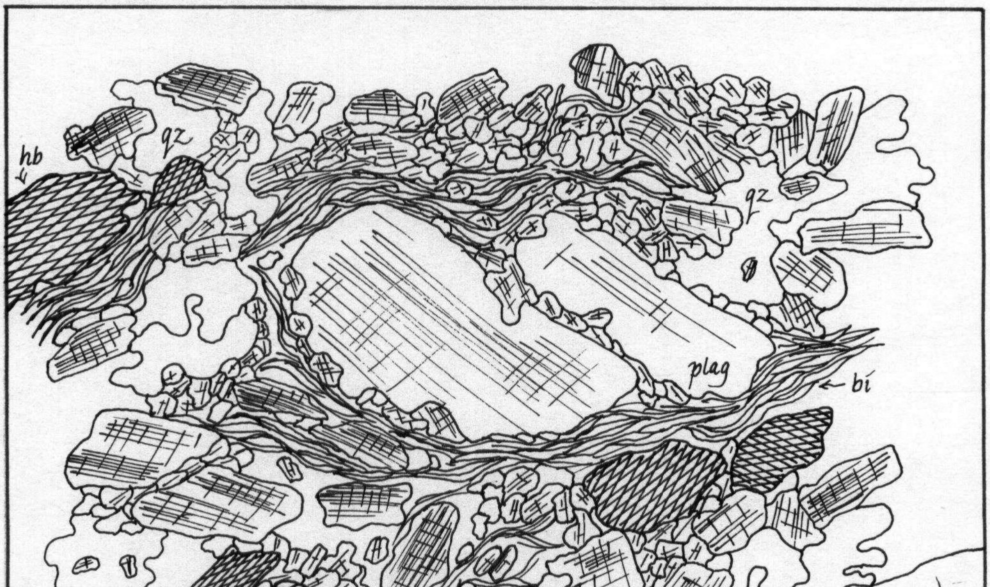



Note flow of dike around xenocryst.

Borders of dike are poikilitic hornblende.

Plag: An 45-76 oscillatory-zoned. clinopyroxene replaced by hornblende.

## B. WELLBORE CHANNEL GNEISSIC QUARTZ DIORITE




Note sutured borders on plagioclase and quartz.

plagioclase : An 50. weak oscillatory zoning.

magnetite. The Miner's Bay body is riddled with small irregular northwest-trending basaltic inliers, which commonly include large square plagioclase crystals. These, along with the obscurity of the contact relations, suggested an origin of the body by dioritization to the Geological Survey reconaissance team (Roddick and Hutchison, in preparation). An intrusive origin is favored here. Microscopically the basaltic inliers are clearly dikes. If they were Karmutsen, the large plagioclases in them should overprint any foliation. On the contrary, the inliers are pilotaxitic; the orientation of small plagioclase and hornblende laths bends around the large plagioclases (Figure 10a). In addition the texture of the Miner's Bay gabbro with its square oscillatory-zoned plagioclase and interstitial mafics bears no resemblance to that of Karmutsen metabasalts, even at the highest grades.

The contact between this body and the younger quartz diorite is an intrusive breccia. Comb-layered veins or dikes in the Miner's Bay body are probably associated with the younger intrusion.

The other older pluton is a gneissic quartz diorite exposed on Hardwicke Island next to Wellbore Channel. In thin section its prominent northeasterly foliation appears to be the result of deformation and recrystallization after initial cooling. (Figure 10b). The foliation is largely due to trains of fine-grained biotite and chlorite which bend around large plagioclase and hornblende crystals. Bent twins are common in plagioclase. Some plagioclases are broken, the fractures healed by recrystallization of small enhedral new plagioclase. Sutured borders surrounded by new crystals are typical. As no severe regional deformation has been inferred for the area, and as the gneissic quartz diorite occurs very near to the youngest intrusion, its deformation and low temperature alteration may be contact effects similar to those seen in the Bonanza Formation. The younger quartz diorite truncates it.

ii. Quartz Diorite and Associated Hornblende Gabbros

The youngest and largest intrusion trends west-northwest, parallel to the major faults and to the northeast coast of Vancouver Island. It consists of two phases: an incomplete sheath of hornblende gabbro (of insignificant volume), and an inner quartz diorite. It is interpreted here as a liquid or mostly liquid intrusion that differentiated below its present exposures to give two phases.

The quartz diorite phase is texturally and compositionally homogeneous except for variations in degree of foliation. Microscopically it is characterized by hornblende laths and large, complexly zoned and twinned plagioclases (Figure 11). Like most Coast Mountains plutons this one contains rounded fine-grained basaltic inclusions. Microscopically the inclusions display a non-directional polygonal texture very like Karmutsen metabasalts in the hornblende-plagioclase zone. They differ from Karmutsen volcanics in containing large oscillatory zoned plagioclase porphyroblasts, probably a result of interdiffusion and minerological

Fig. 11. QUARTZ DIORITE

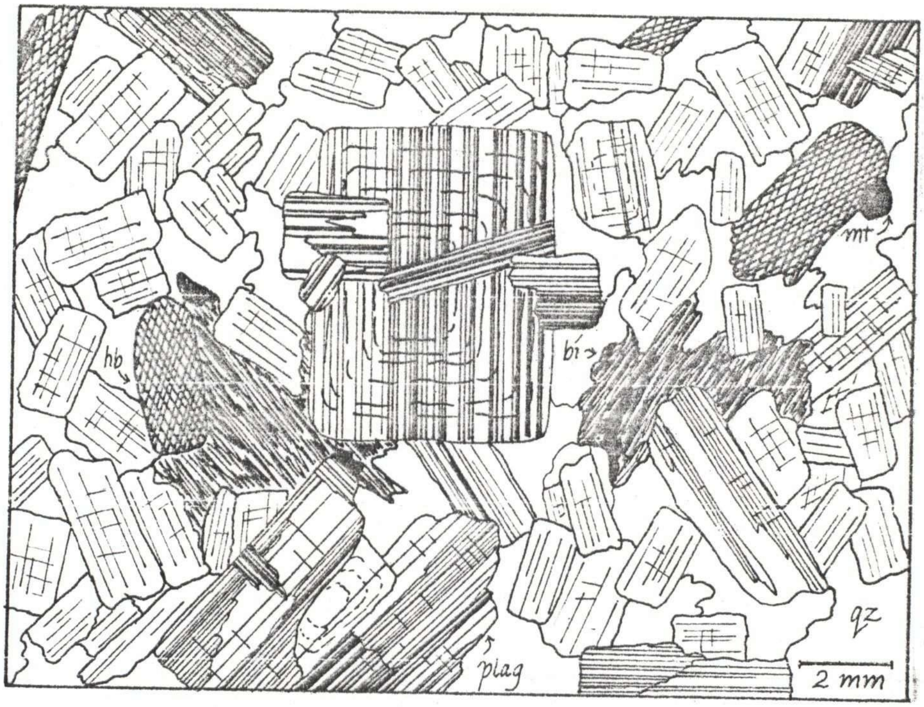



Fig. 12. Large-scale intrusive breccia, Miner's Bay

equilibrium with the enclosing quartz diorite over a considerable period.

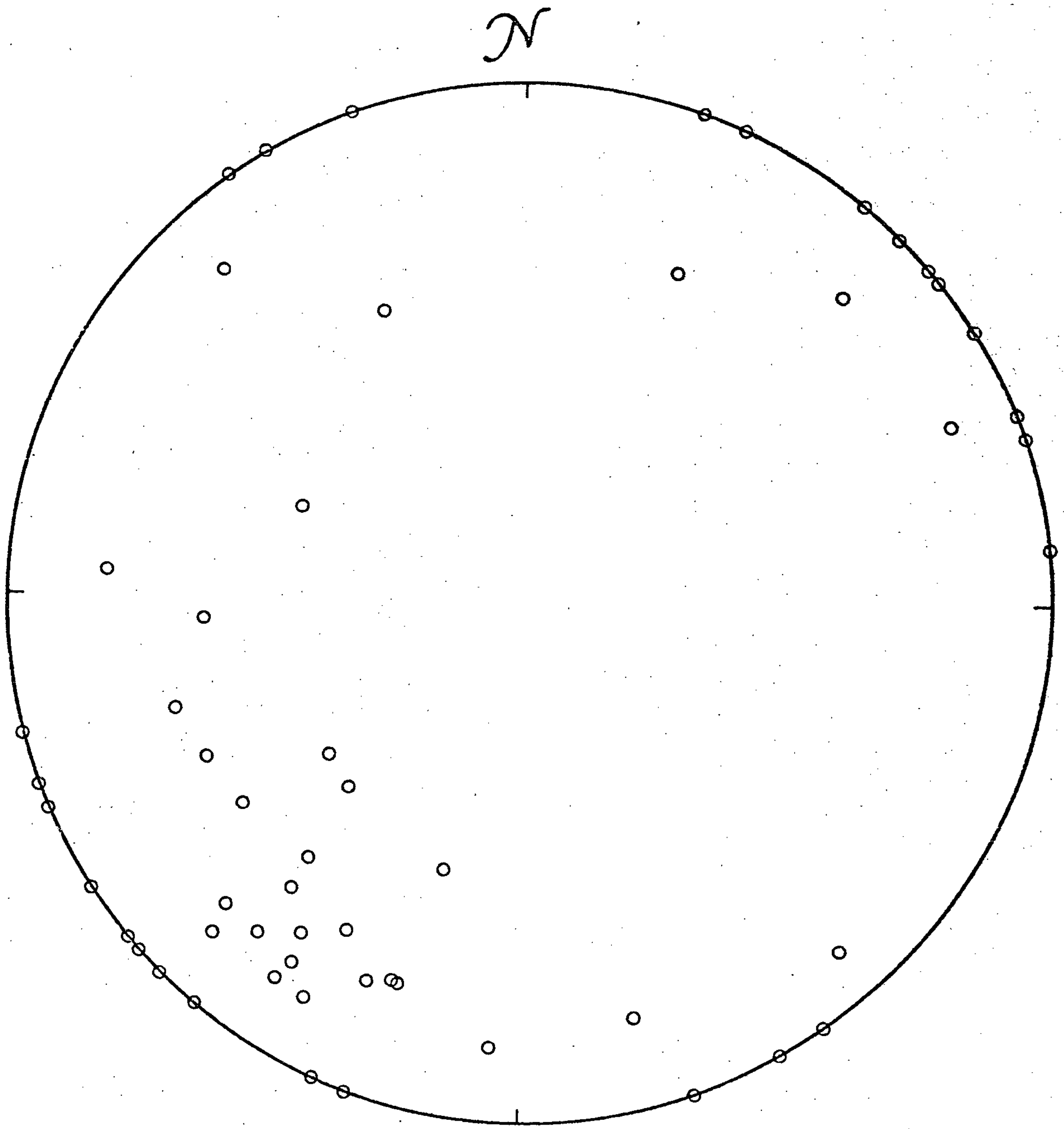
The quartz diorite truncates both the Vancouver Group and the earlier hornblende gabbros along sharp contacts. Dikes on all scales penetrate the country rocks, extending hundreds of meters from the parent body. Intrusive breccias, both mono- and polylithologic, are a common contact feature. The blocks in them are angular and can measure 2 meters across (Figure 12). These features are more indicative of liquid intrusion than of diapir emplacement or passive dioritization. They suggest forcible intrusion and a large viscosity contrast between pluton and country rock.

The foliation in the quartz diorite is best defined by flattened/stretched inclusions; biotite and hornblende are only weakly orientated. The foliation generally trends west-northwest but varies considerably (Figure 13). Near the outside of the body it tends to follow local contacts. The deformation of the inclusions suggests a rather small viscosity(or competency) contrast between them and the surrounding quartz diorite (G.Woodsworth, pers. comm. 1975). The foliation may thus have formed at a late stage of crystallization.

Low temperature alteration of the quartz diorite is ubiquitous but minor except in northeast-trending fracture zones (Chapter VII) and near contacts. The following secondary phases are present: pale green hornblende, chlorite and prehnite in biotite, saussurite in plagioclase, and epidote.

Fig. 13

Poles to foliation in quartz diorite

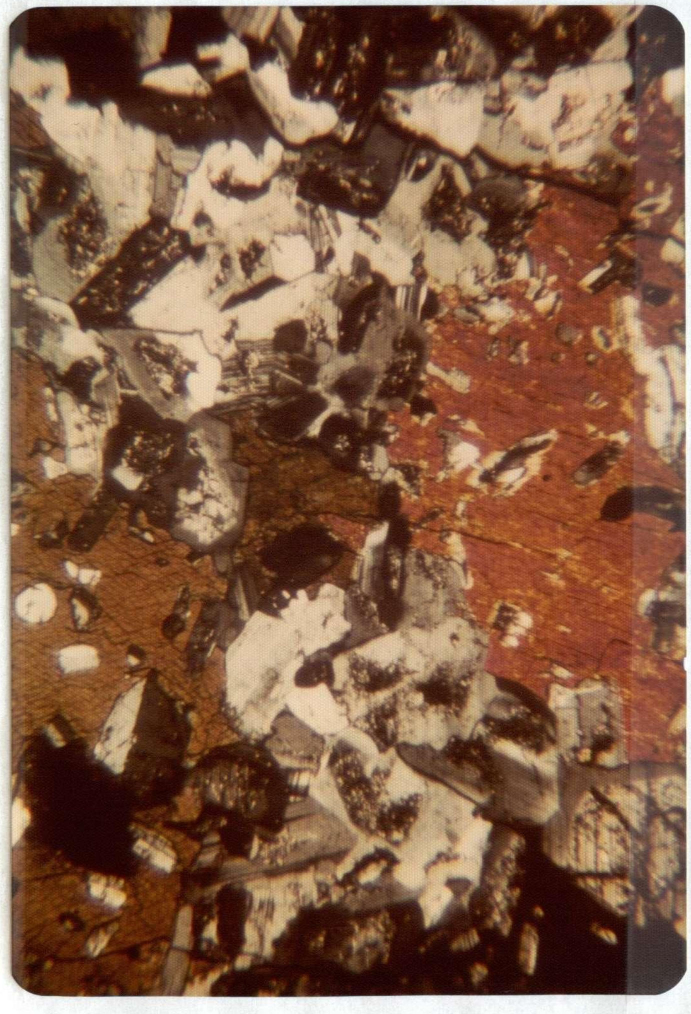


The hornblende gabbros are generally peripheral to the quartz diorite. They tend to be more extensive next to sediments than next to metavolcanics (see Geologic Map, Plate 2). (Several small gabbroic bodies within the quartz diorite may or may not be related to the border phase.) Peripheral gabbros between plutons and the Vancouver Group have been interpreted as dioritized and mobilized Karmutsen (Kuniyoshi 1972; Carlisle pers. comm. 1975). In this study area they are probably intrusive. There are clear textural distinctions between them and the metavolcanics. In addition, contacts with the Vancouver Group are sharp and intrusive breccias and dikes are common, pointing to liquid emplacement. The later quartz diorite sharply intrudes the hornblende gabbro: contacts are never transitional, although a succession of progressively more leucocratic phases may intervene, for example south of Eden Point. In contrast to the quartz diorite foliation is absent; inclusions in the hornblende gabbro are never stretched or flattened.

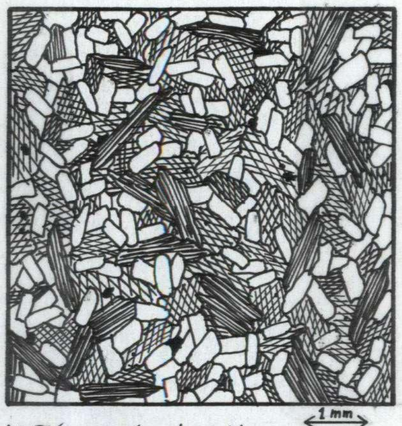
In striking contrast to the homogeneity of the quartz diorite, the hornblende gabbros are highly variable in grainsize, texture and composition. Proportions of the main phases, hornblende and plagioclase, vary both across contacts and gradationally within units. Numerous hornblende habits - laths, needles, anhedra, poikilitic globs or squares, clumps, veins, strings and layers - and sizes (from less than 1 mm. to 5 cm.) principally inform the textures (Figure 14). Plagioclase mainly occurs as small laths with strong normal zoning ($An_{50}$ to $An_{80}$). Clinopyroxene relics occupy the centers of some hornblendes. Magnetite is ubiquitous.

The hornblende gabbros possess several textures typical of the Sierran comb-layered bodies described by Moore and Lockwood (1973). These are: 1) orbicules, 2) comb-layered veins or dikes in the Miner's Bay body next to the hornblende gabbros, and 3) hornblende-plagioclase veins in the gabbro. Most gabbros have been pervasively altered. They contain several phases which are rare or absent in the quartz diorite: pyrite, sphene, large apatite included in hornblende, calcite and pumpellyite, as well as the more common alteration products. This minerology is also typical of the Sierran comb-layered intrusions. Moore and Lockwood (1973) believe comb-layering to form in the presence of a separate fluid phase. Lofgren and Donaldsen (1975) have shown experimentally that comb textures can result from supercooling and rapid crystallization rates such as might obtain near the margin of a plutonic body. The textures in the hornblende gabbros here do not support a firm choice between the two models. However the heterogeneity of the rocks may be explained by large variations in fluid pressure. That they crystallized in an active environment is shown by the deformation of the country rocks, by repeated cross-cutting and brecciation within the gabbros, and by vertical truncations and channels (Figure 15b, 17b, 18b)

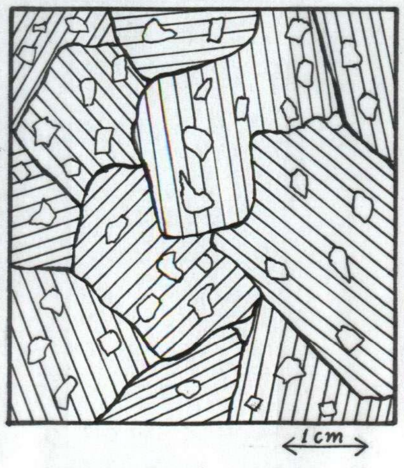
# Fig. 14 Textural variations in hornblende gabbros



a) Typical gabbro : hornblende poikilitic to skeletal ; plagioclase euhedral, strongly normal zoned. × 50




b) Fine-grained gabbro. Both plagioclase and hornblende are euhedral to subhedral.




c) Hornblendite, plagioclase inclusions (southern W. Thurlow Is.)

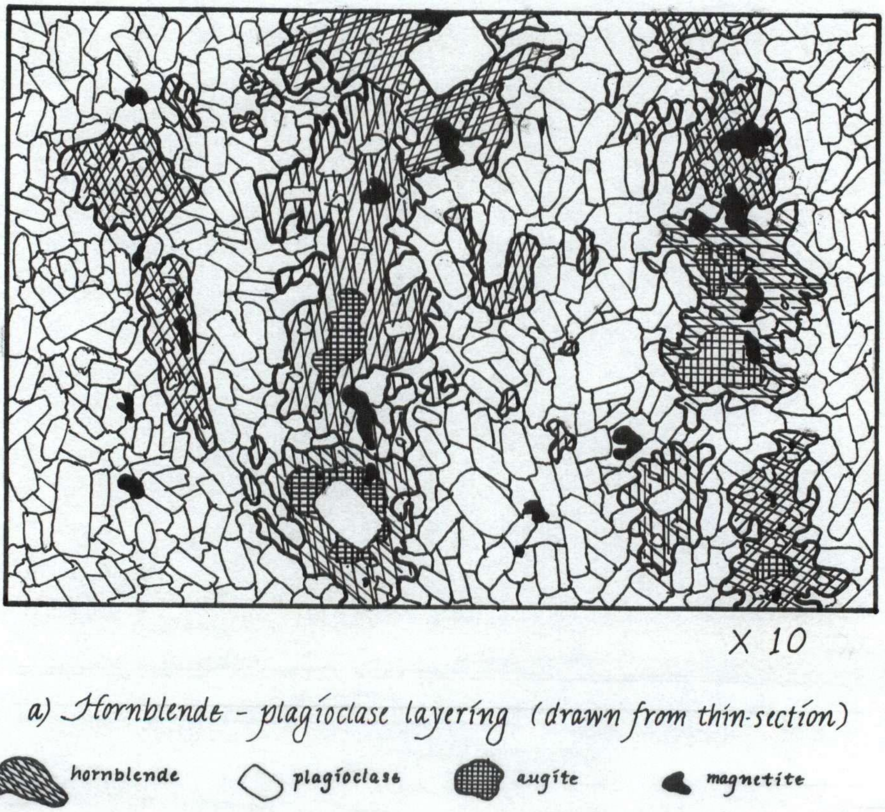
d) Orbicular dike, Clarence Island

Table 1 - Comparison between layering in hornblende gabbros, cumulate and schlieren layering.

| | Hornblende Gabbros | Schlieren Layering (Wilshire 1969) | Cumulate Layering (Wager and Brown 1968) |
|---|---|---|---|
| mineral morphology | Hornblende invariably anhedral - poikilitic, clumped; in some cases interstitial - skeletal. Plagioclase euhedral laths to enhedral. | Mostly euhedral. | Euhedral (cumulate phases) plus overgrowths. Intercumulus minerals may be poikilitic anhedral, skeletal. |
| mineral orientation | Plagioclase may be subparallel to layers. Hornblende never oriented. | Ideally long axes in plant of low pointing in flow direction. | Long axes commonly in plane of layering, but not invariably. Crescumulate textures form nearly perpendicular to layering. |
| shape of layers | Fine, continuous. | Bands or lenses, if crystal segregations. Wispy and discontinuous if due to "smearing" of inhomogeneities. | Fine, continuous. |
| layer orientation | Generally vertical or near vertical (Figure 15b). | Any orientation. | Low dips (gravity controlled). |
| structures | Truncations, channels, cross-beds, "clastic" dikes, grading, intermixing and crossculting of massive gabbros. | Bifurcations, truncations, cross-beds, plastic deformation, grading*, intermixing with massive phases. | Truncations, channels, cross-beds, grading mostly one way. |

* both ways.

Fig 15




a) Hornblende - plagioclase layering (drawn from thin-section)

b) Layered hornblende-rich gabbro truncated by lighter unlayered phase. Note the near-vertical attitude of the layering. - Clarence Island.

As Lofgren and Donaldsen point out, disturbances could cause rapid changes in fluid pressure. The hornblende gabbros are best developed next to or near sedimentary intervals: the high water content of the sediments may have been important in their genesis.

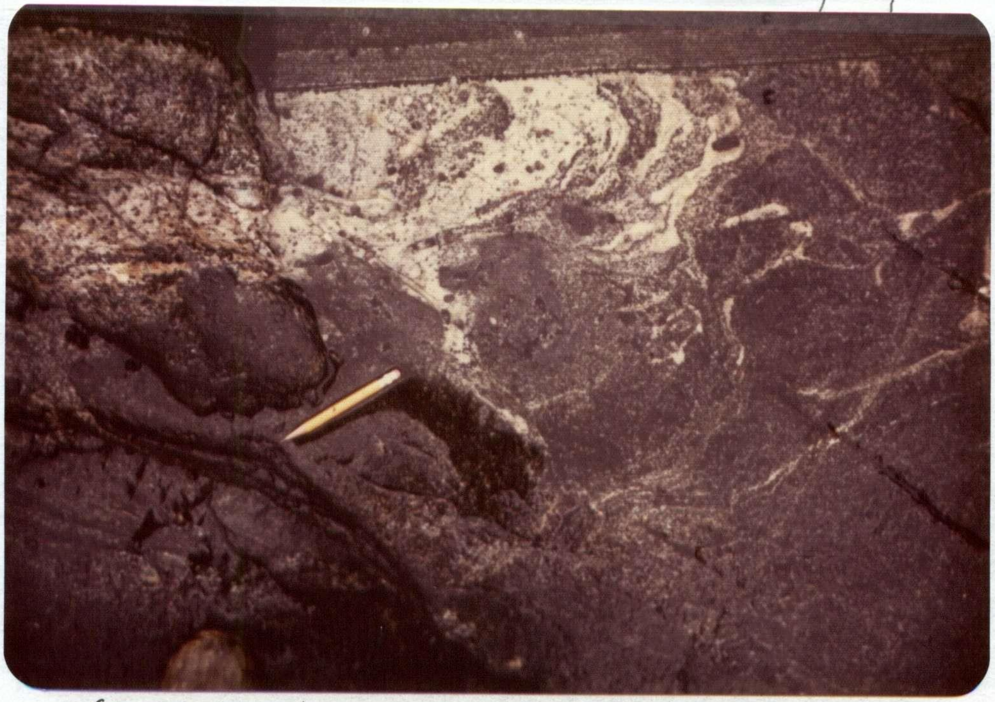
Delicate compositional layering, due to rhythmic variation in plagioclase/hornblende ratios, is seen in gabbros on Clarence Island, northern Yorke Island, and on the southern coast of West Thurlow Island. Plagioclases in the layered gabbros are generally euhedral and are much smaller than the anhedral hornblendes, which range up to 1 centimeter in diameter (Figure 15a). The larger hornblendes poikilitically enclose plagioclase: they are later than plagioclase in the paragenetic sequence. The reverse, plagioclase enclosing hornblende, is never seen. The layering shows certain similarities to schlieren layering and to cumulate layering but differs from both of them in significant respects (Table 1 and Figures 15 - 18). It does not appear to be due to alignment and flow-sorting of plagioclase and hornblende phenocrysts: the hornblende is anything but euhedral. In appearance the layering resembles accumulate layering, but the attitudes of the layers are nearly vertical. (Most layering of this type seen in the Coast Mountains is subvertical - Hutchison, pers. comm. 1975). Postcrystallization rotation from near-horizontal to near-vertical is implausible, as bedding just east of Clarence Island dips less than $40^{\circ}$. It is thus certain layering formed parallel to the pluton wall rather than to its floor.

Fig. 16

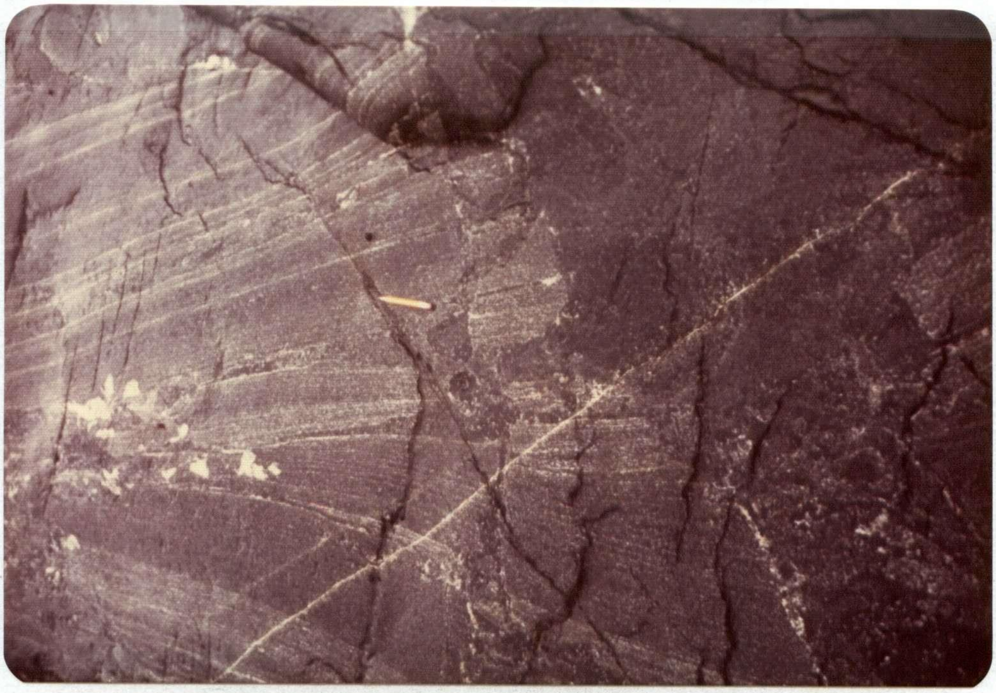
a) Near-vertical cross-bedding, Clarence Island.

b) Agmatitic zone between normal layers, Clarence Island.

Fig. 17



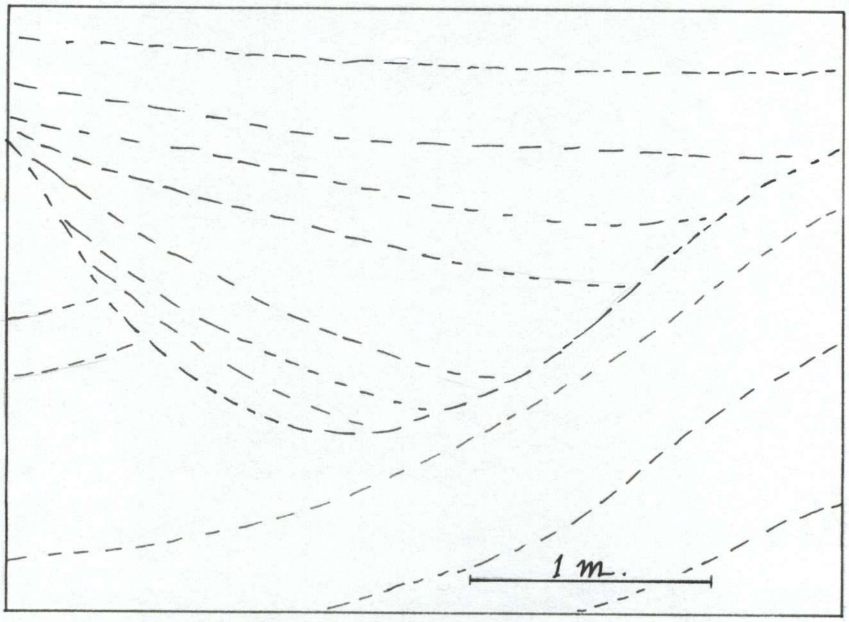
a) deformed layering truncated by planar layers; large euhedral hornblendes along "unconformity" - Clarence Is.



b) layered gabbro on left intruded by gabbro dike; dikelets cut across and between layers. Note diverging layers at lower left. – Clarence Island.

Fig. 18

a) "Clastic dike", Clarence Is. Layers near-vertical.



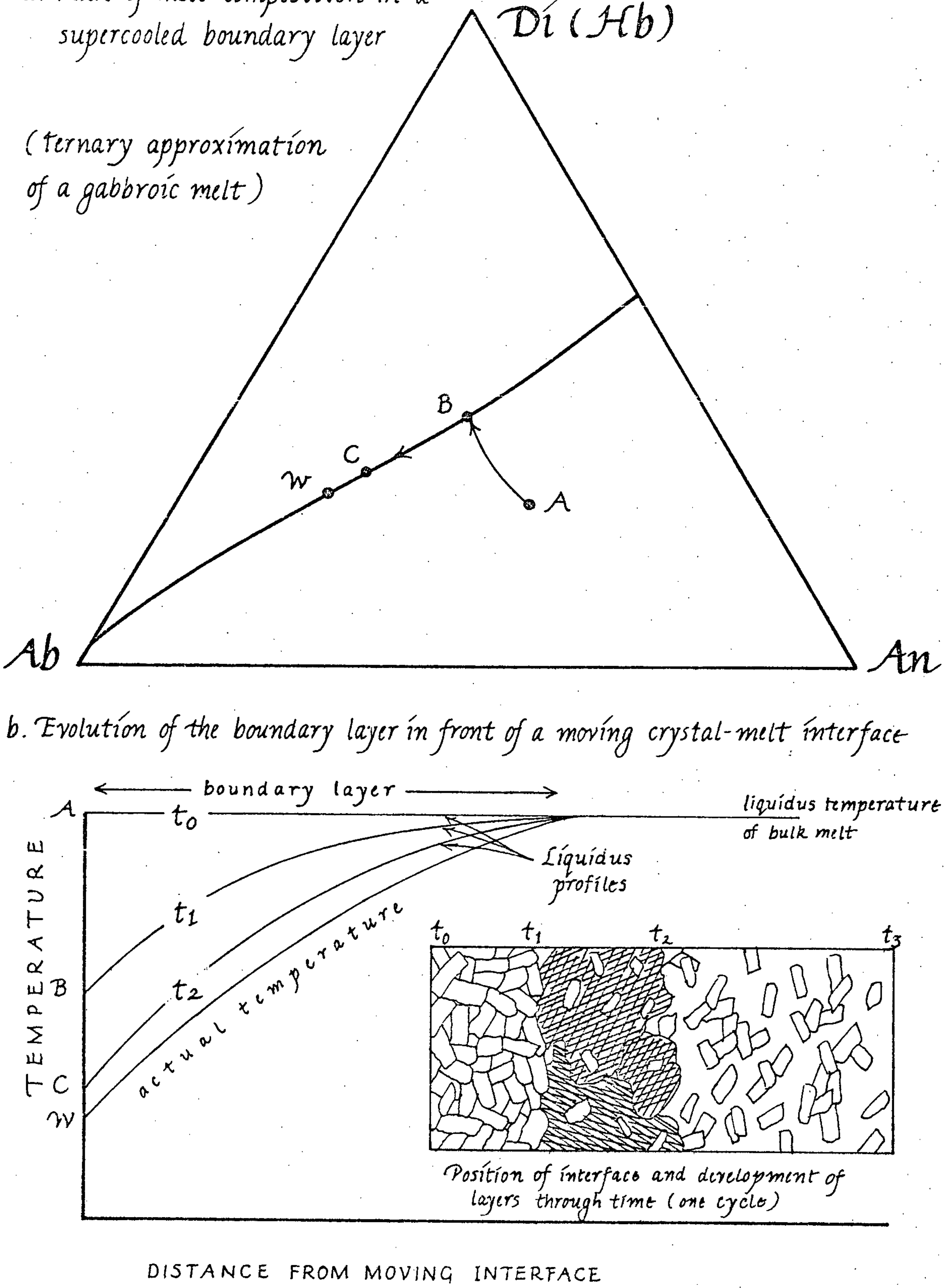

b) Channel, Clarence Is.

