GEOLOGIC AND TECTONIC EVOLUTION OF ANNETTE, GRAVINA, DUKE, AND SOUTHERN PRINCE OF WALES ISLANDS, SOUTHEASTERN ALASKA

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ABSTRACT

Annette, Gravina, Duke, and southern Prince of Wales Islands are underlain primarily by Cambrian (and perhaps Proterozoic) through Triassic volcanic, sedimentary, plutonic, and metamorphic rocks. These rocks belong to the Alexander terrane, which is a coherent tectonic fragment that underlies much of southeastern (SE) Alaska, the Saint Elias Mountains of British Columbia, Yukon, and eastern Alaska, and coastal regions of west-central British Columbia. Geologic mapping combined with U-Pb (zircon) geochronologic studies have delineated the major geologic units and features of these islands, and contribute to our understanding of the geologic and tectonic evolution of the Alexander terrane.

The oldest rocks recognized on Annette, Gravina, Duke, and southern Prince of Wales Islands consist of greenschist- and amphibolite-facies metavolcanic and metasedimentary rocks of the Wales metamorphic suite. These rocks are locally intruded by dioritic and granodioritic metaplutonic rocks which yield U-Pb apparent ages of approximately 540-520 Ma (Middle and Late Cambrian). Rocks in the Wales suite were therefore deposited, at least in part, prior to Late Cambrian time, but their maximum depositional age is not known. The Wales suite and associated metaplutonic rocks are intruded by large dioritic to granitic plutons which yield U-Pb apparent ages in the 475-425 Ma (Middle Ordovician-Early Silurian) range and are probably overlain by Lower Ordovician-Lower Silurian volcanic and sedimentary rocks of the Descon Formation. However, depositional contacts between the Descon Formation and the older metamorphic rocks have not been demonstrated. Deformation, metamorphism, and uplift of rocks in the Wales metamorphic

suite occurred during a Middle Cambrian-Early Ordovician tectonic event which I have referred to as the Wales "orogeny." The term orogeny is used informally in this instance, as little is known about the regional and tectonic significance of this event.

Ordovician-Early Silurian rocks on these islands are interpreted to have formed in an oceanic volcanic arc environment based on similarities with young or presently active volcanic arcs in the Circum-Pacific region. Characteristics of the volcanic-plutonic complex in the Alexander terrane which are similar to those in other magmatic belts include: 1) predominance of basaltic to andesitic volcanic rocks and dioritic to granodioritic and subordinate granitic plutonic rocks, 2) calc-alkaline affinity of the plutonic and volcanic rocks, as defined on AFM, FeO*/MgO versus SiO_2, and La versus Nb diagrams and by an alkali-lime index of 56-62, 3) patterns of strong (50-100 times chondrites) light REE enrichments, moderate (5 to 20 times chondrites) heavy REE enrichments, and strong negative europium anomalies, 4) evolution of the magmatic system over a period of approximately 50 m.y., and 5) increasing potassium content with time in the plutonic rocks.

Facies relations in Ordovician-Silurian strata in the southern part of the terrane generally record northwesterly paleogeographic trends, indicating that the interpreted arc trended oblique to the north-northwesterly elongation of the terrane. Continuation of Ordovician-Silurian shallow-marine strata for over 600 km to the north-northwest indicates that the interpreted arc probably faced to the southwest and that the strata to the north accumulated in a back-arc environment. Protoliths of the Wales metamorphic suite may also have formed in a volcanic arc environment, but the penetrative deformation and regional metamorphism of these rocks precludes detailed analyses of protolith relations and composition.

During middle Silurian-earliest Devonian time the Early Silurian and older rocks in the area were involved in a major tectonic event which I refer to as the Klakas orogeny. Manifestations of this orogeny on Annette, Gravina, Duke, and Prince of Wales Islands include: 1) cessation of the Ordovician-Early Silurian volcanism and plutonism, 2) deposition of middle and Upper Silurian polymictic conglomerate on northern Prince of Wales Island and regions to the north, 3) erosion or non-deposition of Silurian strata on Annette, Gravina, and Duke Islands and on central and southern Prince of Wales Island, 4) southwest-directed movement on thrust faults on southern Prince of Wales Island and perhaps on Annette Island, 5) deposition and deformation of a Lower Devonian talus breccia and penetrative brecciation of Ordovician rocks along thrust faults on southern Prince of Wales Island, 6) greenschist- and local amphibolite-facies regional metamorphism and penetrative deformation of Ordovician-Early Silurian rocks on Annette, Gravina, and Duke Islands, 7) emplacement, and perhaps generation by anatexis, of Late Silurian trondhjemite, sodic leucodiorite, and subordinate granite plutons, 8) several kilometers (perhaps as much as 10 km) of uplift of Late Silurian and older rocks prior to middle Early Devonian time, and 9) deposition of Lower Devonian conglomeratic red beds of the Karheen Formation in topographically rugged subaerial environments in some regions to the south, and in a northward tapering clastic wedge to the north. Previous workers recognized the stratigraphic manifestations of this orogenic event on central and northern Prince of Wales Island, but most relations to the south were recognized initially during this study.

On southern Prince of Wales Island, Lower Devonian conglomeratic strata are overlain by shallow-marine limestone, mudstone, and siltstone of middle Early Devonian age, which are in turn overlain by deeper-water mudstone and graptolitic shale. Subsidence of the region below sea level following the Klakas orogeny therefore occurred during middle Early Devonian time and produced a marine transgression on a north-facing paleoslope. Lower Devonian strata on Annette, Gravina, and Duke Islands were deposited in shallow-marine environments, and only locally include polymictic conglomerate and coarser clastic strata. Andesitic volcanic rocks of probable Early Devonian age locally overlie the marine clastic strata and are the youngest Paleozoic rocks in the study area.

Triassic strata herein referred to as the Hyd Group unconformably overlie the Devonian and older rocks on Annette and Gravina Islands. At the base of the section in most areas is a thick conglomerate or sedimentary breccia with meter-size clasts of Devonian and older rock in a poorly sorted matrix. These strata are overlain by a sequence, from bottom to top, of rhyolite and rhyolitic tuff, shallow-marine limestone, calcareous siltstone and limestone, and basalt flows and breccia. A U-Pb apparent age of 225 ± 3 on the rhyolite combined with megafossil and conodont ages demonstrate that these strata were deposited during Late Carnian to Late Norian time, and place a minimum age constraint of 225 ± 3 Ma on the Carnian-Norian boundary. A large body of pyroxene gabbro on Duke Island yields a U-Pb apparent age of 226 ± 3 Ma, which demonstrates that this gabbro is not genetically related to the zoned ("Alaskan-type") ultramafic bodies on Duke Island (assuming that the ultramafic bodies are indeed Cretaceous in age!). Rather, the pyroxene gabbro is interpreted to be genetically related to the Triassic basaltic rocks.

Intrusive relations indicate that hornblende gabbro on northeastern Duke Island is pre-Late Silurian in age, and is therefore not genetically related to the pyroxene gabbro or to the Cretaceous(?) ultramafic rocks.

The unconformity at the base of the Triassic section records a major latest Paleozoic(?)-Triassic uplift and erosional event in the Alexander terrane, but this event was not associated with regional deformation or metamorphism. This lack of deformation combined with the occurrence of Triassic strata along the eastern margin of the terrane in SE Alaska and the bimodal (basalt-rhyolite) composition of the volcanic rocks suggest that the Triassic strata and their subjacent unconformity formed in an extensional environment. A major low-angle normal fault on southern Prince of Wales Island (the Keete Inlet fault) may also have moved during this interpreted extensional event.

Jurassic and younger rocks intrude and overlie rocks in various terranes in western British Columbia and southern Alaska and demonstrate that the Alexander terrane has been adjacent to Wrangellia since Middle(?) Jurassic time, and to terranes to the east since Late Cretaceous-early Tertiary time. Regional sub-greenschist- to greenschist-facies metamorphism and moderate deformation of Cretaceous and older rocks along the eastern margin of the terrane are interpreted to have occurred during mid-Cretaceous-early Tertiary juxtaposition of the Alexander terrane against terranes to the east.

North of Annette, Gravina, Duke, and southern Prince of Wales Islands the Alexander terrane is underlain primarily by Paleozoic marine clastic strata and limestone. Lower Paleozoic strata in some regions of the Saint Elias Mountains include Cambrian volcanic rocks which may be correlative with rocks in the Wales metamorphic suite, and a thick section of Ordovician-Devonian clastic strata and limestone. Upper Paleozoic clastic strata are widespread in the Saint Elias Mountains region but occur in only a few areas of southeastern Alaska, where they were generally deposited in tectonically stable, shallow-marine environments. Triassic strata to the north are generally similar to rocks on Annette and Gravina Islands, and are interpreted by other workers to have been deposited in a rift environment.

A variety of geologic, paleomagnetic, and paleobiogeographic evidence suggests that the Alexander terrane occupied low paleolatitudes during much of Paleozoic and Mesozoic time, and did not reach its present latitude in the Cordillera until after Early Cretaceous time. Previous hypotheses were that the Alexander terrane was originally adjacent to rocks in the Sierra-Klamath region of California, and that both assemblages formed and evolved adjacent to the California continental margin. Comparison of the geologic and tectonic evolution of the Alexander terrane with that in the Sierra-Klamath region indicates, however, that the two assemblages have little in common and probably were not closely associated during Paleozoic time.

Alternatively, I suggest that the early Paleozoic geologic and tectonic evolution of the Alexander terrane is remarkably similar to that in a dismembered orogenic belt which occurs in southeastern Australia (Lachlan Fold Belt), New Zealand, the Transantarctic Mountains and Byrd Land of Antarctica, and perhaps in tectonic fragments in Asia. Similarities between the Alexander terrane and the Lachlan Fold Belt include: 1) arc-type(?) volcanism and sedimentation during Cambrian time (and perhaps Proterozoic time in the Alexander terrane), 2) regional deformation and metamorphism of the Cambrian and older(?) rocks during Middle Cambrian-Early Ordovician time, 3) evolution of some regions in a volcanic arc environment during Ordovician time (into Early Silurian time in the Alexander terrane), 4) cessation of this volcanic arc activity during the onset of a Silurian-earliest Devonian orogenic event, which is manifest by regional uplift and erosion, deformation and regional metamorphism, anatectic(?) plutonism (and volcanism in the Lachlan Belt), 5) deposition of Lower Devonian and locally Silurian conglomeratic red beds, and 6) evolution in relatively stable marine environments from middle Early through Middle Devonian time.

Comparison of paleolatitudes of the Alexander terrane (determined from paleomagnetic data) with paleolatitudes of eastern Australia (interpreted from continental reconstructions) indicates that the two regions occupied similar paleolatitudes from Ordovician to Late Devonian time. A similar comparison of declination data from the Alexander terrane indicates that both regions also rotated in a clockwise sense during this period. There are also similarities in lower Paleozoic fossils of the two regions, but some faunas from the Alexander terrane apparently bear stronger affinities with North American or Asian fossils.

Based on the geologic, paleomagnetic, and, to some degree the paleobiogeographic similarities, I raise the possibility that the Alexander terrane formed and evolved along the paleo-Pacific margin of Gondwana, perhaps adjacent to rocks in eastern Australia, during early Paleozoic time. The data are not sufficient to draw correlations between the Alexander terrane and specific regions in this complex orogen, although I note that similarities are strongest with the Molong volcanic province in the Lachlan Belt of eastern Australia. The paleomagnetic data indicate that the terrane could have been associated with these rocks or with potential northern correlatives in tectonic fragments that now reside in Asia.

The geologic, paleomagnetic, and paleobiogeographic(?) similarities between the Alexander terrane and the Lachlan Belt end in Middle Devonian-Early Carboniferous time. During this time the Lachlan Belt apparently underwent a major rifting episode, and the Alexander terrane began to evolve in tectonically stable marine environments. The paleolatitudes of the two regions also diverge at this time, with the Alexander terrane migrating northward toward the paleo-equator and eastern Australia continuing its southward movement. Carboniferous fauna from the Alexander terrane are reported by some workers to have "Tethyan" affinities, a fact that is consistent with the low paleolatitudes determined from the paleomagnetic data. Triassic faunas from the terrane are endemic to equatorial or perhaps more southerly regions in the eastern part of the paleo-Pacific basin, and paleomagnetic data from the terrane are most consistent with a paleolatitude of approximately 43° South. In concert with the hypothesis that the terrane was adjacent to the paleo-Pacific margin of Gondwana during early Paleozoic time, I raise the possibility that the terrane was tectonically removed from the Gondwana margin, perhaps by rifting, during Middle Devonian-Early Carboniferous time, and migrated eastward across the paleo-Pacific basin during late Paleozoic time. Northward displacement apparently began after Late Triassic time, and ended during the mid-Cretaceous to early Tertiary juxtaposition of the terrane against fragments previously accreted to western North America.

TABLE OF CONTENTS

	Page
ACKNOWLEDGMENTS	**ii**
ABSTRACT	**iv**
LIST OF FIGURES	**xvi**
LIST OF TABLES	**xix**
LIST OF PLATES	**xx**
CHAPTER 1: INTRODUCTION	**1**
REFERENCES	6
CHAPTER 2: GEOLOGY OF SOUTHERN PRINCE OF WALES ISLAND	8
ABSTRACT	8
INTRODUCTION	9
Previous work	15
GEOLOGIC FRAMEWORK	16
Overview	16
Wales metamorphic suite	17
Descon Formation	23
Middle Ordovician-Early Silurian plutonic rocks	27
Overview	27
Diorite and Granodiorite	27
Quartz diorite	42
Quartz monzonite, granite, and quartz syenite	43
Interpretation of the Ordovician-Early Silurian plutonic rocks	44
Ordovician metaplutonic complex	45
Silurian leucodiorite	52
Karheen Formation	53
Conglomeratic strata (lower part of section)	54
Sandstone, siltstone, and limestone (middle part of section)	58
Mudstone and black shale (upper part of section)	60
Volcanic and hypabyssal rocks	60
Regional variations in the Devonian strata	61
Bokan Mountain Granite (Jurassic)	63
Cretaceous diorite and granodiorite	64
STRUCTURAL GEOLOGY	65
Thrust faults	65
Keete Inlet fault	69
Strike-slip faults	70
Summary of movement on faults	71
DISCUSSION OF U-Pb GEOCHRONOLOGIC DATA	72
GEOLOGIC AND TECTONIC HISTORY	74
ACKNOWLEDGMENTS	77
REFERENCES	78

CHAPTER 3: GEOLOGY OF ANNETTE, GRAVINA, AND DUKE ISLANDS 82
ABSTRACT 82
INTRODUCTION 83
Previous work 85
GEOLOGIC FRAMEWORK 85
Overview 85
Wales metamorphic suite (pre-Ordovician) 86
Cambrian metaplutonic rocks 92
Descon Formation (Ordovician-Early Silurian) 101
Ordovician-Early Silurian diorite 102
Ordovician-Early Silurian hornblende gabbro 111
Trondhjemite to granite (Late Silurian) 112
Klakas orogeny (middle Silurian-earliest Devonian) 113
Karheen Formation and associated volcanic rocks (Lower Devonian) 114
Hyd Group (Upper Triassic) 120
Late Triassic pyroxene gabbro 124
Jurassic-Lower Cretaceous sedimentary and volcanic rocks 125
Cretaceous(?) ultramafic rocks 125
DISCUSSION OF THE U-Pb GEOCHRONOLOGIC DATA 126
GEOLOGIC HISTORY 128
CONCLUSIONS 131
ACKNOWLEDGMENTS 132
REFERENCES 132

CHAPTER 4: GEOLOGIC AND TECTONIC EVOLUTION OF THE ALEXANDER TERRANE IN SOUTHERN SOUTHEASTERN ALASKA 136
ABSTRACT 136
INTRODUCTION 137
GEOLOGY OF ANNETTE, GRAVINA, DUKE, AND SOUTHERN PRINCE OF WALES ISLANDS 141
Overview 141
Pre-Ordovician metamorphic complex 141
Ordovician-Early Silurian rocks 145
Silurian rocks and the Klakas orogeny 148
Lower Devonian strata 153
Upper Triassic rocks 154
Bokan Mountain Granite (Jurassic) 156
Jurassic-Lower Cretaceous strata 156
Late Cretaceous-early Tertiary metamorphism and deformation 157
GEOLOGY OF THE ALEXANDER TERRANE IN THE SAINT ELIAS MOUNTAINS AND IN CENTRAL AND NORTHERN SE ALASKA 157
Overview 157
Lower Paleozoic rocks 159
Upper Paleozoic rocks 160
Upper Triassic strata 161
TECTONIC EVOLUTION OF THE ALEXANDER TERRANE 161
Cambrian (Proterozoic?) through Early Devonian time 161
Middle Devonian through Early Permian time 164
Late Permian and Triassic time 165

DISPLACEMENT HISTORY OF THE ALEXANDER TERRANE 165
Geologic evidence and previous syntheses 165
Evolution during Paleozoic time 165
Evolution during Mesozoic time 169
Paleomagnetic data 169
Paleontologic data 170
Discussion of previous syntheses 170
Derivation from the Sierra-Klamath region 172
Evolution near its present position in the Cordillera 176
Location during Mesozoic time 176
Comparison with other orogenic belts 177
Geologic and tectonic evolution of the Lachlan Fold Belt 178
Comparison between the Alexander terrane and the Lachlan Fold Belt 184
Speculative displacement history 186
SUMMARY 189
Summary of the geologic and tectonic evolution of the Alexander terrane 189
Summary of the displacement history of the Alexander terrane 191
ACKNOWLEDGMENTS 192
REFERENCES 192

APPENDIX 1: GEOLOGIC MAP OF SOUTHERN PRINCE OF WALES ISLAND 200
INTRODUCTION 200
DESCRIPTION OF MAP UNITS 202
Sedimentary and (or) volcanic rocks 205
Regionally metamorphosed and (or) deformed rocks 224
Intrusive rocks 233
STRUCTURAL GEOLOGY 255
Thrust faults 256
Shipwreck Point, Bird Rocks, and Ruth Island faults 256
Frederick Cove fault 258
Anchor Island fault 259
Keete Inlet fault 260
Strike-slip faults 261
Northwest- to north-northwest-striking faults 261
North- to north-northeast-striking faults 264
Cape Chacon fault 264
ACKNOWLEDGMENTS 265
REFERENCES 266

APPENDIX 2: U-PB GEOCHRONOLOGIC METHODS 268
LABORATORY METHODS 268
Zircon separation 268
Zircon dissolution and isolation of Pb and U 268
Mass spectrometry 269
DATA ANALYSIS 269
REFERENCES 273

APPENDIX 3: GEOLOGIC MAP OF SOUTHEASTERN ALASKA 274
INTRODUCTION 274
DESCRIPTION OF MAP UNITS 280
Lithostratigraphic units 280
Undivided map units 296
Lithodemic units 305
Intrusive units 311
ACKNOWLEDGMENTS 321
REFERENCES 321

LIST OF FIGURES

Page

CHAPTER 1

Figure 1-1. Sketch map showing: (a) the distribution of rocks in the Alexander terrane in the North American Cordillera and (b) the locations of Annette, Gravina, Duke, and southern Prince of Wales Islands in the Alexander terrane. 2

CHAPTER 2

Figure 2-1. Sketch map showing the location of the study area and the distribution of rocks belonging to the Alexander terrane (adapted from Monger and Berg, 1984). 10

Figure 2-2. Geologic sketch map of southern Prince of Wales Island (Dixon Entrance D-1, D-2, and C-1 and parts of adjacent 1:63,360 quadrangles). 12

Figure 2-3. Photograph of highly flattened fragments in a greenschist-facies metavolcanic rock of the Wales metamorphic suite. 18

Figure 2-4. Photograph of asymmetric folds in the metamorphic foliation of the Wales metamorphic suite. 20

Figure 2-5. Cross section across southern Klakas Inlet showing relations between faults and geologic units. 22

Figure 2-6. Well-preserved pillows in a basalt belonging to the Descon Formation. 24

Figure 2-7. Concordia diagram of zircon fractions from leucogranodiorite at sample locality 3. 37

Figure 2-8. Concordia diagram of zircon fractions from hornblende diorite at sample locality 4. 40

Figure 2-9. AFM diagram showing my samples of Ordovician-Early Silurian plutonic and volcanic rocks and the tholeiitic and calc-alkaline fields of Irvine and Baragar (1971). 46

Figure 2-10. $(FeO + .9xFe_2O_3)/MgO$ versus SiO_2 diagram for my samples of Ordovician-Early Silurian plutonic and volcanic rocks (all in weight percent). 47

Figure 2-11. La versus Nb (in ppm) for my samples of Ordovician-Early Silurian plutonic and volcanic rocks. 48

Figure 2-12. REE abundances relative to chondrites for my samples of Ordovician-Early Silurian plutonic and volcanic rocks. 49

Figure 2-13. Photograph of foliated diorite and crosscutting but foliated granodiorite. 51

Figure 2-14. Schematic stratigraphic sections and available conodont ages along the north-northwest-trending belt of Lower Devonian strata on southern Prince of Wales Island. 55

Figure 2-15. Large clast in the basal conglomerate of the Lower Devonian Karheen Formation on southernmost Prince of Wales Island. 57

Figure 2-16. Sedimentary breccia that overlies and contains clasts of highly deformed Ordovician volcanic and plutonic rocks, and is overlain by relatively nondeformed middle Lower Devonian mudstone and shale. 59

Figure 2-17. Stereoplot (equal-area) showing the trend and plunge of slickenside striae on the thrust faults and associated minor faults west of the Keete Inlet fault. 67

Figure 2-18. Penetratively brecciated Ordovician leucogranodiorite in southern Klakas Inlet. 68

CHAPTER 3

Figure 3-1. Location map of the study area and of rocks belonging to the Alexander terrane [derived from Berg and others (1978); Monger and Berg (1984)]. 84

Figure 3-2. Geologic sketch map of western Gravina Island and legend for Figures 3-2, 3-3, and 3-4 [in part modified from Berg (1973)]. 87

Figure 3-3. Geologic sketch map of southern Annette Island [in part modified from Berg (1972a)]. 89

Figure 3-4. Geologic sketch map of northern Duke Island, Hotspur Island, and southern Annette Island [in part modified from Irvine (1974)]. 90

Figure 3-5. Photograph of layered and foliated diorite and quartz diorite on small islands south of Gravina Island. 93

Figure 3-6. Sketch map showing the correlation of rocks on southern Gravina Island and on eastern Prince of Wales Island. 100

Figure 3-7. Photograph of a Late Silurian trondhjemite dike cutting across the foliation in diorite of probable Ordovician-Early Silurian age on southern Gravina Island. 104

Figure 3-8. $^{206}Pb^*/^{238}U$ versus $^{207}Pb^*/^{238}U$ 105 concordia diagrams for age samples 2(a), 3(b), 4(c), 6(d), and 7(e).

Figure 3-9. Schematic stratigraphic section of the Lower Devonian rocks on Hotspur Island (Fig. 3-4). 116

Figure 3-10. Photograph of limestone and sandstone olistoliths in the olistostromal layer on northwestern Hotspur Island (see Figures 3-4 and 3-9). 117

Figure 3-11. Schematic stratigraphic section of Upper Triassic strata on Annette and Gravina Islands. 121

Figure 3-12. Large angular clasts of Ordovician-Silurian plutonic rocks in Upper Triassic sedimentary breccia on southwestern Gravina Island. 123

CHAPTER 4

Figure 4-1. Location map of the Alexander terrane and regions referred to in the text. 138

Figure 4-2. Schematic columns depicting the geologic and tectonic evolution of the Alexander terrane in southern SE Alaska. 142

Figure 4-3. Silurian continental reconstruction (from Scotese and others, 1979) showing the distribution of early Paleozoic orogenic belts with which the Alexander terrane may have been associated. 179

Figure 4-4. Comparison of the apparent paleolatitudes of the Alexander terrane (assuming it was in the southern hemisphere) and of eastern Australia. 187

APPENDIX 2

Figure A2-1. Analyses of NBS lead standards showing their measured isotopic composition relative to values certified by the National Bureau of Standards (except for the $^{206}Pb/^{204}Pb$ ratio of NBS SRM 981, which is adapted from Hamelin and others, 1985). 271

APPENDIX 3

Figure A3-1. Index map of southeastern Alaska showing the primary sources of information used in compiling the geologic map. 275

LIST OF TABLES

Page

CHAPTER 2

Table 2-1. U-Pb (zircon) isotopic data and apparent ages. 28

Table 2-2. U-Pb geochronologic sample information. 31

Table 2-3. Major, minor, and trace element data. 33

CHAPTER 3

Table 3-1. U-Pb (zircon) isotopic data and apparent ages. 95

Table 3-2. U-Pb geochronologic sample information. 98

CHAPTER 4

Table 4-1. Fossil groups in the Alexander terrane and their apparent similarities with fossil groups from other regions. 171

APPENDIX 1

Table Al-1. Paleontologic samples and localities. 203

Table Al-2. Geochronologic samples and localities. 204

LIST OF PLATES

Plate 1. Geologic map of southern Prince of Wales Island, southeastern Alaska. (pocket)

Plate 2. Geologic map of southeastern Alaska. (pocket)

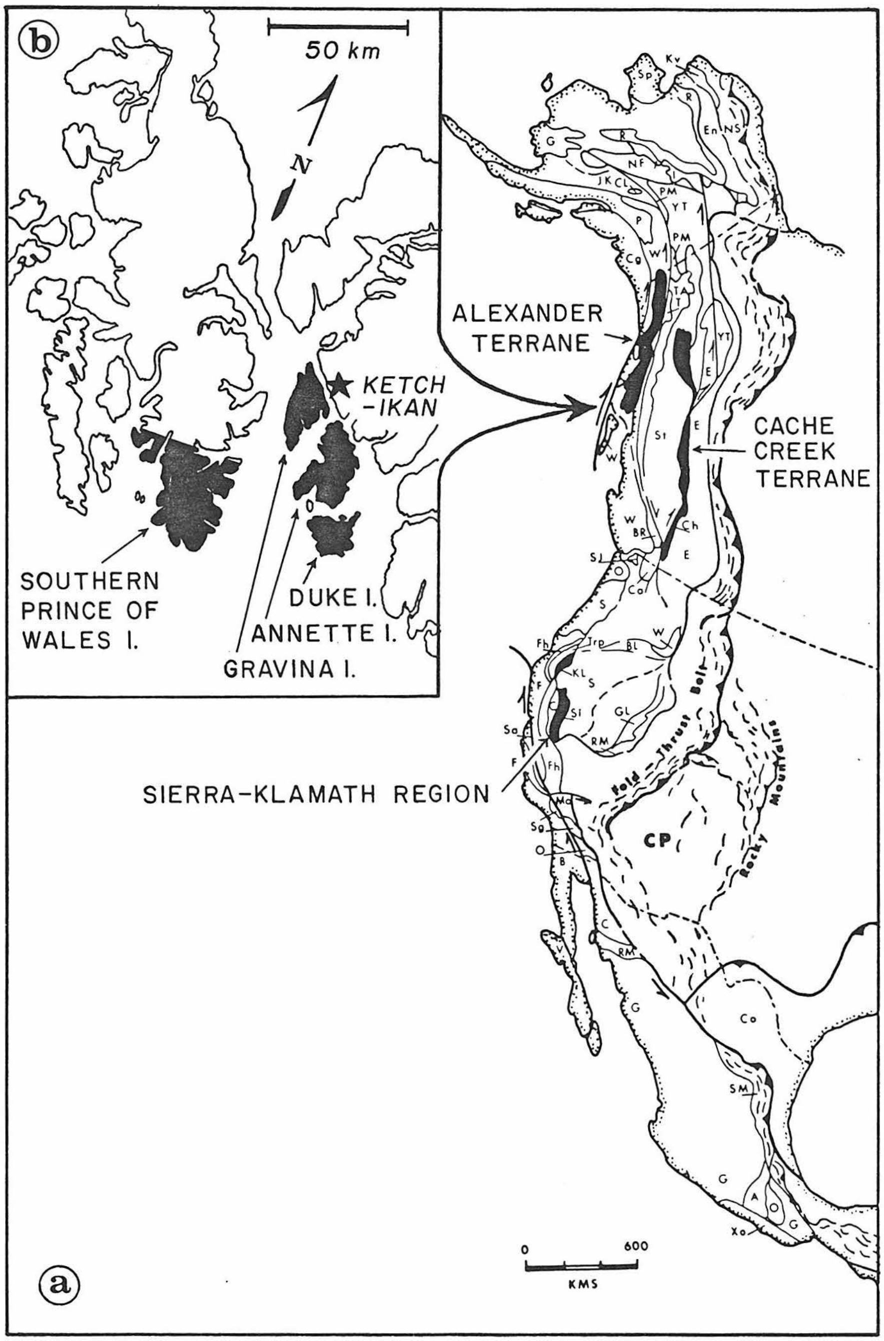
CHAPTER 1

INTRODUCTION

Annette, Gravina, Duke, and southern Prince of Wales Islands are underlain primarily by pre-Ordovician through Triassic plutonic, metamorphic, and stratified rocks that belong to the Alexander terrane (Fig. 1-1). Rocks in this terrane are anomalous in the North American Cordillera in that they occur outboard (west) of large regions in British Columbia in which upper Paleozoic-lower Mesozoic strata are the oldest rocks recognized. This led Schuchert (1923) to suggest that Paleozoic rocks in the Alexander terrane formed in a geosynclinal system (the "Alexandrian Embayment") which was separate from the main Cordilleran geosyncline.

Wilson (1968) initially recognized that the Alexander terrane is a discrete tectonic fragment in the Cordillera, and suggested that it was not accreted to western North America until after early Mesozoic time. The allochthonous nature of the terrane has since been demonstrated by many lines of evidence, including: 1) the occurrence of Permian fusulinid fauna of "Tethyan" affinity in the Cache Creek terrane of central British Columbia (Monger and Ross, 1971: Fig. 1-1), 2) separation of the Alexander terrane from terranes to the east by an assemblage of deep-marine strata of Late Jurassic-Early Cretaceous age (Berg and others, 1972), 3) paleomagnetic data from Paleozoic rocks in the terrane which indicate that it has been displaced northward relative to North America by at least 18° since Pennsylvanian time (Van der Voo and others, 1980), and 4) the equatorial affinity of Triassic bivalves in the Alexander terrane (Tozer, 1982).

Figure 1-1. Sketch map showing: (a) the distribution of rocks in the Alexander terrane in the North American Cordillera and (b) the locations of Annette, Gravina, Duke, and southern Prince of Wales Islands in the Alexander terrane. Also shown are the Cache Creek terrane in central British Columbia and the Sierra-Klamath region of California. Map of western North America is adapted from Coney (1981).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The displacement history of the Alexander terrane has been a subject of considerable controversey. Wilson (1968) originally suggested that the terrane was derived from Asia, but did not cite specific evidence to support this hypothesis. Following a general scenario outlined by Monger and Ross (1971), Jones and others (1972) reported that rocks in the Alexander terrane record a history which is quite similar to that in the Sierra-Klamath region of California, and suggested that rocks in these two regions formed and evolved along the California continental margin. This reported correlation of rocks in the Alexander terrane and the Sierra-Klamath region has influenced most subsequent displacement scenarios (e.g., Schweikert and Snyder, 1981). Comparison of the geology of the Alexander terrane and the Sierra-Klamath region has been inhibited, however, by the scarcity of published information about the geologic and tectonic evolution of the Alexander terrane.

The primary objective of my work in southeastern Alaska has been to reconstruct the geologic and tectonic evolution of the Alexander terrane in an effort to learn about its displacement history. My approach in reconstructing this evolution has centered on geologic mapping of Annette, Gravina, Duke, and southern Prince of Wales Islands (Fig. 1-1) and comparison of the geology of these areas to that in other parts of the terrane. These islands were selected because a variety of lower Paleozoic to Triassic rocks had been recognized by previous workers, and long fjords and rugged shorelines on these islands offer excellent exposures in an otherwise densely vegetated region. These geologic studies have served as the framework for U-Pb geochronologic studies [conducted with Jason Saleeby (Caltech)], analyses of conodonts [by

Norman Savage (University of Oregon)], major, minor, and trace element geochemical analyses [by the U.S. Geological Survey under the supervision of Fred Barker], studies of the occurrence and origin of massive sulfide deposits in the southern part of the terrane [with Henry C. Berg (formerly of the U.S. Geological Survey)], and studies of oxygen isotope variations in the terrane [with Hugh Taylor (Caltech)]. Comparison of the geology of these islands with other regions has been accomplished through compilation of a geologic map of southeastern Alaska under the supervision of Henry C. Berg.

This thesis is organized into three main chapters, which have been written in the form of separate manuscripts, and three appendices -- two of which are separate manuscripts. Each manuscript has a separate abstract, introduction, and reference list. Chapter 2 describes the geologic evolution of southern Prince of Wales Island, presents and discusses the U-Pb geochronologic and the geochemical data, and discusses the stratigraphic implications of the conodont ages. The geologic units and major structures on southern Prince of Wales Island are shown on Plate 1, and are described in detail in Appendix 1. Chapter 3 provides a description of the Triassic and older rocks on Annette, Gravina, and Duke Islands, presents the U-Pb geochronologic data from these islands, and discusses the conodont ages. The laboratory methods used in the geochronologic studies presented in Chapters 2 and 3 are described in Appendix 2. Plate 2 is a geologic map of southeastern Alaska -- the geologic units on this map are described in Appendix 3. Chapter 4 summarizes the geology of Annette, Gravina, Duke, and southern Prince of Wales Islands, integrates the geology of these areas with relations in other parts of the terrane, and discusses the tectonic implications of the geologic relations. This serves as the basis for a discussion of the displacement history of the terrane, which is presented in the second part of Chapter 4.

The fundamental conclusion of my study, as outlined in Chapter 4, is that rocks in the Alexander terrane record a history which is quite different from that in the Sierra-Klamath region, but is remarkably similar to that in parts of the Lachlan Fold Belt of southeastern Australia. Geologic, paleomagnetic, and, to a lesser degree paleobiogeographic data from the Alexander terrane support the hypothesis that the Alexander terrane formed and evolved along the paleo-Pacific margin of Gondwana during early Paleozoic time, migrated eastward across the paleo-Pacific basin during late Paleozoic time, and was juxtaposed against terranes previously accreted to western North America during mid-Cretaceous to early Tertiary time.

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CHAPTER 2

GEOLOGY OF SOUTHERN PRINCE OF WALES ISLAND

ABSTRACT

Geologic mapping combined with U-Pb geochronologic, geochemical, and conodont studies of southern Prince of Wales Island yield new information about the early Paleozoic geologic framework and evolution of the southern Alexander terrane. The oldest rocks recognized on southern Prince of Wales Island consist of late Proterozoic(?)-Cambrian metavolcanic and metasedimentary rocks (Wales metamorphic suite), which are interpreted to have been deposited in a volcanic arc environment. These rocks were deformed and metamorphosed during a Middle Cambrian-Early Ordovician tectonic event referred to informally as the Wales "orogeny." Beginning soon after this orogenic event and continuing into Early Silurian time, basaltic to rhyolitic rocks and subordinate clastic strata (Descon Formation) were deposited and large dioritic to granitic plutons were emplaced. Regional stratigraphic and intrusive relations and the calc-alkaline affinity of the plutonic and volcanic rocks suggest that they formed in an ensimatic volcanic arc along a convergent plate margin.

Volcanic arc activity ceased during Early Silurian time with the onset of a major middle Silurian-earliest Devonian tectonic event referred to as the Klakas orogeny. This orogeny is manifest on southern Prince of Wales Island as a change in style of magmatism, southwest-vergent movement on thrust faults, penetrative brecciation and pervasive hydrothermal alteration of pre-Devonian rocks in some areas, and several kilometers of structural uplift of Silurian and older rocks. Lower

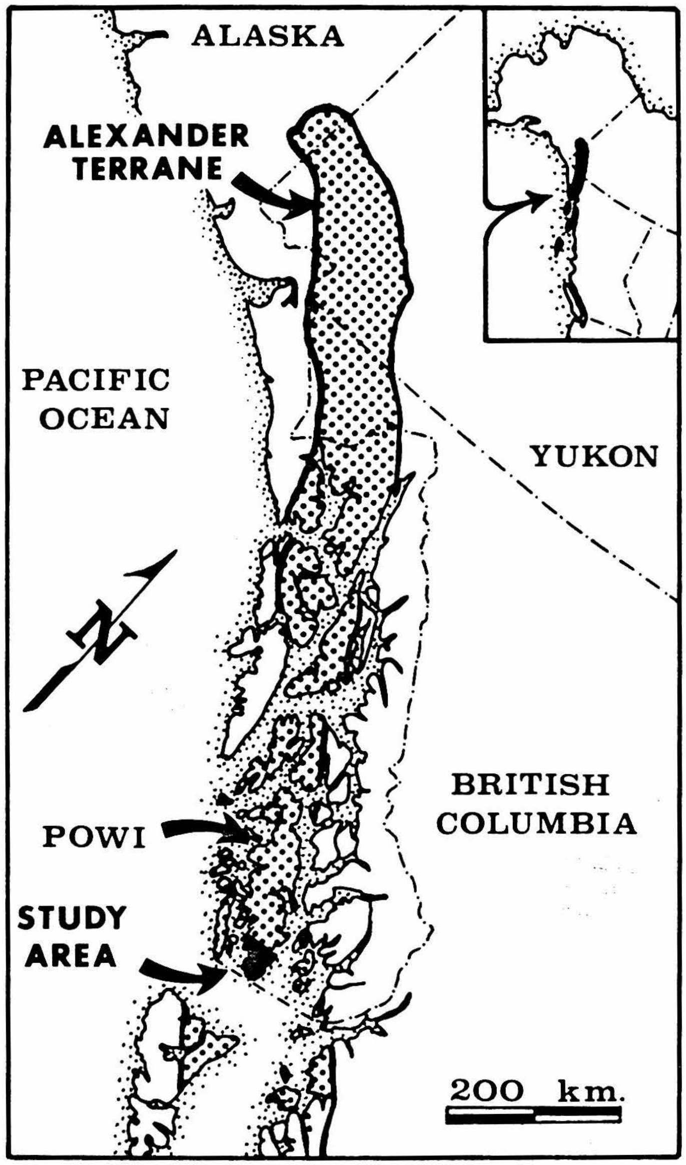
Devonian strata assigned to the Karheen Formation record the waning stages of this event: conglomeratic redbeds low in the section were deposited in topographically rugged, subaerial to shallow-marine environments, whereas middle Lower Devonian strata high in the section were deposited in marine environments after most of the region had subsided below sea level.

Post-Devonian rocks on southern Prince of Wales Island include the Bokan Mountain Granite (Jurassic) and mid-Cretaceous granodiorite and diorite bodies. Prior to emplacement of the mid-Cretaceous plutons, the Devonian and older rocks were offset along several sets of strike-slip faults, and Ordovician and younger rocks were juxtaposed over rocks of the Wales metamorphic suite along the Keete Inlet fault. The regional juxtaposition of younger rocks over older rocks combined with the sinuosity of its trace suggests that the Keete inlet fault is a low-angle normal fault.

INTRODUCTION

Southern Prince of Wales Island is underlain by a variety of sedimentary, volcanic, intrusive, and metamorphic rocks that range in age from late Proterozoic(?)-Cambrian to mid-Cretaceous. These rocks belong to the Alexander terrane, which underlies much of southeastern Alaska, coastal regions of west-central British Columbia, and the Saint Elias Mountains of British Columbia, southwestern Yukon, and eastern Alaska (Fig. 2-1: Berg and others, 1972). Ordovician-Silurian marine clastic strata and limestone underlie much of the Alexander terrane and are in many areas the oldest rocks exposed (Gehrels and Berg, 1984). On southern and central Prince of Wales Island, however, these strata grade into an Ordovician-Early Silurian volcanic-plutonic complex and intrude

Figure 2-1. Sketch map showing the location of the study area and the distribution of rocks belonging to the Alexander terrane (adapted from Monger and Berg, 1984). POWI = Prince of Wales Island.

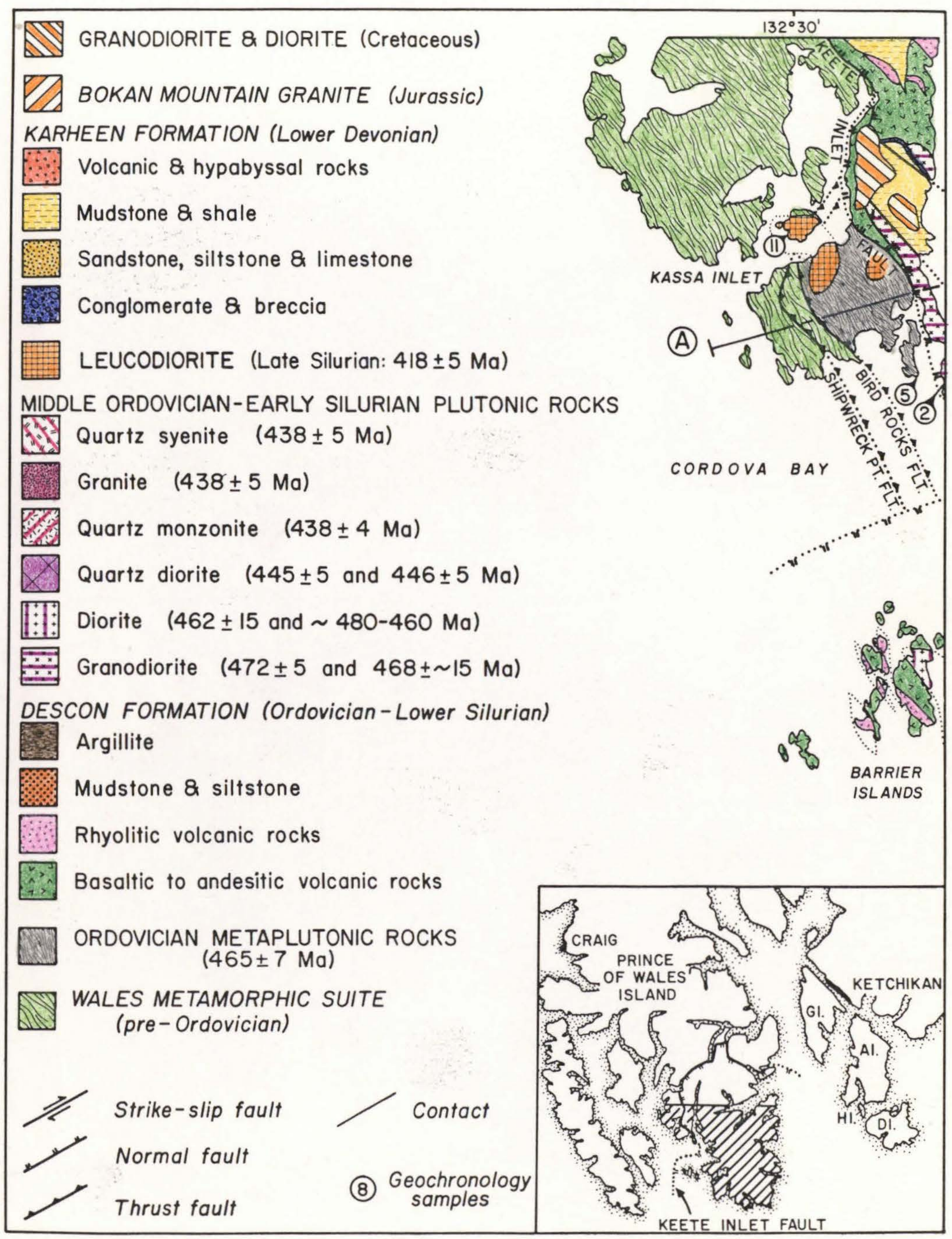


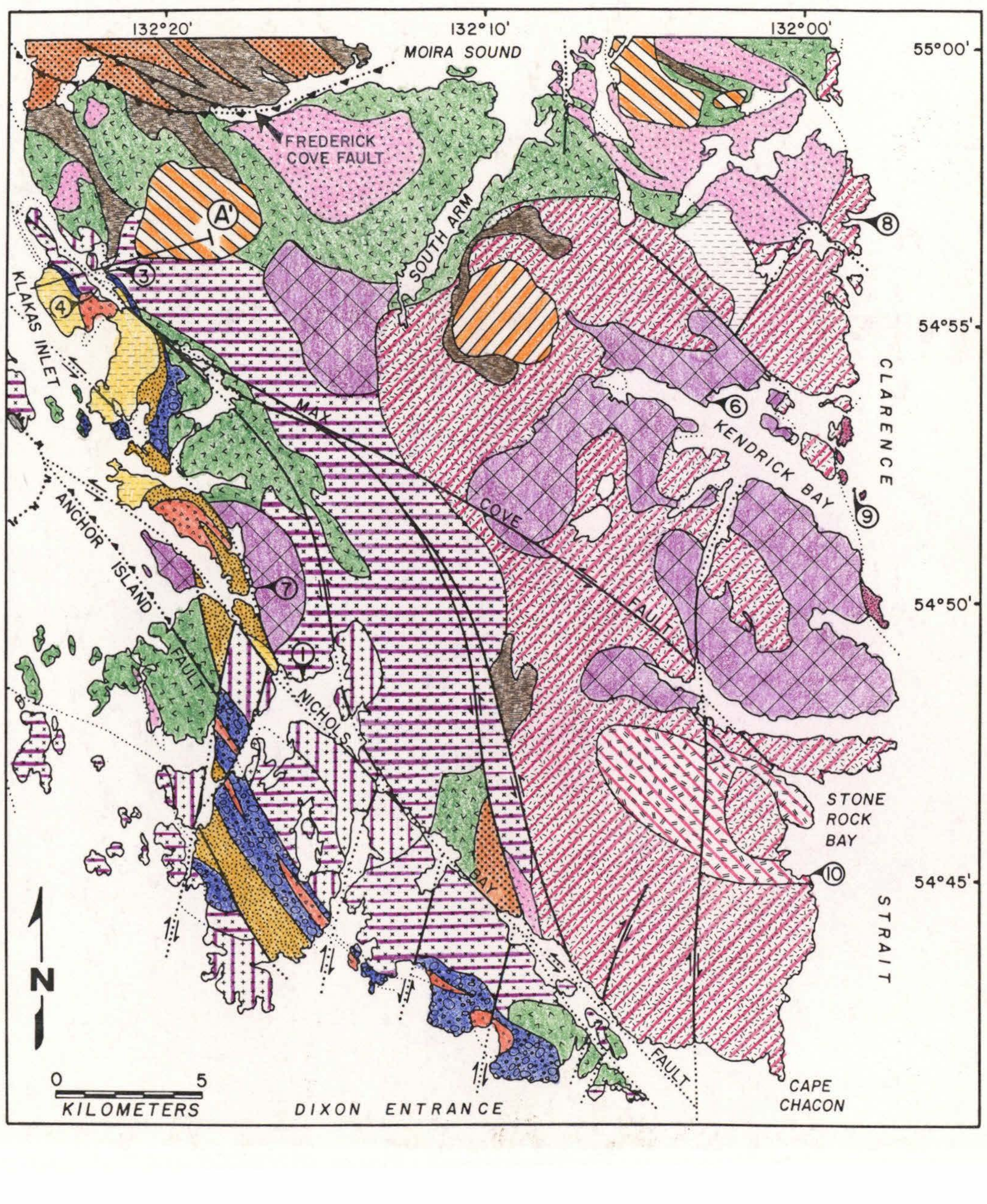
and overlie(?) pre-Ordovician metamorphic rocks. Long inlets and excellent shoreline exposures on southern Prince of Wales Island provide an opportunity to study these lower Paleozoic rocks in detail and reconstruct the early geologic history of the southern Alexander terrane.

During the summers of 1982 and 1983, I mapped the shorelines of southern Prince of Wales Island (south of lat. 55° N) in detail and the interior of the island in reconnaissance fashion (Fig. 2-2). In conjunction with this mapping, I determined U-Pb (zircon) apparent ages of the main plutonic suites, Fred Barker (U.S. Geological Survey) arranged for major, minor, and trace element geochemical analyses to be conducted through the U.S. Geological Survey, and Norman Savage (Univ. of Oregon) studied conodonts from the area. In this report I outline the geologic framework and evolution of southern Prince of Wales Island and present my geochronologic and geochemical data. Conodonts from the area have been described by Savage and Gehrels (1984), and Gehrels and others (1983a) describe several small sulfide-rich deposits discovered during the mapping. A 1:63,360 geologic map of southern Prince of Wales Island with detailed unit descriptions is presented in Gehrels and Saleeby (in review a), and Gehrels and Saleeby (in review b) outline the tectonic implications of my studies.

Radiometric ages cited herein have been calculated or recalculated with revised decay constants (Steiger and Jager, 1977), and their uncertainties are reported at the 95% level unless otherwise indicated. My U-Pb data are reported in the text and in Tables 2-1 and 2-2, and complexities of the data are discussed following the description of geologic relations. I adopt the DNAG Time Scale (Palmer,

Figure 2-2. Geologic sketch map of southern Prince of Wales Island (Dixon Entrance D-1, D-2, and C-1 and parts of adjacent 1:63,360 quadrangles). The map is compiled from my field studies during 1982 and 1983 and from mapping by MacKevett (1963) between South Arm (Moira Sound), Kendrick Bay, and Stone Rock Bay; by Thompson and others (1983) of the Bokan Mountain Granite; and by Herreid and others (1978) in the northern part of Kassa Inlet. The trace of the Keete Inlet fault in the inset map is in part from Redman (1981). On the inset map: GI = Gravina Island, AI = Annette Island, DI = Duke Island, and HI = Hotspur Island.


GRANODIORITE & DIORITE (Cretaceous)
BOKAN MOUNTAIN GRANITE (Jurassic)
KARHEEN FORMATION (Lower Devonian)
Volcanic & hypabyssal rocks
Mudstone & shale
Sandstone, siltstone & limestone
Conglomerate & breccia
LEUCODIORITE (Late Silurian: 418 ± 5 Ma)
MIDDLE ORDOVICIAN-EARLY SILURIAN PLUTONIC ROCKS
Quartz syenite (438 ± 5 Ma)
Granite (438 ± 5 Ma)
Quartz monzonite (438 ± 4 Ma)
Quartz diorite (445 ± 5 and 446 ± 5 Ma)
Diorite (462 ± 15 and ~ 480-460 Ma)
Granodiorite (472 ± 5 and 468 ±~15 Ma)
DESCON FORMATION (Ordovician - Lower Silurian)
Argillite
Mudstone & siltstone
Rhyolitic volcanic rocks
Basaltic to andesitic volcanic rocks
ORDOVICIAN METAPLUTONIC ROCKS (465 ± 7 Ma)
WALES METAMORPHIC SUITE (pre-Ordovician)
Strike-slip fault
Normal fault
Thrust fault
Contact
⑧ Geochronology samples
132°30'
KEETE INLET FAULT
KASSA INLET
11
A
5
2
BIRD ROCKS FLT.
SHIPWRECK PT. FLT.
CORDOVA BAY
BARRIER ISLANDS
CRAIG
PRINCE OF WALES ISLAND
KETCHIKAN
GI.
AI.
HI.
DI.
KEETE INLET FAULT


132°20'
132°10'
132°00'
55°00'
54°55'
54°50'
54°45'
MOIRA SOUND
FREDERICK COVE FAULT
A'
SOUTH ARM
KLAKAS INLET
ANCHOR ISLAND FAULT
NICHOLS BAY
MAY COVE FAULT
KENDRICK BAY
CLARENCE STRAIT
STONE ROCK BAY
CAPE CHACON
DIXON ENTRANCE
FAULT
N
0
5
KILOMETERS
1
3
4
6
7
8
9
10

1983) in relating radiometric ages to geologic time, and note for reference that the Ordovician-Silurian and the Silurian-Devonian boundaries are assigned ages of 438 ± 12 and 408 ± 12 Ma, respectively, on this time scale. Plutonic rocks in the area are classified according to the guidelines recommended by the IUGS Subcommission on Systematics of Igneous Rocks (Streckeisen, 1976).

Previous work

Buddington and Chapin (1929) originally divided the rocks of southern Prince of Wales Island into three main units: pre-Ordovician to Devonian metamorphic rocks of the Wales Group, Devonian sedimentary and volcanic rocks, and Upper Jurassic or Lower Cretaceous intrusive rocks. W. H. Condon, I. L. Tailleur, G. D. Eberlein, and C. D. Reynolds (all currently or formerly with the U.S. Geological Survey) remapped parts of the area in reconnaissance and refined the distribution of these units (Condon and Tailleur, unpub. U.S.G.S. report, 1960). In an effort to determine the geologic setting of U-Th mineralization near Bokan Mountain, MacKevett (1963) mapped the area between South Arm (Moira Sound), northeastern Kendrick Bay, and southern Stone Rock Bay (see Fig. 2-2) and subdivided the intrusive rocks into various compositional units. These intrusive rocks were interpreted to be of Cretaceous age until Lanphere and others (1964) determined Jurassic K-Ar apparent ages on the Bokan Mountain Granite and Paleozoic K-Ar apparent ages on the surrounding quartz monzonite and quartz diorite. The petrography and geochemistry of the Bokan Mountain Granite has been studied subsequently by Thompson and others (1982) and by B. Collot (in Saint-André and others, 1983).

The general stratigraphic framework of Buddington and Chapin (1929) was revised by G. D. Eberlein, M. Churkin Jr., and W. Vennum (unpub. U.S.G.S. mapping) and Herreid and others (1978) during studies of the south-central Prince of Wales Island region. These workers determined that rocks mapped previously as Devonian were in part of Ordovician age, and that rocks mapped as the Wales Group by Buddington and Chapin (1929) are pre-Middle Ordovician in age. My detailed mapping extends this general stratigraphic framework to southern Prince of Wales Island, confirms the Paleozoic age of most intrusive rocks, and provides more detailed information on the nature, age, distribution, and regional significance of these units.

GEOLOGIC FRAMEWORK

Overview

The oldest rocks in the study area consist of pre-Ordovician metavolcanic and metasedimentary rocks referred to herein as the Wales metamorphic suite. These rocks are juxtaposed against an Ordovician metaplutonic complex along the Bird Rocks fault, and against Ordovician-Early Silurian strata (assigned to the Descon Formation) and intrusive rocks along the Keete Inlet fault (Fig. 2-2). Ordovician-Early Silurian rocks east of the Keete Inlet fault are overlain by Lower Devonian strata assigned herein to the Karheen Formation, and are intruded by the Jurassic Bokan Mountain Granite. Granodioritic to dioritic plutons of mid-Cretaceous age intrude the Devonian and older rocks and constitute the youngest regional geologic unit on southern Prince of Wales Island. Faults in the area include the Keete Inlet fault, several thrust faults, and three sets of strike-slip faults.

The nature and age of the geologic units mentioned above are described below and their regional distribution is shown schematically on Figure 2-2. Faults in the area are described in a separate section on structural geology. I discuss complexities of the U-Pb geochronologic data following the geologic descriptions, and conclude with an outline of the geologic and tectonic history recorded by the rocks and structures in the area.

Wales metamorphic suite

The Wales metamorphic suite consists of greenschist- and locally amphibolite-facies schist and gneiss derived from basic to intermediate (basaltic to andesitic?) volcanic rocks and volcaniclastic strata. The dominant rock type in the suite is a fine-grained chlorite-albite-epidote schist in which protolith features are highly flattened and moderately elongated. Where preserved, protolith features include pyroclastic fragments (Fig. 2-3) and pillows in metavolcanic rocks, and rhythmic and locally graded bedding in metagraywacke. Marble and metarhyolite occur locally in the study area and form thick layers in the suite to the north (Herreid and others, 1978; Eberlein and others, 1983). These rocks have been referred to as the Wales Group by Buddington and Chapin (1929) and most subsequent workers. I suggest that these rocks be renamed the "Wales metamorphic suite," however, because deformation and metamorphism have modified their primary stratigraphic relations to a degree that the rocks no longer conform to the Law of Superposition (in accordance with the North American Commission on Stratigraphic Nomenclature, 1983).

Rocks in the Wales metamorphic suite north of Kassa Inlet and in most areas of south-central Prince of Wales Island have been metamorphosed to greenschist facies. South of Kassa Inlet the metamorphic grade increases eastward from greenschist to amphibolite

Figure 2-3. Photograph of highly flattened fragments in a greenschist-facies metavolcanic rock of the Wales metamorphic suite.

facies, and higher-grade rocks are juxtaposed over lower-grade rocks along the east-dipping Shipwreck Point fault (Figures 2-2). Amphibolite-facies rocks to the east consist of hornblende-biotite-garnet-plagioclase gneiss in which protolith features are only locally preserved. These rocks are in turn overthrust along the Bird Rocks fault by metaplutonic rocks of Ordovician age (described below). The age of movement on the Shipwreck Point and Bird Rocks faults is constrained by brecciation of Middle Ordovician metaplutonic rocks along the Bird Rocks fault, and by a cross-cutting leucodiorite body of Late Silurian age (described below) in Kassa Inlet (Fig. 2-2).

The orientation of the metamorphic foliation in the suite is quite variable in and north of the study area (Gehrels and Saleeby, in review a; Herreid and others, 1978). My preliminary structural analyses suggest that these variations define kilometer-scale upright folds that have apparently been refolded around kilometer-scale folds with steeply plunging axes. At outcrop scale, the metamorphic foliation has been folded around shallow-plunging, asymmetric folds (Fig. 2-4) which are coaxial with the upright folds. Where I have studied these rocks in detail, the outcrop-scale folds are overturned in a sense consistent with their having formed as parasitic structures on the limbs of the upright folds, and do not record a regionally consistent sense of asymmetry. These outcrop-scale folds do not have an axial planar foliation and are accordingly interpreted to have formed during the waning stages of, or after, the main phase of deformation and metamorphism. The lack of such folds in late Early and Middle Ordovician strata of the Descon Formation indicates that they formed prior to Middle Ordovician time, and therefore prior to movement on the

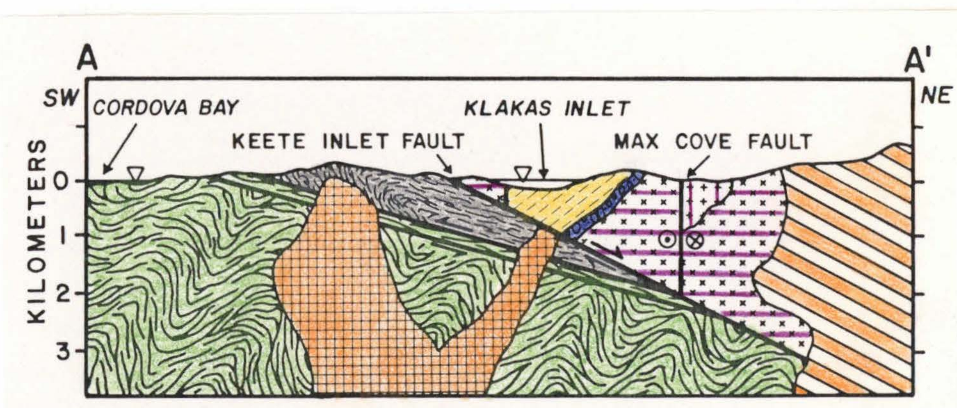
Figure 2-4. Photograph of asymmetric folds in the metamorphic foliation of the Wales metamorphic suite.

thrust faults in the area (Fig. 2-5).

The minimum age of rocks in the Wales suite is constrained by interlayered bodies of Middle and Late Cambrian metadiorite north of the study area (J. Saleeby, unpub. data) and on southern Gravina Island (Gehrels and others, in review). Turner and others (1977) suggest that regional metamorphism of these rocks occurred during Early Ordovician time based on a $^{40}K/^{40}Ar$ isochron apparent age of approximately 483 Ma. Their interpretation is supported by observations that: (1) metaplutonic rocks of Middle and Late Cambrian age have experienced the metamorphism and deformation, (2) late Early Ordovician and younger rocks of the Descon Formation (described below) have not been regionally deformed and metamorphosed, and (3) relatively nondeformed plutons of Ordovician-Early Silurian age intrude the Wales suite on east-central Prince of Wales Island (Saleeby and Gehrels, unpub. mapping). Based on these relations and my interpretation that the Middle and Late Cambrian intrusive rocks are at least in part cogenetic with volcanic rocks in the Wales metamorphic suite, I assign a late Proterozoic(?)-Cambrian age to the Wales suite. We note, however, that the age of these rocks has not yet been documented through geochronologic data or stratigraphic relations.

The tectonic event recorded by metamorphism and deformation of rocks in the Wales suite has been referred to informally as the Wales "orogeny" (Gehrels and Saleeby, 1984 and in review b). Other manifestations of this event include uplift of the metamorphic rocks prior to deposition of strata in the Descon Formation, and accumulation of a thick section of conglomeratic graywacke at the base(?) of the Descon Formation several kilometers north of the area (Herreid and

Figure 2-5. Cross section across southern Klakas Inlet showing relations between faults and geologic units. Unit symbols are the same as on Figure 2-2.



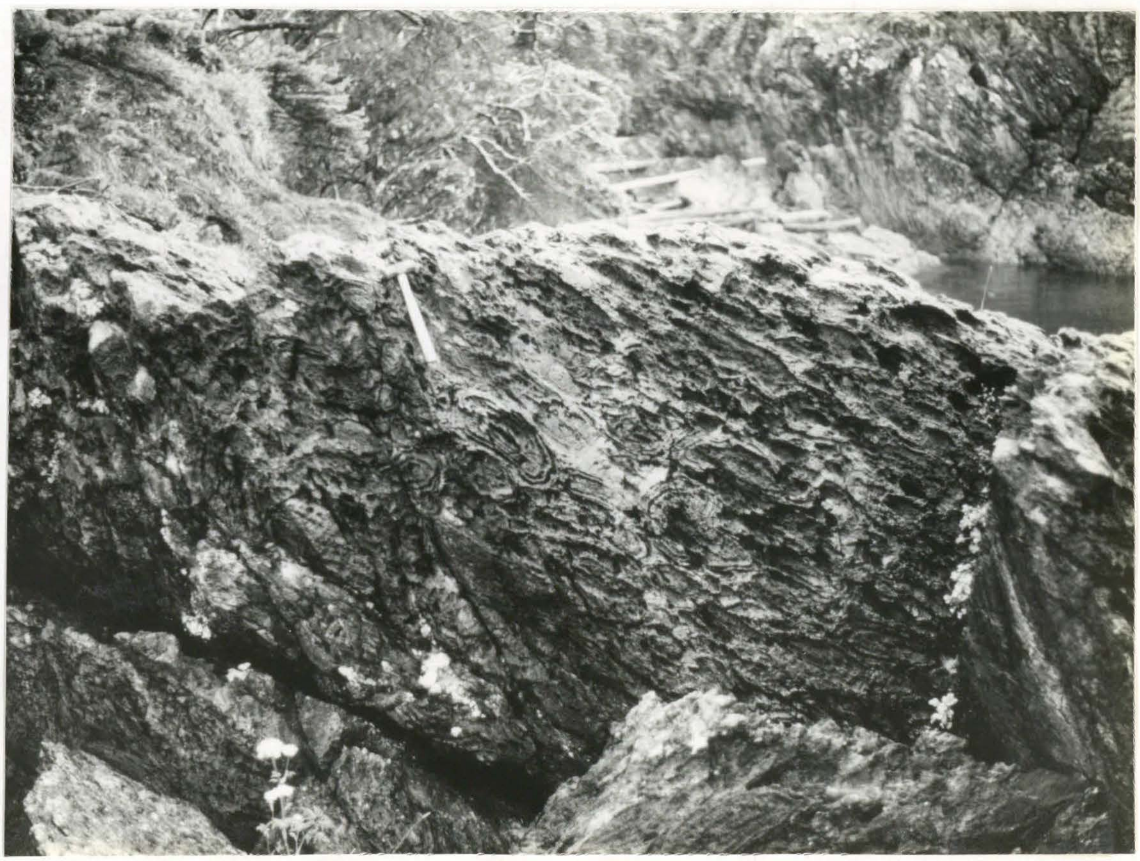
others, 1978). I use the term orogeny informally in this instance because little is known about the regional and tectonic significance of this event.

The tectonic environment in which protoliths of the Wales metamorphic suite accumulated is difficult to determine due to the regional metamorphism and penetrative deformation. The scarcity of metapelite and the abundance of pillow flows and coarse andesitic breccia suggest that most rocks were deposited in a proximal volcanic setting. Comparison of the composition and stratigraphic relations of rocks in the Wales suite with rocks in the Descon Formation and other Circum-Pacific magmatic belts suggests that the Wales suite formed in an ensimatic volcanic arc environment along a convergent margin.

Descon Formation

Much of southern Prince of Wales Island is underlain by basaltic to rhyolitic volcanic rocks, subordinate argillite, mudstone, siltstone, and graywacke, and minor limestone which are referred to herein and by Herreid and others (1978) as the Descon Formation. North of the Frederick Cove fault (Fig. 2-2), the dominant rock types are interbedded argillite and banded mudstone and siltstone. South of the fault, basaltic to andesitic pillow flows (Fig. 2-6), pillow breccia, and tuff breccia predominate, rhyolitic pyroclastic breccia, tuff, and extrusive domes are subordinate, and argillite, mudstone, and limestone are minor constituents. Reconnaissance mapping along the north shore of Moira Sound north of the study area (Fig. 2-2) indicates that the clastic strata north of the Frederick Cove fault overlie basaltic-andesitic volcanic rocks which are interpreted to be correlative with those south of the fault. The clastic strata north of the fault are therefore

Figure 2-6. Well-preserved pillows in a basalt belonging to the Descon Formation.



interpreted to be younger than the volcanic rocks to the south.

The degree of deformation and metamorphism of rocks in the Descon Formation is quite variable, but is nowhere as high as in rocks of the Wales suite. North of the Frederick Cove fault, the marine clastic strata dip and face to the southwest and are only locally folded. South of the fault, northwest-trending, shallow-plunging folds with several-meter-scale wavelengths are common, and larger-scale folds with this orientation apparently control the outcrop pattern of strata in the Descon Formation in most of the area. Protolith features are slightly deformed in most rocks, and the mafic- to intermediate-composition rocks are commonly recrystallized to microcrystalline chlorite, epidote, and albite. Even where deformed and recrystallized, however, rocks in the Descon Formation are readily distinguished from those in the Wales metamorphic suite by the higher grade of metamorphism, stronger flattening and elongation of protolith features, and asymmetric folds in the older rocks.

Rocks belonging to the Descon Formation on central and northern Prince of Wales Island consist of volcanic and volcaniclastic strata which range in age from Early Ordovician to Early Silurian (Eberlein and others, 1983). Fossils recovered from similar strata on southern Prince of wales Island (Klakas Inlet and Moira Sound) include Middle Ordovician conodonts and latest Early and Middle Ordovician graptolites (Eberlein and others, 1983; Gehrels and Saleeby, in review a). I herein correlate the rocks on southern Prince of Wales Island with those to the north and assign the strata in the study area to the Descon Formation based on their similarity in rock types, stratigraphic position, and, at least in part, age. I note, however, that Upper Ordovician-Lower Silurian fossils have not been recovered from strata in the study area.

Depositional relations between rocks of the Descon Formation and the Wales suite have not been documented in the Prince of Wales Island region. Churkin and Eberlein (1977) and Eberlein and others (1983) report that Ordovician strata along the shores of Klakas Inlet belong to a continuous, southward-younging section which overlies rocks of the Wales suite at the north end of the inlet. Because late Early and Middle Ordovician fossils have been recovered from the Descon Formation in southern Klakas Inlet, they argue that strata to the north are probably Cambrian in age, and that rocks in the Wales suite are likely Precambrian in age. In contrast, my mapping has shown that the Ordovician strata along Klakas Inlet are cut by several major thrust and high-angle faults, including the Frederick Cove fault, and are folded at both outcrop and regional scale. Mapping along the north shore of Moira Sound suggests also that clastic strata north of the Frederick Cove fault are younger than the fossil-bearing Ordovician strata to the south. These relations indicate that strata along the shores of Klakas Inlet are structurally complex and are at least in part younger than the late Early and Middle Ordovician strata at the south end of the inlet. In addition, Herreid and others (1978) report that the Wales suite and Descon Formation are separated by the Keete Inlet fault at the head of Klakas Inlet. My reconnaissance mapping north of the Klakas Inlet region indicates that the contact between the Descon Formation and Wales suite is everywhere a fault along the east side of Prince of Wales Island.