Figure 19 shows a possible model for the formation of hornblende-plagioclase layers along the wall of a magma chamber. In Figure 19a the hornblende-plagioclase cotectic in a real system is schematically represented by the plagioclase-diopside cotectic in the water-free ternary system Di - Ab - An. A hypothetical melt whose bulk composition is shown at A contains numerous small plagioclase crystals and under normal crystallization is wholly in the plagioclase field. It comes to rest against a cool wall (wall temperature, W, shown on ternary diagram). If no convection occurs, supercooling will induce rapid accretion to plagioclase nuclei next to the wall. If the rate of crystallization of plagioclase outstrips diffusion rates in the melt, the melt composition near the wall will change from A to B on the cotectic. Plagioclase in the layered gabbros is strongly normal zoned. At B hornblende begins to crystallize along with plagioclase. The compositional gradient which develops between the melt at the wall and the bulk melt is, as shown in Figure 19b, also a liquidus-temperature gradient. Figure 19b shows a series of liquidus profiles developed at successive times in front of a moving crystal - melt interface. From time $t_0$ to $t_2$ the rate of crystal growth is greater than diffusion. It is fastest on the crystal-melt interface because the degree of supercooling is greatest there; however crystallization of plagioclase occurs throughout the boundary layer. Up to time $t_1$ plagioclase alone crystallizes. At $t_1$ hornblende can nucleate

Fig. 19

a. Path of melt composition in a supercooled boundary layer

(ternary approximation of a gabbroic melt)




After Lofgren and Donaldson 1975

near the interface. As the liquidus temperatures in the boundary layer progressively approach the actual temperatures, the rate of crystal growth slows. At $t_2$ the rate of crystallization at the interface is equal to the rate of plagioclase crystallization in the outer boundary layer. Rapid crystallization at the interface no longer dominates; the whole boundary layer crystallizes more slowly as a unit. A new crystal - melt interface is then established to the right of the second plagioclase layer (time $t_3$) and the sequence is repeated. Each cycle consists of three layers; plagioclase-rich, hornblende-rich, plagioclase-rich. Successively younger layers thus form parallel to the initial wall of the pluton and the locus of crystallization moves progressively inward. This model is similar to those presented by Lofgren and Donaldson (1975) and Sibley et al. (1976) to account, respectively, for comb-layering in igneous rocks and for oscillatory zoning of plagioclase.

It should be noted that the wall temperature W (Figure 19a) lies above the gabbro liquidus. The layered gabbros probably were not competely solid when they first formed; rather they were a framework containing some interstitial melt. This interstitial melt would produce clastic dikes (Figure 18a) and aid in deformation of the layers (Figure 17a).

The principal differences between the quartz diorite and the hornblende gabbros are 1) homogeneity vs. heterogeneity, 2) foliation vs. lack of directional fabric, 3) minor vs. pervasive alteration, 4) plagioclase in the quartz diorite is intricately zoned and twinned while in the gabbros it is strongly normal zoned and simply twinned, and 5) hornblende in the quartz diorite is lath-shaped while its most common habit in the gabbro is anhedral poikilitic. Some of these differences have been attributed to the role of the fluid phase. Throughout the main part of the intrusion it would evolve (or diffuse in) slowly; but when the early gabbros encountered wet sediments large amounts of fluid would be available. The lack of foliation and strong plagioclase zoning, may reflect rapid cooling. The gabbro textures would thus be due to chilling of an early phase of the intrusion.

The history of the main intrusion can be summarized as follows:

1. The early gabbros rose from a magma chamber at depth along deep crustal fractures parallel to the Telegraph Hill Fault.

2. Differentiation (crystal fractionation?) gave rise to the quartz diorite which rose along the same fractures. It cut into and brecciated the gabbros as well as the Vancouver Group. At this point the country rocks began to deform.

3. Expansion of the quartz diorite after emplacement due to continued introduction of magma created the foliation within it and persistent deformation outside it.

4. Two possibilities for the late cooling history and alteration of the quartz diorite and the dikes in it will be discussed in the next sections. It may have cooled to an ambient temperature within the prehnite-pumpellyite facies, or the alteration may have been due to heat remaining within the intrusion itself.

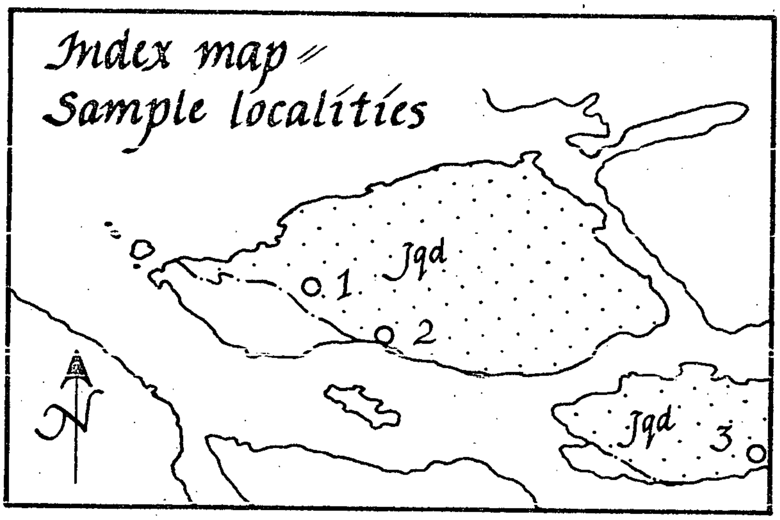
iii. Radiometric Dating: Quartz Diorite

Absolute ages obtained from three localities within the youngest intrusion on Hardwicke and West Thurlow Islands (Figures 20 and 21; Table 2) lie between 147 and 158 m.y. Both potassium-argon mineral dates and rubidium-strontium biotite isochrons are concordant within experimental error except the biotite-hornblende pair from site 1 (discussed below). The apparently simultaneous setting of the various radiometric clocks implies a rapid temperature drop through the critical isotherms. One interpretation which readily explains the concordance is that the clocks were set during crystallization of the pluton and not disturbed since.

The dikes which cut the quartz diorite have been metamorphosed at temperatures above 200°C (see Chapter VI). Hart (1964) showed that significant diffusion of argon and strontium from biotite can occur down to 200°C. Thus it is possible that the intrusion cooled to an ambient temperature above that required for argon and strontium retention in biotite and (because of the observed concordance) above argon retention in hornblende as well. If so, the dates obtained do not record intrusion; rather they mark the end

Fig. 20

# Radiometric dating of quartz diorite



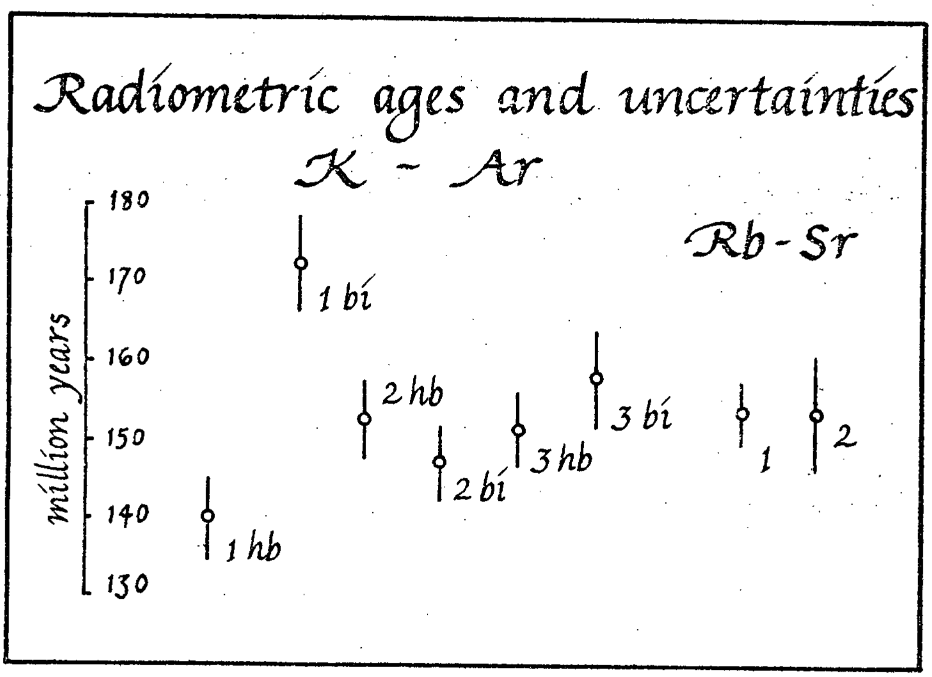
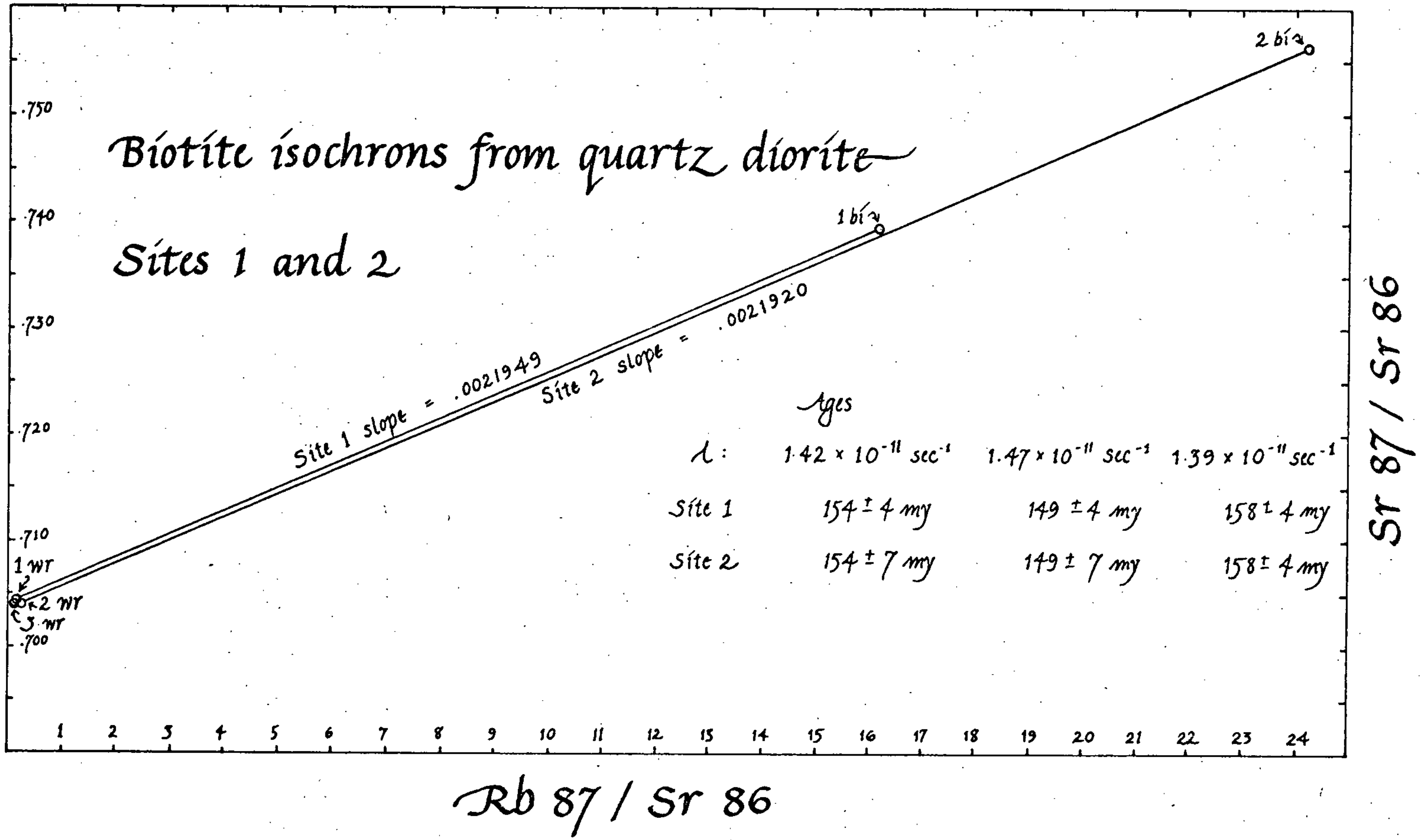



Fig. 21


Biotite isochrons from quartz diorite
Sites 1 and 2
Site 1 slope = .0021949
Site 2 slope = .0021920
2 bi
1 bi
1 wr
2 wr
3 wr
.750
.740
.730
.720
.710
.700
1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21 22 23 24
Ages
λ: 1.42 × 10⁻¹¹ sec⁻¹ 1.47 × 10⁻¹¹ sec⁻¹ 1.39 × 10⁻¹¹ sec⁻¹
Site 1 154 ± 4 my 149 ± 4 my 158 ± 4 my
Site 2 154 ± 7 my 149 ± 7 my 158 ± 4 my
Rb 87 / Sr 86
Sr 87 / Sr 86

## Table 2 - Radiometric Age Data From Quartz Diorite

Rb-Sr

| sample no. | material | Rb(ppm) | Sr(ppm) | Rb/Sr $\pm 1_S$ | Sr87/Sr86 $\pm 1_S$ | slope | int. | apparent age(my) |
|---|---|---|---|---|---|---|---|---|
| 1 | whole rock | 29.8 | 366.6 | $.2351 \pm .00470$ | $.70442 \pm .0016$ | .0021949 | .7039 $\pm$.0002 | $154 \pm 7$ |
| | biotite | 179.1 | 32.07 | $16.2083 \pm .32417$ | $.73948 \pm .0013$ | | | |
| 2 | whole rock | 29.4 | 384.9 | $.2209 \pm .0042$ | $.70425 \pm .00054$ | .0021920 | .7037 $\pm$.0005 | $154 \pm 4$ |
| | biotite | 200.2 | 24.02 | $24.2308 \pm .4846$ | $.75688 \pm .00046$ | | | |

K-Ar

| sample no. | material | K. | moles K40/gm | moles Ar40/gm (rad) | apparent age (my) |
|---|---|---|---|---|---|
| 1 | biotite | 6.26 | $1.908 \times 10^{-7}$ | $2.016 \times 10^{-9}$ | $172 \pm 6$ |
| | hornblende | .584 | $1.777 \times 10^{-8}$ | $1.509 \times 10^{-10}$ | $140 \pm 5$ |
| 2 | biotite | 5.59 | $1.799 \times 10^{-7}$ | $1.522 \times 10^{-9}$ | $147 \pm 5$ |
| | hornblende | .336 | $1.023 \times 10^{-8}$ | $9.446 \times 10^{-11}$ | $152 \pm 5$ |
| 3 | biotite | 5.80 | $1.765 \times 10^{-7}$ | $1.703 \times 10^{-9}$ | $158 \pm 6$ |
| | hornblende | .359 | $1.093 \times 10^{-8}$ | $1.008 \times 10^{-10}$ | $151 \pm 5$ |

Site data and lithologic descriptions for age-dating samples.

| sample no. | field no. | latitude | longitude | lithology |
|---|---|---|---|---|
| 1 | 5.28.1 | 50°26'15"N | 125°54'9"W | coarse grained equigranular slightly foliated biotite-hornblende quartz diorite. |
| 2 | 5.16.1 | 50°25'11"N | 125°51'51"W | same |
| 3 | 8.14.1 | 50°23'5"N | 125°40'40"W | same |

Analytical Methods:

Potassium - Perkin-Elmer atomic absorption spectrophotometer by Dr. R. L. Armstrong

Argon - mass spectrometry, by Mr. Joe Harakal

Rubidium/Strontium - X-ray fluorescence following the method developed by J. Ryan (1973)

Sr87/Sr86 and total Strontium - mass spectrometry, by Dr. P. LeCouteur.

SAMPLE PURITY

| Sample | Percent Impurities | | | | |
|---|---|---|---|---|---|
| | hb | bi | chl | sphene | total |
| 1 hb | | 1.6 | | .6 | 2.2 |
| 1 bi | 4.5 | | 1.4 | | 5.9 |
| 2 hb | | 1 | .2 | 1 | 2.2 |
| 2 bi | .2 | | 2.4 | .2 | 2.8 |
| 3 hb | | .2 | .3 | 3.2 | 3.7 |
| 3 bi | 1.2 | | 1.2 | | 2.4 |

total 500 counts

of a low-grade metamorphic episode of unknown duration. Carson (pers. comm. 1974) considers the younger dates from the Middle Jurassic Island Intrusions (of which this may be one) to record metamorphism after emplacement. In this case uplift would have had to be very rapid in order to set all clocks simultaneously. A preferable explanation, consistent with the observed concordance, is that the dikes were intruded and subsequently metamorphosed while the quartz diorite was solidified but still hot (see Chapter VI).

The 172 m.y. (K/Ar) biotite and the 140 m.y. (K/Ar) hornblende from site 1 are not concordant. Does this biotite preserve an original cooling age, while all other clocks have been reset by low-temperature metamorphism? Or has it acquired excess argon either in the parent rock or during analysis? The latter possibility is more probable. Cases of K/Ar dates exceeding Rb/Sr dates from the same biotite are uncommon (York and Farquhar 1972). The discordance between K/Ar dates on biotite and hornblende is also opposite from the temperature-resetting predicted by Hart's diagram (1964) and normally observed in rocks.

Whether the Upper Jurassic age of the quartz diorite is taken as crystallization or as post-metamorphic cooling it is still one of the oldest ages so far obtained within the southern Coast Plutonic Complex. In fact, since Coast Mountains intrusions are generally Early Cretaceous to Eocene and the Island Intrusions Middle Jurassic, one might prefer to assign the quartz diorite on Hardwicke and West Thurlow Islands to the Insular Belt on the basis of its age rather than to the Coast Plutonic Complex because of its position. The issue involved is whether the western boundary of the Coast Plutonic Complex should be drawn along the western limit of Cretaceous intrusions or along the western limit of predominantly plutonic terrane. The latter interpretation is followed here, since the lack of Jurassic dates in the western Coast Mountains may partly reflect preferential sampling of unaltered bodies (Woodsworth, pers. comm. 1974). This part of the province boundary, then, is at least 148 m.y. old. Of course the present strong distinction between the two provinces did not exist at that time.

The initial Sr87/Sr86 ratios calculated from the isochrons are .7037 and .7039 (Table 2). They are among the most "primitive" observed in continental intrusive rocks; they closely resemble the ratios found in circum-Pacific andesites. Such low radiogenic strontium rules out both the Central Gneiss Complex and older sialic crust as possible parents. It does not, however, prohibit crustal origin. Culbert (1970) observes that the Rb/Sr ratio in typical Coast Mountains intrusions is so low that Sr87/Sr86 only changes by .001 every 550 m.y. Thus two-stage models for the origin of plutons such as this one are just as acceptable isotopically as single-stage derivation from the upper mantle or a subducted slab. The most convincing uses of strontium isotopes as indicators of crustal or subcrustal origin depend on patterns from large areas rather than single values (e.g. Church and Tilton 1973; Kistler and Peterman 1973). Such a synthesis is not yet possible for the Coast Mountains. LeCouteur and Tempelman-Kluit (1976) have assembled the available strontium isotopic data from western British Columbia. Their summary shows clearly the present paucity of data in the Coast Mountains and the vast opportunities for future isotopic work.

## CHAPTER IV PRE-AND SYN-PLUTONIC STRUCTURE AND DIKES

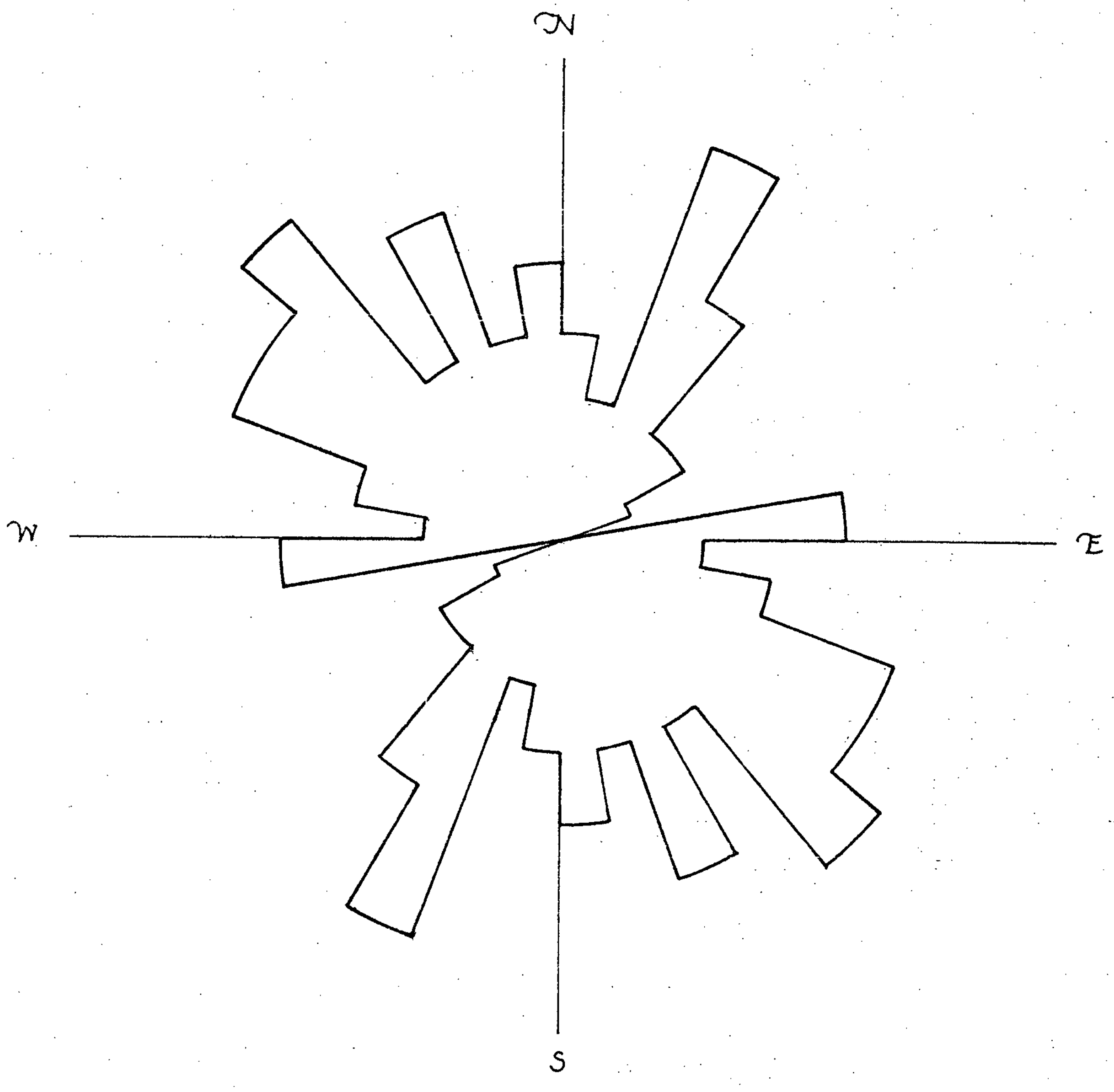
The tectonic history of this area in its simplicity and mildness is more akin to the Insular Belt than to the Coast Plutonic Complex. The first phase of deformation involved pre-plutonic faulting; the second, synplutonic deformation attributed to forceful intrusion; and the third, fracturing and minor faulting which accompanied post-plutonic dike emplacement (Chapter VI).

### i. Pre-Plutonic Dikes

There are two sets of pre-plutonic dikes in the map area. The first includes basalt, andesite and gabbro dikes and sills probably related to the Karmutsen volcanic suite. Wherever cross-cutting relations are seen these are transected by "grey" dikes of the second set which includes andesites, basaltic andesites and hornblende porphyry dikes resembling Bonanza feeders on Vancouver Island. These "grey" dikes are equally abundant in Karmutsen and Bonanza rocks. They are contact metamorphosed to the same grade as their surroundings. They are involved in syn-metamorphic folding; near the intrusion they may be foliated. The "grey" dikes are all steeply dipping to vertical. They show no clearly preferred strike orientation (Figure 22). It is reasonable to infer from this that no directed tectonic stress prevailed during their emplacement. If they are Bonanza dikes, then the beginning of the first deformational episode postdated Bonanza volcanism.

Fig. 22

# Strike orientations of pre-plutonic dikes



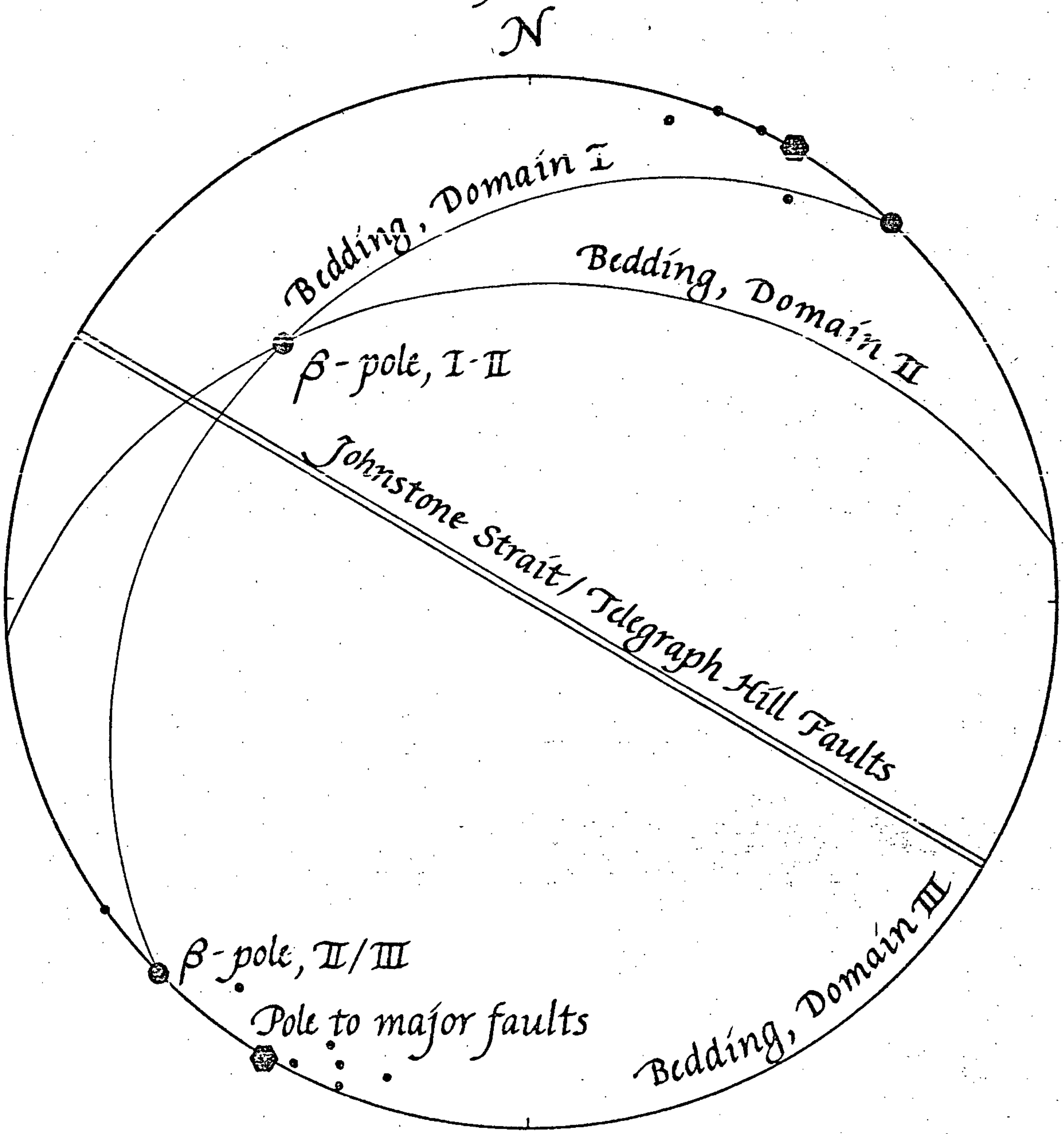

total points 65

ii. Phase One (Pre-Plutonic): The Major Faults

There are two steep to vertical faults of large displacement in the area. They trend approximately 120°, passing to the north and south of Helmcken Island, and lie mostly under water. The Johnstone Strait Fault (Kuniyoshi 1972) forms a prominent topographic depression and slope break on Vancouver Island. The middle and upper Karmutsen Formations lie northeast of it (Domain II); the lower Karmutsen Formation lies to the southwest of it (Domain III). The maximum possible stratigraphic separation across it is 7,000 meters. The Telegraph Hill Fault juxtaposes Bonanza and upper Karmutsen rocks along the northeast side of Telegraph Hill on Hardwicke Island. The maximum possible stratigraphic separation across it is 5,000 meters. In Figure 23 the rotational relationships between bedding in adjacent domains are shown. The Johnstone Strait/Telegraph Hill fault planes are assumed to be vertical. The $\beta$. pole relating Domains II and III is 44/0. The rotation path is nearly in the fault plane, as required in rotation faulting (Ragan 1968). The $\beta$. pole relating Domains I and II is 314/32, lying nearly in the fault plane, as if it were an axial plane. This type of motion, with a rotation path perpendicular to the fault plane, is not typical of slip along a planar fault. The fault surface might curve at depth to become convex towards Domain I. Alternatively the Telegraph Hill Fault may be on axial plane, the broken back of a large fold with planar limbs. The latter

Fig. 23

Rotational relationships between homoclinal domains




• poles to minor faults and shear zones

possibility is likely. The $\beta$. pole determined for the fault coincides with the axes of the mesoscopic folds. Instead of folding under compression the Vancouver Group volcanics fractured. Reverse faulting at the hinge would account for the vertical separation. Steep minor faults on Helmcken and Hardwicke Islands parallel the major faults and are presumably related to them.

Movement on the Telegraph Hill Fault ceased before emplacement, or at least cooling, of the main quartz diorite: it is truncated by an intrusive contact on the ridge southwest of Clam Bay. The parallelism between the major faults and the strike of the plutonic contact suggests that the intrusion rose along deep faults of the same system. Several Insular Belt plutons have utilized pre-existing faults and are consequently elongate parallel to them (Sutherland Brown 1966).

iii. Phase Two: Syn-Plutonic Deformation

The second phase of deformation had a variety of expressions depending on the strain-reponse characteristics of the various units. Vancouver Group volcanics in greenschist facies (outer aureole) were not significantly folded. It is likely that they are cut by numerous synplutonic faults and fractures; however these would have the same trend as phase-one structures and thus cannot be separated from them. The cherty interlava lenses in the upper Karmutsen Formation and thin-bedded Harbledown sediments developed consistent mesoscopic folds. Within 75 meters of the of the quartz diorite/gabbro intrusion chaotic folding affected the Quatsino and Parson Bay Formations. The Bonanza metavolcanics in the inner aureole underwent ductile deformation which produced a strong foliation. All of these features (except the chaotic folding) are northwest to west-northwest trending; most are demonstrably syn-metamorphic.

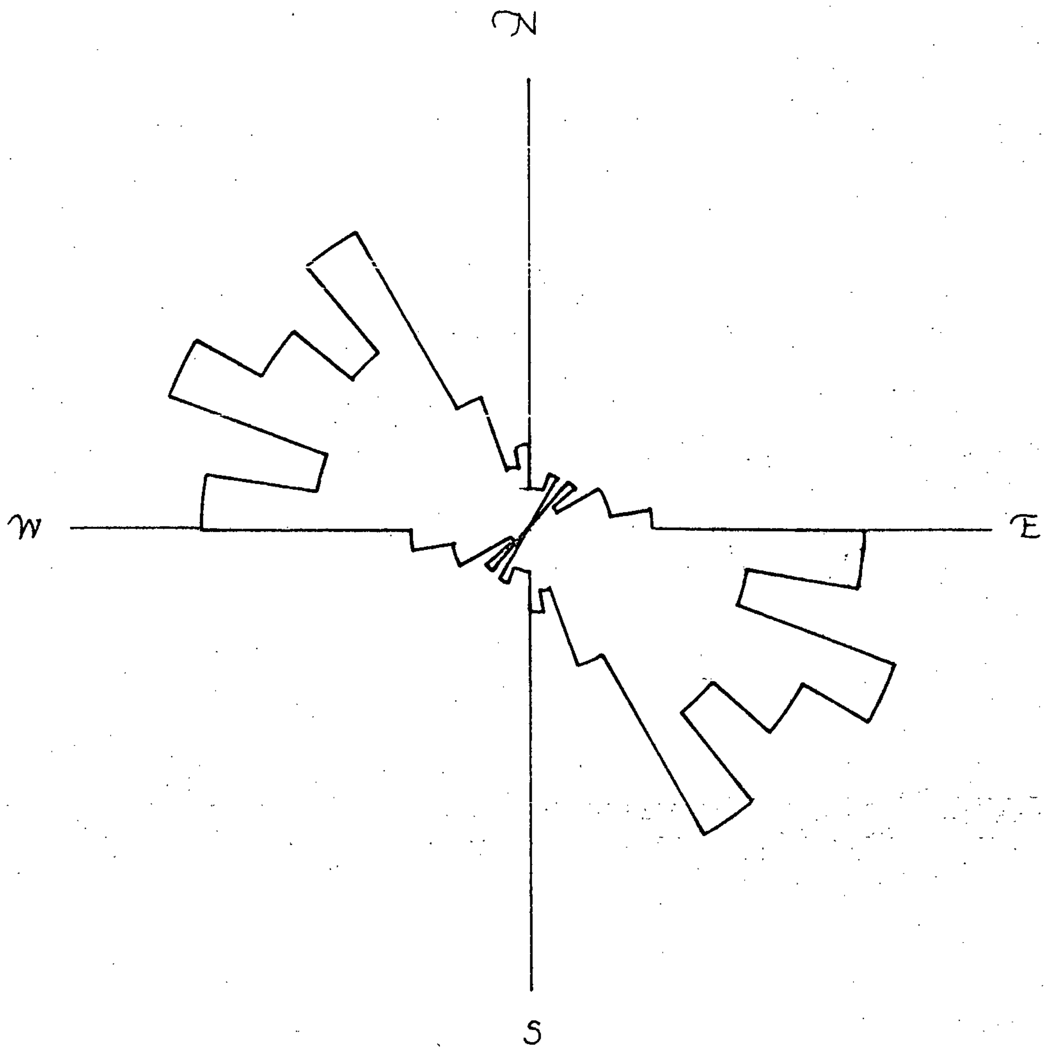
Epidote stringers are abundant in Karmutsen volcanics in the outer aureole of the main intrusion. They fill small fractures and faults which generally strike between $90^{\circ}$ and $150^{\circ}$ (Figure 24). The epidote segregation is probably linked to greenschist contact metamorphism (Chapter V); but many of the fractures may be pre-plutonic.