Intrusive relations and age and petrographic similarities suggest that rocks in the Descon Formation are genetically related to large plutons of Ordovician diorite and granodiorite (described below) which underlie much of southern Prince of Wales Island. Because similar plutonic rocks intrude the Wales suite to the north, it is likely that strata of the Descon Formation were originally in depositional contact with rocks in the Wales suite. Thus, I concur with Churkin and Eberlein (1977) and Eberlein and others (1983) that the Wales suite is probably basement to the younger volcanic-plutonic complex, but note that I have not been able to demonstrate depositional relations between the two units on southern or eastern Prince of Wales Island.

Middle Ordovician-Early Silurian plutonic rocks

Overview

Plutonic rocks of Middle Ordovician-Early Silurian age on southern Prince of Wales Island constitute the largest exposure of lower Paleozoic intrusive rocks in the Alexander terrane. Crosscutting relations and my geochronologic data indicate that the oldest intrusive bodies in this suite consist of Middle Ordovician diorite and granodiorite. Large bodies of quartz diorite were emplaced during Late Ordovician time, and latest Ordovician-earliest Silurian quartz monzonite, granite, and quartz syenite are the youngest members of the plutonic suite. The petrographic characteristics, intrusive relations, and age constraints of these plutonic rocks are described below. My geochronologic data are presented in the text and in Tables 2-1 and 2-2, and are discussed in a following geochronology section. The major, minor, and trace element data from samples of the intrusive rocks and two samples of Descon volcanic rocks are listed in Table 2-3.

Diorite and granodiorite

Large intrusive bodies of diorite, quartz-porphyritic granodiorite,

Table 2-1. U-Pb (zircon) isotopic data and apparent ages.

SMPL #	WT (mg)	SIZE	CONC. (ppm) ^{238}U	CONC. (ppm) $^{206}Pb^*$	$\frac{Pb^*}{Pb}$ (AT%)	ISOTOPIC COMPOSITION $\frac{^{206}Pb}{^{204}Pb}$	$\frac{^{206}Pb}{^{207}Pb}$	$\frac{^{206}Pb}{^{208}Pb}$	APPARENT DATES $\frac{^{206}Pb^*}{^{238}U}$	$\frac{^{207}Pb^*}{^{235}U}$	$\frac{^{207}Pb^*}{^{206}Pb^*}$	APPARENT AGE (Ma)
1	16.3	B	429.2	28.21	99.45	10,690	17.294	7.169	472.0	472.2	472.8	$472 \pm 5^{\alpha}$
2	7.1	C	494.7	31.84	99.48	9,360	17.307	6.247	462.5	462.1	460	$462 \pm 15^{\alpha}$
3a	25.9	C	717.5	40.49	99.81	31,200	17.595	7.3395	407.4	416.7	468.9	$468 \pm \sim 15^{\beta}$
3b	9.6	C	721.1	43.23	99.78	26,600	17.575	7.3504	431.9	437.7	468.3	"
3c	12.3	D	900.4	45.06	99.76	23,500	17.550	7.1197	362.5	377.2	468	"
4a	17.9	AB	808.6	51.70	99.79	26,400	17.566	4.5456	459.6	461.2	469.3	$\sim 480\text{-}460^{\psi}$
4b	15.9	Cs	757.5	48.02	99.82	30,700	17.585	4.8011	455.8	458.1	469.9	"
4c	12.4	Cs	819.1	51.78	99.78	25,540	17.565	4.7251	454.6	456.9	468.7	"
5a	11.4	C	906.2	58.76	99.83	34,700	17.610	5.532	465.8	466.3	468.9	$465 \pm 7^{\alpha}$
5b	11.4	C	871.5	56.15	99.83	32,100	17.610	5.507	463.0	463.7	467.5	"
6	17.0	Cl	456.5	28.22	98.99	5,880	17.155	7.048	444.8	445.3	447.4	$445 \pm 5^{\alpha}$
7	18.0	AB	499.7	30.96	99.76	25,200	17.738	12.978	445.8	445.9	446.4	$446 \pm 5^{\alpha}$
8a	16.1	As	235.1	14.33	99.45	10,540	17.563	7.653	438.7	438.4	436.8	$438 \pm 4^{\alpha}$
8b	17.1	Bl	249.2	15.16	99.12	6,780	17.342	7.697	437.9	437.5	435	"
9a	16.5	AB	913.9	55.73	99.87	47,300	17.880	10.4713	439.0	439.1	439.3	$438 \pm 5^{\alpha}$
9b	11.1	Cl	1,085.7	65.69	99.90	58,100	17.918	10.894	435.7	435.9	436.8	"
10	7.0	BC	1,195.6	72.72	99.18	7,510	17.375	9.508	437.9	438.0	438.3	$438 \pm 5^{\alpha}$
11a	17.8	A	1,157.5	67.08	99.76	24,300	17.953	7.055	417.9	418.0	418.4	$418 \pm 5^{\alpha}$
11b	19.4	Ad	2,195	126.0	99.73	21,800	17.948	6.281	414.1	414.4	416.2	"

* radiogenic lead.

$^{\alpha}$ Uncertainty of age is reported at the 95% level.

$^{\beta}$ Apparent age is interpreted from the upper intercept on Pb*/U concordia diagrams and uncertainty is estimated graphically (Fig. 2-7).

$^{\psi}$ Age is probably within range cited, but data are not sufficient to determine a reliable apparent age (Fig. 2-8)

Notes: Additional sample information is provided in Table 2-2.

Size of zircon fractions: A = 120-160u; B = 80-120u; C = 45-80u; D = < 45u; s = smaller part of fraction, l = larger part of fraction, d = darker part of fraction.

Constants used: $\lambda^{238}U = 1.55125 \times 10^{-10}$; $\lambda^{235}U = 9.8485 \times 10^{-10}$; and $^{238}U/^{235}U$ (atomic) = 137.88 (from Steiger and Jager, 1977).

Uncertainty in concentrations, isotopic composition, and apparent dates is in last two figures cited.

Isotopic compositions cited above have not been adjusted. In calculating concentrations and apparent dates the isotopic data have been adjusted as follows:

1) Mass-dependent correction factors of .09% ± .04% (per AMU) and 0.12 ± .04% (per AMU) have been applied to lead and uranium ratios, respectively. These factors have been determined by replicate analyses of NBS SRM 981, 982, and 983 for Pb, and U-500 and U-930 for U.

2) The $^{207}Pb/^{206}Pb$ of the spiked aliquot has been adjusted for the ^{206}Pb and ^{207}Pb added with the spike.

3) Isotopic composition of the unspiked aliquot has been adjusted for 0.04 ±.02 ng blank lead with a composition of: $^{206}Pb/^{204}Pb$ = 18.78 ± 0.30; $^{207}Pb/^{204}Pb$ = 15.61 ± 0.22; and $^{208}Pb/^{204}Pb$ = 38.5 ± .60. The amount of blank Pb has been determined by isotope dilution analysis of a

typical dissolution and chemical separation procedure conducted without zircon. The blank composition has been determined through isotopic analysis of acids and dust particles in the laboratory.

4) Isotopic composition of U in the spiked aliquot has been adjusted for .07 ± .05 ng blank U, which has been determined by the same procedure as the Pb blank.

5) Isotopic composition of the spiked aliquot has been adjusted for blank Pb by balancing the $^{206}Pb/^{207}Pb$ of the spiked and unspiked aliquots. Blank Pb in the spiked aliquot is assigned the composition cited above.

6) Common Pb remaining after correction for blank Pb is interpreted to be initial Pb and is assigned a composition of: $^{206}Pb/^{204}Pb = 18.0 \pm 1.5$; $^{207}Pb/^{204}Pb = 15.59 \pm 0.4$; and $^{208}Pb/^{204}Pb = 37.8 \pm 2.0$. (Values interpreted from Doe and Zartman, 1979).

The U and Pb concentrations and apparent dates cited above and the Pb*/U ratios, uncertainties, and error correlations shown in Figs. 7 and 8 have been calculated with a computer program written by Ken Ludwig (1983).

Table 2-2. U-Pb geochronologic sample information.

Smpl No.	Station No.	Latitude (N)	Longitude (W)	Magnetic Character
1	82GP702	54°48'41"	132°15'37"	NM 10°/20°

Rock: Moderately altered, coarse-grained, quartz-porphyritic granodiorite.

Zircon: Sb>An=Eu, morphologically heterogeneous population ranging from 2:1 (euhedral) to 1:1 (anhedral), most have inclusions, light pink in color.

2	79AE215	54°53'41"	132°24'54"	NM 5°/20°

Rock: Highly altered, medium-grained, hornblende=biotite diorite.

Zircon: light pinkish-tan in color.

3	82GP48	54°56'03"	132°21'45"	NM 10°/20°

Rock: Highly altered hornblende(?) leucogranodiorite.

Zircon: Eu>Sb>>An, 1:1 to 2:1, most grains have inclusions, very light pinkish color.

4	82GP40	54°55'44"	132°22'10"	NM 2°/20°

Rock: moderately altered, medium- to fine-grained, hornblende diorite.

Zircon: Eu>Sb=An, 2:1 to 4:1 with most grains terminated by fractures, inclusions common, medium to deep pink.

5	79AE114	54°53'53"	132°26'00"	M 5°/20°

Rock: Strongly foliated, fine-grained, biotite granodiorite.

Zircon: An=Sb>Eu, 1:1 to 2:1 elongation with very irregular and angular shapes; almost all grains have inclusions, light pinkish-tan.

6	82GP28	54°53'42"	132°03'00"	NM 10°/20°

Rock: Highly altered, medium-grained, hornblende quartz diorite.

Zircon: Eu>>Sb=An, 2:1 to 3:1, inclusions in most grains, light-medium pink.

7	82GP346	54°50'20"	132°17'08"	NM 2°/20°

Rock: Highly altered, medium-grained, hornblende quartz diorite.

Zircon: Eu=Sb>An, 2:1 to 3:1 elongation, small inclusions in most grains, variable in color from light pink to dark brown with slight maroon tint.

8	83GP255	54°56'57"	131°58'32"	NM 2°/20°

Rock: Moderately altered, medium-grained, hornblende=biotite quartz monzonite.

Zircon: Sb=An>Eu, 1:1 to 2:1 elongated, inclusions are rare, medium to deep pink with slight maroon tint to some grains.

9 83GP335 54°52'14" 131°58'28" NM 2°/20°

Rock: Moderately altered, coarse-grained, biotite leucogranite.

Zircon: Sb>Eu>An, Sb are 1:1, Eu are 2:1, inclusions are common, pink with brownish tint.

10 83GP364 54°45'08" 131°59'56" NM 5°/15°

Rock: Highly altered, fine- to medium-grained, quartz syenite.

Zircon: Sb>Eu>An, Sb are elongate to 2:1 and Eu reach 3:1; inclusions are rare, medium pink with some brownish grains.

11 82GP626 54°56'26" 132°30'07" NM 2°/10°

Rock: Moderately altered, medium-grained, arfvedsonite garnet leucodiorite.

Zircon: An>>Sb>Eu, very angular and irrgular with no elongation, few inclusions, low U-conc. fraction is medium pink with brownish tint; higher U-conc. fraction is a mottled dark brown.

Notes: Magnetic characteristics; NM = Non-magnetic split; M = Magnetic split; first angle is side slope on Frantz Isodynamic Separator; second angle is forward slope. All samples were processed at 1.7 amps current.

Eu = Euhedral, Sb = subhedral, An = Anhedral

2:1 signifies length:width of grains.

Color determined under reflected light.

Inclusions consist of.clear and colorless rods and tiny dark brown specks.

Table 2-3. Major, minor, and trace element data.

SAMPLE:	13	8	3	6	10	5	12	9	14
	MAJOR ELEMENT COMPOSITION (WEIGHT PERCENT)								
SiO_2	51.1	56.3	59.5	61.7	68.7	70.1	74.6	74.6	76.6
Al_2O_3	18.2	17.3	15.7	17.5	13.8	14.5	13.5	12.9	13.7
$FeTO_3$	8.20	7.97	7.46	5.34	3.34	3.38	1.89	1.80	0.54
MgO	4.90	3.06	3.28	2.28	1.11	1.03	0.39	0.40	0.26
CaO	6.43	6.59	4.56	4.70	1.71	2.29	0.38	1.27	0.27
Na_2O	4.93	3.67	3.20	4.65	3.51	3.76	4.07	3.56	7.26
K_2O	2.55	2.65	2.89	1.20	4.58	2.31	3.94	3.95	0.15
TiO	0.96	0.87	0.58	0.44	0.41	0.33	0.24	0.22	0.18
P_2O_5	0.41	0.33	0.14	0.16	0.08	0.09	0.06	0.06	<0.05
MnO	0.13	0.16	0.11	0.12	0.03	0.08	<0.02	0.05	<0.02
LOI	1.21	0.90	1.87	1.34	2.06	1.55	0.96	0.35	0.30
TOTAL:	99.02	99.80	99.29	99.43	99.33	99.42	100.05	99.16	99.33
CO_2	1.06	0.04	0.40	0.22	1.50	0.42	0.08	0.07	0.08
H_2O+	1.51	1.13	1.92	1.38	0.70	1.23	0.74	0.27	0.26
H_2O-	0.07	0.06	0.08	0.04	0.06	0.06	0.04	0.03	0.03
	MINOR AND TRACE ELEMENT CONTENTS (PARTS PER MILLION)								
Nb	6	7	8	6	6	6	9	10	13
Rb	58	73	82	38	123	51	73	97	2

Sr	473	614	397	729	403	391	79	176	47
Zr	65	110	115	95	405	189	215	186	206
Y	24	29	28	15	32	19	23	23	45
Cu	59	40	35	6	49	3	2	5	<2
Ni	99	13	19	10	7	7	<5	<5	<5
Zn	113	87	39	67	25	49	36	22	9
Ba	378	748	687	1670	1490	622	483	962	56
Y	24.0	28.2	22.7	11.8	12.7	13.0	8.4	14.3	22.7
La	14.2	23.3	20.3	10.3	28.9	16.2	22.3	29.0	24.9
Ce	32.1	49.9	41.3	20.8	51.0	28.7	43.7	45.9	54.0
Pr	4.2	6.4	4.7	2.5	5.4	2.7	4.9	4.5	6.8
Nd	20.4	27.6	19.4	11.3	20.1	11.6	17.8	15.2	27.1
Sm	5.0	6.2	4.2	2.3	3.9	2.2	3.4	2.6	6.1
Eu	1.52	1.68	1.00	0.79	0.58	0.73	0.53	0.51	0.83
Gd	6.2	6.3	4.4	2.9	3.3	2.5	2.8	2.7	5.5
Dy	4.5	5.3	3.8	2.3	1.8	2.0	1.3	2.2	3.5
Ho	0.87	1.05	0.81	0.48	0.44	0.45	0.30	0.49	0.88
Er	2.57	3.14	2.44	1.56	1.30	1.34	0.87	1.58	2.75
Tm	0.36	0.47	0.35	0.33	0.17	0.13	0.13	0.26	0.43
Yb	2.33	3.05	2.71	1.29	1.64	1.58	1.16	1.90	3.47
Lu	0.31	0.43	0.42	0.18	0.22	0.22	0.19	0.29	0.54

Notes: Major element data in weight %, minor and trace elements in ppm.

$FeTO_3$ = total iron expressed as Fe_2O_3.

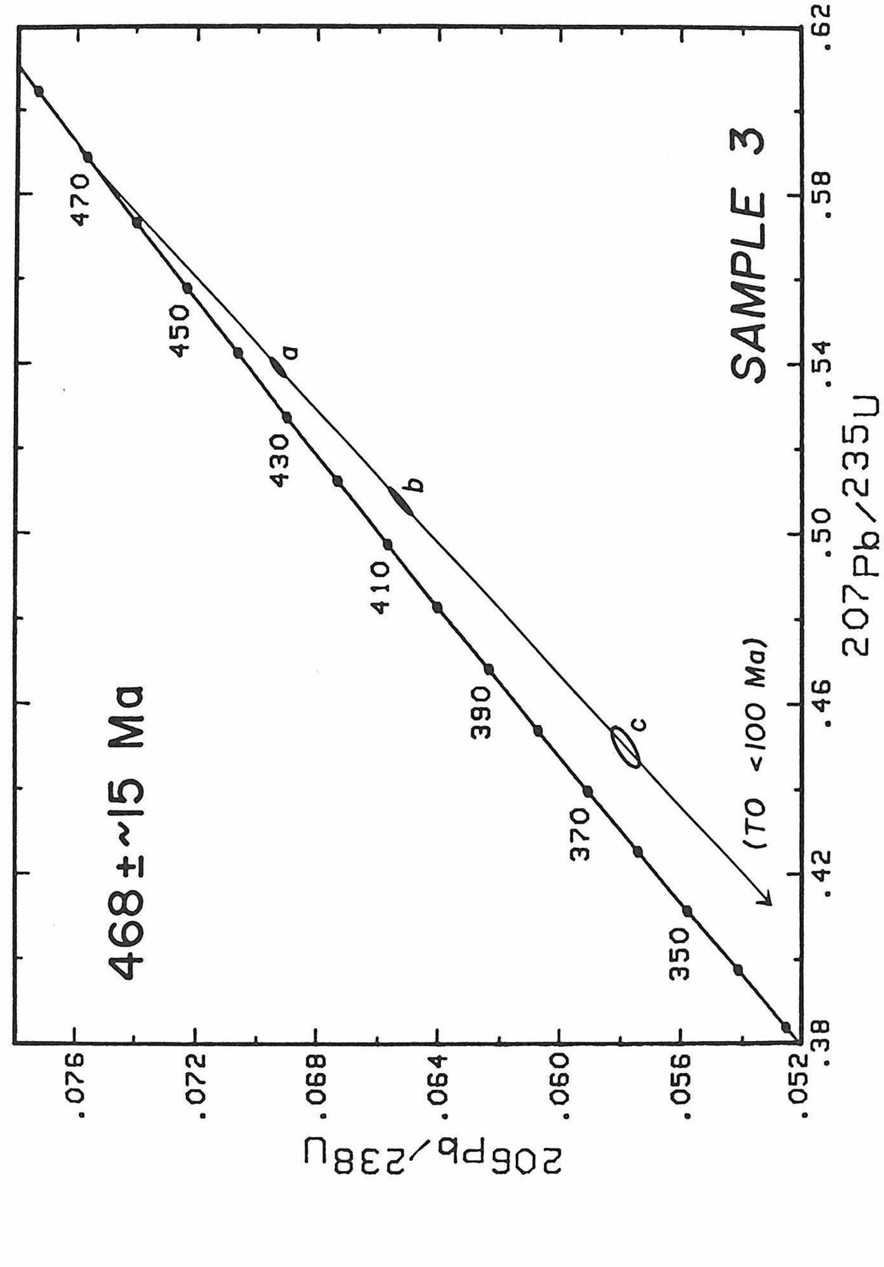
Major and minor elements by XRF; rare earth elements by INAA.

Sample numbers are the same as the geochronologic samples except for 12, which is an Ordovician leucogranodiorite; 13, which is a basaltic to andesitic volcanic rock belonging to the Descon Formation; and 14, which is a rhyolitic volcanic rock belonging to the Descon Formation.

and leucogranodiorite underlie much of the western and southern parts of the study area (Fig. 2-2). Diorite occurs as massive and homogeneous bodies and as heterogenous complexes of layered and foliated diorite, diorite-basite-granodiorite agmatite, and dikes and small intrusive bodies of diorite, microdiorite, and basalt. Most diorites have a color index between 30 and 50 and consist of medium-grained hornblende, andesine, and subordinate biotite and quartz. Quartz-porphyritic granodiorite generally occurs in massive, medium- to coarse-grained, homogeneous bodies that consist primarily of large bluish quartz grains, hornblende, oligoclase, and interstitial microperthite. Color indices range from 5 to 20 in these rocks. The leucogranodiorite (not distinguished from quartz-porphyritic granodiorite on Fig. 2-2) is a distinctive rock that consists of myrmekitic quartz and plagioclase and a variable proportion of interstitial microperthite. Ferromagnesian minerals constitute less than 5% of the leucogranodiorite and occur as small clusters of chlorite presumably derived from hornblende.

The available age constraints indicate that the dioritic and granodioritic rocks are primarily Middle Ordovician in age. Quartz-porphyritic granodiorite (sample 1) yields an apparent age of 472 ± 5 Ma, and diorite from west of Klakas Inlet yields an apparent age of 462 ± 15 Ma (sample 2). Leucogranodiorite (sample 3) from east of Klakas Inlet yields discordant apparent dates, where discordance is recognized on Pb*/U concordia diagrams by a lack of overlap (at the 95% level) of individual zircon fractions and concordia. As shown on Figure 2-7, three zircon fractions from sample 3 yield an upper intercept of 468 ± ~15 Ma and a poorly constrained lower intercept of less than 100 Ma. Discordance of the zircon populations is attributed to isotopic

Figure 2-7. Concordia diagram of zircon fractions from leucogranodiorite at sample locality 3. Error ellipses are at the 95% level. The concordia diagram was generated with computer programs written by Ken Ludwig (1983; written commun., 1984).

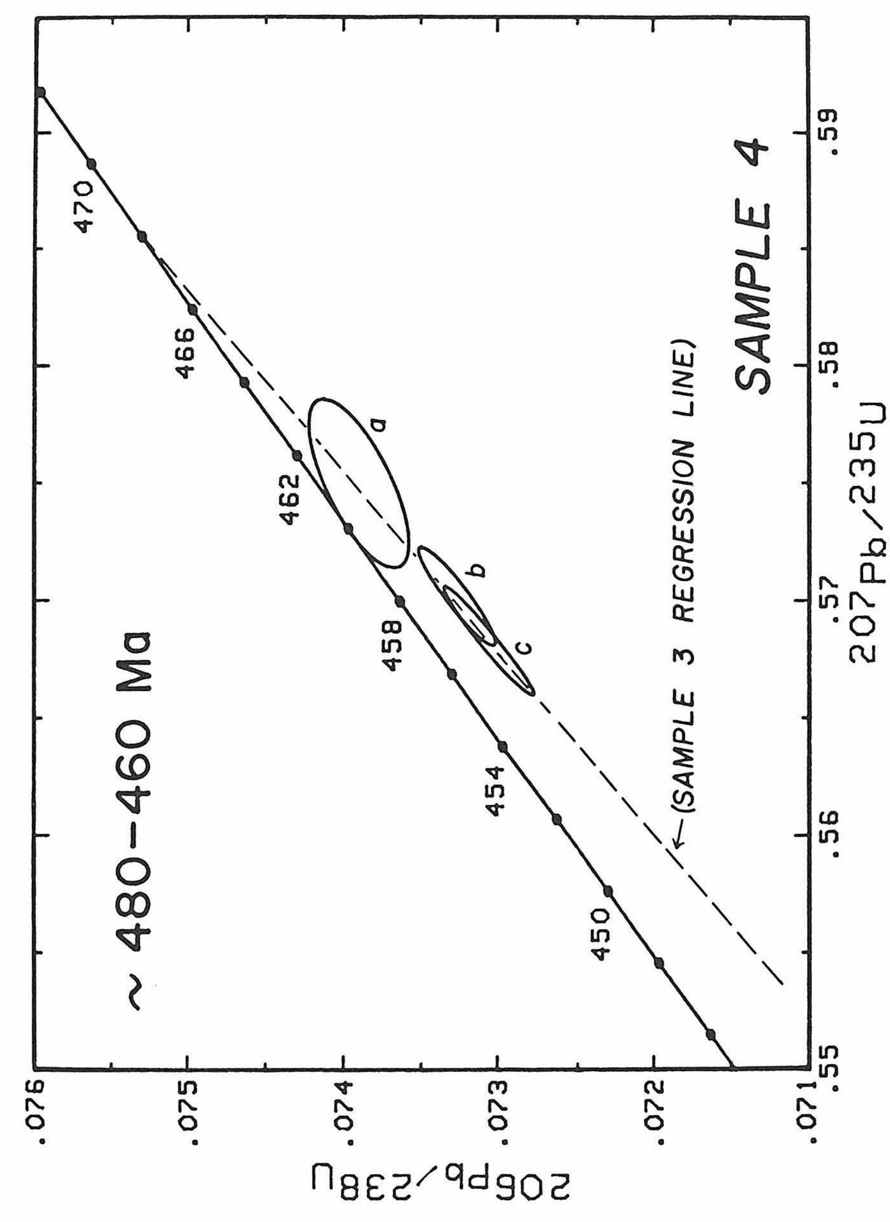

468±~15 Ma
SAMPLE 3
(TO <100 Ma)
a
b
c
350
370
390
410
430
450
470
.076
.072
.068
.064
.060
.056
.052
.38
.42
.46
.50
.54
.58
.62
207Pb/235U
206Pb/238U

disturbance of Paleozoic zircons rather than entrainment of older U-bearing phases in Mesozoic or Cenozoic zircons because the leucograndiorite is overlain by Lower Devonian clastic strata. I accordingly interpret the upper intercept with concordia (468 ± ~15 Ma; Fig. 2-7) as the apparent age for sample 3. Because the process by which the zircon populations from this sample have been isotopically disturbed is not well known (as discussed in the geochronology section below), the uncertainty cited for the age of sample 3 is an estimate and has not been determined statistically.

Diorite from just west of sample locality 3 also yields slightly discordant apparent dates (sample 4: Fig. 2-8). In this case, however, the isotopic data indicate that the rock is older than 460 Ma but are not sufficient to determine a precise apparent age. I note, however, that the zircon fractions from sample 4 plot on the same regression line as fractions from sample 3 (Fig. 2-8). This suggests that zircons in the two rocks formed at approximately the same time and have undergone a similar isotopic disturbance. I accordingly interpret the apparent age of sample 4 to be in the 480-460 Ma range but do not report a specific age.

Turner and others (1977) report a K-Ar (hornblende) date of 428 ± 13 Ma (one-sigma uncertainty) on a small granodiorite body adjacent to the leucogranodiorite at sample locality 3. They interpret this date to be a minimum age based on the penetrative alteration of the rock, and I suggest that the rock is significantly older than 428 Ma because it is intruded by the 468 ± ~15 Ma leucogranodiorite at sample locality 3. Disturbance of the K-Ar system in their sample and of the U-Pb systems in samples 3 and 4 is discussed in the geochronology section below.

Figure 2-8. Concordia diagram of zircon fractions from hornblende diorite at sample locality 4. Regression line shown is that determined from sample 3 and shows that samples 3 and 4 may be approximately coeval and may have a similar history of isotopic disturbance. Error ellipses are at the 95% level. The concordia diagram was generated with programs written by Ken Ludwig (1983, written commun., 1984).


~ 480-460 Ma
SAMPLE 4
(SAMPLE 3 REGRESSION LINE)
a
b
c
450
454
458
462
466
470
.55
.56
.57
.58
.59
$^{207}Pb/^{235}U$
.071
.072
.073
.074
.075
.076
$^{206}Pb/^{238}U$

Churkin and Eberlein (1977, pp. 777) report an age of at least 730 Ma on a trondhjemite body along the west side of Klakas Inlet (Ruth Bay), which they interpret to intrude rocks of the Wales suite. The date they cite is an incorrect interpretation of preliminary discordant Pb/Pb and Pb/U dates provided informally by J. Saleeby and has no bearing on the age of intrusive rocks in the area. In addition, my mapping has shown that these intrusive rocks (which are actually leucogranodiorite in composition) are separated from rocks in the Wales suite by both the Keete Inlet fault and the suite of Ordovician metaplutonic rocks (Fig. 2-2).

Middle Ordovician dioritic and granodioritic rocks on southern Prince of Wales Island are interpreted to be genetically related to volcanic rocks in the Descon Formation based on: (1) their similar composition, (2) a gradation in the southwestern part of the area from massive diorite and granodiorite, through a kilometer-thick zone of hypabyssal dioritic and granodioritic dikes and small intrusive bodies, into interlayered basaltic to andesitic and rhyolitic volcanic rocks of the Descon Formation, (3) the petrographic and compositional similarity of leucogranodiorite bodies and nearby silicic volcanic layers in the Descon Formation along Klakas Inlet, and (4) their similarity in age.

Quartz diorite

Large bodies of massive, fine- to medium-grained quartz diorite and subordinate diorite underlie the east-central part of the study area. These rocks have color indices of approximately 25 and consist of andesine, green hornblende, and subordinate biotite and interstitial quartz and K-feldspar. Rocks in this unit are intruded by quartz monzonite and granite plutons, but intrusive relations with the large diorite and granodiorite bodies to the west have not been determined. Lanphere and others (1964) report a K-Ar (hornblende) apparent age of 439 ± 21 Ma (one-sigma uncertainty) from quartz diorite in Kendrick Bay, and I have determined a single-fraction apparent age of 445 ± 5 Ma (sample 6) from what appears to be the same intrusive body.

A suite of interlayered dikes and small intrusive bodies of foliated quartz diorite and diorite in the west-central part of the area yields an apparent age of 446 ± 5 Ma (sample 7). Intrusive relations suggest that the layering and foliation in this suite result from syn-plutonic flow rather than a regional deformational event, and similarities in mineralogy, composition, and age suggest that these rocks are genetically related to the quartz dioritic rocks to the east. Thus, my age data indicate that the quartz dioritic rocks are predominantly Late Ordovician in age and significantly younger than the large dioritic and granodioritic bodies to the west.

Quartz monzonite, granite, and quartz syenite

The youngest members of the Ordovician-Early Silurian plutonic suite have a greater abundance of K-feldspar than earlier members, and have been divided into quartz monzonite, granite, and quartz syenite units on Figure 2-2. Also associated with these units, but not shown on Figure 2-2, are small intrusive bodies of hornblendite and pyroxenite that locally display gradational relations with quartz syenite. The quartz monzonite unit comprises large bodies of homogeneous and massive quartz monzonite, granite, and quartz monzodiorite that underlie much of the eastern part of the study area. These rocks consist primarily of medium-grained K-feldspar, andesine-oligoclase, quartz, and glomerocrysts of biotite and hornblende. They intrude Late Ordovician quartz diorite and are intruded by both granite and quartz syenite. Lanphere and others (1964) report K-Ar apparent ages of 454 ± 22 Ma on hornblende and 379 ± 18 Ma on biotite from these rocks (one-sigma uncertainties), and I have determined an apparent age of 438 ± 4 Ma on a medium- to coarse-grained biotite quartz monzonite from sample locality 8.

A body of distinctive granite with large microperthite phenocrysts and up to 10% biotite intrudes quartz monzonite along the easternmost shore of the island near Kendrick Bay. Two zircon fractions from granite at sample locality 9 yield an apparent age of 438 ± 5 Ma. This rock may be genetically related to a large body of grayish, reddish, and locally maroon quartz syenite that intrudes quartz monzonite along the east shore of the island. The quartz syenitic rocks consist of fine- to medium-grained microperthite, quartz, oligoclase-andesine, and up to 20% glomerocrysts of hornblende and biotite. I have determined an apparent age of 438 ± 5 Ma on a sample of quartz syenite from sample locality 10.

Interpretation of the Ordovician-Early Silurian plutonic rocks

Middle Ordovician-Early Silurian plutonic rocks and coeval volcanic rocks of the Descon Formation are similar in many respects to modern-day volcanic-plutonic complexes in the Circum-Pacific region. Characteristics of rocks on southern Prince of Wales Island that are shared by other convergent-margin assemblages include: (1) evolution of the magmatic suite during a period of approximately 40 m.y., (2) predominance of dioritic to granitic intrusive rocks and basaltic to rhyolitic volcanic rocks, (3) a general trend toward increasing potassium content with time in the intrusive rocks, (4) the calc-alkaline affinity of the intrusive rocks, as defined by an alkali-lime index of between 56 and 62, and on AFM (Fig. 2-9), FeO*/MgO versus SiO_2 (Fig. 2-10), and a variety of other petrochemical diagrams, (5) the similarity to orogenic andesites as defined on a La versus Nb diagram (Fig. 2-11), and (6) strong (30 to 100 times) enrichment of light REE, moderate (5 to 20 times) enrichment of heavy REE, and a negative Eu anomaly (compared to chondritic abundances: Fig. 2-12). Based on these similarities and a lack of conflicting geological or geochemical evidence, I suggest that the Ordovician-Early Silurian volcanic-plutonic complex on southern Prince of Wales Island formed in an ensimatic volcanic arc environment along a convergent margin.

Ordovician metaplutonic complex

A complex of foliated and layered metaplutonic and minor metavolcanic and metasedimentary rocks is juxtaposed against rocks of the Wales metamorphic suite along the Bird Rocks fault, and against Ordovician-Early Silurian rocks along the Keete Inlet fault (Figs. 2 and 6). Foliated and slightly layered gabbroic rocks constitute approximately 60% of the complex and consist predominantly of elongate and aligned green hornblende and andesine. These rocks are intruded by sills of strongly foliated granodiorite that are generally 10 to 50 cm thick and at least several meters long. The granodioritic rocks generally have a color index of less than 15 and consist of strongly aligned quartz, plagioclase (oligoclase-andesine), brown biotite, and subordinate hornblende. In spite of the penetrative deformation, an igneous texture is apparent in most rocks.

The gabbroic and granodioritic rocks intrude screens of metavolcanic and subordinate metasedimentary rocks which constitute less

Figure 2-9. AFM diagram showing my samples of Ordovician-Early Silurian plutonic and volcanic rocks and the tholeiitic and calc-alkaline fields of Irvine and Baragar (1971). (A = $Na_2O + K_2O$, F = FeO + $.9xFe_2O_3$, and M = MgO, all in weight percent)

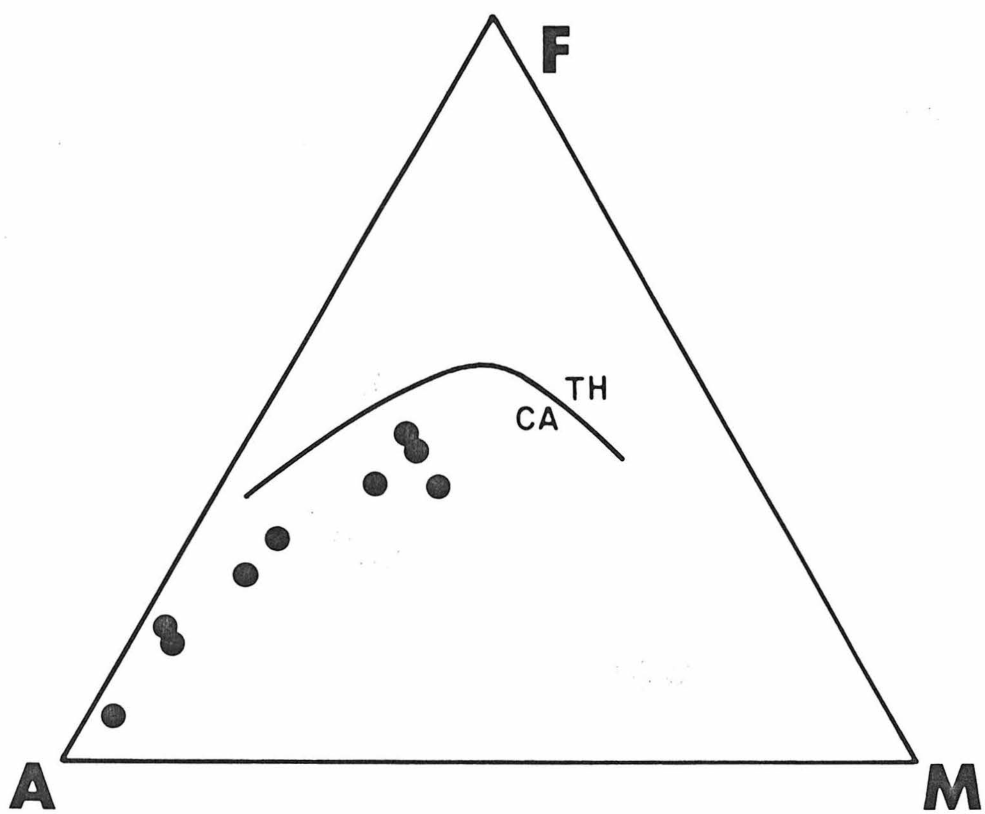


Figure 2-10. $(FeO + .9xFe_2O_3)/MgO$ versus SiO_2 diagram for my samples of Ordovician-Early Silurian plutonic and volcanic rocks (all in weight percent). Calc-alkaline and tholeiitic fields from Miyashiro (1974).

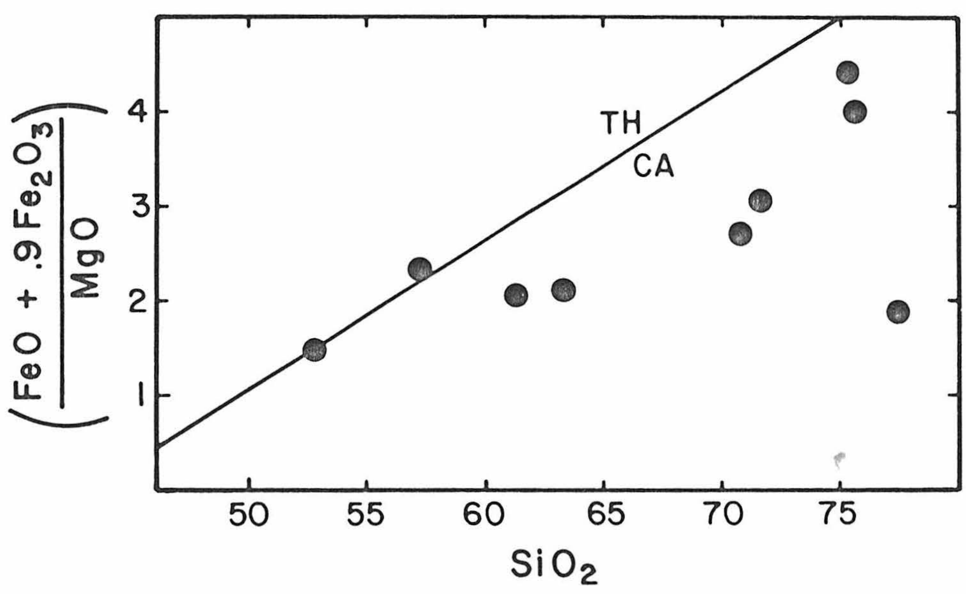


Figure 2-11. La versus Nb (in ppm) for my samples of Ordovician-Early Silurian plutonic and volcanic rocks. Field of orogenic andesites from Gill (1981).

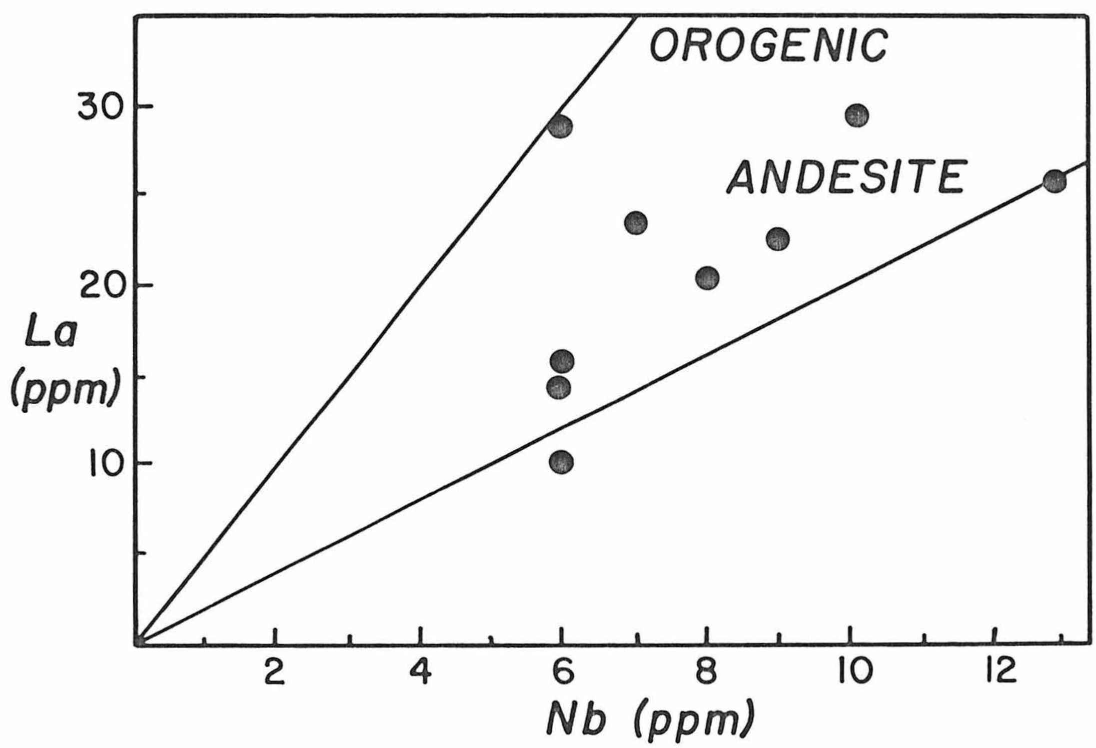
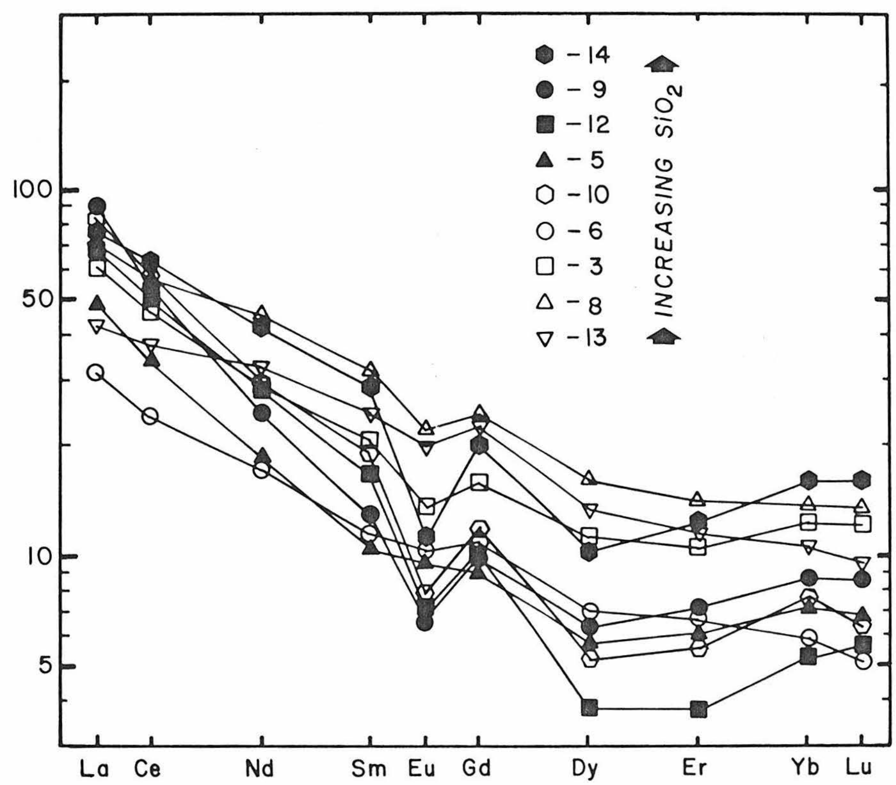


Figure 2-12. REE abundances relative to chondrites for my samples of Ordovician-Early Silurian plutonic and volcanic rocks. Sample numbers correspond to geochronologic samples except for 12 (Ordovician leucogranodiorite), 13 (basaltic-andesitic volcanic rock belonging to the Descon Formation), and 14 (rhyolitic volcanic rock belonging to the Descon Formation). Chondrite abundances from Nakamura (1974).



than 5% of the complex. These rocks generally have a hornblende-plagioclase mineralogy similar to the gabbroic rocks, which suggests that they have been metamorphosed to amphibolite facies. A lack of strong flattening and elongation of protolith features in the metavolcanic rocks indicates that these screens were derived from rocks in the Descon Formation rather than the Wales suite.

Contacts between the granodioritic and gabbroic rocks are generally parallel to the regional north-northwest-striking foliation contained by the intrusive bodies. In some outcrops, however, granodiorite dikes cut across the foliation in older gabbro yet contain the regional foliation (Fig. 2-13). These relations indicate that the granodioritic rocks were emplaced during the deformation. I have determined an apparent age of 465 ± 7 Ma on a granodiorite sill at sample locality 5, which indicates that the granodioritic rocks were emplaced and the complex was deformed during Middle Ordovician time. The rare occurrence of foliated gabbro dikes cutting granodioritic rocks indicates that the gabbroic rocks are probably of Middle Ordovician age as well.

The tectonic significance of the layering, foliation, and metamorphism of rocks in this complex is not well constrained. As described above, intrusive rocks above the Keete Inlet fault are somewhat similar in composition and are of approximately the same age, but do not display the layering, ductile deformational fabrics, or higher grade of metamorphism of rocks below the fault. This indicates that either these aspects of the metaplutonic complex were imparted during a local rather than regional tectonic event, or the Keete Inlet fault represents a major tectonic boundary separating rocks with dissimilar Middle Ordovician histories.

Figure 2-13. Photograph of foliated diorite and cross cutting but foliated granodiorite. A similar granodiorite from sample locality 7 yields a U-Pb apparent age of 465 ± 7 Ma.

In accord with: (1) my conclusion that the layering and foliation in the complex were imparted during emplacement of the intrusive rocks, (2) the lack of Middle Ordovician tectonism in rocks elsewhere in the region, and (3) my interpretation (described below) that the Keete Inlet fault is a normal fault and not a major tectonic boundary, I hypothesize that the layering and foliation formed through magmatic processes at depths where rocks in the Descon Formation were undergoing amphibolite-facies metamorphism. In this scenario, the metaplutonic rocks are interpreted to be deeper-level equivalents of the dioritic and granodioritic rocks that occur above the Keete Inlet fault. Layering in the complex is attributed to preferential alignment of intrusive sheets during emplacement (perhaps in a regime of extension) and reorientation of the bodies into parallelism during flattening. Amphibolite-facies metamorphism of rocks in the Wales suite west of and beneath the Bird Rocks fault may have occurred: (1) during intrusion and metamorphism of rocks in the Ordovician metaplutonic complex, (2) during Silurian time, in response to their being overthrust by deeper-level rocks in the metaplutonic complex, or (3) in response to regional metamorphism associated with the Middle Cambrian-Early Ordovician Wales "orogeny." The lack of a strong linear fabric in both these rocks and the metaplutonic rocks to the east favor the interpretation that the amphibolite-facies metamorphism of rocks in the Wales suite occurred during emplacement and metamorphism of the Middle Ordovician metaplutonic rocks.

Silurian leucodiorite

Leucodiorite bodies near Kassa Inlet are quite different from other Paleozoic intrusive rocks in the study area, and are apparently unique in the Alexander terrane. These rocks have a color index of less than 15 and consist primarily of medium- to coarse-grained oligoclase, arfvedsonite with cores of aegirine-augite, subordinate dark-brown (melanite?) garnet, and large grains of euhedral sphene. Two zircon size fractions from the leucodiorite yield an apparent age of 418 ± 5 Ma (sample 11).

The large leucodiorite body in Kassa Inlet is tectonically significant because it cuts across the Bird Rocks and Shipwreck Point faults (Fig. 2-2), which indicates that the faults moved prior to the end of Silurian time. Plutons belonging to this suite are additionally significant because they record a change from calc-alkaline plutonism during Middle Ordovician-Early Silurian time to generation of sodic magmas during Late Silurian time. As described below, this change in magmatic evolution and pre-latest Silurian movement on the Bird Rocks and Shipwreck Point faults are manifestation of a major middle Silurian-earliest Devonian tectonic event which I have referred to as the Klakas orogeny (Gehrels and others, 1983b).

Karheen Formation

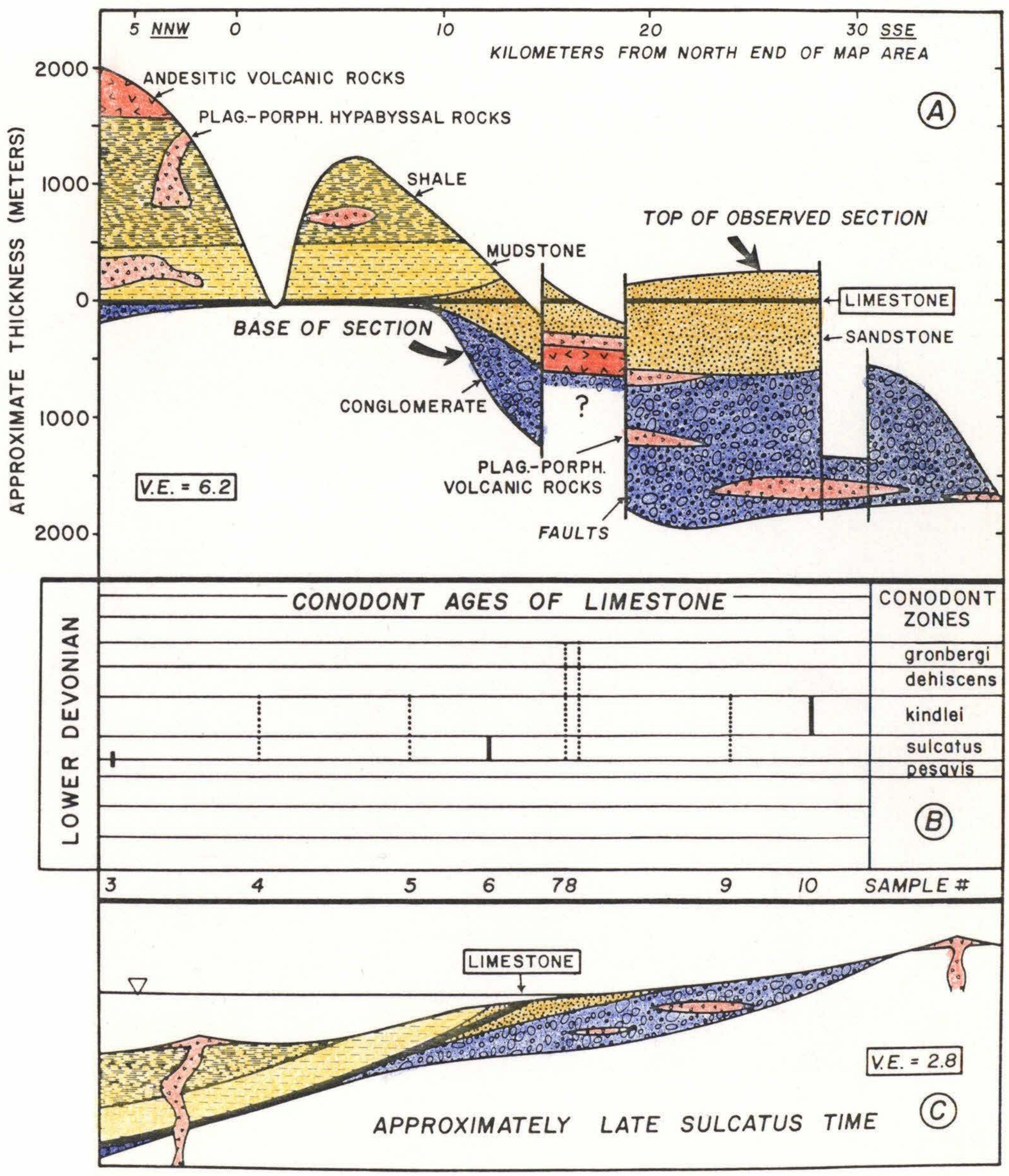
Lower Devonian strata on southern Prince of Wales Island consist of sedimentary and subordinate volcanic rocks which exhibit lateral and vertical facies changes over short distances. In general, the sedimentary sequence comprises conglomeratic strata at the base, sandstone, siltstone, and limestone in the middle, and, in the northern part of the area, laminated mudstone and black shale at the top. I assign these strata to the Karheen Formation based on their similarities in rock-type, stratigraphic position, and age with strata near the type section of the Karheen Formation on west-central Prince of Wales Island

(Eberlein and Churkin, 1970; Eberlein and others, 1983). In the southern part of the study area, plagioclase-porphyritic dacite and microporphyritic andesite are interbedded with the conglomeratic strata and occur as hypabyssal intrusive bodies. These volcanic and hypabyssal rocks and strata in the lower, middle, and upper parts of the sequence are described below. Relations between the various rock types are shown schematically on Figure 2-14 and are discussed in a following section on regional stratigraphic variations.

Conglomeratic strata (lower part of section)

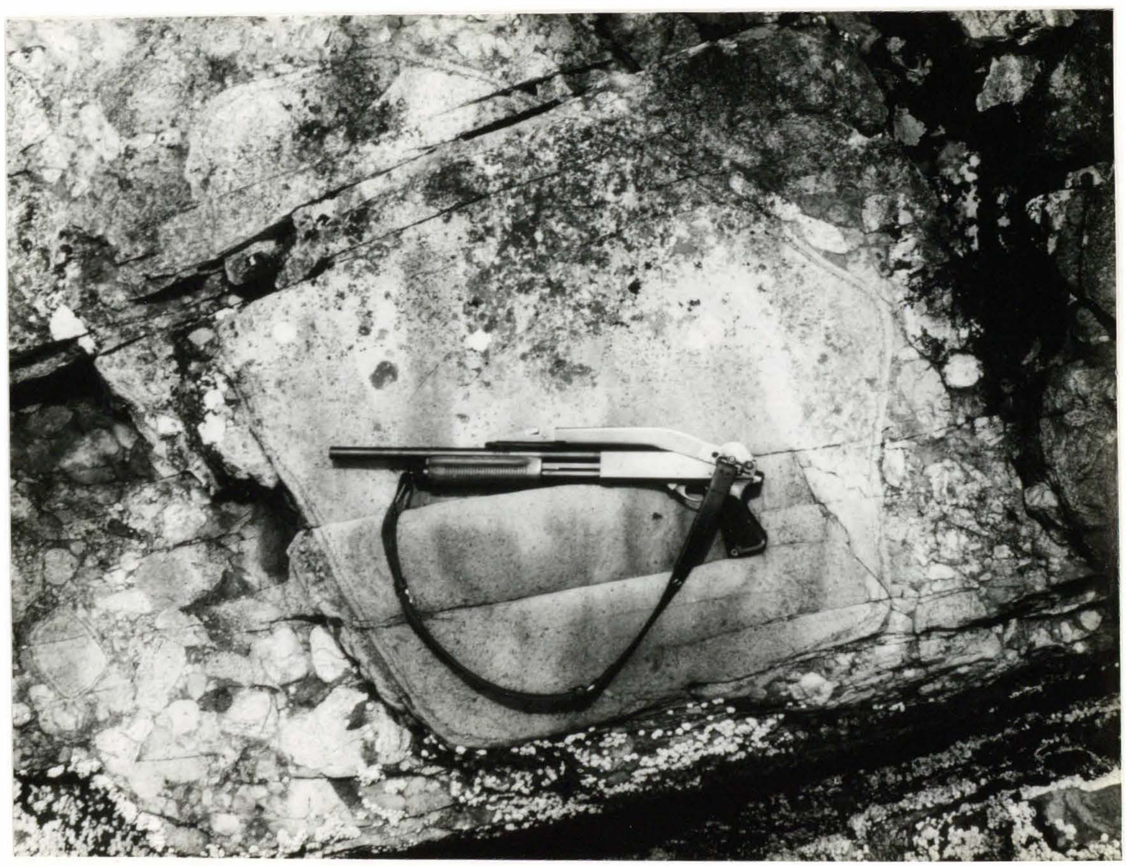
The lower part of the Karheen Formation in the study area consists of reddish-brown cobble to boulder conglomerate, pebbly sandstone, and poorly sorted sedimentary breccia. The thickness of this part of the section varies from several meters at the north end of the area to over a kilometer along the south shore of the island (Fig. 2-14). In the northern part of the area, clasts in the conglomerate are moderately well rounded, generally less than 40 cm in diameter, and consist primarily of Ordovician volcanic and plutonic rocks. Along the south shore of the island, plagioclase-porphyritic dacitic flows and breccia are interbedded with the conglomeratic strata. Conglomeratic strata near these volcanic layers consist of porphyritic dacite clasts in an arkosic matrix. Clasts in the southern part of the area are in general larger and more angular than those to the north and locally reach over a meter in diameter (Fig. 2-15). In some areas the conglomeratic strata display high-angle cross-beds, channels, ripple marks, and other sedimentary structures which indicate deposition in subaerial environments. These relations combined with the thickness of strata in the southern part of the area and the size of clasts suggest that the

Figure 2-14. Schematic stratigraphic sections and available conodont ages along the north-northwest-trending belt of Lower Devonian strata on southern Prince of Wales Island. Data from north of the study area are adapted from Herreid and others (1978). The upper section (a) shows the presently exposed thickness of units above and below the layer of limestone which occurs along most of the length of the section. Conodont ages are shown in part b with a solid vertical line where fauna indicate a well-constrained age, and with a dotted vertical line where they indicate a range in age. These fauna have been described by Savage and Gehrels (1984). The age-diagnostic fauna from localities 3, 6, and 10 apparently indicate that the limestone is younger to the south. The lower section (c) is a schematic reconstruction of the first-order facies variations during deposition of the strata.



5 NNW 0 10 20 30 SSE
KILOMETERS FROM NORTH END OF MAP AREA
APPROXIMATE THICKNESS (METERS)
2000
1000
0
1000
2000
ANDESITIC VOLCANIC ROCKS
PLAG.-PORPH. HYPABYSSAL ROCKS
SHALE
MUDSTONE
TOP OF OBSERVED SECTION
LIMESTONE
SANDSTONE
BASE OF SECTION
CONGLOMERATE
?
PLAG.-PORPH. VOLCANIC ROCKS
FAULTS
V.E. = 6.2
A
LOWER DEVONIAN
CONODONT AGES OF LIMESTONE
CONODONT ZONES
gronbergi
dehiscens
kindlei
sulcatus
pesavis
B
3 4 5 6 78 9 10 SAMPLE #
LIMESTONE
V.E. = 2.8
APPROXIMATELY LATE SULCATUS TIME
C

Figure 2-15. Large clast in the basal conglomerate of the Karheen Formation (Lower Devonian) on southernmost Prince of Wales Island.



region had significant subaerial topographic relief during deposition of the lower part of the section.

Near the south end of Klakas Inlet, the conglomeratic strata grade into a coarse sedimentary breccia that overlies and contains clasts of brecciated Ordovician volcanic and plutonic rocks (Fig. 2-16). This breccia places an important constraint on the age of deformation in the area because underlying Ordovician rocks are penetratively deformed, the breccia itself is moderately deformed, and unconformably overlying mudstone and shale are only slightly deformed. These relations, the poorly sorted and angular nature of the clasts and matrix, and regional structural relations described below, suggest that the breccia was deposited within or adjacent to an active fault system. Movement on this inferred fault system and deposition of the coarse conglomeratic strata are interpreted to be manifestations of the Klakas orogeny.

Sandstone, siltstone, and limestone (middle part of section)

Conglomeratic strata generally grade upsection into a sequence of reddish-brown, well-bedded sandstone, siltstone, and mudstone, subordinate gray fossiliferous limestone, and minor chert-pebble conglomerate and interbedded maroon and green shale. The limestone can be traced nearly continuously across the area and occurs as a single layer up to several meters in thickness or as a pair of thin layers that occur near the top of the middle part of the section. Shelly fauna are abundant in the limestone and age-diagnostic conodonts have been recovered from several localities (Fig. 2-14: Savage and Gehrels, 1984). According to the conodont ages, the limestone was deposited during a brief interval in the middle Early Devonian, and appears to be slightly younger to the south (Fig. 2-14). Strata in this part of the

Figure 2-16. Sedimentary breccia that overlies and contains clasts of highly deformed Ordovician volcanic and plutonic rocks, and is overlain by relatively nondeformed middle Lower Devonian mudstone and shale. The breccia is interpreted to have been deposited and deformed within a regime of active faulting during Early Devonian time.

section are interpreted to have been deposited in shallow-marine to intertidal environments based on the occurrence of cross beds and small channels in the siltstone, interbedded maroon and green shale, and shallow-marine shelly fauna in the limestone.

Mudstone and black shale (upper part of section)

Tan to gray mudstone and overlying graptolitic black shale constitute the upper part of the sedimentary sequence in the northern part of the study area. These strata are laminated, layered, and locally show size-grading, and are interpreted to have been deposited in distal regions of a submarine fan. Tectonic activity within or adjacent to the basin during deposition of the strata is recorded by several-centimeter-thick layers of leucogranodiorite-clast conglomerate in the shale and by the subjacent sedimentary breccia in southern Klakas Inlet. Churkin and others (1970) report that graptolites in black shale near the top of the section are probably late Pragian (middle Early Devonian) in age.

Volcanic and hypabyssal rocks

Two types of volcanic rocks are interbedded with Lower Devonian clastic strata on southern Prince of Wales Island. In the southwestern part of the area, layers of plagioclase-porphyritic dacite(?) up to several hundred meters in thickness are interbedded with conglomeratic and arkosic rocks low in the section. Hypabyssal bodies of this plagioclase (and diopsidic augite) porphyry occur along the length of the belt of Devonian strata in the study area (Gehrels and Saleeby, in review a) and are also widespread in Lower Devonian strata just north of the area (Herreid and others, 1978). Northern Kuiu Island (central southeastern Alaska: Muffler, 1967) is the only other region in the

Alexander terrane where similar volcanic, hypabyssal, and arkosic rocks have been reported. The strong similarity of rocks in these two regions suggests that the arkosic and related rocks on Kuiu Island are Early Devonian in age rather than Late Silurian as inferred by Muffler (1967).

In the west-central part of the area, plagioclase-microporphyritic basalt or andesite pillow flows, volcanic breccia, and hypabyssal rocks occur beneath limestone in the middle part of the section. These volcanic rocks resemble andesitic breccia and pillow flows that overlie Lower Devonian clastic strata just north of the area (Herreid and others, 1978) and to the east on Hotspur Island (Gehrels and others, in review). A thin sequence of pillow flows along the south shore of the island near the base of the sequence may also be correlative with these basaltic to andesitic rocks.

Regional variations in the Devonian strata

The Lower Devonian strata on southern Prince of Wales Island generally define a fining-upward clastic sequence, but, as shown on Figure 2-14, there are significant regional variations in the facies relations and thickness of these strata. The upper cross section of Figure 2-14 (part a) displays the first-order north-south variations using the limestone as a horizontal reference, but does not accurately depict the age relations of different units. As shown in part b of this figure, age-diagnostic conodonts from the limestone are younger to the south, indicating that at least the middle and upper units in the sequence are time-transgressive. Southward-younging of the sequence is also indicated by observations that: (1) plagioclase-porphyritic rocks occur as hypabyssal bodies in the northern part of the area and as porphyritic volcanic flows and breccia low in the section to the south, and (2) basaltic to andesitic pillow flows and breccia occur at the top of the section to the north, beneath limestone in the middle part of the section in the central part of the area, and near the base of the section along the south shore of the island (Fig. 2-14). Assuming that the volcanic rocks correlated above are coeval, these relations indicate that shale at the north end of the area is coeval with or slightly older than much of the conglomeratic strata along the south shore of the island! If this assumption is correct, the age constraints provided by the conodonts indicate that the marine clastic strata above the limestone to the north and most of the conglomeratic strata below the limestone to the south were both deposited during a brief interval between late pesavis and the end of kindlei time (see conodont zones of Fig. 2-14, part b).

The time-transgressive nature of the limestone combined with the change from subaerial, through shallow-marine, to more distal marine environments recorded by the clastic strata indicate that these rocks were deposited during a marine transgression on a generally north-facing paleoslope. This transgression may have occurred in response to regional subsidence following the Klakas orogeny and (or) may reflect a global rise in sea level during middle Early Devonian (sulcatus-kindlei) time (Johnson and others, 1985). A speculative reconstruction of the first-order facies variations during deposition of the strata is shown in the lower cross section of Figure 2-14 (part c).

On central and northern Prince of Wales Island (northwest of Craig on the inset map of Figure 2-2), strata in the Karheen Formation grade northwestward from a >1800-m-thick section of conglomerate to a thinner section of fine-grained clastic strata (Ovenshine, 1975; Ovenshine and others, 1969). Studies by these workers of sedimentary structures and facies relations indicate that the conglomeratic rocks were deposited in shallow-marine to subaerial environments and belong to a clastic wedge that was shed from a source area to the south or southeast. The evidence for middle Silurian-earliest Devonian deformation, faulting, and uplift in the study area suggests that: (1) the source area postulated by Ovenshine and others (1969) extended into the southern Prince of Wales Island region (Gehrels and others, 1983b), and 2) conglomeratic strata on southern Prince of Wales Island belong to a proximal facies of this clastic wedge. The evidence for a northward transition from subaerial to shallow-marine depositional environments both in the study area and on central Prince of Wales Island suggests, however, that the Early Devonian paleotopography was considerably more complex than a uniform north-facing paleoslope. These relations also suggest that the area uplifted during the Klakas orogeny extended considerably south or southeast of southern Prince of Wales Island.

Bokan Mountain Granite (Jurassic)

The Bokan Mountain Granite has received considerable research attention because of its U-Th mineralization and anomalous peralkaline composition. Thompson and others (1982) describe the intrusive body as a peralkaline ring-dike complex that consists of aegirine- and riebekite-bearing granite aplite, porphyry, and pegmatite. B. Collot (*in* Saint-André and others, 1983) reports that the granite is generally zoned from albitic aegirine granite at the center, through albitic arfvedsonite granite, to fine-grained albitic arfvedsonite-aegirine granite around the margin. I have not studied the granite in detail, but examination of several thin sections suggests that the amphibole in the rock is indeed arfvedsonite rather than riebekite.

The Bokan Mountain Granite has been dated by a variety of methods, all of which yield Jurassic dates. Lanphere and others (1964) report K-Ar apparent ages of 185 ± 8 Ma and 190 ± 8 Ma (one-sigma uncertainties) on riebekite (arfvedsonite?) from the granite. Saint-André and others (1983) have analyzed several zircon fractions from two samples near the margin of the body and report an upper-intercept U-Pb apparent age of 171 ± 5 Ma (one-sigma uncertainty?). I consider this date suspect, however, because: (1) their measured $^{206}Pb/^{204}Pb$ is so low that a reasonable error in the assigned common lead composition would significantly change the apparent age, and (2) two of their zircon fractions plot above concordia and the other fractions define a chord with a lower intercept of 0 ± 15 Ma. This indicates that either their zircon populations were not completely dissolved, their sample and spike(s) were not equilibrated, or the natural Pb*/U of the zircon fractions has been altered during dissolution or chemical separation procedures. Rb/Sr analyses of 10 whole-rock samples from the granite yield a minimum isochron apparent age of 156 Ma (Armstrong, in press). The available geochronologic data therefore suggest that the Bokan Mountain Granite is Jurassic in age, but, recognizing the possibility that excess radiogenic argon has affected the K-Ar dates, the geochronometric age of the body has apparently not yet been reliably determined.

Cretaceous diorite and granodiorite

Several large bodies of diorite and granodiorite intrude the Paleozoic rocks on southern Prince of Wales Island. The mineralogy of these intrusive bodies is quite different from rocks in the Paleozoic suite in that large euhedral sphene and anhedral magnetite grains are ubiquitous, augite occurs in most rocks, and plagioclase forms tabular, strongly zoned, and relatively unaltered grains. The similarity in mineralogy and composition between these rocks and granodiorite and diorite of known mid-Cretaceous age north of the study area (Herreid and others, 1978) suggests that these plutons are mid-Cretaceous in age. One such pluton cuts across the Keete Inlet fault north of the study area (Herreid and others, 1978) and another may intrude across a strand of the Keete Inlet fault between Kassa and Klakas Inlets (Fig. 2-2).

STRUCTURAL GEOLOGY

The dominant regional structures on southern Prince of Wales Island include several thrust faults, the Keete Inlet fault, and strike-slip faults belonging to three sets (Fig. 2-2). These faults are described below.

Thrust faults

Thrust faults in the area include the Frederick Cove fault, the Bird Rocks and Shipwreck Point faults, and the Anchor Island fault and its inferred northern continuation in Klakas Inlet. The Frederick Cove fault dips southward at moderate angles and juxtaposes volcanic rocks of the Descon Formation over younger marine clastic strata. This fault is overlain by Lower Devonian strata along the west shore of Klakas Inlet just north of the study area (Herreid and others, 1978), which suggests that it moved prior to middle Early Devonian time.