Minor fold orientations in Domains I and II (Figure 25) are consistent except in very thin interlava lenses and in sediments near the intrusive contact. Vergences of the mesoscopic folds in both domains are uniformly southward, indicating structural culmination towards Vancouver Island. The average axial plane is 123/58 NE., considerably steeper than bedding but not as steep as the foliation or major faults. The minor folds were probably generated by shearing between the massive volcanics that enclose the sediments (Figure 26).

A second phase of deformation affecting the mesoscopic fold orientations is not apparent on Figure 25. Orientations of the few folds observed in Domain II coincide with those in Domain I.

Fig. 24

# Strike orientations of epidote veins




total points 133

Fig. 25

Orientations of minor folds

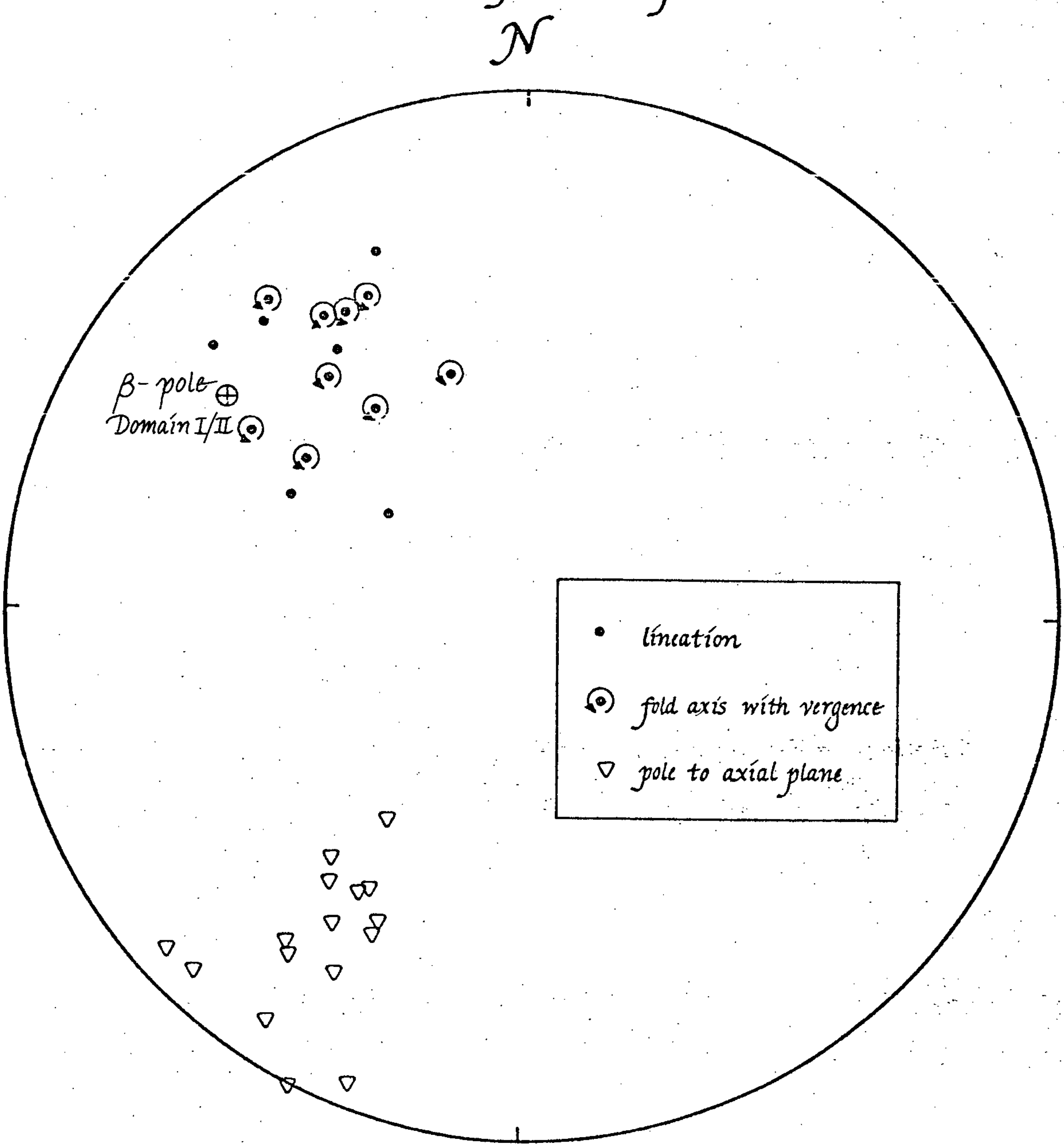


Fig. 26

Minor folds in interlava sediments, upper Karmutsen Formation, West Thurlow Is.

Bonanza (?) dike cuts sediments on left. Cliff in upper left background is Karmutsen flow over-lying sediments.

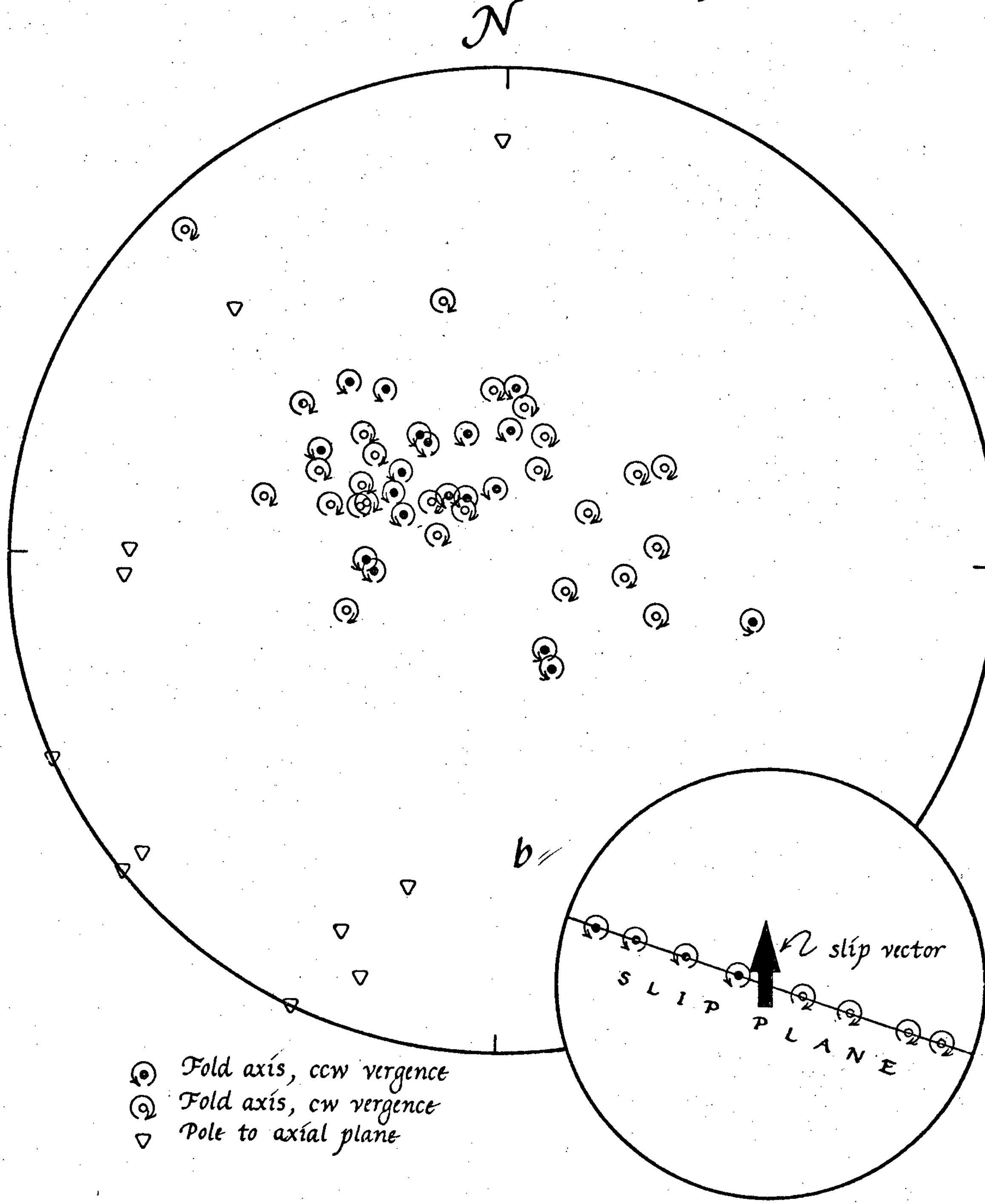
This would imply that rotation on the Telegraph Hill Fault had ceased before development of the minor folds. The style of the folds (Figure 26), not typical of brittle deformation, may indicate that they formed during metamorphism.

Within 75 meters of the quartz diorite intrusion fold orientations lose consistency. A dramatic instance of chaotic folding is seen in the upper Quatsino and lower Parson Bay Formations east of Vansittart Point. Thin cherty beds are chaotically folded and boudinaged; rootless folds of all orientations swim in a marble matrix. Wollastonite laths aligned in the axial plane of at least one fold suggest that deformation occured at the peak of contact metamorphism. Two-centimeter to three-meter gabbroic dikes are folded and boudinaged.

The increasing intensity of deformation towards the intrusion and the coincidence of deformation with contact metamorphism suggests that forceful intrusion may have caused the deformation. If so, fold orientations can show whether the motion between pluton and country rock was essentially vertical, parallel to the contact, or whether it was compressional. Figure 27a is a plot of axial planes and fold axes from this exposure. Figure 27b, shown for comparison, is a theoretical diagram after Hansen (1971) showing the fold axis/vergence distributions given the following conditions: 1) the intrusive contact is vertical and trends $110^{\circ}$ and 2) that folding is completely due to shear between the pluton and the country rock, the pluton moving relatively upwards. In such a situation the mesoscopic folds

Fig. 27

a – Orientations of chaotic folds, east of Vansittart Pt.



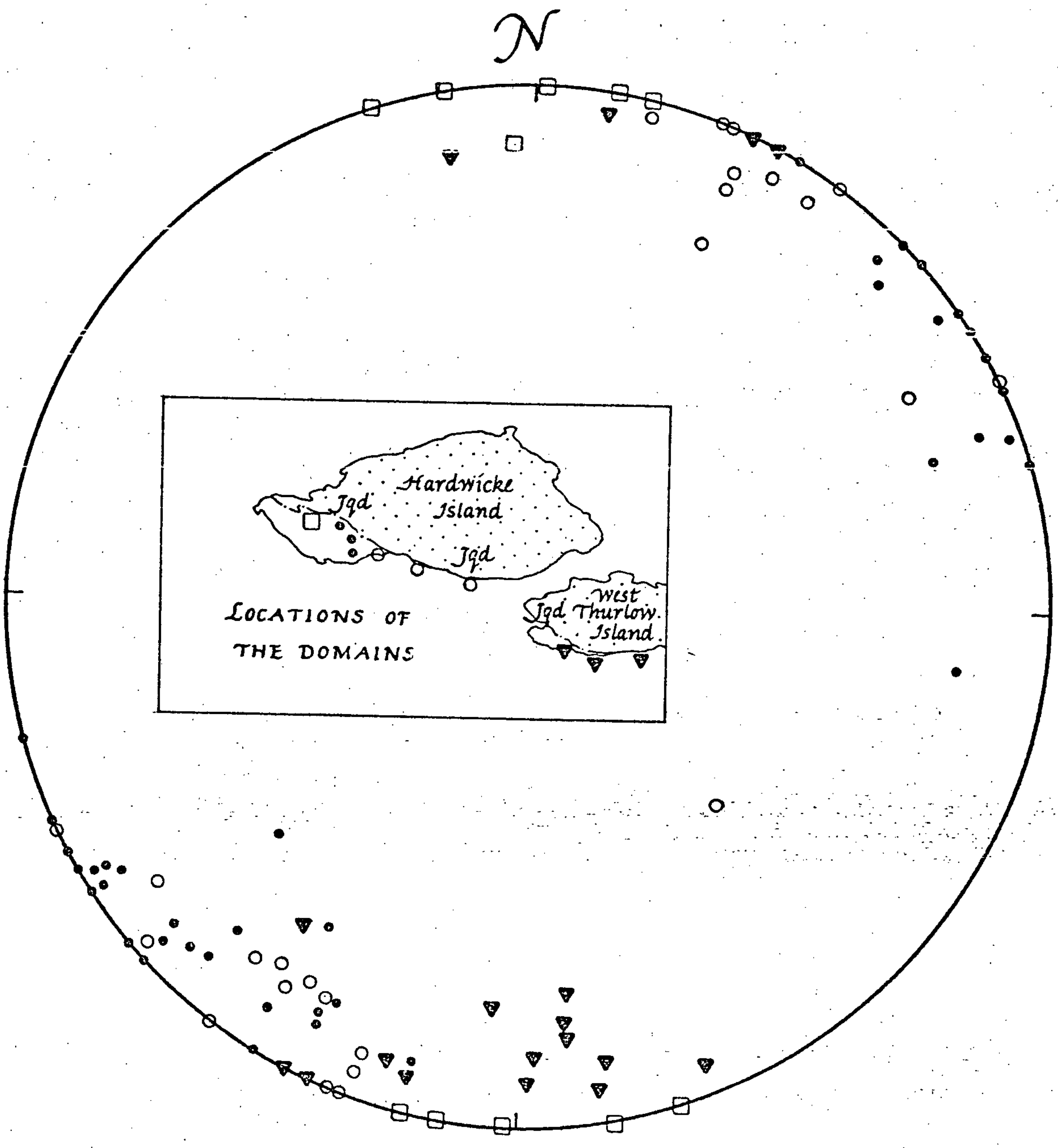
will be drag folds verging away from the pluton. Since the sense of vergence is by convention measured looking down the fold axis, there are two populations of opposite vergences lying in the slip plane, separated by a vertical slip vector.

Clearly the distribution in Figure 27a does not represent a simple case of vertical shear. There are not two distinct populations of clockwise and counter-clockwise vergence, nor do the fold axes lie in the slip plane. Instead they cluster in a subvertical array. Shearing between pluton and country rock may thus have been minor with compression perpendicular to the contact producing the chaotic folds. On the other hand, the steepness of the fold axes suggests extensive flow in a vertical direction.

Near Edwards' Homestead on Hardwicke Island, going upsection from Harbledown to Bonanza and towards the quartz diorite, bedding steepens to vertical and the moderately dipping axial planes in the sediments are replaced by bedding-parallel foliation in the metavolcanics. Foliation is extensively developed in Bonanza volcanics and sediments in the inner aureole, in marbles, and in places in the Karmutsen within a few meters of intrusive contacts. At Miner's Bay, 500 meters from the intrusion, mineral foliation in the marble is considerably steeper than bedding; by contrast the inner aureole (less than 300 - 500 meters from the intrusion) foliation (defined by mineral orientation; flattening of breccia clasts) and bedding (defined by grainsize and textural variations) are parallel and nearly vertical. The strike of the foliation ranges from $70^{o}$ to $160^{o}$ (Figure 28). Somewhat greater

Fig. 28

Poles to foliation in metavolcanics and marbles



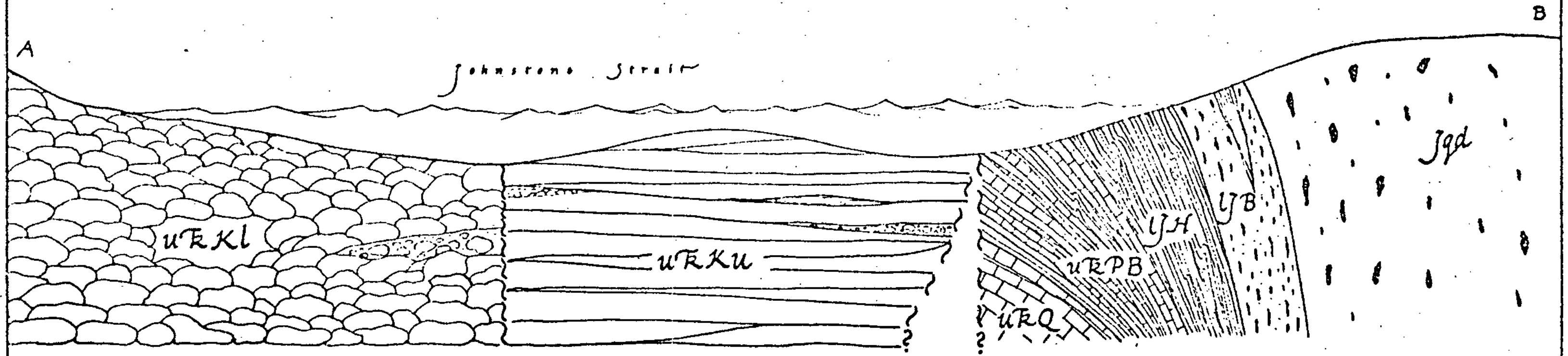
strike consistency is seen in the following areas: southern West Thurlow Island; Hardwicke Island east of Edwards' Homestead; between Edwards' Homestead and Pump Creek; and near Pump Creek. The foliation in these areas roughly parallels the local trend of the contact (inset Figure 28). This pattern resembles the accommodation of an incompetent matrix to competent members during deformation. However the newly intruded pluton while it supplied heat to the recrystallizing metavolcanics was anything but competent; it was still only partly solidified (see page 107). The parallelism of foliation and intrusive contacts may be best explained by considering the intrusion itself to be the source of the deformation. Since the chaotic folds may be interpreted as reflecting compression rather than shear between pluton and country rocks, the deformation may be due to expansion of the intrusion perpendicular to its walls.

iv. Summary

The second phase structures have been linked in timing and orientation to the main quartz diorite intrusion. The relationships are shown in the structural cross-section, Figure 29. Pre-plutonic dikes were emplaced in an unstressed environment, while contact metamorphism proceeded under directed stress. East of Vansittart Point, dikes involved in chaotic folding are traceable to the early gabbroic phase of the intrusion (Chapter III). The intensity of deformation increases towards the pluton. The development of near-vertical foliation in metavolcanics and marbles

Fig. 29

Structural cross-section, Hkusam to Hardwicke Island (perpendicular to intrusive contact)




lJB • Bonanza Volcanics.

lJH • Harbledown Fm.

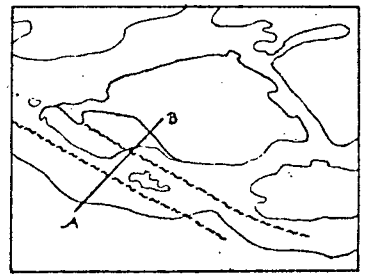
uTrPB • Parson Bay Fm.

uTrQ • Quatsino Fm.

uTrKu • upper Karmutsen Fm.

uTrKl • lower Karmutsen Fm.

Jqd • quartz diorite



may have been caused by outward expansion of the pluton. Clearly, much of the deformation in the map area is the result of forcible intrusion.

# CHAPTER V: METAMORPHISM

Three metamorphic episodes can be distinguished in the map area. The first, in the Vancouver Group, is due to burial; the second is contact metamorphism resulting from emplacement of the quartz diorite; the third, discussed in Chapter VI, encompases low-temperature alteration of the intrusive rocks and post-plutonic dikes as well as the older rocks.

One of the objects of this study has been to determine the maximum depth of burial of the Insular Belt-Coast Plutonic Complex boundary, and the maximum metamorphic temperatures achieved near it. Contact assemblages, in both basic and impure carbonate rocks, have been used. Carbonates are restricted to the innermost aureole of the quartz diorite where their parageneses are useful in establishing maximum contact temperatures. Basic rocks are ubiquitous. They include Karmutsen basalts, Bonanza intermediate volcanics, and pre-plutonic dikes of probable Bonanza affinity. Three metamorphic zones are recognized in them: albite-epidote-actinolite (greenschist), plagioclase-amphibole-epidote-chlorite (greenschist-hornblende hornfels-transitional), and plagioclase-hornblende (hornblende hornfels).

## i. Burial Metamorphism

Burial metamorphism has been overprinted by contact effects except in the area around Kelsey Bay which lies outside or nearly outside the aureole. Pillows in the Kelsey Bay Formation there show, microscopically, perfect preservation of igneous textures.

Secondary minerals generally do not transgress grain boundaries of the original phases. Clinopyroxene phenocrysts are unaltered. Plagioclase phenocrysts are thoroughly albitized and are full of chlorite and pumpellyite microlites. These secondary ferromagnesium minerals, within crystals more than two millimeters across, indicate considerable iron and magnesium mobility and also that the albitization process was more complex than

$$CaAl_2Si_2O_8 + Na^+ + Si^{4+} = NaAlSi_3O_8 + Ca^{2+} + Al^{3+}$$

(see Kuniyoshi 1972, 1972b; Surdam 1973).

Voids between pillows are filled with epidote, ferrian clinozoisite, quartz, calcite, and pumpellyite. The tiny irregular vesicles contain chlorite. Epidote-chlorite-quartz veinlets, while common, are fine and not as prominent as the vein-networks of the albite-epidote-actinolite zone (outer aureole).

The burial metamorphic assemblage near Kelsey Bay is epidote-albite-chlorite-pumpellyite-quartz-calcite-ferrian clinozoisite. The absence of actionolite defines it as sub-greenschist. According to Kuniyoshi (1972) actinolite forms at approximately $400^{\circ}C$. The minimum temperature for this assemblage is defined by the reaction analcime + quartz = albite + $H_2O$ which passes through $200^{\circ}$, 2 kb. $PH_2O$ and $196^{\circ}$, 3 kb. $PH_2O$ (Liou 1971). Kuniyoshi (1972) estimates the temperature of regional metamorphism in this to be approximately $350^{\circ}$, near the high-temperature limit of the prehnite-pumpellyite facies.

Further within the aureole the effects of burial metamorphism are obscured except that ferromagnesian inclusions in original plagioclase phenocrysts persist to the highest temperatures. These microlites, upgraded to amphibole, serve to identify their host feldspars as primary.

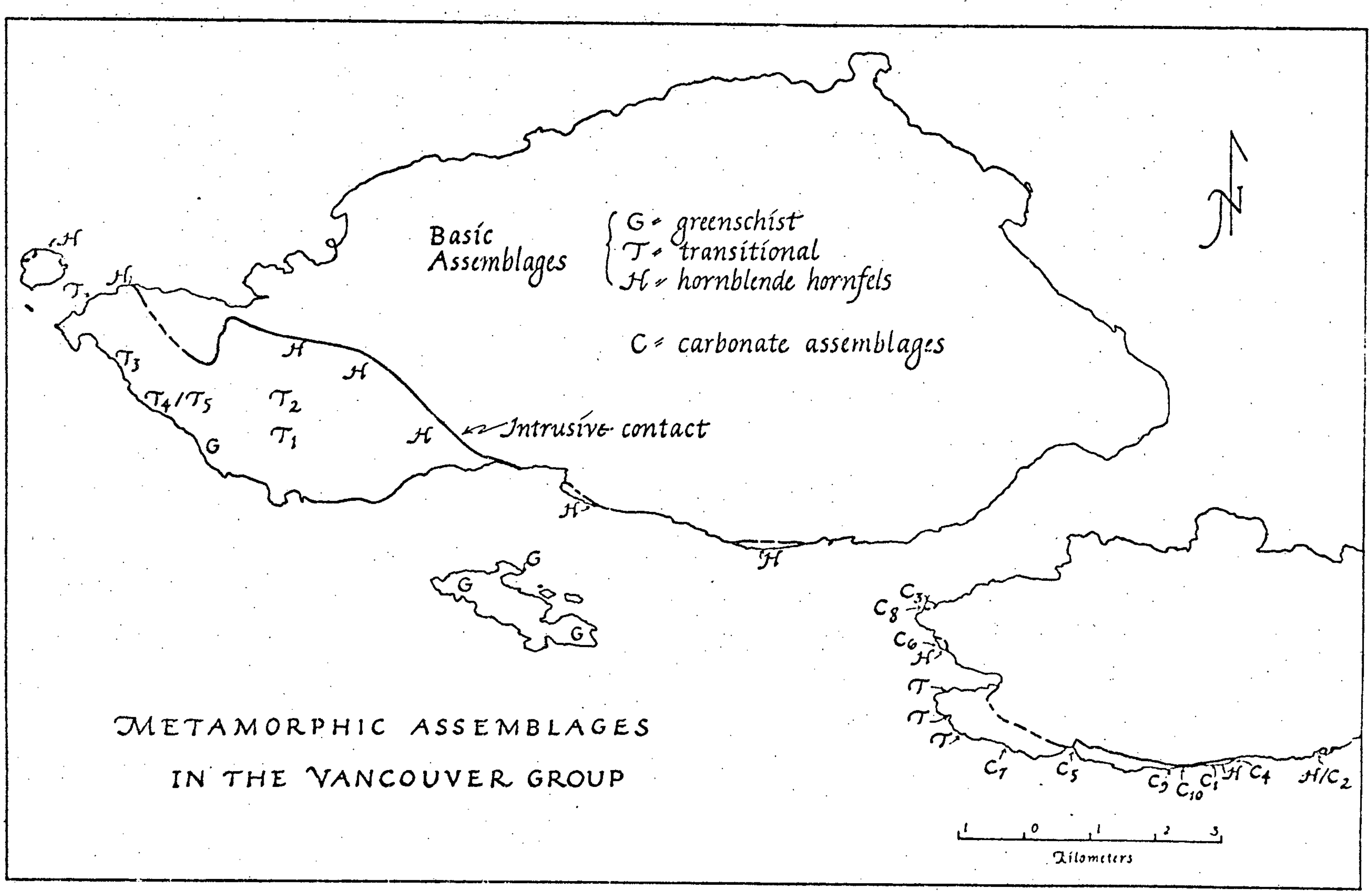
ii. Contact Metamorphism

a) Assemblages in Basic Rocks

Mineral assemblages of the greenschist facies are found on Helmcken Island and on Hardwicke Island near Bendickson Harbor (Figure 30). Albite, epidote, actinolite, chlorite and quartz constitute the essential matrix assemblage in this zone; in addition magnetite, sphene, pyrite and calcite may be present. Vesicles contain epidote, quartz and chlorite (Figure 31a). The chlorite, probably penninite, is nearly isotropic with strong anomalous blue birefringence.

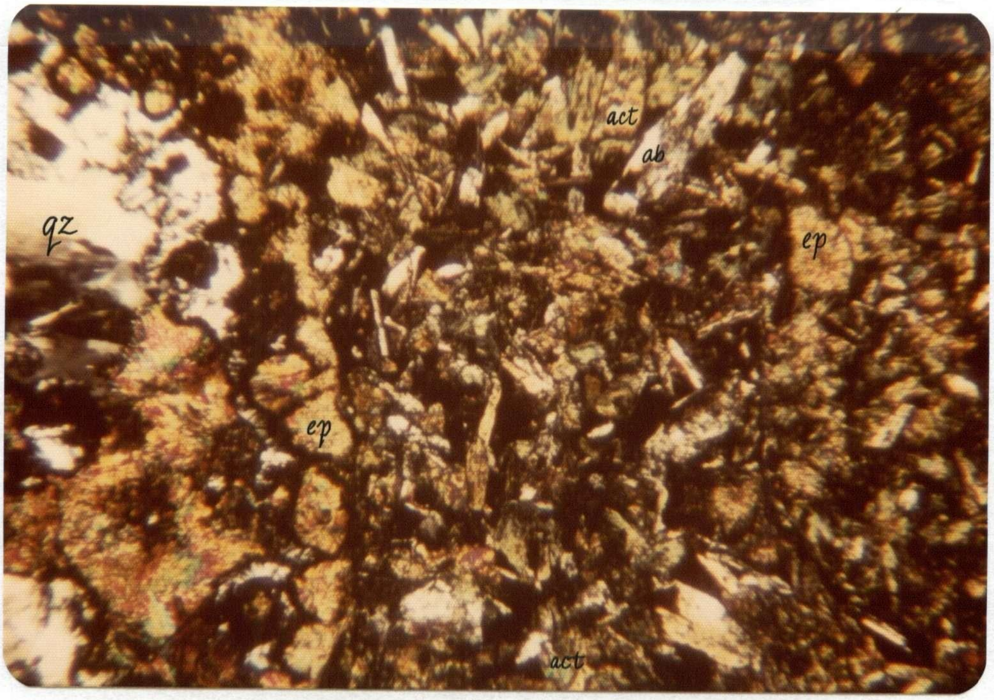
Epidote veins and zones of epidote enrichment are prominent features of this zone. In thin section many are seen to be granular concentrations that replace and grade into the surrounding rock. Some veins follow small faults which displace flow-tops or dikes. Although vein orientations are variable they tend to follow the west-northwest structural grain. Most are vertical or nearly so. The areas of epidote enrichment either are in pods and patches or follow the highly amygdaloidal zones near flow-tops. Microscopically the epidote enrichment consists of two effects - riddling with epidote veinlets and the growth of up to 50% granular epidote in the matrix. Albite and opaque minerals are entirely lacking in these zones. Actinolite and chlorite persist in small quantities. Amygdules contain quartz,

Fig. 30



Basic Assemblages
G = greenschist
T = transitional
H = hornblende hornfels
C = carbonate assemblages
Intrusive contact
METAMORPHIC ASSEMBLAGES
IN THE VANCOUVER GROUP
Kilometers

Fig. 31

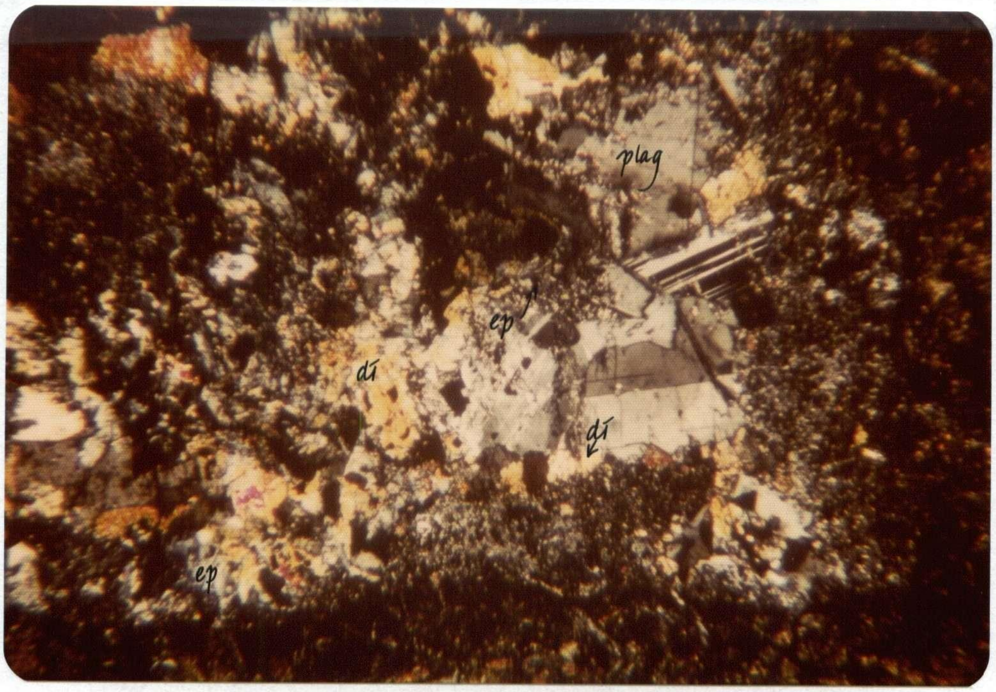



a) Karmutsen greenschist with amygdule, Helmcken Island. (×50)

Matrix: albitized plagioclase laths, actinolite (note ragged terminations), epidote, chlorite, interstitial opaque phase (magnetite-sphene?)

Amygdule: radiating epidote along rim, radiating quartz inside.

Note non-directionality of matrix. 7.21.4.B




b) Greenschist – hornblende hornfels transition in amygdule in upper Karmutsen, West Thurlow Island (×50).

Plagioclase and diopside replace epidote and quartz. The speckled zone around the rim consists of fine-grained epidote surrounded by plagioclase. The clear plagioclases in the centre are after quartz. Compare epidote habits with Figure 31a. 7.8.3

epidote and minor calcite and pyrite. Chlorite, an important amygdular phase elsewhere, is absent. Epidote-rich zones grade into normal albite-epidote-actinolite-chlorite greenschist.

The concentration of epidote in faults, fractures and flowtops suggests a metasomatic process. Vertical faults and fractures form likely passageways for migrating fluids during dehydration. The highly amygdaloidal flowtops must have been relatively permeable, acting as catchment zones sealed above by dense, nearly unvesicular flow bottoms.

Jolly and Smith (1972) have studied a similar pattern of epidote segration in the lower Portage Lake basalts of northern Michigan. As in the present area, they found amygdaloidal flowtops and fractures to be so enriched in epidote as to form "epidote metadomains". Jolly and Smith model flowtops and fractures as permeable zones. During dehydration water would escape from them more readily than from massive basalt. In the epidote metadomains the condition $P_{H_2O} < P_{load}$ would prevail, tending to stablize the relatively anhydrous mineral epidote. Jolly and Smith's analyses for water contents tend to substantiate this model. Unaltered Portage Lake basalts contain an average of 3.8 wt.% water; albite basalt 2.7% and epidote metadomains 1.1%.

The possibility that fluid pressure was less than load pressure in parts of the Karmutsen basalts during greenschist metamorphism affects the application of experimental data to reactions occurring in them as a means of determining depth of burial. A more detailed comparison between conditions in the

Portage Lake Series and the Karmutsen Group is therefore appropriate. Epidote metadomains in the Portage Lake basalts are associated with uppermost zeolite assemblages. In the Karmutsen they occur with the greenschist assemblage albite-epidote-actinolite-chlorite-magnetite-sphene-quartz which is itself relatively anhydrous. Liou et al. (1974) obtained a water content of 1.61 wt. % from a greenschist containing 24% chlorite. Greenschists from Helmcken and southern Hardwicke Islands generally contain less than 10% chlorite; their water contents may approach those of the epidote metadomains. If, as Jolly and Smith propose, present water contents are reliable indicators of relative $P_{H_2O}/P_{load}$ it can be argued that differences between fluid and load pressure in Karmutsen basalts were small. Nevertheless, water-pressure gradients provide an attractive explanation for epidote segregation. Interpretation of parageneses is therefore subject to a dual uncertainty: the exact value of $P_{H_2O}/P_{load}$ in a particular rock, and the effect of this ratio on the reactions observed in it.