Thrust faults west of the Keete Inlet fault dip at moderate angles to the east and juxtapose Ordovician metaplutonic rocks over amphibolite-facies rocks of the Wales suite (Bird Rocks fault), and higher-grade metamorphic rocks of the Wales suite over lower-grade rocks

(Shipwreck Point fault: Figs. 2 and 5). Striae on these and associated minor faults indicate a slip-line of N65E-S65W (± 20 degrees) (Fig. 2-17). A southwestward sense of movement is indicated by an overturned antiform in the foliation and layering of metaplutonic rocks above the Bird Rocks fault (Fig. 2-5), and by juxtaposition of higher-grade over lower-grade rocks (assuming that the higher-grade rocks were at deeper levels prior to movement on the faults). The timing of movement is constrained by the Middle Ordovician age of metaplutonic rocks cut by the Bird Rocks fault, and by cross-cutting Late Silurian leucodiorite in Kassa Inlet.

The Anchor Island fault is recognized as wide zones of brecciation in Ordovician volcanic rocks east of the Barrier Islands and of brecciation and alteration in Ordovician volcanic and plutonic rocks in southern Klakas Inlet. In these two regions, Ordovician leucogranodiorite is penetratively brecciated into angular, centimeter-scale fragments surrounded by finely comminuted quartz and feldspar grains and rock fragments (Fig. 2-18). Ordovician volcanic and dioritic rocks are moderately brecciated and have been recrystallized to chloritic semischist. Throughout the zone in Klakas Inlet these rocks also have a pervasive, orange-weathering dolomitic alteration. The minimum age of deformation along this fault zone is constrained by overlying sedimentary breccia which contains clasts of the highly deformed Ordovician rocks, is itself moderately deformed, and is overlain by relatively nondeformed middle Lower Devonian mudstone and shale. Based on the occurrence of thrust faults with pre-Devonian displacement nearby and projection of the Anchor Island fault toward Klakas Inlet, I suggest that the deformational fabrics and alteration

Figure 2-17. Stereoplot (equal-area) showing the trend and plunge of slickenside striae on thrust faults and associated minor faults west of the Keete Inlet fault. The average slip-line is N65°E-S65°W ± 20°.

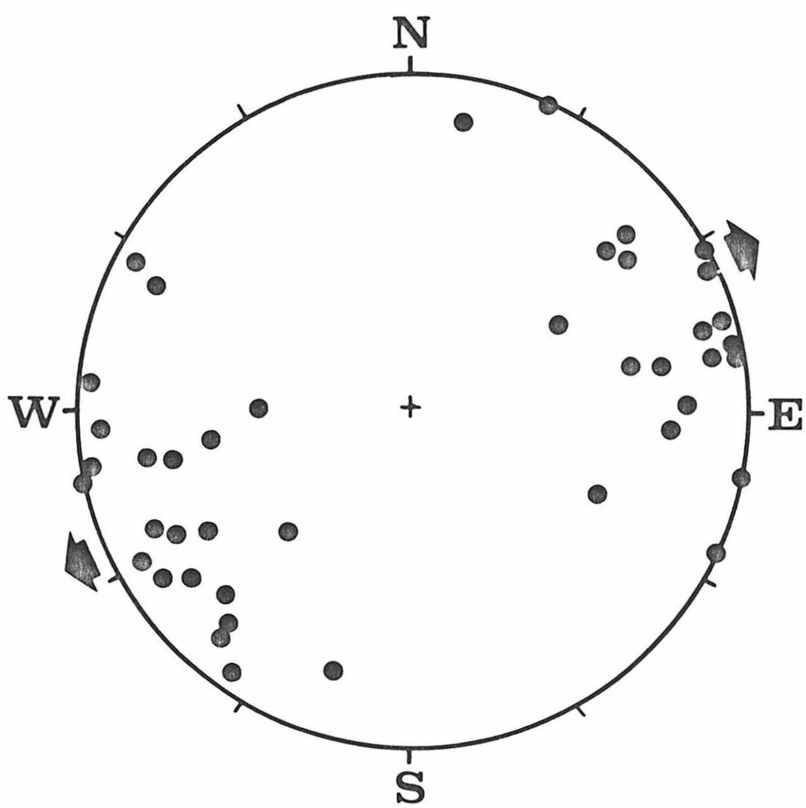
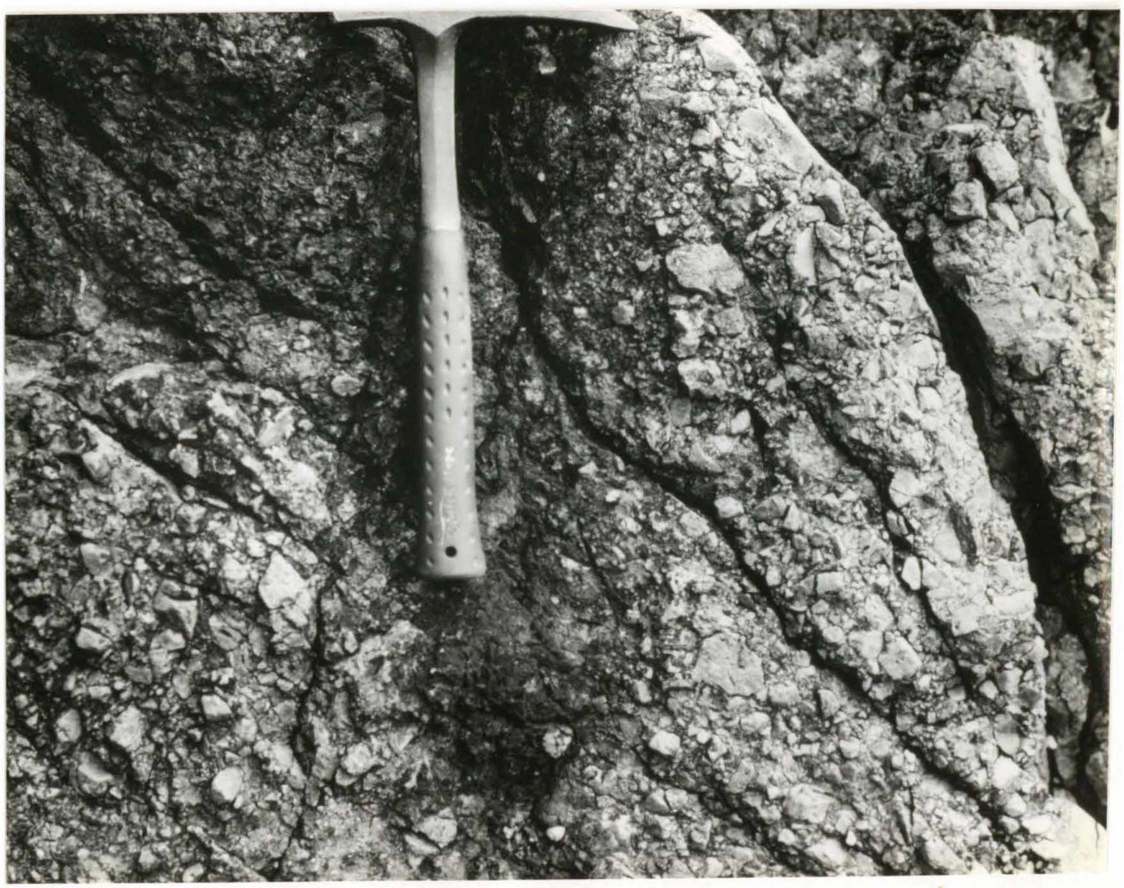


Figure 2-18. Penetratively brecciated Ordovician leucogranodiorite in southern Klakas Inlet. Deformation is known to be pre-middle Early Devonian in age, and is interpreted to have been caused by movement on the Anchor Island and related thrust faults.



are related to movement on the Anchor Island or closely associated thrust faults. The superjacent sedimentary breccia is interpreted to have been deposited as a talus breccia during movement on faults in this thrust system. Toward the north the thrust system and associated zone of deformation are truncated by the Keete Inlet fault.

Keete Inlet fault

The Keete Inlet fault is a major stratigraphic boundary in the southern Alexander terrane (Fig. 2-2). In the study area it dips moderately eastward and juxtaposes Ordovician stratified and intrusive rocks over the Wales metamorphic suite and Ordovician metaplutonic rocks. North of the area the fault juxtaposes Devonian strata against the Wales metamorphic suite and is cut by a mid-Cretaceous pluton (Herreid and others, 1978; Redman, 1981). These relations demonstrate that the fault moved after deposition of the Lower Devonian strata -- much later than movement on the thrust faults in the area. In the northeastern corner of Kassa Inlet, and north and east of the study area, rocks immediately adjacent to the fault are only moderately deformed, and the degree of deformation decreases rapidly away from the fault. Toward the south the fault cuts across deformational fabrics in the upper plate that are presumably related to the Anchor Island and related thrust faults. South of Kassa Inlet the fault strikes north-northwesterly and is parallel to the trace of the older thrust faults, suggesting that the thrust faults may have been reactivated by movement on the Keete Inlet fault. South of Klakas Inlet the fault swings westward into Cordova Bay and is not seen again.

Based on its curviplanar trace and regional juxtaposition of younger rocks over older rocks, I raise the possibility that the Keete

Inlet fault is a low-angle normal fault, rather than a thrust fault as interpreted by Herreid and others (1978) and Redman (1981). This sense of movement is consistent with my interpretation that the Ordovician metaplutonic rocks below the Keete Inlet fault are deeper-level equivalents of the large plutons of Middle Ordovician diorite and granodiorite above the fault. The direction of movement on this fault has not been determined, and its age of movement is constrained only as post-middle Early Devonian and pre-mid-Cretaceous. Movement on the fault may have occurred, however, during a latest Paleozoic(?)-Triassic rifting event in the southern Alexander terrane (Gehrels and others, in review).

Strike-slip faults

Strike-slip faults on southern Prince of Wales Island belong to three sets, including: north-northeast-striking right-lateral faults, a complex set of curviplanar northwest-striking faults with left-lateral displacement, and a north-striking right-lateral fault. North-northeast-striking faults along the southern shore of the island have shallow-plunging slickenside striae and separate Devonian strata up to several kilometers in a dextral sense (Fig. 2-2). These faults are cut by the Nichols Bay fault, which offsets stratigraphic and intrusive units and their contacts by several kilometers in a left-lateral sense and has shallow-plunging slickenside striae. Several curviplanar faults structurally connect the Nichols Bay fault to the Max Cove fault, which has approximately a kilometer of left-lateral offset.

The Max Cove fault and the two northwestern strands of the Nichols Bay fault continue northwestward across Klakas Inlet, but do not offset the Keete Inlet fault. Because the Keete Inlet fault is intruded by a mid-Cretaceous pluton north of the study area, the northwest-striking faults and older north-northeast-striking faults must also have moved prior to mid-Cretaceous time. The youngest strike-slip fault recognized in the area strikes north-south along the southeastern corner of the island (near Cape Chacon) and apparently offsets the Max Cove fault by approximately a kilometer in a right-lateral sense.

Summary of movement on faults

Based on a series of stratigraphic, intrusive, and structural relations, I have been able to establish a chronology for movement on the major sets of faults in the area. Thrust faults clearly offset or deform rocks of Middle Ordovician age and are either overlain by Lower Devonian strata or are intruded by a Late Silurian pluton. If uplift and erosion of the Ordovician-Early Silurian rocks is interpreted to have occurred at the same time as movement on the thrust faults, the occurrence of latest Ordovician-earliest Silurian intrusive rocks beneath Lower Devonian strata indicates that the faults moved during Silurian time. This interpretation is supported by stratigraphic and intrusive relations on Annette, Gravina, and Duke Islands and stratigraphic relations on central Prince of Wales Island which indicate that deformation and uplift associated with the Klakas orogeny began during middle Silurian time (Gehrels and others, in review; Gehrels and Saleeby, in review b).

Devonian strata are offset by north-northeast-striking right-lateral faults, which are cut by northwest-striking left-lateral faults, which in turn are cut by the Keete Inlet fault and by a north-striking, right-lateral fault. The observation that a mid-Cretaceous pluton intrudes the Keete Inlet fault north of the study area demonstrates that all of these faults (except, perhaps, for the north-striking fault) moved prior to mid-Cretaceous time.

DISCUSSION OF U-Pb GEOCHRONOLOGIC DATA

Most samples analyzed from southern Prince of Wales Island yield concordant U-Pb (zircon) apparent ages, where concordance is recognized on Pb*/U diagrams by overlap (at the 95% level) of individual zircon fractions with concordia. Samples which yield discordant dates (lack of overlap of individual zircon fractions and concordia) include the leucogranodiorite from sample locality 3 and the diorite from locality 4 (Figs. 7 and 8). The degree of discordance of all three zircon fractions from sample 3 indicate that the isotopic systems of the zircon have been strongly disturbed, and the discordance pattern suggests that disturbance occurred through loss of Pb relative to U (Fig. 2-7). Sample 4 yields isotopic data which are consistent with a similar isotopic evolution (Fig. 2-8).

Post-Devonian Pb*/U dates and the young (<100 Ma) lower intercept of sample 3 indicate that the isotopic systems of samples 3 and probably 4 were disturbed primarily during Mesozoic-Cenozoic time. This young disturbance is interpreted to have occurred during a regional Late Cretaceous-early Tertiary hydrothermal event, which is recorded by the non-equilibrium oxygen isotope composition of quartz and feldspar pairs in lower Paleozoic and Cretaceous-Tertiary intrusive rocks (Gehrels and Taylor, 1984). Hydrothermal fluids probably reached 300°C during this event, which is the temperature recorded by the color alteration index of Lower Devonian conodonts in the area (Savage and Gehrels, 1984).

Most K-Ar (hornblende) dates from southern Prince of Wales Island are consistent with the available geologic and geochronologic constraints. The K-Ar (hornblende) apparent date of 428 ± 13 Ma from granodioritic rocks near localities 3 and 4 (Turner and others, 1977) is known to be significantly reset, however, because this rock is intruded by the 468 ± ~15 Ma leucogranodiorite at locality 3.

Paleozoic age samples which have been isotopically disturbed are therefore apparently restricted to the Klakas Inlet region, which is also the region in which rocks were strongly deformed and highly altered during the Klakas orogeny. I suggest that the spatial association of samples with isotopically disturbed zircon populations, the sample with a reset K-Ar (hornblende) date, and rocks that have been regionally deformed and altered is not coincidental, and that disturbance of these samples occurred at least in part due to processes operating during the Klakas orogeny. Such processes might include deformation and recrystallization of zircon grains (recorded by fractured and contorted grains), hydrothermal alteration (evidenced by the penetrative dolomitic alteration throughout the zone of deformation), and uplift and weathering of the rocks (indicated by proximity of the sample localities to the basal contact of overlying Lower Devonian strata). The isotopic systems may also have been disturbed by hydrothermal alteration and uplift during a latest Paleozoic(?)-Triassic rifting event (Gehrels and others, in review), which may be manifest on southern Prince of Wales Island by swarms of mafic dikes and normal-slip displacement on the Keete Inlet fault.

Although a unique explanation of the discordance of samples 3 and 4 is not possible with my limited data, the relations described above suggest that their isotopic systems were disturbed initially during deformation, alteration, and uplift associated with the Klakas orogeny, and finally during a Late Cretaceous-early Tertiary hydrothermal event. Latest Paleozoic(?)-Triassic uplift and hydrothermal activity, as well as other mechanisms or episodes of disturbance may have also played a role in the complex isotopic evolution of these samples.

GEOLOGIC AND TECTONIC HISTORY

The recorded geologic history of southern Prince of Wales Island begins in late Proterozoic(?)-Cambrian time with the deposition of basaltic to andesitic volcanic rocks, volcaniclastic graywacke, and subordinate limestone and rhyolitic volcanic rocks that are protoliths of rocks in the Wales metamorphic suite. Similarities between these strata and volcanic and sedimentary rocks in modern-day volcanic arcs indicate that rocks in this suite may have accumulated in a volcanic arc environment along a convergent margin. The Wales suite may thus represent the oldest intact arc assemblage in the Cordillera. During Middle Cambrian-Early Ordovician time these rocks were metamorphosed to greenschist and locally amphibolite facies and were penetratively deformed. Shallow-plunging upright folds with kilometer-scale wavelengths formed during the waning stages of, or after, the main phase of deformation and metamorphism, and prior to the end of Early Ordovician time. Asymmetric outcrop-scale folds probably formed as parasitic folds on the limbs of these regional structures. I have referred informally to the tectonic event manifest by deformation, metamorphism, and uplift of these rocks as the Wales "orogeny" (Gehrels and Saleeby, 1984 and in review b).

Beginning soon after this orogenic event, basaltic to andesitic pillow flows, pillow breccia, and tuff breccia, and subordinate rhyolitic volcanic rocks and marine clastic strata of the Descon

Formation were deposited. These rocks accumulated during late Early and Middle Ordovician time, although deposition of a thick section of argillite, mudstone, and graywacke that extends north of the area may have continued into Late Ordovician-Early Silurian time. Large bodies of Middle Ordovician diorite and granodiorite are interpreted to be subvolcanic to the basaltic to rhyolitic rocks in the Descon Formation. In the eastern part of the area, quartz diorite bodies were emplaced during Late Ordovician time, and quartz monzonite, granite, and quartz syenite plutons were emplaced during latest Ordovician-earliest Silurian time. Comparison of the composition, geochemical characteristics, and regional relations of these Ordovician-Early Silurian plutonic, volcanic, and sedimentary rocks with those in modern-day convergent margin assemblages indicate that they formed in an ensimatic volcanic arc environment.

After Middle Ordovician time, and probably following emplacement of the latest Ordovician-earliest Silurian intrusive rocks, rocks in the Wales suite, Descon Formation, and Ordovician-Early Silurian plutonic suite were imbricated on southwest-vergent thrust faults, and rocks in the southern Klakas Inlet area were penetratively brecciated and highly altered. Movement on the Bird Rocks and Shipwreck Point faults occurred prior to the end of Silurian time, but the Anchor Island thrust fault remained active into earliest Devonian time.

Lower Devonian conglomeratic strata were deposited as part of a regional clastic wedge soon after and in part during the waning stages of movement on these thrust faults (Ovenshine and others, 1969; Gehrels and others, 1983b). In the southern part of the study area, these strata accumulated in subaerial environments which had considerable topographic relief. Toward the north they graded into sandstone, siltstone, and shallow-marine limestone, and into laminated mudstone and graptolitic black shale at the north end of the study area. Southward younging of the limestone combined with facies relations in strata high in the section indicate that the finer-grained clastic strata were deposited during a marine transgression on a north-facing paleoslope. The occurrence of a similar northward transition from subaerial to marine depositional environments on central Prince of Wales Island indicates that the paleotopography during Early Devonian time was more complex than a uniform north-facing paleoslope. A significant amount of structural uplift prior to or during deposition of these strata is recorded by the occurrence of latest Ordovician-earliest Silurian intrusive rocks beneath the Lower Devonian rocks.

The Klakas orogeny is recorded on southern Prince of Wales Island by: (1) a change from calc-alkaline plutonism and volcanism to emplacement of sodic plutons, (2) southwest-directed movement on thrust faults, (3) penetrative deformation and alteration of Ordovician rocks in the southern Klakas Inlet region, (4) at least several kilometers of structural uplift of latest Ordovician-earliest Silurian and older rocks, and (5) deposition of coarse conglomeratic strata as part of a regional clastic wedge. Available age constraints from southern Prince of Wales Island indicate that this event began between 438 ± 4 Ma and 418 ± 5 Ma, continued into earliest Devonian time, and ceased prior to the middle Early Devonian. Relations on central Prince of Wales Island and on Annette, Gravina, and Duke Islands demonstrate that the event began after middle Early Silurian time (Gehrels and Saleeby, in review b; Gehrels and others, in review). Cessation of the Klakas orogeny is recorded by middle Lower Devonian strata which were deposited in tectonically stable marine environments.

Several fault systems became active after deposition of the Lower Devonian strata. North-northeast-striking faults along the south shore of the island moved first with up to several kilometers of right-lateral displacement. Movement on these faults was followed by left-slip displacement of up to several kilometers on northwest-striking faults, and then the Keete Inlet fault moved -- apparently as a low-angle normal fault. Post-Devonian rocks include the Bokan Mountain Granite (Jurassic) and large bodies of mid-Cretaceous granodiorite and diorite. The tectonic environment in which the Bokan Mountain Granite was emplaced is poorly known, as Jurassic intrusive bodies are rare in the southern Alexander terrane. The mid-Cretaceous plutons in the area belong to a belt of Early and mid-Cretaceous intrusive bodies that presumably formed in response to plate convergence along the western margin of the Alexander terrane (Berg and others, 1972). One of these intrusive bodies cuts the Keete Inlet fault north of the study area (Herreid and others, 1978), demonstrating a pre-mid-Cretaceous age for movement on the Keete Inlet fault and the north-northeast-striking and northwest-striking faults.

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CHAPTER 3

GEOLOGY OF ANNETTE, GRAVINA, AND DUKE ISLANDS

ABSTRACT

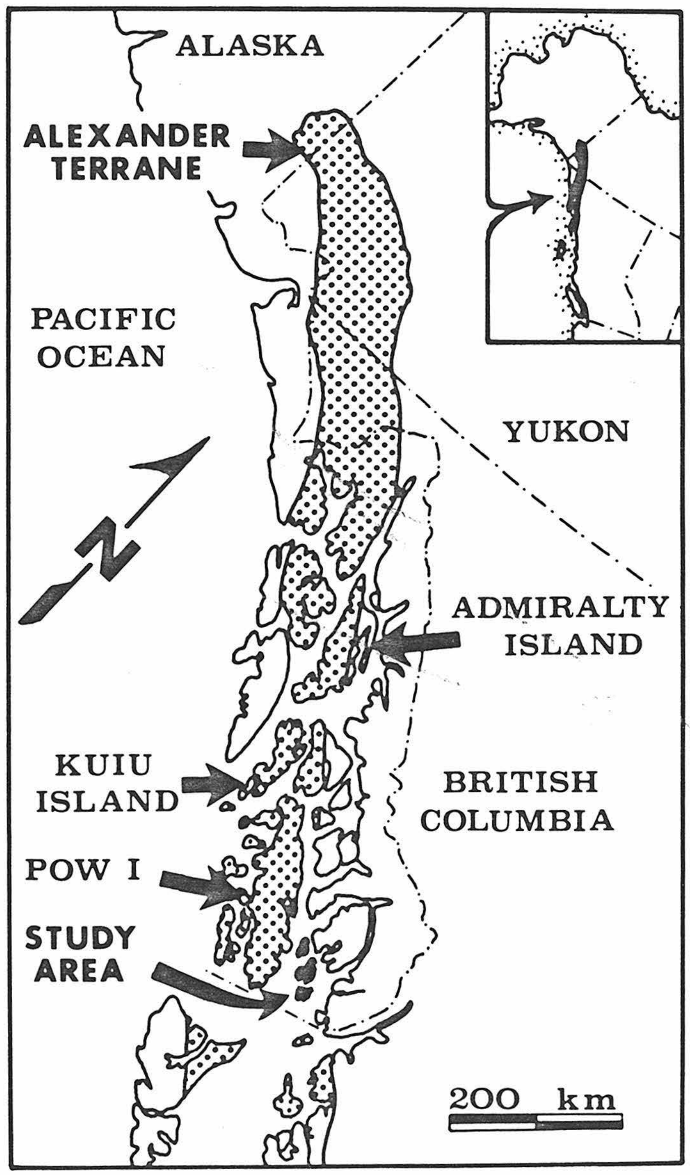
Geologic mapping, U-Pb (zircon) geochronometry, and analyses of conodonts demonstrate that the major pre-Jurassic assemblages on Annette, Gravina, Duke, and adjacent smaller islands include: pre-Ordovician metavolcanic and metasedimentary rocks (Wales metamorphic suite), metaplutonic rocks of Cambrian age, arc-type(?) dioritic rocks and coeval volcanic rocks (Descon Formation) of Ordovician-Early Silurian age, large trondhjemitic plutons of Silurian age, Lower Devonian strata (Karheen Formation), Upper Triassic sedimentary and volcanic rocks (Hyd Group), and a large body of Late Triassic pyroxene gabbro. Stratigraphic, structural, and intrusive relations record episodes of regional deformation, metamorphism, and uplift during Middle Cambrian-Early Ordovician time (Wales "orogeny") and middle Silurian-earliest Devonian time (Klakas orogeny). Upper Triassic strata were apparently deposited during a latest Paleozoic(?)-Triassic rifting event. Comparison with the geology of southern Prince of Wales Island indicates that the Annette and Craig subterranes of the Alexander terrane belong to the same tectonic fragment, and that the Clarence Strait fault has approximately 15 km of right-lateral displacement at this latitude. My geochronologic data also indicate that the pyroxene gabbro on Duke Island is Triassic in age (and therefore not genetically related to the zoned ultramafic bodies) and place a minimum age constraint of 225 ± 3 Ma on the Carnian-Norian boundary.

INTRODUCTION

Triassic and older rocks on Annette, Gravina, and Duke Islands belong to the Alexander terrane, which is a coherent tectonic fragment that underlies much of southeastern (SE) Alaska and parts of western British Columbia, southwestern Yukon, and eastern Alaska (Fig. 3-1: Berg and others, 1972). In most regions, the Alexander terrane is underlain by marine clastic strata and shallow-marine limestone that range in age from early Paleozoic through Triassic. The rocks on Annette, Gravina, and Duke Islands record a somewhat different history, however, in that lower Paleozoic rocks consist predominantly of regionally metamorphosed and deformed volcanic and plutonic rocks, and that upper Paleozoic strata are not present. Based on these differences, Berg and others (1978) subdivided the Alexander terrane into the Annette subterrane, which includes Annette, Gravina, and Duke Islands, and the Craig subterrane, to which rocks in most other areas of the terrane belong.

My studies have focused on the lower Paleozoic rocks on Annette, Gravina, and Duke Islands in an effort to learn about the early Paleozoic evolution of the southern Alexander terrane and evaluate the differences between the Craig and Annette subterranes. I have mapped these rocks in detail along shorelines of the islands, conducted U-Pb (zircon) geochronologic studies of the intrusive rocks, and collected limestone samples for conodont studies. In this report I describe the geologic framework and evolution of Annette, Gravina, Duke, and adjacent smaller islands, and present my U-Pb geochronologic data (Tables 3-1 and 3-2). The regional implications of my work on these islands are reported by Gehrels and Saleeby (in review a), and conodonts from the area are described by Savage and Gehrels (in press).

Fig. 3-1. Location map of the study area and of rocks belonging to the Alexander terrane [derived from Berg and others (1978); Monger and Berg (1984)]. (POW I = Prince of Wales Island).



I adopt the DNAG Time Scale (Palmer, 1983) in relating radiometric ages to geologic time and note, for reference, that the Ordovician-Silurian and Silurian-Devonian boundaries have been assigned ages of 438 ± 12 Ma and 408 ± 12 Ma on this time scale. Plutonic rocks in the area are classified according to guidelines recommended by the IUGS Subcommission on Systematics of Igneous Rocks (Streckeisen, 1976).

Previous work

Buddington and Chapin (1929) originally divided the rocks on Annette, Gravina, and Duke Islands into: metamorphic rocks of Ordovician to Jurassic age; Devonian limestone and clastic strata; Triassic sedimentary and volcanic rocks; and Jurassic or Cretaceous sedimentary, volcanic, and intrusive rocks. Berg (1972a and 1973) mapped the geology of Annette and Gravina Islands in more detail and determined that the granitic intrusive rocks are predominantly pre-Devonian in age, rather than Jura-Cretaceous, and that the metamorphic rocks mapped previously as Ordovician to Jurassic consist of Devonian strata and pre-Devonian metaplutonic and metastratified rocks. Irvine (1974) studied the petrology and field setting of Cretaceous(?) ultramafic rocks on Duke Island in detail and determined the general nature and distribution of pre-Cretaceous rocks on the island.

GEOLOGIC FRAMEWORK

Overview

Annette, Gravina, and Duke Islands are underlain by a variety of sedimentary, volcanic, and intrusive rocks, and their metamorphosed and (or) deformed equivalents, that range in age from Cambrian (and perhaps late Proterozoic) through Cretaceous. These rocks are herein subdivided into geologic units based on their predominant rock type and age. The primary units recognized include: Wales metamorphic suite (pre-Ordovician), Cambrian metaplutonic rocks, Descon Formation (Ordovician-Lower Silurian) and coeval intrusive rocks, Late Silurian trondhjemite to granite, Karheen Formation (Lower Devonian) and associated volcanic rocks, Hyd Group (Upper Triassic) and a coeval gabbro body, Jurassic-Lower Cretaceous strata, and Cretaceous(?) intrusive rocks. Several regional tectonic events are recorded by stratigraphic and structural relations on these islands, including the Wales "orogeny" (Middle Cambrian-Early Ordovician), the Klakas orogeny (middle Silurian-earliest Devonian), an uplift and erosional event during latest Paleozoic(?)-Triassic time, and a major deformational and metamorphic event during Late Cretaceous-early Tertiary time.

Each of the geologic units and tectonic events mentioned above is described in the following sections and the distribution of the main units and structures is shown on Figures 3-2, 3-3, and 3-4. Because my studies focused on the Devonian and older rocks, the reader is referred to Berg (1972a and 1973) and Irvine (1974) for more information on the Mesozoic rocks. My geochronologic data are presented in the text and in Tables 3-1 and 3-2, and complexities of the data are discussed in a separate section on U-Pb geochronology. In the final two sections, I summarize the geologic history recorded by rocks on these islands and outline the main conclusions of my study.

Wales metamorphic suite (pre-Ordovician)

Greenschist- and locally amphibolite-facies metavolcanic and metasedimentary rocks and subordinate marble occur on small islands near the south tip of Gravina Island (Fig. 3-2). These rocks are strongly foliated and layered, and protolith features, where recognizable, are

Fig. 3-2. Geologic sketch map of western Gravina Island and legend for Figures 3-2, 3-3, and 3-4 [in part modified from Berg (1973)].

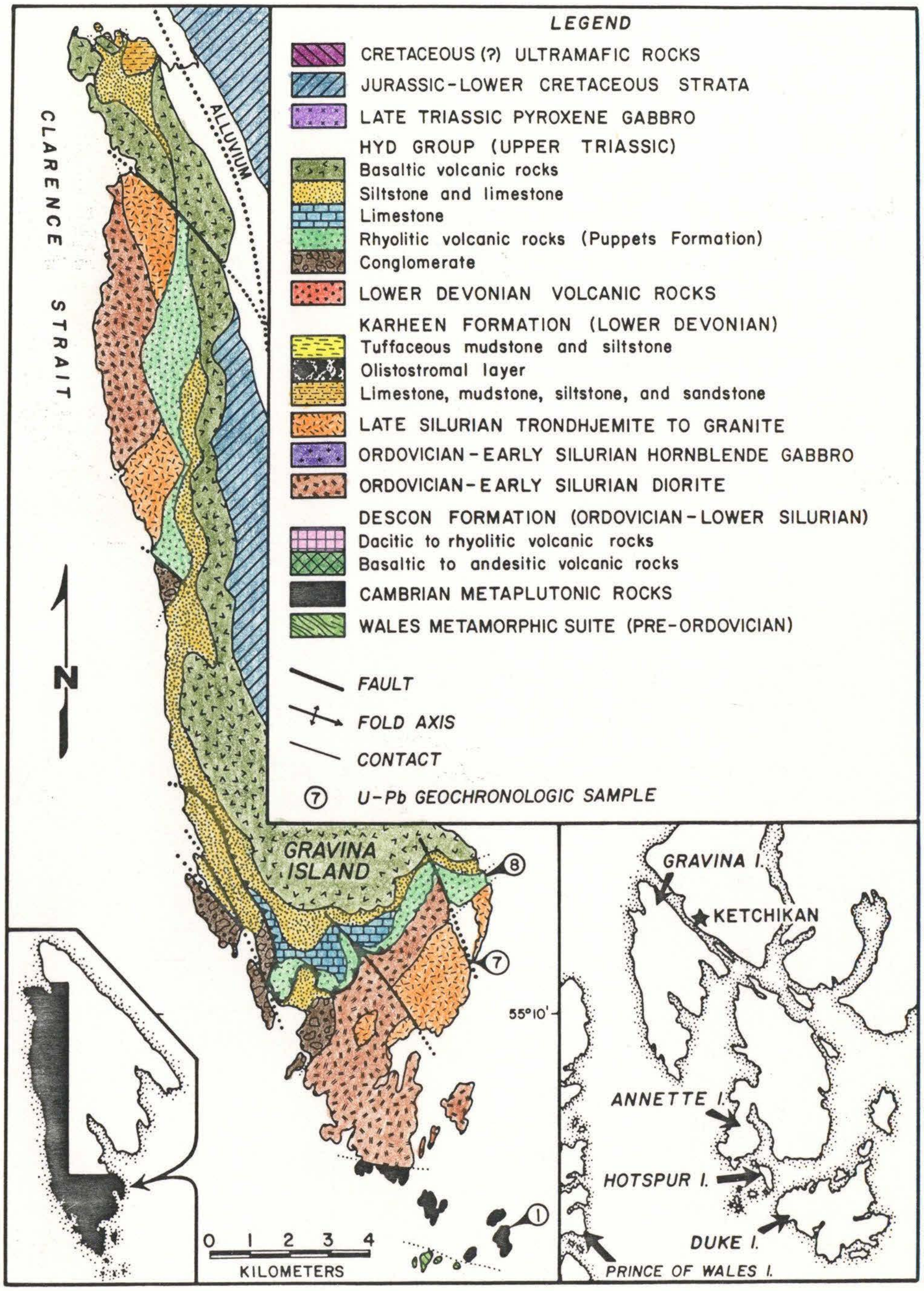

LEGEND
CRETACEOUS (?) ULTRAMAFIC ROCKS
JURASSIC-LOWER CRETACEOUS STRATA
LATE TRIASSIC PYROXENE GABBRO
HYD GROUP (UPPER TRIASSIC)
Basaltic volcanic rocks
Siltstone and limestone
Limestone
Rhyolitic volcanic rocks (Puppets Formation)
Conglomerate
LOWER DEVONIAN VOLCANIC ROCKS
KARHEEN FORMATION (LOWER DEVONIAN)
Tuffaceous mudstone and siltstone
Olistostromal layer
Limestone, mudstone, siltstone, and sandstone
LATE SILURIAN TRONDHJEMITE TO GRANITE
ORDOVICIAN-EARLY SILURIAN HORNBLENDE GABBRO
ORDOVICIAN-EARLY SILURIAN DIORITE
DESCON FORMATION (ORDOVICIAN-LOWER SILURIAN)
Dacitic to rhyolitic volcanic rocks
Basaltic to andesitic volcanic rocks
CAMBRIAN METAPLUTONIC ROCKS
WALES METAMORPHIC SUITE (PRE-ORDOVICIAN)
FAULT
FOLD AXIS
CONTACT
⑦ U-Pb GEOCHRONOLOGIC SAMPLE
CLARENCE STRAIT
ALLUVIUM
N
GRAVINA ISLAND
8
7
1
55°10'
0 1 2 3 4
KILOMETERS
GRAVINA I.
KETCHIKAN
ANNETTE I.
HOTSPUR I.
DUKE I.
PRINCE OF WALES I.

Fig. 3-3. Geologic sketch map of southern Annette Island [in part modified from Berg (1972a)]. Legend is on Figure 3-2.

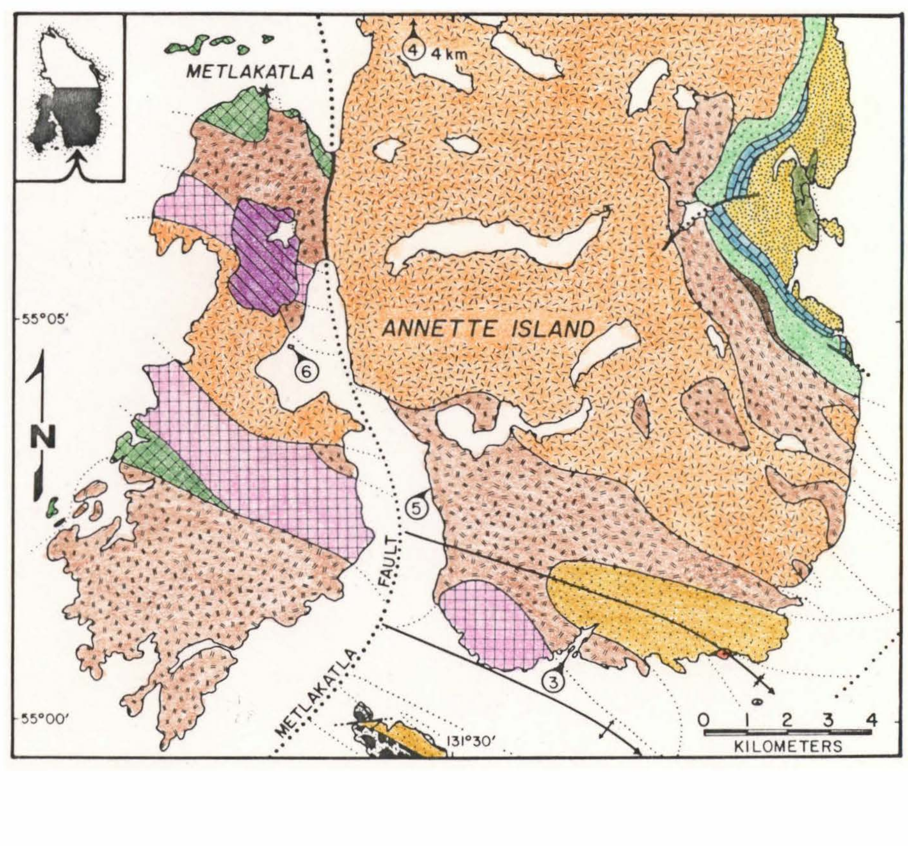
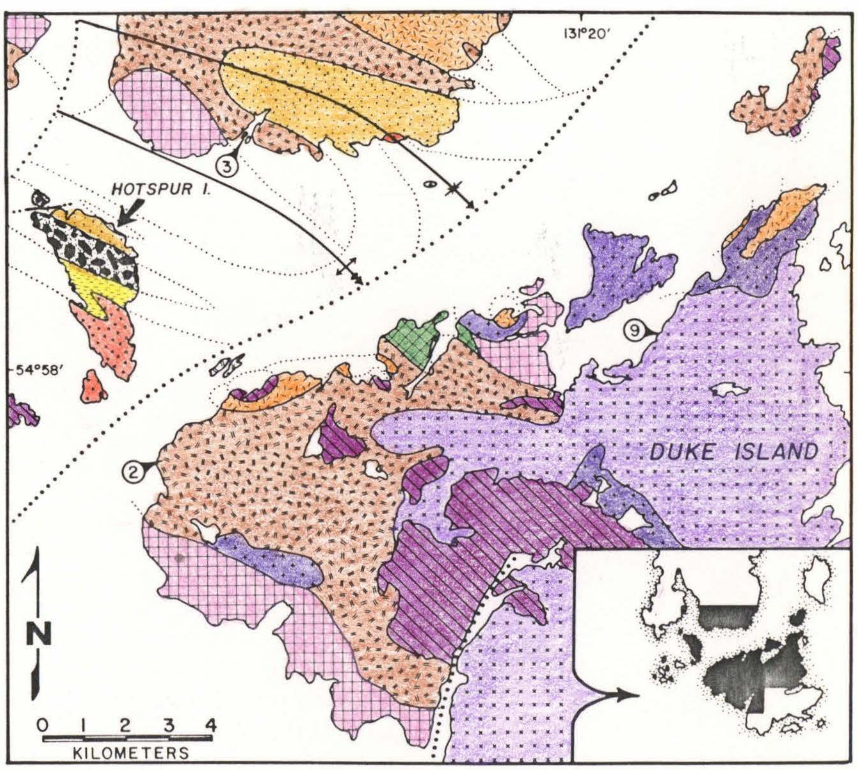


Fig. 3-4. Geologic sketch map of northern Duke Island, Hotspur Island, and southern Annette Island [in part modified from Irvine (1974)]. Legend is on Figure 3-2.



highly flattened and moderately elongated. Their foliation generally strikes west-northwesterly and dips steeply to the south, and the elongation lineation plunges to the west at shallow angles. Meter-scale asymmetric folds in the metamorphic foliation occur locally. These folds do not have an axial planar foliation and are interpreted to have formed during the waning stages of, or after, the metamorphism and deformation. Protolith features include centimeter-scale graded beds in metasedimentary rocks and pyroclastic fragments in metavolcanic rocks. The general composition of these rocks indicates that the protoliths were basic to intermediate (probably basaltic to andesitic) in composition. Most rocks consist primarily of chlorite, actinolite, biotite, epidote, albite-oligoclase, quartz, opaque minerals, and secondary calcite and white mica. The marble is coarsely recrystallized, white to grayish-tan in color, and occurs in highly deformed lenses and layers up to several meters in thickness.

Contact relations between rocks in the Wales metamorphic suite and other units are not exposed on these small islands. On adjacent islands, however, metaplutonic rocks of Middle to Late Cambrian age (described below) have deformational fabrics which are similar in style and orientation to those in the Wales suite. These fabrics are locally cut by dikes of Ordovician-Early Silurian diorite (described below). The deformation and metamorphism of rocks in the Wales suite are accordingly interpreted to have occurred after (or perhaps during) emplacement of the metaplutonic rocks, and prior to emplacement of the Ordovician-Early Silurian diorite.

Pre-Ordovician metavolcanic and metasedimentary rocks on Prince of Wales Island were originally referred to as the Wales Group (Buddington and Chapin, 1929) but have since been renamed the Wales metamorphic suite (Gehrels and Saleeby, in review b). On south-central Prince of Wales Island, these rocks are interlayered with dikes and small intrusive bodies of Middle and Late Cambrian metadiorite (J. Saleeby, unpub. data) and deformational fabrics in both suites are intruded by Ordovician-Early Silurian diorite. Relations on both Gravina and Prince of Wales Islands therefore indicate that rocks in the Wales suite were deposited prior to Late Cambrian time and were metamorphosed and deformed during Middle Cambrian-Early Ordovician time. I refer informally to this phase of deformation, metamorphism, and perhaps uplift as the Wales "orogeny" (Gehrels and Saleeby, 1984, in review a). More complete descriptions of the nature, age, and regional significance of this tectonic event await detailed structural studies of rocks in the Wales suite.

Cambrian metaplutonic rocks

Southernmost Gravina Island and adjacent smaller islands are underlain in part by an intrusive suite consisting predominantly of foliated and slightly layered diorite, elongate meter-scale inclusions of foliated gabbro and hornblendite, and dikes and sills of foliated quartz diorite, microdiorite, and microgabbro (Figures 3-2 and 3-5). Contacts between most intrusive bodies are parallel to the regional west-northwest-striking, steeply-dipping foliation developed within the bodies. Locally, however, quartz dioritic and subordinate dioritic dikes intrude at oblique angles to the foliation in their country rocks. Foliation in the cross-cutting dikes is locally parallel to the dike margins, but is more commonly parallel to the foliation in the country rocks. These intrusive relations indicate that deformational

Fig. 3-5. Photograph of layered and foliated diorite and quartz diorite on small islands south of Gravina Island. Age sample 1 is from the foliated quartz diorite member of this outcrop and yields U-Pb data consistent with an age in the 540-510 Ma range. Approximately 10 m of section are shown on this photograph.

fabrics in the suite were imparted during emplacement of the various members. The minimum age of this deformation is constrained by cross-cutting dikes of Ordovician-Early Silurian diorite (described below) which are less deformed.

A sample of foliated quartz diorite from this intrusive suite yields U-Pb data which are consistent with an apparent age in the 540-510 Ma (Middle-Late Cambrian) range (sample 1). This approximate age is interpreted to represent the time of formation of the suite, as the quartz diorite body sampled intrudes foliated diorite, and is in turn intruded by a folded dike of foliated diorite (Fig. 3-5).

Similar layered and foliated metaplutonic rocks on Prince of Wales Island yield U-Pb isotopic data similar to that of sample 1 and are sufficient to demonstrate Middle and Late Cambrian apparent ages (J. Saleeby, unpub. data). These metaplutonic rocks are intruded by large bodies of less-deformed diorite of Ordovician-Early Silurian age, which demonstrates that they were deformed and metamorphosed during Middle Cambrian-Early Ordovician time. As described above, the structural and intrusive relations on southern Gravina Island are consistent with this timing.

The regional distribution, structural trends, and contact relations of Upper Triassic sedimentary breccia, Ordovician-Early Silurian diorite, Cambrian metaplutonic rocks, and rocks in the Wales suite are quite similar on southern Gravina Island and on eastern Prince of Wales Island (Fig. 3-6). These similarities demonstrate that Prince of Wales Island (Craig subterrane) and Gravina Island (Annette subterrane) belong to the same tectonic fragment and that the Clarence Strait fault has approximately 15 km of right-lateral offset at this latitude (Fig. 3-6).

Table 3-1. U-Pb (zircon) isotopic data and apparent ages.

SMPL #	WT (mg)	SIZE	CONC. (ppm) ^{238}U	CONC. (ppm) $^{206}Pb^*$	$\frac{Pb^*}{Pb}$ AT%	ISOTOPIC COMPOSITION $\frac{^{206}Pb}{^{204}Pb}$	$\frac{^{206}Pb}{^{207}Pb}$	$\frac{^{206}Pb}{^{208}Pb}$	APPARENT DATES (Ma) $\frac{^{206}Pb^*}{^{238}U}$	$\frac{^{207}Pb^*}{^{235}U}$	$\frac{^{207}Pb^*}{^{206}Pb^*}$	APPARENT AGE (Ma)
1	17.3	B	156.6	11.14	99.6	15500	16.981	7.460	509.4	513.2	530	~540-510$^{\psi}$
2a	5.8	A	512.5	28.57	99.3	9610	17.658	13.760	402.5	404.8	417.9	429 ± ~20$^{\beta}$
2b	7.6	B	736.6	41.92	99.7	22800	17.918	13.597	410.6	412.2	421.2	"
2c	16.2	B	593.2	33.44	99.7	22730	17.925	13.591	406.8	408.8	420.3	"
2d	13.2	D	882.0	48.01	99.8	25200	17.989	11.122	393.3	396.4	414.3	"
3a	20.1	A	274.3	14.80	99.4	9610	17.588	9.249	390.0	395.6	428.2	426 ± ~15$^{\beta}$
3b	20.5	A	229.7	12.36	99.3	8590	17.582	9.4702	389.0	393.8	421.8	"
3c	18.5	B	214.8	12.163	98.5	4070	16.990	8.479	408.7	411.3	425.7	"
3d	12.0	C	308.2	16.46	99.5	11020	17.647	8.420	386.0	392.1	428.3	"
3e	5.6	D	402.2	20.88	99.2	7190	17.467	8.438	375.5	382.3	423.6	"
4a	17.0	A	189.8	9.673	97.0	1980	16.148	8.1914	369.0	373.6	402.4	409 ± ~30$^{\beta}$
4b	21.0	A	153.9	7.069	97.5	2374	16.440	8.583	333.5	342.2	402	"
4c	25.3	B	183.9	8.260	97.2	2190	16.39	8.504	326.3	334.2	391	"
4d	26.2	C	209.7	9.065	96.9	1960	16.17	8.181	314	323	391	"
5a	10.2	B	220.4	12.67	98.8	5190	17.297	9.875	414.7	414.7	414.6	415 ± 5$^{\alpha}$
5b	14.4	CD	368.5	19.71	99.5	11200	17.781	9.572	386.6	390.2	411.7	"
6a	6.1	A	95.16	5.239	97.6	2360	16.381	7.119	397.6	399.4	409	~430-390$^{\psi}$
6b	12.8	B	499.5	27.18	99.7	18200	17.985	8.1244	393.2	395.1	406.2	"
6c	14.7	B	489.7	26.60	99.3	8840	17.758	8.077	392.5	393.7	400.7	"
6d	37.9	C	647.6	34.57	98.4	3880	17.057	7.365	386.0	389.3	409.3	"
7a	4.6	AB	348.6	19.38	97.4	2370	16.32	10.67	401	404	420	410 ± ~20$^{\beta}$
7b	5.0	D	501.2	26.21	96.9	2027	16.038	12.069	378.3	384.4	421.6	"
7c	7.7	D	418.6	21.50	96.7	1910	15.820	12.194	371.8	381	437	"
8	16.4	C	492.4	15.12	96.0	1488	16.546	6.182	224.8	224.9	225.3	225 ± 3$^{\alpha}$
9	10.0	A	410.9	12.68	99.7	20400	19.501	9.18	225.9	225.8	224.4	226 ± 3$^{\alpha}$

* radiogenic lead.

$^{\alpha}$ Uncertainty of age is reported at the 95% level.

$^{\beta}$ Apparent age is interpreted from the upper intercept on Pb*/U concordia diagrams and uncertainty is estimated graphically (Fig. 3-8a-c and e).

$^{\psi}$ Age is probably within range cited, but data are not sufficient to determine a reliable apparent age (Fig. 3-8d)

Notes: Additional sample information is provided in Table 3-2.

Size of zircon fractions: A = 120-165u; B = 80-120u; C = 45-80u; D = < 45u.

Constants used: $\lambda^{238}U = 1.55125 \times 10^{-10}$; $\lambda^{235}U = 9.8485 \times 10^{-10}$; and $^{238}U/^{235}U$ (atomic) = 137.88 (from Steiger and Jager, 1977).

95%-level uncertainty in concentrations, isotopic composition, and apparent dates is in last two figures cited.

Isotopic compositions cited above have not been adjusted. In calculating concentrations and apparent dates the isotopic data have been adjusted as follows:

1) Mass-dependent correction factors of .09% ± .04% (per AMU) and 0.12 ± .04% (per AMU) have been applied to lead and uranium ratios, respectively. These factors have been determined by replicate analyses of NBS SRM 981, 982, 983 for Pb and U-500 and U-930 for U.

2) The $^{207}Pb/^{206}Pb$ of the spiked aliquot has been adjusted for the ^{206}Pb and ^{207}Pb added with the spike.

3) Isotopic composition of the unspiked aliquot has been adjusted for 0.04 ±.02 ng blank lead with a composition of: $^{206}Pb/^{204}Pb$ = 18.78 ± 0.30; $^{207}Pb/^{204}Pb$ = 15.61 ± 0.22; and $^{208}Pb/^{204}Pb$ = 38.5 ± .60. The amount of blank Pb has been determined by isotope dilution

analysis of a typical dissolution and chemical separation procedure conducted without zircon. The blank composition has been determined through isotopic analysis of acids and dust particles in the laboratory.

4) Isotopic composition of U in the spiked aliquot has been adjusted for .07 ± .05 ng blank U, which has been determined by the same procedure as the Pb blank.

5) Isotopic composition of the spiked aliquot has been adjusted for blank Pb by balancing the $^{206}Pb/^{207}Pb$ of the spiked and unspiked aliquots. Blank Pb in the spiked aliquot is assigned the composition cited above.

6) Initial lead is assigned a composition of: $^{206}Pb/^{204}Pb$ = 18.0 ± 1.5; $^{207}Pb/^{204}Pb$ = 15.59 ± 0.4; and $^{208}Pb/^{204}Pb$ = 37.8 ± 2.0 for Paleozoic samples, and 18.2, 15.57, and 38.0 (with the same uncertainties) for Triassic samples. (Values interpreted from Doe and Zartman, 1979).

The concentrations and apparent dates cited above and the Pb*/U ratios, uncertainties, and error correlations shown in Figure 3-8 have been calculated with a computer program written by Ken Ludwig (1983).

Table 3-2. U-Pb geochronologic sample information.

Smpl No.	Station No.	Latitude (N)	Longitude (W)	Magnetic Character
1	84GG49	55°07'10"	131°42'21"	NM 2°/20°

Rock: Strongly foliated, medium- to coarse-grained hornblende quartz diorite.

Zircon: Eu=Sb>An, 2:1 to 3:1, well-rounded forms, inclusions in some grains, medium to dark pink in color.

2	81GD20	54°56'49"	131°29'20"	M 0°/15°

Rock: Strongly foliated medium-grained biotite>hornblende diorite.

Zircon: Sb=Eu>An, 1.5:1 to 2.5:1, tiny inclusions in most grains.

3	81GA3	55°00'55"	131°27'21"	NM 3°/15°

Rock: Extremely altered, medium-grained, hornblende(?) diorite.

Zircon: Sb>Eu>An, 1:1 to 2:1, most grains have many inclusions, light to medium pinkish lavender.

4	79JD975	55°10'56"	131°30'57"	NM 10°/20°

Rock: Moderately altered, coarse-grained, hornblende quartz diorite.

Zircon: An>Sb>>Eu, 1:1 with very irregular shapes, inclusions in most grains, light pinkish lavender.

5	80JA3	55°02'52"	131°30'27"	NM 10°/20°

Rock: Slightly altered, coarse-grained, hornblende(?) leucotrondhjemite.

Zircon: Eu>Sb>An, 1.5:1 to 2.5:1, generally rounded, lots of inclusions, light lavender in color.

6	80JA11	55°04'40"	131°33'37"	NM 10°/20°

Rock: Foliated, moderately altered, medium-grained biotite trondhjemite.

Zircon: Eu=Sb=An, 1:1 to 2:1, very heterogeneous population, most grains have inclusions, slight yellowish tint.

7 80JG6 55°10'20" 131°43'15" NM 10°/20°

Rock: Moderately altered, coarse-grained, leucotrondhjemite.

Zircon: Eu>Sb>>An, 2:1 average, small inclusions in about half the grains, light pink with slight lavender tint.

8 81GG6 55°11'49" 131°42'51" NM 10°/20°

Rock: Moderately altered aphyric rhyolite.

Zircon: Sb>Eu>>An, 1:1 and generally tetrahedral, few inclusions, colorless.

9 81GD38 54°58'11" 131°18'22" NM 2°/10°

Rock: Unaltered, medium-grained, hypersthene-augite-biotite gabbro.

Zircon: An>Sb>>Eu, 1:1, very irregular shapes, colorless.

Notes: Magnetic characteristics; NM = Non-magnetic split; M = Magnetic split; first angle is side slope on Frantz Isodynamic Separator; second angle is forward slope. All samples were processed at 1.7 amps current.

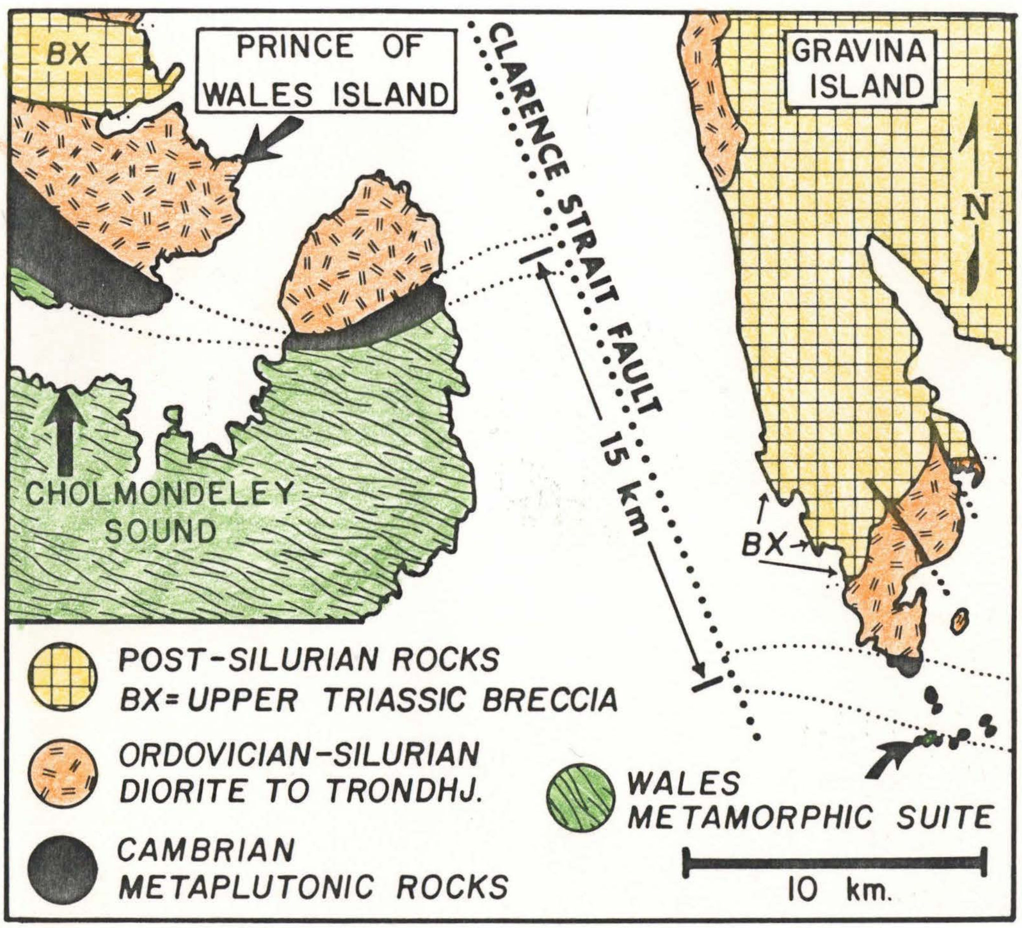
Eu = Euhedral, Sb = Subhedral, An = Anhedral.

2:1 = length:width of grains.

Color determined under reflected light.

Inclusions generally consist of colorless narrow rods and tiny dark-brown grains.

Fig. 3-6. Sketch map showing the correlation of rocks on southern Gravina Island and on eastern Prince of Wales Island. Based on this correlation the Clarence Strait fault is interpreted to have approximately 15 km of right-lateral offset. Geologic relations on eastern Prince of Wales Island are from Gehrels and Saleeby (unpub. mapping, 1984), Eberlein and others (1983), and Buddington and Chapin (1929, pp. 313).



Descon Formation (Ordovician-Early Silurian)

Western Annette and Duke Islands are underlain in part by moderately deformed and metamorphosed dacitic to rhyolitic tuff and tuff breccia, basaltic to andesitic pillow flows, pillow breccia, and tuff breccia, and subordinate volcaniclastic graywacke, black argillite, and limestone (Figures 3-3 and 3-4). Although fossils have not been recovered from these rocks, their minimum age is constrained by cross-cutting bodies of Ordovician-Early Silurian diorite, and their maximum age by the lack of Middle Cambrian-Early Ordovician deformational fabrics characteristic of rocks in the Wales metamorphic suite. Similar volcanic and sedimentary rocks on Prince of Wales Island range in age from Early Ordovician to Early Silurian and have been assigned in most areas to the Descon Formation (Eberlein and others, 1983; Herreid and others, 1978; Gehrels and Saleeby, in review b). Based on their similarity in constituent rock types, stratigraphic position, and age, I herein assign the pre-Late Silurian, post-Cambrian volcanic and sedimentary rocks on Annette and Duke Islands to the Descon Formation.

On Annette Island the Descon Formation consists primarily of dacitic to rhyolitic tuff and tuff breccia with well-preserved pyroclastic fragments and tuffaceous lamination and layering. Basaltic to andesitic pillow flows and tuff breccia are interbedded with the silicic volcanic rocks in most areas and are the predominant member near Metlakatla (Fig. 3-3). These rocks rarely preserve protolith features and are generally recrystallized to semi-schistose chlorite, albite, and epidote. On small islands north of Metlakatla (Fig. 3-3), however, well-preserved basaltic to andesitic tuff and tuff breccia are interbedded with black argillite and layered and finely laminated mudstone. These rocks are only slightly deformed and have clearly not experienced the metamorphism and deformation of nearby rocks in the Wales metamorphic suite. The metamorphic grade of rocks in the Descon Formation increases southward from Metlakatla to amphibolite facies (Berg, 1972a).

The dominant rock type in the Descon Formation on Duke Island is moderately silicic, finely laminated tuff with layers of tuff breccia. Individual fragments in the breccia range up to 20 cm in length and are in most rocks only moderately flattened. Rocks along the western shore of the island are slightly schistose, cut by many faults, and are commonly folded at outcrop scale. A 50-meter-thick layer of quartz-porphyritic dacite can be traced along this shoreline for over 4 km, however, which demonstrates that the rocks belong to a stratigraphic sequence which has not been highly folded or disrupted. Interbedded with the dacitic to rhyolitic volcanic rocks, particularly along the north shore of the island, are basaltic to andesitic pillow flows and breccia, volcaniclastic graywacke and mudstone turbidites, and several lenses of silicic breccia that may have been extrusive domes. Thin sulfide-rich layers occur locally in the silicic rocks.

Ordovician-Early Silurian diorite

Dioritic rocks on Annette, Gravina, and Duke Islands are quite variable in their style and degree of deformation and metamorphism and occur in homogeneous bodies and in heterogenous intrusive suites. Where protolith relations are preserved, the rocks commonly consist of: massive and homogeneous diorite; agmatite with several-centimeter- to several-meter-scale inclusions of hornblendite, gabbro, and microgabbro; or heterogeneous complexes of diorite, quartz diorite, and gabbro dikes and small intrusive bodies. Hornblende, andesine, quartz, K-feldspar, and biotite are the dominant minerals in the rocks, and secondary epidote, chlorite, calcite, and white mica are widespread. In most areas the dioritic rocks have a moderate foliation defined by flattening of protolith features and alignment and elongation of ferromagnesian minerals. Some rocks have a compositional layering parallel to the foliation. In contrast to the regionally consistent foliation in the pre-Ordovician rocks, the foliation in the dioritic rocks is variable in orientation, both regionally and at outcrop scale.

On Gravina Island, dioritic dikes with a slight foliation cut across deformational fabrics in the Cambrian metaplutonic rocks, and the dioritic rocks are in turn cut by nondeformed dikes of Late Silurian trondhjemite (Fig. 3-7). Dioritic rocks on Duke Island are in most areas foliated and layered -- locally to the degree that primary igneous textures have been obliterated. Irvine (1974) mapped the strongly layered diorite on parts of Duke Island as metasedimentary and metavolcanic rock, but more detailed mapping demonstrates that these rocks grade with decreasing deformation into foliated diorite. A sample of foliated and layered diorite from northwestern Duke Island yields a U-Pb apparent age of 429 ± ~20 Ma (Fig. 3-8a: sample 2).

On Annette Island the dioritic rocks are quite variable in composition and in style and degree of deformation. The large diorite body south of Metlakatla is foliated and layered and resembles the diorite on northwestern Duke Island. Dioritic rocks along the south shore of the main part of the island are schistose and highly altered, and most of the primary igneous minerals have been recrystallized to chlorite, white mica, epidote, calcite, and quartz. Age sample 3 is

Fig. 3-7. Photograph of a Late Silurian trondhjemite dike cutting across the foliation in diorite of probable Ordovician-Early Silurian age on southern Gravina Island. Trondhjemite nearby yields a U-Pb apparent age of 410 ± ~20 Ma (sample 7).

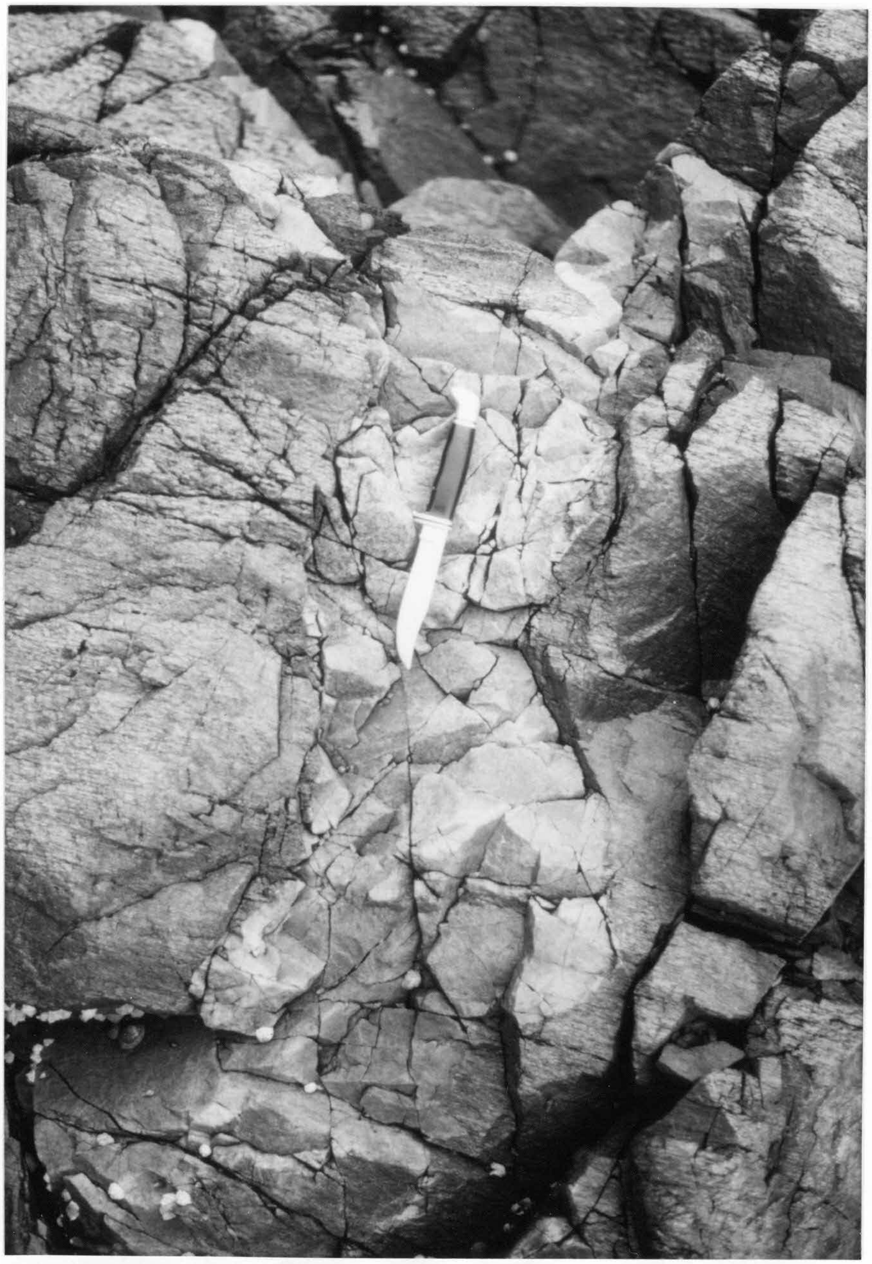
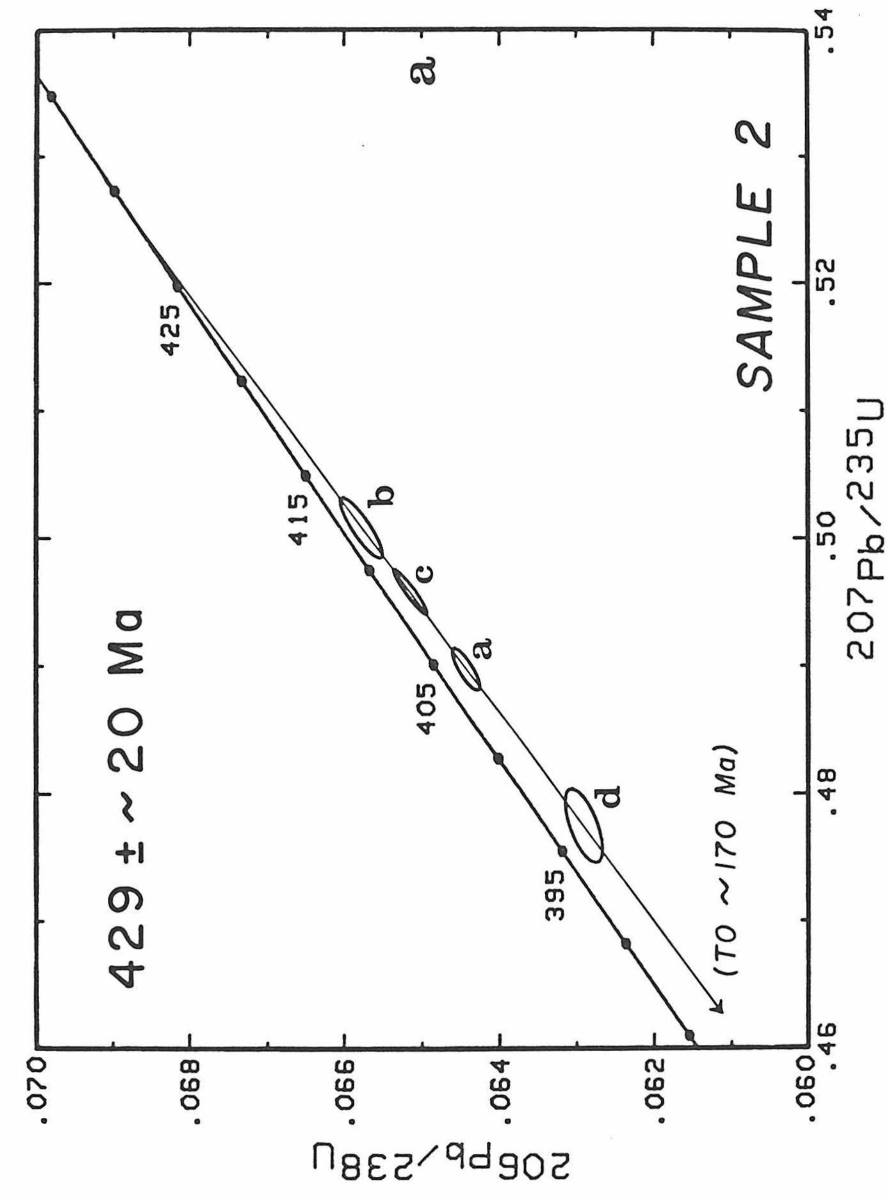
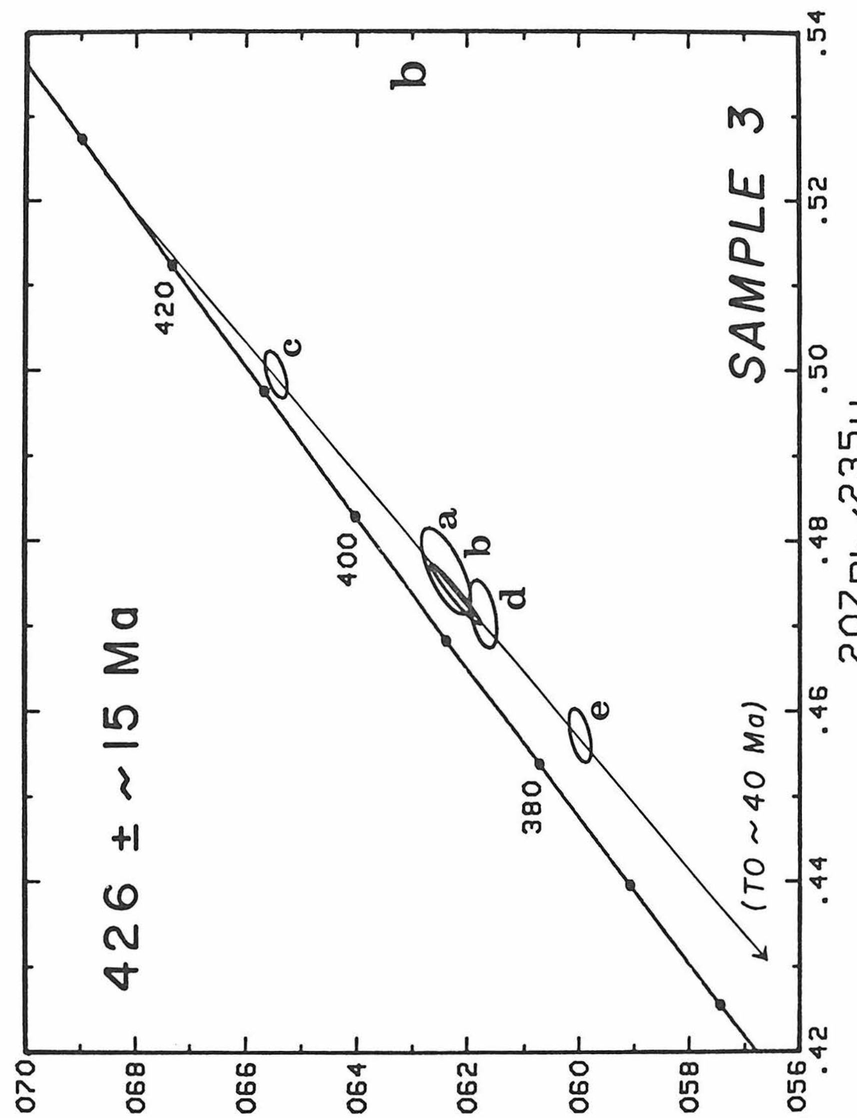
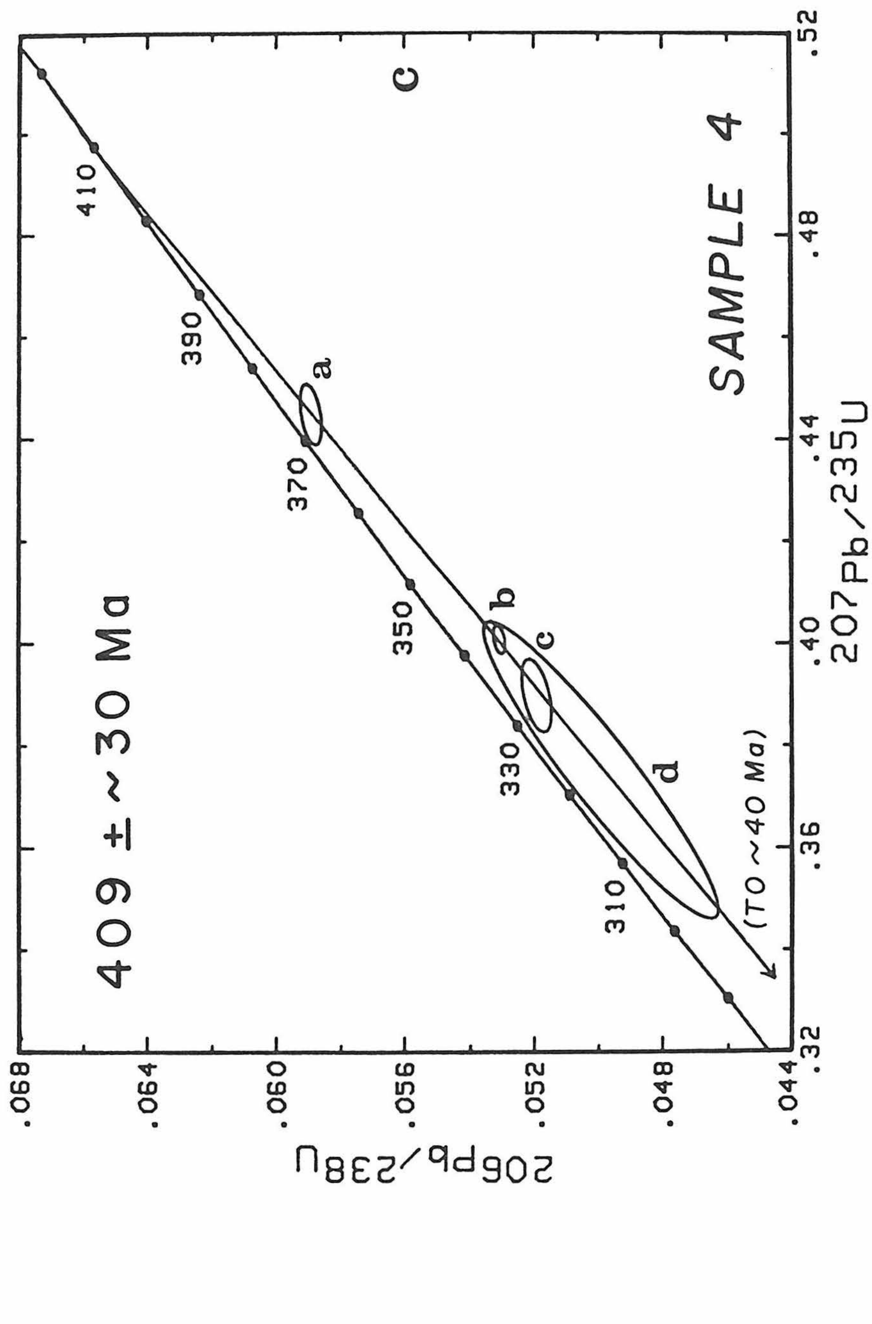
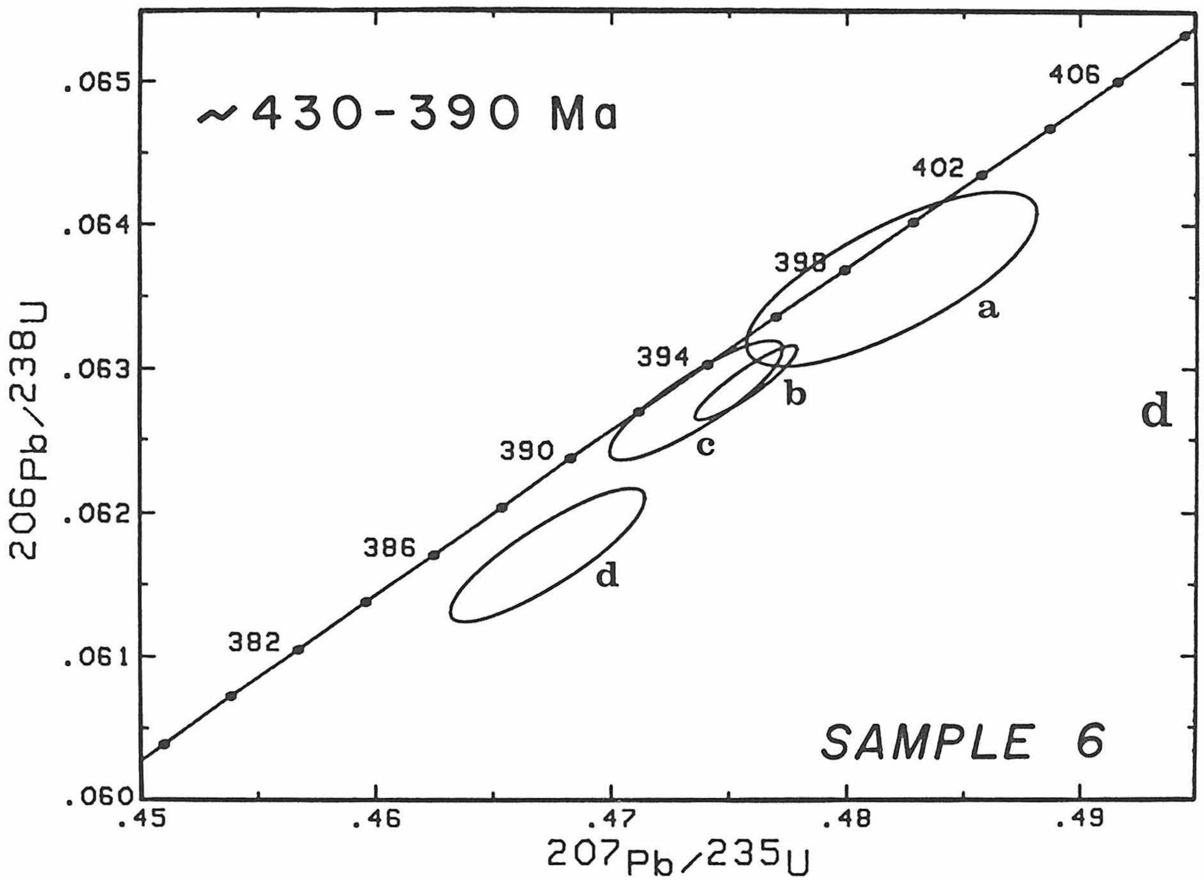


Fig. 3-8. $^{206}Pb^*/^{238}U$ versus $^{207}Pb^*/^{238}U$ concordia diagrams for age samples 2(a), 3(b), 4(c), 6(d), and 7(e). The error ellipses are at the 95% level and incorporate all known significant errors. The error ellipses have been calculated and the concordia diagrams have been drawn with programs written by Ken Ludwig (1983 and written commun., 1984). Apparent ages cited in 8a, b, c, and e are interpreted from the upper intercepts of concordia and the regression line, which has been calculated using the equations of York (1969) and a program written by Ken Ludwig (written commun., 1984). The associated errors have been estimated graphically by passing a family of lines through the error ellipses: these approximate errors are generally similar to statistically determined errors at the 95% level. The data from sample 6 (Fig. 3-8d) are insufficient to determine a reliable apparent age, but are consistent with the interpretation that these trondhjemitic rocks are generally coeval with samples 4, 5, and 7. I accordingly suggest that their age is in the 430-390 Ma range.

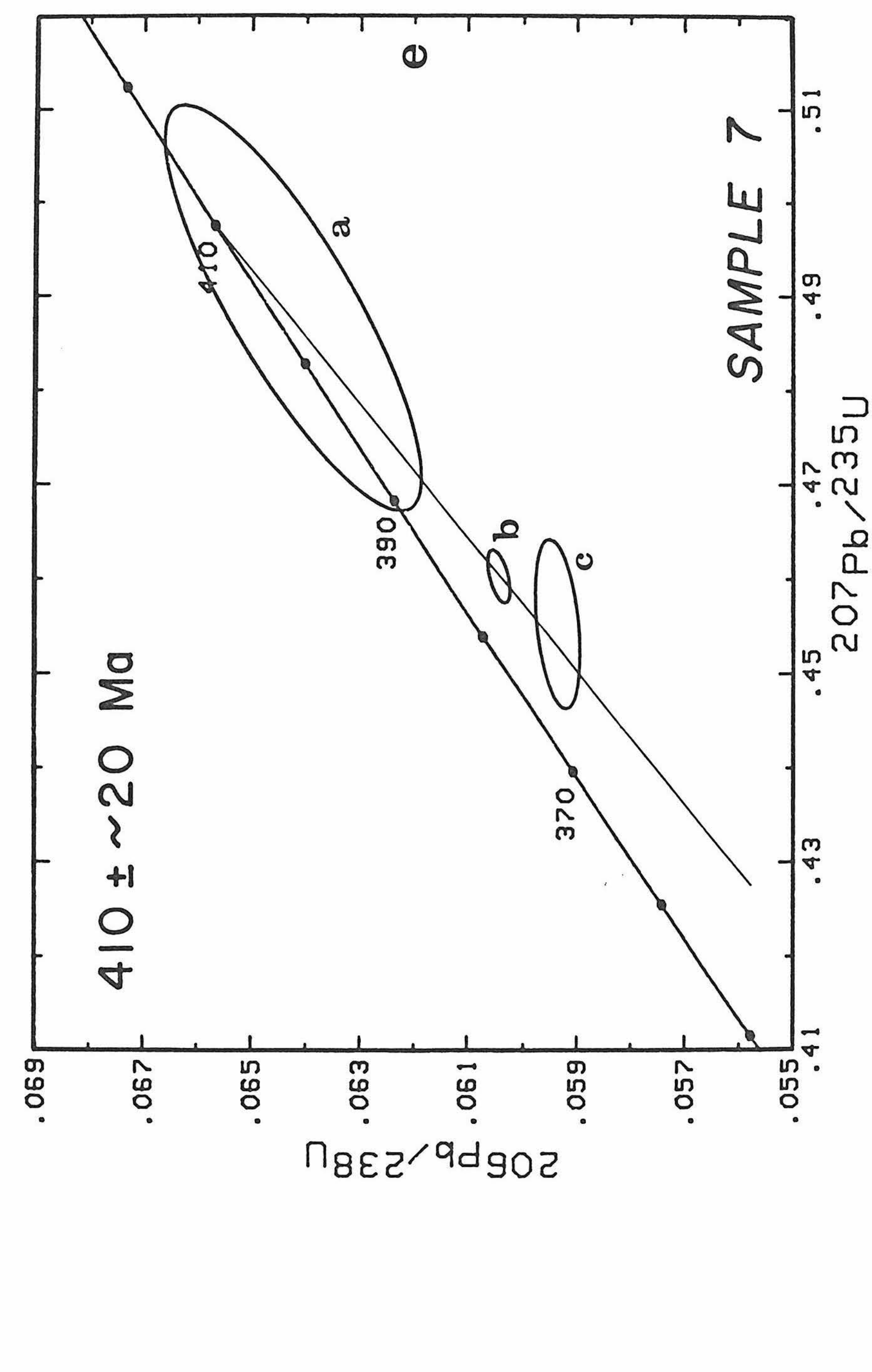

429 ± ~20 Ma
SAMPLE 2
a
$^{207}Pb/^{235}U$
$^{206}Pb/^{238}U$
(TO ~170 Ma)
395
405
415
425
a
b
c
d
.46
.48
.50
.52
.54
.060
.062
.064
.066
.068
.070


426 ± ~15 Ma
SAMPLE 3
b
420
400
380
a
b
c
d
e
(TO ~40 Ma)
.070
.068
.066
.064
.062
.060
.058
.056
.42
.44
.46
.48
.50
.52
.54
206Pb/238U
207Pb/235U


409 ± ~30 Ma
SAMPLE 4
C
a
b
c
d
(TO ~40 Ma)
410
390
370
350
330
310
.068
.064
.060
.056
.052
.048
.044
.32
.36
.40
.44
.48
.52
207Pb/235U
206Pb/238U



~430-390 Ma
SAMPLE 6
d
a
b
c
d
406
402
398
394
390
386
382
.065
.064
.063
.062
.061
.060
.45
.46
.47
.48
.49
$^{206}Pb/^{238}U$
$^{207}Pb/^{235}U$


410 ± ~20 Ma
SAMPLE 7
e
a
b
c
410
390
370
206Pb/238U
207Pb/235U
.069
.067
.065
.063
.061
.059
.057
.055
.41
.43
.45
.47
.49
.51

from a moderately schistose diorite within this zone and yields a U-Pb apparent age of 426 ± ~15 Ma (Fig. 3-8b). These rocks are overlain by Lower Devonian strata of the Karheen Formation (described below) and are intruded by dikes of nondeformed Late Silurian trondhjemite.

The age data and intrusive relations on Annette, Gravina, and Duke Islands indicate that the dioritic rocks were emplaced between Cambrian and Late Silurian time, and that in some (most?) areas the diorite is Early Silurian in age. I assign an Ordovician-Early Silurian age to the dioritic rocks on these islands based on the Early Silurian apparent ages and the occurrence of similar dioritic rocks on southern Prince of Wales Island with Ordovician U-Pb apparent ages (Gehrels and Saleeby, in review b).

Ordovician-Early Silurian hornblende gabbro

Hornblende gabbro occurs locally on Duke Island (Fig. 3-4) and consists predominantly of green hornblende, brown biotite, and highly altered plagioclase. On northeastern Duke Island the gabbroic rocks have a moderate foliation which is cut by a trondhjemitic body that resembles the Late Silurian trondhjemite on Annette Island (described below). Textural and mineralogical similarities between the hornblende gabbro in this region and dioritic rocks that occur nearby, combined with the cross-cutting Late Silurian(?) trondhjemite, indicate that this hornblende gabbro is probably Ordovician-Early Silurian in age. In contrast, Irvine (1974) reports that hornblende gabbro on central and southern Duke Island was derived from the adjacent pyroxene gabbro (now known to be Triassic in age) during metamorphism associated with emplacement of the Cretaceous(?) ultramafic rocks. The hornblende gabbro unit shown on Figure 3-4 may therefore include rocks of both

Ordovician-Early Silurian and Mesozoic age.

Trondhjemite to granite (Late Silurian)

Much of Annette Island is underlain by a composite stock of leucotrondhjemite and subordinate trondhjemite, quartz diorite, diorite, and leucocratic granite, quartz monzonite, and granodiorite (Berg, 1972a). These rocks consist of varying proportions of plagioclase (albite to oligoclase), large bluish quartz grains, microperthite, green hornblende, brown biotite, and chlorite. M.A. Lanphere has determined a K-Ar (hornblende) apparent age of 424 ± 12 Ma (one-sigma uncertainty) on a leucocratic quartz diorite member of the pluton (cited in Berg, 1972a) and I have determined a U-Pb apparent age of 409 ± ~30 Ma (Fig. 3-8c) on the same leucocratic quartz diorite (sample 4). A trondhjemite dike (sample 5) near the southwestern margin of the stock (Fig. 3-3) yields an apparent age of 415 ± 5 Ma. This dike is also significant because it cuts across the foliation in the adjacent diorite. In most areas the trondhjemitic to granitic rocks are moderately brecciated, and along the Metlakatla fault the rocks are moderately to strongly foliated.

Trondhjemitic to quartz dioritic rocks south of Metlakatla are quite similar to rocks in the large stock to the east (Berg, 1972a) and yield U-Pb data consistent with an age in the 430-390 Ma range (sample 6: Fig. 3-8d). Trondhjemitic rocks on Gravina Island are also similar (Berg, 1973) and yield an apparent age of 410 ± ~20 Ma (sample 7: Fig. 3-8e). Dikes of trondhjemite near the western margin of the body on southern Gravina Island clearly cut across the deformational fabrics in adjacent Ordovician-Early Silurian diorite (Fig. 3-7). On Duke Island, rocks in this unit range from granite to trondhjemite and resemble members of the large stock on Annette Island. They consist of medium-to coarse-grained microperthite, plagioclase, bluish quartz, and minor biotite and hornblende. A foliation is present in some rocks, but most are massive and homogeneous. Dikes of trondhjemite on northeastern Duke Island cut across deformational fabrics in the adjacent hornblende gabbro and demonstrate, as on Annette and Gravina islands, that the trondhjemitic to granitic rocks were emplaced after the Ordovician-Early Silurian diorite and subordinate gabbro were deformed.

Klakas orogeny (middle Silurian-earliest Devonian)

Early Silurian and older rocks on Annette, Gravina, and Duke Islands were moderately deformed, metamorphosed to greenschist and locally amphibolite facies, and uplifted prior to emplacement of Late Silurian trondhjemitic rocks. The age of this deformation and metamorphism is tightly constrained by relations on southern Annette Island, where a 415 ± 5 Ma trondhjemite dike (sample 5) cuts the semi-schistose foliation in a metadiorite which is similar to and probably continuous with the 426 ± ~15 Ma diorite at locality 3. Deformation and uplift are known to have continued into latest Silurian-earliest Devonian time, however, as rocks in the large trondhjemitic stock on Annette Island are regionally brecciated and locally foliated, and large trondhjemite clasts occur in superjacent Lower Devonian strata of the Karheen Formation (described below).

Tectonic activity during middle Silurian-earliest Devonian time on southern Prince of Wales Island is recorded by southwest-vergent movement on thrust faults, penetrative deformation of Ordovician-Early Silurian(?) rocks, regional uplift and erosion of Early Silurian and older rocks, and deposition of a Lower Devonian clastic wedge (Ovenshine and others, 1969; Gehrels and Saleeby, in review b). This tectonic event has been referred to as the Klakas orogeny because thrust faults, deformational fabrics, and Lower Devonian conglomeratic strata are well exposed in the Klakas Inlet region of southern Prince of Wales Island (Gehrels and others, 1983; Gehrels and Saleeby, in review a, b).

The Metlakatla fault on western Annette Island (Fig. 3-3) may also have moved as a southwest-vergent thrust fault during the Klakas orogeny. Berg (1972b) described this fault as an east-dipping thrust that moved initially during late Mesozoic or Tertiary time and was later reactivated by movement on high-angle faults. Structural analyses along the fault zone during this study indicate that older, east-dipping slickensides have predominantly east-plunging striae, and younger, high-angle slickensides have both shallow- and steeply plunging striae. North of the area shown on Figure 3-3, Ordovician-Early Silurian rocks are penetratively brecciated, whereas Late Silurian trondhjemite and Upper Triassic strata are only moderately deformed. These relations are consistent with the interpretation that the Metlakatla fault moved initially during the Klakas orogeny as an east-dipping thrust fault. Dip-slip and (or) strike-slip movement along the trace shown on Figure 3-3 and on a fault south of Annette Island (Fig. 3-4) apparently occurred during or after Late Triassic time.

Karheen Formation and associated volcanic rocks (Lower Devonian)

Lower Devonian sedimentary and volcanic rocks underlie Hotspur Island (Fig. 3-4) and occur on northwestern Gravina and southern Annette Islands. The basal contact of these rocks is exposed on southern Annette Island, where a meter-thick layer of pebbly sandstone and siltstone unconformably overlies highly deformed Early Silurian diorite. This basal clastic layer is overlain by ~50 m of interbedded dark-gray siltstone, mudstone, and limestone, which are juxtaposed against a sequence of interbedded dark-gray mudstone and siltstone, gray limestone, conglomerate, and approximately 20 m of andesitic breccia at the top of the section. Clasts in the conglomerate consist primarily of schistose diorite derived from the subjacent Early Silurian metadiorite and of nondeformed Late Silurian trondhjemite. Lower Devonian strata on southern Annette Island are generally deformed around open, shallow-plunging, outcrop-scale folds, and have been recrystallized to phyllite during sub-greenschist- to greenschist-facies metamorphism. Stratigraphic and structural relations indicate that these strata occur in the trough of a regional east-southeast-plunging syncline of probable Late Cretaceous age (Fig. 3-3).

Hotspur Island is underlain by a thick section of Lower Devonian sedimentary and volcanic rocks which, in contrast to the strata on southern Annette Island, are only slightly deformed and metamorphosed. As shown on Figure 3-9, the lowest rocks exposed consist of interbedded gray siltstone and mudstone, brownish-gray sandstone with cross beds and small channels, brown to gray limestone with abundant shallow-marine megafossils (Berg, 1972a), and gray tuffaceous mudstone. The clastic strata grade upsection into a several-hundred-meter-thick olistostromal layer that consists of meter-scale olistoliths enclosed in a matrix of chaotic sandstone, siltstone, and mudstone (Fig. 3-10). Olistoliths consist primarily of intraformational limestone, siltstone, and sandstone: exotic blocks have not been recognized. Many of the olistoliths show evidence of soft-sediment deformation, which suggests that they were incorporated into the olistostrome prior to lithification. The olistostromal unit grades upsection into interbedded

Fig. 3-9. Schematic stratigraphic section of the Lower Devonian rocks on Hotspur Island (Fig. 3-4).

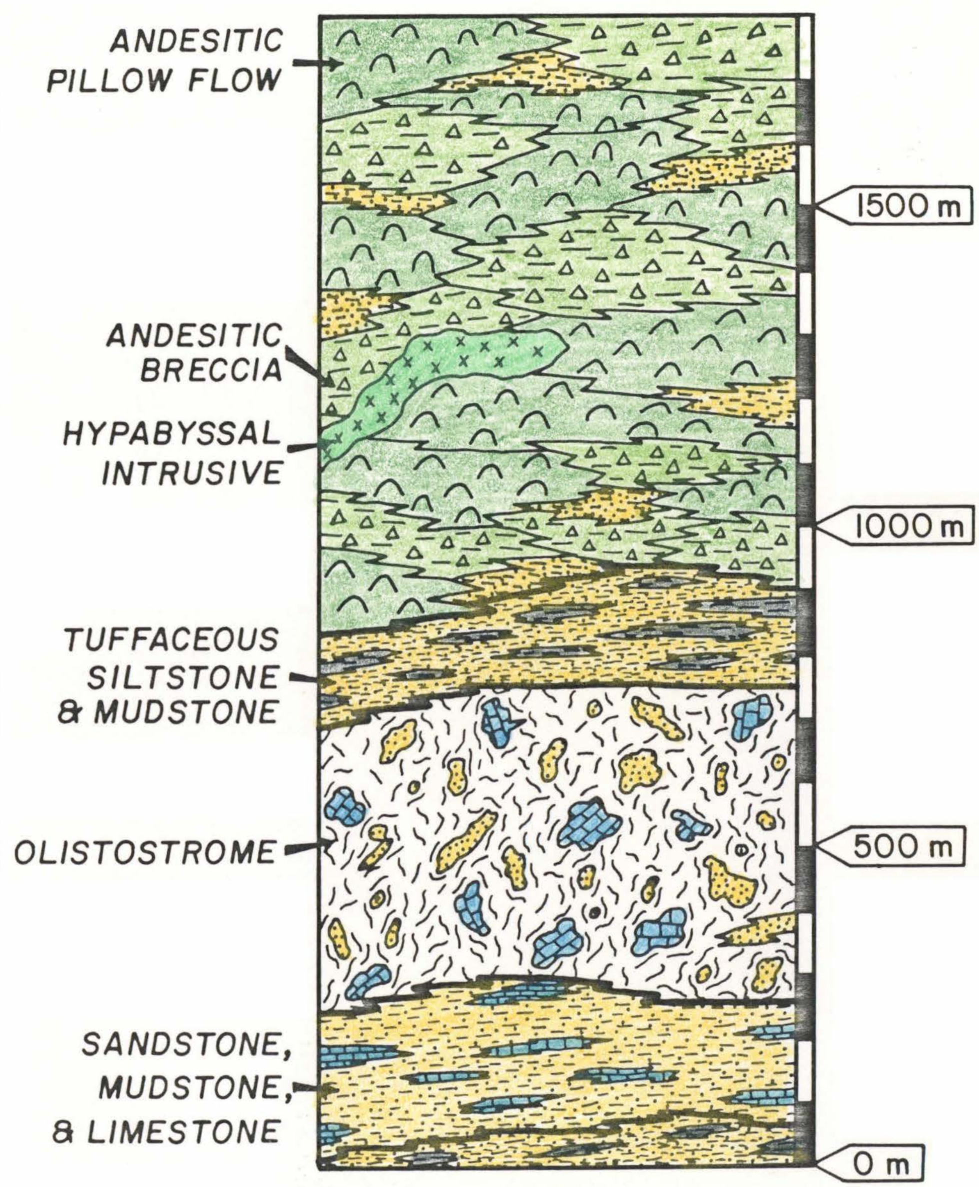


Fig. 3-10. Photograph of limestone and sandstone olistoliths in the olistostromal layer on northwestern Hotspur Island (see Figures 3-4 and 3-9). Olistoliths range up to several meters in diameter and show evidence of soft-sediment deformation along the eastern shore of the island.

mudstone, tuff, and brownish-weathering siltstone, which in turn grades into a thick section of andesitic pillow flows, tuff breccia, tuff, massive flows, hypabyssal intrusive rocks, and subordinate clastic strata (Fig. 3-9).

Interbedded gray limestone and dark-gray siltstone occur on the northwestern tip of Gravina Island (Fig. 3-2). These strata are moderately deformed and metamorphosed, but primary sedimentary structures and megafossils are well preserved. Triassic sedimentary and volcanic rocks apparently surround the limestone and siltstone (Berg, 1973), which suggests that these strata may belong to a large slide(?) block in the Triassic section.

Fossils recovered from strata in this unit have been assigned to the Early or Middle Devonian, although the ages of some fossils range into the Silurian (Berg, 1972a, 1973). I collected conodont samples from limestones throughout the area, but the only age-diagnostic fauna were recovered from a limestone olistolith on eastern Hotspur Island. As described by Savage and Gehrels (in press), these fauna are representative of the middle Early Devonian and demonstrate that the olistostrome and overlying strata were deposited during or after middle Early Devonian time. Because the olistoliths are similar to the underlying and overlying strata, and at least in some cases were incorporated into the olistostrome prior to lithification, I suggest that most of the clastic strata on Hotspur Island are of middle Early Devonian age. The overlying volcanic rocks are gradational with these strata and are accordingly also interpreted to be Lower Devonian.

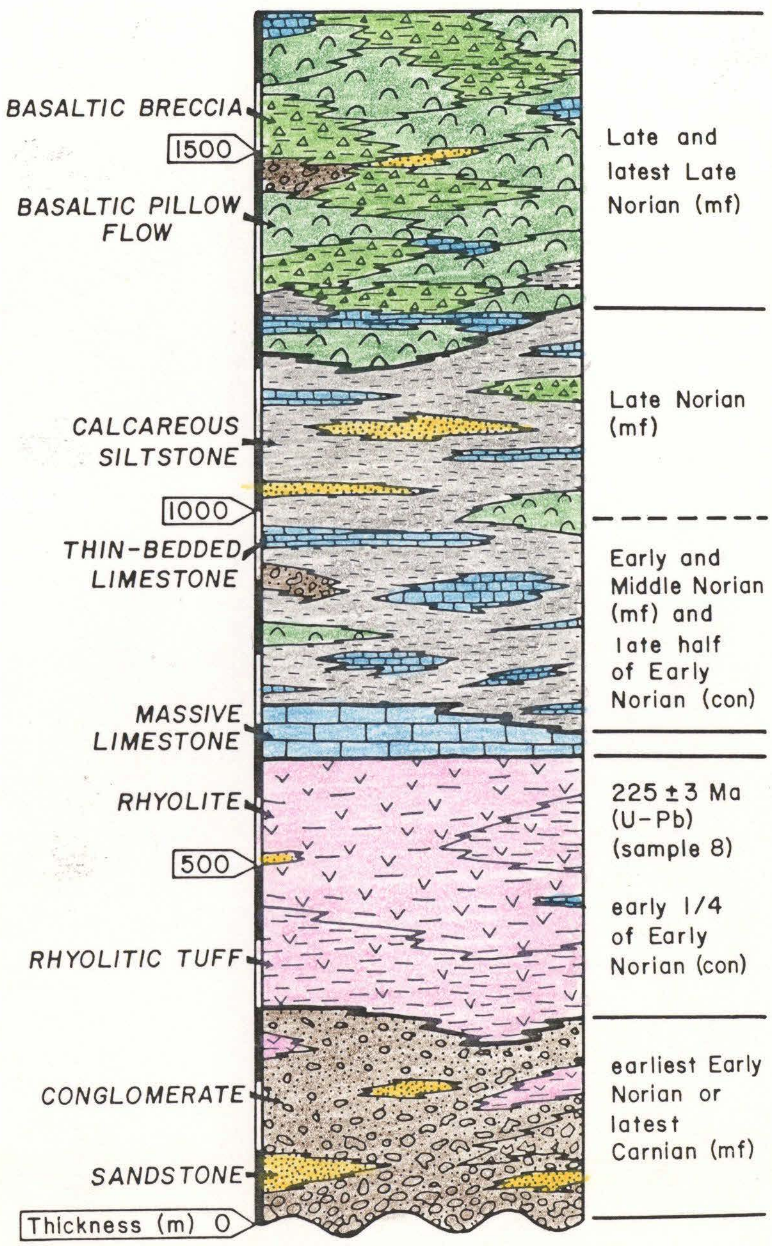
Lower Devonian strata of the Karheen Formation on Prince of Wales Island (Eberlein and Churkin, 1970; Eberlein and others, 1983; Gehrels and Saleeby, in review b) are similar to the Lower Devonian strata on Annette and Hotspur Islands. At the type locality on northwestern Prince of Wales Island, the Karheen Formation consists of graywacke and subordinate siltstone, shale, limestone, and conglomerate (Eberlein and Churkin, 1970) which yield middle Early Devonian conodonts (Savage, 1977). Toward the south these strata grade into a thick clastic wedge of conglomeratic strata which Ovenshine and others (1969) suggest was shed from uplifted regions to the south. On southern Prince of Wales Island the formation consists of a highly variable section of conglomerate, sandstone, siltstone, mudstone, graptolitic shale, and middle Lower Devonian limestone (Gehrels and Saleeby, in review b; Savage and Gehrels, 1984; Herreid and others, 1978). These strata overlie locally deformed Ordovician-Silurian volcanic, sedimentary, and plutonic rocks, and are overlain by and locally interbedded with andesitic flows and breccia (Gehrels and Saleeby, in review b; Herreid and others, 1978). Structural and stratigraphic relations indicate that the strata on southern Prince of Wales Island were deposited within or adjacent to regions uplifted during the Klakas orogeny and that the strata described by Ovenshine and others (1969) were deposited along the northern flank of the uplifted orogen (Gehrels and others, 1983; Gehrels and Saleeby, in review a).

I assign the Lower Devonian strata on Annette and Hotspur Islands to the Karheen Formation based on similarities in rock type, stratigraphic position, and age with strata near the type section of the Karheen Formation on Prince of Wales Island. The strata on these islands are also similar in that their basal unconformity and superjacent conglomeratic strata record uplift and erosion of Silurian and older rocks during the Klakas orogeny. The amount of uplift in the Annette-Gravina-Duke region is interpreted to be approximately 5-10 km based on the occurrence of greenschist- and locally amphibolite-facies rocks beneath the Devonian strata on southern Annette Island. The occurrence of Late Silurian(?) trondhjemite boulders in the lower part of the section on Annette Island demonstrates that the uplift continued into latest Silurian-earliest Devonian time.

Hyd Group (Upper Triassic)

Upper Triassic strata on Annette and Gravina Islands consist of volcanic and clastic sedimentary rocks and limestone that overlie Devonian and older rocks on a major unconformity (Berg, 1972a and 1973). As shown on Figure 3-11, the stratigraphic section generally consists, from bottom to top, of: coarse polymictic conglomerate and sedimentary breccia; rhyolite flows, breccia, and tuff; massive bluish-gray limestone and minor dolomite; interbedded gray to brown limestone, siltstone, and sandstone; and basaltic pillow flows, breccia, and tuff. On Gravina Island the rhyolitic rocks are referred to as the Puppets Formation; most basaltic rocks belong to the Chapin Peak Formation; and the sequence of interbedded siltstone and limestone is assigned to the Nehenta Formation (Berg, 1973; Berg, 1982, pp. 5-7). Similar sedimentary and volcanic rocks on northern Kuiu Island (Fig. 3-1) belong to the Hyd Group, although rhyolitic rocks in the section on Kuiu Island have been excluded from the group and assigned to the Keku Volcanics (Muffler, 1967). Based on similarities in rock types, stratigraphic position, and age, I assign the Upper Triassic strata on Annette and Gravina Islands to the Hyd Group. I note, however, that rhyolite similar to the Keku Volcanics is included in the group on

Fig. 3-11. Schematic stratigraphic section of Upper Triassic strata on Annette and Gravina Islands. Megafossil ages (mf) are from Berg (1973) and Berg and Cruz (1982), and conodont ages (con) are from Savage and Gehrels (in press).



Annette and Gravina Islands.

The age of these strata is tightly constrained by megafossils (Berg, 1972a and 1973; Berg and Cruz, 1982), conodonts (Savage and Gehrels, in press), and by my U-Pb apparent age of 225 ± 3 Ma (sample 8) on rhyolite of the Puppets Formation (Fig. 3-11). The apparent age of the Puppets rhyolite is additionally significant because it places a minimum age of 225 ± 3 Ma on the Carnian-Norian boundary (the boundary is assigned an age of 225 ± 8 Ma on the DNAG Time Scale: Palmer, 1983).

A variety of stratigraphic relations suggest that these Upper Triassic strata were deposited during a phase of uplift, erosion, and faulting in the southern Alexander terrane. In most areas the base of the section is a massive to thick-bedded sedimentary breccia or conglomerate that locally reaches several hundred meters in thickness. Clasts are generally angular, poorly sorted, and 50 cm to 1 meter in diameter (Fig. 3-12), although clasts greater than two meters across have been recognized locally. Ordovician-Early Silurian diorite and trondhjemite constitute most of the clasts, but Devonian and Triassic stratified rocks also occur. This basal unit is interpreted to have been deposited as a talus breccia in a regime of active faulting based on the: 1) great thickness of breccia and conglomerate in some areas, 2) abrupt lateral variations in thickness of the section, 3) size, angularity, and lack of sorting of clasts, and 4) massive to thick-bedded nature of the strata. Similar breccia of probable Triassic age occurs in the Clover Bay area of east-central Prince of Wales Island (Fig. 3-6: Buddington and Chapin, 1929, pp. 313; Eberlein and others, 1983).

Uplift and erosion during and prior to deposition of the strata is

Fig. 3-12. Large angular clasts of Ordovician-Silurian plutonic rocks in Upper Triassic sedimentary breccia on southwestern Gravina Island. The largest boulder in the photograph is approximately 80 cm across.

recorded by: 1) the regional unconformity at the base of the Upper Triassic section, 2) a lack of Upper Permian to Middle Triassic rocks anywhere in the Alexander terrane, and 3) occurrence of rocks ranging in age from Ordovician-Early Silurian (Annette and Gravina Islands) through Early Permian (Kuiu and Admiralty Islands) beneath the unconformity. This uplift and erosion does not appear to be associated with regional deformation or metamorphism, however, as Devonian strata on Annette Island are not more highly deformed than Triassic strata, and the color alteration indices of Devonian and Triassic conodonts are similar (Savage and Gehrels, in press). I alternatively suggest that this tectonism is related to a latest Paleozoic(?)-Triassic rifting event based on the relations described above, the bimodal (basalt-rhyolite) composition of the volcanic rocks, and occurrence of the Upper Triassic strata in a fairly narrow belt along the eastern margin of the Alexander terrane (Gehrels and Berg, 1984). Steeply dipping mafic dikes of probable Mesozoic age that occur throughout the southern Alexander terrane (Gehrels and Saleeby, in review b, and unpub. mapping) may also have been emplaced during this interpreted rifting event.

Late Triassic pyroxene gabbro

Much of central and eastern Duke Island is underlain by a large body of pyroxene gabbro which consists of varying proportions of olivine, hypersthene, augite, plagioclase, hornblende, and minor opaque minerals (Irvine, 1974). The gabbro has traditionally been regarded as Jurassic in age based on a K-Ar (biotite) date of 177 ± 5.3 Ma (one-sigma uncertainty: Irvine, 1974; Smith and Diggles, 1981). Murray (1972) and James (1971) argue that this pyroxene gabbro is genetically related to Cretaceous(?) ultramafic rocks on Duke Island based on their close proximity and approximately similar age. Taylor (1967) and Irvine (1967) report, however, that the petrologic and geochemical characteristics of the two suites are quite different and conclude that they are unrelated.

Age sample 9 is a pyroxene gabbro which consists of medium-grained hypersthene, labradorite, and subordinate reddish biotite, light-brown augite, and minor opaque minerals. This sample yields a U-Pb apparent age of 226 ± 3 Ma, which demonstrates that the pyroxene gabbro is considerably older than the interpreted age of the zoned ultramafic bodies. Assuming that the ultramafic rocks are indeed Cretaceous in age, this Late Triassic apparent age for the gabbro demonstrates that the two suites are not genetically related.

Jurassic-Lower Cretaceous sedimentary and volcanic rocks

Jurassic-Lower Cretaceous strata overlie Triassic strata on northeastern Gravina and Annette Islands disconformably (locally unconformably) and generally consist of sub-greenschist- to greenschist-facies andesitic to basaltic tuff and breccia, slate, conglomerate, sandstone, and siltstone (Berg, 1972a and 1973). These strata belong to the Gravina-Nutzotin belt, which overlies the eastern margin of the Alexander terrane in much of southeastern Alaska (Berg and others, 1972).

Cretaceous(?) ultramafic rocks

Ultramafic rocks on Annette and Duke Islands consist of pyroxenite, peridotite, dunite, and hornblendite that have been studied and described by Taylor (1967), Berg (1972a), and Irvine (1974). K-Ar apparent dates of the ultramafic rocks range from 100 to 135 Ma (Lanphere and Eberlein, 1966; Smith and Diggles, 1981) but these dates may have been at least in part reset during Late Cretaceous-early Tertiary metamorphism.

DISCUSSION OF THE U-Pb GEOCHRONOLOGIC DATA

Nearly all of the zircon fractions from Paleozoic samples yield discordant U-Pb and Pb-Pb dates, where discordance is recognized on Pb*/U concordia diagrams by a lack of overlap of individual zircon fractions (with 95%-level uncertainties) and concordia (Figures 3-8a-e). As shown on Figure 3-8, samples 2, 3, 4, and 7 are discordant, with upper intercepts in the 450-400 Ma range and lower intercepts of less than 200 Ma (except for sample 7, the data from which are not sufficient to define a meaningful lower intercept). Because these samples belong to units that are either intruded by Late Silurian trondhjemite or overlain by Lower Devonian strata, their discordance is attributed to isotopic disturbance of Paleozoic zircon populations rather than incorporation of older uranium-bearing phases in Mesozoic or Cenozoic zircons. This interpretation is consistent with the: 1) clustering of most fractions near the upper intercept with concordia, 2) correlation in most samples between the degree of discordance and the size range and uranium concentration of individual zircon fractions, and 3) lack of optically distinguishable cores or compositional zoning in the zircons. A likely explanation of this discordance is that lead has been lost relative to uranium from the zircon populations during Mesozoic-Cenozoic time. As described below, however, isotopic disturbance of the zircon populations may have been considerably more complex than a single-stage lead-loss event.

K-Ar dates from Paleozoic rocks on these islands are in almost all cases significantly younger than their geologically constrained ages.

As reported by Smith and Diggles (1981), dioritic rocks yield dates of 182 ± 5.5 Ma (biotite) and 205 ± 16 Ma (hornblende) on Annette Island, and 315 ± 9.4 Ma (biotite), 117 ± 3.5 Ma (muscovite), and 259 ± 7.8 Ma (hornblende) on Duke Island. Also reported by these workers are dates of 306 ± 9.2 Ma (biotite) from a trondhjemitic body and 79.3 ± 1.6 Ma (hornblende) on quartz-biotite schist (derived from metavolcanic rocks of the Descon Formation) south of Metlakatla. (Uncertainties in the K-Ar dates cited above are reported at the one-sigma level.)

Late Cretaceous-early Tertiary metamorphism and hydrothermal alteration have apparently played a major role in disturbance of the K-Ar and U-Pb isotopic systems. Metamorphism during this time is indicated by greenschist-facies mineral assemblages in Cretaceous and older rocks along the eastern side of the islands, and sub-greenschist-facies assemblages to the west. Pervasive interaction with hot hydrothermal fluids during this time is recorded by the oxygen isotopic composition of quartz and feldspar pairs in Ordovician-Silurian and Cretaceous-Tertiary intrusive rocks (Gehrels and Taylor, 1984). The color alteration indices of conodonts in Devonian and Triassic strata on Hotspur and western Annette Islands suggest that temperatures of approximately 300°C were reached during this event (Savage and Gehrels, in press). Resetting of the K-Ar apparent dates to as young as 79 Ma and isotopic disturbance of the U-Pb systems after 200 Ma are likely manifestations of this regional metamorphic-hydrothermal event.

The lack of discordance in U-Pb apparent ages of Triassic rocks (samples 8 and 9; Table 3-1) indicates that the isotopic systems of Paleozoic zircon populations may have been disturbed initially prior to or during Late Triassic time. This early isotopic disturbance could have occurred initially during metamorphism, deformation, and uplift related to the Klakas orogeny, and again during hydrothermal activity and uplift related to the latest Paleozoic(?)-Triassic rifting(?) event. It is also possible that lead has diffused continuously from the zircon populations following the speculated Silurian-Early Devonian disturbance. Although my isotopic data are not sufficient to discern between these various mechanisms and episodes of disturbance, the relations described above suggest that the Paleozoic zircon populations have undergone a complex isotopic evolution. In light of this probable complexity, the uncertainties cited for apparent ages of samples 2, 3, 4, 6, and 7 are estimates and have not been determined statistically.

GEOLOGIC HISTORY

The recorded geologic history on Annette, Gravina, and Duke Islands begins with deposition of volcanic and sedimentary rocks of the Wales metamorphic suite prior to Late Cambrian time. Dioritic and subordinate quartz dioritic rocks were emplaced during Middle-Late Cambrian time, and both suites of rocks were metamorphosed, deformed, and subsequently uplifted(?) during the Middle Cambrian-Early Ordovician Wales "orogeny" (Gehrels and Saleeby, 1984; Gehrels and Saleeby, in review a). Soon after this "orogenic" event, basaltic to rhyolitic flows, breccia, and tuff of the Descon Formation were deposited and large bodies of diorite and subordinate gabbro were emplaced. Similar rocks on southern Prince of Wales Island have calc-alkaline affinities and are interpreted to have formed in an ensimatic volcanic arc (Gehrels and Saleeby, in review b). Preliminary studies of the geochemistry of these rocks on Annette, Gravina, and Duke Islands (Berg, unpub. data) and their modal mineralogy and composition are consistent with the interpretation that they also formed in a volcanic arc environment.

Stratigraphic and structural relations on Annette, Gravina, and Duke Islands record a major middle Silurian-earliest Devonian tectonic event, referred to as the Klakas orogeny, which has been recognized throughout the southern Alexander terrane (Gehrels and others, 1983; Gehrels and Saleeby, in review a). On Annette, Gravina, and Duke Islands this orogeny is manifest by 1) cessation of the Ordovician-Early Silurian arc-type(?) plutonism and volcanism, 2) regional metamorphism and deformation after 426 ± ~15 Ma and prior to 415 ± 5 Ma, 3) probable west-vergent movement along the Metlakatla fault zone on western Annette Island, and 4) uplift (probably 5-10 km) and erosion of Silurian and older rocks prior to middle Early Devonian time. Lower Devonian clastic strata of the Karheen Formation were deposited unconformably over the Late Silurian and older rocks and record a period of relative tectonic stability after the Klakas orogeny.

Relations between the Klakas orogeny and the generation and emplacement of Late Silurian trondhjemitic rocks on Annette, Gravina, and Duke Islands remain uncertain. It is possible that these rocks were generated by partial melting of rocks at greater depth during regional metamorphism, but this has not been demonstrated through petrochemical or isotopic studies. I note, however, that zircon populations from the trondhjemitic rocks generally have lower Pb*/Pb than fractions from the older arc-type(?) diorites (Table 3-1), which supports the interpretation that the trondhjemitic rocks belong to a different petrogenetic suite.

Upper Triassic strata of the Hyd Group were deposited during the waning stages of a major uplift and erosional event along the eastern margin of the Alexander terrane. The thick section of sedimentary breccia and conglomerate at the base of the section indicates that tectonic activity continued into Late Triassic time, and the occurrence of Lower Permian strata unconformably beneath Upper Triassic strata on Kuiu (Muffler, 1967) and Admiralty (Lathram and others, 1965) Islands indicates that uplift began after Early Permian time. Although regionally significant, this uplift and erosional event does not appear to have been associated with deformation or metamorphism of the pre-Late Triassic rocks. Rather, I suggest that this uplift and erosion are manifestations of a latest Paleozoic(?)-Triassic rifting event based on the stratigraphic relations described above, occurrence of the strata in a fairly narrow belt along the eastern margin of the terrane, and the bimodal (basalt-rhyolite) composition of the volcanic rocks. Jurassic-Lower Cretaceous strata of the Gravina-Nutzotin belt were apparently deposited in the marine basin formed during this rifting(?) event.

During Late Cretaceous-early Tertiary time the pre-Tertiary rocks on Annette, Gravina, and Duke Islands were regionally metamorphosed as high as greenschist facies, moderately deformed, and subjected to hydrothermal fluids of at least 300°C (Gehrels and Taylor, 1984; Savage and Gehrels, in press). Reset K-Ar dates and discordant U-Pb dates with lower intercepts of less than 200 Ma indicate that the K-Ar and U-Pb isotopic systems in most rocks were also disturbed during this event. However, the lack of discordance in U-Pb apparent ages of Triassic rocks suggests that the Paleozoic zircon populations have undergone a more complex isotopic evolution. Disturbance may have begun during metamorphism, deformation, and uplift associated with the Klakas orogeny, recurred during latest Paleozoic(?)-Triassic uplift and hydrothermal alteration, and culminated with Late Cretaceous-early Tertiary metamorphism, hydrothermal alteration, and uplift. My isotopic data do not, however, preclude the existence of other mechanisms or episodes of isotopic disturbance.

CONCLUSIONS

My studies of the pre-Jurassic rocks of Annette, Gravina, and Duke Islands yield several regionally significant conclusions regarding the tectonic evolution of the southern Alexander terrane:

1) The Devonian and older rocks on these islands are quite similar to rocks on southern Prince of Wales Island, indicating that the Annette and Craig subterranes have similar early Paleozoic geologic histories and belong to the same tectonic fragment.

2) Intrusive and structural relations on southern Gravina Island and eastern Prince of Wales Island demonstrate that the Clarence Strait fault has approximately 15 km of right-lateral displacement at this latitude.

3) The age of the Klakas orogeny has been tightly constrained: the main phase of deformation and metamorphism occurred between 426 ± ~15 Ma and 415 ± 5 Ma; uplift of the pre-Devonian rocks occurred after 426 ± ~15 Ma, at least in part after 415 ± 5 Ma, and prior to middle Early Devonian time and; by middle Early Devonian time shallow-marine sediments were being deposited in a tectonically stable environment.

4) I suggest that the Upper Triassic strata and their subjacent unconformity record a latest Paleozoic(?)-Triassic rifting event based on the: a) evidence for uplift and erosion but not deformation or metamorphism between Early Permian and Late Triassic time, b) stratigraphic relations which record syn-depositional faulting, c)

occurrence of the Upper Triassic strata in a fairly narrow belt along the eastern margin of the terrane in SE Alaska, and d) bimodal composition of Upper Triassic volcanic rocks.

5) Gabbroic rocks on Duke Island consist primarily of Late Triassic pyroxene gabbro and subordinate Ordovician-Early Silurian hornblende gabbro. This indicates that these rocks are not genetically related to the zoned ultramafic bodies on Duke Island (assuming that the ultramafic rocks are indeed Cretaceous in age).

6) My apparent age of 225 ± 3 Ma for the Puppets rhyolite combined with the age of megafossils and conodonts from adjacent strata confirm a Late Triassic age for the Puppets Formation, as suggested by Berg (1982). These relations also place a minimum age constraint of 225 ± 3 Ma on the Carnian-Norian boundary, which is currently assigned an age of 225 ± 8 Ma on the DNAG Time Scale (Palmer, 1983).

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CHAPTER 4

GEOLOGIC AND TECTONIC EVOLUTION OF THE ALEXANDER TERRANE IN SOUTHERN SOUTHEASTERN ALASKA

ABSTRACT

The Alexander terrane consists of Paleozoic, Triassic, and perhaps Proterozoic rocks that underlie much of southeastern (SE) Alaska and parts of eastern Alaska, western British Columbia, and southwestern Yukon. Regional relations indicate that these rocks have been displaced considerable distances from their sites of origin, and were not accreted to their present position in the North American Cordillera until after Early Cretaceous time.

My geologic and U-Pb geochronologic studies in southern SE Alaska and the work of others to the north indicate that the geologic and tectonic evolution of the terrane can be subdivided into three distinct phases. During the initial phase, from Cambrian (Proterozoic?) through Early Silurian time, the terrane evolved in a volcanic arc environment along a convergent plate margin. This magmatic activity was interrupted during a Middle Cambrian-Early Ordovician tectonic disturbance (Wales "orogeny") and ceased during the middle Silurian-earliest Devonian Klakas orogeny. The second phase is marked by Middle Devonian through Lower Permian strata that accumulated in tectonically stable marine environments. Devonian and Lower Permian volcanic rocks and late Pennsylvanian-Early Permian syenitic to dioritic intrusive bodies occur locally but do not appear to represent major magmatic systems. The third phase in the evolution of the terrane is marked by Triassic volcanic and sedimentary rocks and by a sub-Triassic unconformity, which

are interpreted to have formed in a rift environment.

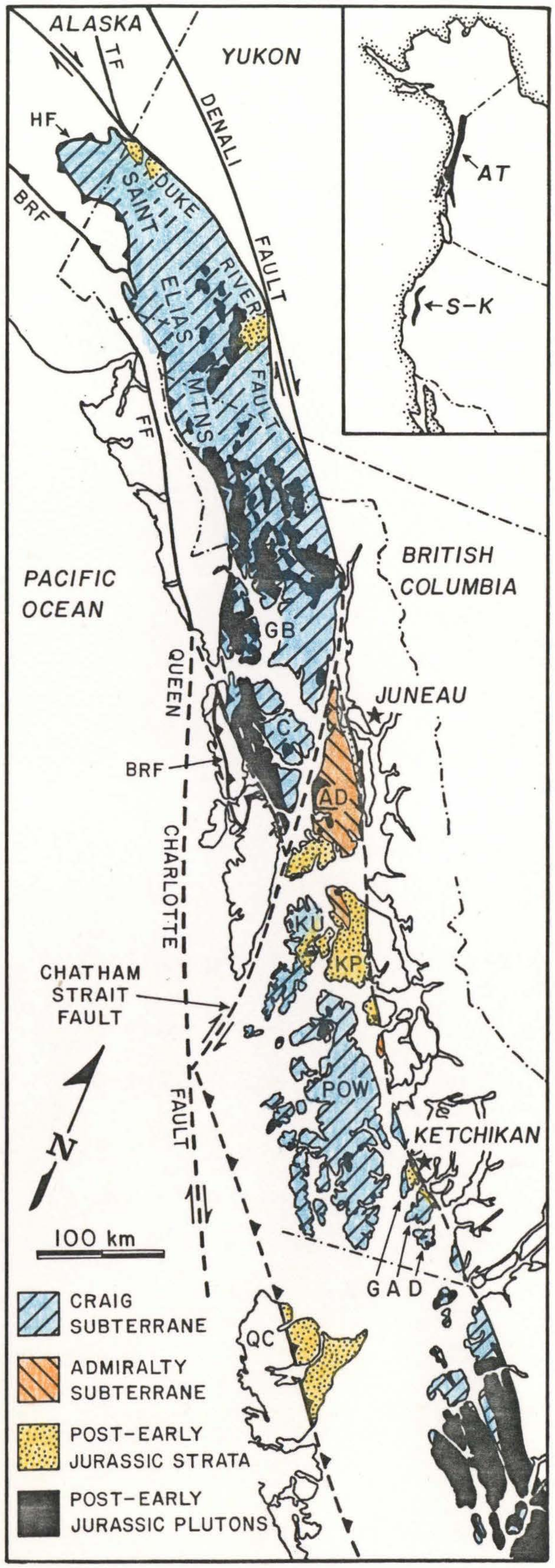
Previous syntheses of the displacement history of the terrane emphasize apparent similarities with rocks in the Sierra-Klamath region and suggest that the Alexander terrane formed in proximity to the California continental margin. My studies indicate, however, that the geologic and tectonic evolution of the Alexander terrane is quite different from that in the Sierra-Klamath region, and I conclude that the two regions were probably not closely associated during Paleozoic time. The available geologic, paleomagnetic, and paleontologic data suggest that a more likely scenario involves: 1) origin and evolution of the Alexander terrane along the paleo-Pacific margin of Australia or adjacent fragments of Gondwana during early Paleozoic time, 2) rifting of the terrane from this margin during Middle Devonian-Early Carboniferous time, 3) migration of the terrane across the paleo-Pacific basin in low southerly paleolatitudes during late Paleozoic time, and 4) location of the terrane south of the equator in the eastern part of the paleo-Pacific basin, perhaps in proximity with rocks now in Peru, during Triassic time. Northward displacement apparently began after Late Triassic time and continued until mid-Cretaceous time, when the Alexander terrane was juxtaposed against terranes previously accreted to the western margin of North America.

INTRODUCTION

The Alexander terrane is a coherent tectonic fragment that underlies much of southeastern (SE) Alaska, the Saint Elias Mountains of eastern Alaska, Yukon, and British Columbia, and part of the coastal region of west-central British Columbia (Fig. 4-1). This terrane is anomalous in the North American Cordillera in that its Phanerozoic

Figure 4-1. Location map of the Alexander terrane and regions referred to in the text. Adapted from Monger and Berg (1984), Tipper and others (1981), MacKevett (1978), Campbell and Dodds (1982a-c, 1983a-c), and Gehrels and Berg (1984).

TF = Totschunda fault; HF = Hubbard fault; BRF = Border Ranges fault; FF = Fairweather fault, DRF = Duke River fault. GB = Glacier Bay; C = Chichagof Island, AD = Admiralty Island; KU = Kuiu Island; KP = Kupreanof Island; POW = Prince of Wales Island; G = Gravina Island; A = Annette Island; D = Duke Island; QC = Queen Charlotte Islands. Inset Map: AT = Alexander terrane; S-K = Sierra-Klamath region.


ALASKA
YUKON
TF
HF
DENALI
DUKE
SAINT
ELIAS
MTNS
RIVER
FAULT
FAULT
BRF
AT
S-K
FF
PACIFIC
OCEAN
BRITISH
COLUMBIA
GB
JUNEAU
QUEEN
CHARLOTTE
FAULT
C
BRF
AD
KU
KP
CHATHAM
STRAIT
FAULT
POW
KETCHIKAN
N
100 km
G A D
QC
CRAIG
SUBTERRANE
ADMIRALTY
SUBTERRANE
POST-EARLY
JURASSIC STRATA
POST-EARLY
JURASSIC PLUTONS

geologic record is suprisingly long and complete: rocks are known from every epoch of the Paleozoic except the Late Permian and Early Cambrian. In addition to these Paleozoic rocks, the Alexander terrane also includes rocks of Early Cambrian and (or) Proterozoic and of Late Triassic age. Jurassic and younger rocks overlie and intrude these Triassic and older rocks, but are not included in the Alexander terrane because they also overlie and intrude adjacent terranes.

The anomalous nature of rocks rocks belonging to the Alexander terrane was initially recognized by Schuchert (1923), who suggested that they formed in a geosynclinal system (the "Alexandrian embayment") which was separate from the main Cordilleran geosyncline. Wilson (1968) recognized that these rocks constitute a distinct tectonic fragment and, based on their occurrence outboard of coeval rocks in the Cordilleran miogeocline, suggested that they were exotic to North America. Berg and others (1972) named this tectonic fragment the Alexander terrane, delineated the boundaries of the terrane, and described its main geologic components.

I have conducted detailed field and U-Pb geochronologic studies in the southern part of the Alexander terrane (Annette, Gravina, Duke, and southern Prince of Wales Islands) and have reviewed the work of others to the north (Figure 1) in an effort to reconstruct the geologic and tectonic evolution of the Alexander terrane. In this report I begin with an outline of the geology of Annette, Gravina, Duke, and southern Prince of Wales Islands, review the regional geologic framework of the Alexander terrane, and discuss the tectonic evolution of the terrane. I then compare the geologic and tectonic history of the terrane to that in other orogenic belts, and evaluate various scenarios for its displacement history. The DNAG Time Scale (Palmer, 1983) is used in this report.

GEOLOGY OF ANNETTE, GRAVINA, DUKE, AND SOUTHERN PRINCE OF WALES ISLANDS

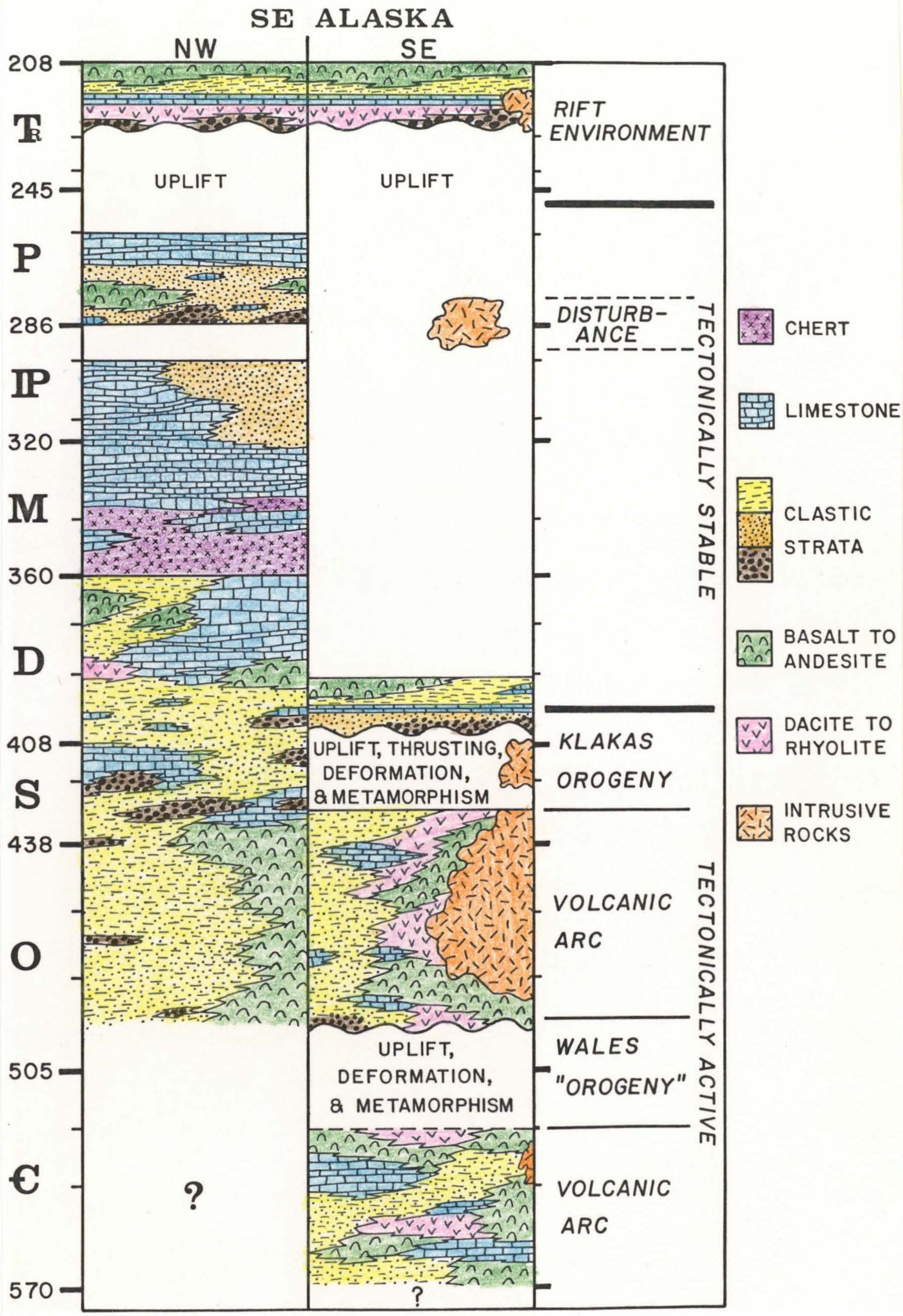
Overview

In this section I describe the geology of Annette, Gravina, Duke, and southern Prince of Wales Islands and discuss the tectonic implications of the geologic relations. The geology of these islands is summarized in Figure 4-2 and is depicted on the geologic map of Gehrels and Berg (1984). More detailed decriptions are provided by Gehrels and others (in review: Annette, Gravina, and Duke Islands) and Gehrels and Saleeby (in review a, b: southern Prince of Wales Island).

Pre-Ordovician metamorphic complex

The oldest rocks recognized in the southern Alexander terrane consist of metavolcanic, metasedimentary, and metaplutonic rocks that occur on south-central Prince of Wales Island (Herreid and others, 1978; Eberlein and others, 1983; Gehrels and Saleeby, in review a, b) and on small islands near the south tip of Gravina Island (Gehrels and others, in review; Gehrels and Berg, 1984). Protoliths of the metavolcanic rocks include basaltic to andesitic pillow flows, breccia, and tuff, and subordinate rhyolitic breccia and tuff. Metasedimentary rocks were derived from banded graywacke and mudstone turbidites and from thick layers of limestone (now marble). In most areas protolith features are highly flattened and are moderately elongated along a penetrative metamorphic foliation. This foliation is in turn commonly folded around shallow-plunging, asymmetric folds that do not have an axial-planar foliation. These rocks were originally referred to as the Wales Group

Figure 4-2. Schematic columns depicting the geologic and tectonic evolution of the Alexander terrane in southern SE Alaska. Adapted from Eberlein and others (1983), Brew and others (1984), Gehrels and Saleeby (in review a, b), and Gehrels and others (in review). NW Column depicts relations in the central Prince of Wales-Kuiu Islands region, and SE column summarizes relations on Annette, Gravina, Duke, and southern Prince of Wales Islands. The geologic time scale is from Palmer (1983).


SE ALASKA
NW
SE
208
Ꞧ
245
P
286
IP
320
M
360
D
408
S
438
O
505
Є
570
UPLIFT
UPLIFT
RIFT ENVIRONMENT
DISTURB- ANCE
TECTONICALLY STABLE
UPLIFT, THRUSTING, DEFORMATION, & METAMORPHISM
KLAKAS OROGENY
VOLCANIC ARC
TECTONICALLY ACTIVE
UPLIFT, DEFORMATION, & METAMORPHISM
WALES "OROGENY"
VOLCANIC ARC
?
?
CHERT
LIMESTONE
CLASTIC STRATA
BASALT TO ANDESITE
DACITE TO RHYOLITE
INTRUSIVE ROCKS

(Buddington and Chapin, 1929) and have since been renamed the "Wales metamorphic suite" (Gehrels and Saleeby, in review b).