Less than 1250 meters from the quartz diorite, greenschist assemblages give way to greenschist-amphibolite transitional assemblages (Figure 30). At the outer limit of the transition anorthite contents in plagioclase increase sharply from near zero to greater than 20%. Epidote, quartz, and chlorite are replaced by idioblastic amphibole and new plagioclase (Figure 31b). Blue-green hornblende replaces or grows side-by-side with pale green actinolite. Quartz in amygdules decreases and finally disappears. The inner limit of the transition zone, approximately 500 meters from the intrusion, is marked by the disappearance of epidote and chlorite.

The significant transitional assemblages observed in rocks from Hardwicke and West Thurlow Islands are the following: (see Figure 30 for locations)

$T_1$ : plag($An_{40}$)-act-hb-mt-ep-qz-ap-py-sphene-chl (andesite dike)

$T_2$ : plag($An_{40}$)-actinolitic hornblende-chl-bi(5%)-mt-py-qz (Bonanza)

$T_3$ : plag(An 20)-hb-chl-ep-qz-mt-py (Karmutsen)

$T_4$ : plag(An 20)-hb-act-ep-mt (Karmutsen)

$T_5$ : plag (An30)-hb-act-chl-ep-mt-ilm; in vein, ep-qz-ga (Karmutsen)

(plag=plagioclase, act=actinolite, hb=hornblende, mt=magnetite, ep=epidote, qz=quartz, ap=apite, py=pyrite, chl=chlorite, bi=biotite, ilm=ilmenite, ga=garnett)

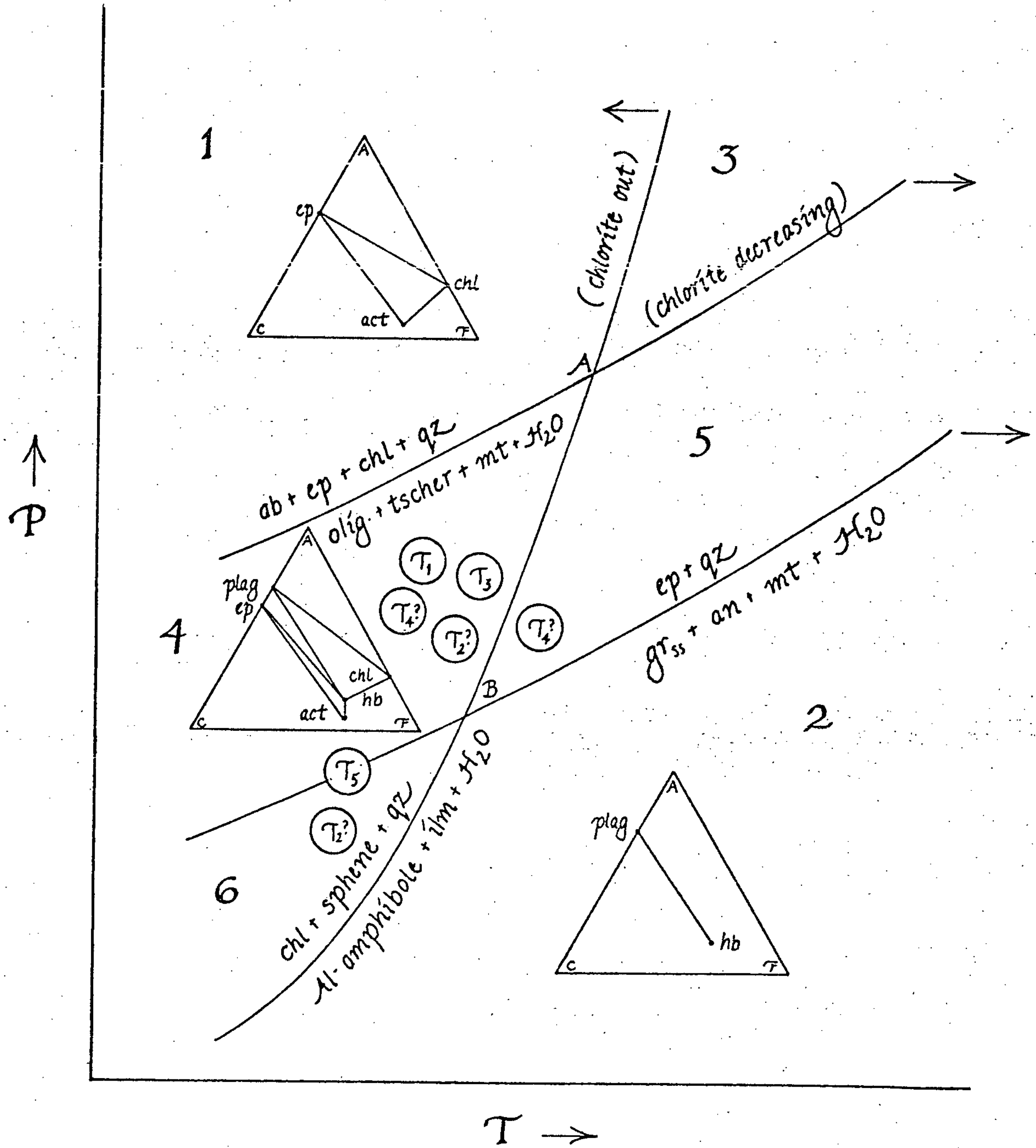
The transition from greenschist to amphibolite in Karmutsen basalts has been investigated experimentally by Liou et al. (1974) at $P_{H_2O} = P_{total}$ and oxygen fugacities defined by the quartz-fayalite-magnetite buffer. They consider the transition to be bounded by two reactions:

1) chl + ep + ab + qz = olig + tschermakite + $Fe_3O_4$ + $H_2O$ and

2. chl + sphene + qz = Al-amphibole + ilm + $H_2O$ (upper). The first involves a decrease in chlorite with rising temperature; the second its elimination. These reactions are shown schematically in Figure 32 along with the reaction ep + qz = $\text{grandite}_{ss}$ + an + mt (Liou 1973) which may occur in epidote-quartz veins. Field 1 is greenschist; Field 2 is amphibolite. The transitional fields 3 to 6 are of greatest interest. Field 3 is albite-epidote amphi-

Fig. 32

Schematic diagram of the greenschist-amphibolite transition (After Liou et al 1974)



bolite. In Field 4 transitional assemblages and epidote-quartz segregations coexist; in Field 5 amphibolites may contain epidote-quartz veins. Field 6 includes chlorite-bearing assemblages in which epidote-quartz is unstable.

Because of the differing slopes of the chlorite-decreasing and chlorite-out curves greenschist-amphibolite transitional assemblages can be used as pressure indicators. On Hardwicke and West Thurlow Islands albite amphibolite is conspicuously absent; plagioclase replaces albite at the outer border of the transition zone. Pressure certainly did not exceed that at point A on Figure 32.

Epidote + quartz is a stable vein and vesicle assemblage in $T_1$ and $T_3$. They therefore lie in Field 4. The reaction ep + qz = $gr_{ss}$ = an + mt + $H_2O$ is inferred in $T_5$, as garnet occurs in contact with epidote and quartz in the wider veins. (In veinlets less than 1 mm. across amphibole has replaced ep-qz, indicating exchange with the matrix over short distances.) Epidote and chlorite in the matrix of this rock are sparse and texturally unstable. Epidote in vesicles is embayed and engulfed by amphibole. Small patches of chlorite are transected by idioblastic amphibole crystals.

$T_2$ and $T_4$ are more difficult to interpret. $T_2$ is a Bonanza basaltic andesite containing no epidote. It might for that reason fit in Field 6. However the ACF diagram approximating assemblages in Field 4 shows that epidote may be eliminated early in Al-rich, Ca-poor rocks. Typical Bonanza volcanics contain far less calcium than Karmutsen basalts; some contain appreciably more aluminum (Muller et al 1974 Tables 2 and 4). The position of $T_2$ is uncertain. It is shown in both Fields 2 and 6. $T_4$ contains epidote but no chlorite. These characteristics place it in Field 5. However $T_4$ is highly amygdaloidal and contains enough epidote to be an epidote metadomain, which as described above contain little or no chlorite even in greenschist facies. The close spatial association of $T_3$ and $T_4$ suggest that $T_4$ belongs in Field 4.

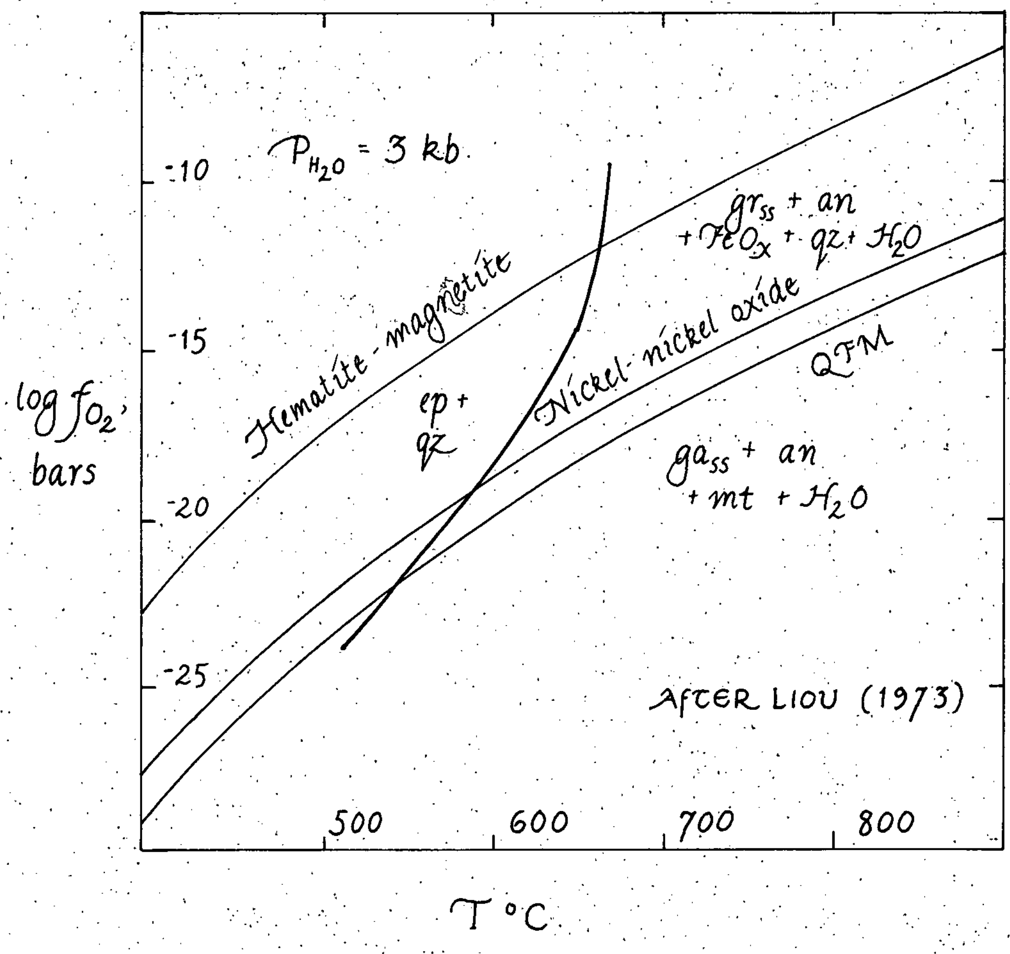
The above assemblages all lie below point A. $T_5$ and possibly $T_2$ indicate pressures below point B. Putting a numerical value on the pressures at points A and B requires a knowledge of the oxygen fugacities in the rocks during their metamorphism. The positions of the curves, which involve oxidation-reduction reactions, and thus of their intersections, are very sensitive to oxygen fugacity. The displacement vectors shown on Figure 32 for increasing $f_{O_2}$ shift the intersection points to lower pressure (Liou et al 1974). The value of $f_{O_2}$ in Karmutsen metabasalts is uncertain. Liou et al take the ubiquity of magnetite + quartz in many aureoles to signify $f_{O_2}$ higher than that defined by the QFM buffer. This association is commonly observed here. However, small amounts of titanium in magnetite tend to stabilize it relative to fayalite (I. Duncan, pers. comm. 1974) rendering quartz-magnetite an unreliable indicator of oxygen fugacity. The application of the magnetite-ilmenite geothermometer (Buddington and Lindsley 1964) to Karmutsen metabasalts would prove a useful study.

While $f_{O_2}$ greater than QFM cannot be demonstrated, $f_{O_2}$ less than QFM is unlikely. In natural systems oxygen fugacity is controlled by the phases and reactions present. The breakdown of epidote involves reduction of $Fe^{3+}$ to $Fe^{2+}$. It has a positive slope on a temperature-$F_{O_2}$ diagram (Liou 1973): as temperature increased the reaction would move through increasing oxygen fugacities (Figure 33). The chlorite-decreasing reaction would behave similarly. It is possible, then, that the greenschist-amphibolite transition itself acted as an oxygen buffer and that $f_{O_2}$ increased with rising temperature. With the QFM buffer point B lies slightly above 3 kb (Liou et al 1974). This provides a maximum estimate for load pressure (hydrostatic) during contact metamorphism. The inner limit of the transition is about 500 meters from the quartz diorite. The temperature here, assuming QFM conditions, was 550° (Liou et al 1974).

Basic assemblages typical of the hornblende-hornfels facies are developed in Karmutsen and Bonanza volcanics less than 500 meters from the intrusion. Textures in the Karmutsen are typically granoblastic, or sublepidoblastic if dominated by hornblende. This hornblende lacks the ragged (001) terminations of greenschist and transition zone amphiboles. It tends to form straight, rational boundaries, particularily the prism faces, although polygonal grains with 120° intersections are not uncommon. Hornblende colors vary from brownish green to brown. Rims and new crystals of prograde texture are browner than the amphibole which they supercede. Within a few centimeters of the quartz diorite the hornblende in the Karmutsen is green, as is that in the intrusion

Fig. 33

Epidote breakdown, T - $f_{O_2}$ space



(Figure 34). Compositional equilibrium between magma and country rock probably obtained over a short distance. Poikilitic hornblende prophyroblasts are developed in the innermost aureole. Intergrown magnetite and ilmenite occur as scattered round grains or clumps inside hornblende. Patterns of opaque microlites in hornblende may be relics of the destruction of igneous clinopyroxene. Poikilitic metamorphic clinopyroxene occurs in some parageneses. It seems to concentrate in and around amygdules. It apparently reflects exceptionally calcic composition due to local abundance of epidote. Biotite and cummingtonite porphyroblasts have developed in the Karmutsen metabasalts only adjacent to intrusive contacts (Figure 34). Like green hornblende they are probably the result of small-scale metasomatism.

Plagioclase compositions in Karmutsen rocks of this zone, determined from extinction angles, are $An_{40}$ to $An_{80}$. Their zoning patterns are normal, patchy, or irregular. Except rarely in amygdules plagioclase does not form porphyroblasts. On the contrary, phenocrysts tend to recrystallize to small, xenoblastic grains. New crystals create a polygonal granoblastic fabric which bears no resemblance either to the mat of plagioclase laths in the original rock or to the euhedral to subhedral strongly normal-zoned plagioclases in the peripheral gabbro phase of the main intrusion. Polygonal fabric is retained even within 2 mm. of intrusive contacts. Plagioclase compositions next to the quartz diorite, ranging from $An_{36}$ to $An_{48}$ in strongly normal-zoned

Fig. 34

x 50

Contact between Late Jurassic quartz diorite and Karmutsen hornfels, Hardwicke Island.

The large plagioclases at the top of the photograph are part of the quartz diorite.
Note the green hornblende and the large skeletal biotite (labelled) growing around small plagioclases in the hornfels. The hornfels has a polygonal texture and a faint foliation not visible in the photograph. Plagioclase is strongly normal zoned.

7.28.3

crystals, are equivalent to and probably in equilibrium with compositions in the intrusion.

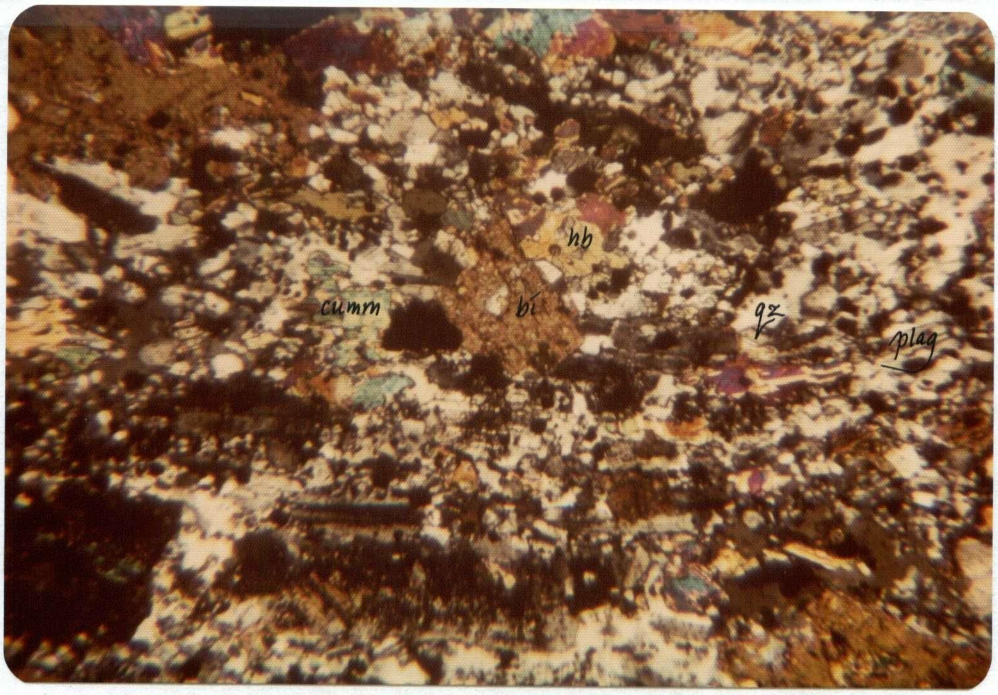
Bonanza metavolcanics, unlike Karmutsen rocks, are foliated everywhere within the hornblende-plagioclase zone. Bonanza hornblende is green. If acicular it lies in the foliation (Figure 35a). Its most common habit is xenoblastic to subidioblastic and equant. Such grains, although lying in compositional bands, are not individually oriented.

As in the Karmutsen, large plagioclase grains are blastoporphyritic. In thin section they are seen to contain patches or symmetrical zones of amphibole microlites, a probable relic of burial metamorphism. The plagioclases are randomly oriented and commonly broken and recrystallized. The foliation bends around them (Figure 35a). In the outer part of the hornblende-plagioclase zone some are still oscillatory-zoned. This is a relict, not a prograde feature, absent closer to the intrusion. Stable metamorphic zoning modes are normal, patchy, or irregular. Anorthite contents of metamorphic plagioclase range from $An_{28}$ to $An_{70}$. This variability probably reflects differences in bulk composition; there is no correlation with distance from the intrusion. New plagioclases are small round or polygonal grains, or subidioblastic laths lying in the foliation.

Within 70 meters of plutonic contacts the growth of large poikiloblastic to skeletal biotite, hornblende, and cummingtonite partly obscures the foliation (Figure 35b). Cummingtonite is noteworthy in that it has not been previously reported from Insular Belt contact aureoles. It is colorless to slightly green

a) Bonanza tuff (?) in hornblende hornfels zone ( x50 ) 7.25.9A

The foliation, defined by biotite and actinolite, bends sharply around plagioclase porphyroclasts, which tend to lie with their long axes in the foliation. The large plagioclase in the centre shows relict oscillatory zoning. Actinolite needles grow in all plagioclase porphyroclasts. They concentrate in the core of the one at the top of the photograph, where albitization would have been most severe.




b) Bonanza hornfels ( x 50 )

Large biotite, hornblende and cummingtonite grow across a foliation defined by plagioclase laths. The large plagioclases have embayed borders and are in the process of recrystallizing to small xenoblastic grains.

7.19.5

with numerous lamellar twins. Two sets of green pleochroic lamellae, presumably of calcic amphibole, are discernible. The combination of large positive 2V and relief similar to hornblende serve to distinguish cummingtonite from clinopyroxene.

Shido (1958) notes the presence of cummingtonite in basic rocks from contact aureoles in the Abukuma Plateau. She considers it a low pressure phase for two reasons. The first is its association with other low-pressure indicators, andalusite in pelitic hornfelses and the actinolite-plagioclase transitional assemblage in metavolcanic rocks. Secondly, a likely reaction for its formation is:

$$7 \text{ tscher} + 10 \text{ qz} = 14 \text{ an} + 3 \text{ cumm} + 4 H_2O$$

There is a considerable volume increase from left to right (2044 vs. 2221 cc. for the solid phases). (Note that abundant free quartz favors cummingtonite development, accounting for its presence in Bonanza but not Karmutsen hornfelses.) The presence of cummingtonite corroborates the low-pressure condition inferred above from greenschist-amphibolite transitional parageneses.

Several textural features of Karmutsen and Bonanza metavolcanics in the hornblende-plagioclase zone deserve reiteration as they bear on the origin of the plutonic rocks, particularily the peripheral gabbros. Rocks in this zone all have polygonal granoblastic or lepidoblastic fabrics (Figures 34, 35). Near the plutonic contacts amphiboles and biotite develop as large poikiloblastic grains which appear to postdate the foliation.

Plagioclase exhibits no such tendency. It typically forms as small xenoblastic to subidioblastic grains with simple zoning and twinning patterns. It is distinguishable from the plagioclase in the peripheral gabbros, and bears no resemblance to that in the quartz diorite. Convincing arguments for dioritization based on transitional metavolcanic-plutonic textures cannot be made.

b) Carbonate Metamorphic Assemblages

Impure carbonate layers are common in the intravolcanic sediments of the upper Karmutsen Formation and in the Quatsino and Parson Bay Formations on West Thurlow Island. Carbonates also occur on Yorke Island and Hardwicke Island. Fanny Island is an isolated marble block surrounded by gabbro. Metamorphic minerals in the impure carbonates include calcite, quartz, wollastonite, clinozoisite, epidote, garnet, diopside, idocrase, sphene, and plagioclase. A number of carbonate parageneses were found to be useful in determining temperature and $P_{CO_2}/P_{total}$. They can be approximated by parageneses in the system $CaO - Al_2O_3 - SiO_2 - H_2O - CO_2$ investigated by Gordon and Greenwood (1971) and Storre and Nitsch (1972). They are listed in Table 3 and plotted on a simplified version of Gordon and Greenwood's phase diagram (Figure 36). Total pressure in their experiments was 2 kb., close to the estimated total pressure in the aureole.

It will be noted that several plotted assemblages either lie on or bracket reaction boundaries. The most significant of these is the reaction grossular + quartz = anorthite + wollastonite. As

## Table 3 = Carbonate Parageneses

$C_1$. plag - ga - qz  plag = $An_{65}$  ga = $gros_{50}$ $andr_{50}$
(di-sphene-ap)  (refractive index)

$C_2$. ga - plag (An 28-36) - qz; ga - ep (layered)

$C_3$. ga - wo - plag
(sphene - py - di)

$C_4$. plag ($An_{68}$) - wo - ga
(sphene - py - di)

$C_5$. cc - wo - qz

$C_6$. cc - wo - qz - ga (ga & qz not touching.qz in wo)

$C_7$. cc - qz - wo - ep - ga

$C_8$. cc - czt - plag; cc - wo - ga (idocrase - di) (layered)

$C_9$. wo - ga (idocrase - di)

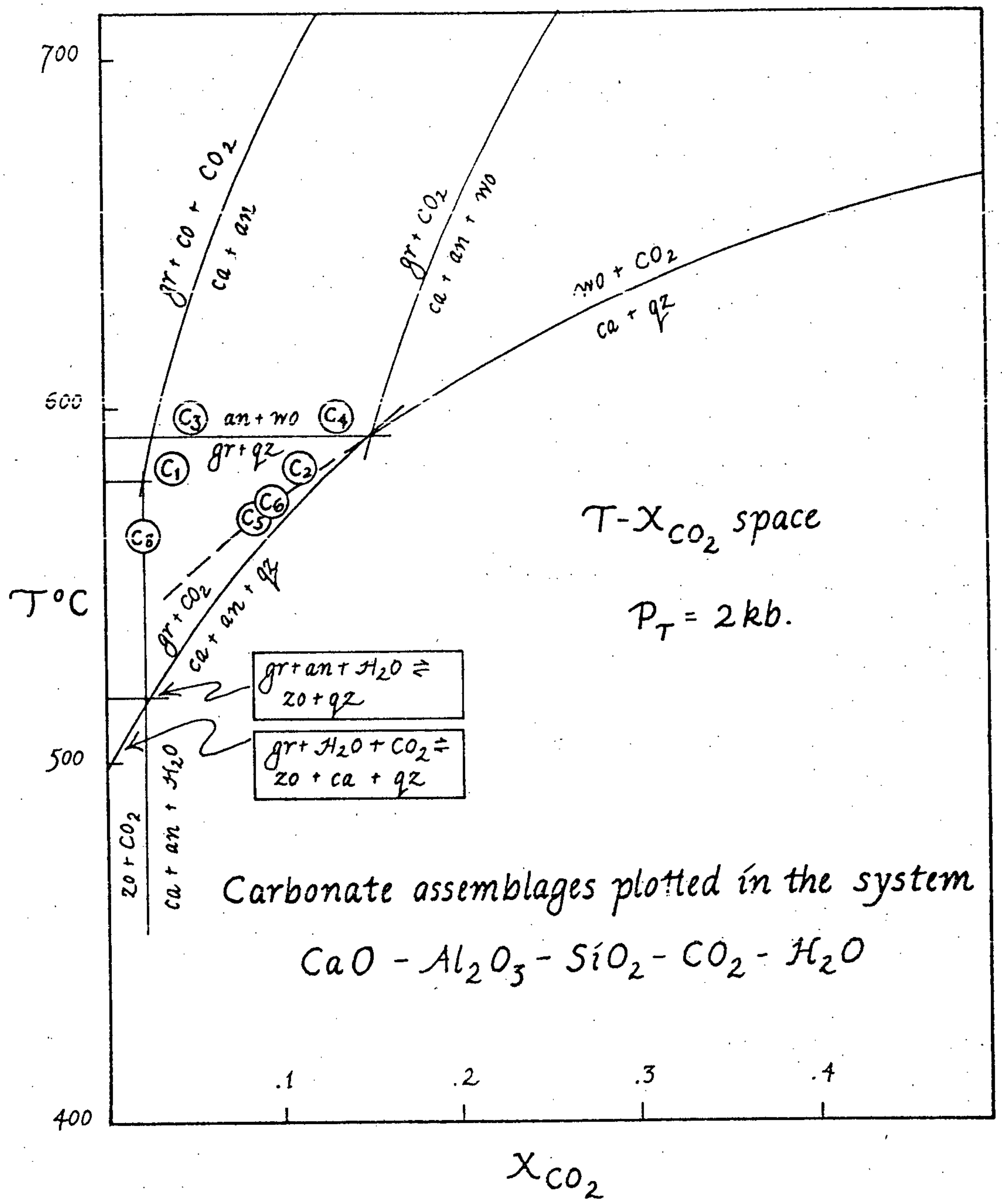
not diagnostic

$C_{10}$. cc - wo - ga (di)

### Abreviations

| | |
|---|---|
| plag | plagioclase |
| ga | garnet |
| qz | quartz |
| di | diopside |
| ap | apatite |
| ep | epidote |
| wo | wollastonite |
| y | pyrite |
| cc | calcite |
| czt | clinozoisite |

Fig. 36




Gordon and Greenwood 1971

$zo + CO_2 \rightleftharpoons an + ca + H_2O$ from Storre and Nitsch 1972 only solid phases are involved the position of the curve does not change with fludi composition: the temperature of the reaction is a function of total pressure alone. The assemblages which bracket this reaction were all collected with 30 meters of the main intrusion; therefore it can be used to estimate the maximum temperature in the country rock next to the contact. On Gordon and Greenwood's diagram for 2 kb. total pressure the reaction temperature is slightly less than 600°C. The reaction boundaries depicted on Figure 36 are not quantitatively applicable to the assemblages at hand. It is evident from the phases present (i.e. diopside, epidote, idocrase, plagioclase instead of anorthite, and grandite solid-solution instead of grossular) that Fe, Mg, and Na are present in addition tothe five components of Gordon and Greenwood's system. To be useful, the positions of the reactions must be adjusted to account for mixing in the participating phases.

The displacements of the line gros + gz = an + wo due to andradite in the garnet and albite in the plagioclase can be calculated from the equilibrium constant if ideal mixing is assumed (Greenwood lecture 1974). The balanced reaction is:

gros + qz = an + 2 wo.

The equilbrium constant $K = \dfrac{(a_{an})(a_{wo})^2}{(a_{gros})(a_{qz})}$

The equation for this constant, using the data of Newton (1966), is:

$$\log_{10}K = \frac{3492.2}{T} + 4.4464 - \frac{.17060\ (P - 1)}{T}$$

or $$T^{\circ K} = \frac{3492.2 + .17060\ (P - 1)}{4.4464 - \log_{10} K}$$ (P is in bars)

In the assemblage $C_1$ $a_{an}$ = .65 and $a_{gros}$ = .25 therefore $\log_{10}K$ = .415. The reaction proceeds at 656.7°C. at 1.5 kb.; 677.8° at 2 kb.; and 720.2° at 3 kb. Temperatures at the contact, then, lay between 650 and 720°C.

If considerable $Fe_2O_3$ is added to the system $CaO - Al_2O_3 - SiO_2 - H_2O - CO_2$ the low-temperature limit of garnet stability at very low $X_{CO_2}$ is defined by:

2 ep + 2 cc + 3 wo = $\text{grandite}_{ss}$ + 2 $CO_2$ + 1$H_2O$ (Chatterjee 1966)

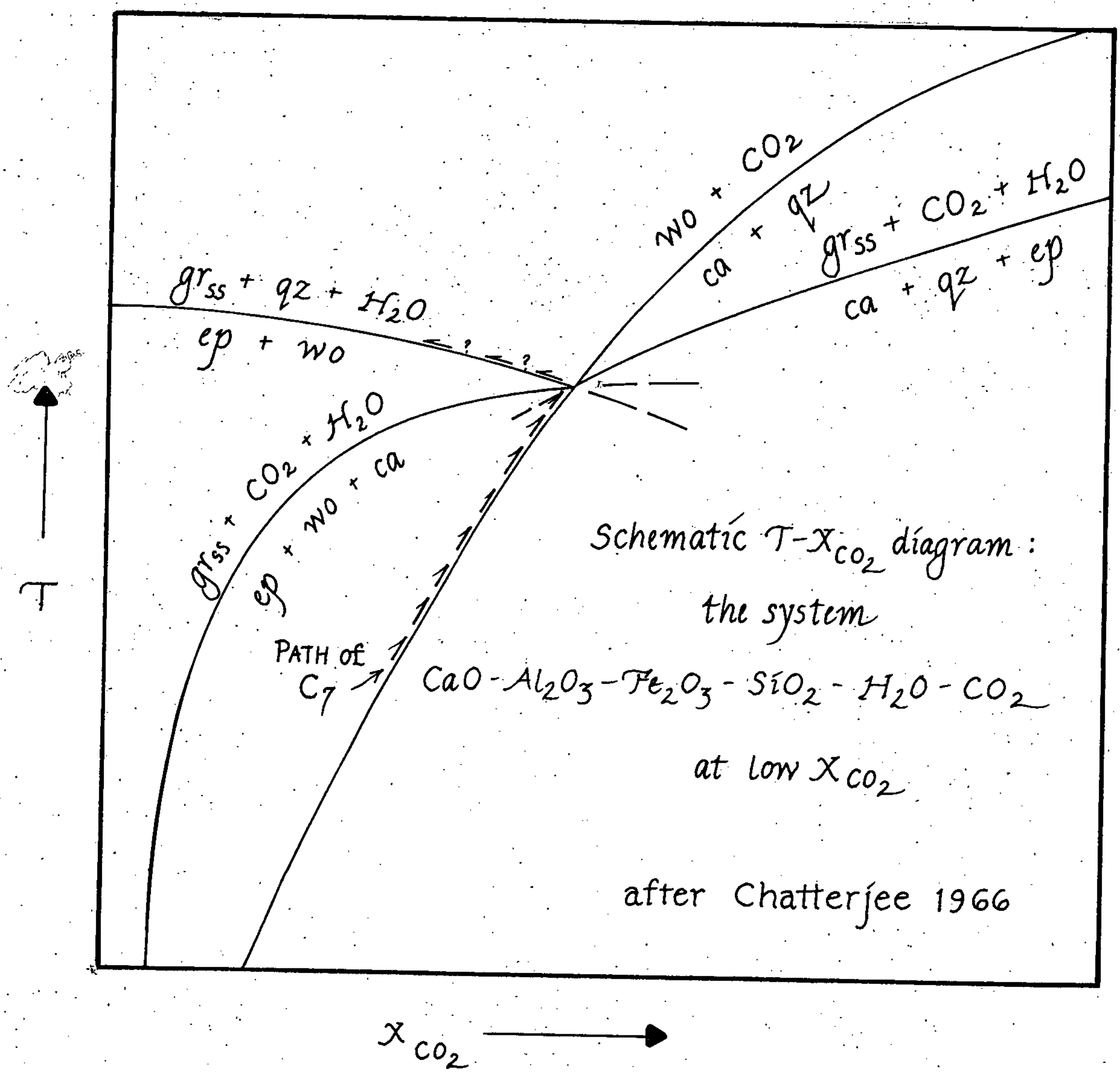
rather than by:

2 zo + 5 cc + 3 qz = 3 gros + 5 $CO_2$ + 2 $H_2O$.

Chatterjee's numerical data for grandite breakdown are not applicable because $f_{O_2}$ was unbuffered in his runs and $P_{total}$ was held at 1 kb. The geometries of his reactions are, however, interesting. They are reproduced on Figure 37 which shows schematically the effect of ferric iron on Gordon and Greenwood's system. In the iron-free system zoisite and wollastonite cannot coexist (Kerrick et al 1973). Epidote and wollastonite can, and are observed here.

The assemblage cc-qz-wo-ep-ga falls on an invariant point on Chatterjee's diagram. Calcite and quartz in it are in the process of reacting to form wollastonite, buffering $X_{CO_2}$ to the

Fig. 37




Schematic $T$-$X_{CO_2}$ diagram: the system $CaO-Al_2O_3-Fe_2O_3-SiO_2-H_2O-CO_2$ at low $X_{CO_2}$

after Chatterjee 1966

invariant point. the reactions cp + cc + wo = $Gr_{ss}$ + $CO_2$ + $H_2O$ and ep + wo = $gr_{ss}$ + gz + $H_2O$ would then have proceeded simultaneously until one reactant was used up. This pathway indicates an initially very water-rich fluid becoming somewhat less so with rising temperature. High $X_{H_2O}$ is also evident in the other hydrous phases present in the aureole - clinozoisite and idocrase- and in the presence of garnet which requires low $X_{CO_2}$ (Figure 36). A hydrous fluid phase favours the formation of wollastonite at relatively low temperatures; the first appearance of wollastonite here ( $C_7$ ) is approximately 500 meters from the intrusion where the temperature was near 550°C. The occurrence of the assemblage cc-qz-wo at several distances from the intrusion suggests possible buffering of $X_{CO_2}$ by this reaction. With all these indicators it is safe to say that $X_{H_2O}$ in most carbonates was at least 0.8.

c. Conclusions

This study of contact metamorphic assemblages has produced five quantitative results.They are as follows:

$P_{total} \leq$ 3 kb.

$X_{CO_2}$ in carbonates $\leq$ 0.2

Temperatures during contact metamorphism (maxima):

between 200° and 400°, probably close to 350° (Kuniyoshi 1972) 6 km from the main intrusion

550° 500 meters from the intrusion (upper limit of greenschist - amphibolite transition, QFM conditions)

650 - 720° at the contact ( $gr_{ss}$ + qz = an + wo )

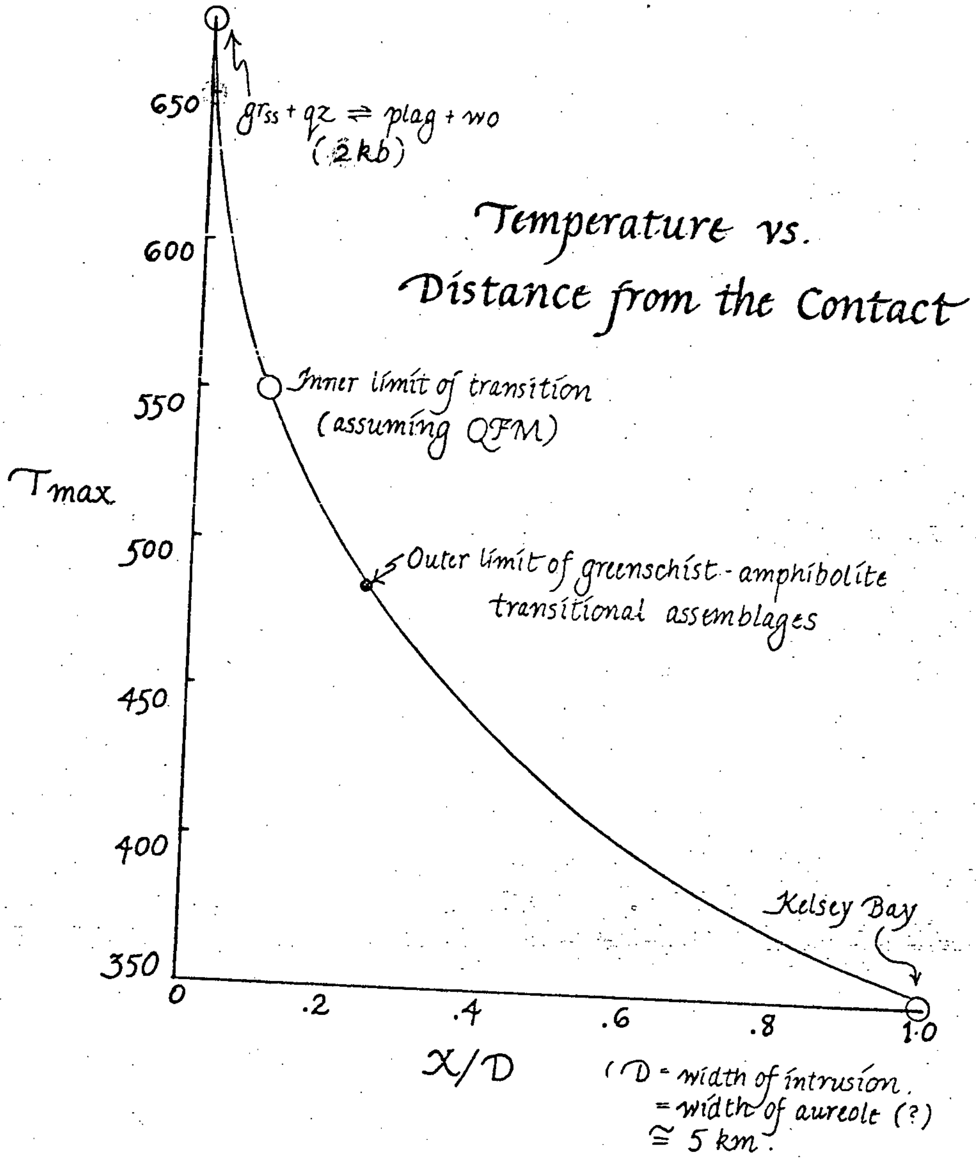
Since these results will be used to make important inferences a caveat should be offered before proceeding. They are subject to several uncertainties: oxygen fugacity, phase compositions, the assumption of ideal mixing, $P_{H_2O}/P_{load}$ in the metavolcanics and experimental uncertainties. The first two could be eliminated by extensive microprobe work.

The three temperatures determined in the aureole define a curve (Figure 38) like those computed by Jaeger (1959). A further restriction on pressure can be drawn from this temperature-distance curve. The greenschist-amphibolite transition commences 1.3 km. from the intrusion. The temperature at that point according to Figure 38 is less than or equal to $500^{\circ}C$. For QFM conditions such a situation obtains only if $P_{total} \leq 2.5$ kb. (Liou et al 1974).

The maximum stratigraphic load at the top of the Karmutsen Formation on Vancouver Island is 1.5 kb. If the maximum pressure in the present area did not exceed 2.5 kb. it is reasonably certain that we are dealing with the Vancouver Island stratigraphic column. Therefore here, at the edge of the Coast Plutonic Complex, neither depth of burial or depth of erosion have significantly exceeded conditions anywhere in the Insular Belt.

Jaeger's calculations allow the initial temperature of an intrusion to be inferred from the temperature at its contacts. It is assumed that intrusion is instantaneous. If further it is assumed that there is no latent heat of crystallization then the simple relation $( T_{contact} - T_{initial} ) = \frac{1}{2} ( T_{intrusion} - T_{initial} )$ holds, where the initial temperature is that of

Fig. 38


Temperature vs.
Distance from the Contact
Tmax
650
600
550
500
450
400
350
$gr_{ss}$ + qz ⇌ plag + wo
(2 kb)
Inner limit of transition
(assuming QFM)
Outer limit of greenschist-amphibolite
transitional assemblages
Kelsey Bay
0
.2
.4
.6
.8
1.0
X/D
(D = width of intrusion
= width of aureole (?)
≅ 5 km.

the country rocks before emplacement of the pluton. $T_{contact}$ is at most $700^{o}C$. The ambient temperature, using Kuniyoshi's estimate (1972), is $350^{o}$. $T_{intrusion}$ is then $1050^{o}$. If latent heat is allowed for it is approximately $1000^{o}$.

d) Structure and Metamorphic Textures

Pervasive deformation in the albite-epidote-actinolite zone is confined to close-spaced shear-planes in the fault zones. Stress-induced recrystallization has not occurred. Vesicles are equant; relict phenocrysts, even if bent or broken, retain their original form. Actinolite needles form a non-directional matte. These textures indicate sluggish recrystallization. Under greenschist metamorphism the Karmutsen Group behaved as competent and brittle blocks. Its response to tectonic stress was limited to faulting, as described in the last chapter.

Foliation is rarely observed in Karmutsen metabasalts. Bonanza rocks, by contrast, acquire a foliation in the innermost greenschist-amphibolite transition zone which characterises them throughout the inner aureole. This planar fabric can be attributed to three features: the flattening of breccia clasts, alignment of biotite and acicular hornblende, and the accentuation of delicate primary minerological/textural layering in tuffaceous units by recrystallization. Why Bonanza rocks were able to develop foliation in response to stress while Karmutsen rocks generally were not may be explained in several ways.

1. Biotite, which readily assumes a planar orientation, is abundant in Bonanza metavolcanics.

2. Quartz is also abundant. It deforms with ease at much lower temperatures than hornblende and plagioclase, the principal constituents of Karmutsen hornfelses.
3. Some of the "foliation" observed in the field may actually be bedding features transposed parallel to the biotite schistosity.

According to structural evidence (Chapter IV) this foliation resulted from forcible emplacement of the quartz diorite. Textural relationships show that it accompanied contact metamorphism. Quartz-biotite-plagioclase and hornblende veins are folded with axial planes parallel to the foliation. Small biotite plates are oriented. In one example of refolding in a siltstone lens a second axial-plane alignment supercedes biotite bedding-parallel alignment.

Except immediately adjacent to the quartz diorite static recrystallization has not overprinted the foliation. This suggests that deformation affected the metavolcanics at the maximum temperatures achieved.According to Jaeger's data (1957 Figure 1), the temperature maximum at the outer limit of developed foliation, 500 meters from the intrusion, persisted to about 320,000 years after its emplacement. During this time-interval, of course, the pluton itself was cooling. The distance of the plane of solidification from the contact 320,000 years after intrusion is about 750 meters (from Jaeger 1957 equation (9)). It is thus probable that solidus or near-solidus conditions prevailed in the outer part of the intrusion while the country rocks were still deforming. This is substantiated by the deformation of inclusions in the quartz diorite (Chapter III), which probably occurred at a late stage of crystallization. The simultaneous development of foliation in the quartz diorite and its country rocks may have resulted from continued uprise of magma into the center of the intrusion, causing it to expand outwards.

## CHAPTER VI: POST-PLUTONIC STRUCTURE AND DIKES

Fracturing, faulting dike emplacement, and low-temperature metamorphism followed crystallization of the quartz diorite. These post-plutonic structures trend northeasterly, nearly perpendicular to the earlier structural grain. Northeast-trending joint sets are widespread both in the pluton and in the Vancouver Group. Post-plutonic dikes tend to strike parallel to local joint sets (Figure 39). A plot of dike orientations (Figure 40) has a well-defined maximum near N 45 E. The pattern of straight east-northeast valleys on Hardwicke Island follows a system of chloritized fracture zones. The border of the quartz diorite is not significantly offset across these linears. Faulting was generally less important than fracturing. The only major northeast trending fault in the area offsets the intrusive contact 200 meters in a right-lateral sense and juxtaposes Quatsino and upper Karmutsen Formations in Miner's Bay; its probable extension south of Johnstone Strait juxtaposes upper and middle Karmutsen Formations (Kuniyoshi 1972). Post-plutonic dikes are disrupted in a number of localities, suggesting small-scale movements.

Post-plutonic dikes constitute five percent of outcrop volume along the coast between Patterson Bay and the southeastern tip of Hardwicke Island: their emplacement involved a five percent northwest-southeast crustal extension. Because of pervasive epidote-chlorite alteration the dikes are typically green. They crosscut each other at many localities (Figure 41) but in no apparent petrologic sequence. They range from hornblende basalt to dacite with andesite predominant (Table 4).

Fig. 39

Poles to post-plutonic dikes and joints

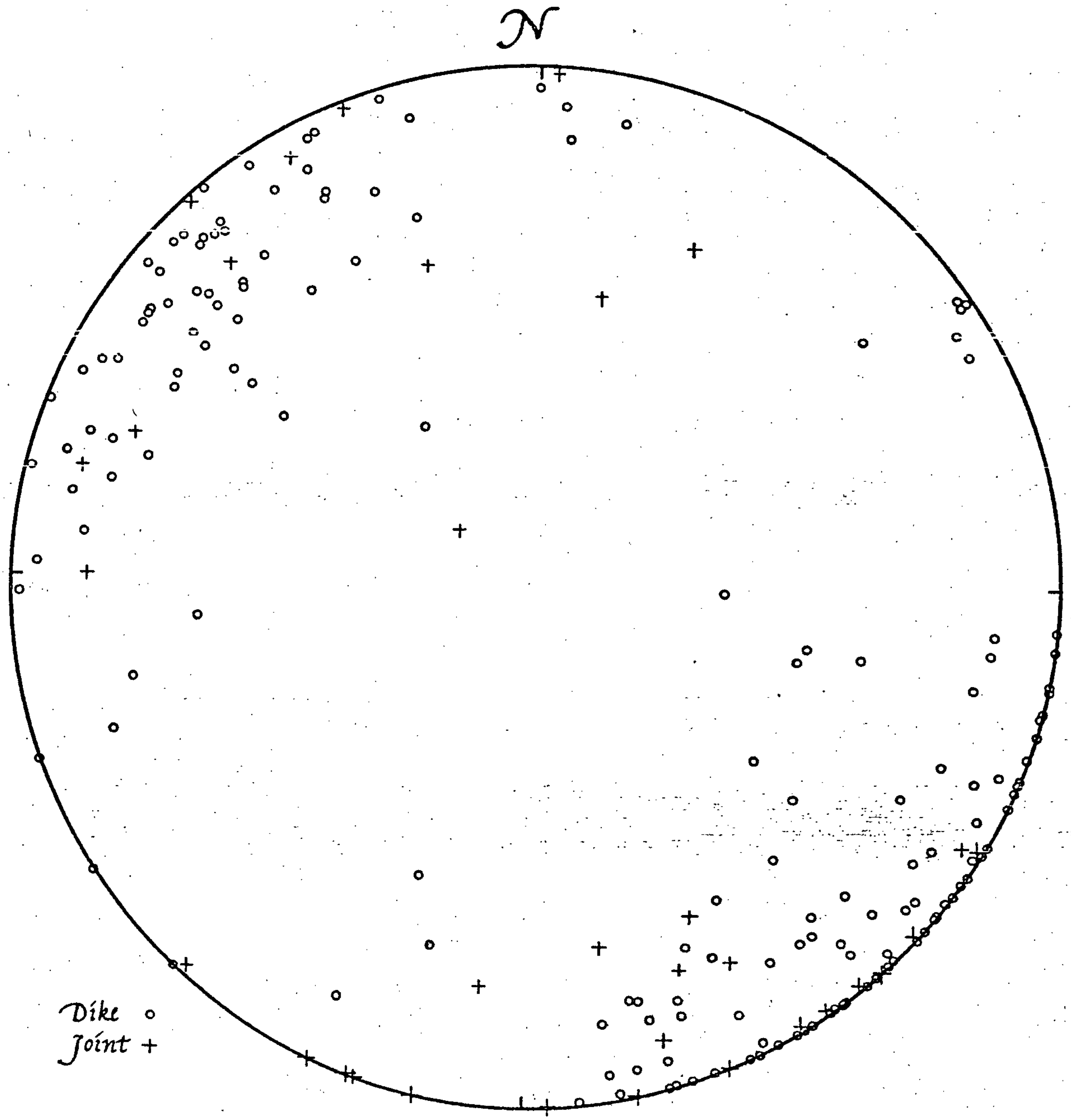
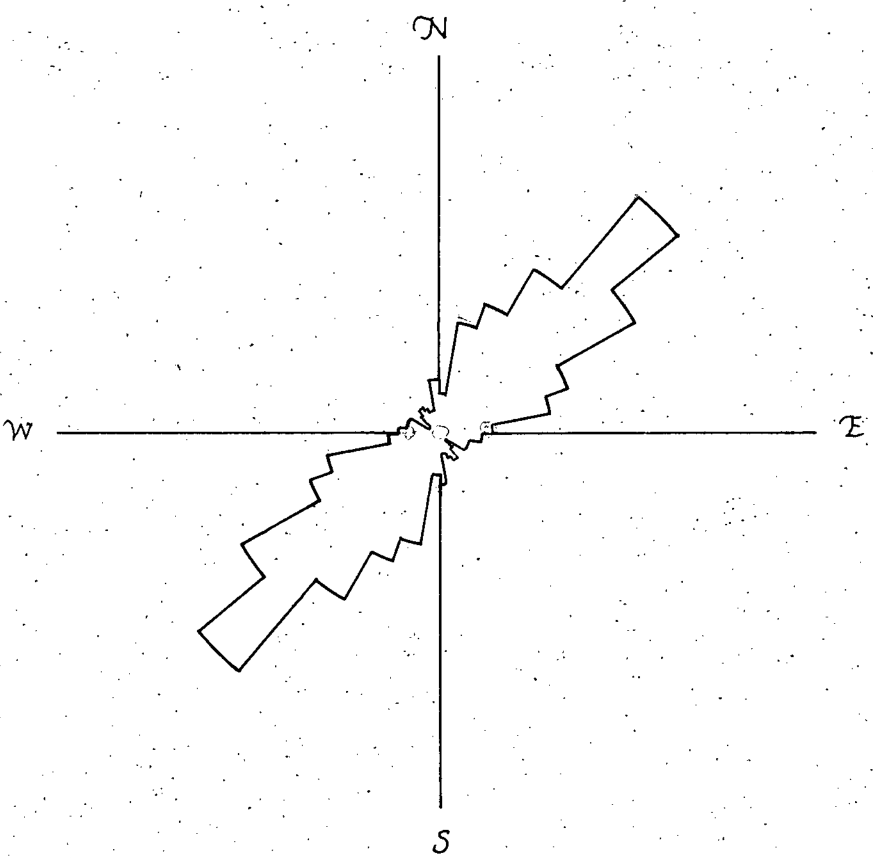


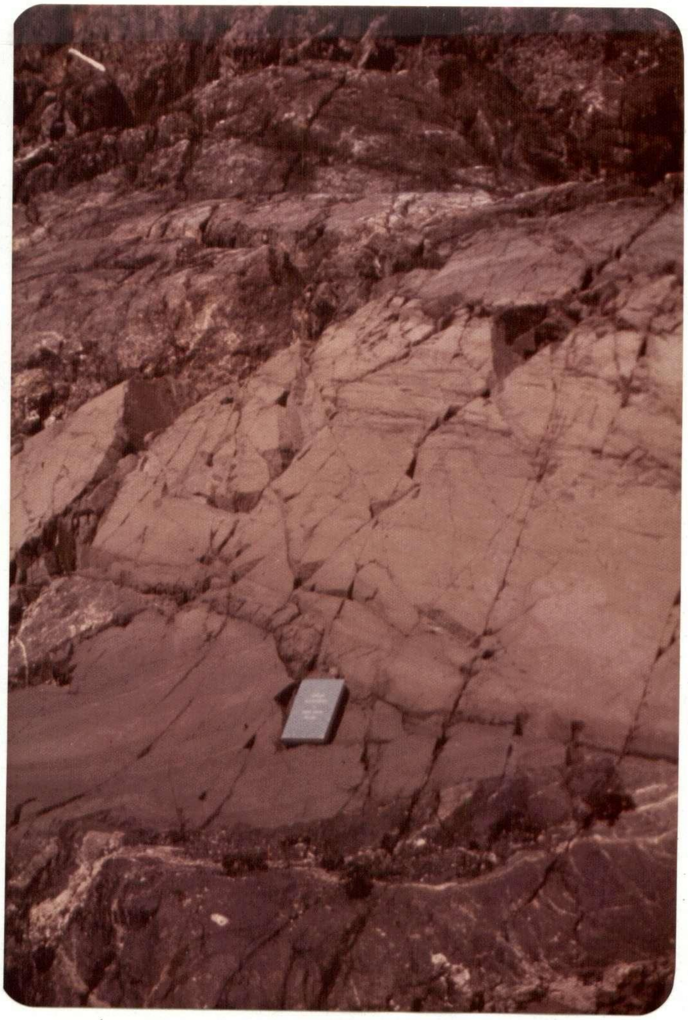
Fig. 40

# Strike orientations of post-plutonic dikes




total points : 411

Fig. 41



Cross-cutting post-plutonic dikes

Chilled margin visible on later dike.

## Table 4 = Petrographic Data = Post Plutonic Dikes

| Sample Number | #1 | #2 | #3 | #4 | #5 | #6 |
|---|---|---|---|---|---|---|
| **Primary:** | | | | | | |
| clinopyroxene | 1 | 1 | 5 | 3 | 0 | 0 |
| hornblende | 0 | 0 | 0 | 0 | 0 | 40 |
| plagioclase | 10(An54) | 0 | 40(An60-63) | 10 | 0 | 0 |
| **Secondary:** | | | | | | |
| albite(+saussurite+sericite) | 55 | 50 | 25 | 60 | 75 | 35 |
| epidote | 1 | 5 | 2 | 5 | 2 | 10 |
| chlorite | 15 | 30 | 15 | 10 | 5 | 10 |
| prehnite | 0 | 0 | 0 | 2 | 3 | 0 |
| pumpellyite | 0 | 1 | 0 | 0 | 0 | 0 |
| calcite | 2 | 0 | 0 | 0 | tr | 0 |
| sphene | 1 | 1 | 0 | 1 | 0 | 0 |
| **Either:** | | | | | | |
| quartz | 5 | 2 | 3 | 5 | 15(phenocrysts) | 2 |
| magnetite | 10 | 1 | 10 | 5 | 1 | 3 |
| ilmenite | 0 | 10 | 0 | 0 | 0 | 0 |
| **Percentage of secondary minerals:** | 74 | 87 | 42 | 78 | 85 | 55 |
| **Rock name:** | andesite | andesite | andesite | andesite | dacite | hornblende basalt |

The large proportions of secondary phases shown in Table 4 attest to the high degree of alteration.

By contrast the quartz diorite and metavolcanics intruded by these dikes are generally quite fresh. They contain only small quantities of low-temperature phases, including chlorite in biotite and hornblende, prehnite in biotite, saussurite, epidote, and pumpellyite. Significant low-temperature alteration is confined to joint surfaces (mainly epidote) and to the east-northeast fracture zones (chlorite-albite, quartz-albite-epidote-chlorite). Thus, although the entire area has been subjected to late prehnite-pumpellyite metamorphism, pervasive alteration has occurred only in fractures and in the dikes which fill them. This association of abundant low-temperature phases with fractures suggests that fluid circulation in them exerted a controlling influence on the alteration process.

Prehnite-pumpellyite facies metamorphism at low pressure requires temperatures between 200°C (analcime + qz = albite + $H_2O$ - Liou 1971) and 400°C (prehnite breakdown - Liou 1970). If the fractures were formed and the dikes emplaced shortly after crystallization of the pluton the remaining heat would have been sufficient to metamorphose them. This model is somewhat preferable to that of a post-plutonic burial-metamorphic episode because of the concordance of the radiometric dates (see Chapter III).

The northeast fracturing and dike emplacement are the latest tectonic events recorded in the map area. As they are probably reflected in the radiometric dates on the quartz diorite, no post-

Jurassic tectonic events of any significance are seen at this part of the province boundary. In particular there are no northwest-trending faults or dike swarms that might be associated with a northwest-trending Upper Cretaceous rift zone. This of course does not eliminate the possibility, it merely fails to support it.

SUMMARY OF FIELD RESULTS

The history of the map area began with the deposition of the Vancouver Group - the Karmutsen basalts, the Quatsino limestone, the limestones and fine siltstones of the Parson Bay and Harbledown Formations,and last the volcanic breccias and siltstones of the Bonanza Formation. Extensive diking accompanied both Karmutsen and Bonanza volcanism. Following their deposition, rocks of the Vancouver Group were subjected to prehnite-pumpellyite burial metamorphism at pressures less than 3 kb. Two early plutonic phases have been identified at Miner's Bay and on eastern Hardwicke Island. The main intrusion, a homogeneous quartz diorite with a gabbroic rim, dated as Later Jurassic, was probably emplaced along deep west-northwest fractures parallel to the Telegraph Hill and Johnstone Strait faults. The Telegraph Hill Fault is demonstrably pre-plutonic. The existence of these large faults provides an explanation for the anomalous west-northwest trend of the province boundary along Johnstone Strait. The quartz diorite produced a thermal aureole about 4 km. wide; maximum temperatures at the contact reached 650°C. It also generated all the important structures in the area except the monoclines between the major faults and the northeast-trending fracture system, which is post-plutonic. Post-plutonic dikes indicate 5% northwest-southeast extension, i.e. parallel to the regional tectonic trend. The dikes have been metamorphosed in prehnite-pumpellyite facies. They probably predate the 150 m.y. radiometric age of the quartz diorite.

No support was found in the area for a Lower Jurassic suture between the two provinces; in fact arguments can be made against it. No strong pre-plutonic deformation can be demonstrated and no "trench" sediments occur in the Vancouver Group. Bonanza breccias are monolithologic and purely volcanic; the fine greywackes in the Harbledown Formation are most likely accumulations of local basalt weathering debris (Carlisle and Suzuki 1974). No evidence was found either to support or refute the existence of a northwest-trending rift zone.

The Telegraph Hill Fault with a maximum stratigraphic separation of 5000 meters, and the Johnstone Strait Fault with a maximum of 7000 meters, are the only major tectonic features in the area not directly related to the intrusion. Faults of similar magnitude are common on Vancouver Island. They cannot be said to constitute a major tectonic break between the Insular Belt and the Coast Plutonic Complex. In fact the most significant feature of the boundary here is an intrusive contact.

The province boundary has also been mapped on Quadra Island by Carlisle and Suzuki (1965) and on the northeast shoulder of Vancouver Island by Kuniyoshi (1972) (Figure 5), who similarily describe it as an intrusive contact. The plutonic rocks involved are generally Early Cretaceous. Metamorphic conditions resemble those found here: prehnite-pumpellyite regional metamorphism preceded contact metamorphism on northeastern

Vancouver Island (Kuniyoshi 1972) and just west of the province boundary on Texada Island (Asihene 1970). Kuniyoshi determined a total pressure of approximately 2 kb.; Asihene, assuming 2 kb., estimated a maximum contact metamorphic temperature of 700$^{o}$C.

On Quadra Island intense pre-plutonic deformation - folding of the limestones and foliation in the metavolcanics - was interpreted by Carlisle as due to regional deformation in a shear zone extending onto Vancouver Island (Kuniyoshi 1972). On Hardwicke and West Thurlow Islands the foliation and intense folding are interpreted as due to forcible intrusion. No northwest regional deformation is evident prior to the west-northwest synplutonic phase. The case of Hardwicke and West Thurlow Islands points out the importance of intrusion-related deformation, which may be of significance elsewhere along the province boundary.

This field study has verified on a small scale what is apparent on large-scale geologic maps, that the boundary between the Insular Belt and the Coast Plutonic Complex is simply the western limit of voluminous intrusion. While this limit may in part be controlled by faults, it is essentially Dickinson's magmatic front (1971; 1973).

It has shown so far that Georgia Depression lies west of a mainly Early Cretaceous magmatic front, and that the depression formed in the Upper Cretaceous, tentatively as a complement to the uplift of the Cretaceous arc. It has been proposed that the whole Coastal Trough has had a common mode of origin. If so, it too should lie immediately west of magmatic arcs and form at the same time that they rose. In the following sections on Hecate Depression, the Puget and Williamette Lowlands and the Great Valley these relationships will be established.

# SECTION III: THE COASTAL TROUGH

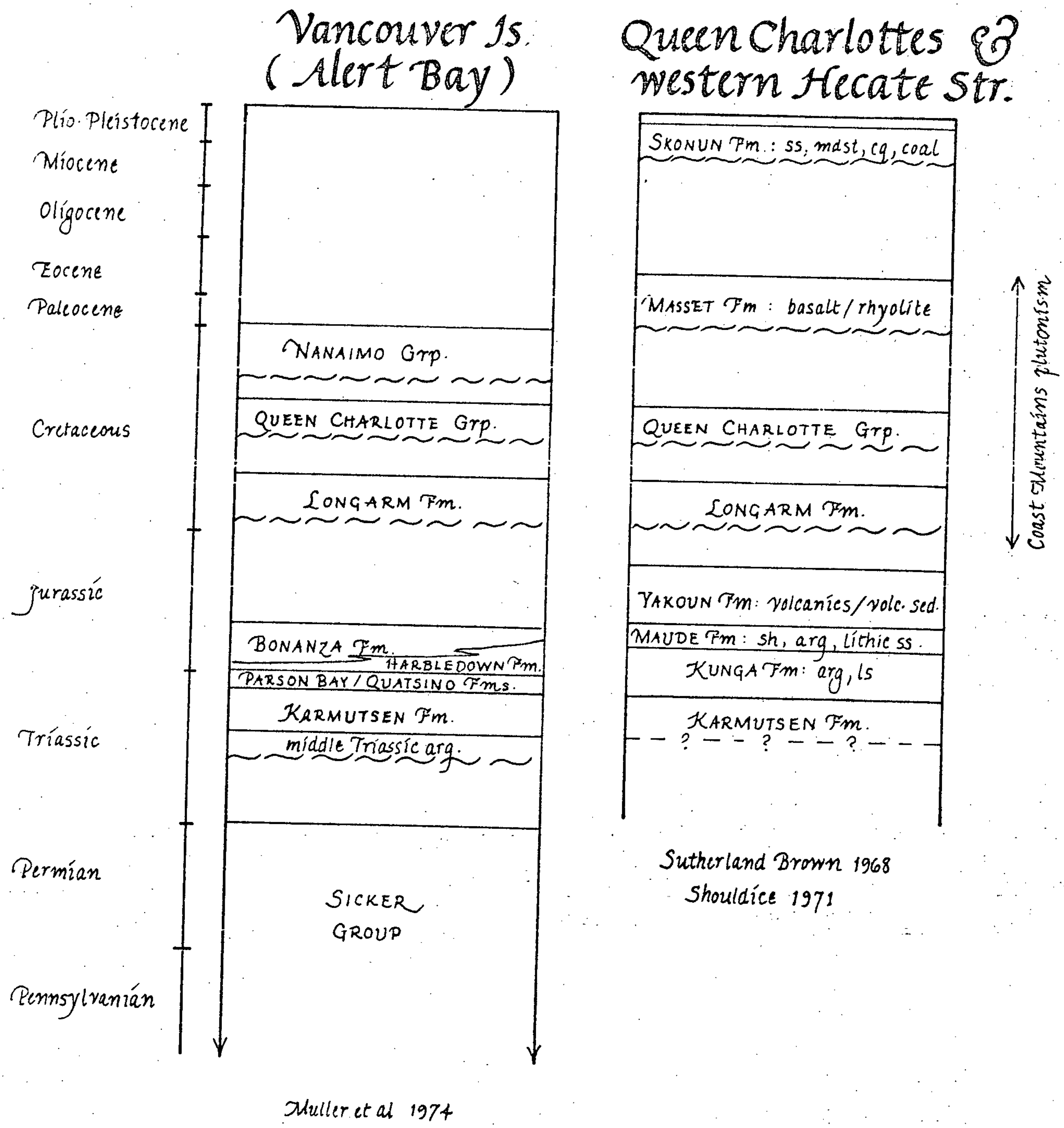
## CHAPTER VII: GEOLOGIC SETTING OF THE COASTAL TROUGH

### i. Hecate Depression

Hecate Depression is the northernmost sub-basin of the Coastal Trough. Like Georgia Depression it lies across the Coast Plutonic Complex/Insular Belt boundary. The exact position of the province boundary under Hecate Strait is uncertain. Triassic and Jurassic volcanics, including probable Karmutsen equivalents, occur near Prince Rupert (Skeena River 1 : 1,000,000 map sheet Hutchison et al. 1973). They may represent the eastern margin of the Insular Belt, in which case it extends considerably east of the axis of the depression, as it does under Georgia Strait.

The paleogeographic evolution of Hecate Depression is not well known. Most of it lies under water and is only accessible to drilling and seismic methods. The adjacent Coast Mountains have been mapped on a regional scale by the Geological Survey (Hutchison 1970; Baer 1973; Roddick and Hutchison 1974). However stratified inliers are too scarce to support a detailed history. The Queen Charlotte Islands have been mapped on a regional scale by Sutherland Brown (1968). The Triassic to mid-Cretaceous history of the Queen Charlottes resembles that of the southern Insular Belt (Figure 42). The Lower to Middle Cretaceous clastic sequences - the Longarm Formation and Queen Charlotte Group - are chronologically and lithologically equivalent to those on

Fig. 42

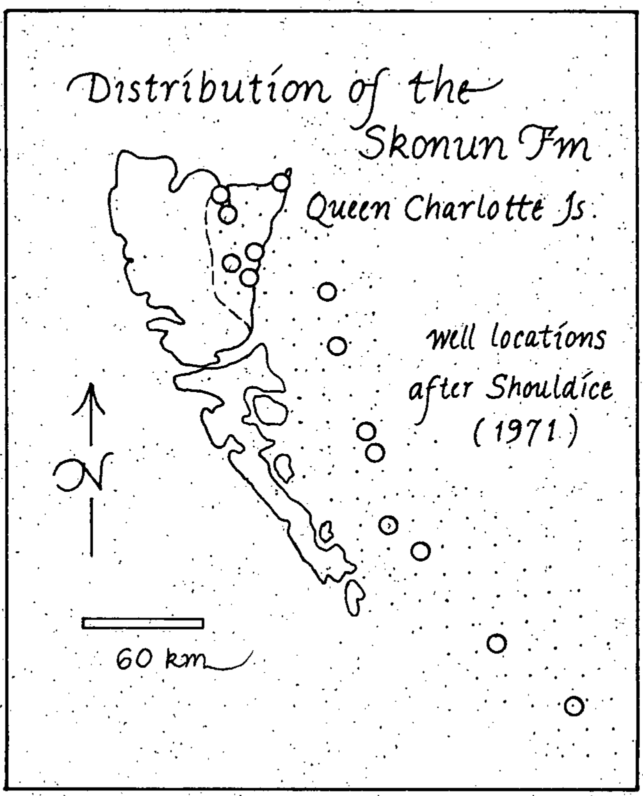



Comparative Insular Belt Stratigraphies

Vancouver Island, but while the Queen Charlotte Group on Vancouver Island seems to be derived from an early Insular Belt uplift (Muller et al. 1974), sediment patterns in the Queen Charlottes are more complex and insufficiently studied (Sutherland Brown 1968). The presence of Middle Cretaceous land in the place of Hecate Depression has yet to be established.