Interlayered with the metasedimentary and metavolcanic rocks are dioritic to granodioritic metaplutonic rocks that have been regionally metamorphosed and deformed along with rocks in the Wales suite. These metaplutonic rocks yield U-Pb (zircon) apparent ages of approximately 540 to 520 Ma (Middle and Late Cambrian: Gehrels and others, in review; J. Saleeby, unpub. data), which demonstrates that rocks in the Wales suite were deposited prior to Late Cambrian time. The protolith age of rocks in the suite has not yet been determined. Intrusive relations indicate that the metaplutonic rocks were emplaced during and perhaps slightly after the main phase of deformation and metamorphism in the Wales suite. Cross-cutting bodies of non-deformed Middle Ordovician-Early Silurian dioritic to granitic intrusive rocks (described below) constrain the age of deformation and metamorphism to pre-Middle Ordovician.

Relatively undeformed Lower Ordovician-Lower Silurian volcanic and sedimentary rocks of the Descon Formation (described below) occur near rocks in the Wales metamorphic suite, but depositional relations between the two units have not been demonstrated. Their close association suggests, however, that metamorphism and deformation of rocks in the Wales suite occurred prior to the end of Early Ordovician time. This interpretation is consistent with the intrusive relations described above and with a K-Ar isochron apparent age of 483 Ma (Early Ordovician) determined from metamorphic minerals in the Wales suite (Turner and others, 1977). The available data therefore indicate that metamorphism and deformation of rocks in the Wales metamorphic suite occurred during

Middle Cambrian-Early Ordovician time. I have referred to this event informally as the Wales "orogeny" (Gehrels and Saleeby, 1984), but I note that little is known about the tectonic and regional significance of this deformational and metamorphic event.

The tectonic environment in which these pre-Ordovician rocks formed is difficult to determine because of the penetrative deformation and regional metamorphism. The general basalt-andesite-rhyolite composition of the volcanic rocks and the presence of interlayered graywacke and thick layers of limestone are consistent with the interpretation that these rocks formed in a marine volcanic arc environment, but this interpretation has not been tested through geochemical or petrologic studies. The Wales "orogeny" apparently marks the end of this early volcanic-plutonic phase in the evolution of the southern part of the terrane.

Ordovician-Early Silurian rocks

Basaltic to rhyolitic volcanic rocks, volcaniclastic and fine-grained marine strata, and subordinate limestone of the Descon Formation were deposited in much of the southern Alexander terrane during Ordovician-Early Silurian time (Eberlein and others, 1983; Herreid and others, 1978; Gehrels and Saleeby, in review b; Gehrels and others, in review). The volcanic rocks consist of basaltic to andesitic pillow flows, pillow breccia, tuff breccia, and tuff, and subordinate rhyolitic to dacitic tuff, tuff breccia, and extrusive domes. The more silicic rocks increase in abundance southward on Prince of Wales Island and constitute most of the formation on Annette and Duke Islands. Small sulfide-rich deposits are locally present in the silicic volcanic rocks (Gehrels and others, 1983b) and these rocks may also be the host for larger mineral deposits at Niblack Anchorage (southeast-central Prince of Wales Island) (Herreid, 1964). Thin-bedded to massive mudstone and graywacke turbidites and subordinate argillite and conglomerate are regionally interlayered with the volcanic rocks. Fossils in the Descon Formation range from Early Ordovician to middle Early Silurian in age (Eberlein and others, 1983).

Intrusive bodies of predominantly dioritic to granodioritic composition were emplaced during deposition of volcanic rocks belonging to the Descon Formation. Rock types in the intrusive suite consist of hornblende diorite and quartz diorite, biotite-hornblende granodiorite and leucogranodiorite, hornblende-biotite quartz monzonite, subordinate gabbro, trondhjemite, granite, and quartz syenite, and minor hornblendite and pyroxenite. Intrusive relations and my U-Pb geochronologic data indicate that the dioritic, granodioritic, and quartz dioritic rocks are Middle and Late Ordovician in age (475-440 Ma) on Prince of Wales Island and Early Silurian (approximately 430 Ma) on Annette, Gravina, and Duke Islands. The more potassic rocks occur on southern Prince of Wales Island and are latest Ordovician-earliest Silurian in age (440-435 Ma). Major, minor, and trace element geochemical data (provided by Fred Barker of the U.S. Geological Survey) from intrusive and volcanic rocks on southern Prince of Wales Island indicate that they have have calc-alkaline affinities (Gehrels and Saleeby, in review b).

Volcanic rocks of the Descon Formation were apparently deposited in a marine environment near basaltic to rhyolitic volcanic centers; volcaniclastic graywacke and mudstone were probably deposited in small basins adjacent to the volcanic centers; and limestone layers may have accumulated on volcanic highs between eruptive periods. On southern Prince of Wales Island the volcanic rocks locally grade downsection into swarms of diorite, quartz diorite, and granodiorite dikes and hypabyssal bodies, which in turn grade into the large diorite and granodiorite plutons that underlie much of southern Prince of Wales Island (Gehrels and Saleeby, in review a). These gradational relations combined with similarities in composition and age indicate that the dioritic to granodioritic rocks are genetically related to volcanic rocks in the Descon Formation. Comparison of this Ordovician-Early Silurian sedimentary-volcanic-plutonic complex with modern-day tectonic environments suggests that these rocks formed in a marine volcanic arc along a convergent plate margin.

Depositional relations between Ordovician-Lower Silurian strata and the deformed and metamorphosed pre-Ordovician rocks have not been demonstrated, but Middle Ordovician-Early Silurian diorite and quartz diorite locally intrude the older rocks (Gehrels and others, in review). This critical intrusive relationship demonstrates that the pre-Ordovician metamorphic complex forms the basement to Ordovician and younger rocks in the terrane, and that the Wales "orogeny" occurred prior to formation of the Ordovician-Early Silurian volcanic-plutonic complex. The Wales "orogeny" was therefore not a result of deformation and metamorphism at deep structural levels within the Ordovician-Early Silurian magmatic system. Conglomeratic strata that occur at the base(?) of the Descon Formation on southern Prince of Wales Island (conglomeratic graywacke of Herreid and others, 1978) may have been shed from regions uplifted during this Middle Cambrian-Early Ordovician tectonic event.

Silurian rocks and the Klakas orogeny

A variety of stratigraphic, intrusive, and structural relations in the southern Alexander terrane record a major middle Silurian-earliest Devonian orogenic event. Buddington and Chapin (1929, pp. 281-289) originally recognized this disturbance as a regional unconformity that separates Devonian strata from Silurian and older rocks. Brew and others (1966) recognized that Silurian-Devonian conglomeratic strata on Prince of Wales, Kuiu, and Chichagof Islands were related to this disturbance and suggested that they were shed from source areas uplifted at the time of the Cariboo orogeny, which affected rocks in British Columbia (White, 1959).

More detailed studies of Lower Devonian conglomeratic strata on central Prince of Wales Island led Ovenshine and others (1969) to suggest that they were a "basinward manifestation of Late Silurian to pre-Middle Devonian diastrophism in southern southeastern Alaska." Loney and others (1975, pp. 92) report that similar conglomeratic strata on northeastern Chichagof Island were shed from a source area to the southwest, and note that uplift of the source area and perhaps folding of the pre-Late Silurian rocks began earlier in Silurian time. Monger and others (1972) suggest that the stratigraphic relations mentioned above, metamorphism of pre-Silurian rocks, and Ordovician-Silurian plutonism were evidence of an early to mid-Paleozoic "orogeny." My studies of the Devonian and older rocks on Annette, Gravina, Duke, and southern Prince of Wales Islands indicate that the unconformities and conglomeratic strata recognized by previous workers are stratigraphic manifestations of a major middle Silurian-earliest Devonian orogenic event in the southern part of the terrane, which I refer to as the

Klakas orogeny (Gehrels and others, 1983a). This event is referred to as the Klakas orogeny because the thrust faults, deformational fabrics, and Lower Devonian talus breccia and conglomerate are well exposed in the Klakas Inlet region of southern Prince of Wales Island (Gehrels and others, 1983a).

On southern Prince of Wales Island this orogenic event is manifest as: 1) cessation of Ordovician-Early Silurian arc-type volcanism and plutonism, 2) imbrication of Silurian and older rocks on southwest-vergent thrust faults, 3) emplacement of Late Silurian leucodiorite bodies after movement on some thrust faults but during or prior to movement on others, 4) penetrative brecciation of Ordovician volcanic and intrusive rocks in some areas, 5) at least several kilometers of uplift of Ordovician-earliest Silurian rocks, 6) deposition and deformation of a Lower Devonian talus breccia in a regime of active faulting, and 7) deposition of Lower Devonian conglomeratic strata (described below) in topographically rugged, subaerial environments (Gehrels and Saleeby, in review b).

The minimum age of movement on the thrust faults is constrained by a 418 $\pm$ 5 Ma leucodiorite pluton which intrudes two faults, and by middle Lower Devonian strata that unconformably overlie a third fault and rocks containing the deformational fabrics mentioned above. The occurrence along one of the faults of deformed Lower Devonian talus breccia overlain by undeformed middle Lower Devonian mudstone and shale indicates that movement on some faults continued into earliest Devonian time. The maximum age of movement is constrained by the Middle and Late(?) Ordovician age of rocks cut by the faults. Assuming that uplift of the pre-Devonian rocks occurred during movement on the thrust faults, the presence of intrusive rocks as young as 438 ± 4 Ma overlain by Lower Devonian strata indicates that the faults moved after earliest Silurian time.

On Annette, Gravina, and Duke Islands the Klakas orogeny is associated with: 1) cessation of Ordovician-Early Silurian volcanic-plutonic activity, 2) regional greenschist- and locally amphibolite-facies metamorphism and deformation of Ordovician-Early Silurian rocks, 3) emplacement of large bodies of Late Silurian trondhjemite and subordinate granite, and 4) uplift and erosion of Late Silurian and older rocks prior to deposition of middle Lower Devonian clastic strata (Gehrels and others, in review). The occurrence of rocks of greenschist and perhaps amphibolite facies beneath Devonian strata on southern Annette Island indicates that the amount of structural uplift was approximately 5 to 10 km.

The age of deformation and metamorphism on Annette, Gravina, and Duke Islands is constrained by 426 ± 15 and 429 ± 20 Ma apparent ages on dioritic rocks that are penetratively deformed, by 415 ± 5 Ma, 409 ± 20 Ma, and 410 ± 20 Ma ages on trondhjemitic rocks that cut the deformational fabrics, and by clasts of metadiorite in superjacent middle(?) Lower Devonian conglomerate. The occurrence of Late Silurian trondhjemite boulders in the Devonian conglomerate indicates that uplift of the pre-Devonian rocks continued into latest Silurian-earliest Devonian time.

Relations between the Klakas orogeny and the generation and emplacement of Late Silurian trondhjemitic to granitic bodies on Annette, Gravina, and Duke Islands and of Late Silurian leucodiorite bodies on southern Prince of Wales Island are uncertain. Trondhjemitic rocks consist almost entirely of quartz and sodic plagioclase (Berg, 1972) and clearly cut across deformational fabrics in the Early Silurian rocks. The undeformed leucodiorite on southern Prince of Wales Island consists of oligoclase and a few percent aegirine-augite, arfvedsonite, and garnet, but differs from the trondhjemite in that quartz is not present. These intrusive rocks are clearly different compositionally and mineralogically from the earlier arc-type(?) intrusive rocks, and they were emplaced approximately 20 m.y. after the youngest known member of the older suite. I speculate that these sodic intrusive rocks may have been generated by partial melting of rocks at greater depth during middle to Late Silurian metamorphism, but this has not been demonstrated by petrochemical or isotopic studies.

On central and northern Prince of Wales Island the Klakas orogeny is recorded by stratigraphic relations. On northern Prince of Wales Island and on smaller islands to the west and southwest, Ordovician-Lower Silurian strata of the Descon Formation are conformably overlain by or grade into a sequence of interbedded turbidites and shallow marine limestone of Silurian age (Eberlein and others, 1983; Brew and others, 1984), which are in turn conformably overlain by Lower Devonian fine-grained clastic strata of the Karheen Formation (Ovenshine, 1975). Initiation of the Klakas orogeny is recorded in these regions by layers and lenses of polymictic conglomerate which are interbedded with the Silurian strata. Clasts in the conglomerate consist of limestone, chert, graywacke, volcanic rock, and gabbroic, dioritic, granitic, and syenitic intrusive rocks (Eberlein and Churkin, 1970; Ovenshine and Webster, 1970; Brew and others, 1984). The available paleontologic data indicate that the conglomeratic strata were deposited during latest

Early and Late Silurian time (Eberlein and others, 1983).

Toward the south the Silurian limestone and clastic strata pinch out and Lower Devonian strata of the Karheen Formation unconformably overlie rocks of the Descon Formation (Eberlein and others, 1983; Ovenshine and others, 1969). According to Ovenshine and others (1969) and Ovenshine (1975), the Karheen Formation thickens and becomes more conglomeratic southward, and was deposited on a northwest-dipping paleoslope. These stratigraphic relations combined with the evidence of Silurian-earliest Devonian faulting and uplift to the south indicate that the Lower Devonian conglomeratic strata were probably derived from an uplifted source area in the Annette-Gravina-Duke-southern Prince of Wales Islands region. The southward disappearance of Silurian strata may be a result of post-depositional erosion from the uplifted region, or the strata may never have been deposited to the south. The occurrence of conglomerate with volcanic and plutonic clasts in Silurian strata to the north indicates that uplift of the source area may have begun in late Early Silurian time, in which case Silurian strata were probably never deposited to the south. This argument is supported by the close timing between the initiation of conglomerate deposition on northern and central Prince of Wales Island and the onset of metamorphism, deformation, and uplift of the Early Silurian and older rocks to the south.

Alternatively, the conglomeratic strata to the north could have been derived from northeastern Chichagof Island, where Loney and others (1975) report that rocks of Ordovician-Silurian(?) age were uplifted prior to Late Silurian time. Restoration of approximately 150 km of dextral displacement on the Chatham Strait fault (Hudson and others,

1981) indicates that northeastern Chichagof Island was probably west of northern Prince of Wales Island during Silurian time. Occurrence of uplifted regions on southern Prince of Wales Island and west of northern Prince of Wales Island indicates that the Silurian paleogeography may have had a general northwesterly trend. This trend is also recorded in Silurian-Devonian facies relations in the area (Eberlein and others, 1983; Loney and others, 1975).

Lower Devonian strata

Lower Devonian strata on southern Prince of Wales Island generally belong to a fining-upward section of conglomeratic red beds at the base, siltstone, mudstone, and limestone in the middle, and laminated mudstone and black shale at the top (Gehrels and Saleeby, in review a, b). Sedimentary structures and stratigraphic relations indicate that the conglomeratic strata low in the section were deposited in subaerial to intertidal environments during or soon after the Klakas orogeny. Subsidence of the region below sea level after the Klakas orogeny is recorded by middle Early Devonian conodonts in superjacent shallow-marine limestone (Savage and Gehrels, 1984).

On Annette and Hotspur Islands Lower Devonian clastic strata consist of interbedded mudstone, shale, limestone, siltsone, and subordinate sandstone, olistostromal layers, conglomerate, and tuffaceous mudstone (Gehrels and others, in review). Reefoidal fossils in the limestone record deposition in shallow-marine environments, and conodonts indicate that the strata were deposited during middle(?) Early Devonian time (Savage and Gehrels, in review). Latest Silurian-earliest Devonian uplift and erosion of the Late Silurian and older rocks are recorded by clasts of metadiorite and nondeformed trondhjemite in a conglomerate layer on southern Annette Island. The clastic strata on Hotspur Island and locally on Annette and southern Prince of Wales Islands are conformably overlain by andesitic breccia and pillow flows (Gehrels and others, in review; Herreid and others, 1978; Gehrels and Saleeby, in review b).

Lower Devonian strata on central Prince of Wales Island include clastic strata of the Karheen Formation, and, in restricted areas, basaltic and rhyolitic volcanic rocks, sedimentary breccia, and limestone (Eberlein and others, 1983). As described by Ovenshine and others (1969), strata in the Karheen Formation to the north belong to a regional clastic wedge that was derived from uplifted regions to the south. The coarse clastic strata on southern Prince of Wales Island probably belong to a more proximal facies of this clastic wedge, as they were apparently deposited within the topographically rugged source region. Strata on Annette and Hotspur Islands only locally record deposition in such high-energy environments, and probably accumulated after much of the area had subsided below sea level. The tectonic significance of the superjacent andesitic rocks is not known.

Upper Triassic rocks

Upper Triassic volcanic and sedimentary rocks of the Hyd Group overlie Devonian and older rocks on Annette and Gravina Islands on a regional unconformity (Gehrels and others, in review). The base of the section is in most areas a polymictic conglomerate or breccia which contains large clasts of Devonian and older rocks. These conglomeratic strata grade upsection into several hundred meters of rhyolite and rhyolitic tuff, which is conformably overlain by up to 100 m of thick-bedded to massive limestone. A thick section of calcareous siltstone and thin-bedded limestone overlies the massive limestone, and grades upsection into a thick sequence of basaltic flows and breccia. The unconformity at the base of the section records regional uplift and erosion, but is apparently not the result of a major deformational event as underlying Devonian strata in the region are not more highly deformed and metamorphosed than superjacent Triassic strata.

Occurrence of the Upper Triassic strata in SE Alaska in a fairly narrow belt along the eastern margin of the terrane (Gehrels and Berg, 1984) combined with the bimodal (basalt-rhyolite) composition of the volcanic rocks indicate that the strata may have been deposited in a rift environment (Gehrels and others, in review). Movement on a regional low-angle fault (Keete Inlet fault) on southern Prince of Wales Island, and emplacement of steeply dipping mafic dikes in much of the southern part of the terrane may have been related to this interpreted rifting event (Gehrels and others, in review; Gehrels and Saleeby, in review b).

A large body of pyroxene gabbro was emplaced on Duke Island during deposition of the Upper Triassic strata on Annette and Gravina Islands. This gabbro was originally assigned a Jurassic age based on a K-Ar date of 177 ± 3 Ma by M.A. Lanphere (in Smith and Diggles, 1981). Based in large part on the proximity of this gabbro to the Cretaceous zoned ultramafic bodies on Duke Island, Murray (1972) and James (1971) argued that the gabbro and the ultramafic rocks were genetically related. A U-Pb (zircon) apparent age of 226 ± 3 Ma on the gabbro demonstrates, however, that the two suites cannot be genetically related (assuming that the ultramafic rocks are indeed Cretaceous in age!), as Taylor (1967) argued based on petrologic and geochemical differences. I suggest instead that the gabbro may be subvolcanic to basaltic rocks in the Hyd Group.

Bokan Mountain Granite (Jurassic)

The Bokan Mountain Granite is a peralkaline ring-dike complex on southern Prince of Wales Island that consists of aegirine- and arfvedsonite-bearing granite, aplite, porphyry, and pegmatite (Thompson and others, 1982; Saint-André and others, 1983). This body has recieved considerable research attention because of its anomalous composition and significant U-Th mineralization, and has not been mapped during my study. K-Ar, Rb-Sr, and U-Pb (zircon) geochronologic analyses indicate that the body is probably Jurassic in age, but dates range from 190 ± 8 Ma to a minimum age of 156 Ma (Saint-André and others, 1983; Lanphere and others, 1964; Armstrong, in press). The regional and tectonic significance of the Bokan Mountain Granite is unknown, as it is the only such intrusive body known in the Alexander or adjacent terranes. The alkalic composition of the pluton is consistent, however, with the interpretation that it was generated in response to the hypothesized latest Paleozoic-Triassic rifting event.

Jurassic-Lower Cretaceous strata

Jurassic-Lower Cretaceous strata consist of andesitic flows, breccia, and tuff, volcaniclastic strata, and fine-grained marine clastic strata that overlie the Upper Triassic strata on eastern Annette and Gravina Islands. These strata belong to a belt of upper Mesozoic sedimentary and volcanic rocks that occurs along the eastern margin of the Alexander and adjacent terranes and has been referred to as the Gravina-Nutzotin belt (Berg and others, 1972). The regional and tectonic significance of these strata is discussed by Gehrels and

Saleeby (1985, in review c).

Late Cretaceous-early Tertiary metamorphism and deformation

Rocks on Annette, Gravina, and Duke Islands were involved in a regional metamorphic and deformational event during Late Cretaceous-early Tertiary time. This event resulted in greenschist-facies metamorphism of rocks along the eastern sides of the islands and sub-greenschist-facies metamorphism to the west. A regional sub-greenschist-facies overprint of rocks on southern Prince of Wales Island may also be related to this event. As described by Gehrels and Saleeby (1985; in review c), this deformation and metamorphism are interpreted to have occurred in response to juxtaposition of the Alexander terrane against terranes to the east, which began in mid-Cretaceous time and continued into early Tertiary time.

GEOLOGY OF THE ALEXANDER TERRANE IN THE SAINT ELIAS MOUNTAINS AND IN CENTRAL AND NORTHERN SE ALASKA

Overview

The Alexander terrane has been subdivided by Berg and others (1978) and Monger and Berg (1984) into the Craig, Annette, and Admiralty subterranes based on apparent differences in their geologic history. The Craig subterrane consists of Cambrian (and perhaps Proterozoic) to Triassic rocks that underlie Prince of Wales Island, much of SE Alaska to the north, and all of the terrane in the Saint Elias Mountains. Annette, Gravina, and Duke Islands and part of the mainland to the southeast have been assigned to the Annette subterrane, which is distinguished from the Craig subterrane based on the absence of upper Paleozoic strata and on apparent dissimilarities in the Devonian and older rocks. My mapping has shown, however, that the Devonian and older rocks in the two subterranes are directly correlative, and that upper Paleozoic strata are also absent in adjacent parts of the Craig subterrane (Gehrels and others, in review). I therefore assign the Paleozoic and Triassic rocks on Annette, Gravina, and Duke Islands and regions to the southeast to the Craig subterrane and abandon the term Annette subterrane.

The Admiralty subterrane is distinguished from the Craig subterrane based on apparent dissimilarities in the Carboniferous and older rocks. Permian and Triassic strata in the Admiralty and Craig subterranes are similar, however, indicating that the subterranes have been closely associated since at least Early Permian time (Jones and others, 1981). Discussions of the geology of the Admiralty subterrane and comparisons with the geology of the Craig subterrane are inhibited by the lack of detailed mapping and modern geologic studies in the pre-Permian rocks of the Admiralty subterrane. I accordingly restrict my discussion of pre-Permian rocks in this report to rocks in the Craig subterrane (Fig. 4-1).

Rocks in the northern and central parts of the Alexander terrane have been mapped and described primarily by Campbell and Dodds (1982a-c and 1983a-c) in the Saint Elias Mountains, and by Brew and others (1978) in Glacier Bay, Loney and others (1975) on Chichagof Island, Lathram and others (1965) on Admiralty Island, Brew and others (1984) on Kuiu, Kupreanof, and northern Prince of Wales Islands, and Eberlein and others (1983) on central and northern Prince of Wales Island (Fig. 4-1). The geologic relations in SE Alaska have been summarized on the geologic map and unit descriptions of Gehrels and Berg (1984). Rocks that probably belong to the Alexander terrane also occur to the southeast along the coast of British Columbia (Tipper and others, 1981: Fig. 4-1). These rocks are currently being studied by Peter Van der Heyden (University of British Columbia) and Glenn Woodsworth (Geological Survey of Canada) and are not described herein.

In this section I outline the geology of the Alexander terrane north of southern SE Alaska. The following descriptions are derived primarily from maps and reports of Campbell and Dodds in the Saint Elias Mountains, and the compilation map of SE Alaska by Gehrels and Berg (1984).

Lower Paleozoic rocks

The oldest rocks recognized in the central and northern parts of the Alexander terrane consist of basaltic to andesitic volcanic rocks, volcaniclastic strata, limestone, and gabbro of Cambrian age that occur in the eastern Saint Elias Mountains. Similarities in rock types and stratigraphic position suggest that these rocks may be at least in part equivalent to rocks in the Wales metamorphic suite to the south, and may therefore underlie much of the terrane in between. In the Saint Elias Mountains the Cambrian rocks are overlain by a thick and apparently continous sequence of Lower Ordovician to Upper Devonian-Lower Mississippian limestone and marine clastic strata which do not record either the Wales "orogeny" or the Klakas orogeny.

In much of central and northern SE Alaska, Silurian turbidites and shallow-marine limestone are the oldest rocks exposed. Facies relations and detritus in the turbiditic strata suggest that they were derived both from Ordovician(?)-Silurian plutonic rocks on northeastern Chichagof Island and from volcanic rocks belonging to or coeval with the Descon Formation on Prince of Wales Island (Brew and others, 1984; Loney and others, 1975; Eberlein and others, 1983). As noted above, uplift of these plutonic and volcanic source regions is interpreted to mark the onset of the Klakas orogeny. Other manifestations of this middle Silurian-earliest Devonian orogenic event to the north include deposition of 1) Silurian-Lower Devonian conglomeratic strata on central and northern Prince of Wales Island, as described above, 2) Lower Devonian(?) arkosic strata on Kuiu Island (Brew and others, 1984), and 3) Lower(?) Devonian conglomeratic strata of the Cedar Cove Formation on Chichagof Island, which were apparently shed from an uplifted landmass to the southwest (Loney and others, 1975). In some parts of SE Alaska the Silurian strata are overlain by a variety of clastic strata, volcanic rocks, and limestone of Devonian age.

Upper Paleozoic rocks

Upper Paleozoic strata in the Saint Elias Mountains constitute a thick and laterally extensive section of argillite, shale, siltstone, sandstone, and limestone. In SE Alaska, upper Paleozoic strata occur only locally and consist of Mississippian and Pennsylvanian limestone, chert, calcareous sandstone and siltstone, and locally gypsum, and Lower Permian basaltic rocks, marine clastic strata, and limestone. These strata generally record deposition in tectonically stable marine environments, although the Devonian and Lower Permian volcanic rocks indicate that magmatic activity occurred at least locally during late Paleozoic time.

The only late Paleozoic plutonic rocks in the Alexander terrane are large dioritic to syenitic bodies of late Pennsylvanian-Early Permian age in the northernmost part of the terrane and locally in southern SE Alaska. Emplacement of these intrusive bodies occurred at approximately the same time that deeper-water strata of the Admiralty subterrane were juxtaposed against shallow-marine strata of the Craig subterrane (Jones and others, 1981). These relations apparently record a minor late Pennsylvanian-Early Permian tectonic disturbance in the Alexander terrane.

Upper Triassic strata

Upper Triassic strata occur along the eastern margin of the terrane in SE Alaska and have only locally been distinguished from upper Paleozoic strata in the Saint Elias Mountains. In SE Alaska the section is generally similar to that described on Annette and Gravina Islands and is everywhere separated from subjacent Permian and older rocks by an unconformity. Upper Triassic strata in the Saint Elias Mountains consist of a thick section of interbedded basaltic and more felsic flows and breccia, limestone, and marine clastic strata which are interpreted to have formed in a rift environment (MacIntyre, 1984).

TECTONIC EVOLUTION OF THE ALEXANDER TERRANE

Our detailed work in southern SE Alaska combined with the work of others to the north indicates that the Alexander terrane evolved in a tectonically active environment during Cambrian (Proterozoic) through Early Devonian time, in a tectonically stable environment during Middle Devonian through Early Permian time, and in a rift environment during Late Permian(?) and Triassic time. The tectonic evolution during each of these three phases is outlined below and is shown schematically on Figure 4-2.

Cambrian (Proterozoic?) through Early Devonian time

The initial phase in the evolution of the Alexander terrane is characterized by Cambrian volcanism in the Saint Elias Mountains and by

Cambrian and (or) Proterozoic through Early Devonian volcanism, plutonism, and orogenic activity in the southern part of the terrane. This tectonically active phase begins with the deposition of volcanic and sedimentary rocks of Cambrian age in the Saint Elias Mountains and of Cambrian or Proterozoic age in the Wales metamorphic suite of southern SE Alaska. Comparison with modern-day magmatic systems suggests that these rocks accumulated in a volcanic arc environment. The apparent lack of older rocks and the occurrence of correlative(?) rocks in the Saint Elias Mountains and in southern SE Alaska suggests that these arc-type rocks may form the crystalline basement beneath much of the terrane.

The arc-type(?) rocks in southern SE Alaska were penetratively deformed, regionally metamorphosed, and uplifted(?) during the Middle Cambrian-Early Ordovician Wales "orogeny," but deposition of marine clastic strata and limestone was continuous from Late Cambrian to Early Ordovician time in the Saint Elias Mountains. Structures associated with the Wales "orogeny" generally record northeast-southwest-directed shortening, but it is not known whether this tectonic event occurred in response in interplate or intraplate processes.

Volcanism and plutonism resumed soon after the Wales "orogeny" in the southern part of the terrane and continued until approximately middle Early Silurian time. These rocks have strong similarities to rocks in young and presently active volcanic arcs and are accordingly interepreted to have formed in a volcanic arc environment along a convergent plate margin. Facies relations in Ordovician-Silurian strata and structures that formed during Silurian-Devonian time indicate that this arc system trended northwesterly, at an oblique angle to the present north-northwesterly elongation of the terrane. The occurrence of shallow-marine limestone and clastic strata for over 600 km to the north-northwest indicates that this interpreted volcanic arc system faced to the southwest rather than northeast, and that the Ordovician-Silurian strata to the north-northwest accumulated in a back-arc environment.

The tectonic evolution of the southern Alexander terrane changed dramatically during the middle Silurian-earliest Devonian Klakas orogeny (Gehrels and others, 1983a). Early manifestations of the orogenic event include cessation of the arc-type volcanism and plutonism, uplift of volcanic and plutonic source areas, and deposition of polymictic conglomerate in shallow-marine environments. Regional metamorphism, penetrative deformation, southwest-directed movement on thrust faults, and structural uplift of as much as 10 km began prior to the end of Silurian time, and topographically rugged subaerial regions were uplifted prior to middle Early Devonian time. Late Silurian trondhjemitic and sodic leucodiorite bodies record a significant change from the pre-Late Silurian arc-type magmatism, and may have been generated by anatexis during the regional metamorphism. By middle Early Devonian time most regions had subsided below sea level and the terrane began to evolve in a tectonically stable marine environment.

Structures associated with the Klakas orogeny record northeast-southwest-directed shortening and, at least locally, southwestward movement of higher-grade rocks over lower-grade rocks. It is not known, however, whether this orogenic event occurred in response to interplate processes along a convergent plate boundary, or intraplate processes within the interpreted volcanic arc.

Middle Devonian through Early Permian time

The tectonic evolution of the Alexander terrane from Middle Devonian through Permian time is difficult to interpret because strata of this age occur in only a few areas of SE Alaska and have not been mapped in detail in the Saint Elias Mountains. In SE Alaska, upper Paleozoic strata of the Craig subterrane generally formed in tectonically stable, shallow-marine environments, although Devonian and Lower Permian volcanic rocks record magmatic activity in some areas. The restricted distribution of the volcanic rocks and the lack of Devonian plutons in the older rocks suggest that this volcanic activity was not of the same regional and tectonic significance as the early Paleozoic magmatism.

Upper Paleozoic strata in the Saint Elias Mountains apparently accumulated in tectonically stable marine environments, and volcanic rocks have not been recognized in the section. Although little is known about the late Pennsylvanian-Early Permian dioritic to syenitic bodies in the Saint Elias Mountains, their emplacement does not appear to represent a major tectonic event in the Alexander terrane (Fig. 4-2). Emplacement of these intrusive bodies and juxtaposition of deeper-water strata of the Admiralty subterrane against shallow-marine strata of the Craig subterrane are accordingly interpreted to represent a minor tectonic disturbance (Fig. 4-2).

Based on the available evidence, the Alexander terrane appears to have evolved in a tectonically stable environment during much of late Paleozoic time, with volcanism in restricted(?) areas during Devonian and Early Permian time, and a minor(?) disturbance occurring during late Pennsylvanian-Early Permian time. In contrast to early Paleozoic time, geologic relations are consistent with the interpretation that the terrane was in an intraplate oceanic environment during much of late Paleozoic time.

Late Permian and Triassic time

The third phase in the evolution of the Alexander terrane is recorded by Upper Triassic rocks and by a major unconformity which separates these strata from Early Permian and older rocks in SE Alaska. The unconformity at the base of the section represents a major uplift and erosional event which began after deposition of the Lower Permian strata and continued during deposition of the lower part of the Upper Triassic section. Based on the bimodal (basalt-rhyolite) composition of volcanic rocks in the section, the linear trace of the strata along the eastern margin of the terrane in SE Alaska, and the lack of evidence of Late Permian-Triassic deformation and metamorphism, I suggest that the Upper Triassic strata formed in a rift environment. Structural and stratigraphic relations in the Saint Elias Mountains indicate that Upper Triassic strata to the north were also deposited in a rift environment (MacIntyre, 1984). Occurrence of these strata along the eastern margin of the terrane in SE Alaska indicates that the terrane may have been much larger prior to Late Triassic time. Jurassic-Lower Cretaceous strata of the Gravina-Nutzotin belt were apparently deposited along this rifted(?) margin (Gehrels and Saleeby, in review c).

DISPLACEMENT HISTORY OF THE ALEXANDER TERRANE

Geologic evidence and previous syntheses

Evolution during Paleozoic time

Wilson (1968) originally hypothesized that the Alexander terrane is a tectonic fragment which did not form in its present position in the North American Cordillera and implied that it did not arrive until after early Mesozoic time. Monger and Ross (1971) provided the first documentation of far-travelled components in the Cordillera through their recognition of Permian fusulinid fauna with "Tethyan" affinities in the Cache Creek terrane of central British Columbia. They suggested that rocks in western British Columbia and SE Alaska belong to an island arc that was juxtaposed against the Cache Creek terrane, and they recognized that Permian fauna in this arc terrane are similar to fauna in terranes inboard of the Cache Creek. Monger and Ross (1971) presented two tectonic scenarios for this juxtaposition. In their first scenario, the outboard fragment is envisioned as an exotic component in the Cordillera, with no primary ties to rocks to the east. This hypothesis was not favored by these workers because of the faunal similarity of the outboard and inboard terranes. The second scenario involves derivation of the outboard terrane from an original position south of and along trend with the inboard terrane (Monger and Ross, 1971, their Fig. 4-5). They hypothesized that the outboard arc was juxtaposed against the western margin of the Cache Creek terrane by right-lateral movement on a strike-slip fault oriented oblique to the continental margin.

Jones and others (1972) adopted this second scenario and suggested more specifically that the Alexander terrane was originally contiguous with Paleozoic rocks in the Sierra-Klamath region of California and Oregon. Their argument is based on: 1) deposition of Silurian rocks in both regions across a northwestward transition from shallow-marine carbonates to deeper-water graywacke and flysch-like deposits in the

Cordilleran miogeocline, 2) the occurrence of similar Permian fusulinid fauna in the Alexander terrane and in the Sierra-Klamath region, 3) general similarity of the Paleozoic geologic records of the two regions, 4) termination of the central (Cache Creek-affinity) and inboard (eastern) terranes of Monger and Ross (1971) in the Sierra-Klamath region, and 5) dissimilarity of rocks in the Alexander terrane with Paleozoic rocks elsewhere along the continental margin.

Monger and others (1972) recognized that the Alexander terrane formed in a volcanic arc environment, rather than along a passive continental margin, and demonstrated that the arc must have been displaced a significant distance from its site of origin. They acknowledged that the original position of the terrane is unknown, but noted that formation in the vicinity of northern California, as suggested by Jones and others (1972), is "compatible with what is known of the later history of the Canadian Cordillera" (Monger and others, 1972, pp. 584).

Churkin (1974) proposed a very different model in which the terrane formed as a volcanic arc near its present position in the Cordillera and has moved toward and away from North America during the closing and opening of marginal basins, but has not been displaced along the continental margin. Churkin and Eberlein (1977) emphasize that the Alexander terrane was probably not adjacent to rocks in the Sierra-Klamath region during Paleozoic time, pointing out specifically that Silurian facies relations in the Alexander terrane are much more complex than the simple northwestward transition suggested by Jones and others (1972). Churkin and Eberlein (1977, pp. 784) conclude with a scenario similar to that presented by Churkin (1974) in which the Alexander terrane evolved during early Paleozoic time as a west-facing volcanic arc near its present position in the Cordillera, switched polarity prior to or during Early Devonian time, and impinged on continental margin assemblages to the east during the Late Devonian-Early Mississippian Antler orogeny.

Based on an analysis of the Mesozoic evolution of the California margin, Schweikert (1976) suggested that the Alexander terrane formed in a volcanic arc environment outboard of California, and was displaced northward during Triassic time. Schweikert and Snyder (1981) present a more specific model in which the Alexander terrane evolved as an east-facing volcanic arc outboard of California during Ordovician-Silurian time, and coeval rocks in the Sierra-Klamath region accumulated in an accretionary prism along the leading edge of this arc. They suggest that the arc magmatism migrated eastward over the older accretionary prism during Devonian time, and then this composite arc collided with the western margin of North America during the Antler orogeny. According to their scenario the Alexander terrane remained adjacent to the Sierra-Klamath region during late Paleozoic time and the two assemblages evolved in a west-facing volcanic arc. The Alexander arc and a significant portion of the lower Paleozoic accretionary wedge not found in California are interpreted to have been displaced northward during Triassic time.

Girty and Wardlaw (1984) analyzed detrital zircons in a pre-Late Devonian feldspathic sandstone from the Shoo Fly melange in the Sierra Nevada and concluded that they were shed from a source terrane which contained igneous rocks of approximately 506 ± 22 Ma (Late Cambrian-Early Ordovician). They cite the euhedral shape of the zircons and the composition and textural immaturity of the feldspathic sandstone as evidence that the detritus was derived during a single cycle of erosion from a volcanic-plutonic provenance. Based on previous suggestions that the Alexander terrane may have been in the Sierra Nevada region during Paleozoic time, paleomagnetic evidence cited below, and the occurrence of early Paleozoic volcanic and plutonic rocks in the Alexander terrane, Girty and Wardlaw (1984) suggest that the detritus in the Shoo Fly melange may have been derived from the Alexander terrane during its residence along the California margin.

Evolution during Mesozoic time

Accretion of the Alexander terrane to its present position in the Cordillera was initially interpreted by Monger and others (1972) to have occurred during Middle Triassic time based on apparent similarity of Late Triassic volcanic rocks across the northwestern Canadian Cordillera. Schweikert (1976) and Schweikert and Snyder (1981) suggest a similar timing based on analyses of the Triassic evolution of California. In contrast to these interpretations, Berg and others (1972) report that a deep marine basin existed along the eastern margin of the Alexander terrane until the end of Early Cretaceous time, and suggest that the terrane was not accreted until Late Cretaceous time.

Paleomagnetic data

The interpretation that the Alexander terrane was in the northern California region during Paleozoic time received significant support from paleomagnetic studies of Paleozoic rocks on Prince of Wales and adjacent islands (Van der Voo and others, 1980). These workers report that Ordovician through Pennsylvanian strata yield paleolatitudes which are consistent with evolution in the northern California region, and that the declination data indicate post-Pennsylvanian counter-clockwise rotation (relative to North America) of approximately 25° to 35°. In contrast, Upper Triassic strata on northern Kuiu Island yield paleolatitudes which are consistent with the interpretation that the terrane was in approximately its present position with respect to North America by Late Triassic time (Hillhouse and Grommé, 1980). These workers report that the declinations measured on Triassic rocks indicate approximately 100° of counter-clockwise rotation (relative to North America) since Late Triassic time. The conclusions reached by Van der Voo and others (1980) and Hillhouse and Grommé (1980) are both based on the assumption that the terrane was located in the northern hemisphere during Paleozoic and Triassic time.

Paleontologic data

Fauna and flora from the Alexander terrane have been described in many reports, but the paleobiogeographic implications of these fossils have not been synthesized. In Table 4-1 I note fossils from the terrane that are reported to have similarities with fauna or flora in other regions. Although these similarities do not clearly demonstrate primary relations with any single region, most fauna are apparently endemic to the Circum-Pacific realm. In addition, early Paleozoic fauna commonly have Asiatic affinities, Carboniferous fossils show "Tethyan" affinities, and Mesozoic fossils were endemic to equatorial regions in the eastern part of the paleo-Pacific basin.

Discussion of previous syntheses

The scenarios outlined above offer five fundamentally different alternatives for the tectonic evolution of the Alexander terrane during Paleozoic time. These include:

Table 4-1. Fossil groups in the Alexander terrane and their apparent similarities with fossil groups from other regions.

Ordovician: Conodonts - similarities with fauna in the North Atlantic conodont province (Savage, 1984)

Silurian: Brachiopods - (Late Silurian) - similar to fauna in the eastern Urals (Kirk and Amsden, 1952).

Early Devonian: Conodonts - similar to fauna in western North America, eastern Australia, and central Asia (Savage, 1984).

Land plants (Baragwanatia) and graptolites - similar to an occurrence in eastern Australia (Churkin and others, 1969, p. 567)

Corals - similar to fauna in asiatic U.S.S.R. (Churkin and others, 1970)

Middle Devonian: Conodonts - Cordilleran North American affinity (Savage, 1984)

Corals - Similar to fauna in the U.S.S.R. (Tchudinova and others, 1974)

Late Devonian: Brachiopods - some degree of isolation(?) (Savage and others, 1978)

Corals - Similar to fauna in the U.S.S.R. (Tchudinova and others, 1974)

Mississippian: Conodonts - western North American affinity (Savage, 1984)

Foraminifera - Characteristic of a tropical oceanic arc or plateau environment, and not found in the Cordillera (Mamet, 1976); North American and Tethyan affinity (Dutro and others, 1981)

Pennsylvanian: Fusulinids - similar to fauna in Japan and the Cache Creek terrane (British Columbia) (Douglass, 1971)

Various fauna indicate a position near the middle of the paleo-Pacific and south of the paleoequator during early Pennsylvanian time, and closer to the paleoequator during late Pennsylvanian time (Ross and Ross, 1983, 1985)

Early Permian: Brachiopods - Similar fauna occur in the Urals, southern China, eastern Alaska, the Canadian Arctic, Yugoslavia, and in a melange in central Oregon (Grant, 1971)

Various fauna indicate a position near the middle of the paleo-Pacific ocean basin and near the paleoequator (Ross and Ross, 1983, 1985).

Triassic: Bivalves - eastern part of the paleo-Pacific ocean basin and near the paleoequator (Tozer, 1982); similar to fauna in Peru(?) (Newton, 1983).

1) derivation of the Alexander terrane from another continent - probably Asia (Wilson, 1968; Moores, 1970),

2) deposition of strata belonging to the Alexander terrane in the Cordilleran miogeocline, adjacent to rocks in the Sierra-Klamath region (Jones and others, 1972),

or formation of the Alexander terrane as a volcanic arc:

3) in an unknown region of the paleo-Pacific basin (Monger and Ross, 1971; Monger and others, 1972),

4) near its present position in the Candian Cordillera (Churkin, 1974; Churkin and Eberlein, 1977), or

5) outboard of the California-Nevada continental margin (Monger and others, 1972; Schweikert, 1976; Schweikert and Snyder, 1981; Girty and Wardlaw, 1984).

Derivation from the Sierra-Klamath region

Hypotheses in which the Alexander terrane is interpreted to have been contiguous with rocks in the Sierra-Klamath region during Paleozoic time are not supported by a comparison of the geologic records of the two regions. Jones and others (1972) suggest that rocks in these two regions were contiguous primarily on the basis of apparent similarities in their Silurian strata and facies relations. More recent studies in SE Alaska have shown, however, that the Silurian strata do not record a simple northwestward transition from shallow-marine limestone to deeper-water clastic strata as reported by Jones and others (1972). In contrast, the rocks record a northwestward transition from a sedimentary-volcanic-plutonic complex in southern SE Alaska to interbedded clastic strata and shallow-marine limestone that continue north-northwestward for at least 600 km. Facies relations in these strata indicate that Silurian paleogeographic trends were oriented northwesterly, rather than northeasterly as in the California-Nevada region. In addition, Ordovician-Silurian clastic strata in the Alexander terrane consist of volcaniclastic turbidites and of subordinate quartzo-feldspathic turbidites derived from Ordovician-Silurian syenitic bodies (Loney and others, 1975; Brew and others, 1984), whereas the reportedly correlative clastic strata in the Sierra-Klamath region contain units such as the Antelope Mountain quartzite (Klamath Mountains) and quartz-rich parts of the Shoo Fly Complex (Sierra Nevada), which were apparently derived from a continental source area.

Upper Paleozoic rocks in the Alexander terrane and in the Sierra-Klamath region are also quite different. Upper Paleozoic strata in the Alexander terrane were generally deposited in relatively stable marine environments, with volcanism occurring locally(?) in SE Alaska during Devonian and Early Permian time. The regional extent and significance of this volcanism are difficult to determine because of the restricted occurrence of upper Paleozoic strata in SE Alaska. In the Saint Elias Mountains, where upper Paleozoic strata are widepread, volcanic rocks have not been recognized (Campbell and Dodds, 1982a-c, 1983a-c). In contrast, upper Paleozoic volcanic rocks constitute much of the section in the Sierra-Klamath region, and are generally interpreted to have formed in a volcanic arc environment (Schweikert and Snyder, 1981).

Schweikert and Snyder (1981) base their interpretation that the Alexander terrane was adjacent to rocks in the Sierra-Klamath region primarily on a plate tectonic model for the evolution of California and Nevada: they apparently involve the Alexander terrane only to serve as an Ordovician-Silurian volcanic arc against which to accumulate rocks in their interpreted accretionary prism. Their model is difficult to evaluate using geologic criteria because they do not cite geologic relations which indicate that there are primary ties between the two regions. As described above, I suggest that the geologic histories of these two regions are quite different, and are not consistent with their having evolved adjacent to one another. I also note that the scenario of Schweikert and Snyder (1981) is not consistent with my conclusion that the Ordovician-Silurian arc in the Alexander terrane faced to the southwest rather than east, and with my observation that a lower Paleozoic accretionary wedge which they suggest was displaced northward with the terrane has not been recognized in western Canada or southern Alaska.

The only tangible geologic evidence which has been presented in support of a primary tie between the two regions is the occurrence of detrital zircon populations of apparent Late Cambrian-Early Ordovician age in the Shoo Fly melange (Sierra Nevada), which Girty and Wardlaw (1984) suggest were derived from the Alexander terrane. Although intriguingly consistent with previous scenarios, source rocks of the appropriate composition and age have not been recognized in the Alexander terrane. Rocks in the terrane that could reasonably yield zircon populations with the isotopic characteristics reported by Girty and Wardlaw (1984, their Figure 3) include: 1) silicic metavolcanic rocks of the Wales metamorphic suite, 2) Middle Ordovician-Early Silurian plutonic rocks that contain zircons with a significant component of inherited pre-Ordovician zircon, and 3) Middle and Late Cambrian metadiorite layers in the Wales metamorphic suite. I submit that the silicic metavolcanic rocks in the Wales suite are not a likely source for the zircon because they are intruded by metadiorite layers of Middle Cambrian age, and are therefore too old, and because they are very poor in zircon. The Ordovician-Early Silurian rocks are also an unlikely source in that they are too young, and only rarely contain detectable quantities of inherited zircon.

Middle and Late Cambrian metadiorite layers associated with the Wales suite yield zircon of nearly the appropriate age (approximately 540-520 Ma; Gehrels and others, in review), but these layers are too restricted in regional extent to have been a reasonable source region for the sandstone described by Girty and Wardlaw (1984). Larger plutons of this age could conceivably occur beneath the thick cover of Ordovician and younger strata in other parts of the terrane, but the existence of such plutons remains speculative. A more reasonable source region for the detritus described by Girty and Wardlaw (1984) might be the alkalic plutons of Cambrian(?)-Ordovician age in east-central Idaho (Evans, 1984), which is near the interpreted source area for much of the detritus in the Shoo Fly Complex (Schweikert and Snyder, 1981, their Figure 7-5).

The proposed occurrence of similar Permian fauna in the Alexander terrane and in the Sierra-Klamath region (Jones and others, 1972) remains a significant line of evidence in support of proximity of the two regions during Permian time. However, Ross and Ross (1983, 1985) have recently re-evaluated faunal data from various Paleozoic terranes in the Cordillera and conclude that the Alexander terrane and rocks in the inboard (eastern) belt of Monger and Ross (1971) and Jones and others (1972) were both located just south of the equator during Permian time but were separated by a fair longitudinal distance (their Figure 3). The occurrence of similar Early Permian fauna in the two regions therefore apparently indicates that they were in the same faunal regime, but not adjacent to one another or to the North American continental margin.

Evolution near its present position in the Cordillera

The hypothesis by Churkin (1974) and Churkin and Eberlein (1977) that the Alexander terrane evolved near its present position in the Cordillera is regarded as untenable based on the occurrence of Permian fauna with "Tethyan" affinity inboard of the terrane (Monger and Ross, 1971; Monger and others, 1972; Ross and Ross, 1983, 1985). In addition, paleomagnetic data from Paleozoic rocks in the Alexander terrane indicate that it has been displaced northward by at least 18° relative to North America since Pennsylvanian time (Van der Voo and others, 1980).

Location during Mesozoic time

Hillhouse and Grommé (1980) report that the Alexander terrane has been in its present position with respect to North America since Late Triassic time based on paleomagnetic data from Upper Triassic rocks in the terrane. This interpretation is inconsistent, however, with the observation that the Alexander terrane was separated from inboard terranes by a marine basin until mid-Cretaceous time (Berg and others, 1972), and the geologic, paleomagnetic, and paleobiogeographic evidence which suggests that the Alexander terrane and terranes to the east were farther south during Late Triassic time.

Tozer (1982) recognized in particular that Late Triassic fauna in the Alexander terrane are endemic to low paleolatitudes, and concluded that the interpretation of Hillhouse and Grommé (1980) is unlikely. Based on an analysis of paleomagnetic data from a variety of terranes in the western Cordillera, Stone and others (1982) propose that the paleomagnetic data of Hillhouse and Grommé (1980) record a southern paleopole and that the Alexander terrane was in the southern hemisphere (at approximately 43°S) during Late Triassic time. Newton (1983) reached the same conclusion based on the occurrence of similar Late Triassic bivalves in the Alexander terrane and in Peru. As described by Gehrels and Saleeby (in review c), the available geologic, paleontologic, and paleomagentic data are most consistent with location of the terrane in the eastern part of the paleo-Pacific basin and south of the equator during Late Triassic time. The occurrence of Upper Jurassic-Lower Cretaceous marine strata along the eastern margin of the terrane indicates that it was not accreted until after Early Cretaceous time (Berg and others, 1972; Gehrels and Saleeby, 1985, in review c).

Comparison with other orogenic belts

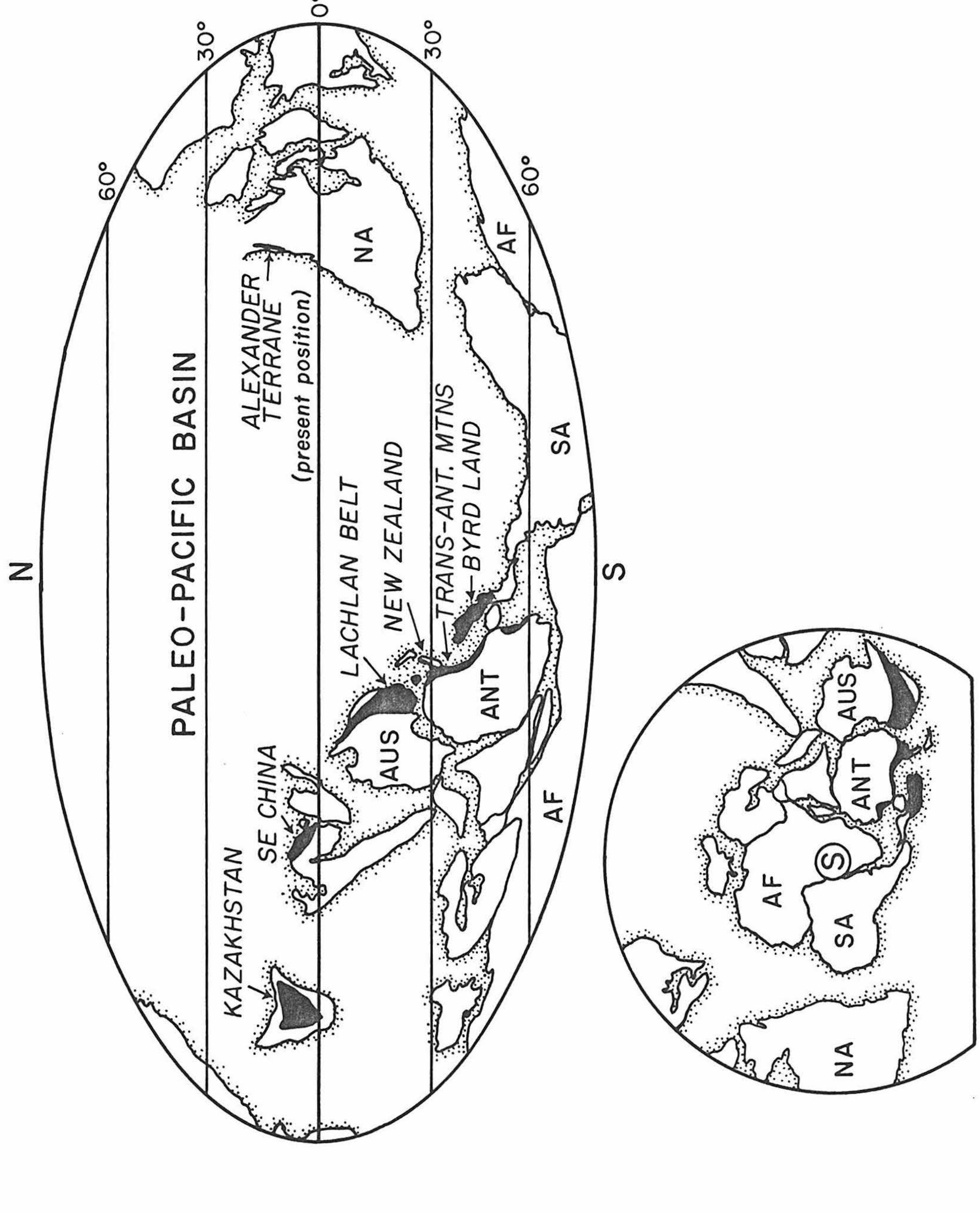
Our geologic and geochronologic studies in southern SE Alaska indicate that the Alexander terrane formed and evolved in a long-lived and regionally extensive volcanic arc system during early Paleozoic time. The distinct and well-constrained history of Cambrian (and perhaps Proterozoic) through Early Silurian magmatic activity combined with Middle Cambrian-Early Ordovician and middle Silurian-earliest Devonian orogenic events serves as a basis for comparisons between the Alexander terrane and other orogenic belts. Additional criteria used in delineating potential correlatives include the apparent Circum-Pacific affinity of fauna and flora of the Alexander terrane, and the low paleolatitudes indicated by the paleomagnetic data of Van der Voo and others (1980). The late Paleozoic geologic and faunal characteristics of the terrane are equivocal regarding proximity to other terranes or landmasses, and are generally consistent with my interpretation that the terrane was in an intraplate oceanic setting.

Orogenic systems which apparently satisfy these criteria occur in eastern Australia, New Zealand, the Transarctic Mountains and Byrd Land of Antarctica, and in various tectonic fragments in Asia. The approximate distribution of these orogenic systems during Silurian time is shown on Figure 4-3, which indicates that these presently disparate early Paleozoic orogens formed a nearly continuous belt along the paleo-Pacific margin of Gondwana. Cooper and Grindley (1982) have recently summarized the geologic and tectonic evolution of parts of this orogenic system in Antarctica, New Zealand, Tasmania, and southeastern Australia, and Cas (1983) reviewed in detail the geologic evolution of the orogen in southeastern Australia, where it is referred to as the Lachlan Fold Belt. Because the Lachlan Belt has been studied in greater detail than the other regions, the following outline of the geologic and tectonic evolution of the orogen focuses on the history of the Lachlan Belt and is derived primarily from Cas (1983).

Geologic and tectonic evolution of the Lachlan Fold Belt

The Phanerozoic geologic history of the Lachlan Belt begins in Cambrian time with the widespread deposition of Lower, Middle, and locally Upper Cambrian mafic to intermediate volcanic rocks, marine clastic strata, and subordinate limestone in western parts of the orogen (also referred to as the Kanmantoo Fold Belt). Most volcanic rocks have calc-alkaline affinities and are generally interpreted to have formed in a convergent-margin environment. Others are alkaline in composition and

Figure 4-3. Silurian continental reconstruction (from Scotese and others, 1979) showing the distribution of early Paleozoic orogenic belts with which the Alexander terrane may have been associated.


PALEO-PACIFIC BASIN
N
S
60°
30°
0°
30°
60°
KAZAKHSTAN
SE CHINA
AUS
LACHLAN BELT
NEW ZEALAND
TRANS-ANT. MTNS
BYRD LAND
ANT
AF
SA
AF
NA
ALEXANDER
TERRANE
(present position)
AF
SA
ANT
AUS
NA
S

may record rift-related volcanism. In some areas the Cambrian strata overlie Precambrian crystalline or stratified rocks or ophiolitic(?) ultramafic rocks, but in most areas the depositional basement is unknown. Beginning in Late and locally Middle Cambrian time, and continuing into the Early Ordovician, rocks in many areas were penetratively deformed and metamorphosed and were intruded by granitic plutons. This orogenic event has been referred to as the Delamerian orogeny in the Lachlan Fold Belt and the Ross orogeny in Antarctica.

Following the Middle Cambrian-Early Ordovician tectonic activity, Lower Ordovician conglomerate and quartz-rich clastic strata, and Middle and Upper Ordovician shallow-marine limestone and clastic strata were deposited in the western part of the belt. These strata grade eastward into deeper-water clastic strata which were deposited throughout much of the central Lachlan Belt. In the eastern part of the belt, mafic to intermediate volcanic rocks were deposited, in some areas in association with shallow-marine limestone. The geochemical affinity of these volcanic rocks suggest that they formed in a volcanic arc environment, and regional relations indicate that this arc faced to the east (Cas and others, 1980). Powell (1983) suggests that rocks along the eastern margin of the volcanic arc (Molong arc) constitute a fore-arc basin and accretionary wedge (Monaro trough and Narooma terrane), and Cas (1983) and Scheibner (1984) suggest that the marine clastic strata west of the arc were deposited in a marginal basin (Wagga trough). Granitic rocks of Ordovician age occur only locally in the western part of the belt.

The relatively simple Ordovician paleogeography became considerably more complex during Silurian time. Volcanism ceased in the Molong arc at the end of Ordovician time, and deep-marine strata characteristic of the Ordovician were deposited in only a few areas during Early Silurian time. Beginning in middle Silurian time and reaching a peak during the latest Silurian-earliest Devonian, many regions were uplifted above sea level and covered by terrigeneous red beds and silicic volcanic rocks. Deeper-water strata were deposited in relatively narrow troughs separated by shallow-marine or terrestrial regions, and plutons of predominantly S-type were emplaced in much of the Lachlan Belt. Emergence during Early Silurian time marks the onset of the Benambran and Quidongan orogenies, which are manifest in some areas by penetrative deformation and regional metamorphism. This orogenic activity apparently continued into latest Silurian-middle Early Devonian time, when rocks in much of the Lachlan Belt were deformed and metamorphosed during the Late Silurian-middle Early Devonian Bowning orogeny. This event is referred to as the Bindian orogeny in part of the Lachlan Belt, and similar deformational and metamorphic events occurred during this time in Antarctica (Borchgrevink orogeny) and New Zealand (early phase of Tuhuan orogeny). Most workers envision the region as a wide silicic volcanic arc in a convergent margin environment during Silurian time, and Scheibner (1984) ascribes the orogenic activity to collision of the Ordovician volcanic arc (Molong arc) against the marginal basin (Wagga trough) to the west.

Middle Early to late Middle Devonian rocks are preserved in only a few areas, where they consist of shallow-marine limestone, clastic strata, and subordinate volcanic rocks. Granitic plutons were emplaced in the southeasternmost part of the belt and are of I-type. This interval is interpreted by Cas (1983) to record a transition from a tectonically active phase during Silurian-Early Devonian time to another tectonically active phase beginning in late Middle Devonian time. The onset of the next phase of tectonism is marked by the Middle to Late Devonian Tabberabberan orogeny, which is manifest as regional deformation and metamorphism in the Lachlan Belt and adjacent regions. During Late Devonian-Early Carboniferous time conglomeratic red beds were deposited in much of the Lachlan Belt and silicic and subordinate basaltic volcanic rocks were also widespread. Granitic plutons of both S- and I-type were emplaced in the southern part of the belt. The tectonic environment in which the Tabberabberan orogeny occurred is not known, although Cas (1983) suggests that the succeeding Upper Devonian-Lower Carboniferous sedimentary and bimodal volcanic rocks were deposited in an extensional environment. He cites the Basin and range Province of the western U.S.A. as a tectonic analogue. Deformation occurred in some regions of the belt during late Early Carboniferous time, marking the Kanimbran orogeny, and the Lachlan Belt has been tectonically stable since this event.

The tectonic evolution of the Lachlan Fold Belt can be summarized as follows. Early through Middle and locally Late Cambrian volcanic and sedimentary rocks are the oldest rocks in most of the orogen and record evolution in a volcanic arc environment. These rocks were deformed and metamorphosed during the Late Cambrian-Early Ordovician Delamerian orogeny. Following this orogenic event, an east-facing volcanic arc formed offshore and was separated from the Australian continental margin by a marginal basin. This arc began to impinge against the marginal basin and a wide shelf region to the west, initiating a period of regional uplift, deformation, metamorphism, anatectic(?) volcanism and plutonism, and deposition of terrigeneous strata that lasted from Early

Silurian through middle Early Devonian time. This orogenic activity has been subdivided into the Benambran (primarily Early Silurian), Quidongan (middle Silurian), and Bowning (Late Silurian-middle Early Devonian) orogenies. Following a period of relative stability, orogenic activity resumed in late Middle Devonian time with the onset of the Tabberabberan orogeny. Upper Devonian-Lower Carboniferous rocks are interpreted to have formed in an extensional regime, perhaps analogous to the Basin and Range province of the western U.S.A. The orogen experienced the minor Kanimblan orogeny near the end of Early Carboniferous time and has remained tectonically stable since.

Comparison between the Alexander terrane and the Lachlan Fold Belt

I conclude from my work in SE Alaska and these syntheses of the Lachlan Belt that the pre-Middle Devonian evolution of the Lachlan Fold Belt and of the southern Alexander terrane are remarkably similar. The apparent similarities include not only the nature and age of geologic units, but also the tectonic environments in which the units formed, and the timing and to some degree the style of the major orogenic events. These similarities include:

1) Arc-type volcanism during Cambrian time (and perhaps Proterozoic time in the Alexander terrane),

2) Cessation of this arc-type volcanism during the Middle Cambrian-Early Ordovician Wales "orogeny" (Alexander terrane) and Delamerian orogeny (Lachlan Belt),

3) Arc-type volcanism and plutonism resuming soon after the Middle Cambrian-Early Ordovician orogenic events and continuing through Ordovician time in the Lachlan Belt and into Early Silurian time in the

Alexander terrane,

4) Cessation of this arc-type magmatic activity during major Silurian-Early Devonian orogenic events (Benambran, Quidongan, and Bowning orogenies in the Lachlan Belt and Klakas orogeny in the Alexander terrane). Orogenic events in both regions are manifest by uplift and erosion of the older volcanic, plutonic, and sedimentary rocks, regional deformation and metamorphism in some areas, movement on thrust faults, anatectic(?) plutonism (and volcanism in the Lachlan Belt), and deposition of conglomeratic red beds, and

5) Deposition of strata in tectonically stable marine environments during middle Early to Middle Devonian time.

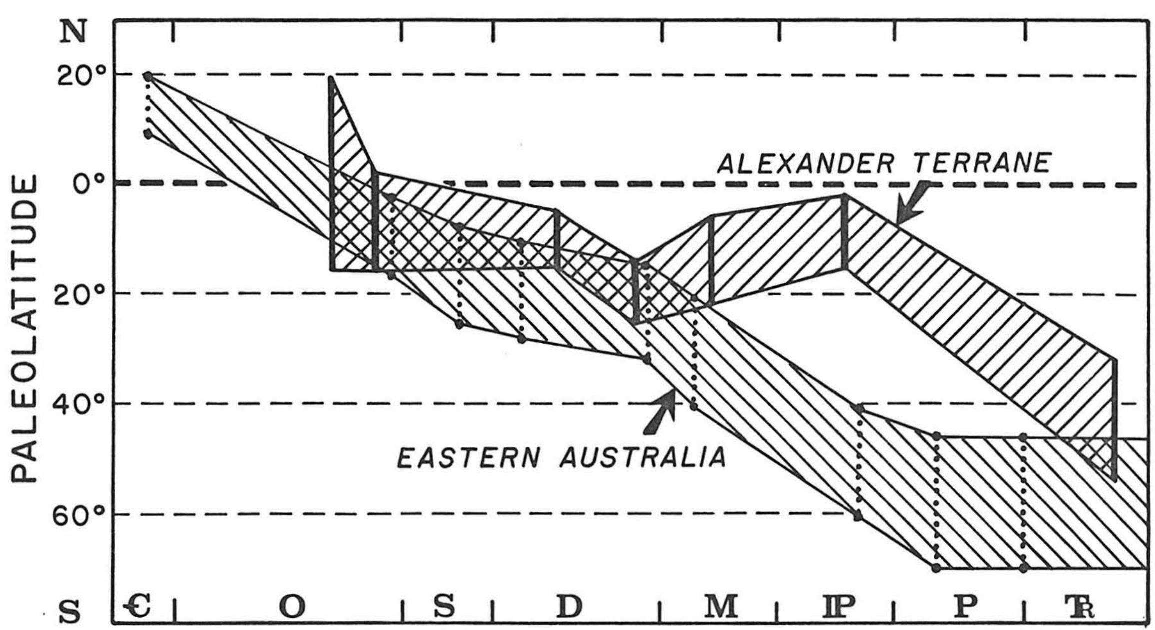
Based on these similarities I raise the possibilty that the Alexander terrane may have evolved from Cambrian (and perhaps Proterozoic) through Early Devonian time in a convergent margin environment along the paleo-Pacific margin of the Australia-Antarctica-New Zealand part of Gondwana. Although I do not draw direct correlations between the Alexander terrane and specific regions in the Lachlan orogen, similarities are strongest with rocks in the Molong arc in the eastern part of the fold belt.

Middle Devonian time marks the end of apparent similarities in the Alexander terrane and the Lachlan Belt, and, as described above, also marks a major change in the evolution of the Alexander terrane. Beginning in Middle Devonian time and continuing through the Carboniferous, the Alexander terrane evolved in a fairly stable marine environment, whereas the Lachlan Belt remained tectonically active. As suggested by Cas (1983), however, the Middle Devonian-Early Carboniferous tectonic activity in the Lachlan Belt may reflect an extensional event rather convergence along a plate margin. The available geologic evidence is therefore consistent with the interpretation that the terrane was tectonically separated from the Australia-Antarctica-New Zealand orogen during a rifting event which began in Middle-Late Devonian time.