Hecate Depression probably did not develop at the same time as Georgia Depression. There is not known equivalent of the Nanaimo Group either on the Queen Charlottes or below Hecate Strait, so far as can be ascertained from well and seismic reflection data (Shouldice 1971), although the Skidegate Formation (upper Queen Charlotte Group) may extend into the Upper Cretaceous (Muller pers comm. 1976). A pre-Tertiary unconformity separates the Paleocene-Eocene Masset Volcanics from all other formations. The Masset Volcanics themselves are predominantly continental. On the eastern Queen Charlottes and under Hecate Strait they are overlain, again unconformably, by Upper Miocene and Pliocene marine and non-marine sediments of the Skonun Formation, the first clear indication of a coastal trough in northern British Columbia (Figure 43). The Masset volcanics and the erosion surfaces on top of them dip gently into Hecate Depression. The overlying Skonun Formation thickens basinwards; it changes from dominantly continental to dominantly marine in a series of wells from northern Graham Island southeast into Hecate Strait (Shouldice 1971). Sutherland Brown (1968) describes the Mio-Pliocene basin as "...moderately rapidly sinking...possibly fault controlled at its eastern and northern margins...filled to capacity with detritus from the proto-Queen Charlotte Islands and probably the Mainland."

Fig. 43


Distribution of the Skonun Fm
Queen Charlotte Is.
well locations after Shouldice (1971)
N
60 km

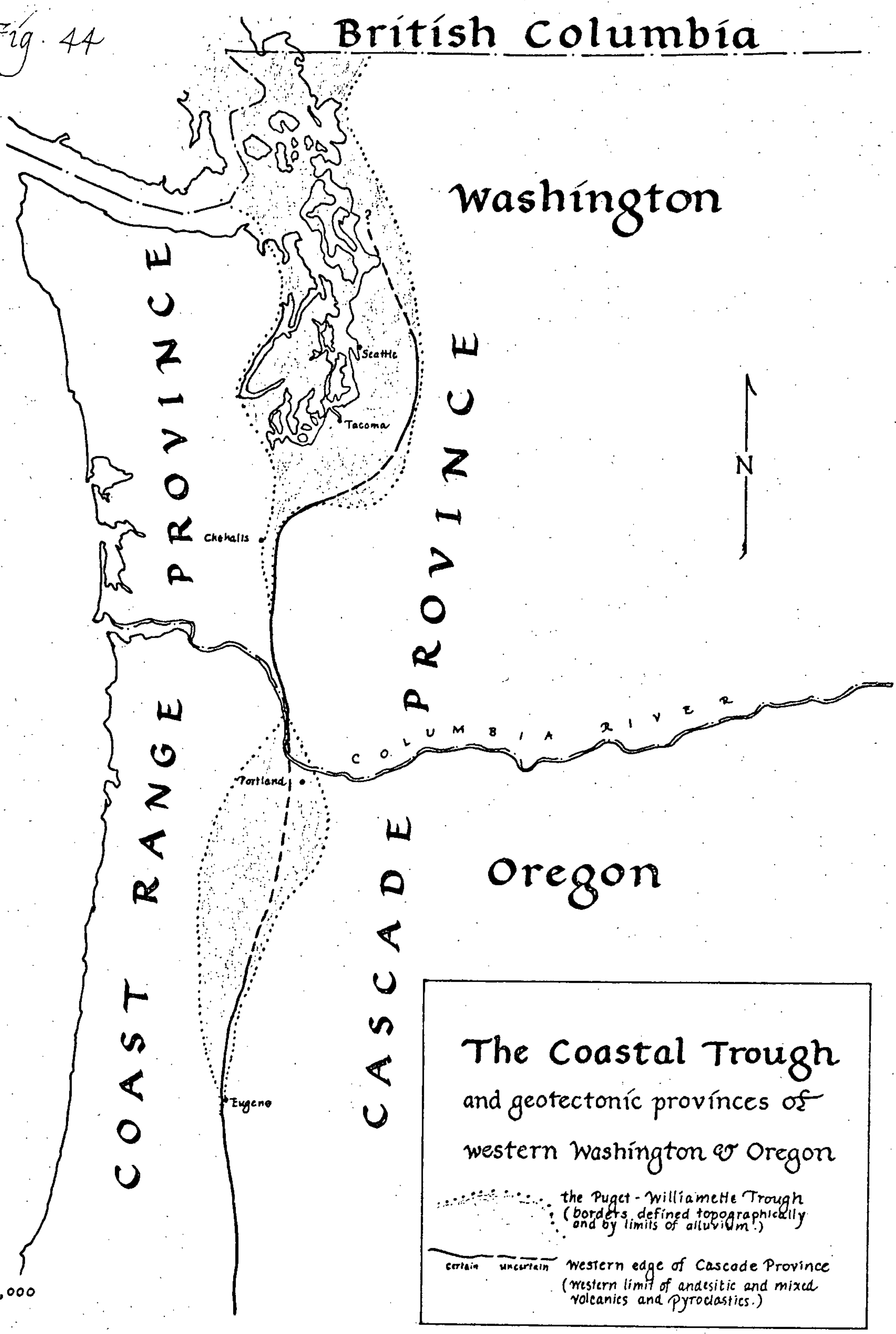
Two events in the history of the Coast Plutonic Complex need to be emphasized here - the dates of intrusion and the timing of uplift. Potassium-argon dates follow an eastward younging trend from Early Cretaceous to Eocene (Woodsworth pers. comm.), like those in the southern Coast Mountains. Eisbacher (1974) attributes pre-Cenomanian deformation of the Bowser Group and the advent of westerly-derived clastics into the Sustut Basin 54 m.y. ago to pulses of uplift in the northern Coast Plutonic Complex. The latest phase has involved as much as 1000 meters of uplift of Miocene erosion surfaces (Baer 1973); it is roughly coeval with the deposition of the Skonun Formation.

Hecate and Georgia Depressions, while occupying similar positions with respect to the tectonic belts, are not time-equivalent. Georgia Depression formed in the Upper Cretaceous while the Coast Plutonic Complex was an active magmatic arc. Hecate Depression formed in the Upper Miocene, long after plutonism had ceased. Yet both accompanied significant pulses of uplift of the Coast Plutonic Complex.

ii. <u>The Puget and Williamette Depressions</u>

The first evidence for a coastal trough in western Washington and Oregon is in the Upper Miocene, contemporary with the beginning of uplift of the Cascades. The present trough comprises two elongate depressions, the Puget and Williamette Lowlands (Figure 44). They are bounded on the east by the Cascade Range and on the west by the Coast Ranges and Olympic Mountains. These flanking ranges belong to two distinct geotectonic provinces (Figures

Fig. 44

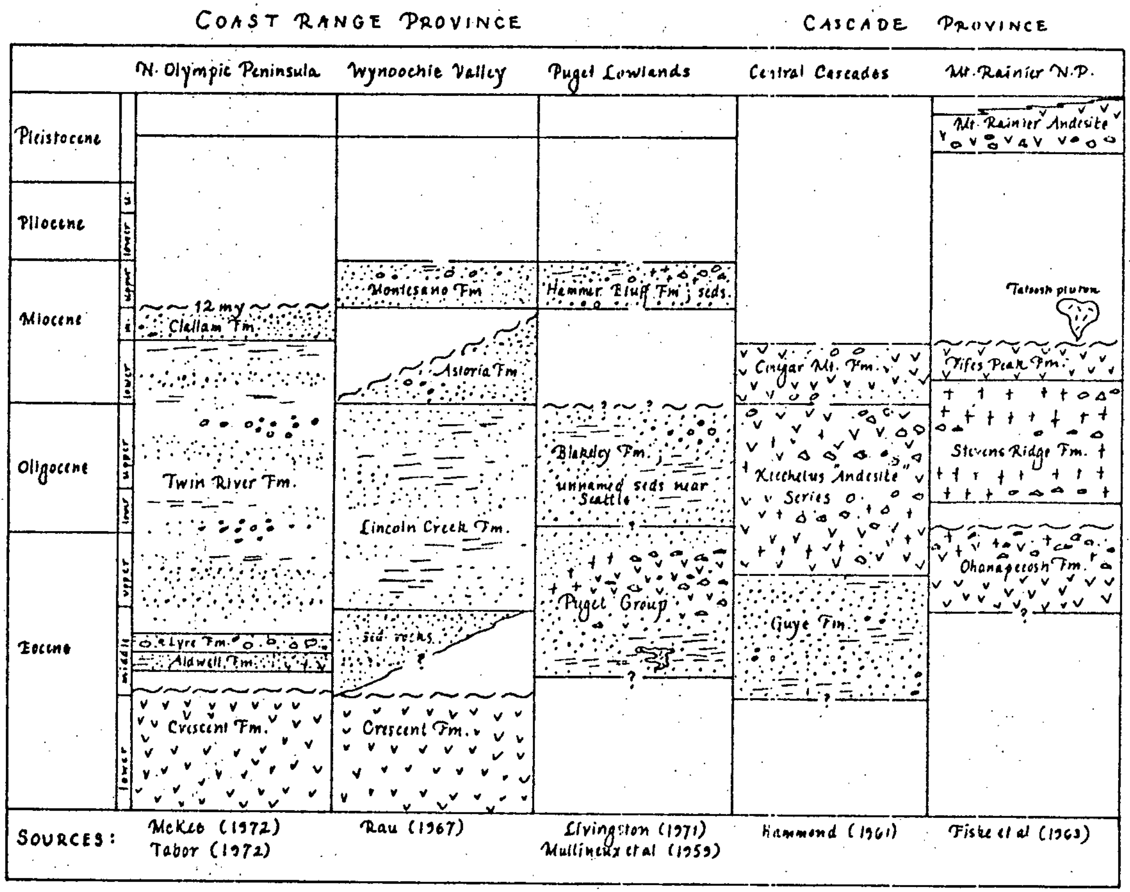

British Columbia
Washington
Oregon
COAST RANGE PROVINCE
CASCADE PROVINCE
Seattle
Tacoma
Chehalis
Portland
Eugene
COLUMBIA RIVER
N
scale 1:7,500,000
The Coastal Trough
and geotectonic provinces of
western Washington & Oregon
the Puget - Williamette Trough
(borders defined topographically and by limits of alluvium.)
certain
uncertain
Western edge of Cascade Province
(western limit of andesitic and mixed volcanics and pyroclastics.)

45 and 46). The Coast Range Province includes both the Coast Ranges and most of the Coastal Trough. The Cascade Province coincides with the Cascade Range, an uplifted Cenozoic magmatic arc.

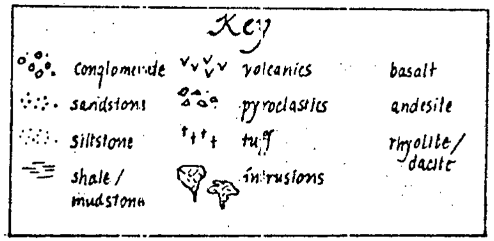
The backbone of the Coast Ranges is a north-south trending chain of Eocene volcanic centers which produced the voluminous tholeiitic and alkalic basalts of the Siletz River, Crescent, Metchosin and Umpqua Formations. Chemically they resemble mid-ocean ridge and ocean island basalts (Snavely and Wagner 1968). The Eocene basalts intertongue with and are overlain by middle Eocene to Pleistocene clastic sediments. The sediments are derived both from within the Columbian Embayment and from its pre-Cenozoic margins, the Klamath Mountains and the Insular Belt. Tuffs and tuffaceous sediments from the Cascade Province form an important part of the section, particularily the Oligocene (Snavely and Wagner 1963). After the Eocene, flows and intrusions within the Coast Range Province become rare.

The Cascade Province is a north-south trending linear volcanic belt approximately 150 km. wide lying about 200 to 250 km. from the western edge of the North American plate. It may be superimposed on Eocene Coast Range Province rocks (Snavely and Wagner 1963; Baldwin 1964). Its northern end laps onto the pre-Tertiary North Cascades, its southern end onto the Klamaths. While Cascade vulcanism spans the entire Cenozoic era since the middle Eocene, it has been distinctly episodic. (Sutter and McBirney 1974; McBirney et al 1974). Andesite is most abundant with lesser amounts of basalt, dacite and rhyodacite. Church and

Fig 45

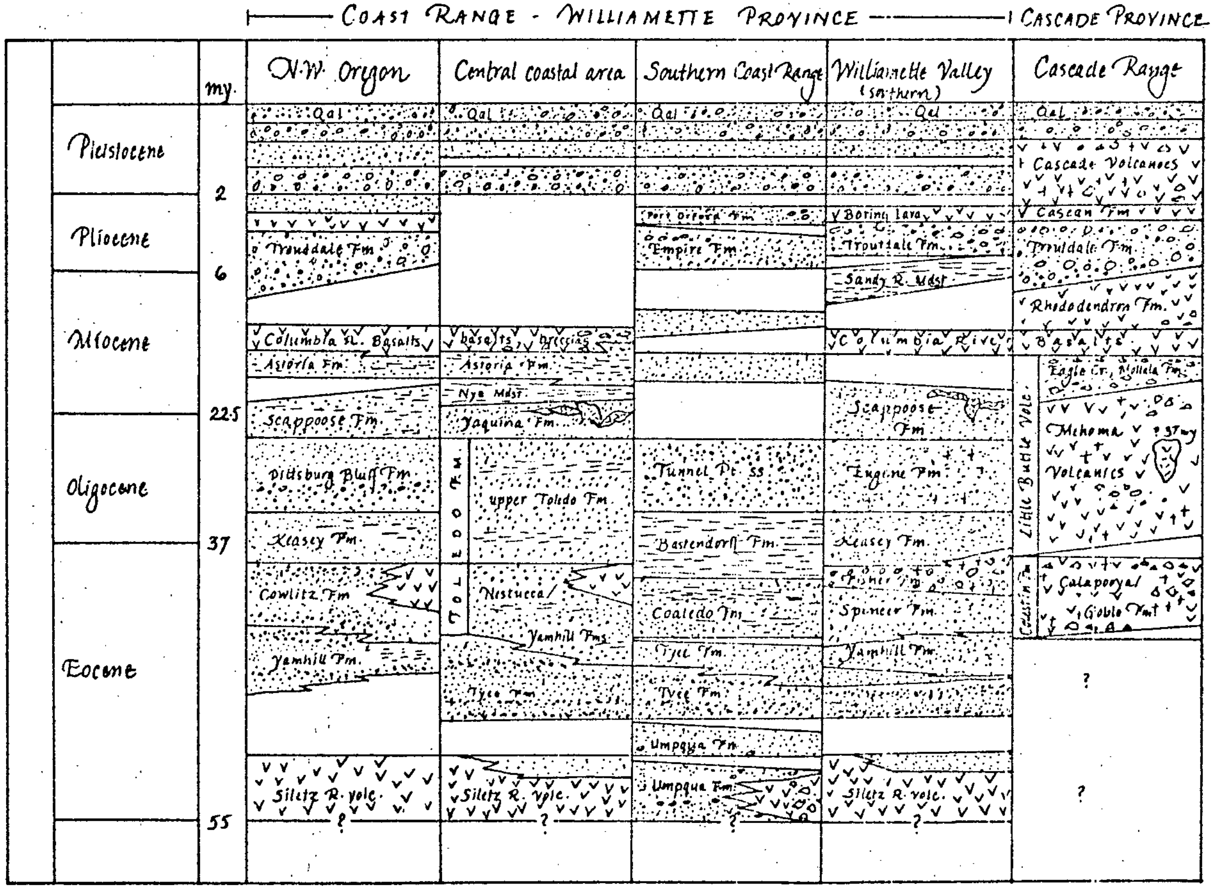


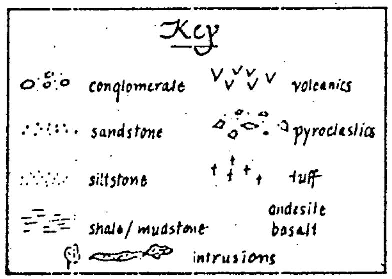

Lithologic chart : western Wash.




137.

Fig. 46




Lithologic chart : western Oregon

after Baldwin (1964)

p. 139 does not exist



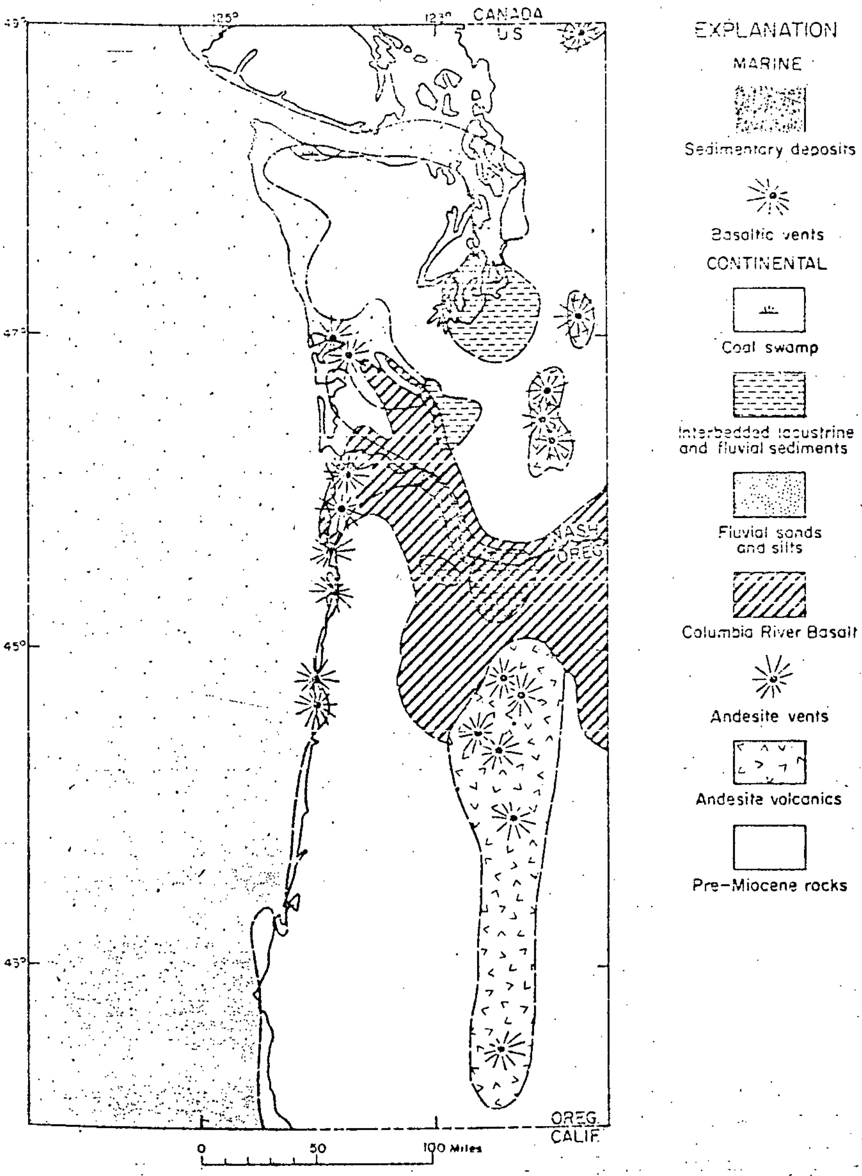
Tilton (1972) have used lead isotopic ratios and low, relatively constant Sr87/Sr86 (.7037 ave.) in Pleistocene Cascade volcanics to support a subcrustal origin. In addition to volcanics, Miocene intrusions (granodiorite with subordinate diorite, quartz monzonite, granite, etc.) are exposed in the Washington Cascades.

The boundary between the Cascade and Coast Range Provinces (i.e. the western limit of Cascade volcanism) lies within the Coastal Trough. Its exact position is largely obscured by drift and alluvium. In places it is expressed as a volcanic/sedimentary facies change near the eastern side of the trough. In King County, Washington, Tertiary volcanic rocks, mostly andesite breccias and flows, extend westward to the Cascade foothills (Livingston 1971). Tertiary sediments, arkose, volcanic sand, tuff, clay and coal, predominate in the Puget Lowland. The transition (for example the zone of interfingering between Eocene and Puget Group sediments and the Keechelus Volcanic Series) occurs near the eastern margin of the lowland and in the Cascade foothills. The position of the volcanic front in the Miocene is not as well bracketed as the Eocene. Cascade stratigraphy is not yet precise enough to define the western limit of Miocene volcanics; however it must have lain between the westernmost exposure of Miocene Snoqualmie Granodiorite five kilometers east of North Bend and sediments of the Miocene Hammer Bluff Formation (Glover 1936) along the Green River south of Black Diamond.

The Oligocene volcanic front in Oregon is exposed on the eastern side of the Williamette Lowland. The Eugene Formation (Middle Oligocene) and the Butte Creek beds (Upper Oligocene), outcropping within the Lowland, are predominantly marine volcanic sandstone and waterlain tuff. They interfinger along its eastern side with the Mehama Volcanics, a series of andesitic and dacitic tuffs, breccias, and flows (Baldwin 1964; Beaulieu 1971). These examples should not be taken to imply that the western limit of Cascade volcanism remained fixed at the eastern edge of the present Coastal Trough. It is evident from paleogeographic reconstructions (Snavely and Wagner 1963) that the position of the volcanic front varied with time. However only infrequently and only in places did it advance west of the front of the Cascade Range, which accordingly corresponds to a significant decrease in volcanic volume. The location of the Coastal Trough, then, is linked to the distribution of Cascade volcanics. But while Cascade volcanism began in the Middle Eocene, the first evidence for a north-south trending depression on the site of the present trough is shown by the distribution of the Columbia Basalt west of the Cascades and the fluvial-lacustrine Wilkes Formation (Figure 47). The Columbia Basalts are dated as 13-16 m.y. (McBirney et al. 1974); the Wilkes Formation is Upper Miocene and possibly Pliocene (Roberts 1958).

Sinking of the trough accompanied uplift of the Cascades. As the Cascades rose they became a barrier to moisture from the Pacific Ocean. The progressive lowering of rainfall east of

Fig. 47




Upper Miocene paleogeography of the Columbian Embayment

— Snavely and Wagner 1963
Fig. 16

the Cascades caused an evolution of the flora from moist-to-dry climate assemblages. The first manifestation of this trend is in the Upper Miocene Mascall flora, which shows a decrease in abundance of Metasequoia from earlier floras (Chaney 1925). The xeric trend culminated in the middle Pliocene (Detling 1968).

Distribution of the Pliocene Troutdale Formation provides and estimate of the amount of differential movement between the Oregon Cascades and the Williamette Lowland in the late Pliocene-Pleistocene. Fluvial in origin, it is interpreted as the remains of a Pliocene Columbia River alluvial plain which extended from the Portland Basin eastward to The Dalles (Baldwin 1964). Near Portland the Troutdale Formation outcrops a few hundred feet above sea level; at Nesmith Point above the Columbia Gorge it has been elevated to 700 meters (Allen 1958). The Oregon Coast Ranges have also risen markedly in the late Cenozoic. Pleistocene beach terraces lie as much as 300 meters above sea level (Baldwin 1964). Plio-Pleistocene uplift of the Cascades in central Washington exceeds 1500 meters (McKee 1972). The great thickness of Quaternary sediments near Seattle, more than 500 meters (Rogers 1970), attests to considerable sinking of the Puget Lowland.

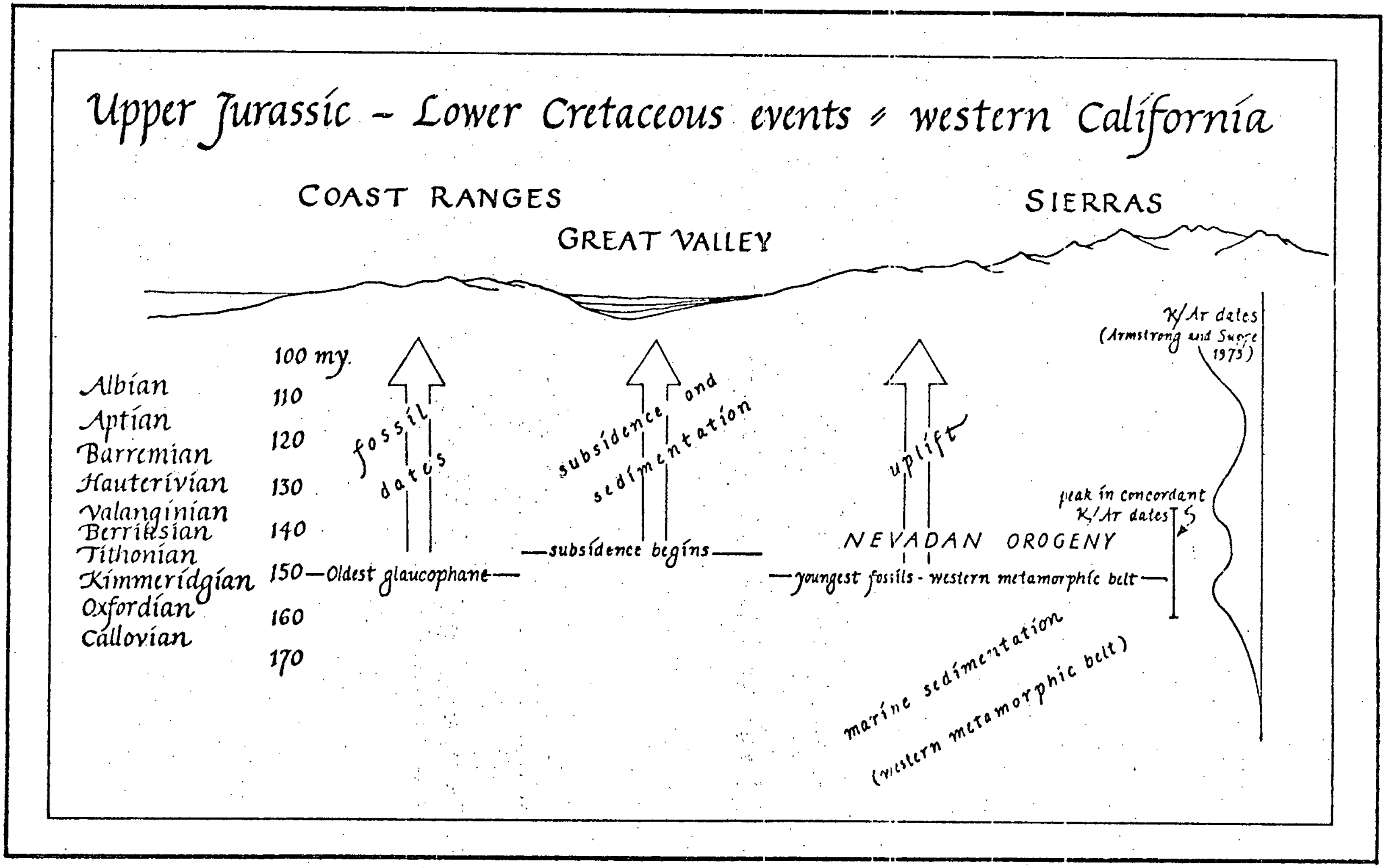
The question of warping versus faulting is not easily resolved: examples of both can be found. The Columbia Basalts along the Columbia River Gorge form a simple upwarp of 1000 meters amplitude, somewhat modified by open folds and minor faults. The geologic profile across northwestern Oregon by Wells and Peck (1961) shows the Coast Range-Williamette Lowland-Cascade Range as upwarp-downwarp-upwarp; however the discontinuity of units across the Lowland may obscure significant faults. Bromery and Snavely (1964) infer the presence of a major normal fault from steep gravity gradients on the western side of the Williamette Lowland. Beaulieu (1971) points out that large faults are possible along the eastern margin of the Lowland. Near Seattle the case is clearer for fault control. The west face of Mount Si is a fault scarp of considerable magnitude (Rogers 1970); the Mount Si fault can be traced north to the San Juans. Earthquakes along it demonstrate continued motion (Danner pers. comm. 1975). According to Danes et al. (1965), the Hood Canal Lineament is a major fault between the Puget Lowland and the Olympic Mountains. Rogers (1970) interprets the Bouguer gravity field in the Puget Lowland as the expression of large fault blocks concealed by alluvium and drift. He could not determine whether the sediments were involved in faulting or merely draped over basement blocks. Since his work several Pleistocene-Recent fault scarps have been identified near the Hood Canal Lineament and near Seattle (Livingston, Gower 1975).

The Puget and Williamette Lowlands fit the model for the Coastal Trough suggested by Georgia and Hecate Depressions. They lie immediately west of the Cascade volcanic front. They have formed simultaneously with the uplift of the Cascade Range. The Coast Range also rose at the same time, making the trough an area of subsidence between two active uplifts. It is significant to note that both faulting and warping were involved.

iii. The Great Valley

The Great Valley of California is a gentle syclinorium of Upper Mesozoic and Cenozoic sediments. On the east they onlap onto the Sierra Nevada, an uplifted Jura-Cretaceous composite magmatic arc. On the west they are in probable fault contact with the Franciscan terrane, and archtypical trench complex (Hsu 1971; Ernst 1971). The Cretaceous Great Valley is one of the examples of arc-trench gaps cited by Dickinson (1971), a zone of tectonic quiet between the arc and the trench. The basal unit of the Great Valley Sequence is the Tithonian Knoxville Formation, which marks the beginning of subsidence. At the same time the Great Valley also became an arc-trench gap (Figure 48). There is evidence for a different arc-trench configuration prior to that time. Volcanic flows, pyroclastic debris and marine volcanic sediments ranging in age from Middle Triassic to Kimmeridgian are exposed in the Klamath Mountains and the Sierra Nevada. They were severely deformed and metamorphosed in Kimmeridgian-Tithonian time. A peak in plutonic dates, 158-132 m.y. (Lanphere and Reed 1973), coincides with this late Jurassic tectonic episode. Moores (1972) interprets the event as the result of an arc-arc collision. Schweickert and Cowan(1974) envisage a Triassic-Jurassic trench immediately west of the Sierran foothills which became choked and stepped outward to the Francisan about 150 million years ago. The beginning of subsidence of the Great Valley also coincided with the initial uplift of the Sierran arc above wavebase: the youngest marine sediments in the Sierra

Fig. 48



Upper Jurassic - Lower Cretaceous events - western California
COAST RANGES
GREAT VALLEY
SIERRAS
K/Ar dates (Armstrong and Suppe 1973)
100 my.
Albian
110
Aptian
120
Barremian
130
Hauterivian
Valanginian
140
Berriasian
Tithonian
150
Kimmeridgian
160
Oxfordian
Callovian
170
fossil dates
subsidence and sedimentation
uplift
peak in concordant K/Ar dates
NEVADAN OROGENY
Oldest glaucophane
subsidence begins
youngest fossils - western metamorphic belt
marine sedimentation
(western metamorphic belt)

are Kimmeridgian (Clark (1964); the first sediments in the Great Valley, derived from the Sierran/Klamath uplift, are Tithonian. The uplift continued as the Great Valley subsided.

The large volume of sediments in the Great Valley, as much as 12-15 km. on the western side (Repenning 1960; Bailey et al. 1964) raises the possibility that subsidence might have been due solely to sediment loading. Although significant, it was not the sole cause. Ojakangas (1968) reports a progression from shallow-water to bathyal deposition in the western part of the Great Valley Sequence between the late Jurassic and the middle Cretaceous, showing that subsidence exceeded sediment loading during that time. In Walcott's (1972) model for crustal downwarping under sedimentary prisms the two processes keep pace.

The structure of the Great Valley is generally simple. The eastern limb of the synclinorium is underlain by gently sloping Sierran basement. The western limb, against the Tesla-Ortigalita Fault, is steep. Faulting within the Great Valley is considered unimportant, although Brown and Rich (1967) and Berry (1973) postulate buried faults. One unresolved issue concerns the geometry of the Great Valley in the Cretaceous: was it a trough, or a shelf and slope? The present Coast Range is of Eocene origin. Was there an appreciable basement uplift west of the Great Valley before that time? Dickinson (1971,1972) has argued for a trough configuration based on longitudinal paleocurrents in the Great Valley Sequence. Ojakangas (1968), on the other hand, explains the current pattern by postulating a submarine fan at the northern end of the Great Valley. Bailey et al. (1964) favor a ridge separating the Great Valley and Franciscan depositional environments because of the comparative lack of detrital potassium feldspar in Franciscan greywackes, but Barbat (1971) considers this to be a diagenetic/metamorphic effect. Westerly-derived sand units have been noted in a few places in the Great Valley Sequence, as has detrital serpentinite (Hackel 1966; Raney 1973; Berry 1973). These need not have come from a major basement ridge: seismic profiling of modern trenches has shown accumulations of deformed sediments piled against the landward side (Le Pichon et al. 1973). Such a welt could supply minor amounts of sediment to the Great Valley sequence.

The Great Valley in the Cretaceous was a zone of subsidence bordering the rising Sierran magmatic arc. It may or may not have been a trough at that time. Its synclinal structure was created after cessation of subduction by movement on the Tesla-Ortigalita Fault. In timing it corresponds to the other sections of the trough, then, but its structure in the Cretaceous may have been open to the west.

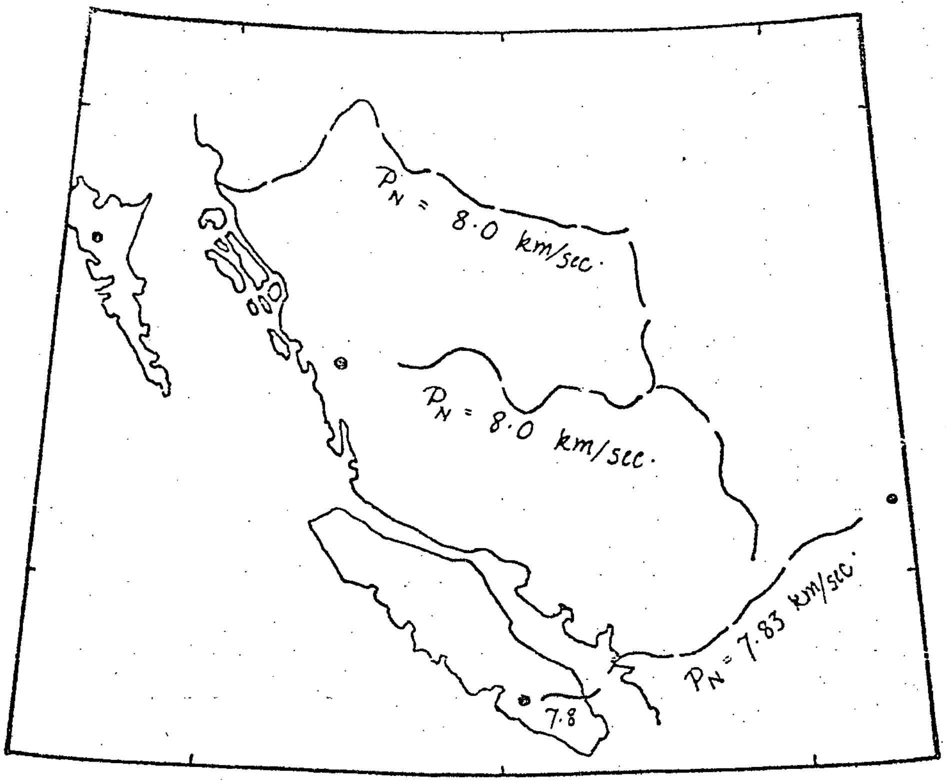
## CHAPTER VIII: DEEP STRUCTURE OF THE COASTAL TROUGH

Refraction seismology and gravity data can aid in illuminating the deep crustal and upper mantle structure under the Coastal Trough. Although these studies are still in early stages, they give several important pieces of information. Seismic work in British Columbia and California indicates the possibility of thinned crust under the Coastal Trough. Hecate Depression and the Great Valley of California are close to isostatic, while the Puget Lowland has one of the largest negative free-air and isostatic anomalies in the world.