Speculative displacement history

The hypothesis that the Alexander terrane formed and evolved along the paleo-Pacific margin of Australia and adjacent regions of Gondwana during early Paleozoic time is consistent with the paleomagnetic data of Van der Voo (1980) if a southern paleopole is selected. On Figure 4-4 I compare the apparent paleolatitudinal positions of eastern Australia (derived from continental reconstructions of Scotese, 1984) and the Alexander terrane (from the paleomagnetic data of Van der Voo and others, 1980) assuming that the Alexander terrane was in the southern hemisphere. Because the paleo-Pacific margin of Gondwana was oriented northwest-southeast during part of early Paleozoic time (Fig. 4-3), the data shown on Figure 4-4 indicate that the Alexander terrane could have been associated with only the northern part of the orogen -- perhaps near northeastern Australia or fragments now in Asia. The declination data of Van der Voo and others (1980) are also consistent with the hypothesis that the two regions were closely associated during early Paleozoic time. The paleomagnetic data from the Alexander terrane require approximately 100° of clockwise rotation (relative to the paleopole) since Pennsylvanian time, and show rapid, although not well constrained, clockwise rotation during Ordovician-Devonian time. As shown on the reconstructions of Scotese and others (1979) and Scotese (1984), eastern Australia also rotated clockwise during Ordovician-

Figure 4-4. Comparison of the apparent paleolatitudes of the Alexander terrane (assuming it was in the southern hemisphere) and of eastern Australia. Paleolatitudes and their uncertainties for the Alexander terrane are shown with solid bars (from Van der Voo, 1980), and have been connected to yield a generalized paleolatitudinal path for the Alexander terrane. The paleolatitudes for eastern Australia (dotted vertical lines) represent the latitudinal range of the eastern shoreline of Australia on paleocontinental reconstructions by Scotese (1984), and these vertical lines have been connected to yield an approximate paleolatitudinal path for eastern Australia. Note that the uncertainty in the paleolatitude of eastern Australia is not shown on this diagram, and that, according to Scotese (1984), the Late Devonian paleolatitude of eastern Australia is not well constrained.



Devonian time. The amount and sense of rotation recorded by the paleomagnetic data indicate that the interpreted Ordovician-Early Silurian arc in the Alexander terrane originally faced to the southeast, assuming that I am correct in my interpretation that it faces southwesterly today.

Figure 4-4 also shows that during Late Devonian-Early Mississippian time eastern Australia continued moving southward whereas the Alexander terrane began to move northward toward the paleoequator. The timing of this apparent divergence is consistent with my geologically derived interpretation that the Alexander terrane may have been rifted from the orogen in Middle Devonian-Early Carboniferous time. During late Carboniferous time the terrane was apparently just south of the equator and in the central paleo-Pacific basin according to both paleomagnetic and paleobiogeographic data (Van der Voo and others, 1980; Ross and Ross, 1983, 1985). I envision the terrane as an isolated microcontinental fragment in an intraplate oceanic environment during this time.

Geologic, palemagnetic, and paleontologic data from Triassic rocks suggest that the Alexander terrane was in the eastern part of the paleo-Pacific basin south of the paleoequator during Late Triassic time (Gehrels and Saleeby, 1985, in review c). Between Pennsylvanian and Late Triassic time the Alexander terrane therefore apparently remained south of the paleo-equator and migrated eastward across the paleo-Pacific basin. The declination data reported by Hillhouse and Grommé (1980) require the terrane to have undergone approximately 50° of clockwise rotation (relative to the paleopole) since Late Triassic time. I note that this is generally consistent with the 100° of post-

Pennsylvanian clockwise rotation required by the data of Van der Voo and others (1980), and suggests that part of the rotation occurred between Pennsylvanian and Late Triassic time, and part has occurred since Late Triassic time. A scenario for the post-Triassic northward displacement and accretion of the Alexander terrane is presented in Gehrels and Saleeby (1985, in review c).

SUMMARY

The Alexander terrane is a unique tectonic fragment in the North American Cordillera in that its geologic record is suprisingly long, complete, and well preserved. The Triassic and every geologic period of the Paleozoic are represented in the terrane by clastic strata, limestone, and volcanic rocks, except for volcanic rocks of Carboniferous age, and plutonic rocks were emplaced during every period except the Devonian and Mississippian. In most areas these rocks are only slightly to moderately deformed and are metamorphosed below greenschist facies. The available geologic, paleomagnetic, and paleontologic evidence suggest that the terrane was not accreted to its present position in the Cordillera until after Early Cretaceous time: hence, the rocks described herein have been displaced large distances from the regions in which they formed. Based on my geologic and geochronologic studies in southern SE Alaska and syntheses of the work of others to the north, I have attempted to 1) reconstruct the geologic and tectonic evolution of the Alexander terrane and 2) compare the evolution of the terrane to other regions in hopes of determining its displacement history.

Summary of the geologic and tectonic evolution of the Alexander terrane

The tectonic evolution of the terrane is subdivided into three distinct phases, including Cambrian (Proterozoic?) through Early Devonian time, Middle Devonian through Early Permian time, and Late Permian and Triassic time. The first phase records the origin and evolution of the terrane in a volcanic arc environment during Cambrian (and perhaps Proterozoic) time, and continued arc-type volcanism and plutonism in the southern part of the terrane until Early Silurian time. The arc activity was interrupted during the Middle Cambrian-Early Ordovician Wales "orogeny," and was terminated during the middle Silurian-earliest Devonian Klakas orogeny. These orogenic events apparently formed in a compressional regime, but it is not known whether they occurred in response to interplate or intraplate processes. Rocks of early Paleozoic age in the central and northern parts of the terrane form a thick section of limestone and clastic strata that have not experienced either of the early Paleozoic disturbances.

Rocks of Middle Devonian through Permian age were deposited in tectonically stable shallow-marine environments, and record the second phase in the evolution of the Alexander terrane. Volcanic rocks of Devonian and Permian age occur locally, but they do not appear to have formed in long-lived or regionally significant magmatic systems. Late Pennsylvanian-Early Permian juxtaposition of shallow-marine strata of the Craig subterrane against deeper-water strata of the Admiralty subterrane in central SE Alaska occurred during emplacement of large syenitic to dioritic intrusive bodies in the northern part of the terrane. This tectonic disturbance is apparently not associated with regional deformation or metamorphism.

The third stage in the evolution of the Alexander terrane is recorded by deposition of Triassic volcanic and sedimentary rocks in an interpreted rift environment and by a major unconformity at the base of the Triassic section. Occurrence of the Triassic strata along the eastern margin of the terrane in SE Alaska suggests that this interpreted rifting may record detachment of the terrane from a northeastern continuation of the terrane or from another tectonic fragment.

Summary of the displacement history of the Alexander terrane

Previous discussions of the displacement history of the Alexander terrane have centered around correlations with Paleozoic rocks in the Sierra-Klamath region of California. My analysis of the Paleozoic geologic record of the Alexander terrane suggests, however, that there are few similarities in the geologic and tectonic history of the two regions. Based on a review of the geology of other Paleozoic orogens, I raise the possibility that the Alexander terrane formed and evolved along the paleo-Pacific margin of Australia or adjacent fragments of Gondwana during early Paleozoic time. The existing paleomagnetic data and the paleobiogeographic affinities of several fossil groups from the Alexander terrane are apparently consistent with this hypothesis, although some lower Paleozoic fossils in the terrane apparently have stronger affinities with other parts of the Circum-Pacific region.

The available geologic, paleomagnetic, and paleontologic evidence from the terrane suggest that it separated from the Gondwana margin in Middle Devonian-Early Carboniferous time, perhaps during an interpreted rifting event, and migrated across the paleo-Pacific ocean basin south of the paleo-equator during late Paleozoic time. I envision the terrane to have been a microcontinental fragment in an intraplate oceanic environment during this time. By the end of Triassic time the terrane was apparently south of the paleoequator in the eastern part of the paleo-Pacific basin -- perhaps in association with rocks now found in Peru. As interpreted by Gehrels and Saleeby (1985, in review c), northward displacement of the terrane along the eastern margin of the paleo-Pacific basin apparently began during or soon after Late Triassic time and continued until mid-Cretaceous time when the terrane was juxtaposed against terranes previously accreted to the western margin of North America.

ACKNOWLEDGMENTS

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APPENDIX 1

GEOLOGIC MAP OF SOUTHERN PRINCE OF WALES ISLAND

INTRODUCTION

Southern Prince of Wales Island is underlain by surficial deposits and by stratified and intrusive rocks (and their metamorphosed and deformed equivalents) of Cretaceous through pre-Middle Ordovician age. These rocks have been studied during a geologic mapping project of the Dixon Entrance C-1, D-1, D-2, and parts of adjacent 1:63,360 quadrangles, and as part of regional geologic study of the Prince of Wales Island region. Southern Prince of Wales Island was mapped originally in reconnaissance fashion by Buddington and Chapin (1929) as part of a geologic study of southeastern Alaska. The region was mapped subsequently by W.H. Condon and I.L. Tailleur (unpublished U.S.G.S. report, 1960) based primarily on analyses of aerial photographs. More recently, MacKevett (1963) studied the geology of the Bokan Mountain-Stone Rock Bay area, Herreid and others (1978) and G. Donald Eberlein, Michael Churkin Jr., and Walter Vennum (unpublished data) mapped Kassa and Klakas Inlets and regions to the north, and Thompson and others (1982) and B. Collot (in Saint-André and others, 1983) conducted research on the Bokan Mountain Granite. The inset map of Plate 1 shows where this published and unpublished mapping has been incorporated into my geologic map (Plate 1). Also shown on this inset map are the locations of my foot traverses, skiff traverses, and helicopter landings.

The rocks on southern Prince of Wales Island have been subdivided on the geologic map (Plate 1) and in this text into map units based on guidelines recommended by the North American Commission on Stratigraphic Nomenclature (1983). The main categories of map units include:

SEDIMENTARY AND (OR) VOLCANIC ROCKS, which consist of stratified rocks, and in some areas their moderately deformed and (or) metamorphosed equivalents, that obey the law of superposition. These units include sedimentary and (or) volcanic rocks herein assigned to the Karheen Formation (Early Devonian) and to the Descon Formation (Early Silurian and Ordovician).

REGIONALLY METAMORPHOSED AND (OR) DEFORMED ROCKS, which consist of metasedimentary, metavolcanic, and (or) metaplutonic rocks that do not obey the law of superposition. These rocks include the Wales metamorphic suite (pmOw), and rocks herein assigned to the complex of Klakas Inlet (DSk), complex of Kendrick Bay (SOk), and complex of Ruth Bay (Or). The ages assigned to these units reflect the age of formation of the complex or suite rather than the protolith age of the rocks included.

INTRUSIVE ROCKS, which consist of plutonic and hypabyssal rocks that have been subdivided on the basis of their composition and emplacement age. Compositional terms follow recommendations by the International Union of Geological Sciences Subcommission on Systematics of Igneous Rocks (Streckeisen, 1976), except for the use of adamellite in place of their granite 3b. All radiometric ages have been calculated or recalculated with the decay constants and abundances recommended by the International Union of Geological Sciences Subcommission on Geochronology (Steiger and Jager, 1977). The Decade of North American Geology Time Scale (Palmer, 1983) has been used in relating these radiometric dates to the geologic time scale.

The rocks belonging to each map unit are described below in the Description of Map Units. Geologic units that have not been assigned to map units include surficial deposits and widespread basaltic to dacitic(?) dikes. Surficial deposits are described at the beginning of the Description of Map Units and the dikes are described at the beginning of the section on intrusive rocks. Major faults in the area are described in the following section on Structural Geology. Tables A1-1 and A1-2 outline the basic information about each of the paleontologic and geochronologic samples analyzed during this study. Discussion of the geologic and tectonic history of the region, tables with major, minor, and trace element geochemical data collected from samples of volcanic and intrusive rocks, and documentation of my U-Pb geochronologic data are presented by Gehrels and Saleeby (in review). The Early Devonian conodonts identified from samples collected during this study have been described by Savage and Gehrels (1984).

DESCRIPTION OF MAP UNITS

COLLUVIUM, ALLUVIUM, AND BEACH DEPOSITS (QUATERNARY) -- Surficial deposits consist of beach deposits along the shores of many bays, colluvium and soil in the interior of the island, and alluvium along some creeks. Bedrock is moderately to well exposed along most beaches: beach deposits are extensive only in the interior parts of some bays that have very gentle topography. These beach deposits are shown with a stipple pattern on Plate 1 where the bedrock is obscured. Surficial deposits are widespread in the interior of the island but are generally less than a few meters thick. These deposits have not been shown on Plate 1 because of their limited thickness and discontinuous distribution.

Table A1-1. Paleontologic samples and localities.

1 = Early Devonian (Pragian?) graptolites in black shale (localities 64ACn1171, 1172, 1182, 1183, and 1191 of Churkin and others, 1970)

2 = Early Devonian conodonts in limestone layer (N.M. Savage, verbal commun., 1985)

3 = latest Early Ordovician graptolites in shale (locality G-13 of Eberlein and others, 1983) and early Middle Ordovician conodonts in limestone lens in shale (Anita Harris, written commun., 1980) (both at station 72AE221)

4 = middle Early Devonian conodonts in limestone layer (locality 4 of Savage and Gehrels, 1984)

5 = middle Early Devonian conodonts in limestone layer (locality 5 of Savage and Gehrels, 1984)

6 = middle Early Devonian conodonts in limestone layer (locality 6 of Savage and Gehrels, 1984)

7 = Early Devonian conodonts in limestone layer (locality 7 of Savage and Gehrels, 1984)

8 = Early Devonian conodonts in limestone layer (locality 8 of Savage and Gehrels, 1984)

9 = middle Early Devonian conodonts in limestone layer (locality 9 of Savage and Gehrels, 1984)

10 = middle Early Devonian conodonts in limestone layer (locality 10 of Savage and Gehrels, 1984)

11 = middle Early Devonian conodonts in limestone float block (locality 11 of Savage and Gehrels, 1984)

12 = Middle Ordovician graptolites (locality G-14 of Eberlein and others, 1983) (station 74AE167)

13 = Middle Ordovician graptolites (locality G-14 of Eberlein and others, 1983) (station 74AE168)

Table A1-2. Geochronologic samples and localities.

1 = 472 ± 5 Ma (U-Pb zircon): Middle Ordovician quartz-porphyritic granodiorite (Oqgd) (station 82GP702)

2 = 462 ± 15 Ma (U-Pb zircon): Middle Ordovician diorite (Od) (station 72AE215)

3 = 468 ± ~15 Ma (U-Pb zircon): Middle Ordovician leucogranodiorite (Ogd) (station 82GP48)

4 = ~480-460 Ma (U-Pb zircon): Ordovician diorite (Od) (station 82GP40)

5 = 465 ± 7 Ma (U-Pb zircon): Middle Ordovician foliated granodiorite (Or) (station 79AE114)

6 = 445 ± 5 Ma (U-Pb zircon): Late Ordovician quartz diorite (Oqd) (station 82GP28)

7 = 446 ± 5 Ma (U-Pb zircon): foliated and layered Ordovician quartz diorite (Ofd) (station 82GP346)

8 = 438 ± 4 Ma (U-Pb zircon): Early Silurian and (or) Late Ordovician quartz monzonite (SOqm) (station 83GP255)

9 = 438 ± 5 Ma (U-Pb zircon): Early Silurian and (or) Late Ordovician granite (SOgr) (station 83GP335)

10 = 438 ± 5 Ma (U-Pb zircon): Early Silurian and (or) Late Ordovician quartz syenite (SOsy) (station 83GP364)

11 = 418 ± 5 Ma (U-Pb zircon): Late Silurian leucodiorite (Sd) (station 82GP626)

12 = 428 ± 13 Ma (K-Ar hornbl. minimum age): locality 5 of Turner and others (1977) (station number DT72-58c)

13* = 454 ± 22 Ma (K-Ar hornbl.): locality 7 of Lanphere and others (1964)

14* = 379 ± 18 Ma (K-Ar biot.): locality 6 of Lanphere and others 1964)

15* = 439 ± 21 Ma (K-Ar hornbl.): locality 5 of Lanphere and others (1964)

16* = 185 ± 8 Ma (K-Ar riebekite): locality 1 of Lanphere and others (1964), and 190 ± 8 Ma (K-Ar riebekite): locality 4 of Lanphere and others (1964)

17* = Bokan Mountain Granite (U-Pb): Saint-Andre and others (1983)

* = sample location on Plate 1 is approximate.
[See Gehrels and Saleeby (in review) for more information.]

Sedimentary and (or) volcanic rocks

KARHEEN FORMATION (EARLY DEVONIAN) -- Strata assigned herein to the Karheen Formation constitute a fining-upward sequence of clastic sedimentary and subordinate volcanic rocks. These strata unconformably overlie Early Silurian and Ordovician rocks, and are the youngest sedimentary and volcanic rocks in the map area. The stratigraphic sequence generally ranges from coarse conglomerate (Dcg) or breccia (Dbx) at the base, through sandstone (Ds), mudstone and siltstone (Dms), and limestone (Dls) in the middle, into laminated black graptolitic shale (Dsh) at the top. Plagioclase-porphyritic and basaltic to andesitic volcanic rocks (Dpv and Dbv) are interlayered with the coarser clastic strata in the southern part of the area.

A middle Early Devonian age for rocks in upper parts of the section is indicated by conodonts in limestone layers (Savage and Gehrels, 1984) and graptolites in black shale (Churkin and others, 1970). Coarser clastic rocks in the lower part of the section grade into these strata and are interpreted to be Early Devonian in age (Gehrels and Saleeby, in review). Similarities in lithic types, stratigraphic position, and depositional age indicate that these strata are correlative with the Early Devonian Karheen Formation on west-central Prince of Wales Island (Eberlein and Churkin, 1970; Eberlein and others, 1983; Gehrels and Saleeby, in review). The thickness of the section in the area averages approximately 2 km but is quite variable.

$\boxed{\text{Dsh}}$ SHALE (EARLY DEVONIAN) -- Dark-gray to black shale and subordinate slate and slaty argillite with mm-scale laminations and slight cm-scale compositional layering. These strata are distinguished from Early Silurian(?) and Ordovician argillite and shale (SOa) by the more siliceous composition of the older strata and greater fissility of the Devonian rocks. Layers up to several tens of centimeters in thickness of gray mudstone, brown carbonate-rich siltstone, and leucogranodiorite-clast conglomerate locally occur in the section. Fine laminations and a slight fining upwards in individual shale layers record transportation by turbidite flows and deposition in a submarine fan system. The presence of conglomeratic horizons and a large slide(?) block of leucogranodiorite in the section indicate that there was significant topographic relief within or adjacent to the basinal system.

In most areas the shale grades downsection into tan to gray mudstone and siltstone (Dms) over a stratigraphic distance of several meters: the contact is drawn where shale dominates over mudstone and siltstone. The stratigraphic thickness is known to be at least 250 m along the east shore of Klakas Inlet but could be considerably greater along the west shore. North of the study area, the shale and underlying mudstone and siltstone (Dms) have been mapped together as a section of dark-gray siltstone (about 600 m) and dark-gray argillite (about 1200 m) by Herreid and others (1978).

The age of strata in this unit is indicated by several occurrences of graptolites of probable Pragian age (paleontologic sample localities labeled "1": Churkin and others, 1970). Morphologically similar graptolites were recognized at the localities indicated with a "G" on Plate 1, but these have not been formally identified as the same fauna described by Churkin and others (1970).

$\boxed{\text{Dms}}$ MUDSTONE AND SILTSTONE (EARLY DEVONIAN) -- Tan, dark- to light-gray, and bluish-gray mudstone and siltstone in beds from 2 to 20 cm in thickness. Individual beds commonly show a slight size gradation from silty to more shaly detritus upsection, and low-angle cross-stratification is common. These characteristics and the general stratigraphic relations of this unit indicate transportation of detritus by turbidity currents and deposition in distal marine environments -- probably in a submarine fan. Orange- to brown-weathering beds up to 10 cm in thickness constitute about 10% of the section and consist of sandy siltstone with a carbonate-rich matrix. These beds commonly weather more rapidly than the adjacent mudstone and siltstone, forming recessed horizons. In some layers the sandy siltstone changes along strike from this orange-brown, deeply recessed character to a gray siltstone with weathering characteristics similar to that in the adjacent mudstone and siltstone. Cobbles and lenses of sandy siltstone that occur within the mudstone and siltstone also show this style of weathering. The difference in weathering characteristics is interpreted to be the result of a greater amount of carbonate matrix material in the sandy siltstone layers.

These strata grade upsection into black shale (Dsh) over a distance of several meters: the contact is drawn where shale dominates over mudstone and siltstone. Downsection the strata grade into coarser-grained siltstone and locally into sandstone. The basal contact is drawn where mudstone becomes subordinate to sandstone. In Max Cove and west of Klakas Inlet the underlying Devonian sandstone (Ds) is too thin to distinguish on the map and is included in this unit. The thickness of the mudstone and siltstone is approximately 400 m between Max Cove and Klakas Inlet: to the south and on the west side of Klakas Inlet the thickness may be considerably greater. North of the study area these strata and the overlying shale (Dsh) are included in an argillite and siltstone map unit consisting of about 600 m of dark-gray siltstone overlain by about 1200 m of dark-gray argillite (Herreid and others, 1978). On the unnamed island west of Klinkwan in southern Klakas Inlet, nondeformed mudstone, siltstone, and shale overlies a thin layer of cross-bedded sandstone which overlies moderately deformed Devonian sedimentary breccia (Dbx). This relationship places an important constraint on the age of movement on the Anchor Island thrust fault and deformation in the complex of Klakas Inlet (DSk), as described below.

The age of strata in this unit is constrained by graptolites of probable Pragian (middle Early Devonian) age in overlying shale (Dsh), and by middle Early Devonian conodonts in limestone (Dls) lower in the section.

Ds SANDSTONE (EARLY DEVONIAN) -- Tan- to reddish-brown-weathering sandstone, siltstone, and subordinate mudstone and pebbly conglomerate interbedded locally with volcanic rocks (Dpv and Dbv) and fossiliferous limestone (Dls). The sandstone varies from massive beds several meters in thickness to thin-bedded sandstone and siltstone with ripple marks and high-angle cross beds and channels. Plagioclase-porphyritic volcanic rocks (Dpv) are widespread south of Hunter Bay. Some volcanic rocks, particularly along the southern shoreline of the island, have not been mapped separately from the clastic strata. North of Tah Bay, detritus in these strata consists predominantly of monocrystalline quartz and plagioclase with subordinate lithic fragments of fine-grained quartz and plagioclase (leucogranodiorite?) and of a cryptocrystalline rock with trachytic feldspar microphenocrysts (volcanic rock?). To the south the lithic grains and their matrix are more highly altered with secondary carbonate and clay minerals. Lithic fragments from southern areas consist of feldspar phenocrysts in an altered cryptocrystalline matrix (hypabyssal or volcanic rock?) or highly altered cryptocrystalline rock of intermediate to basic composition (volcanic rocks?). Monocrystalline quartz and plagioclase grains are subordinate. The dominant source for the detritus in this unit is interpreted to have varied from Early Silurian and Ordovician intrusive and stratified rocks in the northern part of the area to intraformational volcanic rocks along the southern shoreline of the island.

In most areas this unit contains or is interbedded with a thin horizon of fossiliferous limestone (Dls). Associated with the limestone in a few areas are sections of interbedded maroon and green shale and chert-pebble conglomerate. Megafossils in the limestone, the presence of interbedded maroon and green shale, and sedimentary structures in the sandstone all suggest that these strata were deposited in shallow marine to fluvial environments.

Strata in this unit generally coarsen downsection, grading through pebbly sandstone into cobble and boulder conglomerate. The basal contact is drawn where conglomerate becomes the dominant lithology. Upsection siltstone predominates, and the contact with overlying mudstone and siltstone (Dms) is drawn where mudstone dominates over siltstone. The thickness of this unit varies from over a kilometer at the south end of the island to several meters or less in the Max Cove region (Gehrels and Saleeby, in review). Much of this thinning apparently occurs near Klinkwan Cove.

The age of the sandstone is well constrained by the occurrence of interbedded limestone (Dls) with middle Early Devonian conodonts, and graptolites of probable Pragian (middle Early Devonian) age in overlying shale (Dsh).

| Dls | LIMESTONE (EARLY DEVONIAN) -- Light- to medium-gray fossiliferous limestone breccia and subordinate massive to thin-bedded limestone. Thin layers of darker gray siliceous limestone are common in both the breccia and the bedded limestone. North of Tah Bay the limestone forms a single bed 1 to 3 m in thickness. In Max Cove it is underlain by, and locally interbedded with, chert-pebble conglomerate with cm-scale rounded clasts of black to dark-gray chert. Strata above and below the limestone in Max Cove consist of sandstone and siltstone. South of Tah Bay there are commonly two layers of limestone up to several meters in thickness which are interbedded with and separated by sandstone and siltstone. Limestone in this unit apparently belongs to a single horizon that extends continuously throughout the study area: limestone has not been recognized in other parts of the Early Devonian section. In contrast, Herreid and others (1978) report that similar limestone just north of the study area is laterally discontinuous.

Conodonts recovered from limestone at several localities (paleontologic samples 4 through 11) demonstrate a middle Early Devonian age for the limestone throughout the area (Savage and Gehrels, 1984) and may record a slight younging of the section to the south (Gehrels and Saleeby, in review). Megafossils recovered by Theodore Chapin and identified by Edwin Kirk (cited in Buddington and Chapin, 1929) were assigned a Middle Devonian age. These fauna include:

Cladopora sp. and crinoid columnals from a locality in Hessa Inlet (which may be the same as paleontologic sample locality 9 in Buschman

Pass),

Atrypa reticularis, Favosites hemisphericus, Alveolites sp., Cyathophyllum sp., and Syringopora sp. from paleontologic sample locality 7 in Hunter Bay,

Favosites hemisphericus and Alveolites sp. from paleontologic sample locality 5 in Max Cove.

Based on the diagnostic conodont fauna and the probable Pragian (middle Early Devonian) graptolites in overlying shale (Churkin and others, 1970), the limestone is assigned a middle Early Devonian age.

Dpv PLAGIOCLASE-PORPHYRITIC VOLCANIC ROCKS (EARLY DEVONIAN) -- Volcanic and subordinate hypabyssal rocks that consist of tan, reddish, and gray plagioclase-porphyritic dacite(?). These rocks occur as porphyritic tuff, tuff breccia, and flows interbedded with Devonian conglomerate (Dcg) and sandstone (Ds), and as hypabyssal dikes, sills, and small intrusive bodies. Dominant minerals include subhedral altered grains of andesine in a matrix of altered plagioclase microlites, opaque minerals, calcite, iron oxides, and other secondary minerals. Small, highly altered ferromagnesian grains with cores of pigeonite are preserved in a few samples. The age of these rocks is constrained by middle Early Devonian fossils in adjacent clastic strata. Hypabyssal plagioclase-porphyritic diorite (Dph) bodies in the Klakas Inlet region and north of the study area (Herreid and others, 1978) are interpreted to be subvolcanic to these rocks based on their similar mineralogy and texture, and the close spatial association of the hypbyssal bodies and Devonian strata.

Dbv BASALTIC TO ANDESITIC ROCKS (EARLY DEVONIAN) -- Basaltic to andesitic rocks that occur as flows, breccia, dikes, and small intrusive bodies in the Tah Bay region and as a thin pillow flow just west of Bert Millar Cutoff. In thin section these rocks consist of highly altered, millimeter-scale plagioclase phenocrysts in a matrix of very fine grained opaque minerals, chlorite, calcite, epidote, and other secondary minerals. These rocks resemble andesitic breccia and flows that overlie the Early Devonian clastic strata north of the area (Herreid and others, 1978; Gehrels and Saleeby, in review). Their age in the study area is constrained by middle Early Devonian fossils in adjacent clastic strata.

$\boxed{\text{Dcg}}$ CONGLOMERATIC STRATA (EARLY DEVONIAN) -- Tan- to reddish-brown-weathering pebble, cobble, and boulder conglomerate, conglomeratic sandstone, and subordinate sandstone, siltstone, and undifferentiated volcanic rocks. The stratigraphic section varies considerably in thickness and in clast composition from north to south. On the west shore of Klakas Inlet and in Max Cove the conglomerate is several meters thick and consists of well-rounded cobbles of intrusive, volcanic, and hypabyssal rocks in a sandstone matrix. In both areas the conglomerate is massive and overlies brecciated leucogranodiorite (Ogd). The basal contact can be located to within a meter at its northern extent in Max Cove, but is not actually exposed. In the Klinkwan Cove region the conglomerate thickens rapidly southward, reaching a thickness of over 500 m on the northeast side of the Klinkwan Cove fault. Clasts in this area are well rounded and consist of approximately equal proportions of intrusive and volcanic rocks. High-angle cross-beds and channels and the presence of clasts over a meter in diameter suggest that these strata were deposited in subaerial to shallow marine environments within or adjacent to an area with significant topographic relief.

South of Hunter Bay, volcanic clasts become predominant, the section increases in thickness to over 1500 m, and plagioclase-porphyritic volcanic rocks (Dpv) constitute a locally significant part of the section. The composition of the volcanic clasts changes southward also, from aphyric basalt-andesite to the north, to plagioclase-porphyritic andesite-dacite to the south. The volcanic and plutonic clasts to the north were apparently derived from underlying Early Silurian and Ordovician rocks, whereas the porphyritic clasts to the south were derived from intraformational volcanic rocks (Dpv and Dbv). The basal contact of conglomerate overlying Early Silurian(?) and Ordovician rocks is exposed just east of Seagull Island (southern Buschmann Pass), along the west shore of Brownson Bay, and near Bert Millar Cutoff. In most areas the conglomerate grades upsection into sandstone: the contact is drawn where conglomerate dominates over sandstone.

The minimum age of the conglomeratic strata is constrained by middle Early Devonian fossils in limestone (Dls) which locally grades into strata in the upper part of this unit.

Dbx SEDIMENTARY BRECCIA (EARLY DEVONIAN) -- The basal Devonian unit in southern Klakas Inlet is a sedimentary breccia consisting of nonsorted angular clasts (up to 50 cm across) of brecciated leucogranodiorite (Ogd) and diorite (Od), and highly deformed volcanic, sedimentary, and intrusive rocks derived from the subjacent complex of Klakas Inlet (DSk). The breccia is moderately flattened, brecciated, and locally semischistose, but is not as strongly deformed or altered as rocks in the underlying structural complex. The basal contact of the breccia is exposed in several areas on these islands, but the degree of deformation in both overlying and underlying rocks makes its recognition difficult. The breccia is locally overlain unconformably by up to a meter of cross-bedded sandstone, which is in turn overlain by Devonian mudstone, siltstone, and shale (Dms). These strata are only slightly deformed, which demonstrates that deformation in the breccia and the underlying structural complex occurred prior to middle Early Devonian time. The breccia is interpreted to be Early Devonian in age because it grades laterally into conglomerate (Dcg) which locally contains limestone (Dls) with middle Early Devonian fossils.

The stratigraphic and structural relations described above suggest that deformation in the southern Klakas Inlet region continued into Early Devonian time, but ceased prior to the deposition of middle Early Devonian sandstone, siltstone, mudstone, and shale. Because deformation in this zone is attributed to movement on the Anchor Island thrust system (as described in the Structural Geology section) I interpret the breccia to have been deposited and deformed as a talus breccia within this active system. These relations indicate that movement on the Anchor Island fault system continued into earliest Devonian time.

DESCON FORMATION (EARLY SILURIAN AND ORDOVICIAN) -- Strata herein mapped as the Descon Formation include volcanic and sedimentary rocks that are similar in lithic types, stratigraphic position, and age to Early Silurian and Ordovician rocks in the Descon Formation on central and northern Prince of Wales Island (Eberlein and Churkin, 1970; Eberlein and others, 1983; Gehrels and Saleeby, in review). Volcanic rocks dominate in most of the study area and consist of basaltic to andesitic pillow flows, pillow breccia, and tuff breccia (SObv) with subordinate intermediate-composition and silicic tuff and tuff breccia (SOiv and SOsv). These rocks locally interfinger with marine clastic strata south of the Frederick Cove fault: north of the fault the strata consist primarily of interbedded argillite and shale (SOa), banded mudstone and siltstone (SOms), and graywacke (SOgw). Limestone (SOls) is a minor component of the Descon Formation in the study area.

Sedimentary and volcanic rocks of the Descon Formation are overlain unconformably by Early Devonian clastic strata of the Karheen Formation, and are in fault contact with pre-Middle Ordovician rocks of the Wales metamorphic suite. The age of the clastic strata south of the Frederick Cove fault is indicated by early Middle Ordovician conodonts (Anita Harris, written commun., 1980) and latest Early Ordovician graptolites (Eberlein and others, 1983) recovered from the east shore of Klakas Inlet. Middle Ordovician graptolites have also been recovered from argillite interbedded with basaltic to andesitic volcanic rocks in Moira Sound (Eberlein and others, 1983). Volcanic rocks south of the Klakas Inlet-Moira Sound region are interpreted to be Ordovician and perhaps locally Early Silurian in age based on correlation with rocks of known Ordovician age, and intrusive relations which suggest that they are coeval with Ordovician plutonic rocks.

SOa ARGILLITE AND SHALE (EARLY SILURIAN AND ORDOVICIAN) -- Black, well-bedded siliceous argillite and dark-gray to black shale. North of the Frederick Cove thrust fault, black shale and argillite with millimeter-scale laminations and 1- to 3-cm-scale layering are the dominant lithic types. Interbedded with these strata are layers of dark-gray mudstone and siltstone, and several beds up to a meter in thickness of gray laminated limestone. Along strike to the northwest the argillite and shale are interbedded with dark-gray and brownish-gray banded mudstone and siltstone (SOms). The contact between the two units is drawn where shale and argillite dominate over mudstone. The rhythmically bedded nature of these strata, combined with graded beds in adjacent mudstone and siltstone, indicate that they were deposited in a basin plain to distal submarine fan system environment. Sedimentary structures indicate that these strata face to the southwest.

South of the Frederick Cove fault, siliceous argillite is interbedded with volcanic rocks of the Descon Formation in many areas. Between Klakas Inlet and West Arm (Moira Sound), thick sections of argillite and shale are interlayered with basaltic to andesitic and minor silicic volcanic rocks. Thin beds of siliceous (locally cherty) argillite are common in the volcanic rocks in Winter Bay, Johnson Cove, South Arm (Moira Sound), and in the Barrier Islands. In the Barrier Islands and in Winter Bay, these beds commonly contain disseminated pyrite and are stained with red- and orange-weathering iron oxides (Gehrels and others, 1983a)

The age of the fine-grained clastic strata south of the Frederick Cove fault is known from graptolites of latest Early Ordovician age (paleontologic sample locality 3) and Middle Ordovician age (localities 12 and 13) (Eberlein and others, 1983). Early Middle Ordovician conodonts have also been recovered from a limestone lens at locality 3 (Anita Harris, written commun., 1980). North of the Frederick Cove fault the strata are interbedded with limestone layers that have been sampled for conodonts but none have been recovered (N.M. Savage, written commun., 1985). These strata are interpreted to belong to the Descon Formation based on similarities in lithic types and their occurrence beneath strata of known Devonian age (Herreid and others, 1978; Gehrels and Saleeby, in review).

SOms BANDED MUDSTONE AND SILTSTONE (EARLY SILURIAN AND ORDOVICIAN) -- Rhythmically bedded gray, greenish gray, light-green, and locally tan mudstone and siltstone turbidites with well-developed size grading. Graywacke and argillite also occur in this unit in some areas. Individual beds generally range from 2 to 6 cm in thickness and are finely laminated and laterally continuous. These stratigraphic characteristics are typical in both the thick sections of mudstone and siltstone north of the Frederick Cove fault, and also in thin layers interbedded with other clastic strata or volcanic rocks to the south. The greenish color, strong compositional layering, and lack of tan-weathering carbonate-rich layers distinguish these strata from Devonian mudstone and siltstone (Dms). Size grading, low-angle cross stratification, and the rhythmically bedded nature of the strata record deposition by turbidity currents in a submarine fan system.

These strata are interlayered with and grade into argillite and shale (SOa) north of West Arm (Moira Sound) and along the east shore of Klakas Inlet. The contact is drawn where banded mudstone and siltstone become the dominant lithology. In Hunter Bay and south of Nichols Bay, mudstone and siltstone are interbedded with layers of light-colored limestone (SOls) in a section of predominantly basaltic to andesitic volcanic rocks. At the north end of Nichols Bay, mudstone is interbedded with gray, carbonate-rich mudstone (mapped as SOls on Plate 1) and silicic volcanic rocks (SOsv). Thin sulfide-rich layers occur in the mudstone and siltstone and in the adjacent volcanic rocks in Nichols Bay (Gehrels and others, 1983a). The strata in Hunter Bay and to the south are moderately deformed and are metamorphosed to low greenschist facies. Metamorphism and deformation of these rocks occurred prior to deposition of the superjacent Early Devonian strata. In most regions these strata display meter-scale open folds with shallow-plunging, northwest-trending axes.

The age of strata in this unit south of the Frederick Cove fault is constrained by latest Early and Middle Ordovician fossils in interbedded argillite (SOa). To the north the strata are interpreted to be correlative with the Descon Formation based on similarities in lithic types and stratigraphic position: they are overlain unconformably by Early Devonian strata (Herreid and others, 1978), and have not experienced the Early Ordovician-Middle Cambrian metamorphism and deformation of rocks in the Wales metamorphic suite (pmOw).

SOgw GRAYWACKE (EARLY SILURIAN AND ORDOVICIAN) -- Gray and greenish-gray graywacke consisting of silt-, sand-, and pebble-size grains of volcanic lithic fragments and subordinate feldspar. Individual beds range from several centimeters to over a meter in thickness and commonly are size graded. These strata are interbedded with and grade into banded mudstone, siltstone, argillite, and shale (SOms and SOa) north of the Frederick Cove fault, and are interbedded with basaltic to andesitic volcanic rocks of the Descon Formation in Moira Sound. The age of the graywacke along South Arm (Moira Sound) is indicated by Middle Ordovician graptolites in argillite (SOa) interbedded with volcanic rocks (paleontologic sample localities 12 and 13: Eberlein and others, 1983). North of the study area, massive to thick-bedded graywacke constitutes a major component of the Descon Formation (Herreid and others, 1978).

SOls LIMESTONE (EARLY SILURIAN AND ORDOVICIAN) -- Limestone is a subordinate component of the Descon Formation in the study area and is quite variable in character. Along the north shore of West Arm (Moira Sound), limestone occurs as light- to dark-gray beds up to a meter in thickness of cross-laminated sandy limestone in a section of shale, argillite, and subordinate mudstone and siltstone. Sedimentary structures in these beds and in adjacent clastic strata suggest that they were deposited by turbidity flows. Limestone along the south shore of West Arm (Moira Sound) and the east shore of Klakas Inlet consists of thin layers of massive, dark-gray limestone interbedded with siliceous black argillite. In Hunter Bay and southern Nichols Bay, massive, light-colored layers of recrystallized limestone up to several meters in thickness are interlayered with volcanic and subordinate volcaniclastic rocks. Limestone along the north shore of Nichols Bay consists of gray calcareous mudstone interbedded with moderately deformed and metamorphosed mudstone and siltstone (SOms) and silicic volcanic rocks (SOsv). Dark-gray limestone along the east shore of Klakas Inlet yields Middle Ordovician conodonts (Anita Harris, written commun., 1980). Samples of limestone from West Arm (Moira Sound), Hunter Bay, and southern Nichols Bay have not yielded conodonts (N.M. Savage, written commun., 1985): the assigned age of these strata is based on the interpreted age of the adjacent sedimentary and volcanic rocks.

SObv BASALTIC TO ANDESITIC VOLCANIC ROCKS (EARLY SILURIAN AND ORDOVICIAN) -- Dark- to light-green basaltic to andesitic pillow flows, pillow breccia, and tuff breccia with subordinate tuff, volcaniclastic strata, and volcanic rocks of more silicic composition. Pillow flows (P on Plate 1) display variably deformed pillows that average 30 cm across (ranging from 10 cm to as much as 1 m). Interpillow carbonate and chert are quite rare. Of approximately equal abundance is volcanic breccia locally distinguished as pillow breccia (PB) or tuff breccia (TB) on Plate 1. Bedding in the breccia is well developed and is commonly parallel to a slight flattening of protolith features. Thin layers of tuff (T on Plate 1), volcaniclastic graywacke (SOgw), and banded mudstone and siltstone (SOms) are interbedded with the volcanic breccia throughout the area. Also present are layers of silicic volcanic rock (SOsv) and siliceous argillite (SOa) that contain disseminated and locally massive sulfides.

Rocks in this unit generally consist of spilitic basalt to andesite with highly altered plagioclase phenocrysts. The plagioclase grains are of andesine composition, have variable albitic and sericitic alteration, and range up to 4 mm in length. Millimeter-scale clots of chlorite, epidote, and opaque minerals are probably alteration products of ferromagnesian phenocrysts. Matrix minerals consist of plagioclase microlites, chlorite, epidote, leucoxene, and opaque minerals that formed during lower greenschist- to sub-greenschist-facies metamorphism. Adjacent to the Early Silurian and Ordovician intrusive bodies, the rocks are locally metamorphosed to biotite- or hornblende-bearing hornfels. Deformation of the volcanic rocks is quite variable: in many areas protolith features such as pillows and breccia fragments are only slightly flattened, the rocks do not have a foliation, and plagioclase microlites in the matrix are randomly oriented. In other areas, particularly along the trace of the Anchor Island fault, the volcanic rocks are brecciated and semischistose, and protolith features have been obscured.

Features of these rocks that distinguish them from metavolcanic rocks of the Wales metamorphic suite (pmOw) include: 1) the lack of both schistosity and outcrop-scale, asymmetric folds 2) the abundance of well-preserved protolith features such as pillows, breccia fragments, etc., and 3) the tendancy for pillows and breccia fragments to weather differentially from their matrix.

The depositional age of basaltic to andesitic volcanic rocks in the Klakas Inlet-Moira Sound region is indicated by latest Early and Middle Ordovician fossils recovered from interbedded argillite (SOa). A minimum age in other parts of the map area is derived from cross-cutting Early Silurian and (or) Late Ordovician quartz monzonite and adamellite (SOqm). Dioritic and quartz dioritic intrusive rocks of Middle and Late Ordovician age (Od and Oqd) are interpreted to be the intrusive equivalents of these volcanic rocks.

SOiv INTERMEDIATE COMPOSITION (ANDESITIC TO DACITIC) VOLCANIC ROCKS (EARLY SILURIAN AND ORDOVICIAN) -- Gray, greenish-gray, greenish-white, and tan plagioclase- and quartz-porphyritic tuff and subordinate tuff breccia. Banded mudstone and siltstone (SOms), volcanic rocks of basaltic to andesitic and more silicic composition, and hypabyssal bodies of intermediate composition are also common in this unit. Widespread cm-scale layering in the tuff is locally due to flow banding, but is more commonly due to a change in grain size and a slight change in composition within the individual layers. Compositionally these volcanic rocks probably fall within the andesite to dacite range. They have been distinguished from the generally coeval basaltic to andesitic (SObv) volcanic rocks on the basis of their more silicic composition, abundant plagioclase and quartz phenocrysts, and characteristic cm-scale layering. The silicic volcanic rocks (SOsv) are more silicic and are commonly fragmental. Volcanic rocks of intermediate composition constitute a major component of the Descon Formation in the northeastern part of the study area, and occur as thin layers in the volcanic rocks in other parts of the area.

Most rocks belonging to this unit are plagioclase- and quartz-porphyritic keratophyre. Plagioclase phenocrysts constitute 5% to 15% of the rock and commonly occur as glomerocrysts of large plagioclase and smaller opaque crystals. Individual plagioclase grains range up to 6 mm in length and are generally tabular, unzoned, albitized, and moderately sericitized. Most grains were originally of andesine composition. Quartz phenocrysts are subordinate to plagioclase in abundance, and occur as sub-rounded grains up to 5 mm in diameter. Anhedral opaque grains up to 3 mm long constitute up to several percent of the rock. Matrix minerals consist of microcrystalline quartz, plagioclase, and alteration products including chlorite, leucoxene, epidote, opaque minerals, and calcite.

These rocks are interbedded with clastic strata in Moira Sound from which Eberlein and others (1983) have recovered Middle Ordovician graptolites. Cross-cutting intrusive bodies of Early Silurian and (or) Late Ordovician quartz monzonite and adamellite (SOqm) indicate a minimum depositional age of Early Silurian. Based on compositional similarities, these volcanic rocks may be correlative with the breccia of Luck Creek on central Prince of Wales Island (Eberlein and others, 1983), and may be genetically related to the leucogranodioritic (Ogd and Oqgd) and quartz dioritic intrusives (Oqd) in the study area.

$\boxed{\text{SOsv}}$ SILICIC (DACITIC TO RHYOLITIC) VOLCANIC ROCKS (EARLY SILURIAN AND ORDOVICIAN) -- Light-green, light-gray, tan, and white plagioclase-porphyritic silicic tuff, tuff breccia, and breccia.

Tuffaceous strata commonly have mm-scale laminations formed by slight changes in composition and probably grain size. Tuff breccia occurs both as m-scale layers throughout the Descon volcanics in the area, and as km-thick layers in the northeastern part of the area. Clasts in the breccia range from several cm to 50 cm in length and consist of massive to finely laminated silicic and subordinate intermediate-composition volcanic rock. The clasts are supported in a matrix of laminated tuff and locally argillaceous strata. Silicic breccia also occurs in bodies that map out as lenses up to several hundred meters in length in both the Barrier Islands and the Ingraham Bay region. These breccias consist entirely of 10- to 50-cm-scale angular blocks of laminated silicic volcanic rock and are interpreted to have been extrusive domes. Layers of intermediate-composition (SOiv) and basaltic to andesitic (SObv) volcanic rocks are common in this map unit, as are sedimentary interbeds of argillite and shale (SOa) and banded mudstone and siltstone (SOms).

Sulfide-rich horizons are common in these silicic volcanic rocks and in adjacent sedimentary and volcanic strata (Gehrels and others, 1983a). In the Barrier Islands, several thin sulfide-rich layers occur at or near contacts between silicic breccia and either basaltic to andesitic volcanic rocks or argillaceous strata. The sulfides tend to occur in thin rinds around pillows or breccia fragments or in thin layers in the argillaceous strata. These relations and an occurrence in Nichols Bay where massive sulfide minerals show sedimentary layering and lamination similar to that in the banded siltstone host-rock indicate that the mineralization was syn-volcanic and presumably volcanogenic.

Most rocks in this unit consist of plagioclase-porphyritic quartz keratophyre. Plagioclase phenocrysts range up to 8 mm in length, are andesine in composition, and commonly occur in glomerocrysts with smaller opaque grains. The phenocrysts are generally tabular in shape and nonzoned, and have thin albitic rims and variable secondary alteration. Phenocrysts constitute up to 20% of some rocks but are more commonly less than 10%. Matrix minerals include primary microcrystalline quartz and albitized plagioclase, and secondary chlorite, epidote, leucoxene, calcite, sericite, and opaque minerals. Although these volcanic rocks are more silicic and leucocratic than the intermediate-composition volcanic rocks (SOiv) with which they are interbedded, quartz phenocrysts have not been recognized in the more silicic rocks.

The silicic volcanic rocks are regionally interbedded with strata in Moira Sound and in Klakas Inlet from which Eberlein and others (1983) have recovered latest Early and Middle Ordovician fossils. Cross-cutting intrusive bodies of Early Silurian and (or) Late Ordovician quartz monzonite and adamellite (SOqm) indicate a minimum depositional age of Early Silurian for these rocks. The silicic volcanic rocks in the Barrier Islands and along Klakas Inlet are interbedded with basaltic to andesitic volcanic rocks (SObv) of probable Ordovician age, and may be genetically related to the Ordovician quartz-porphyritic granodiorite (Oqgd) and leucogranodiorite (Ogd) in the area.

Regionally metamorphosed and (or) deformed rocks

DSk	COMPLEX OF KLAKAS INLET (EARLY DEVONIAN AND SILURIAN) --

Highly altered and brecciated intermediate-composition semischist, greenstone, silicic semischist, and leucogranodiorite derived from stratified and intrusive rocks of Early Silurian(?) and Ordovician age. This complex occurs in the southern Klakas Inlet region and is referred to informally herein as the complex of Klakas Inlet. Much of the unit consists of orange- to gray-weathering, green to tan semischist surrounding cm-scale elongate blocks of fractured greenstone. Relict volcanic or sedimentary structures and textures in some blocks indicate protoliths of basaltic to andesitic volcanic rock (SObv), mudstone and siltstone (SOms), and graywacke (SOgw). Protoliths of the intermediate-composition semischist and greenstone are intruded by dikes of leucogranodiorite (Ogd) that have been deformed into fine-grained silicic semischist and leucogranodiorite breccia. A foliation is only rarely developed in the various rock types due to the cataclastic style of deformation, and to the pervasive orange-weathering dolomitic alteration. These rocks are mapped as a complex because the original depositional and intrusive relations between constituent rock types have been obliterated by the penetrative deformation and alteration.

The age of deformation is constrained by the Ordovician age of the intrusive and stratified rocks included, and by the nondeformed middle Early Devonian sandstone, mudstone, and shale (Dms) that overlies the complex. Deformation of rocks in this complex is interpreted to have been related to earliest Devonian-Silurian movement on the Anchor Island and associated thrust faults (Gehrels and Saleeby, in review).

SOk COMPLEX OF KENDRICK BAY (EARLY SILURIAN AND ORDOVICIAN) -- A heterogeneous complex of dark- to medium-brown and gray schistose gneiss in the Kendrick Bay-McLean Arm area that was mapped originally by MacKevett (1963) and is referred to informally herein as the complex of Kendrick Bay. Layering in the gneiss ranges from several mm to 50 cm in thickness and is defined by variations in the abundance of ferromagnesian minerals versus quartz and plagioclase. Ferromagnesian minerals consist of green hornblende and brown biotite which record amphibolite-facies metamorphism. The composition of the gneiss, combined with rare exposures of deformed pillows, volcanic breccia fragments, rhythmic sedimentary layering, and plutonic textures indicate that these metamorphic rocks have been derived from basaltic to andesitic volcanic rocks (SObv), volcaniclastic strata (SOgw and SOms), and subordinate dioritic to quartz dioritic intrusive rocks (Od and Oqd).

Sills of variably layered and foliated quartz monzonite (SOqm) and quartz diorite (Oqd) are also common. Layering and foliation in these sills and in adjacent quartz monzonite and adamellite (SOqm), quartz diorite (Oqd), and granite (SOgr) bodies are nearly everywhere parallel to the compositional layering and foliation in the complex. These relations plus gradational contacts between metadiorite belonging to this unit and adjacent quartz dioritic and dioritic rocks (Oqd) indicate that the metamorphism and deformation occurred during, and probably as a result of, emplacement of the Early Silurian and Ordovician intrusive rocks. Regional relations and relict protolith features indicate that the metasedimentary and metavolcanic components of this complex were derived from rocks of the Descon Formation rather than the Wales suite.

Or COMPLEX OF RUTH BAY (ORDOVICIAN) -- A distinctive complex of metagabbro-metadiorite with subordinate metagranodiorite sills (approximately 20%) and screens of amphibolite-facies metavolcanic and metasedimentary rocks (<5%). The complex occurs in the Ruth Bay region west of Klakas Inlet and is referred to informally herein as the complex of Ruth Bay. The relative proportions of various rock types are quite consistent between the Keete Inlet fault and the Bird Rocks fault and on southern Klakas Island. On small islands south of Klakas Island, however, the complex consists almost entirely of 10- to 50-cm-thick layers of foliated gabbro, diorite, and quartz diorite. The complex is juxtaposed to the east along the Keete Inlet fault against brecciated and deformed Early Silurian(?) and Ordovician stratified and intrusive rocks (Ogd, Od, SObv, and DSk) and to the west along the Bird Rocks fault against amphibolite-facies rocks of the Wales metamorphic suite (pmOw). The Bird Rocks thrust fault and rocks on either side of the fault are intruded by Silurian leucodiorite (Sd) and Cretaceous granodiorite (Kgd), which indicates that the Bird Rocks fault moved prior to the end of Late Silurian time. The age of movement on the Keete Inlet fault is constrained only as post-middle Early Devonian and pre-mid-Cretaceous (Herreid and others, 1978; Gehrels and Saleeby, in review).

Metavolcanic and metasedimentary rocks occur in this complex as several-m-thick, lens-shaped screens that are enveloped in metagabbro, metadiorite, and metagranodiorite. These metamorphic rocks are recognized by their compositional layering, relict pyroclastic fragments, and, in Clam Cove, the existence of a thin marble layer. The best exposures of protolith features occur near the narrow part of the large island in Ruth Bay. In most areas the metasedimentary and metavolcanic rocks consist of medium-grained green hornblende, brown biotite, plagioclase, opaque minerals, and minor quartz. The lack of penetrative deformation of protolith features in metavolcanic rocks suggests that these rocks were derived from the Descon Formation rather than the Wales suite.

The metagabbro-metadiorite generally consists of dark-gray to black, highly foliated and locally schistose amphibolite. Its color index ranges from 40 to 70, consisting primarily of green hornblende, plagioclase, sphene, opaque minerals, and minor quartz. The rocks are more highly altered to the east and are overprinted by secondary chlorite, epidote, calcite, and white mica. Contacts between various intrusive components in the complex are generally parallel to the pervasive regional foliation, which is defined by elongation and alignment of metamorphic minerals, and by slight cm-scale variations in the relative proportions of hornblende and plagioclase. The degree to which the ferromagnesian minerals are elongated and aligned varies, and in some rocks a plutonic texture is well preserved.

Both the metasedimentary-metavolcanic rocks and the metagabbro-metadiorite are intruded by sills and dikes of metagranodiorite which are generally 10 to 40 cm in thickness and up to several tens of meters in length. The sills commonly taper out along strike, but locally have abrupt, angular terminations or are truncated by ductile shear zones. These sills commonly intrude the other rocks along contacts that are parallel to the foliation in the country rocks and in the granodiorite itself. The metagranodiorite locally cuts across the foliation in adjacent rocks, but in such cases the granodiorite has a foliation which is parallel to, but not as strongly developed as the foliation in the country rocks. These rocks consist predominantly of quartz, plagioclase, K-feldspar, 5% to 15% brown biotite, up to 5% green hornblende that occurs in long narrow grains, sphene, and trains of opaque grains -- all of which are elongated and aligned parallel to the foliation. Although the rocks commonly have a penetrative foliation defined by alignment of ferromagnesian minerals and elongation of quartz and plagioclase, a relict plutonic texture is preserved in most areas. Toward the east these rocks are moderately altered, with chlorite and epidote replacing ferromagnesian minerals, and secondary calcite, white mica, and epidote overprinting plagioclase.

The intrusive and structural relations described above indicate that the metaplutonic rocks were emplaced during deformation and amphibolite-facies metamorphism of the metasedimentary and metavolcanic rocks. The age of this deformation and metamorphism, and formation of the complex, is constrained by a U-Pb apparent age of 465 ± 7 Ma (Middle Ordovician) on a metagranodiorite sill (sample locality 5) which cuts across the foliation in metagabbro-metadiorite, yet contains the regional foliation. Age relations and compositional similarities suggest that the foliated intrusive rocks in this complex may be the deeper-level equivalents of Ordovician diorite (Od) and leucogranodiorite (Ogd) above the Keete Inlet fault (Gehrels and Saleeby, in review).

$\boxed{\text{pmOw}}$ WALES METAMORPHIC SUITE (PRE-MIDDLE ORDOVICIAN) -- A metamorphic suite of greenschist- to amphibolite-facies metavolcanic and metasedimentary rocks referred to originally as the Wales Group by Buddington and Chapin (1929). I have renamed these rocks the Wales metamorphic suite in accordance with guidelines recommended by the North American Commission on Stratigraphic Nomenclature (1983) (Gehrels and Saleeby, in review). In most areas this suite consists of light- to dark-green, fine-grained greenschist and greenstone derived from basaltic to andesitic volcanic rocks and volcaniclastic strata. Pillows and cm-scale pyroclastic fragments are locally preserved in the metavolcanic rocks, and metasedimentary rocks locally show relicts of rhythmic and graded bedding. The metavolcanic and metasedimentary rocks are interlayered over structural thicknesses of tens of meters. In most areas protolith features are obscured by metamorphic recrystallization, penetrative foliation, a high degree of flattening, and moderate elongation. Black phyllite and schist derived from argillaceous strata (A on Plate 1) are interlayered with the greenschist and greenstone along the shoreline between Kassa Inlet and Mabel Bay. Meter-thick layers of silicic metavolcanic rocks (S on Plate 1) and light-colored, coarsely recrystallized marble (L on Plate 1) constitute a minor part of this metamorphic suite.

The Wales suite is metamorphosed to greenschist facies north of the study area (Eberlein and others, 1983; Herreid and others, 1978) and north and west of Kassa Inlet. The dominant metamorphic mineral assemblage in the greenschist-facies rocks consists of chlorite, actinolite, albite, epidote, and opaque minerals. South of Kassa Inlet the metamorphic grade increases eastward from greenschist to amphibolite facies. Along Ship Island Passage the rocks are similar in metamorphic grade and structural style to rocks north and west of Kassa Inlet. To the east the rocks become progressively higher in metamorphic grade, with brown biotite replacing chlorite, and almandine garnet occurring just west of the Shipwreck Point fault. Between the Shipwreck Point and Bird Rocks faults, the rocks are amphibolite facies and consist of fine- to medium-grained garnet, plagioclase, and hornblende and (or) biotite. In addition to this regional metamorphism, rocks in the suite have been metamorphosed to hornblende-hornfels facies adjacent to the Silurian leucodiorite (Sd) body in Kassa Inlet.

Rocks in the Wales metamorphic suite have a penetrative metamorphic foliation which is generally parallel to compositional layering and to the flattening of protolith features. Most rocks also have a strong linear fabric defined by the elongation of protolith features and by strong mineral lineations along foliation surfaces. Isoclinal folds in protolith features occur in some outcrops and have axes which are parallel to the mineral lineation and the elongation direction. The regional foliation generally forms the axial surface of these folds. Relations between the dominant fabric elements and these isoclinal folds, combined with the high degree of flattening of protolith features, indicate that the primary stratigraphic relations in these rocks have been transposed into the metamorphic foliation.

Superimposed on this metamorphic foliation and lineation are several sets of folds that do not have an axial planar foliation, and are interpreted to have formed after the main phase of deformation and metamorphism. Outcrop-scale folds generally plunge less than 30°, have wavelengths of 10 cm to several meters, and are highly asymmetric. Along the shoreline northwest of Shipwreck Point the metamorphic foliation dips to either the southeast or northwest in domains that define a synform-antiform pair with shallow-plunging, northeast-trending axes (Plate 1). Asymmetric folds along this shoreline are coaxial with the synform-antiform pair and show both "s" and "z" asymmetry, with the direction of overturning in the up-dip direction of the foliation. This suggests that the outcrop-scale "s" and "z" folds along this shoreline, and common throughout the Wales metamorphic suite, may be parasitic to a series of shallow-plunging, upright antiforms and synforms with wavelengths of several tens to several hundreds of meters. These preliminary observations indicate that the outcrop-scale asymmetric folds common throughout the Wales suite may not have direct regional kinematic significance.

Relations north of the study area indicate that rocks in the Wales metamorphic suite were metamorphosed and deformed during Early Ordovician-Middle Cambrian time, and that their protoliths are pre-Late Cambrian in age. The maximum age of metamorphism and deformation has been determined north of the study area (in Cholmondeley Sound), where Middle and Late Cambrian metaplutonic rocks have been metamorphosed and deformed along with rocks in the Wales metamorphic suite (J. Saleeby, unpublished data; Gehrels and Saleeby, in review). These relations also demonstrate that the protoliths of rocks in the Wales metamorphic suite must be pre-Late Cambrian in age. An Early Ordovician age of metamorphism has been proposed by Turner and others (1977) based on an ^{40}Ar-^{40}K isochron apparent age of 483 Ma determined from hornblende with tremolite overgrowths. The occurrence of relatively nondeformed latest Early Ordovician and younger strata in the Descon Formation in many areas of southern Prince of Wales Island suggests that the metamorphism and deformation occurred prior to the end of Early Ordovician time. Thus, available constraints indicate that rocks in the Wales suite were deposited prior to Late Cambrian time, and were regionally deformed and metamorphosed during Middle Cambrian-Early Ordovician time.

Eberlein and others (1983) report that the Wales metamorphic suite may be overlain by a several-kilometer-thick section of south-dipping strata near the head of Klakas Inlet (several kilometers north of the study area). They argue that these strata are Cambrian in age and that the Wales suite is at least in part Precambrian because late Early to

Middle Ordovician fossils have been recovered from the top of the section in southern Klakas Inlet (paleontologic sample locality 3). My mapping in Klakas Inlet has shown, however, that the Ordovician strata at sample locality 3 are separated from rocks to the north by the Frederick Cove thrust fault, and that the strata north of the fault are moderately deformed, highly folded, and cut by many high- and low-angle faults. In addition, Herreid and others (1978) report that these strata are separated from the Wales suite near the head of Klakas Inlet by the Keete Inlet fault. Thus, a Precambrian age for rocks in the Wales metamorphic suite has not been demonstrated by stratigraphic relations or geochronometric determinations.

Intrusive rocks

BASALTIC TO DACITIC(?) DIKES (CRETACEOUS TO ORDOVICIAN) -- Dikes of basaltic to dacitic(?) composition are widespread in the study area. They have not been mapped separately on Plate 1 because of their narrow width and poorly constrained length, and also because of their abundance: in many areas there are several tens of dikes per kilometer. There are at least two sets of dikes in the area: an early set that is coeval with the Silurian and (or) Ordovician intrusive rocks, and a younger set that intrudes Devonian strata. The older dikes commonly have schistose borders and are tabular to irregular in shape, up to a meter in thickness, and laterally discontinuous. They consist predominantly of fine-grained, moderately chloritized hornblende, highly altered plagioclase, and abundant secondary epidote, chlorite, white mica, calcite, and other minerals. Younger dikes are generally tabular, laterally continuous, up to several meters in thickness, and steeply dipping, and they have sharp planar contacts with their country rocks.

They commonly are more resistant to erosion than their country rocks and consist of fine-grained green hornblende and plagioclase (andesine) with minor opaque minerals and secondary epidote, chlorite, and calcite in the interstices. Hornblende phenocrysts are common in the younger dikes.

Cross-cutting relations indicate that dikes belonging to the second set were emplaced after middle Early Devonian time. These dikes apparently do not intrude mid-Cretaceous intrusive rocks (Herreid and others, 1983) and their relations with the Jurassic Bokan Mountain Granite (Jgr) are ambiguous. MacKevett (1963) reports that dikes of this set are less common in the granite than in adjacent Paleozoic country rocks, which indicates that they may be in part pre-Jurassic in age. Some dikes are apparently post-Jurassic, however, as albitization associated with the Jurassic granite has affected some dike rocks. These relations suggest that most dikes in the area were emplaced between middle Early Devonian and Jurassic time. I speculate that they may have been emplaced during a latest Paleozoic(?)-Triassic rifting event which affected rocks throughout the Prince of Wales Island region (Gehrels and Saleeby, in review).

$\boxed{\text{Kgd}}$ GRANODIORITE (CRETACEOUS) -- Massive, medium-grained granodiorite that intrudes Devonian and older rocks and several thrust faults west of Klakas Inlet. The color index of these rocks ranges from 5 to 25, with ferromagnesian minerals consisting of green hornblende and subordinate light-green augite and chloritized brown biotite. Anhedral opaque minerals (generally magnetite) constitute up to several percent of some samples, are commonly anhedral, and range up to a millimeter in diameter. Sphene is ubiquitous and locally over a millimeter in length. Plagioclase forms 2- to 5-mm-long, tabular, strongly zoned grains with moderately altered cores. The outer parts of the grains are oligoclase in composition. K-feldspar commonly grows in large (locally over a centimeter in diameter) anhedral grains around plagioclase, and also occurs as small grains intergrown with quartz in the interstices of larger plagioclase grains. Locally associated with these bodies are small masses of gabbro similar to rocks in the Cretaceous gabbro unit (Kgb).

These granodioritic rocks are readily distinguished from the Silurian and Ordovician plutonic rocks by their tabular and strongly zoned plagioclase, large anhedral K-feldspar, and abundant magnetite and sphene. I correlate these rocks with mid-Cretaceous intrusive rocks north of the study area (Herreid and others, 1978) based on their similar mineralogy and texture.

Kd DIORITE (CRETACEOUS) -- Massive, medium- to coarse-grained diorite of probable Cretaceous age east of Klakas Inlet. The color index ranges from 15 to 40, with ferromagnesian minerals including moderately chloritized brown biotite (up to 15%), diopsidic(?) augite (5% to 20%) grains up to 5 mm in length, and green hornblende (10% to 25%) locally seen to be in a reaction relationship with clinopyroxene. Large euhedral grains of sphene, anhedral and irregular opaque (primarily magnetite) grains, and a minor proportion of small anhedral apatite constitute up to 5% of the rock. Quartz is a minor constituent, plagioclase occurs in euhedral to subhedral, tabular, strongly zoned grains of andesine composition, and K-feldspar forms large interstitial grains around the other minerals. This intrusive body is similar in mineralogy, texture, weathering characteristics, and aeromagnetic signature to mid-Cretaceous granodiorite and diorite bodies north of the study area (Eberlein and others, 1983; Rossman and others, 1956; Herreid and others, 1978) and is accordingly interpreted herein to be Cretaceous in age.

Kgb GABBRO (CRETACEOUS) -- A small body of medium- to coarse-grained gabbro that intrudes the Shipwreck Point fault and rocks belonging to the Wales metamorphic suite (pmOw) west of Ruth Bay. Green hornblende, moderately altered plagioclase, and chloritized brown biotite constitute most of the rock. Subordinate granodiorite pods are associated with this intrusive body. Dikes of granodiorite and gabbro cut the metamorphic foliation in the Wales suite and also cut cataclastic fabrics in rocks along the trace of the Shipwreck Point fault. This intrusive body is interpreted to be Cretaceous in age based on mineralogical and textural similarities with gabbroic phases of the mid-Cretaceous intrusive bodies north of the study area (Herreid and others, 1978; Eberlein and others, 1983).

Jgr BOKAN MOUNTAIN GRANITE (JURASSIC) -- A fine- to coarse-grained stock of peralkaline granite that occurs between Kendrick Bay and South Arm (Moira Sound). A considerable amount of research has been conducted on the petrology and geochemistry of this body because of its unusual composition and its association with uranium, thorium, and rare-earth-element deposits in the area. MacKevett (1963) originally mapped the stock as silica-rich, riebekite- and aegirine-bearing peralkaline granite. Thompson and others (1982) interpret the body to be a ring-dike complex and subdivide it into twelve separate phases of aegirine- and riebekite-bearing granite, aplite, porphyry, and pegmatite. The four major phases recognized by Thompson and others (1982: their figure

2) are subdivided on Plate 1 as follows: riebekite granite porphyry (Jgr-r), aegirine granite porphyry (Jgr-a), fine-grained riebekite granite porphyry (Jgr-fr), and felty-aegirine granite (Jgr-fa). The other rock-types are grouped into an undivided granite unit (Jgr).

Thompson and others (1982) describe these four phases as follows: Riebekite granite porphyry (Jgr-r) consists of quartz phenocrysts (up to 5 mm in diameter), subhedral riebekite phenocrysts (up to 4 mm in length), albite, and K-feldspar. This rock grades outward through a transition zone of aegirine- and riebekite-bearing granite into aegirine granite porphyry (Jgr-a). Phenocrysts in the aegirine porphyry include quartz and microperthite grains up to 8 mm in diameter. Groundmass and accessory minerals include quartz, microcline, albite, aegirine, sphene, zircon, monazite, muscovite, and fluorite. The fine-grained riebekite granite porphyry (Jgr-fr) consists of microcline, quartz, and riebekite phenocrysts in a groundmass of albite, microcline, quartz, and riebekite. Felty-aegirine granite (Jgr-fa) occurs in the center of the body and consists of fine aegirine needles in an aplitic groundmass of quartz, K-feldspar, albite, and accessory sphene and fluorite.

As reported by Saint-André and others (1983), Bernard Collot has divided the granite into three main phases: 1) albitic aegirine granite around the margin of the body, 2) fine-grained albitic arfvedsonite-aegirine granite in the center, and 3) albitic arfvedsonite granite between them. I concur that the sodic amphibole in the rock is arfvedsonite rather than riebekite.

The Bokan Mountain Granite has been dated by a variety of isotopic methods, all of which yield Jurassic apparent dates. Lanphere and others (1964) report K-Ar dates of 185 ± 8 Ma and 190 ± 8 Ma (Middle to

Early Jurassic) on riebekite (arfvedsonite?), and Saint-André and others (1983) report a U-Pb apparent age of 171 ± 5 Ma (Middle Jurassic) based on analyses of 10 zircon fractions. I consider this U-Pb age suspect, however, because: 1) their (radiogenic lead)/(common lead) is so low that even a small error in the assigned composition of common lead results in a large uncertainty in apparent age, and 2) several fractions plot above concordia on a $^{206}Pb^{*}/^{238}U$ versus $^{207}Pb^{*}/^{235}U$ concordia diagram, which indicates that their concentrations of uranium and lead may be incorrect due to incomplete dissolution of the zircon and (or) incomplete equilibration of the sample and spike. This interpretation is supported by their lower intercept with concordia of 0 ± 15 Ma. Rb/Sr analyses of ten whole-rock samples from various phases of the granite yield an $^{87}Sr/^{86}Sr$ versus $^{87}Rb/^{86}Sr$ isochron apparent age of less than 156 Ma (Armstrong, in press). A selection of seven samples yields an isochron apparent age of 151 ± 5 Ma (Late Jurassic) which Armstrong (in press) interprets to be a minimum age for the granite.

Dph PLAGIOCLASE-PORPHYRITIC HYPABYSSAL ROCKS (EARLY DEVONIAN) -- Small intrusive bodies of plagioclase porphyry that occur in association with Devonian strata in the Klakas Inlet region. Similar bodies have been mapped by Herreid and others (1978) in Devonian strata to the north near Keete Inlet. Large (.4 to 1.5 cm long), euhedral, moderately altered, interlocking grains of labradorite and andesine constitute most of the rock. Light-green to light pinkish brown diopsidic(?) augite constitutes from 15% to 30% of most samples and occurs as several-mm-scale, subhedral grains intergrown with plagioclase. In some samples these clinopyroxene grains have rims of green or reddish-brown hornblende and opaque minerals. Microcline and microperthite generally occur as sub-mm-scale grains in the interstices of plagioclase grains. The abundance of K-feldspar is quite variable and opaque minerals occur as skeletal grains up to 3 mm across. Millimeter-scale angular clots of chlorite and subordinate opaque minerals, calcite, and white mica have filled miarolitic cavities in the rock. These cavities and the lack of hornfels aureoles around the intrusive bodies indicate that they were emplaced at shallow levels.

Cross-cutting relations with Early Devonian strata indicate that these rocks are Devonian or younger, and mineralogical and textural similarities suggest that they are related to the Early Devonian plagioclase-porphyritic volcanic rocks (Dpv) in the southwestern part of the area.

Sd LEUCODIORITE (LATE SILURIAN) -- Massive, medium- to coarse-grained leucocratic diorite and subordinate monzodiorite and monzonite in Kassa Inlet and inland to the southeast. On Kassa Island the body intrudes across metamorphic fabrics in the Wales metamorphic suite (pmOw), and has a hornblende-hornfels contact aureole which extends for distances of several meters. Dikes also cut the deformational fabrics in Ordovician metaplutonic rocks belonging to the complex of Ruth Bay (Or) south of Kassa Inlet. Map patterns indicate that the body cuts across the northern continuations of the Bird Rocks and Shipwreck Point faults.

These rocks lack quartz, have a color index of less than 25, and include 1- to 6-mm-long anhedral grains of arfvedsonite that in some samples contain cores of aegirine-augite. Melanite(?) garnet constitutes up to 10% of the rock, and is generally medium- to dark-brown, moderately zoned, subhedral to euhedral, and up to 4 mm in diameter. Sphene occurs as euhedral to subhedral grains up to 5 mm in length that are commonly intergrown with garnet and arfvedsonite. Opaque minerals are rare. The grains listed above tend to occur in glomerocrysts up to a centimeter in diameter. Moderately zoned, tabular plagioclase grains up to 7 mm in length constitute most of the rock. The cores of these grains are highly altered whereas the outer parts are fresh and oligoclase in composition. Microperthite is generally less abundant than plagioclase, and forms in the interstices between plagioclase grains and glomerocrysts of the other minerals.

A U-Pb apparent age of 418 ± 5 Ma (Late Silurian) has been determined on a sample of leucodiorite from Kassa Island (geochronologic sample 11: Gehrels and Saleeby, in review).

SOsy QUARTZ SYENITE AND GRANITE (EARLY SILURIAN AND (OR) LATE ORDOVICIAN) -- A heterogeneous suite of massive, fine- to medium-grained, gray-, reddish gray-, and maroon-weathering quartz syenite and granite in the McLean Arm-Nichols Bay region. In Stone Rock Bay these rocks intrude, and locally appear to grade into, rocks belonging to the pyroxenite and hornblendite (SOpx) unit. Along the shoreline between Cape Chacon and Nichols Bay, and in the interior of the island between McLean Arm and Nichols Bay, these rocks are difficult to distinguish from rocks belonging to the quartz monzonite and adamellite (SOqm) unit. More detailed mapping in this area would probably result in designation of a map unit consisting of rocks intermediate in composition and texture between the quartz syenitic and quartz monzonitic rocks. On Plate 1 these intermediate rocks have been grouped with rocks in the quartz monzonite and adamellite (SOqm) map unit. Dikes of maroon to red feldspar-porphyritic syenite cut rocks belonging to all of the units in the McLean Arm-Nichols Bay area and represent the latest phase of syenitic intrusive activity.

The quartz syenitic and granitic rocks have a low color index (generally less than 20) and consist primarily of interlocking, fine- to medium-grained microperthite, quartz, and plagioclase. Plagioclase forms tabular grains up to 3 mm in length that are generally the largest grains in the rock. They are highly altered with secondary white mica, calcite, and epidote, and have albitized rims. Compositionally they range from low-calcium andesine to oligoclase and are not zoned. K-feldspar occurs as sub-mm-scale anhedral grains of microperthite intergrown with quartz around plagioclase grains. These grains are not as highly altered as the plagioclase, but have a slight overprint of fine-grained white mica, calcite, and epidote. The ratio of microperthite to plagioclase is generally about 2:1. Quartz grains are also sub-millimeter in size and constitute from 5% to 30% of most rocks. Ferromagnesian minerals include glomerocrysts of green hornblende and subordinate brown biotite, opaque minerals, and sphene. In most samples the hornblende and biotite are moderately altered to chlorite and opaque minerals. Zircon is an abundant accessory mineral, and medium-brown garnet is locally present. Syenitic rocks described by MacKevett (1963) are quite similar to the rocks described above, except for a greater abundance of quartz in my samples.

I have determined a U-Pb apparent age of 438 ± 5 Ma (earliest Silurian and (or) latest Ordovician) on a sample of quartz syenite from sample locality 10 (Gehrels and Saleeby, in review).

SOpx PYROXENITE AND HORNBLENDITE (EARLY SILURIAN AND (OR) LATE ORDOVICIAN) -- Coarse-grained to pegmatitic pyroxenite, hornblende pyroxenite, and hornblendite that occurs along the eastern shore of southernmost Prince of Wales Island. In McLean Arm, Stone Rock Bay, and at the western end of the elongate body north of Cape Chacon this unit consists of medium- to coarse-grained pyroxenite and hornblende pyroxenite. Hornblende generally occurs around cores of augite grains and is associated with abundant opaque minerals. Plagioclase locally occurs in the interstices of the large grains, and MacKevett (1963) reports that biotite occurs in some samples. The ultramafic body along the shoreline north of Cape Chacon is a coarse-grained hornblendite to hornblende pegmatite with hornblende crystals up to 10 cm in length. Large grains of anhedral opaque minerals are common in this rock and plagioclase occurs in the interstices of the hornblende crystals. The pyroxenite and hornblendite are intruded by quartz monzonite (SOqm) along the shoreline north of Cape Chacon, and by quartz syenite (SOsy) at several localities in Stone Rock Bay. Also in Stone Rock Bay, however, is an exposure of quartz syenite grading into fine-grained pyroxenite over a distance of several meters. MacKevett (1963, pp. 17) reports that the contacts between the pyroxenite and quartz monzonite are "gradational through a zone of hybrid rock or intrusive breccia." These relations, plus the close spatial association of ultramafic rocks with quartz syenite and quartz monzonite suggest that rocks in the three units may be coeval and genetically related. I have accordingly assigned an Early Silurian and (or) Late Ordovician age to the ultramafic rocks.

$\boxed{\text{SOgr}}$ GRANITE (EARLY SILURIAN AND (OR) LATE ORDOVICIAN) -- Weakly foliated, light-pink to light-gray, coarse-grained granite that occurs along the eastern shore of Prince of Wales Island between the Kendrick

Islands and McLean Arm. The foliation in this rock is generally parallel to the strong foliation and layering in the complex of Kendrick Bay (SOk) and the weak foliation in adjacent bodies of quartz monzonite and adamellite (SOqm). Foliated sills of granite are generally parallel to the foliation in these rocks, suggesting that the granite was emplaced during the waning stages of deformation and metamorphism.

The granite has a color index of less than 10 and consists primarily of interlocking plagioclase, quartz, and large K-feldspar grains. Plagioclase is generally oligoclase with moderate zoning from more calcic interiors to more sodic rims. Some grains show oscillatory zoning as well. Their interiors are moderately altered with secondary white mica, calcite, and epidote, and the rims are slightly albitized. They occur both as small grains enclosed in K-feldspar, and as subhedral grains up to 5 mm long that are intergrown with K-feldspar and quartz. Quartz occurs as large anhedral grains intergrown with feldspar and constitutes up to 40% of some samples. K-feldspar occurs both as poikilitic microperthite grains up to 1.5 cm across and as microcline grains less than several millimeters in diameter. Ferromagnesian minerals have been altered to chlorite and opaque minerals in most samples, although large books of brown biotite are preserved in some rocks. Based on the shape of the chlorite masses, it appears that biotite was originally dominant over hornblende. Large subhedral sphene and small euhedral zircon grains are common accessory phases.

A U-Pb apparent age of 438 ± 5 Ma (geochronologic sample 9) indicates that rocks in this map unit were emplaced during earliest Silurian and (or) latest Ordovician time (Gehrels and Saleeby, in review).

SOqm QUARTZ MONZONITE AND ADAMELLITE (EARLY SILURIAN AND (OR) LATE ORDOVICIAN) -- Medium-grained quartz monzonite and adamellite that underlies much of the eastern and central parts of southern Prince of Wales Island. In most areas these rocks are massive, although along the eastern shore of Prince of Wales Island the rocks are locally foliated. Between Kendrick and Ingraham Bays the foliated rocks are mapped as a separate unit of foliated quartz monzonite (SOfqm). Mafic dikes are widespread in the quartz monzonitic rocks, and locally show evidence of emplacement prior to crystallization of the country rocks. Agmatite with blocks of microdiorite in a quartz monzonite host is also common. The quartz monzonitic rocks intrude the slightly older quartz diorite (Oqd) and are in turn intruded by granite (SOgr) and quartz syenite (SOsy).

Rocks in this unit are quite variable in color index, relative proportions of plagioclase versus K-feldspar, and abundance of quartz, but generally are quartz monzonite, adamellite, granodiorite, or quartz monzodiorite in composition. Their color index ranges from 5 to as much as 50, with an average of approximately 25. More mafic and more leucocratic phases are in most cases gradational with quartz monzonite, and are interpreted to be cogenetic.

Most rocks consist primarily of interlocking plagioclase, K-feldspar, and quartz. Plagioclase occurs as tabular and subhedral grains that have moderate compositional zoning. Their cores generally are highly altered with secondary white mica, calcite, and epidote, and outer parts of the grains are sodic andesine to calcic oligoclase in composition. In highly altered rocks the plagioclase has albitized rims, but the widespread albitization reported by MacKevett (1963) has not been observed in samples from outside the western Kendrick Bay region. The plagioclase grains are locally up to 7 mm in length and appear to have been among the early minerals to crystallize. K-feldspar occurs both as subhedral grains up to several millimeters in length, and as smaller microperthite grains in the interstices of the ferromagnesian minerals and larger feldspars. K-feldspar is slightly subordinate to plagioclase in abundance in most rocks. Quartz constitutes between 5% and 40% of the rocks and occurs as anhedral grains in the interstices of feldspar and ferromagnesian minerals. Green hornblende (commonly with cores of light-green augite) is slightly more abundant than brown biotite, and both occur together in glomerocrysts up to a centimeter in diameter. Opaque minerals and sphene are common in these glomerocrysts as well. Apatite and zircon are common accessory minerals.

Biotite quartz monzonite from sample locality 8 yields a U-Pb apparent age of 438 ± 4 Ma (earliest Silurian and (or) latest Ordovician: Gehrels and Saleeby, in review). Lanphere and others (1964) report K-Ar apparent ages of 454 ± 22 Ma on hornblende (geochronology sample locality 13) and 379 ± 18 Ma on biotite (geochronology sample locality 14) from quartz monzonite in the South Arm of Kendrick Bay. Armstrong (in press) reports an $^{87}Rb/^{86}Sr$ versus $^{87}Sr/^{86}Sr$ isochron apparent age of 432 ± 19 Ma based on analyses of rocks collected from this map unit, a more foliated quartz monzonite that may or may not belong to this unit, and gabbro which probably belongs with rocks in the diorite (Od) map unit. His samples of quartz monzonite alone do not have a sufficient range in $^{87}Rb/^{86}Sr$ to define a meaningful isochron, and are therefore not reliably dated by this method. Assuming that the biotite date of Lanphere (1964) is not a crystallization age, the various geochronological data and field relations are consistent with emplacement of these rocks during Early Silurian and (or) Late Ordovician time.

SOfqm FOLIATED QUARTZ MONZONITE (EARLY SILURIAN AND (OR) LATE ORDOVICIAN) -- Foliated and locally layered quartz monzonite that occurs near large quartz monzonite and adamellite (SOqm) bodies in the Kendrick Bay region. Contacts between foliated quartz monzonite (SOfqm) and non-foliated quartz monzonite (SOqm) are gradational where exposed, suggesting that rocks in the two units are coeval and genetically related. Between Ingraham and Kendrick Bays these foliated rocks intrude and produce hornblende-hornfels aureoles in adjacent volcanic rocks of the Descon Formation. The foliation in the intrusive rocks is defined by elongation of the quartzo-feldspathic minerals and alignment of hornblende and biotite. On the Kendrick Islands and locally to the north these rocks have a faint cm-scale layering defined by variations in the relative proportion of quartzo-feldspathic versus ferromagnesian minerals.

The mineralogy of these rocks is similar to that in associated quartz monzonite and adamellite (SOqm): dominant minerals include mm-scale subhedral plagioclase, quartz, K-feldspar, and approximately 25% hornblende and biotite. The primary differences are that quartz tends to be finer-grained and elongate, and hornblende and biotite are generally aligned and commonly occur in elongate masses.

Oqd QUARTZ DIORITE AND DIORITE (LATE ORDOVICIAN) -- Fine- to medium-grained quartz diorite and subordinate diorite and quartz monzonite that occurs primarily in the Kendrick Bay region (MacKevett, 1963). Quartz diorite and diorite have not been subdivided in the area mapped by MacKevett (1963) -- outside this area I have mapped them separately into quartz diorite (Oqd) and diorite (Od) units. In the area mapped by MacKevett (1963), the two rock types are gradational and generally similar in mineralogy. They differ primarily in color index, which ranges from 40 to 60 for diorite and from 20 to 40 for quartz diorite. Quartz diorite is commonly massive, although mafic dikes and agmatitic zones are widespread. Dioritic rocks are more heterogenous, with common zones of diorite-basite migmatite and agmatite, and comb-textured dikes of hornblende pegmatite. Rocks in this unit intrude the Descon Formation and are intruded by Early Silurian and (or) Late Ordovician quartz monzonite and adamellite (SOqm).

Quartz diorite consists primarily of interlocking plagioclase, hornblende, and quartz. Plagioclase occurs as tabular grains of andesine that range in length from less than a millimeter to 6 mm. The grains are slightly zoned, and their interiors commonly are more highly altered with secondary white mica, calcite, and epidote than the margins. K-feldspar is generally absent, although a few samples contain interstitial microperthite. Quartz generally constitutes less that 15% of the rock and occurs as small interstitial grains. Green hornblende forms several-mm-long grains intergrown with subordinate brown biotite (generally chloritized) and opaque minerals. Hornblende grains are commonly poikilitic with many small inclusions of quartz, plagioclase, and accessory minerals. Opaque minerals are locally over a millimeter in diameter and constitute up to several percent of some samples. Sphene occurs as fairly large subhedral grains, and apatite and zircon are minor accessory phases.

The age of the quartz diorite is constrained by a U-Pb apparent age of 445 ± 5 Ma (Late Ordovician: Gehrels and Saleeby, in review) on a sample from Kendrick Bay (geochronologic sample 6). Lanphere and others (1964) report a K-Ar apparent age of 439 ± 21 Ma (Late Ordovician) on hornblende from quartz diorite in western Kendrick Bay (geochronologic sample locality 15).

$\boxed{\text{Ofgd}}$ FOLIATED GRANODIORITE (ORDOVICIAN) -- Foliated and locally layered leucocratic granodiorite that occurs in the vicinity of Tah Island. These rocks intrude and are interlayered with foliated and layered Ordovician dioritic rocks (Ofd). Contacts between the granodiorite and diorite are nearly everywhere parallel to the foliation and layering in the diorite and the foliation in the granodiorite. Some rocks have only a slight foliation, yielding well-preserved plutonic textures. Highly foliated rocks generally occur in narrow domains interpreted to have been ductile shear zones. Contact relations with the dioritic rocks combined with the style and variability of the foliation in the granodiorite indicate that the granodiorite was emplaced and deformed prior to complete crystallization of the diorite.

Highly foliated members of this unit consist of fine-grained, interlocking quartz, plagioclase, and subordinate K-feldspar and subhedral green hornblende. Their foliation is defined by alignment of elongate hornblende and quartz grains, thin layers of chlorite and epidote aligned parallel to the foliation, and slight cm-scale variations in grain size and relative abundance of the minerals. Opaque minerals tend to be more common in the coarser-grained quartzo-feldspathic layers. The less-foliated rocks retain their primary texture and consist of medium-grained plagioclase (oligoclase), quartz, and microperthite. Up to 15% of the rock consists of hornblende grains that are moderately recrystallized to chlorite and opaque minerals.

The age of rocks in this unit has not been determined directly, as a geochronologic sample of foliated leucogranodiorite from the larger island west of Tah Island did not yield zircons. Intrusive and structural relations suggest, however, that these rocks are generally coeval with the foliated and layered diorite and quartz diorite (Ofd) of known Ordovician age.

$\boxed{\text{Ofd}}$ FOLIATED AND LAYERED DIORITE AND QUARTZ DIORITE (ORDOVICIAN) -- Moderately to highly foliated and layered diorite and quartz diorite that occurs near Tah Bay and in Max Cove. There is a continuous gradation in this unit from rocks that are moderately foliated and display intrusive relations, to highly layered and foliated rocks that only locally display protolith relations. In highly deformed domains the dioritic and quartz dioritic rocks form strongly foliated, several-cm-thick layers in which the compositional layering is nearly everywhere parallel to the penetrative foliation. Locally, however, quartz dioritic layers intrude at an acute angle across the foliation in more dioritic rocks, yet contain a foliation parallel to that in their country rocks. Cross-cutting relations are common in the moderately deformed rocks, where more silicic rocks generally intrude more basic rocks. In most cases the foliation in the younger intrusive rocks is not as strongly developed as in the rocks they intrude.

A continuous gradation between highly and moderately deformed rocks is seen along the eastern shore of the narrow bay east of Tah Bay. In the northern part of the bay (and inland to the east) the rocks have a strong and laterally continuous foliation and layering. To the south the rocks become less foliated, intrusive contacts between the various layers are well preserved, and the layering and foliation are variable in orientation. In the southern part of this narrow bay the layers are clearly intrusive dikes and sills with well-preserved igneous fabrics, the rocks are not as highly foliated, and the average thickness of compositional units increases to several tens of centimeters. Quartz diorite also replaces diorite as the dominant rock type.

Dioritic to quartz dioritic rocks belonging to this unit consist of varying proportions of green hornblende, plagioclase, and quartz. In the dioritic rocks, elongate green hornblende grains up to 5 mm in length constitute 25% to 50% of the rock. Anhedral plagioclase grains of andesine composition are generally smaller than hornblende. Quartz and K-feldspar are minor components of the dioritic rocks and occur as small grains intergrown with plagioclase. Most rocks are fairly fresh, although in some samples hornblende has been replaced by chlorite, epidote, and opaque grains, and plagioclase has been altered to white mica, epidote, and calcite. Hornblendite is a minor component and consists of interlocking, subhedral, green hornblende grains that are locally over a centimeter in length. Plagioclase occurs in the interstices of the hornblende grains. Quartz dioritic rocks are similar to the diorite except for greater proportions of quartz and K-feldspar, and the presence of large anhedral sphene grains.

A moderately foliated quartz diorite in Tah Bay (geochronologic sample 7) yields a U-Pb apparent age of 446 ± 5 Ma (Late Ordovician: Gehrels and Saleeby, in review). This is interpreted to be the approximate age of formation of the unit based on the intrusive and structural relations described above. Similarities in mineralogy, composition, and apparent age indicate that these rocks may be related to rocks in the quartz diorite and diorite (Oqd) unit.

Intrusive relations indicate that these rocks were deformed prior to and during crystallization. I interpret this deformation to have occurred in response to primarily magmatic processes, as there is no evidence in the area of regional deformation during Late Ordovician time.

Ogd LEUCOGRANODIORITE (ORDOVICIAN) -- Leucogranodiorite in the Klakas Inlet region which intrudes Ordovician diorite (Od) and rocks of the Descon Formation (SObv and SOa), and is locally overlain by Early Devonian clastic strata (Dcg and Dbx). Contacts between leucogranodiorite and older rocks are commonly agmatitic, with 10-cm-scale clasts of country rock enveloped in leucogranodiorite. In most areas of Klakas Inlet and Max Cove the rocks in this unit are penetratively brecciated and consist of cm-scale angular clasts in a matrix of fine-grained rock fragments, quartz, and plagioclase. Adjacent to the brecciated domains are regions in which the rocks are highly fractured -- only rarely are they not deformed.

Rocks in this unit consist of interlocking and commonly myrmekitic plagioclase, quartz and subordinate interstitial microperthite. Plagioclase occurs as subhedral, nonzoned grains ranging from a few millimeters to over a centimeter in length. Compositionally the plagioclase is oligoclase to calcic albite. In most rocks the plagioclase has been overprinted by fine-grained, secondary white mica, calcite, and epidote. Quartz occurs as large grains with myrmekitic or sutured boundaries against plagioclase, and as small interstitial grains. The leucogranodiorite contains up to 20% microperthite that occurs in most samples as small grains in the interstices of plagioclase and quartz. The original ferromagnesian minerals are totally altered to blocky masses of chlorite and tiny opaque grains, the shape of which suggests that hornblende was probably the original ferromagnesian mineral. These chlorite masses constitute less than 5% of most rocks.

The age of these rocks has been determined by a U-Pb apparent age of 468 ± ~15 Ma (Middle Ordovician) from a sample in Max Cove (sample locality 3: Gehrels and Saleeby, in review). Turner and others (1977) report a K-Ar (hornblende) apparent date of 428 ± 13 on granodiorite at locality 12, which they interpret to be a minimum age. This granodiorite is intruded by the leucogranodiorite which I have dated at locality 3, which suggests that 428 ± 13 Ma is considerably younger than the emplacement age of the rock. I suggest that their date may alternatively record the timing of deformation and alteration of the pre-middle Early Devonian rocks in the southern Klakas Inlet region (Gehrels and Saleeby, in review).

Od DIORITE (ORDOVICIAN) -- A heterogeneous suite of diorite and subordinate hypabyssal diorite and gabbro that occurs in the western and southern parts of the study area. On Klakas Island, in Ruth Bay, and along the western shore of Max Cove the rocks are medium grained and homogeneous but have a penetrative cataclastic fabric in many areas. Along the east shore of Max Cove they are massive, agmatitic with diorite clasts in a quartz diorite or leucogranodiorite matrix, or moderately layered and foliated. On Middle Island (Barrier Islands), dioritic rocks occur as dikes and as small intrusive bodies of diorite, quartz diorite, and subordinate gabbro that are interpreted to be subvolcanic to adjacent basaltic to andesitic rocks of the Descon Formation (SObv). On southernmost Prince of Wales Island the rocks are massive, homogeneous, and both intrude and are intruded by quartz-porphyritic granodiorite (Oqgd). In the Kendrick Bay-McLean Arm region, MacKevett (1963) maps the dioritic rocks together with quartz diorite, and both are included in the quartz diorite (Oqd) map unit on Plate 1. The rocks shown as diorite (Od) in the Kendrick Bay-McLean Arm area were mapped by MacKevett (1963) as gabbro.

Dioritic rocks belonging to this unit have a color index of approximately 40, with a range from 65 for the more gabbroic rocks to as low as 25. Most rocks are medium grained and consist primarily of green hornblende, moderately altered plagioclase, and subordinate quartz and K-feldspar. Hornblende is several millimeters or less in length, subhedral, and only slightly altered in most rocks. In a few samples, particularly near Max Cove, a small amount of biotite is intergrown with hornblende, and some hornblende grains contain small cores of augite. Small grains of opaque minerals, apatite, and sphene are associated with the ferromagnesian minerals. Plagioclase occurs as slightly zoned, moderately to highly altered subhedral grains up to 5 mm in length. The outer parts of most grains are andesine in composition: their interiors are too highly altered for compositional analysis. Quartz occurs as small interstitial grains in most samples and rarely constitutes over 10% of the rock. K-feldspar is rare except near Max Cove where samples contain up to 10% interstitial microperthite.

Dioritic rocks yield U-Pb apparent ages of 462 ± 15 Ma (Middle Ordovician: sample 2) from Klakas Island and ~480-460 Ma from Max Cove (sample locality 4: Gehrels and Saleeby, in review). These apparent ages combined with the close association between dioritic rocks and quartz-porphyritic granodiorite (Oqgd) and quartz diorite (Oqd) indicate that the diorite is of Middle and perhaps locally Late Ordovician age.

$\boxed{\text{Oqgd}}$ QUARTZ-PORPHYRITIC GRANODIORITE (ORDOVICIAN) -- Large bodies of medium-grained, quartz-porphyritic granodiorite that occur on southwesternmost Prince of Wales Island. In the Barrier Islands and along the southeastern shore of Hessa Inlet, the granodiorite contains many small bodies of diorite, the larger ones of which are shown on Plate 1. In proximity to these dioritic bodies the granodiorite is agmatitic, with angular and lens-shaped blocks of diorite, microdiorite, or microgabbro. These rocks are also intruded by dikes and small bodies of fine-grained quartz diorite, diorite, gabbro, and basalt in many areas. Intrusive relations suggest that the dikes were emplaced prior to solidification of the granodiorite. In many areas the granodiorite is also intruded by diorite (Od). Contradictory intrusive relations with diorite and the evidence for syn-igneous emplacement of dioritic dikes indicate that the quartz-porphyritic granodiorite (Oqgd) and diorite (Od) are at least in part coeval.

Rocks in this unit consist primarily of large plagioclase and quartz grains, small, highly altered ferromagnesian minerals, and interstitial microperthite. Plagioclase is generally subhedral, tabular, 2 to 6 mm in length, and oligoclase in composition. Secondary white mica, calcite, and epidote are widespread in the interiors of the grains, and the margins are commonly albitized. Myrmekitic intergrowths with quartz are common along the margins of the grains, and in some samples all of the plagioclase is myrmekitic. In most samples plagioclase occurs in irregular clusters separated by large (up to a centimeter in diameter) bluish quartz grains. Microperthite and subordinate microcline occur in the interstices of plagioclase and quartz. These grains constitute 10% to 20% of most samples, rendering a plagioclase to K-feldspar proportion of about 3:1. In a few samples microperthite grains are quite large and envelop plagioclase. The original ferromagnesian minerals have been entirely altered to millimeter-scale blocky masses of chlorite and opaque minerals. The shape of the masses suggests that the secondary minerals have replaced hornblende rather than biotite. The color index of most rocks ranges from 5 to 20. Sphene and zircon occur as tiny accessory phases in most samples.

A U-Pb apparent age of 472 ± 5 Ma (Middle Ordovician) on a sample from Hessa Inlet (geochronologic sample 1: Gehrels and Saleeby, in review) indicates that these rocks were emplaced during Middle Ordovician time.

pmOgb METAGABBRO (PRE-MIDDLE ORDOVICIAN) -- A heterogeneous body of metamorphosed and deformed gabbro that intrudes protoliths of the Wales metamorphic suite (pmOw) near Ship Island Passage (Plate 1). The rocks generally consist of several-mm- to cm-scale grains of clinopyroxene, hornblende, chlorite, and opaque minerals, mm-scale grains of zoned and highly altered plagioclase, and abundant small grains of chlorite, epidote, calcite, and white mica. These rocks intrude protoliths of the Wales metamorphic suite (pmOw) and have experienced the regional greenschist-facies metamorphism. Their minimum age is constrained by the interpretation that regional metamorphism occurred during Early Ordovician-Middle Cambrian time (Gehrels and Saleeby, in review).

STRUCTURAL GEOLOGY

The dominant regional structures in the area include thrust faults, the Keete Inlet fault, and several sets of strike-slip faults. The nature of the faults and their sense and age of displacement are described below. Structures that are restricted to a particular map unit (e.g., folds and deformational fabrics in the Wales metamorphic suite) are described for each unit in the Description of Map Units.

Thrust faults

Shipwreck Point, Bird Rocks, and Ruth Island faults

Thrust faults west of and structurally beneath the Keete Inlet fault dip toward the east at moderate angles and imbricate a variety of Ordovician and older rocks. These faults have been studied primarily along the south shore of the peninsula south of Kassa Inlet -- their trace to the north is inferred from topography and from strong lineaments on aerial photographs. The southern shore of Kassa Inlet has not been revisited to check for the existence of these faults.

The Bird Rocks fault forms a major tectonic boundary in that it juxtaposes Ordovician metaplutonic and subordinate metavolcanic and metasedimentary rocks (complex of Ruth Bay: Or) against amphibolite-facies rocks of the Wales metamorphic suite (pmOw). The fault is recognized along the shoreline west of Ruth Bay as a wide zone of brecciation which separates amphibolite-facies rocks of the Wales suite (pmOw) from rocks belonging to the complex of Ruth Bay. The Ruth Island fault is recognized as a several-m-wide zone in which rocks belonging to the complex of Ruth Bay are brecciated into 10-cm-scale angular blocks. Zones of brecciation associated with both the Ruth Island and Bird Rocks faults are intruded by Cretaceous granodiorite (Kgd) along the shoreline west of Ruth Bay.

The Shipwreck Point fault occurs within rocks of the Wales metamorphic suite (pmOw) and juxtaposes amphibolite-facies rocks to the east against greenschist-facies rocks to the west. Structural trends on either side of the fault are also quite different: to the west the foliation and axes of asymmetric folds strike and trend to the northeast, whereas to the east the structural grain is northwesterly. The fault outcrops in the bay east of Shipwreck Point as a several-m-wide zone of breccia intruded by swarms of gabbro (Kgb) dikes.

These faults are interpreted to dip easterly because slickenside surfaces within the fault zones and in rocks on either side generally dip at moderate angles toward the east, and because their traces inland indicate gently eastward-dipping fault planes. Their slip-line is recorded by an abundance of east-northeasterly trending slickenside striae on minor fault surfaces (Gehrels and Saleeby, in review). A southwestward direction of movement is indicated by: 1) juxtaposition of higher-grade rocks over lower-grade rocks along the Shipwreck Point fault (assuming that higher-grade rocks were at greater depth than lower-grade rocks prior to movement on the fault), and 2) southwestward overturning of a regional antiform in the complex of Ruth Bay (Or) above the above the Ruth Island fault. A detailed study of minor folds along the shoreline west of the Bird Rocks fault demonstrates that asymmetric folds in the Wales metamorphic suite (pmOw) formed prior to movement on the thrust faults.

Toward the north the Shipwreck Point, Bird Rocks, and Ruth Island faults are intruded by a large body of Silurian leucodiorite (Sd) in Kassa Inlet and are apparently cut by the Keete Inlet fault. Rocks of the Wales metamorphic suite (pmOw) north and west of Kassa Inlet and on the large island north of Kassa Island are greenschist facies, and are interpreted to occur structurally beneath the Shipwreck Point fault. In contrast, rocks on the northeastern corner of Kassa Island were apparently of regional amphibolite facies prior to contact metamorphism related to the Silurian leucodiorite (Sd) body. The Shipwreck Point fault is therefore tentatively shown separating these amphibolite-facies(?) rocks from greenschist-facies rocks to the north and west. Because metaplutonic rocks belonging to the complex of Ruth Bay (Or) do not occur on Kassa Island, the northward continuation of the Bird Rocks fault is drawn south and east of Kassa Island.

The minimum age of movement on these faults is constrained by the cross-cutting body of Silurian leucodiorite (Sd) on and adjacent to Kassa Island. Their maximum age is indicated by exposures of brecciated Middle Ordovician metagranodiorite along the Ruth Island and Bird Rocks faults west of Ruth Bay. Regional relations suggest that thrust faults in this system moved primarily during earliest Devonian-Silurian time (Gehrels and Saleeby, in review).

Frederick Cove fault

The Frederick Cove fault juxtaposes two different stratigraphic sections of the Descon Formation along the northern edge of the study area. North of the fault, Early Silurian(?) and Ordovician argillite, shale, mudstone, and siltstone generally dip and face to the southwest. Rocks south of the fault consist of Ordovician volcanic rocks and subordinate marine clastic strata which, based on reconnaissance mapping along the north shore of Moira Sound, probably underlie the clastic strata north of the fault.

The fault is not exposed at the head of Frederick Cove, but strong topographic and vegetational lineaments can be traced inland on aerial photographs from the northern and southern shorelines at the head of the bay. Volcanic rocks along the shore of Frederick Cove near the southern lineament have a strong foliation that dips moderately to the south-southwest. This fabric is interpreted to have formed during movement on the Frederick Cove fault. The two lineaments can be traced westward across the island to the east shore of Klakas Inlet, where a steeply dipping, highly altered shear zone is exposed. Slickenside striae in this shear zone generally plunge steeply to the south or southwest. The westward continuation of the fault has been mapped by Herreid and others (1978) along the west shore of Klakas Inlet just north of the map area. Herreid and others (1978) show this fault as a southwest-dipping thrust that separates Descon Formation volcanic rocks to the south from marine clastic strata to the north. On the ridge west of Klakas Inlet the fault is overlain by Early Devonian clastic strata (Herreid and others, 1978). This fault is significant because it demonstrates that northern Klakas Inlet is not controlled by a major strike-slip fault (as suggested by Herreid and others, 1978), and that the strata along Klakas Inlet do not belong to a continuous, south-dipping section (as suggested by Eberlein and others, 1983).

Anchor Island fault

The Anchor Island fault is recognized as a wide zone of penetrative brecciation in Ordovician rocks south of Tah Island, in the southern Klakas Inlet region, and along the east shore of Kassa Inlet. Along the shoreline south of Tah Island, Ordovician volcanic rocks are strongly brecciated, moderately foliated, and dip at moderate angles to the northeast. In southern Klakas Inlet, Ordovician volcanic, sedimentary, and intrusive rocks are strongly brecciated, locally semischistose, and pervasively altered in a zone which is several kilometers wide. These rocks are mapped as the complex of Klakas Inlet (DSk) where deformation and alteration have obliterated their primary intrusive and stratigraphic relations. The zone of brecciation extends northwestward to the east shore of Kassa Inlet, where it is truncated by the Keete Inlet fault. The distribution of brecciation along the Anchor Island fault zone is shown in a general fashion with small "x"'s, and the distribution and structural grain of the semischistose fabric is shown with a wavy symbol on Plate 1.

The age of movement on this fault zone is constrained by the Middle and Late(?) Ordovician age of rocks which are deformed, and by stratigraphic relations with Early Devonian sedimentary breccia (Dbx) in southern Klakas Inlet. As described above, this breccia unconformably overlies rocks belonging to the complex of Klakas Inlet (DSk), is moderately deformed, and is overlain by nondeformed middle Early Devonian sandstone, mudstone, and shale (Dms). These relations suggest that the fault zone moved prior to, during, and perhaps after deposition of the sedimentary breccia, which is at least in part of Early Devonian age. Regional relations suggest that movement along the fault zone began after middle Early Silurian time (Gehrels and Saleeby, in review).

Keete Inlet fault

The Keete Inlet fault is a major structural and stratigraphic boundary in the study area and to the north on south-central Prince of Wales Island (Redman, 1981). In the map area the fault dips moderately to the northeast and juxtaposes Early Silurian(?) and Ordovician stratified and intrusive rocks against rocks in both the Wales metamorphic suite (pmOw) and the complex of Ruth Bay (Or). The fault continues north of the study area through Keete Inlet (Herreid and others, 1978) and then swings eastward toward the North Arm of Moira Sound (Redman, 1981). South of Klakas Island the fault bends westward into Cordova Bay and then turns south toward Dixon Entrance.

The age of movement on the Keete Inlet fault is constrained as post-middle Early Devonian and pre-mid-Cretaceous because the fault cuts Devonian strata and is intruded by Cretaceous granodiorite. The sinuosity of the fault in and north of the study area combined with its regional juxtaposition of younger rocks over older rocks suggests that it is a normal fault (Gehrels and Saleeby, in review) rather than a thrust fault (as suggested by Herreid and others, 1978, and Redman, 1981). Eastward movement on this fault is suggested by my interpretation that rocks in the complex of Ruth Bay (Or) are deeper-level equivalents of Early Silurian(?) and Ordovician rocks in the upper plate to the east. Regional evidence for a latest Paleozoic(?)-Triassic rifting event in the area (Gehrels and Saleeeby, in review) indicates that the fault may have moved during latest Paleozoic(?)-Triassic time.

Strike-slip faults

Northwest- to north-northwest-striking faults

The dominant set of strike-slip faults in the area consists of anastomosing, curviplanar, and structurally interconnected northwest- to north-northwest-striking faults that have a left-lateral sense of displacement. These faults control the major northwest-trending inlets and valleys on the western and southern parts of the island. Where exposed, the fault zones generally consist of several parallel strands separated by moderately deformed and brecciated rocks. In stratified rocks the zone of deformation along each strand is generally several meters wide, and is manifest as a steeply dipping phyllonitic foliation within which protolith features are disrupted and (or) highly deformed. Intrusive rocks generally show intense cataclastic brecciation in a zone up to several tens of meters in width, and a wider zone in which the rocks are fractured and cut by many narrow shear zones. As seen on Plate 1, the Max Cove and the Tah Bay-Klinkwan Cove-Nichols Bay fault zones consist of several subparallel strands, and are structurally connected by the Biscuit Lagoon fault and by the Hunter Creek, Feikert Claims, and Billy Claims faults.

These faults are known to have predominantly strike-slip displacement because slickenside striae exposed in the fault zones plunge consistently within 20° of horizontal. Offsets of outcrop-scale stratigraphic markers within these zones are common, but indicate both right-lateral and left-lateral senses of offset. Regional offsets of contacts and map units in many areas indicate that the faults have predominantly left-lateral displacement. Along the Max Cove fault in Max Cove the leucogranodiorite-diorite contact near the end of the peninsula and the base of the Devonian section in several areas are offset approximately 1 km in a left-lateral sense. The Klinkwan Cove fault offsets the Devonian sandstone-conglomerate contact by up to a kilometer in a left-lateral sense south of Klinkwan Cove. In Hunter Bay and for over 8 km to the south the Klinkwan Cove fault juxtaposes Devonian strata against pre-Devonian rocks: this apparent offset, combined with shallow-plunging slickenside striae along the fault, indicates at least a component of left-lateral displacement.

The Nichols Bay fault appears to have a larger amount of strike-slip displacement, as indicated by the juxtaposition of different Early

Silurian and Ordovician rocks in Nichols Bay. Assuming that the volcanic and sedimentary rocks between Nichols Lake and northern Nichols Bay are the offset equivalents of volcanic and sedimentary rocks in southern Nichols Bay, the horizontal separation is approximately 7 to 10 km in a left-lateral sense. This sense and amount of offset are also indicated by the separation of the large mass of quartz-porphyritic granodiorite (Oqgd) across the Nichols Bay fault. The lack of syenitic rocks on the southwest side of the fault in Nichols Bay constrains the amount of displacement to greater than 4 km. Patterns of dioritic rocks in Hessa Inlet are most consistent with approximately 4 km of left-slip on the Nichols Bay zone. The stratigraphic relations along the Klinkwan Cove fault are such that most of the displacement on the Nichols Bay zone must continue northward along the Tah Bay fault, the amount of displacement along which is not constrained.

Dioritic intrusive rocks in Hessa Inlet appear to be offset by a kilometer or more in a left-lateral sense along the Biscuit Lagoon fault. The distribution of Early Silurian(?) and Ordovician stratified rocks along the Feikert Claims and Billy Claims faults is consistent with several kilometers of cumulative left-lateral displacement.

The age of displacement on these faults is constrained by offsets of Devonian strata and by the observation that tabular basaltic dikes of probable pre-mid-Cretaceous age locally intrude the fault zones. This is consistent with the observation that the Max Cove, Klinkwan Cove, and Tah Bay faults do not offset the the Keete Inlet fault, which is known to have pre-mid-Cretaceous displacement. Faults in this set cut the north- to north-northeast-striking faults in the area, and are apparently cut by the Cape Chacon fault.

North- to north-northeast-striking faults

These faults control the major north- to north-northeast-trending inlets and valleys along the southern shore of Prince of Wales Island. The Buschmann Pass and Hessa Narrows faults have a structural style similar to the northwest- to north-northwest-striking faults described above, but the other faults in this set have not been studied in detail. Abundant shallow-plunging slickenside striae on the Buschmann Pass and Hessa Narrows faults indicate that they have predominantly strike-slip displacement. A right-lateral sense of displacement on these and most other faults in the set is indicated by the horizontal separation of both Devonian units and the basal contact of the Devonian section. The amounts of right-lateral separation across the major faults are as follows: Buschmann Pass fault north of Buschmann Pass = 4 to 7 km; Hessa Narrows fault = 1 km; Brownson Bay fault = approx. 500 m; Surf Point fault = a few hundred meters(?); and Bert Millar Cutoff fault = over 1 km. In contrast, the Nichols Lake fault has a separation of between .5 and 1.5 km in a left-lateral sense. The only fault belonging to this set that has been recognized on the northeast side of the Nichols Bay fault is the Alice Claims fault, which has a kilometer or less of right-lateral displacement. Other faults belonging to this set probably exist northeast of the Nichols Bay fault but have not been recognized due to lack of geologic control.

Faults belonging to this set cut Devonian strata and are cut by the northwest- to north-northwest-striking faults described above, which are interpreted to have post-middle Early Devonian, pre-mid-Cretaceous displacement.

Cape Chacon fault

This fault is a major north-striking structure that extends from west of Cape Chacon to Kendrick Bay. In McLean Arm it apparently offsets both syenitic rocks and the Max Cove fault by several hundred meters in a right-lateral sense. This indicates that it is younger than the two main sets of strike-slip faults in the area, which moved between middle Early Devonian and mid-Cretaceous time.

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APPENDIX 2

U-Pb GEOCHRONOLOGIC METHODS

LABORATORY METHODS

Zircon separation

Geochronologic samples collected from Annette, Gravina, Duke, and southern Prince of Wales Islands ranged from 30 to 60 kg of homogeneous rock that, for each sample, were collected from a single outcrop. Zircons were separated from the rock mechanically and with heavy liquids, sieved to less than 165 microns, and separated magnetically on a Frantz isodynamic separator. The least-magnetic fractions (side-slopes of 5° or less with a forward slope of 10° to 15°) were acid-washed in warm concentrated HNO_3 for 30 minutes, rinsed several times in warm distilled water, and rinsed in acetone. Depending on the amount of zircon, the population was sieved into various size-fractions using silk screens that were discarded after each use. Each fraction analyzed was then hand-picked under a binocular microscope two times, resulting in a population consisting of greater than 99% zircon. Zircon grains with inclusions, fractures, attached fragments of other grains, etc. were removed from the population where possible. The zircons were weighed (to within 2%), put in a previously cleaned teflon capsule, and rinsed in warm concentrated HNO_3 for 10 minutes.

Zircon dissolution and isolation of Pb and U

Our procedure for dissolving zircon and isolating Pb and U is similar to that described by Krogh (1973). Significant deviations from this procedure include:

1) Rather than adding the uranium tracer prior to redissolving the

zircon-fluoride precipitate in HCl, we use a mixed ^{235}U-^{208}Pb tracer which is added after the HCl solution has been aliquoted. The main drawback to this procedure is the possibility of non-equilibration of the uranium tracer and the sample solution, but we have not seen evidence of this in our U-Pb concentration data.

2) Uranium was removed from the columns with either 2N HCl-1N HF or 1N HBr, rather than H_2O.

Mass spectrometry

Our isotopic analyses were conducted on a 30.48 cm, 60°-sector mass spectrometer with an accelerating voltage of 10 KV. Pb was loaded with H_3PO_4 and silica gel onto a previously outgassed rhenium filament and pre-heated in a laminar-flow hood. Isotopic data were collected during a three- to four-hour period during which $^{206}Pb/^{208}Pb$, and $^{206}Pb/^{207}Pb$, were measured on a Faraday cup collector and $^{207}Pb/^{204}Pb$ was measured with a secondary electron multiplier. Filament temperatures averaged 1075° C. During most runs the signal current averaged 10^{-11} amps (10^{11} ohm resistor) of ^{206}Pb and decreased slightly during the course of the run. Uranium was loaded with H_3PO_4 and graphite and analyzed as uranium metal. Filament temperatures were approximately 1400° C. Uranium isotopic data were generally collected during a period of two to three hours, with signal strength ranging from 10^{-11} to 10^{-12} amps of ^{238}U (10^{11} ohm resistor). 30 to 50 sets of isotope ratios were measured for each sample, with each measurement representing a 10 second integration.

DATA ANALYSIS

U-Pb concentrations and apparent dates and their uncertainties and error correlations have been calculated using a program written by Ken Ludwig (1983). The following values have been used in adjusting the measured isotopic ratios and in the calculations:

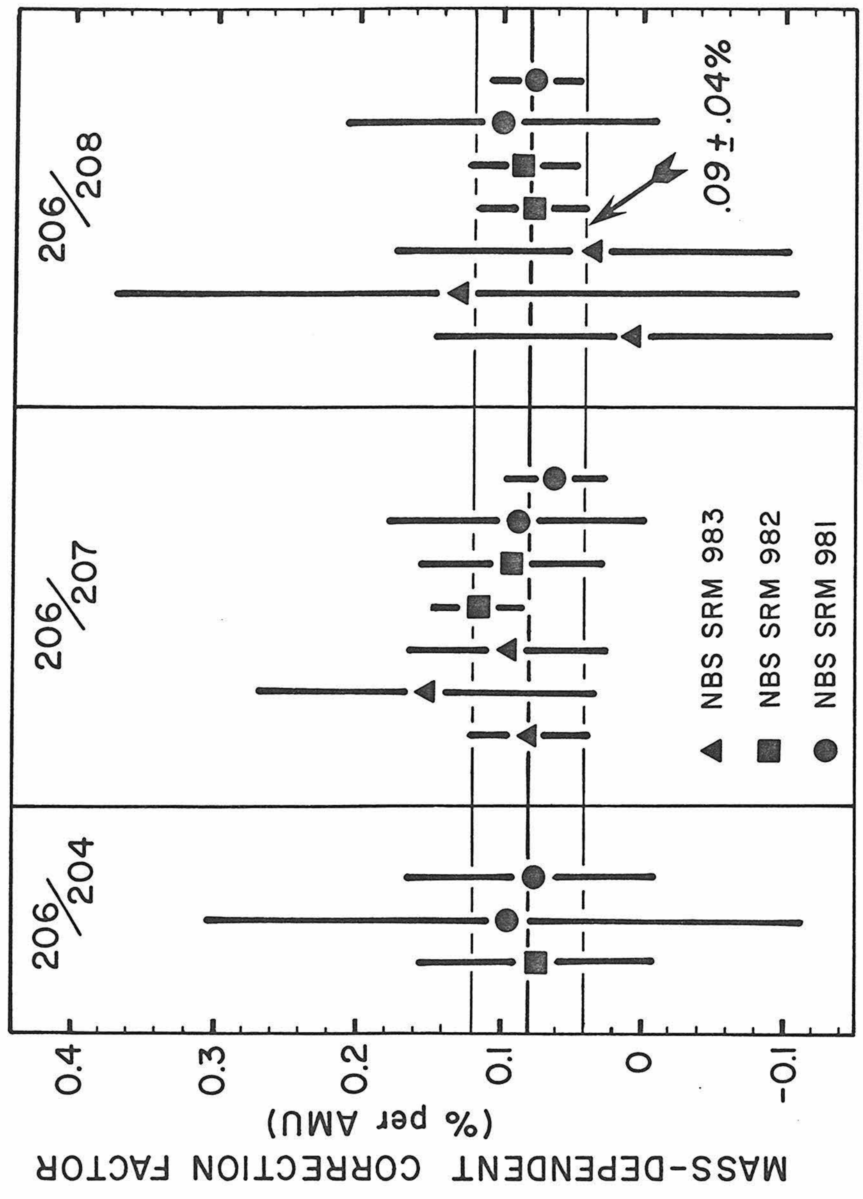
1) Mass-dependent correction factors of .09% ± .04% (per AMU) and 0.12 ± .04% (per AMU) have been applied to lead and uranium ratios, respectively. These factors were determined by replicate analyses of NBS SRM 981, 982, 983 for Pb (Fig. A2-1) and U-500 and U-930 for U. Analyses of the Pb standards also demonstrate that there are no significant systematic errors in the measurement of any particular isotope ratio.

2) Based on repeated measurements, our tracer is assigned a composition of: $^{208}Pb/^{206}Pb$ = 1848 ± 1.8%, $^{208}Pb/^{207}Pb$ = 4025 ± 6%, $^{208}Pb/^{204}Pb$ = 94,000 ± 12%, and $^{235}U/^{238}U$ = 17.330 ± .3%. The concentrations of Pb and U have been determined by Dr. L.T. Silver as: 3.6997×10^{-9} ± .3% moles Pb per gram of tracer and 1.9804×10^{-8} ± .3% moles U per gram of tracer (uncertainties are 2-sigma). Our spiked aliquots from Prince of Wales Island generally have $^{208}Pb/^{206}Pb$ = .45 ± .18 and $^{238}U/^{235}U$ = .10 ± .06, and samples from Annette, Gravina, and Duke Islands have $^{208}Pb/^{206}Pb$ = .93 ± .86 and $^{238}U/^{235}U$ = .24 ± .22 (uncertainties are 1-sigma).

3) The $^{207}Pb/^{206}Pb$ ratio of the spiked aliquot has been adjusted for the ^{206}Pb and the ^{207}Pb added with the spike.

4) The isotopic composition of the unspiked aliquot has been adjusted for 0.04 ±.02 ng blank lead with the following composition: $^{206}Pb/^{204}Pb$ = 18.78 ± 0.30; $^{207}Pb/^{204}Pb$ = 15.61 ± 0.22; and $^{208}Pb/^{204}Pb$ = 38.5 ± .60. The amount of blank Pb has been determined by isotope dilution analysis of a dissolution and isolation procedure conducted without zircon. I have determined the blank composition through isotopic analysis of laboratory water, acids, and dust particles.

Figure A2-1. Analyses of NBS lead standards showing their measured isotopic composition relative to values certified by the National Bureau of Standards (except for the $^{206}Pb/^{204}Pb$ ratio of NBS SRM 981, which is adapted from Hamelin and others, 1985). The uncertainties show the precision of each analysis at the 95% level, and do not incorporate the uncertainty in the certified isotopic composition. Based on these analyses, we assign a mass-dependent correction factor of .09 ± .04 % per AMU to our measured lead isotopic analyses. Systematic errors in any particular isotope ratio are not apparent in the data set.


206/204
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.09 ± .04%
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NBS SRM 982
NBS SRM 981
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0.3
0.2
0.1
0
-0.1
MASS-DEPENDENT CORRECTION FACTOR
(% per AMU)

5) The isotopic composition of U in the spiked aliquot has been adjusted for .07 ± .05 ng blank U, which has been determined by the same procedure as the Pb blank.

6) The isotopic composition of the spiked aliquot has been adjusted for blank Pb by balancing the $^{206}Pb/^{207}Pb$ of the spiked and unspiked aliquots. Additional blank Pb in the spiked aliquot is assigned the composition cited above.

7) Common Pb remaining after correction for blank Pb is interpreted to be initial Pb and is assigned a composition of: $^{206}Pb/^{204}Pb = 18.0 \pm 1.5$; $^{207}Pb/^{204}Pb = 15.59 \pm 0.4$; and $^{208}Pb/^{204}Pb = 37.8 \pm 2.0$ for Paleozoic samples and 18.2 ± 1.5, 15.57 ± 0.40, and 38.0 ± 2.0 for Triassic samples. (Values interpreted from Doe and Zartman, 1979).

8) Constants used: $\lambda^{238}U = 1.55125 \times 10^{-10}$; $\lambda^{235}U = 9.8485 \times 10^{-10}$; and $^{238}U/^{235}U$ (atomic) = 137.88 (from Steiger and Jager, 1977).

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APPENDIX 3

GEOLOGIC MAP OF SOUTHEASTERN ALASKA

INTRODUCTION

Southeastern Alaska is underlain by sedimentary, volcanic, intrusive, and metamorphic rocks of Quaternary to Cambrian, and probable Proterozoic age. These rocks have been classified on the geologic map and in this text into units that emphasize the regional distribution of lithically similar and generally coeval geologic units. The distribution, age, stratigraphic or intrusive relations, and metamorphic and structural characteristics of the map units are described below in the Description of Map Units, and the temporal relations between units are shown on the Correlation of Map Units (Plate 2). Figure A3-1 shows the primary sources of information used in compiling the geologic map.

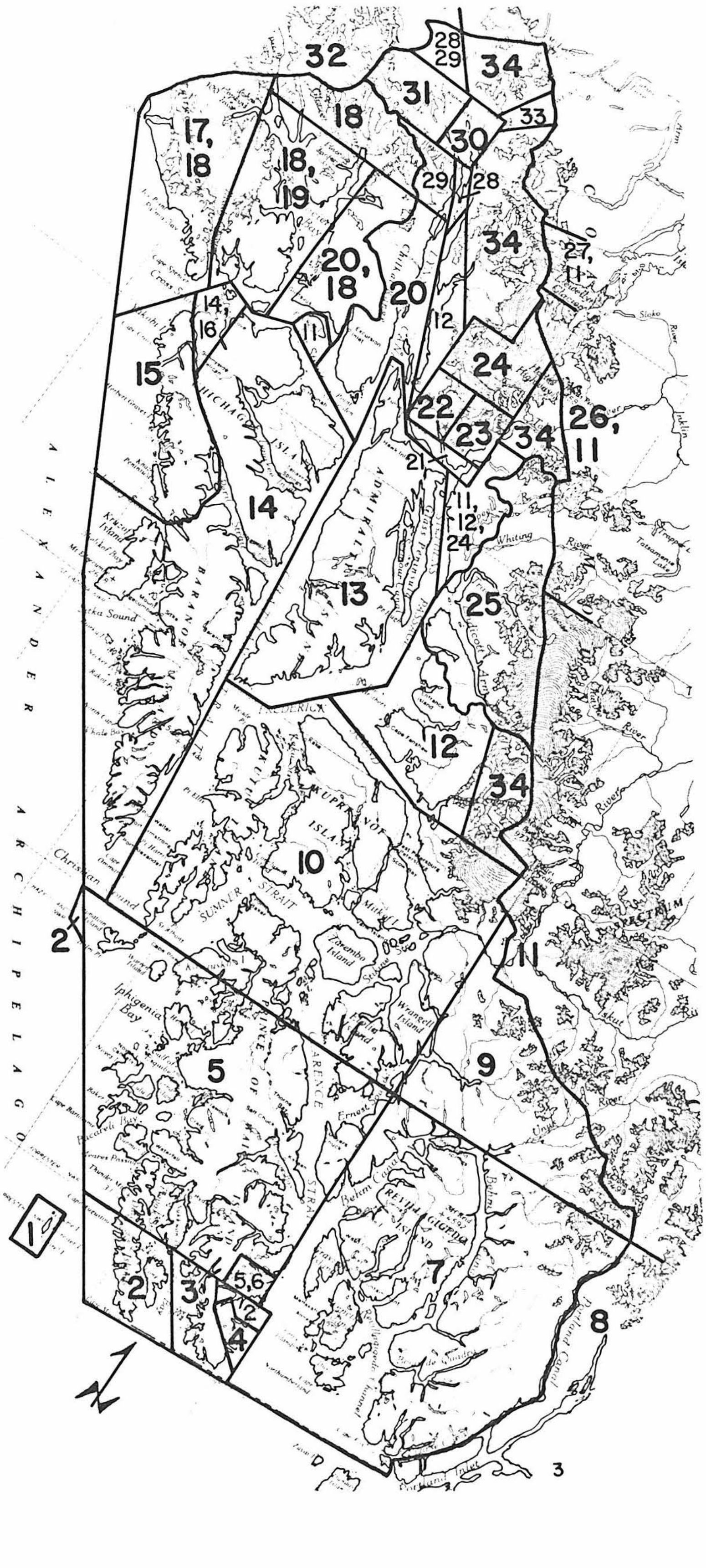
The rocks of southeastern Alaska have been classified into four main categories of map units, including: LITHOSTRATIGRAPHIC UNITS, UNDIVIDED UNITS, LITHODEMIC UNITS, and INTRUSIVE UNITS. These types of units are defined as follows:

LITHOSTRATIGRAPHIC UNITS consist of sedimentary and volcanic rocks, and their metamorphosed and (or) deformed equivalents, that generally conform to the Law of Superposition (North American Commission on Stratigraphic Nomenclature, 1983). Strata that belong to a lithostratigraphic unit are similar in general lithic type, stratigraphic position, and age of deposition, and are, unless otherwise noted, interpreted to be correlative. Such strata comprise formations or groups, lithic components of formations or groups, or unnamed geologic units. We have classified these strata according to their

Figure A3-1. Index map of southeastern Alaska showing the primary sources of information used in compiling the geologic map.

KEY TO SOURCES OF INFORMATION

1 - Clark and others, 1971
2 - G.E. Gehrels, unpub. mapping
3 - Gehrels and others, 1983
4 - MacKevett, 1963
5 - Eberlein and others, 1983
6 - Redman, 1981
7 - Berg and others, 1978
8 - Hutchison and others, 1979
9 - Elliott and Koch, 1981
10 - Brew and others, 1984
11 - Souther and others, 1979
12 - Buddington and Chapin, 1929
13 - Lathram and others, 1965
14 - Loney and others, 1975
15 - Johnson and Karl, 1982
16 - Decker and Plafker, 1982
17 - G. Plafker and T. Hudson, unpub. mapping
18 - Brew and others, 1978
19 - Rossman, 1963
20 - Lathram and others, 1959
21 - Ford and Brew, 1977
22 - Ford and Brew, 1973
23 - Brew and Ford, 1977
24 - Brew and Morrell, 1980
25 - Brew and Grybeck, 1984
26 - Souther, 1971
27 - Werner, 1978
28 - Plafker and Hudson, 1980
29 - Robertson, 1959
30 - Redman and others, 1984
31 - MacKevett and others, 1974
32 - Campbell and Dodds, 1983
33 - F. Barker and J. Arth, unpub. mapping
34 - unmapped



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ALEXANDER ARCHIPELAGO
SUMNER STRAIT
Iphigenia Bay
Sitka Sound
Zarembo Island
Wrangell Island

general lithic type(s), rather than by specific or local rock types, or by formal stratigraphic nomenclature (such as formation names), in an effort to emphasize the regional distribution of correlative geologic units. The general lithic types (and corresponding unit symbols) are listed in the Definition of Map Units (Plate 2). An example of a lithostratigraphic unit is "Sc," which consists of Silurian carbonate rocks that are the predominant rock-type in the Heceta Limestone, Kuiu Limestone, Kennel Creek Limestone, Willoughby Limestone, Pyramid Peak Limestone, and unnamed limestone, and are a subordinate component in units consisting predominantly of Silurian clastic strata (Bay of Pillars Formation, Point Augusta Formation, Tidal Formation, Rendu Formation, and unnamed units).

UNDIVIDED UNITS consist of sedimentary and volcanic rocks, and their metamorphosed and (or) deformed equivalents, that have not been assigned to lithostratigraphic or lithodemic units because their age, stratigraphic relations, and (or) degree of deformation or metamorphism are uncertain, or because they occur in geologic units that grade from relatively intact stratigraphic sequences to metamorphic or structural complexes. Most undivided map units probably consist of, or were derived from, several lithostratigraphic units. These units are classified according to the probable minimum and maximum depositional age and the main lithic type(s) of their constituent strata. An example of an undivided map unit is "R̄Osv," which consists of metasedimentary and metavolcanic rocks of Triassic, Permian, Devonian, and probably Silurian and Ordovician age.

LITHODEMIC UNITS comprise rocks that have been regionally metamorphosed and (or) deformed to a degree that their primary stratigraphic relations are not preserved (i.e., the rocks do not conform to the Law of Superposition) (North American Commission on Stratigraphic Nomenclature, 1983). Two types of lithodemic units have been recognized in southeastern Alaska:

Metamorphic complexes, which are denoted by a lower case "m" preceding the lithic term in the unit symbol. These units are classified according to the probable depositional age and lithic type(s) of the rocks included. The metamorphic rock types and the nature and age of metamorphism are discussed in the Description of Map Units. For example, unit "pTmsv" represents a metamorphic complex derived from sedimentary and volcanic strata of pre-Tertiary age.

Structural complexes, including a melange (denoted by a lower case "m" at the end of the unit symbol), and a complex of regionally disrupted strata (denoted by a lower-case "d" at the end of the unit symbol). Map units in the melange are classified according to the main lithic type(s) of the constituent strata and the age of formation of the melange (rather than the age of the constituent strata). The nature of the melange, and the age of the blocks and the matrix are discussed in the Description of Map Units. For example, unit "Ksvm" represents sedimentary and volcanic rocks (and their metamorphic equivalents) belonging to a melange that formed during Cretaceous time. The complex of disrupted strata has some characteristics of a melange, but the relations between blocks and matrix material and the degree of disruption in the matrix are uncertain. Map units in this disrupted complex are classified according to the main lithic type(s) of the constituent strata, and the interpreted age of formation of the complex. The nature of the complex is described in the Description of Map Units. For example, unit "KJsd" represents a complex of regionally disrupted clastic sedimentary rocks that formed during Cretaceous and Jurassic time.

INTRUSIVE UNITS consist of intrusive rocks that are classified into map units according to their age of emplacement [following the Decade of North American Geology Time Scale (Palmer, 1983)], and their predominant rock types (following the IUGS classification scheme [Streckheisen, 1976)]. The categories of intrusive rock types (and their corresponding unit symbols) are listed in the Definition of Map Units (Plate 2).

Areas in which intrusive rocks have not been mapped separately from their country rocks are shown with a stipple pattern on the geologic map (Plate 2). These types of mixed rocks are referred to as "migmatites" in some of the references cited below. Where younger intrusive rock predominates, the rocks are assigned to an intrusive unit and the stipple pattern indicates that a significant proportion of the area is underlain by older country rocks. Intrusive unit "TKg," for example, is patterned where the intrusive bodies contain abundant unmapped inclusions or pendants of metamorphic rocks elsewhere assigned to unit "pTmsv." Where country rocks predominate, the rocks are assigned to the map unit of the country rocks, and the stipple pattern indicates that a significant proportion of the area is underlain by younger intrusive rock. On Admiralty Island, for example, undivided unit "TROsv" is patterned where it contains a significant proportion of younger intrusive rocks.

DESCRIPTION OF MAP UNITS

Lithostratigraphic units

Qs SURFICIAL DEPOSITS (QUATERNARY) -- Undivided lacustrine, fluvial, colluvial, glacial, beach, and marine deposits that occur in many areas of southeastern Alaska.

QTs SEDIMENTARY ROCKS (QUATERNARY? AND TERTIARY) -- Pleistocene(?) to Miocene marine mudstone, siltstone, sandstone, and conglomerate (Yakataga Formation) northwest of Cross Sound (Plafker and Addicott, 1976; George Plafker, written commun., 1984).

QTv VOLCANIC ROCKS (QUATERNARY AND TERTIARY) -- Unnamed volcanic rocks of basaltic to rhyolitic composition on islands west of Prince of Wales Island (Eberlein and others, 1983), on Revillagigedo Island and the mainland to the east (Berg and others, 1978), near the Alaska-British Columbia border east of Wrangell Island (Elliott and Koch, 1981), and on Zarembo, Kuiu, and Kupreanof Islands (Brew and others, 1984); and the Edgecumbe Volcanics (basalt, andesite, and dacite) on Kruzof Island (Loney and others, 1975). Rocks of Holocene age have been recognized east of Wrangell Island (Elliott and others, 1981) and on Kruzof Island (Loney and others, 1975), and basaltic rocks of Holocene and (or) Pleistocene age occur on southern Kupreanof Island (Brew and others, 1984), and perhaps at other localities. Most of the volcanic rocks in this unit on Zarembo, Kupreanof, and Kuiu Islands are Tertiary in age (Brew and others, 1984) and are probably correlative with volcanic rocks in unit "Tv."

Ts SEDIMENTARY ROCKS (TERTIARY) -- Nonmarine sandstone, shale, and conglomerate (Kootznahoo Formation) in the Zarembo-Kuiu Islands region (Brew and others, 1984) and on Admiralty Island (Lathram and others, 1965), and marine calcareous sandstone and siltstone (Topsy Formation) of Miocene and Oligocene age northwest of Cross Sound (Brew and others, 1978). The age of the Kootznahoo Formation is reported to be Eocene and Paleocene in the Zarembo-Kuiu Islands region and Miocene to Eocene on Admiralty Island (Brew and others, 1984).

Tv VOLCANIC ROCKS (TERTIARY) -- Oligocene and Eocene basalt and andesite (Admiralty Island Volcanics) on Admiralty Island (Lathram and others, 1965); unnamed Oligocene basaltic rocks on islands in Icy Strait (Plafker and others, 1982); and basaltic rocks (Cenotaph Volcanics) of Miocene and Oligocene age along the coast northwest of Cross Sound (Brew and others, 1978). The volcanic rocks northwest of Cross Sound are interbedded with clastic strata belonging to the Topsy Formation (unit Ts) (Brew and others, 1978), and the Admiralty Island Volcanics contain clastic strata that may be in part correlative with the Kootznahoo Formation (unit Ts) (Lathram and others, 1965).

Tsv SEDIMENTARY AND VOLCANIC ROCKS (TERTIARY) -- Unnamed carbonaceous shale, sandstone, and conglomerate of probable Paleogene age, and subordinate Oligocene volcanic rocks near Haines (George Plafker, written commun., 1984; Robertson, 1959; Campbell and Dodds, 1983).

Ks SEDIMENTARY ROCKS (CRETACEOUS) -- Sandstone and mudstone turbidites and subordinate conglomerate (Sitka Graywacke) on Baranof, Chichagof, Kruzof, and Yakobi Islands (Loney and others, 1975; Johnson and Karl, 1982; Decker, 1980), and regionally metamorphosed carbonaceous siltstone, volcanogenic graywacke, mudstone, and minor conglomerate (Valdez Group) north of Cross Sound (Brew and others, 1978; George Plafker, written commun., 1984). These strata are interpreted to be deep-marine trench, slope basin, and fan deposits. The Sitka Graywacke is moderately deformed and disrupted, regionally metamorphosed as high as greenschist facies in some areas, and thermally upgraded to hornblende hornfels facies locally (Decker and others, 1979). Common rock types in the metamorphosed regions south of Cross Sound include metagraywacke and argillite. Rocks north of Cross Sound are regionally metamorphosed from sub-greenschist facies to as high as greenschist to amphibolite facies, producing rock types of graywacke semischist, phyllite, slate, and layered schist, semischist, and gneiss (Brew and others, 1978). These metasedimentary rocks are correlated with strata in the Valdez Group to the northwest by lithic similarity (Brew and Morrell, 1979) and by stratigraphic continuity (Plafker and Campbell, 1979; Campbell and Dodds, 1983). The Sitka Graywacke is unfossiliferous: its minimum age is constrained by an Eocene pluton on Baranof Island (Loney and others, 1975). The Sitka Graywacke and the rocks north of Cross Sound are considered to be Cretaceous in age based on correlation with lithically similar rocks in the Yakutat Group and Valdez Group (Plafker and others, 1977; Brew and Morrell, 1979).