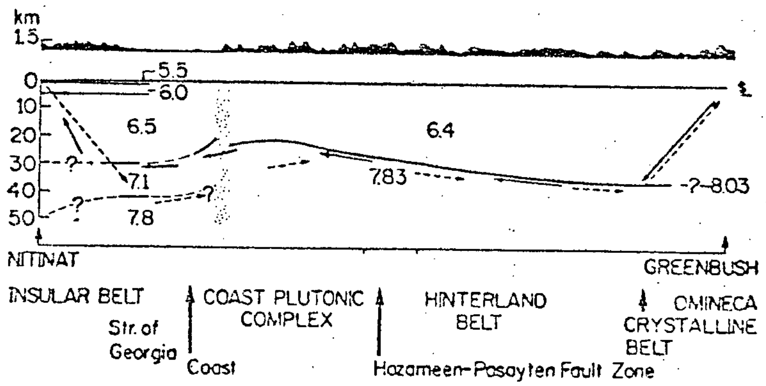
### i. British Columbia

The structure of the lithosphere under northern and southern British Columbia differs greatly. Figure 49a outlines the paths of the seismic profiles used by Berry and Forsyth (1975) in their regional analysis. The northern profile (Forsyth et al. 1974) shows a Moho depth of 26 km. under Graham Island thickening eastward to 35 km. under the eastern Coast Mountains. Upper mantle velocity ($P_N$) is 8.0 km/sec. The central profile (Johnson et al. 1972) also shows $P_N$ = 8.0 km/sec. and crustal thickness increasing eastward from 23 km. at the shot point on the west side of the Coast Mountains. The average free-air anomaly over Hecate Depression is -8 mgal, indicating approximate isostatic equilibrium (calculated from Gravity data for B.C., compiled by Gravity Division, EMR). The southern profile (White and Savage 1965; Tseng 1968; Berry and Forsyth 1975) is characterized by anomalously low $P_N$ under Vancouver Island and the Coast Mountains.

Fig. 49




a) Shot points, recording stations, and upper mantle velocities for reversed seismic profiles in western B.C.



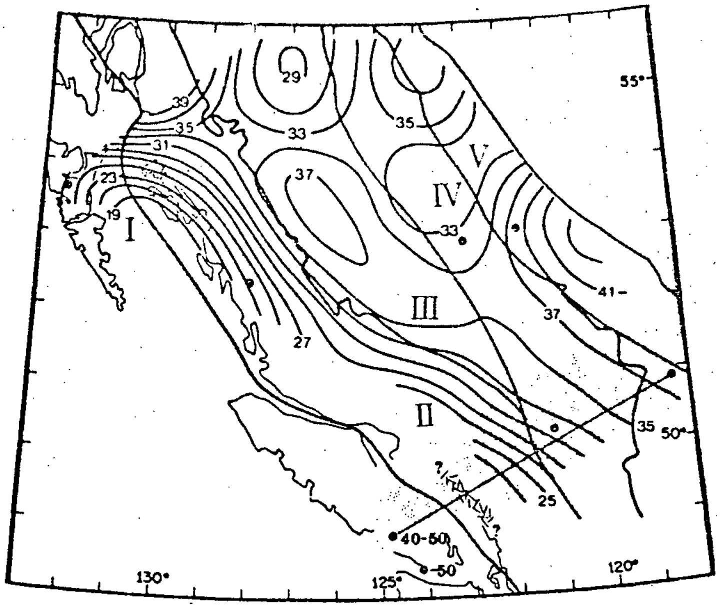

b) Crustal profile across southwestern B.C.

Berry and Forsyth 1975

The crust under the Coast Mountains resembles that in the northern and central profiles; but the crust under Vancouver Island may be 40 to 50 km. thick (Figure 49b). The paucity of stations within Georgia and Hecate Depressions makes detailed reconstructions uncertain. There is some indication of crustal thinning from the Insular highlands into them (Figures 49b and 50a). The seismic scattering zone under eastern Georgia Strait shown in Figure 49b is interpreted by Berry and Forsyth (1975) as either a major fault or an abrupt discontinuity in crustal structure. The marked difference between Vancouver Island and Coast Mountains lithosphere has been documented by the gravity study of Stacey (1973). The lithosphere under Vancouver Island is unusually dense; the lithosphere under the Coast Mountains is unusually light. Figure 49b shows that Georgia Depression overlies a profound change in lithospheric structure, but does not discriminate between the depression and the rest of the Insular Belt. Berry and Forsyth (1975) have calculated $\rho_c$ (crustal density) using the thicknesses shown on Figure 50a. They assume isostasy; and that differences between observed Bouguer anomalies and calculated root anomalies (using a constant crust/ mantle density contrast) are due solely to variations in $\rho_c$. The crustal-density map determined this way shows a strong increase in $\rho_c$ westward across the Coast Mountains with a maximum in the Coastal Trough (Figure 50b). This may be real, or it may be due to other variations in the lithosphere, for instance increase of mantle density under the trough or significant decrease in crustal thickness. Further seismic studies in the trough would aid in establishing which of these models is

Fig. 50

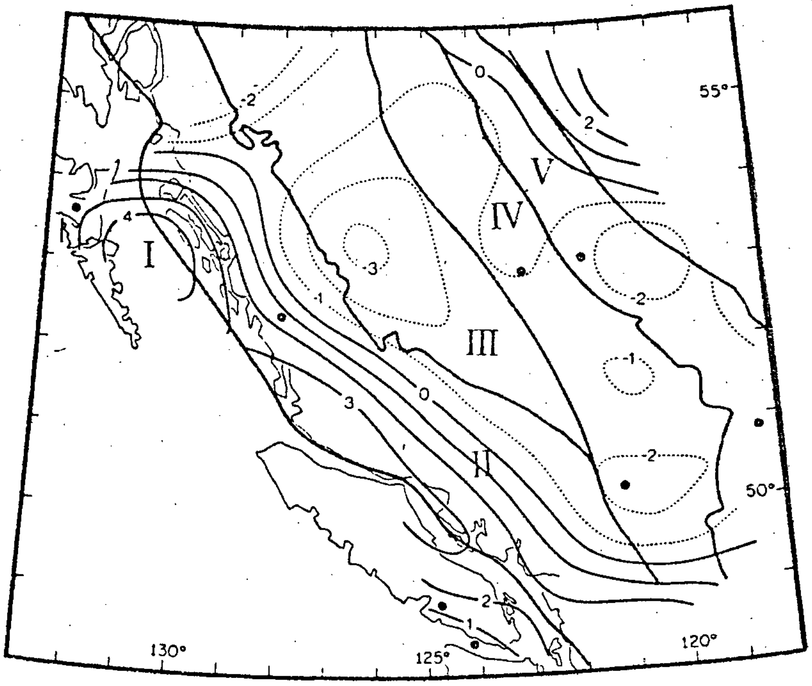
a.




Variations in crustal thickness
British Columbia

Berry and Forsyth 1975

b.




Relative crustal densities, B.C.

Berry and Forsyth 1975

most appropriate. Most importantly, these preliminary data suggest that the Coastal Trough in British Columbia is in some way isostatically compensated. The near zero average free air anomaly over Hecate Depression further suggests isostatic equilibrium.

ii. Washington and Oregon

Seismic studies in Washington and Oregon are indexed on Figure 51. Like the Coast Mountains of British Columbia, the Washington Cascades have no deep root (Johnson and Couch 1970); the Moho slopes under them from 28 km. east of Seattle to 31 km. at 119.5°W. (Figure 52), which also shows other seismically determined depths to the Moho for western Washington projected onto Johnson and Couch's profile. The crustal structure depicted on Figure 52, thickening from the coast to the eastern side of the Cascades, resembles the crust in northern B.C. $P_N$ is low under the entire profile; published estimates are listed on the same figure. The seismic work, as in British Columbia, has not delimited Moho behaviour under the Coastal Trough. Gravity studies provide some constraints. Thiruvathukal (1968) and Thiruvathukal et al. (1970) have presented interpretations of the Bouguer and free-air gravity fields of Oregon. Both Bouguer and free-air highs prevail over the Coast Ranges of southern Washington and Oregon. The Bouguer anomalies pass through zero over the Coastal Trough and become large negative over the Cascades. This trend is due either to crustal

Fig 51

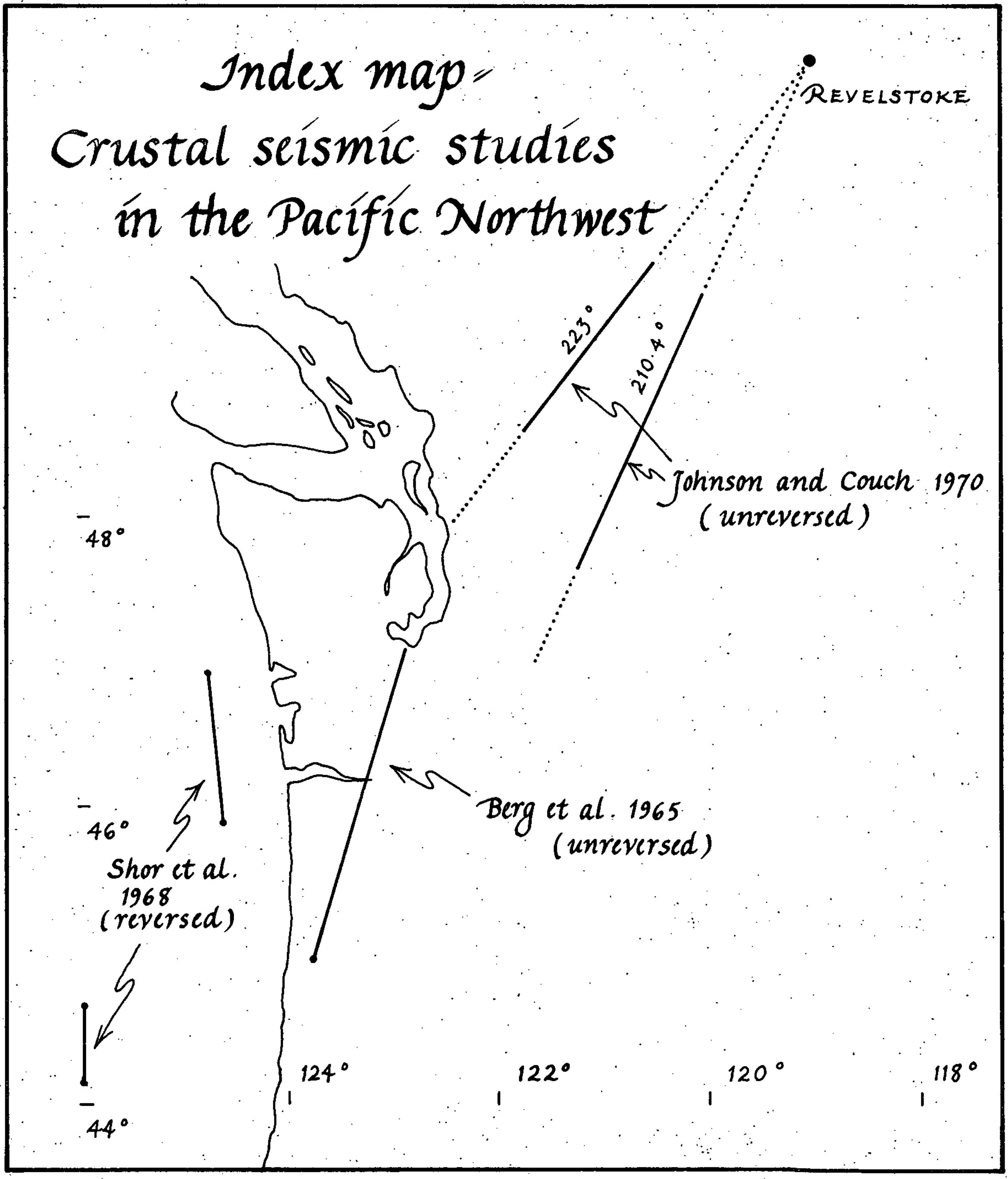
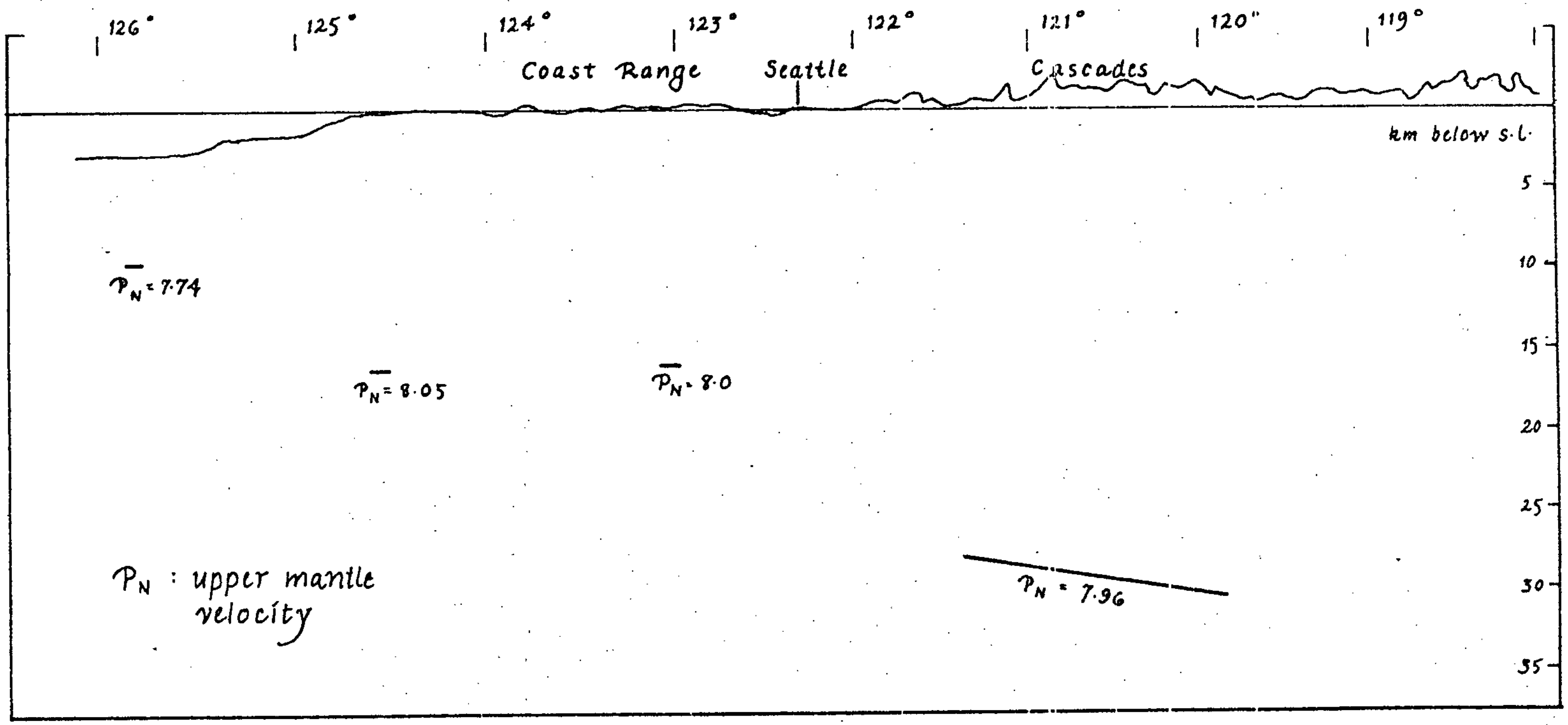

Index map -
Crustal seismic studies
in the Pacific Northwest
REVELSTOKE
223°
210.4°
Johnson and Couch 1970
(unreversed)
48°
46°
Berg et al. 1965
(unreversed)
Shor et al.
1968
(reversed)
124°
122°
120°
118°
44°

Fig. 52




# Crustal thicknesses beneath western Wash.

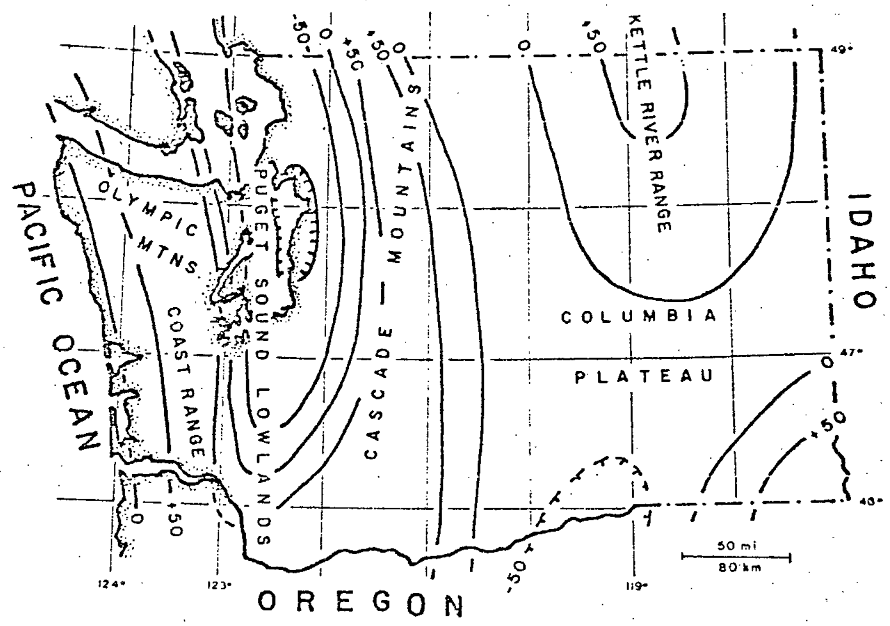
(Projected onto the 223° refraction line of Johnson and Couch 1970) thickening, to lightening of the lithosphere, or both. The free air anomalies show an apparent mass excess associated with the Coast Ranges which Thiruvathukal et al. (1970) attribute either to dense near-surface rocks or shallow mantle or both. The Williamette Depression and the Cascades lie in a zone of negative free-air anomalies which may indicate light crust, light mantle, or a mass deficiency in the total column.

The Puget Lowland is remarkable in several respects. It lies east of the Olympics, the highest and most deformed section of the Coast Ranges; it is underlain by as much as 10 km. of sediments, and it is characterized by large delays in seismic arrivals and one of the largest Bouguer/free-air lows in the world (Woollard 1969). Figure 53, after Daneš (1969), shows this low flanked by free-air highs over the Coast Ranges and the Cascades. Danes believes that the anomalies are maintained by hydrodynamic stresses which counter the tendency of Puget Lowland to rise and the adjacent ranges to sink.

iii. California

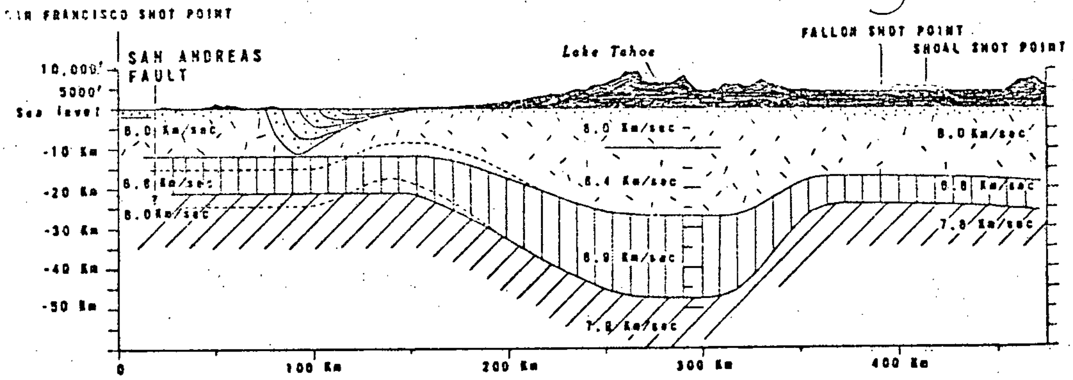
The crustal structure of western California is still subject to major uncertainties. Eaton (1966), based on profiles along and across strike, has determined a 50 km. root under the Sierra Nevada. Carder (1970, 1973) interpreted early arrival times across the Sierras from the Nevada Test Site as evidence against a deep root. Pakiser (1974) has reinterpreted Carder's data to be compatible with thickened crust under the Sierras.

Fig. 53

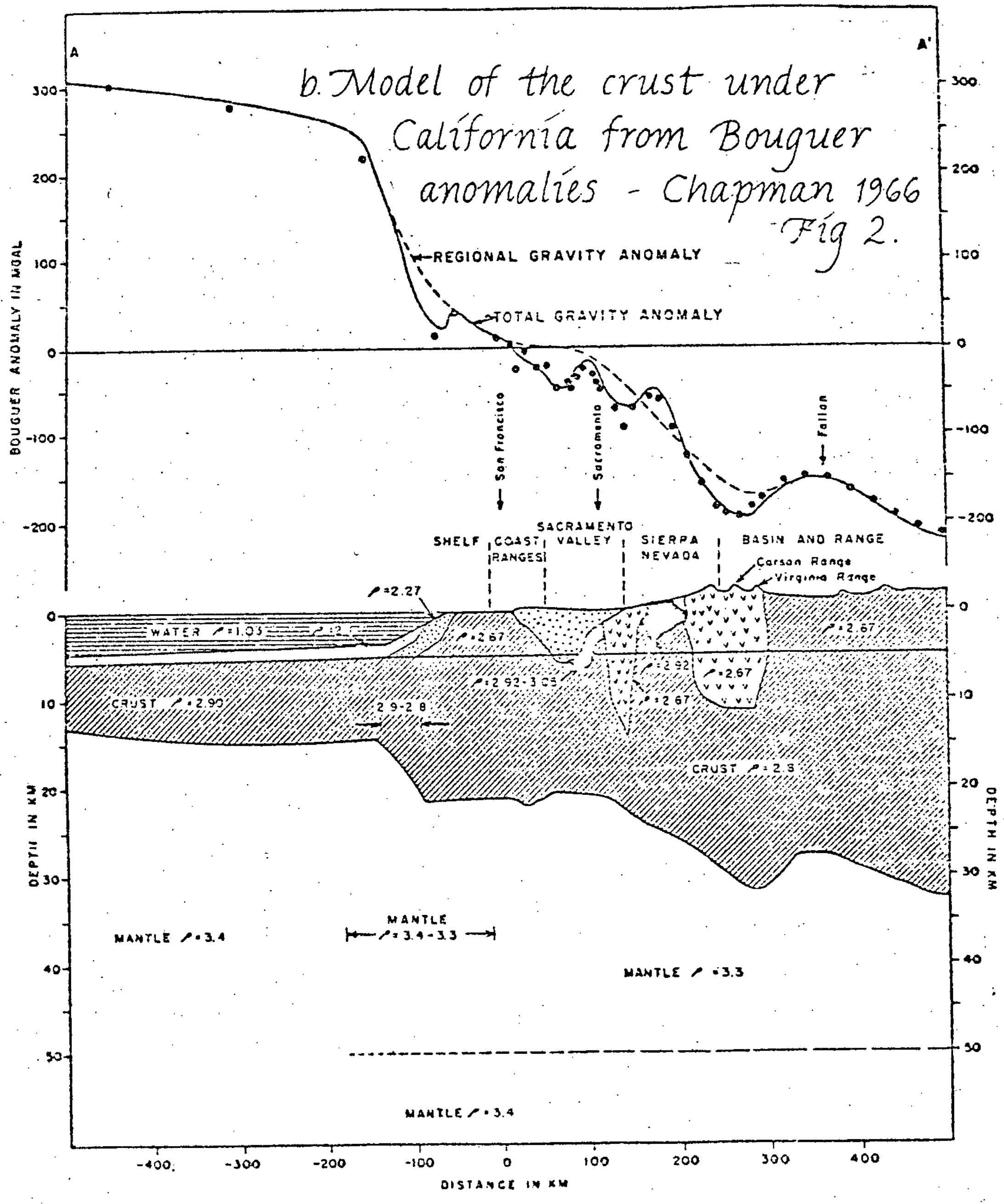



Isostatic anomaly map
Washington

Daneš 1969 Fig 2.




a. Seismic profile across western California
Eaton 1966 Fig 5




b. Model of the crust under California from Bouguer anomalies - Chapman 1966 Fig 2.

Figure 54a from Eaton (1966) shows two alternate interpretations of early $P_N$ arrivals across the Great Valley. One involves thinning of the whole crust; the other a tapering out of the "granitic"layer towards the western side of the valley. Bouguer data (Figure 54b) indicate crustal thicknesses under the Great Valley that are compatible with the seismic data. The low over the western side of the valley and the high over the east are probably the result of variations in the upper crust. The Bouguer field does not point to major departures from isostasy such as those observed over the Puget Lowland.

iv. Summary

The western seaboard of the United States and Canada is generally anomalous in lithospheric structure. Upper mantle velocities tend to be low - 7.7 to 7.9. The crust in many areas is much thinner than average (Woollard's standard sea-level crust is 33 km. thick). There are no roots under the Cascades or Coast Mountains. The Coastal Trough generally seems compensated in British Columbia and California: crustal thinning may be responsible. Negative free-air anomalies over the trough in Washington and Oregon may indicate that it is undercompensated, i.e. that a mass deficiency underlies it. The possible implications of these data are discussed in the final chapter.

## CHAPTER IX: SYNTHESIS AND CONCLUSIONS

Three aspects of the origin of Georgia Depression - timing, structural style, and cause - have been considered in this study. Of the three, only timing is well-known. Georgia Depression began to subside in the Upper Cretaceous, as is shown by the distribution of the Nanaimo Group. The structures that governed the Upper Cretaceous subsidence have so far been impossible to separate from post-Nanaimo Group structures (c.f. Muller and Jeletzky 1972; Rimme 1973; Simmons 1973; Hudson 1974; Sturdavant 1975); therefore the structural style of the original depression remains obscure.

This study has focussed on the cause of the Upper Cretaceous depression and its relation to the tectonic history of southwestern British Columbia. Hypotheses for its origin include rifting, lithospheric collapse over a paleosuture, and subsidence coupled to the uplift of the Coast Plutonic Complex. There is no evidence for a rift zone below Georgia Depression. Heat flow in the Georgia Strait area is subnormal to normal at present (Hyndman 1976); metamorphic assemblages in Upper Cretaceous - Tertiary rocks are not indicative of high geothermal gradients. Volcanics form an insignificant part of the Upper Cretaceous - Tertiary section in Georgia Depression.

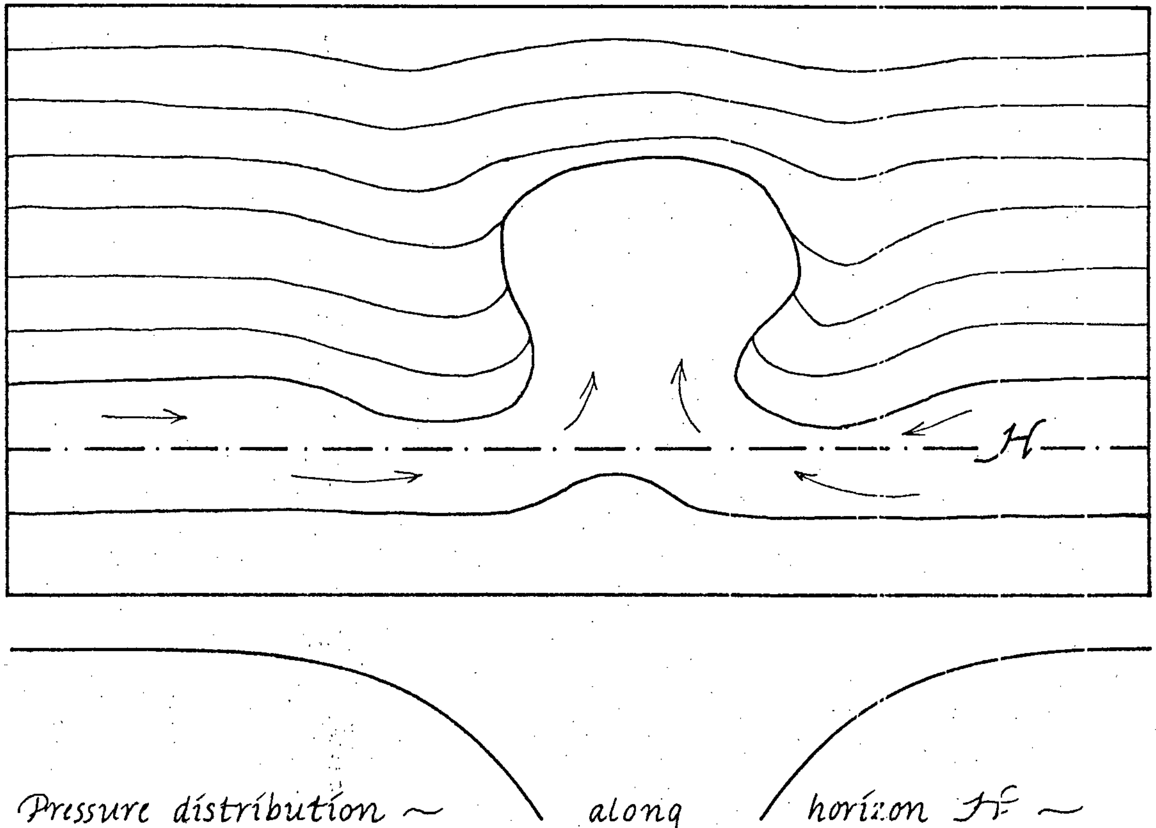
The major argument for a suture near the Insular Belt/ Coast Plutonic Complex province boundary is that the Paleozoic and Triassic of Vancouver Island seem to be allochthonous with respect to North America. However, such a suture more probably lies in the Shulaps Terrane, which is characterized by strong deformation, juxtaposition of "oceanic" and terrigenous sediments, and Upper Triassic/Lower Jurassic ultramafite emplacement, rather than in the eastern Insular Belt. The boundary between the Insular Belt and the Coast Plutonic Complex on Hardwicke and West Thurlow Islands shows no Upper Triassic-Lower Jurassic terrigenous sediments or strong deformation ascribable to a pre-plutonic suturing event. Instead, it is a straightforward and relatively undisturbed intrusive contact. In general the boundary between the two provinces is the magmatic front of the Coast Plutonic Complex.

Georgia Depression lies immediately west of the western limit of dominantly plutonic terrane. It began to subside in the Upper Cretaceous, approximately at the same time that the southern Coast Plutonic Complex began to rise, as shown by lack of post-Albian marine sediments within it and westerly-derived Albian clastics in the Tyaughton Trough. Prior to the Upper Cretaceous the site of Georgia Depression was a generally positive area undergoing erosion. The coincidence of opposite vertical movements in Georgia Depression and the Coast Plutonic Complex is strong evidence that they were coupled.

This conclusion also fits the other segments of the Coastal Trough - Hecate Depression, the Puget-Williamette Lowlands, and the Great Valley. The Coastal Trough lies between a series of colinear magmatic arcs - the Coast Plutonic Complex, the Cascade Province, and the Sierra Nevada - and the western edge of the North American Plate. Its various segments do not have a common structural style - the Puget Lowland is largely fault controlled, while only gentle tilting affected the eastern Great Valley. Neither faulting or warping emerges as the dominant mechanism of subsidence. The sub-basins of the Coastal Trough originated at different times, the Great Valley in the Upper Jurassic - Lower Cretaceous, Georgia Depression in the Upper Cretaceous, and Hecate Depression and the Puget - Williamette Lowlands in the Late Cenozoic. Yet in each case the beginning of subsidence roughly coincided with the uplift of the arc immediately to the east. This coincidence is best explained by conpled vertical movements of arc and trough. The Central Graben of Chile, lying west of the Andean arc on Precambrian-Paleozoic basement, may have a similar origin.

The most obvious mechanism of vertical movement coupling is horst-and graben tectonics, as exemplified by the Tetons and Jackson Hole. The Coastal Trough is not a convincing graben, however. Although major faults are possible, downwarping has played an important role in its formation, at least in Oregon and California. Thus a more complicated mechanism of interaction is likely. One possibility is suggested by the experiments of Ramberg (1967). In his centrifuged models, domes forming from a flat source region are accompanied by marginal synclines (Figure 55). The rise of material into the dome creates a pressure deficiency in its vicinity into which the overlying layer collapses. A rising mountain system could create a similar

Fig. 55 Rising diapir with marginal synclines




Pressure distribution ~ along horizon H ~

pressure deficiency around itself. It is proposed here that the subsidence of the Coastal Trough was initiated by lithospheric collapse into the "hole" generated during the uplift of the adjacent arc. Such a depression, since it is driven by a pressure/mass deficiency, is not isostatic. Negative free-air and isostatic anomalies should be associated with it. It will be recalled that negative free-air/isostatic anomalies are observed over the Puget and Williamette Lowlands, the youngest sections of the trough. This anisostatic condition should be eliminated in time by continued mantle flow. Yet the Great Valley and Georgia Depression have been in existence since the Cretaceous. Moreover Hecate Depression and possibly the Great Valley are approximately in isostatic equilibrium: some mechanism of compensation must perpetuate the trough once it is created. Besides sediment loading, two such mechanisms are possible: crustal thinning either by erosion at the base or by extension (Artemjev and Artyushkov 1971), and phase changes, for example the gabbro-eclogite transformation (Green and Ringwood 1967; Collette 1968). The seismic evidence so far available from British Columbia is compatible with either of both of these compensation models.

Elongate basins are a characteristic feature of arc-trench gaps. So far, vertical movements, uplifts, and depressions in arc-trench gaps have been explained in relation to the process of subduction (c.f. Lomnitz 1969; Fyfe and McBirney 1975). This study of the Coastal Trough has shown that some forearc basins are generated, not by subduction, but by events within the overriding plate itself. A better understanding of the relationship between arc magmatism and uplift and of the mechanisms of arc uplift will doubtless shed more light on the genesis of the Coastal Trough.

## BIBLIOGRAPHY

Allen, J.E., 1958. Columbia River Gorge - Portland to the Dalles, Guidebook for Field Trips. Geol. Soc. Am. Cordilleran Section, Eugene, Oregon, p.4.

Armstrong, R.L., J. Suppe, 1973. Potassium-argon geochronometry of Mesozoic igneous rocks in Nevada, Utah and southern California. Geol. Soc. Am. Bull.84, p.1375.

Artemjev, M.E., E.V. Artyushkov, 1971. Structure and isostasy of the Baikal Rift and the mechanism of rifting. Jour. Geophys. Res. 76, p.1197.

Asihene, K.A.B., 1970. The Texada Formation of B.C. and its associated magnetite concentrations. unpub. PhD - UCLA.

Baer, A.J., 1973. Bella Coola - Laredo Sound map-areas, B.C. Geol. Surv. Can. Memoir 372.

Bailey, E.H., W.P. Irwin, D.L. Jones, 1964. Franciscan and related rocks, and their significance in the geology of western California. Calif. Div. Mines Geol. Bull. 183.

Baker, B.H., P.A. Mohr, L.A.J. Williams, 1972. Geology of the eastern rift system of Africa. Geol. Soc. Am. Special Paper 136, p.67.

Baldwin, Ewart, 1964. Geology of Oregon. 2nd edition. J.W. Edwards, pub.

Barbat, W.F., 1971. Megatectonics of the Coast Ranges, Calif. Geol. Soc. Am. Bull. 82, p. 1541.

Beaulieu, J.D., 1971. Geologic formations of western Oregon. Ore. Dept. Geol. Min. Res. Bull. 70.

Beloussov, V.V., 1969. Continental rifts in P.J. Hart,ed., The Earth's Crust and Upper Mantle. Amer. Geophys. Un. monograph 13. p. 539.

Berg, J.W., Lynn Trembly, D.A. Emilia, J.R. Hult, J.M. King, L.T. Long, W.R. McKnight, S.K. Sarmah, R. Sonders, J.V. Thiruvathukal, D.A. Vassler, 1966. Crustal refraction profile, Oregon Coast Range. Seis. Soc. Am. Bull. 56, p.1357.

Berry, F.A., 1973. High fluid potentials in the Calif. Coast Ranges and their tectonic significance. A.A.P.G. Bull.57, p.1219.

Berry, M.J., W.R. Jacoby, E.R. Niblett, R.A. Stacey, 1971. A review of geophysical studies in the Canadian Cordillera. Can. Jour. Earth Sci. 8,p.788.

Berry, M.J., D.A. Forsyth, 1975. Structure of the Canadian Cordillera from seismic refraction and other data. Can. Jour. Earth Sci. 12, p.2.

Blunden, R.H. Vancouver's Downtown (Coal) Peninsula. Unpub. B.Sc. UBC 1971

Bostrom, R.C., 1968. The ocean ridge system in the northeast Pacific Ocean. Pac. Geol. 1, p.21.

Branery, R.W., P.D. Snavely, 1964. Geologic interpretation of reconaissance gravity and aeromagnetic surveys in northwestern Oregon USGS Bull. 1181 - N.

Brown, R.D. and E.I. Rich, 1967. Implications of two Cretaceous mass transport deposits, Sacramento Valley, California. Jour. Sed. Pet. p.37.

Buddington, A.F., D.H. Lindsley, 1964. Iron titanium oxide minerals and synthetic equivalents. Jour. Petrol. 5, p.310.

Cameron, B.E.B. and J.W.H. Monger,1971. Middle Triassic conodonts from the Fergusson Group, N.E. Pemberton map-area, B.C. Geol. Surv. Can. paper 71.1B, p.94.

Cameron, B.E.B., 1975. Geology of the Tertiary rocks north of latitude 49°, West Coast of Vancouver Island. Geol. Surv. Can. paper 75.1A, p.17.

Carder, D.S.,1970. Trans-California seismic profile; Owens Valley to Monterey. Geol. Soc. Am. Abs. V2, p.2.

Carder, D.S.,1973. Trans-California seismic profile, Death Valley to Monterey Bay. Seis. Soc. Am. Bull. 63, p.571.

Carlisle, D., T. Suzuki,1965.Structure, stratigraphy and paleontology of an upper Triassic section on the West Coast of B.C. Can. Jour. Earth Sci. 2, p. 445.

Carlisle, D., T. Suzuki, 1974. Emergent basalt and submergent carbonate-clastic sequences including the Upper Triassic Dilleri and Welleri zones on Vancouver Island. Can. Jour. Earth Sci. 11, p. 254.

Caron, Mike, 1974. Geology and geochronology of the Strachan Creek area, Howe Sound, S.W. British Columbia. Unpub. B.Sc., UBC.

Cashman, P.H., 1974. Melange terrane in the western Paleozoic and Triassic subprovince, Kalamath Mountains, California. Geol. Surv. Am. abs. 6, p.153.

Chaney, R.W., 1925. The Mascall Flora - its distribution and climatic relation. Carnegie Inst. Pub. 349, p.23.

Chapman, R.H., 1966. The gravity field in northern California. In: Geology of Northern California, Bull. 190, Calif.Div. Mines, Geol. p.395.

Chatterjee, N.D., 1967. Experiments on the phase transition cc + wo + ep = gros - and + $CO_2$ + $H_2O$. Contrib. Mineral. Petr.

Church, S.E., G.R. Tilton, 1973. Lead and strontium isotopic studies in the Cascade Mountains, bearing on andesite genesis. Geol. Soc. Am. Bull. 84, p.431.

Clark, E.D., 1964. Stratigraphy and structure of part of the western Sierra Nevada metamorphic belt, California. U.S.G.S. Prof. Paper 410.

Coates, J.A., 1974. Geology of the Manning Park area, British Columbia. Geol. Surv. Can., Bull. 238.

Cole, M., 1973. Petrology and dispersal patterns of Jurassic and Cretaceous sedimentary rocks in the Methow River area, North Cascades, Washington. Unpub. Ph.D., U. of Washington.

Collette, B.J., 1968. On the subsidence of the North Sea area. Geology of Shelf Seas (Proc. 14th Inter-University Geol. Cong.), D.T. Donovan, ed. Oliver and Boyd, Edinburgh.

Coulbourn, W.J., R. Moberly, 1975. Structure of a basin of the upper continental margin of southern Peru and northern Chile. 13th Pac. Sci. Congress proc. p. 395.

Culbert, R.R., 1972. Abnormalities in the distribution of K, Rb and Sr in the Coast Mountains batholith, British Columbia. Geochim, Cosmochim. Acta 36, p.1091.

Daly, R.A., 1912. Geology of the North American Cordillera at the 49th parallel. Geol. Surv. Can. Memoir 38, p.840.

Danes, L.F., 1969. Gravity results in western Washington. E.O.S. 50, p.548.

Danes, L.F., M. Bonno, E. Brau, W.D. Gilman, T.F. Hoffma, D. Johansen, M.H. Jones, B. Malfait, J. Masten, G. Hagne, 1965. Geophysical investigation of the southern Puget Sound area, Washington. Jour. Geophys. Res. 70, p.5573.

Dehlinger, P., E.F. Chiburts, M.M. Collver, Local travel-time curves and their geologic implications for the Pacific N.W. states. Seis. Soc. Am., Bull.55, p.587.

Detling, L.E., 1968. Historical background of the flora of the Pacific Northwest. Bull. 13, Museum of Natural History, University of Oregon.

Dewey, J.F. and J.M. Bird, 1970. Mountain belts and the new global tectonics. Jour. Geophys. Res. 75, p.2625.

Dickinson, W.R., 1971. Clastic sedimentary sequences deposited in shelf, slope and trough settings between magmatic arcs and associated trenches. Pac. Geol. 3, p.15.

Dickinson, W.R., 1973. Reconstruction of past arc-trench systems from petrotectonic assemblages in the island arcs of the western Pacific. In The Western Pacific, P.J. Coleman, ed., p.569.

Dickinson, W.R., 1974. Plate tectonics and sedimentation, in Tectonics and Sedimentation, W.R. Dickinson, ed. Soc. Econ. Paleontol. Mineral. Spec. Pub. 22, p.1.

Eaton, J.P., 1966. Crustal structure in northern and central California from seismic evidence. Calif. Div. Mines. Geol. Bull. 190, p.419.

Ernst, W.G., 1971. Metamorphic zonations on presumably subducted lithospheric plates from Japan, California and the Alps. Contrib. Mineral Petr. 34, p. 43.

Fenneman, N.M., 1931. Physiography of the Western United States (McGraw-Hill).

Fiske, R.S., C.A. Hopson, A.C. Waters, 1963. Geology of Mt. Rainier National Park, Washington. U.S.G.S. prof. paper 444.

Forsyth, D.A.F., M.J. Berry, R.M. Ellis, 1974. A refraction survey across the Canadian Cordillera at $54^{o}$N. Can. Jour. Earth Sci. 11, p.533.

Fyfe, W.S., A.R. McBirney, 1975. Subduction and the structure of andesitie volcanic belts. Am. Jour. Sci. 275-A, p.285.

Girdler, R.W., J.D. Fairhead, R.C. Searlt, W.T.C. Sowerbutts, 1969. Evolution of rifting in Africa. Nature 224 - Dec.20, p. 1178.

Glover, S.L., 1936. Hammer Bluff Formation of western Washington. Abs. Pan - Amer. Geol. 65-1, p.77.

Gordon, T.M., H.J. Greenwood, 1971. The stability of grossularite in $H_2O$ - $CO_2$ mixtures. Amer. Mineral. 56, p.1674.

Green, D.H., A.E. Ringwood, 1967. An experimental investigation of the gabbro to eclogite transformation and its petrological implications. Geochim. Cosmochim. Acta 31, p.767.

Greenwood, H.J., 1975. Genesis of magmas in intrusive rocks and related mineralization of the Canadian Cordillera. abs. Geol. Surv. Can. Cordilleran Section Symposium. p.13.

Hackel, O., 1966. Summary of the geology of the Great Valley. California Div. Mines. Geol. Bull. 190, p.217.

Hamilton, Warren, 1974. Subduction - melange wedges of modern Circum. Pacific arcs and of Cretaceous California. Geol. Soc. Am. abs. 6, p.187.

Hamilton, W. The Volcanic Central Andes - a modern model for the Cretaceous batholiths and tectonics of W. North America - in Proc. Andesite Conf. Oregon Dept. Mines Bull. 65, p.175. A.R. McBirney, ed. 1969.

Hammond, P., 1961. Structure and stratigraphy of the Keechelus Volcanic Group and associated Tertiary volcanic rocks in the west-central Cascade Range, Washington. Unpub. PhD, U.Washington.

Hart, S.R., 1964. The petrology and isotopic - mineral age relations of a contact zone in the Front Range, Colorado. Jour. Geol. 72, p.493.

Holdaway, M.J., 1971. Stability of andalusite and the aluminum silicate phase diagram. Am. Jour. Sci. 271, p.97.

Holland, S.S., 1964. Landforms of B.C. - a physiographic outline. B.C. Dept. Mines. Bull. 48.

Hollister, L.S., 1969. Metastable paragenetic sequence of andalusite, kyanite and sillimanite, Kwoick area, British Columbia. Am. Jour. Sci., p. 267.

Hopkins, V.S. Jr. 1966. Palynology of Tertiary rocks of Whatcom Basin, southwestern British Columbia and northwestern Washington. Unpub. PhD, UBC.

Hsu, K.J.,1971. Franciscan melanges as a model for eugeosynclinal sedimentation and underthrusting tectonics. Jour. Geophys. Res. 76, p. 1162.

Hudson, Jon Park, 1974. Stratigraphy and paleoenvironments of the Cretaceous rocks, North and South Pender Islands, British Columbia. Unpub. M.Sc. Oregon State University.

Hutchison, W.W., 1970. Metamorphic framework and plutonic styles in the Prince Rupert region of the Central Coast Mountains, British Columbia. Can. Jour. Earth Sci. 7, p.376.

Hutchison, W.W., A.V. Okulitch, H.C. Berg, 1973. Skeena River 1:1,000,000 map sheet, Geol. Surv. Can. open file map.

Hyndman, R.D., 1976. Heat flow measurements in the inlets of southwestern B.C. Jour. Geophys. Res. 81, p.337.

Irving, E. and R.W. Yole, 1972. Paleomagnetism and the kinematic history of mafic and ultramafic rocks in fold mountain belts. In: The Ancient Oceanic Lithosphere, Pub. Earth Phys. Branch. Energy, Mines, Resources V 42 - n 3, p.87.

Irwin, W.P., 1973. Sequential minimum ages of oceanic crust in accreted tectonic plates of northern California and southern Oregon. Geol. Soc. Am. abs. 5, p.62.

Jaeger, J.C., 1957. The temperature in the neighborhood of a cooling intrusive sheet. Am. Jour. Sci. 255, p.306.

Jaeger, J.C., 1959. Temperatures outside a cooling intrusive sheet. Am. J. Sci. 257, p.44.

Jeletzky, J.A., 1965. Age and tectonic nature of the Georgia Strait seaway. Geol. Surv. Can. paper 65 - 2, p.72.

Johnson, S.H., R.W. Couch, 1970. Crustal structure in the North Cascades of Washington and British Columbia from seismic refraction measurements. Seis. Soc. Am. Bull. 60, p. 1259.

Johnson, S.H., R.W. Couch, M. Gemperle, E.R. Banks, 1972. Seismic refraction measurements in S.E. Alaska and western British Columbia. Can. Jour. Earth Sci 0, p. 1756.

Jolly, W.J., R.E. Smith, 1972. Degradation and metamorphic differentiation of the Keweenawan tholeitic lavas of northern Michigan, U.S.A. Jour. Pet 13, p.273.

Kerrick, D.M., 1970. Contact metamorphism in some areas of the Sierra Nevada, Calif. Geol. Soc. Am. Bull. 81, p.2913.

Kerrick, D.M., K.E. Crawford, A.F. Randazzo, 1973. Metamorphism of calcareous rocks in three roof pendants in the Sierra Nevada, California.Jour. Petr. 14, p.303.

Kuniyoshi, S., 1972a. Petrology of the Karmutsen (volcanic) Group, northeastern Vancouver Island, British Columbia. Unpub. PhD, UCLA.

Lanphere, M.A., B.L. Reed, 1973. Timing of Mesozoic and Cenozoic plutonic events in Circum-Pacific North America. Geol. Surv. Am. abs. 5, p.70.

Lecouteur, P.C. and D.J. Templeman-Kluit, 1976. Rb/Sr ages and a profile of initial $Sr^{87}/Sr^{86}$ ratios for plutonic rocks across the Yukon Crystalline Terrane. Can. Jour. Earth Sci. 13-2, p. 319.

Leech, G. B., 1953. Geology and mineral deposits of the Shulaps Range, B.C. Dept. of Mines, Bull. 32.

LePichon, X., J. Francheteau, J. Bonnin, 1973. Plate Tectonics. Elsevier, Amsterdam.

Liou, J.G., 1970. Stability relations of zeolites and related minerals. Unpub. PhD, UCLA

Liou, J.G., 1973. Synthesis and stability relations of epidote, $Ca_2 Al_2 Fe Si_3 O_{12} (OH)$. Jour. Pet. 14, p.381.

Liou, J.G., S. Kuniyoshi, K. Ito, 1974. Experimental studies of the phase relations between greenschist and amphibolite in a basaltic system. Am. Jour. Sci. 274, p. 613.

Livingston,V.E.,1971. Geology and mineral Resources of King County, Washington. Wash. Div. Mines Geol. Bull. 63.

Livingston, V.E., H. Gower, 1975. quoted in The Seattle Times Feb. 23, 1975, p.B.1.

Lofgren, G.E., C.H. Donaldsen, 1975. Curved branching crystals and differentiation in comb layered rocks. Contrib. Mineral. Petr. 49, p.309.

Lomnitz, C., 1969. Seafloor spreading as a factor of tectonic evolution in southern Chile. Nature p. 222.

Lowes, B.E., 1972. Metamorphic petrology and structural geology of the area east of Harrison Lake, B.C. Unpub. PhD, U. Washington.

Mathews, W., 1972. Geology of the Vancouver area of British Columbia. I.G.C. Field Excursion A05-C05, I.G.C. 24th Session Montreal.

Mathews, W. H., J.G. Fyles, H.W. Nasmith, 1970. Post-glacial crustal movements in southwestern British Columbia and adjacent Washington State. Can. Jour. Earth Sci. 7, p.690.

Mattinson, J.M., 1973. Ages of zircons from the North Cascade Mountains, Washington. Geol. Surv. Can. B-83, p.2769.

McBirney, A.R., J.F. Sutter, H.R. Naslund, 1974. Episodic volcanism in the Central Oregon Cascade Range. Geology V.no.12, p.585.

McKee, Bates, 1972. Cascadia: The Geologic Evolution of the Pacific Northwest. McGraw-Hill, New York.

McKillop, G.R., 1973. Geology of southwestern Gambier Island, Howe Sound, British Columbia. Unpub. B.Sc., U.B.C.

Milanovsky, E.E., 1972. Continental rift zones - their arrangement and development. Tectonophys. 15, p.65.

Misch, P. 1966. Tectonic evolution of the Northern Cascades of Washington State. Can. Inst. Min. Met., spec. vol. 8, p. 101.

Miyashiro, A., 1961. Evolution of metamorphic belts. Jour. Petrol. 2, p.277.

Monger, J.W.H., 1970. Hope map-area, west half, British Columbia. Geol. Surv. Can. paper 69-47, p.75.

Monger, J.W.H., C.A. Ross, 1971. Distribution of fusilinaceans in the Western Canadian Cordillera. Can. Journ. Earth Sci. 8, p. 2.

Monger, J.W.H., J.G. Souther, H. Gabrielse, 1972. Evolution of the Canadian Cordillera - a plate-tectonic model. Am. Journ.Sci. 272, p.577.

Moore, J.G., J.P. Lockwood, 1973. Origin of comb layering and orbicular structure, Sierra Nevada Batholith, California. Geol. Soc. Am Bull. 84, p.1.

Moores, E.M., 1972. Model for Jurassic island arc - continental margin collision in California. Geol. Soc. Am. abs 4, p.202.

Muller, J.E., 1972. Geological Recon. Map of Vancouver Island and Gulf Islands, Geol. Surv. Can. open file map 61.

Muller, J.E., 1973. Lower Cretaceous flysch sequence on the west coast of Vancouver Island. Geol. Soc. Am. abs. 5, p.84.

Muller, J.E. 1974. Victoria map-area, Pacific Rim National Park, Vancouver Island, British Columbia in Geol. Surv. Can, paper 74-1, p. 21.

Muller, J.E., 1976. Cape Flattery map area, British Columbia. Geol. Surv. Can. paper 76-1A, p. 107.

Muller, J.E., J.A. Jeletsky, 1970. Geology of the Upper Cretaceous Nanaimo Group,Vancouver Island and Gulf Islands, British Columbia. Geol. Surv. Can.paper 69, p.25.

Mullineux, D.R., L.M. Gard, D.R. Crandell, 1959. Continental sediments of Miocene age in Puget Sound lowlands, Washington. A.A.P.G. Bull. 43, p.688.

Newton, R.C., 1966. Some calc-silicate equilibrium reactions. Am. Jour. Sci. 264, p.204.

Northcote, K.E., J.E. Muller, 1972. Volcanism, plutonism and mineralization - Vancouver Island. Can. Mining Met. Bull. October 1972 p.1.

Ojakangas, R.W., 1968. Cretaceous sedimentation, Sacramento Valley, Calif. Geol. Soc. Am. Bull. 79, p.973.

Page, R.J., 1974. A tectonic melange on the west coast of Vancouver Island, British Columbia. Geol. Soc. Am. abs. 6, p.233.

Pakiser, L.C., 1974. Root of the Sierra Nevada. Geol.Soc. Am., Cordilleran Section. abs. 6-3, p.234.

Peck, D.L., A.B. Griggs, H.G. Schlicker, F.G. Wells, H.M. Dolo, 1964. Geology of the central and northern parts of the western Cascade Range in Oregon. USGS prof. paper 449.

Pigage, L.C., 1973. Metamorphism southwest of Yale, British Columbia. Unpub. M.Sc., University of British Columbia.

Presnall, D.C., P.C. Bateman, 1973. Fusion relations in the system $NaAl Si_3O$ - $CaAl_2Si_2O_8$ - $KALSi_3O_8$ - $SiO_2$ - $H_2O$, and the generation of granitie magmas in the Sierra Nevada batholith. Geol. Soc. Am. Bull. 84, p.3181.

Ragan, D.M., 1968. Structural Geology - An Introduction to Geometrical Techniques, Wiley and Sons, New York.

Ramberg, H., 1967. Gravity, Deformation and the Earth's Crust. Academic Press, N.Y.

Raney, J.A., 1973. Great Valley - Coast Range contact - An alternate hypothesis. Geol. Soc. Am. abstr. with prog. 1973, p.93.

Rau, W.W., 1967. Geology of the Wynoochee Valley quadrangle, Grays Harbor Co., Washington. Washington Dept.Nat.Res. Bull.56.

Reamsbottom, S., 1974. Geology and metamorphism of the Mount Breakenridge area, Harrison Lake, British Columbia. Unpub. Ph.D. University of British Columbia.

Repenning, C.A., 1960. Geologic summary of the Great Valley of California with reference to disposal of liquid radioactive waste. U.S.G.S., T.E.I. Rept. 769.

Rinne, R.W., 1973. Geology of the Duke Point - Kulleet Bay Area, Vancouver Island, British Columbia. Unpub. M.Sc. Oregon State.

Roberts, A.E., 1958. Geology and coal resources of the Toledo - Castle Rock district, Cowlitz and Lewis Counties, Washington. U.S.G.S. Bull. 1062.

Roddick, J.A., 1965. Vancouver North, Coquitlam and Pitt Lake map areas, with special emphasis on the evolution of plutonic rocks. Geol.Surv.Can. Memoir 335.

Roddick, J.A., 1966. The coast crystalline belt of B.C. Tectonic History and Mineral Deposits of the Western Cordillera. Can.Inst.Min.Met. spec. vol.8, p.73.

Roddick, J.A., W.W. Hutchison, 1972. Plutonic and associated rocks of the Coast Mountains of British Columbia. 15th Geol. Cong. 24th Can. Guidebook for Field Excursions A.C.O. 4.

Roddick, J.A., W.W. Hutchison, 1973. Examination of the shoreline between Knight Inlet and Howe Sound. Geol. Surv. Can. paper 73-1A p.42.

Roddick, J.A., W.W. Hutchison, 1973. Pemberton (East-half) map area, B.C. (92 J.E. ½). Geol. Surv. Can. paper 73. p. 17.

Roddick, J.A., W.W. Hutchison, 1974. Setting of the Coast Plutonic Complex, B.C. Pacific Geol. 8, p. 91.

Roddick, J.A., J.E. Muller, A.V. Okulitch, 1973. Fraser River 1:1,000,000 map sheet. Geol. Surv. Can. open file map.

Rogers, W.P., 1970. A geological and geophysical study of the central Puget Lowland. unpub. Ph.D. U. Washington.

Ross, D.A., G.G. Shor, Jr., 1965. Reflection profiles across the Middle America Trench. Jour. Geophys. Res. 70, p.5551.

Rouse, G.E., W.H. Mathews, R.H. Blunden, 1975. The Lions Gate member: a new Late Cretaceous sedimentary subdivision in the Vancouver area of B.C. Can. Jour. Earth Sci. 12, p.3.

Ryan, B., 1973. Geology and Rb/Sr geochronology of the Anarchist Mountain area, south central British Columbia. Unpub. PhD., U.B.C.

Schweickert, R.A., D.S. Cowan, 1974. Pre-Tithonian magmatic arcs and subduction zones, Sierra Nevada, California. Geol. Soc. Am. abs. 6-3, p.251.

Shido, F., 1958. Plutonic and metamorphic rocks of the Nakoso and Iritono districts in the central Abukuma Plateau. Jour. Fac. Sci. Tokyo University XI, pt. 2, p.131.

Shouldice, D.H., 1971. Geology of the Western Canadian Continental Shelf. Can. Petroleum Geol. Bull. 19-2, p.405.

Sibley, D.F., T.A. Vogel, B.M. Walker, G. Byerly, 1976. The origin of oscillatory zoning in plagioclase: a diffusion and growth controlled model. Am. Jour. Sci. 276-3, p.273.

Simmons, Michael Lee,1973. The stratigraphy and paleoenvironment of Thetis, Kuper and adjacent islands, British Columbia. unpub. M.Sc., Oregon State.

Snavely, P.D., H.C. Wagner, 1963. Tertiary geologic history of Western Oregon and Washington. Washington Div. Mines. Geol. Rept. Inv. 22.

Snavely, P.D., N.S. MacLeod, H.C. Wagner, 1968. Tholeiitic and alkalic basalts of the Eocene Siletz River Volcanics, Oregon Coast Range. Am.Jour.Sci. p.266.

Stacey, R.A., 1973. Gravity anomalies, crustal structure, and plate tectonics in the Canadian Cordillera. Can.Jour.Earth Sci. 10, p.615.

Stacey, R.A., L.E. Stevens, 1969. An interpretation of gravity measurements on the west coast of Canada. Can.Jour.Earth Sci. 6, p. 463.

Stewart, R.J., R.J. Page, 1974. Zeolite facies metamorphism of the Late Cretaceous Nanaimo Group, Vancouver Island and Gulf Islands, British Columbia. Can.Jour.Earth Sci. 11, p.280.

Storre, B., K. Nitsch, 1972. Die reaktion 2zo + 1 $CO_2 \rightleftharpoons$ 3an + 1 cc + 1 $H_2O$. Contrib. Mineral Petrol. 35, p.1.

Sturdavant, C.D., 1975. Sedimentary environments and structure of the Cretaceous rocks of Saturna and Tumbo Islands, British Columbia. Unpub. M.Sc. Oregon State.

Surdam, R.C., 1973. Low-grade metamorphism of tuffaceous rocks in the Karmutsen group, Vancouver Island, British Columbia, Bull. 84, p.1911.

Sutherland Brown, A., 1966. Tectonic History of the Insular Belt of British Columbia. Can.Inst.Min.Met. Spec. Vol. 8, p. 83.

Sutherland Brown, A., 1968. Geology of the Queen Charlotte Islands. B.C. Dept. Mines, Bull. 54.

Sutter, J.F., A.R. McBirney, 1974. Periods of Cenozoic volcanism in the Cascade Province of Oregon. Geol. Soc. Am. abs. 6-3, p. 264.

Symons, D.T.A., 1971. Paleomagnetism of the Jurassic Island Intrusions of Vancouver Island, British Columbia. Geol. Sur. Can. paper 70-63, p.1.

Symons, D.T.A., 1971. Paleomagnetic notes on the Karmutsen basalts, Vancouver Island, British Columbia. Geol. Surv. Can. paper 71-24, p.9.

Symons, D.T.A., 1972. Paleomagnetism of the Triassic Guichon Batholith and rotation in the Interior Plateau, British Columbia. Can.Jour.Earth Sci. 8, p.1388.

Symons, D.T.A., 1973. Unit correlations and tectonic rotation from paleomagnetism of the Traissic Copper Mountain Intrusions, British Columbia. Geol. Surv. Can. paper 73-19, p.11.

Tabor, R.W., 1972. Age of the Olympic metamorphism, Washington: K-Ar dating of low-grade metamorphic rocks, Geol. Soc. Am. Bull. 83, p.1805.

Tatel, H.E., M.A. Tuve, 1955. Seismic exploration of a continental crust.in: Crust of the Earth, A, Poldervaart, ed. Geol. Soc. Am. spec. paper 62, p.35.

Tennyson, M. L., 1974. Stratigraphy, structure and tectonic setting of Jurassic and Cretaceous sedimentary rocks in the west-central Methow-Pasayten area, N.E. Cascade Range, Washington and British Columbia. unpub. PhD, U.Washington.

Thiruvathukal, J.V., 1968. Regional gravity of Oregon. unpub. PhD, Oregon State.

Thiruvathukal, J.V., J.W. Berg, D.F. Heinrichs, 1970. Regional gravity of Oregon. Bull. Geol. Soc. Am. p. 725.

Tiffin, D. L. 1969. Continuous seismic reflection profiling in the Strait of Georgia, British Columbia. unpub. PhD, UBC

Tipper, H.W., 1969. Mesozoic and Cenozoic geology of the N.E. part of Mount Waddington map-area, Coast-District, British Columbia. Geol. Surv. Can. paper 68, p.33.

Tseng, K. 1968. A new model for the crust in the vicinity of Vancouver Island. unpub. MSc., U.B.C.

Verhoogen, John, F.J. Turner, L.E. Weiss, C. Wahrhaftig, 1970. The Earth. Holt, Rinehart and Winston, New York.

Wager, L.R., and Brown, G.M., 1968. Layered Igneous Rocks. Oliver and Boyd. Edinburgh.

Walcott, R.L., 1972. Gravity, flexure and the growth of sedimentary basins at a continental edge. Geol. Soc. Am. 83, p.1845. .

Wells, F.G., D.L. Peck, 1961. Geologic map of Oregon west of the 121 st. meridian, State of Oregon Dept. Geol. Mineral Res. Misc. Geol. Investigations Map 1-325.

White, W.R.H., J.C. Savage, 1965. A seismic refraction and gravity study of the earth's crust in B.C. Seis. Soc. Am. Bull. 55, p.463.

Williams, Harold, M.J. Kennedy, E.R.W. Neale, 1972. The Appalachian Structural Province, in: Variations in Tectonic Styles in Canada . Geol. Ass. Can. Spec. Paper 11, p. 118.

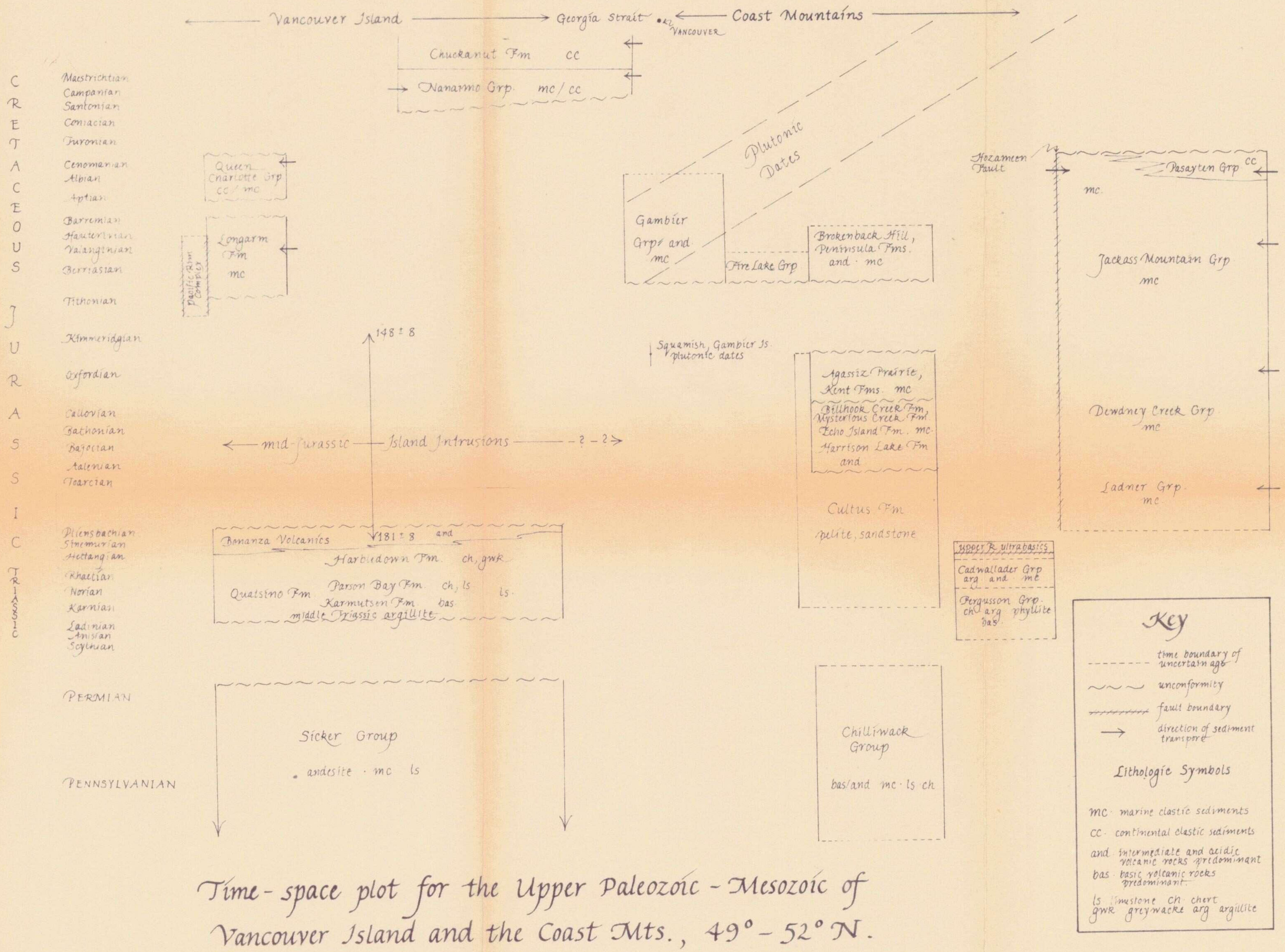
Wilshire, H.G., 1969. Mineral layering in the Twin Lakes Granodiorite, Colorado. Geol. Soc. Am. Memoir 115, p.235.

Woodsworth, G., 1974. Geology of the Mount Raleigh area, British Columbia. unpub. PhD, Princeton University.

Woollard, G. P. 1969. Regional variations in gravity. A.G.U. memo. 13, p.320.

Yorke, D. and Farguhar, R.M., 1972. The Earth's Age and Geochronology. Pergamon Press, Oxford.

# Plate 1




Time-space plot for the Upper Paleozoic - Mesozoic of Vancouver Island and the Coast Mts., 49° - 52° N.

Plate 2


GEOLOGY of HARDWICKE and WEST THURLOW ISLANDS, B.C.

J. Nelson 1974