Kv VOLCANIC ROCKS (CRETACEOUS) -- Basalt and basaltic tuff north of Cross Sound that have been deformed and regionally metamorphosed as high as greenschist to amphibolite facies (Brew and others, 1978; George Plafker, written commun., 1984). Common rock types include schist, gneiss, and amphibolite (Brew and others, 1978). Correlation with volcanic rocks in the Valdez Group to the northwest indicates a Cretaceous age (Plafker and Campbell, 1979; Brew and Morrell, 1979).

Ksv SEDIMENTARY AND VOLCANIC ROCKS (CRETACEOUS) -- Undivided metasedimentary rocks that are correlative and continuous with strata in unit "Ks," and metavolcanic rocks that are correlative with rocks in unit "Kv." This unit occurs north of Cross Sound (Brew and others, 1978; Brew and Morrell, 1979; Plafker and Campbell, 1979). The age constraints, lithic types, metamorphic characteristics, and regional correlations are similar to those described above for units "Ks" and "Kv."

KJs SEDIMENTARY ROCKS (EARLY CRETACEOUS, AND LATE AND MIDDLE? JURASSIC) -- Marine graywacke, mudstone, subordinate conglomerate and andesitic to basaltic volcanic rocks, minor limestone, and the regionally metamorphosed and deformed equivalents of these strata. Metamorphic grade generally increases from sub-greenschist facies or non-metamorphosed on the southwest, to greenschist and in some areas amphibolite facies toward the northeast. Rock types in the higher grade parts of the unit are primarily phyllite, schist, and gneiss. The strata are queried on the geologic map where metamorphism and deformation, or lack of stratigraphic information, make correlation with sedimentary rocks of known Cretaceous and Jurassic age uncertain. Regional metamorphism and deformation occurred after Early Cretaceous time, and prior to the deposition of the Kootznahoo Formation (Ts) of Paleocene to Miocene age (Buddington and Chapin, 1929). Fossils recovered range in age from Late and perhaps Middle Jurassic, to Albian and perhaps Cenomanian (Berg and others, 1972; Brew and others, 1984). This unit consists of: the Gravina Island Formation and unnamed strata on Gravina Island (Berg and others, 1978), unnamed strata on Annette, Revillagigedo, and southern Etolin Islands, and adjacent parts of the mainland (Berg and others, 1978; Eberlein and others, 1983); the Seymour

Canal Formation and its metamorphic equivalents near Etolin-Kupreanof Islands and adjacent areas of the mainland (Buddington and Chapin, 1929; Brew and others, 1984); an unnamed unit of sandstone and mudstone in Keku Strait (Muffler, 1967); the Seymour Canal Formation on Admiralty Island (Lathram and others, 1965), northern Kupreanof Island (Brew and others, 1984), and perhaps near Cape Fanshaw (Muffler, 1967; Buddington and Chapin, 1929); the Symonds and Shelter Formations on islands in southern Lynn Canal (Barker, 1957); the Treadwell Slate and unnamed strata near Juneau (Ford and Brew, 1973 and 1977); part of the Berners Formation (now obsolete) northwest of Juneau (Knopf, 1911 and 1912; Martin, 1926; Buddington and Chapin, 1929; Redman, 1984a); and argillite, siltstone, and sandstone of probable Cretaceous and Jurassic age south of Haines (George Plafker, written commun., 1984).

KJv VOLCANIC ROCKS (EARLY CRETACEOUS, AND LATE AND MIDDLE? JURASSIC) -- Andesitic to basaltic flows, flow breccia, agglomerate, and tuff (generally with conspicuous clinopyroxene phenocrysts), subordinate graywacke and mudstone, and the regionally metamorphosed and deformed equivalents of these strata. Greenschist-facies metamorphism in some areas has yielded common rock types of greenstone and greenschist. Age is constrained by intertonguing stratigraphic relations with the Cretaceous and Jurassic sedimentary rocks (KJs) described above. Geologic and geochemical considerations suggest that these volcanic rocks are genetically related to the Early Cretaceous zoned ultramafic bodies in unit "Kum" (Irvine, 1974), and possibly to the Early Cretaceous and (or) Jurassic diorite and gabbro in units "KJd" and "KJgb" (Berg and others, 1978). This unit consists of: volcanic rocks in the Gravina Island Formation on Gravina Island (Berg and others, 1978); unnamed rocks on Annette and Revillagigedo Islands and adjacent areas of the mainland (Berg and others, 1978; Eberlein and others, 1983); unnamed rocks in the Etolin-Kupreanof Islands region and adjacent areas of the mainland (Brew and others, 1984; Buddington and Chapin, 1929); the Douglas Island Volcanics and the Brothers Volcanics on and near Admiralty Island (Lathram and others, 1965) and in the Juneau area (Ford and Brew, 1973 and 1977); and part of the Berners Formation (now obsolete) northwest of Juneau (Knopf, 1912; Martin, 1926).

$\boxed{\text{KJsv}}$ SEDIMENTARY AND VOLCANIC ROCKS (EARLY CRETACEOUS, AND LATE AND MIDDLE? JURASSIC) -- Moderately deformed and metamorphosed graywacke, mudstone, sandstone, and andesitic to basaltic rocks near Juneau that are the undifferentiated equivalents of the Treadwell Slate (KJs) and the Douglas Island Volcanics (KJv) (Ford and Brew, 1973 and 1977).

$\boxed{\text{R̄s}}$ SEDIMENTARY ROCKS (LATE AND MIDDLE TRIASSIC) -- Unnamed, moderately deformed and metamorphosed graphitic limestone and slate of late Ladinian (latest Middle Triassic) age on Revillagigedo Island (Silberling and others, 1982); Late Triassic conglomerate, limestone, and calcareous siltstone and sandstone (Burnt Island Conglomerate) west of Etolin Island (Karl, 1984) and on northwestern Kupreanof Island (Muffler, 1967); and unnamed silty limestone of Late Triassic age on small islands between Etolin Island and northern Prince of Wales Island (Brew and others, 1984). Based on regional stratigraphic relations and the slightly older age, Silberling and others (1982) follow the suggestion of Berg and others (1972) that these strata on Revillagigedo Island are not correlative with other Triassic sedimentary rocks assigned to this unit or unit "R̄sv."

$\boxed{\text{ŦRv}}$ VOLCANIC ROCKS (RHYOLITIC TO BASALTIC) (LATE TRIASSIC) -- Basaltic rocks are generally pillow flows, pillow breccia, and breccia, except near Haines where massive flows predominate; rhyolitic rocks include tuff with calcareous interbeds, flow breccia, and banded ash-flow tuff; and subordinate andesitic rocks include breccia and aquagene tuff. Rocks in most areas are moderately recrystallized and deformed. Fossils recovered from near the top of the section near Haines are Karnian in age (Plafker and Hudson, 1980), and fauna from interbedded rocks or from conformable adjacent strata in other areas indicate deposition during Karnian and Norian time. Based on regional stratigraphic relations, Plafker and Hudson (1980) argue that the strata near Haines are probably not correlative with the other volcanic rocks in this unit. This unit consists of: unnamed rhyolite on Annette Island and the Puppets Formation (rhyolite) on Gravina Island (Berg, 1982); the Chapin Peak Formation (basalt) on Gravina Island (Berg and others, 1978); the Keku Volcanics (rhyolite and subordinate basalt) and the Hound Island Volcanics (basalt and andesite) in Keku Strait (Muffler, 1967; Brew and others, 1984); and unnamed basalt near Haines (Plafker and Hudson, 1980; Redman and others, 1984; Robertson, 1959).

$\boxed{\text{ŦRsv}}$ SEDIMENTARY AND BASALTIC ROCKS (LATE TRIASSIC) -- Black shale, shaly limestone, siltstone, sandstone, conglomerate, and limestone; felsic, intermediate, and basaltic flows; and minor black chert. Rocks range from relatively unmetamorphosed to as high as greenschist facies. Common rock types in the higher grade areas include phyllite, slate, semischist, marble, greenschist, and greenstone. This unit consists of unnamed strata on Annette and Gravina Islands (Berg and others, 1978); the Nehenta Formation on Gravina Island (Berg and others, 1978); unnamed strata interpreted to occur as blocks in a disrupted sedimentary unit (KJsd) on Kupreanof and Zarembo Islands (D.A. Brew, *in* U.S. Geological Survey, 1982; Brew and others, 1984; Berg and Grybeck, 1980); and the part of the Hyd Formation on Admiralty Island that has been reliably separated from the Cannery Formation (Ps) (Lathram and others, 1965).

$\boxed{\text{TRc}}$ CARBONATE ROCKS (LATE TRIASSIC) -- Massive to thick-bedded limestone and minor dolomite in most areas, but in Keku Strait it also includes thin-bedded dark gray limestone, and on Chichagof Island it consists of massive white to gray marble (Whitestripe Marble). Diagnostic fossils recovered from the limestone and dolomite are Late Triassic in age; the marble on Chichagof Island is interpreted to be Late Triassic by correlation with the Chitistone Limestone in the Wrangell Mountains of southern Alaska (Plafker and others, 1976: Jones and others, 1977). These workers also suggest that the Whitestripe Marble is not correlative with the other carbonate rocks in this unit. This unit consists of unnamed, generally recrystallized limestone on Annette and Gravina Islands (Berg, 1982), the Cornwallis Limestone and Hamilton Island Limestone on northern Kuiu and Kupreanof Islands (Muffler, 1967; Brew and others, 1984), and the Whitestripe Marble on Chichagof Island (Loney and others, 1975; Johnson and Karl, 1982).

$\boxed{\text{TRb}}$ BASALTIC ROCKS (LATE AND (OR) MIDDLE TRIASSIC) -- Basaltic flows and flow breccia (Goon Dip Greenstone) on Chichagof and Yakobi Islands that were deposited, at least in part, in subaerial environments (Loney and others, 1975; Johnson and Karl, 1982). Plafker and others (1976) and Jones and others (1977) correlate the Goon Dip Greenstone and overlying Whitestripe Marble with the Late and (or) Middle Triassic

Nikolai Greenstone and overlying Late Triassic Chitistone Limestone in the Wrangell Mountains of southern Alaska.

| Ps | SEDIMENTARY ROCKS (EARLY PERMIAN AND PERMIAN) -- Black argillite, graywacke, calcareous siltstone, chert, and minor basaltic rocks, limestone, and conglomerate. This unit consists of part of the Halleck Formation of Early Permian age near Keku Strait (Muffler, 1967; Brew and others, 1984; Jones and others, 1981), the Cannery Formation on southern Admiralty Island, where it is reported to be Early Permian in age (Lathram and others, 1965), and unnamed Permian strata northeast of Glacier Bay (Brew and others, 1978). Strata are locally metamorphosed to slate and phyllite.

| Pv | VOLCANIC ROCKS (EARLY PERMIAN AND PERMIAN) -- Early Permian basalt (part of the Halleck Formation) on northern Kuiu Island (Muffler, 1967) and unnamed Permian basaltic(?) rocks northeast of Glacier Bay (Brew and others, 1978). Age is constrained by interbedded Early Permian clastic strata (Ps) on Kuiu Island and interbedded Permian carbonate rocks (Pc) in the Glacier Bay area.

| Pc | CARBONATE ROCKS (EARLY PERMIAN AND PERMIAN) -- Greenschist facies, massive to thin-bedded marble of Early Permian age on Revillagigedo Island (Silberling and others, 1982; Berg and others, 1978) and of Permian(?) age on the mainland east of Admiralty Island (Brew and Grybeck, 1984); medium-bedded dolomite, limestone, and subordinate gray chert beds and nodules (Pybus Formation) on northern Kuiu and Kupreanof Islands (Early Permian) (Muffler, 1967; Brew and others, 1984), Admiralty Island (Permian) (Lathram and others, 1965), and a small island between northern Prince of Wales Island and Etolin Island (Early Permian) (Brew and others, 1984); and unnamed

fossiliferous gray limestone, cherty limestone, and limestone conglomerate of Permian age northeast of Glacier Bay (Brew and others, 1978) and along the west shore of Lynn Canal (Lathram and others, 1959). The marble in this unit on Revillagigedo Island and on the mainland east of Admiralty Island may or may not be correlative with Permian carbonate rocks elsewhere in the map area (Silberling and others, 1982).

| ℙs | SEDIMENTARY ROCKS (PENNSYLVANIAN) -- Early and middle Pennsylvanian sandstone and siltstone with minor limestone and chert-pebble conglomerate (Klawak Formation) on west-central Prince of Wales Island (Eberlein and others, 1983).

| ℙc | CARBONATE ROCKS (PENNSYLVANIAN) -- Massive limestone and minor dolomite with light-gray chert nodules (Ladrones Limestone) on west-central Prince of Wales Island (Eberlein and others, 1983), and unnamed, medium-bedded to massive crinoidal limestone on northern Kuiu Island (Brew and others, 1984). Fossils recovered indicate deposition during early and middle Pennsylvanian time.

| Mc | CARBONATE ROCKS (MISSISSIPPIAN) -- Thin- to thick-bedded dark gray limestone, subordinate beds and nodules of light gray chert, and minor shale interbeds and gypsum. This unit consists of the Peratrovich Formation on western Prince of Wales Island (Eberlein and others, 1983), and the locally gypsum-bearing Iyoukeen Formation on Chichagof Island (Loney and others, 1975). Strata were deposited during early and late Mississippian time.

| MDsv | SEDIMENTARY AND VOLCANIC ROCKS (MISSISSIPPIAN AND LATE DEVONIAN) -- Tuffaceous argillite and graywacke, and subordinate chert, limestone, and andesitic volcanic rocks (Cannery Formation) on northern

Kupreanof Island (Muffler, 1967; Brew and others, 1984; Jones and others, 1981).

| Ds | SEDIMENTARY ROCKS (EARLY DEVONIAN, DEVONIAN, AND DEVONIAN?) -- Siltstone, shale, sandstone, graywacke, and subordinate limestone breccia, arkose, conglomerate, and volcanic rocks. This unit consists of: unnamed Early Devonian strata in the Annette-Gravina Islands area (Berg and others, 1978; Gehrels and others, 1984), on southern Prince of Wales Island (Herreid and others, 1978; Gehrels and others, 1983; Savage and Gehrels, 1984), on west-central Prince of Wales Island (Eberlein and others, 1983), and on smaller islands west of Prince of Wales Island (Eberlein and others, 1983); unnamed Devonian(?) graywacke, argillite, and arkose on northern Kuiu Island (Brew and others, 1984); and the undivided part of the Devonian Cedar Cove Formation on northeastern Chichagof Island (Loney and others, 1975).

| Dv | VOLCANIC ROCKS (LATE DEVONIAN, EARLY DEVONIAN, AND DEVONIAN) -- Basaltic and subordinate andesitic pillow flows, breccia, aquagene tuff, and minor sedimentary interbeds. This unit consists of: the volcanic part of the Port Refugio Formation (Late Devonian) on west-central Prince of Wales Island (Eberlein and others, 1983); Early Devonian rhyolite in Kasaan Bay on east-central Prince of Wales Island (Eberlein and others, 1983); the Late Devonian Freshwater Bay Formation on Chichagof Island (Loney and others, 1975); unnamed volcanic rocks in Glacier Bay (Brew and others, 1978); and unnamed Silurian(?) to Devonian rocks in the Chilkat Range (Lathram and others, 1959). We interpret the volcanic rocks in Glacier Bay and in the Chilkat Range to be correlative with the Freshwater Bay Formation on Chichagof Island based on their similarity in lithology and apparent stratigraphic position.

Dsv SEDIMENTARY AND BASALTIC ROCKS (LATE DEVONIAN, DEVONIAN, AND DEVONIAN?) -- Siltstone, shale, volcanogenic graywacke, conglomerate, and minor limestone that are interbedded with basaltic pillow flows, breccia, and aquagene tuff. This unit consists of the Coronados Island Volcanics (Devonian), the St. Joseph Island Volcanics (Devonian?), and part of the Late Devonian Port Refugio Formation, all of which occur on western Prince of Wales Island and adjacent islands (Eberlein and others, 1983).

Dcg CONGLOMERATIC ROCKS (EARLY DEVONIAN AND DEVONIAN) -- Conglomerate and sedimentary breccia interbedded with sandstone, siltstone, shale, and minor graywacke, limestone, and volcanic rocks. Sedimentary structures and the occurrence of red-beds and thick sections of coarse conglomerate suggest that the strata in the Prince of Wales Island region were deposited in subaerial to shallow-marine environments (Ovenshine, 1975; Gehrels and others, 1983). The argillite-rich section on Chichagof Island was probably deposited in a shallow marine environment (Loney and others, 1975). This unit consists of: unnamed Early Devonian strata on southern Prince of Wales Island (Gehrels and others, 1983; Savage and Gehrels, 1984); the Early Devonian Karheen Formation on central Prince of Wales Island (Eberlein and others, 1983); unnamed Devonian strata on east-central Prince of Wales Island (Eberlein and others, 1983); and the lower part of the Cedar Cove Formation on Chichagof Island (Loney and others, 1975).

Dc CARBONATE ROCKS (DEVONIAN AND DEVONIAN?) -- Thin-bedded to massive gray limestone with minor shale interbeds. This unit consists of: the late Early to Late Devonian Wadleigh Limestone on west-central Prince of Wales Island and unnamed Early Devonian limestone on east-central Prince of Wales Island (Eberlein and others, 1983; Savage and Gehrels, 1984); unnamed Early and Middle Devonian strata interpreted to occur as blocks in a disrupted sedimentary unit (KJsd) on Kupreanof Island (D.A. Brew, in U.S. Geological Survey, 1982; Brew and others, 1984); the upper part of the Cedar Cove Formation (Middle and Late Devonian) on Chichagof Island (Loney and others, 1975); part of the Black Cap Limestone (Middle Devonian) in Glacier Bay (Rossman, 1963; Brew and others, 1978); unnamed Devonian strata on the west shore of Lynn Canal (Lathram and others, 1959; Loney and others, 1975); and unnamed strata in the Chilkat range (Lathram and others, 1959; Brew and others, 1978) that we interpret to be Devonian in age based on similarities in lithology and stratigraphic position with rocks on northeastern Chichagof Island.

Sc CARBONATE ROCKS (SILURIAN) -- Massive, thin- to thick-bedded, and locally reefoidal light-gray limestone, and subordinate shale interbeds and layers and lenses of polymictic conglomerate. This unit consists of: unnamed Silurian limestone and marble on Dall and Long Islands that are locally metamorphosed to greenschist facies (Eberlein and others, 1983; G.E. Gehrels, unpub. mapping, 1984); part of the Late and Early Silurian Heceta Limestone on Prince of Wales Island (Eberlein and others, 1983; Brew and others, 1984); the Late Silurian Kuiu Limestone and limestone layers in the Late(?) and Early Silurian Bay of Pillars Formation on Kuiu Island (Muffler, 1967; Brew and others, 1984); the Kennel Creek Limestone (Devonian and (or) Silurian: probably Silurian) and limestone layers in the Late(?) Silurian Point Augusta Formation on Chichagof Island (Loney and others, 1975); the Willoughby Limestone, Pyramid Peak Limestone, and limestone layers in the Tidal

Formation (all considered to be Silurian by Rossman, 1963) in the Glacier Bay area (Rossman, 1963; Brew and others, 1978); and unnamed Silurian(?) to Devonian limestone in the Chilkat Range (Lathram and others, 1959). Loney and others (1975) correlate limestones in the Chilkat Range along the west shore of Lynn Canal with the Willoughby Limestone and the Pyramid Peak Limestone in Glacier Bay, and the Kennel Creek Limestone on northeastern Chichagof Island. Based on these correlations and the apparent similarity in stratigraphic position, we correlate the limestones in both the "limestone and marble" unit and the "siliceous argillite and volcanic" unit of Lathram and others (1959) with the Kennel Creek Limestone and limestone layers in the Point Augusta Formation on Chichagof Island. The stratigraphic relations and fossil content of rocks in this map unit indicate deposition primarily during Silurian time. Strata in some areas, however, may be in part of Early Devonian age.

| Scg | CONGLOMERATIC ROCKS (SILURIAN) -- Polymictic pebble and cobble conglomerate and subordinate sedimentary breccia, olistostromal deposits, sandstone, graywacke, mudstone, and limestone. Clasts consist of porphyritic andesite, limestone, graywacke, mudstone, granitic to gabbroic intrusive rocks, chert, and other rocks derived from various Silurian, and Silurian and Ordovician units. This unit consists of unnamed polymictic conglomerate on northern Prince of Wales Island (Brew and others, 1984) and conglomeratic layers and lenses in: the Heceta Limestone and the Bay of Pillars Formation on Prince of Wales Island (Eberlein and others, 1983; Brew and others, 1984), the Kuiu Limestone and the Bay of Pillars Formation on Kuiu Island (Brew and others, 1984), and the Kennel Creek Limestone on Chichagof Island (Loney and others,

1975). Interbedded Silurian carbonate rocks (Sc) and clastic sedimentary rocks (Ss) indicate deposition during Silurian time.

Ss SEDIMENTARY ROCKS (SILURIAN) -- Graywacke and mudstone turbidites, and subordinate olistostromal deposits and layers and lenses of limestone and conglomerate. This unit consists of most of the Bay of Pillars Formation (Late and Early Silurian) on northern Prince of Wales Island and Kuiu Island (Brew and others, 1984), the Late(?) Silurian Point Augusta Formation on Chichagof Island (Loney and others, 1975), the Late Silurian Tidal and Rendu Formations in Glacier Bay (Rossman, 1963; Brew and others, 1978), and the Silurian(?) to Devonian "graywacke, argillite, and limestone" unit of Lathram and others (1959) in the Chilkat Range (interpreted to be Late Silurian by Loney and others, 1975). Although stratigraphic relations and fossils indicate that most of the strata in this unit are Silurian in age, deposition of strata in some areas may have continued into Early Devonian time.

Sv VOLCANIC ROCKS (SILURIAN AND SILURIAN?) -- Mafic to intermediate-composition volcanic breccia, agglomerate, and flows, and greenschist and greenstone derived from these rocks. Strata are known to be Silurian in age where they occur in the Bay of Pillars Formation (Ss and Sc) on southern Kuiu Island. On northern Kuiu Island rocks in this unit occur in a fault slice and are interpreted to be Silurian in age (Brew and others, 1984).

SOs SEDIMENTARY ROCKS (EARLY SILURIAN TO EARLY ORDOVICIAN) -- Mudstone and graywacke turbidites, subordinate conglomerate, sandstone, and shale, and minor limestone, chert, and basalt flows and breccia. This unit consists of: unnamed strata of Early Silurian to Middle Ordovician age on Sukkwan and Dall Islands (Eberlein and others, 1983;

G.E. Gehrels, unpub. mapping, 1984); unnamed strata of Silurian and (or) Ordovician age on Forrester Island (Clark and others, 1971); unnamed strata of Early Silurian to Early Ordovician age on southern Prince of Wales Island (Eberlein and others, 1983; Gehrels and others, 1983; G.E. Gehrels, unpub. mapping); part of the Early Silurian to Early Ordovician Descon Formation on northern Prince of Wales Island (Eberlein and others, 1983; Brew and others, 1984); and thin-bedded black argillite, black chert, and black impure limestone of Ordovician age (Hood Bay Formation) on southern Admiralty Island (Lathram and others, 1965; Carter, 1977).

SOv VOLCANIC ROCKS (EARLY SILURIAN TO EARLY ORDOVICIAN) -- Basaltic pillow flows, pillow breccia, and aquagene tuff; massive andesitic pyroclastic breccia; felsic breccia and tuff; subordinate interbeds of mudstone and graywacke turbidites; and the metamorphic equivalents of these strata. This unit consists of: unnamed basaltic to felsic rocks on southern Prince of Wales Island (Gehrels and others, 1983; G.E. Gehrels, unpub. mapping); unnamed felsic and basaltic rocks and their greenschist- and amphibolite-facies equivalents on Long and southern Dall Islands (G.E. Gehrels, unpub. mapping, 1984); and basaltic rocks (part of the Descon Formation) and unnamed andesitic breccia on central Prince of Wales Island (Eberlein and others, 1983). Age is constrained by stratigraphic relations with Early Silurian to Early Ordovician strata, intrusive relations with plutons of Early Silurian and Late Ordovician age, and Late Ordovician and Early Silurian K-Ar apparent ages of the volcanic rocks (Eberlein and others, 1983; Gehrels and others, 1983; G.E. Gehrels and J.B. Saleeby, unpub. data). The higher grade metamorphic rocks on southern Dall Island are primarily schist and gneiss, and may be in part correlative with the pre-Ordovician Wales Group (pOmsv).

SOsv SEDIMENTARY AND VOLCANIC ROCKS (EARLY SILURIAN TO EARLY ORDOVICIAN) -- Undivided mudstone and graywacke turbidites, basaltic and andesitic volcanic rocks, and the metamorphic equivalents of these strata. This unit consists of: part of the Descon Formation on Prince of Wales Island (Eberlein and others, 1983); unnamed rocks on Annette and Duke Islands and adjacent parts of the mainland that are metamorphosed in most areas to greenschist facies (Berg and others, 1978; Gehrels and others, 1984; G.E. Gehrels and J.B. Saleeby, unpub. data); conglomerate, agglomerate, and volcanic breccia belonging to the Bay of Pillars Formation (probable Early Silurian age) on northern Prince of Wales Island (Brew and others, 1984); and chlorite schist, sericite schist, phyllite, and black slate in the Chilkat Range (Lathram and others, 1959). The rocks in the Chilkat Range are interpreted herein to occupy a similar stratigraphic position to the Early Silurian to Early Ordovician Descon Formation of Prince of Wales Island.

SOsvc SEDIMENTARY, VOLCANIC, AND CARBONATE ROCKS (EARLY SILURIAN AND ORDOVICIAN) -- Interbedded graywacke, mudstone, basaltic to felsic volcanic rocks, and limestone, that have been regionally metamorphosed as high as greenschist facies (G.E. Gehrels, unpub. data, 1984). This unit occurs on the west coast of Dall Island and on Long Island.

Undivided map units

KPs SEDIMENTARY ROCKS (CRETACEOUS? AND JURASSIC?, TRIASSIC, TRIASSIC?, PERMIAN, AND PERMIAN?) -- Carbonaceous shale, mudstone, graywacke, and subordinate limestone, chert, conglomerate, and andesitic or basaltic and minor felsic volcanic rocks, that have been metamorphosed and deformed in much of the area. Regional metamorphic grade in these strata and in associated map units (KPv, KPsv, and KPc) generally increases from sub-greenschist or greenschist facies on the southwest, to amphibolite facies toward the northeast. There are also significant changes in metamorphic grade along the northwesterly trend of these map units. Common rock types in the higher grade strata include phyllite, schist, and gneiss. Late Triassic fossils have been recovered from carbonaceous slate and limestone assigned to this map unit near Juneau (Ford and Brew, 1977; H.C. Berg, unpub. data, 1981), and similar strata (locally subdivided into unit "R̄s") on Revillagigedo Island have yielded latest Middle Triassic fauna (Silberling and others, 1982). On Revillagigedo Island phyllite and metagraywacke are locally in depositional contact with crinoidal marble of known Permian age (unit Pc), and are interbedded with crinoidal marble of probable Permian age (Berg and others, 1978; H.C. Berg, unpub. data, 1975). Some strata included in this unit are probably Cretaceous or Jurassic in age: Berg and others (1978) report that metasedimentary rocks in this unit on Revillagigedo Island are locally identical in protolith to Cretaceous and Jurassic strata (KJs) on Annette and Gravina Islands; Brew and others (1984) suggest that some rocks in this unit on the mainland east of Kupreanof Island may have Cretaceous or Jurassic protolith ages; and Buddington and Chapin (1929, pp. 74) report that conglomerate and phyllite in this unit on the mainland east of Admiralty Island are lithologically similar to conglomerate and slate of Cretaceous and Jurassic age (KJs) on eastern Admiralty Island. The contact between rocks of probable Cretaceous and Jurassic age (units KJs, KJv, and KJsv) and rocks in this and associated map units (KPv, KPsv, KPc) is drawn along the northeastern contact of relatively homogeneous meta-graywacke and phyllite. Intrusive bodies of Late Cretaceous age (Kgt) place a younger age limit on the depositional age of these strata. The age of metamorphism and deformation has traditionally been interpreted to be Late Cretaceous and (or) early Tertiary (Buddington and Chapin, 1929). However, recognition of: 1) an unconformity separating metamorphosed strata belonging to associated map unit "KPsv" from less-metamorphosed Cretaceous and Jurassic strata northwest of Juneau (Redman, 1984a), and 2) a greater degree of metamorphism and deformation in this and associated units than in strata of probable Cretaceous and Jurassic age on Revillagigedo Island (G.E. Gehrels and J.B. Saleeby, unpub. mapping, 1984) suggest that Triassic and Permian rocks in this unit may have been initially metamorphosed and deformed prior to the deposition of the Cretaceous and Jurassic strata. This unit consists of unnamed rocks on Revillagigedo Island (Berg and others, 1978), and on the mainland east of Kupreanof Island (Brew and others, 1984) and east of Admiralty Island (Brew and Grybeck, 1984; Buddington and Chapin, 1929; Ford and Brew, 1973 and 1977; Brew and Ford, 1977).

$\boxed{\text{KPv}}$ VOLCANIC ROCKS (CRETACEOUS? AND JURASSIC?, TRIASSIC, TRIASSIC?, PERMIAN, AND PERMIAN?) -- Andesitic or basaltic flows and locally fragmental rocks, subordinate clastic strata and carbonate, and the metamorphosed and deformed equivalents of these rocks. Regional metamorphism of these strata is similar to that described above for unit "KPs." Common rock types in higher grade regions include schist, gneiss, and amphibolite, although some of this amphibolite may be meta-gabbro (see unit "KPsv" below). Graphitic slate assigned to this unit contains Late Triassic fossils near Juneau (Brew and Grybeck, 1984, pp.

31), and marble intercalated with metavolcanic rocks contains Permian and Permian(?) fossils near Juneau, along the coast southeast of Juneau, and in Endicott Arm (Buddington and Chapin, 1929, pp. 73 and 119; Ford and Brew, 1973; Brew and Grybeck, 1984, pp. 30-31). The minimum depositional age is constrained by interbedded rocks belonging to unit "KPs" (described above), and by crosscutting Late Cretaceous intrusive bodies (Kgt). This unit consists of unnamed rocks on Revillagigedo Island (Berg and others, 1978), and on the mainland east of Kupreanof Island (Brew and others, 1984) and east of Admiralty Island (Brew and Grybeck, 1984; Buddington and Chapin, 1929; Ford and Brew, 1973 and 1977; Brew and Ford, 1977).

KPsv SEDIMENTARY AND VOLCANIC ROCKS (CRETACEOUS? AND JURASSIC?, TRIASSIC, TRIASSIC?, PERMIAN, and PERMIAN?) -- Undivided shale, mudstone, graywacke, andesitic or basaltic rocks, subordinate chert and carbonate rocks, and the regionally metamorphosed equivalents of these strata. The nature and age of metamorphism is similar to that described above for unit "KPs." Protolith age is indicated by stratigraphic relations with metasedimentary and metavolcanic rocks of Cretaceous(?) to Permian age (KPs and KPv) described above. Metasedimentary and metavolcanic rocks in the stippled area in this unit southeast of Revillagigedo Island are intruded by a deformed and metamorphosed gabbro (recrystallized to amphibolite) that is apparently related to the foliated tonalitic rocks in unit "TKt" (G.E. Gehrels and J.B. Saleeby, unpub. mapping, 1984). The Work Channel amphibolite to the southeast in British Columbia (Hutchison, 1982), and some of the amphibolite in Cretaceous(?) to Permian map units to the northwest, may consist in part of correlative meta-gabbro. Unit occurs on Revillagigedo Island and on adjacent areas of the mainland to the southeast (Berg and others, 1978) and the northwest (Elliott and Koch, 1981), on the mainland east of Admiralty Island (Buddington and Chapin, 1929; Brew and Grybeck, 1984; Brew and Ford 1977; Ford and Brew, 1973 and 1977), and on the east side of Lynn Canal (Buddington and Chapin, 1929; Redman, 1984a).

$\boxed{\text{KPc}}$ CARBONATE ROCKS (CRETACEOUS? AND JURASSIC?, TRIASSIC?, AND PERMIAN?) -- Unfossiliferous, metamorphosed and deformed carbonate rocks that are intercalated with andesitic or basaltic metavolcanic rocks of Cretaceous(?) and Jurassic(?) age on the mainland northwest of Revillagigedo Island (H.C. Berg, unpub. data, 1975), and are associated with metasedimentary and metavolcanic rocks of Cretaceous(?) to Permian age east of Etolin and Zarembo Islands (Brew and others, 1984).

$\boxed{\text{KPsvc}}$ SEDIMENTARY, VOLCANIC, AND CARBONATE ROCKS (CRETACEOUS TO PERMIAN?, AND PRE-PERMIAN?) -- This unit consists of: unnamed fossiliferous argillite of Cretaceous age; variably metamorphosed chert, tuffaceous sandstone, felsic tuff, argillite, and light-gray thinly bedded limestone of Mesozoic or Paleozoic age; highly sheared metasedimentary, metavolcanic, and granodioritic rocks of Mesozoic and Paleozoic(?) age; and amphibolite, gneiss, schist, and marble of Mesozoic or Paleozoic age (Johnson and Karl, 1982; Loney and others, 1975). This unit occurs on Chichagof, Baranof, and Yakobi Islands. Areas that may be underlain by a significant proportion of intrusive rock are shown with a stipple pattern.

$\boxed{\text{JMsv}}$ SEDIMENTARY AND VOLCANIC ROCKS (JURASSIC TO MISSISSIPPIAN?) -- Diverse assemblage of sedimentary and volcanic rocks and their metamorphic equivalents. This unit occurs in the Coast Mountains near the British Columbia-Alaska border northeast of Revillagigedo Island

(Berg and others, 1978; Elliott and Koch, 1981), and east of Juneau (Souther and others, 1979; Souther, 1971). The unit consists of: unnamed Middle Jurassic rhyolite and andesite; Early Jurassic andesite, basalt, conglomerate, and sandstone of the Hazelton Group; Late Triassic andesite and clastic sedimentary rocks; Permian limestone; and Carboniferous greenstone, limestone, and clastic sedimentary rocks (Hutchison and others, 1979; Berg and others, 1978; Souther and others, 1979; Souther, 1971). Rocks in this unit are locally metamorphosed to semischist, phyllite, and schist. Map unit "pTmsv" probably consists in large part of the higher grade equivalents of these strata.

$\boxed{\overline{\mathrm{R}}\mathrm{Psv}}$ SEDIMENTARY AND VOLCANIC ROCKS (TRIASSIC AND PERMIAN) -- Andesitic to basaltic pillow flows, breccia, and tuff, felsic tuff, graywacke, mudstone, shale, chert, conglomerate, carbonate, and the metamorphic equivalents of these strata. This unit occurs only on Admiralty Island and consists of the Barlow Cove Formation (Barker, 1957) and of strata originally mapped by Lathram and others (1965) as parts of the Hyd Formation (Late Triassic) and the Cannery Formation (Early Permian). Phyllite, greenstone, greenschist, and marble are common rock types in this unit, but the nature and age of the metamorphism are poorly known.

$\boxed{\overline{\mathrm{R}}\mathrm{Osv}}$ SEDIMENTARY AND VOLCANIC ROCKS (TRIASSIC TO ORDOVICIAN) -- Clastic sedimentary rocks, subordinate mafic to felsic volcanic rocks, thin- to thick-bedded gray carbonate, chert, and minor ultramafic rocks, that have been regionally metamorposed to slate, phyllite, greenschist, schist, gneiss, and marble in many areas. The age and grades of this metamorphism have not been reliably determined. Rocks assigned to this unit on Admiralty Island belong to the Gambier Bay Formation, the

Retreat Group, and the "undifferentiated metamorphic rocks" and the "migmatite, gneiss, and feldspathic schist" units of Lathram and others (1965). Devonian fossils have been recovered from marble in the Gambier Bay Formation; the Retreat Group has been inferred to be Devonian in age based on correlation with the Gambier Bay Formation; and the undivided metamorphic rocks are undated (Lathram and others, 1965). The Triassic to Ordovician age assignment on Admiralty Island reflects our interpretation that this map unit consists primarily of regionally metamorphosed and deformed strata correlative with the Late Triassic Hyd Formation (R̄sv), the Permian Pybus Dolomite (Pc) and Cannery Formation (Ps), the Ordovician Hood Bay Formation (SOs), and various strata elsewhere assigned to Devonian, Silurian, and Silurian and Ordovician units. Rocks in part of this unit may also belong to the complex of disrupted strata (KJsd) in the Kupreanof Island region (D.A. Brew, *in* U.S. Geological Survey, 1982). Areas on Admiralty Island that may be underlain by a significant proportion of younger intrusive rock are shown with a stipple pattern.

In the Chilkat Range west of Haines this unit consists of unnamed metasedimentary and metavolcanic rocks (MacKevett and others, 1974; Robertson, 1959; Redman, 1984b) that are of mid to late Paleozoic and probable Triassic age (Berg and Grybeck, 1980; Redman, 1984b). These rocks herein are interpreted to be correlative with sedimentary and volcanic rocks of Permian, Devonian, and Silurian age in Glacier Bay (Brew and others, 1978) and in the Chilkat Range (Lathram and others, 1959), and Triassic to Ordovician strata that adjoin rocks in this unit along the British Columbia-Alaska border north of Glacier Bay (Campbell and Dodds, 1983).

ŦOc CARBONATE ROCKS (TRIASSIC? TO ORDOVICIAN?) -- Carbonate rocks that are regionally metamorphosed to gray marble (MacKevett and others, 1974; Lathram and others, 1959). Fossils of Pennsylvanian(?) or Permian(?), Paleozoic(?), and possibly of Silurian or Devonian age have been recovered from this unit west of Haines (MacKevett and others, 1974). Adjacent units in British Columbia consist of carbonate rocks of Ordovician, Silurian, Devonian, and Triassic age (Campbell and Dodds, 1983). The age of the strata in this unit on Admiralty Island is constrained by stratigraphic relations with Triassic to Ordovician clastic strata (ŦOsv). Regional map patterns suggest that most of the carbonate rocks in this map unit are of Silurian age, and are correlative with the Kennel Creek Limestone, Pyramid Peak Limestone, and Willoughby Limestone (Sc) in the Chichagof Island-Glacier Bay area.

ℙDs SEDIMENTARY ROCKS (PENNSYLVANIAN AND DEVONIAN) -- Saginaw Bay Formation on Kuiu Island, which consists, from youngest to oldest, of: silty limestone, calcareous chert and limestone, black chert, and massive aquagene tuff and pillow breccia (Muffler, 1967). The silty limestone and the calcareous chert and limestone are known to be Pennsylvanian; the age of the black chert is not known directly; and the volcanic rocks have yielded earliest Late to latest Early Devonian conodonts (Brew and others, 1984).

DSsc CARBONATE AND CLASTIC SEDIMENTARY ROCKS (DEVONIAN AND SILURIAN) -- On north-central Prince of Wales Island this unit consists of unnamed limestone, sandstone, calcareous mudstone, and polymictic conglomerate that are interpreted to be facies-equivalents of the Silurian Heceta Limestone and associated clastic strata (Sc, Ss, and Scg), and the Early Devonian Karheen Formation (Dcg) (Eberlein and others, 1983). In the Chilkat Range, this unit consists of unnamed siliceous argillite, conglomerate, graywacke, and subordinate thin limestone beds and basalt and andesite flows, agglomerate, and tuff (Lathram and others, 1959). In Glacier Bay, the unit consists of unnamed clastic sedimentary and minor carbonate and volcanic rocks (Brew and others, 1978). The strata in this unit in the Chilkat Range and in Glacier Bay are probably correlative with the Devonian Cedar Cove Formation (units Ds, Dcg, and Dc) and the Late? Silurian Point Augusta Formation (Ss and Sc) on northeastern Chichagof Island, and the Tidal and Rendu Formations (Ss and Sc) in Glacier Bay.

$\boxed{\text{DOsv}}$ SEDIMENTARY AND VOLCANIC ROCKS (DEVONIAN TO ORDOVICIAN) -- Graywacke, mudstone, shale, limestone, and subordinate mafic to intermediate-composition volcanic rocks, that have been metamorphosed in most areas to hornfels, schist, amphibolite, marble, gneiss, and granofels. This unit consists of unnamed rocks on northeastern Chichagof Island and in the Glacier Bay area. Loney and others (1975) suggest that the metamorphic rocks on Chichagof Island were derived mainly from the Point Augusta Formation (Ss) and from Devonian volcanic rocks of the Freshwater Bay Formation (Dv). In the Glacier Bay area the rocks are probably the metamorphic equivalents of Devonian and Silurian strata that occur nearby (units Ds, Dv, Dcg, Ss, and Sc) (Brew and others, 1978). Ordovician strata occur in contiguous map units northwest of Glacier Bay in British Columbia (Campbell and Dodds, 1983): we therefore suggest that this unit probably includes rocks of Ordovician age as well. On northeastern Chichagof Island the strata were metamorphosed to hornblende hornfels facies during emplacement of the adjacent Cretaceous intrusive rocks (Loney and others, 1975).

Metamorphism in the Glacier Bay area is also interpreted to have occurred during Cretaceous time (Brew and others, 1978).

| DOc | CARBONATE ROCKS (DEVONIAN TO ORDOVICIAN) -- Carbonate rocks that have been metamorphosed to thin- to thick-bedded, dark gray to white marble. This unit consists of unnamed rocks on northeastern Chichagof Island and in the Glacier Bay area. Loney and others (1975) suggest that the marble on Chichagof Island was derived primarily from limestone in the Point Augusta Formation (Sc). The marble in the Glacier Bay area was derived in large part from limestone of Devonian (Dc) and Silurian (Sc) age (Brew and others, 1978). Some marble in this unit may be correlative with Ordovician limestone in map units along the British Columbia-Alaska Border northwest of Glacier Bay (Campbell and Dodds, 1983). Metamorphic relations are the same as in unit "DOsv," described above.

Lithodemic units

| Ksvm | MELANGE (SEDIMENTARY AND MAFIC VOLCANIC ROCKS) (CRETACEOUS) -- Rocks in this map unit and in associated units "Ksm" and "Kvm" comprise a melange of regionally disrupted, deformed, and metamorphosed marine sedimentary and volcanic rocks. This unit consists of: 1) blocks up to several kilometers in length of marine sedimentary and volcanic rocks and their metamorphic equivalents, that are bounded by faults or shear zones, and 2) blocks up to several tens of meters in length, and perhaps larger, of marine sedimentary and volcanic rocks and their metamorphic equivalents, that are enclosed in a matrix of moderately metamorphosed and penetratively sheared graywacke, argillite, tuff, and pillow breccia. Common lithic types in the blocks include tuffaceous argillite, tuff, greenstone, graywacke, chert, limestone, marble,

phyllite, schist, and minor granitic rocks. Fossils recovered from blocks in this unit are Early Cretaceous, Late Jurassic, and possibly Jurassic or Triassic in age (Loney and others, 1975; Johnson and Karl, 1982). The minimum age of the unit is constrained by: 1) mid-Cretaceous K-Ar apparent ages on metamorphic minerals in the melange on Chichagof Island (Decker and others, 1980), 2) Early Cretaceous fossils in sedimentary matrix material near Sitka on Baranof Island (Plafker and others, 1976), and 3) Early Cretaceous fossils in matrix material near the coast northwest of Cross Sound (Plafker and others, 1977; George Plafker, written commun., 1984). The Cretaceous age assigned to this unit therefore refers to the time during which the melange is interpreted to have formed. Metamorphic grade is primarily sub-greenschist, greenschist, and, south of Cross Sound, glaucophane-schist facies (Decker and others, 1980). On Chichagof Island this unit consists of the Khaz Formation, which is a melange of blocks enclosed in a sedimentary and volcanic matrix, and the Freeburn assemblage, which consists of fault-bounded blocks of sedimentary and volcanic rocks (Decker, 1980; Johnson and Karl, 1982; Karl and others, 1982). The Freeburn assemblage comprises rocks described by Loney and others (1975) as the Pinnacle Peak Phyllite and an unnamed schist unit in the Kelp Bay Group. On Baranof Island this unit consists of rocks described by Loney and others (1975) as the Khaz Formation, part of the undivided Kelp Bay Group, and an unnamed "schist, gneiss, amphibolite, and greenschist" unit (Karl and others, 1982). Northwest of Cross Sound the unit consists of: melange between the Brady Glacier and Border Ranges faults (Tarr Inlet suture zone) that is continuous and probably correlative with part of the melange on Chichagof and Baranof Islands (Decker and

Plafker, 1982); and the melange facies of the Yakutat Group west of the Fairweather fault (Plafker and others, 1977; George Plafker, written commun., 1984).

Ksm MELANGE (SEDIMENTARY ROCKS) (CRETACEOUS) -- Melange of predominantly metasedimentary rocks that resembles unit "Ksvm" (described above) in structural and metamorphic characteristics. Common lithic types include dark-gray phyllite, light-gray quartzite, and graywacke semischist. The age of formation of this melange unit is constrained by close stratigraphic and structural association with rocks in unit "Ksvm." Fossils have not been recovered from these metasedimentary rocks. This unit consists of unnamed rocks mapped by Loney and others (1975) as the "phyllite," and the "graywacke semischist" units on Baranof Island.

Kvm MELANGE (MAFIC VOLCANIC ROCKS) (CRETACEOUS) -- Melange of predominantly mafic metavolcanic rocks that resembles unit "Ksvm" (described above) in structural and metamorphic characteristics. Common lithic types include greenstone, greenschist, and minor phyllite, graywacke semischist, marble, and metachert. The age of formation of this melange unit is constrained by close stratigraphic and structural association with rocks in units "Ksvm" amd "Ksm." This unit consists of the Waterfall Greenstone on western Chichagof Island, which has yielded radiolaria of Early Cretaceous age (Johnson and Karl, 1982), and unnamed metavolcanic rocks on Baranof Island that are referred to as the "greenschist and greenstone" unit by Loney and others (1975).

pTmsv METAMORPHIC COMPLEX (DERIVED FROM SEDIMENTARY AND VOLCANIC ROCKS) (PRE-TERTIARY) -- Amphibolite- and locally granulite-facies metasedimentary, metavolcanic, and subordinate metaplutonic rocks, that belong to a metamorphic complex. Common rock types include pelitic, semipelitic, and quartzo-feldspathic schist and gneiss, and subordinate amphibolite, quartzite, marble, and calc-silicate. Protoliths are interpreted to have been argillaceous marine strata, limestone, chert, subordinate mafic to felsic volcanic rocks, and minor intrusive rocks. Areas in which there may be a significant proportion of unmapped Cretaceous or Tertiary intrusive rocks are shown with a stipple pattern. Such areas comprise "migmatite" map units in some reports listed below. The minimum age of the rocks in this unit is constrained by cross-cutting tonalitic bodies of Paleocene and Late Cretaceous age (TKt) in the Coast Mountains (Gehrels and others, 1985). Protolith ages of Cretaceous(?), Jurassic(?), Triassic, Permian(?), Carboniferous(?), and Proterozoic(?) are interpreted from: 1) relations along the British Columbia-Alaska border east of Juneau, which suggest that rocks in this unit grade into strata of Triassic and older age, and that these Triassic rocks locally contain clasts of older metamorphic rocks (Souther, 1971), 2) a preliminary Rb/Sr apparent isochron of Proterozoic age (R.L. Armstrong and L.J. Werner, oral commun., 1984) determined on high-grade metamorphic rocks along the British Columbia-Alaska border north of Juneau (Souther and others, 1979; Werner, 1978), 3) relations in the Coast Mountains between Portland Canal and Terrace (southeast of the map area) which suggest that analogous metamorphic rocks were derived from Cretaceous(?) and Jurassic(?) strata (Douglas, 1983), the Jurassic Bowser Lake Group (Woodsworth and others, 1983, pp. 13), and perhaps pre-Permian strata (Hutchison, 1982, pp. 23-24), and 4) the probability that rocks in this unit were derived in part from Jurassic to Mississippian(?) strata in unit "JMsv." Regional metamorphism of rocks in this complex occurred during Late Cretaceous to Early Tertiary time (Forbes and Engels, 1970; Smith and Diggles, 1981; Gehrels and others, 1985), in part prior to Late Triassic time (Souther, 1971), and perhaps at other times. This unit consists of unnamed metamorphic rocks in the Coast Mountains east of Ketchikan (Berg and others, 1978; Elliott and Koch, 1981), east of Petersburg (Brew and others, 1984; Souther and others, 1979), near Tracy Arm (Brew and Grybeck, 1984), near Juneau (Ford and Brew, 1973; Brew and Ford, 1977; Souther and others, 1979; Brew and Morrell, 1980), southeast of Haines (Souther and others, 1979; Werner, 1978), and near Skagway (Redman and others, 1984).

pTmc METAMORPHIC COMPLEX (DERIVED FROM CARBONATE ROCKS) (PRE-TERTIARY) -- Amphibolite- and locally granulite-facies carbonate rocks and carbonate-rich clastic strata that occur as discontinuous marble lenses, as thick, continuous marble layers, and as calc-silicate gneiss. Age is constrained by intercalation with the pre-Tertiary metasedimentary and metavolcanic rocks (pTmsv) described above. This unit occurs in the Coast Mountains northeast of Petersburg (Brew and others, 1984), and southeast of Juneau (Brew and Grybeck, 1984).

KJsd DISRUPTED UNIT (SEDIMENTARY ROCKS) (CRETACEOUS AND JURASSIC) -- Regionally deformed, disrupted, and metamorphosed graywacke, siltstone, mudstone, and subordinate chert, limestone, and volcanic and intrusive rocks, that belong to a structural complex in the Kupreanof-Etolin Islands area (Brew and others, 1984). Common metamorphic rock types include sub-greenschist- to greenschist-facies graywacke semischist, phyllite, argillite, slate, and subordinate greenstone, greenschist, and marble. The complex consists of blocks up to several kilometers in length of Triassic sedimentary and volcanic rocks (Ŧsv), Permian carbonate (Pc), and Devonian carbonate (Dc) enclosed in a matrix of strata belonging to the Stephens Passage Group (Cretaceous and Jurassic), and perhaps the Cannery Formation (Mississippian and Late Devonian: Jones and others, 1981) and other Mesozoic or Paleozoic units (D.A. Brew, in U.S. Geological Survey, 1982; Brew and others, 1984; H.C. Berg, unpub. field data, 1978). The complex probably formed by tectonic and (or) sedimentary processes operating during deposition of the Stephens Passage Group (KJs) (D.A. Brew, in U.S. Geological Survey, 1982). We have therefore assigned an age of Cretaceous and Jurassic for the formation of the complex. Faulting during Tertiary time may have further disrupted the complex. These rocks are not referred to as a melange because the structural and stratigraphic relations between the blocks and the matrix, and the degree to which the matrix strata are disrupted, are uncertain.

KJvd DISRUPTED UNIT (VOLCANIC ROCKS) (CRETACEOUS AND JURASSIC) -- Regionally deformed, disrupted, and metamorphosed intermediate to mafic and minor felsic volcanic rocks that belong to a structural complex on Kupreanof, Zarembo, and adjacent smaller Islands (Brew and others, 1984). Rocks are generally metamorphosed to greenschist and greenstone. Relict pyroxene phenocrysts suggest that the metavolcanic rocks were probably derived from Cretaceous and Jurassic volcanic rocks in unit "KJv" (Brew and others, 1984). Metavolcanic rocks of Triassic (Ŧv) and Mississippian and Late Devonian (MDsv) age may also be included.

pOmsv METAMORPHIC COMPLEX (DERIVED FROM SEDIMENTARY AND VOLCANIC ROCKS) (PRE-ORDOVICIAN) -- Greenschist- and locally amphibolite-facies basaltic to andesitic pillow flows, pillow breccia, and tuff breccia, volcaniclastic graywacke and mudstone, and minor limestone and felsic volcanic breccia and tuff, that belong to a metamorphic complex. Common rock types include greenschist, greenstone, quartzo-feldspathic schist, and marble. Radiometric age constraints suggest that the metamorphism occurred during Early Ordovician to Middle Cambrian time (Turner and others, 1977; J.B. Saleeby and G.E. Gehrels, unpub. data; M.A. Lanphere, oral commun., 1984). A pre-Ordovician depositional age is indicated by metaplutonic rocks (O€dg) on eastern Prince of Wales Island that have yielded preliminary U/Pb (zircon) apparent ages of Cambrian (J.B. Saleeby, oral commun., 1983). The unit consists of part of the Wales Group on south-central Prince of Wales Island (Eberlein and others, 1983; Gehrels and others, 1983; Redman, 1981), and unnamed rocks on small islands south of Gravina Island (G.E. Gehrels, unpub. mapping, 1984).

| pOmc | METAMORPHIC COMPLEX (DERIVED FROM CARBONATE ROCKS) (PRE-ORDOVICIAN) -- Greenschist- and locally amphibolite-facies marble layers and lenses that belong to a metamorphic complex. A pre-Ordovician depositional age and an Early Ordovician to Late Cambrian metamorphic age are indicated by intercalation with the metamorphic rocks described above in unit "pTmsv." This unit consists of part of the Wales Group on southern Prince of Wales Island (Eberlein and others, 1983).

Intrusive units

| Tgb | GABBRO (MIOCENE AND OLIGOCENE) -- Layered and locally zoned bodies of two-pyroxene ± olivine ± biotite ± hornblende ± quartz gabbro, and subordinate troctolite, peridotite, leucogabbro, diorite, and tonalite. This unit consists of stocks on Revillagigedo Island and the adjacent mainland (Berg and others, 1978; Koch and Elliott, 1984),

northern Kupreanof and Kuiu Islands (Brew and others, 1984), Chichagof Island (Johnson and Karl, 1982; Loney and others, 1975), and north of Cross Sound (Brew and others, 1978). The stocks on Revillagigedo Island and adjacent areas of the mainland have yielded K-Ar apparent ages of late Oligocene and early Miocene (Smith and Diggles, 1981), and the large body (La Perouse gabbro) northwest of Cross Sound has yielded an Oligocene Ar^{40}/Ar^{39} apparent age (Loney and Himmelberg, 1983).

Tgr GRANITE (MIOCENE AND OLIGOCENE) -- Biotite ± hornblende ± pyroxene granite, alkalai granite, quartz monzonite, and subordinate syenite, granodiorite, and diorite. This unit consists of stocks on the mainland southeast of Revillagigedo Island that have yielded K-Ar apparent ages of early Miocene and Oligocene (Berg and others, 1978), large plutons of Miocene and (or) Oligocene age on Etolin, Zarembo, Kupreanof, and Kuiu Islands (Brew and others, 1984), and part of the Oligocene Tkope River intrusions along the Alaska-British Columbia border north of Glacier Bay (MacKevett and others, 1974; Campbell and Dodds, 1983).

Tgd GRANODIORITE (OLIGOCENE AND EOCENE) -- Biotite ± hornblende ± muscovite ± garnet granodiorite, granite, quartz monzonite, tonalite, and quartz diorite. This unit consists of muscovite- and locally garnet-bearing granodiorite, granite, and tonalite on Baranof and Chichagof Islands and in Glacier Bay. Where dated, these intrusive bodies have yielded Eocene K-Ar apparent ages (Loney and others, 1975; Decker and Plafker, 1982; Johnson and Karl, 1982; Brew and others, 1978). In the Chilkat Range and on southeastern Baranof Island the intrusive bodies consist of biotite- and hornblende-bearing quartz diorite and granodiorite that have yielded Oligocene K-Ar apparent ages

(Loney and others, 1975; MacKevett and others, 1974).

Tg GRANODIORITE (EOCENE) -- Biotite-dominant, hornblende- and sphene-bearing granodiorite and subordinate quartz monzonite, quartz diorite, and leucogranite that occurs in the Coast Mountains (Berg and others, 1978; Elliott and Koch, 1981; Brew and others, 1984; Brew and Grybeck, 1984; Brew and Ford, 1977; Brew and Morrell, 1980; Souther and others, 1979; Redman and others, 1984; Fred Barker and Joe Arth, written commun., 1984). Concordant K-Ar apparent ages on hornblende and biotite (Forbes and Engels, 1970; Smith and others, 1979) and preliminary U/Pb (zircon) apparent ages (Gehrels and others, 1985) indicate emplacement during Eocene time.

Tgt GRANODIORITE AND TONALITE (PALEOCENE) -- Biotite-dominant, hornblende-bearing granodiorite and tonalite with local foliation and layering. Preliminary U/Pb (zircon) geochronometry indicates emplacement during Paleocene time (Gehrels and others, 1985). This unit occurs in the Coast Mountains east of Petersburg (Brew and others, 1984), east of Juneau (Brew and Ford, 1977; Gehrels and others, 1985), and near Skagway (Fred Barker and Joe Arth, written commun., 1984).

TKg GRANODIORITE (PALEOCENE? AND CRETACEOUS?) -- Diverse assemblage of generally foliated and layered granodiorite, quartz monzonite, and tonalite, and their metamorphic equivalents. May include a significant component of pre-Tertiary metasedimentary and metavolcanic rocks (pTmsv) in some areas. Where recognized, such areas are indicated by a stipple pattern. Constraints on the age of these rocks are provided by cross-cutting Eocene granodiorite plutons, and the interpretation that some rocks in this unit have experienced a regional Late Cretaceous and early Tertiary metamorphic event (Forbes and Engels, 1970; Smith and Diggles, 1981). We suspect, however, that most rocks in this unit are Paleocene in age and correlative with rocks in unit "Tgt." The unit occurs in the Coast Mountains east and north of Ketchikan (Berg and others, 1978; Elliott and Koch, 1981), east of Petersburg (Brew and others, 1984), east of Admiralty Island (Brew and Grybeck, 1984), and in the Skagway area (Fred Barker and Joe Arth, written commun., 1984).

TKt TONALITE (PALEOCENE AND LATE CRETACEOUS) -- Hornblende-dominant, biotite-bearing tonalite and subordinate quartz diorite in steeply dipping, foliated, and locally lineated sills in the Coast Mountains (Brew and Ford, 1981; Brew and Morrell, 1980). This unit occurs east and north of Ketchikan (Berg and others, 1978; Elliott and Koch, 1981), east of Petersburg (Brew and others, 1984), east of Admiralty Island (Brew and Grybeck, 1984; Brew and Ford, 1977; Brew and Morrell, 1980), and north of Haines (MacKevett and others, 1974; Redman and others, 1984; Robertson, 1959). Field and preliminary U/Pb (zircon) geochronologic data indicate emplacement in Late Cretaceous and early Tertiary time, during the waning stages of deformation and metamorphism in the Coast Mountains (Gehrels and others, 1985; Brew and Ford, 1981).

Kgt GRANODIORITE AND TONALITE (LATE CRETACEOUS) -- Small plutons to batholiths of granodiorite, tonalite, and subordinate quartz monzonite to quartz diorite and diorite. Most bodies contain biotite and hornblende, many have magmatic epidote and garnet and are plagioclase porphyritic, and some contain pyroxene and (or) muscovite. K-Ar apparent ages of these bodies are generally Late Cretaceous (Smith and Diggles, 1981; Brew and others, 1984). This unit consists of plutons on Revillagigedo Island and the adjacent mainland (Berg and others, 1978; Eberlein and others, 1983), in the Etolin-Kupreanof Islands area (Brew and others, 1984; Burrell, 1984; Buddington and Chapin, 1929), on the mainland east of Admiralty Island (Buddington and Chapin, 1929; Brew and Grybeck, 1984; Souther and others, 1979), and perhaps in the Haines region (MacKevett and others, 1974; Redman and others, 1984). Plutons that are mineralogically or compositionally different from the main suite of intrusive bodies are queried on the geologic map.

$\boxed{\text{Kum}}$ ULTRAMAFIC ROCKS (EARLY CRETACEOUS AND CRETACEOUS?) -- Ultramafic intrusive bodies of magnetite-bearing hornblende clinopyroxenite and subordinate dunite, peridotite, and hornblendite (Taylor, 1967). Several of the complexes are concentrically zoned from a core of dunite to rocks containing progressively less olivine and more hornblende and magnetite. The zoned bodies commonly intrude a two-pyroxene gabbro that may be genetically related (Irvine, 1974) or unrelated (Taylor, 1967) to the ultramafic rocks. Geological and geochemical considerations suggest that rocks in these bodies may be genetically related to Cretaceous and Jurassic volcanic rocks (KJv) (Berg and others, 1972; Irvine, 1973). K-Ar apparent ages of the ultramafic rocks indicate emplacement during Early Cretaceous time (Lanphere and Eberlein, 1966). Bodies belonging to this suite occur on Duke Island (Irvine, 1974), Annette and Revillagigedo Islands (Berg and others, 1978), small islands west of Etolin Island and Kupreanof Island (Brew and others, 1984), and on the mainland near Myers Chuck (Ruckmick and Noble, 1959), Tracy Arm (Brew and Grybeck, 1984), and Klukwan (MacKevett and others, 1974).

Undated ultramafic rocks provisionally assigned to this unit on the basis of similar lithology are queried on the geologic map and consist of: hornblendite and hornblende pyroxenite on Revillagigedo Island (Berg and others, 1978); pyroxenite, gabbro, and hornblendite on east-central Prince of Wales Island (Eberlein and others, 1983); peridotite, dunite, and pyroxenite in the Coast Mountains near Tracy Arm (Brew and Grybeck, 1984; Grybeck and others, 1977); hornblendite and pyroxene- and hornblende-bearing gabbro on eastern Admiralty Island (Lathram and others, 1965); serpentinized peridotite or pyroxenite on north-central Admiralty Island (Lathram and others, 1965); and peridotite and serpentinite on eastern Baranof Island (Loney and others, 1975).

Kg GRANODIORITE (EARLY CRETACEOUS) -- A heterogeneous suite of plutons consisting primarily of biotite, hornblende, magnetite ± pyroxene ± garnet granodiorite and subordinate quartz monzonite, tonalite, and quartz diorite. On Chichagof Island these plutons are associated with Early Cretaceous gabbro and diorite. Many intrusive bodies show a crude zoning from leucocratic rocks in the interior, to progressively more basic rocks toward the margins. K-Ar apparent ages record emplacement primarily during Early Cretaceous time, although similar intrusive bodies northwest of Glacier Bay in Canada have also yielded Late Jurassic K-Ar apparent ages (Chris Dodds, oral commun., 1984). This unit occurs on Dall Island (G.E. Gehrels, unpub. mapping, 1984), Prince of Wales Island and adjacent smaller islands (Eberlein and others, 1983; Brew and others, 1984), Kuiu Island (Brew and others, 1984), Admiralty Island (Lathram and others, 1965), Chichagof Island (Loney and others, 1975; Johnson and Karl, 1982), north of Cross Sound (Decker and Plafker, 1982; George Plafker, written commun., 1984), in the Glacier Bay area (Brew and others, 1978), and in the Chilkat Range (Lathram and others, 1959; Sonnevil, 1981; MacKevett and others, 1974; Robertson, 1959). Also included is the Moth Bay pluton on southern Revillagigedo Island (Berg and others, 1978), which may alternatively belong to unit "Kgt."

$\boxed{\text{Kd}}$ DIORITE (EARLY CRETACEOUS) -- Primarily hornblende ± biotite ± clinopyroxene diorite with subordinate quartz diorite and gabbro. Generally occurs with Early Cretaceous granodiorite and gabbro. This unit occurs on Prince of Wales Island (Eberlein and others, 1983; Gehrels and others, 1983), and on Chichagof Island (Loney and others, 1975; Johnson and Karl, 1982). As described above for unit "Kg," some rocks in this unit may be Late Jurassic in age.

$\boxed{\text{Kgb}}$ GABBRO (EARLY CRETACEOUS) -- Primarily clinopyroxene (generally augite) ± hornblende ± biotite ± olivine gabbro, leucogabbro, and subordinate norite, syenite and pyroxenite. This unit occurs on Prince of Wales Island (Eberlein and others, 1983), Kuiu Island (Brew and others, 1984), and Chichagof Island, where it is associated with Early Cretaceous granodiorite and diorite (Loney and others, 1975; Johnson and Karl, 1982). As described above for unit "Kg," some rocks in this unit may be Late Jurassic in age.

$\boxed{\text{KJd}}$ DIORITE (CRETACEOUS AND JURASSIC?) -- Metamorphosed and moderately deformed diorite and subordinate quartz diorite and gabbro on Annette Island, Revillagigedo Island, and the mainland to the northwest (Berg and others, 1978). A small quartz diorite body near northern Annette Island has yielded a Cretaceous K-Ar apparent age: the other bodies are undated. We have assigned a Cretaceous and Jurassic(?) age based on the interpretation that rocks in this unit are genetically related to Cretaceous and Jurassic volcanic rocks in units "KJv" and "KJsv."

KJgb GABBRO (EARLY CRETACEOUS AND JURASSIC?) -- Two-pyroxene gabbro and subordinate hornblende ± biotite gabbro and diorite. These intrusive rocks generally occur adjacent to the Early Cretaceous zoned ultramafic bodies. Gabbro on Duke Island has yielded a Middle Jurassic K-Ar apparent age on biotite (Smith and Diggles, 1981), although field relations suggest that this gabbro may be in part of Silurian or Ordovician age (G.E. Gehrels, unpub. mapping, 1984). This unit occurs on Duke Island (Irvine, 1974; Taylor, 1967) and on the mainland near Myers Chuck (Eberlein and others, 1983). We tentatively include the large gabbro and diorite body near Klukwan based on its close association with an Early Cretaceous ultramafic body (MacKevett and others, 1974; Taylor, 1967; Redman and others, 1984).

Jgr GRANITE (MIDDLE JURASSIC) -- Peralkaline aegerine and arfvedsonite granite at Bokan Mountain on southern Prince of Wales Island, which has yielded a Middle Jurassic U/Pb (zircon) apparent age (MacKevett, 1963; Saint-André and others, 1983).

Jt TONALITE (MIDDLE? JURASSIC) -- Biotite and hornblende tonalite and quartz diorite on Chichagof Island that has yielded a Middle Jurassic K-Ar apparent age on hornblende (Loney and others, 1975).

J$\overline{\mathrm{R}}$d DIORITE (JURASSIC OR TRIASSIC) -- Hornblende and biotite diorite (Jualin Diorite) along the east shore of Lynn Canal (Knopf, 1911; Redman, 1984a). The diorite hornfelses adjacent metasedimentary and metavolcanic rocks that are probably Triassic or Permian in age, and is overlain unconformably by less metamorphosed strata belonging to unit

"KJs" (Redman, 1984a). These relations suggest an emplacement age of Jurassic or Triassic for the Jualin diorite. As discussed above under unit "KPsv," these relations also suggest that some rocks in "KPsv" and associated units were metamorphosed and deformed prior to the deposition of strata of Cretaceous and Jurassic age.

| T̶Rg | GRANODIORITE (TRIASSIC) -- Hornblende and biotite granodiorite and minor quartz diorite (Texas Creek granodiorite) northeast of Revillagigedo Island (Berg and others, 1978) that has yielded a minimum K-Ar apparent age of Triassic from hornblende (Smith and Diggles, 1981).

| PIPsy | SYENITE (EARLY PERMIAN AND (OR) LATE PENNSYLVANIAN) -- Biotite ± amphibole ± aegerine ± augite syenite on Sukkwan Island that has yielded K-Ar apparent ages of late Pennsylvanian from biotite and Early Permian from hornblende, and an undated but lithically similar body on south-central Prince of Wales Island (Eberlein and others, 1983).

| Sst | SYENITE AND TRONDHJEMITE (SILURIAN) — Leucocratic biotite ± aegerine ± arfvedsonite ± garnet syenite and subordinate diorite on southern Prince of Wales Island (Gehrels and others, 1983); biotite and hornblende trondhjemite on Annette and Gravina Islands and the mainland to the southeast (Gehrels and others, 1983; Berg and others, 1978); and undivided biotite and (or) hornblende syenite and trondhjemite on Chichagof Island (Loney and others, 1975). A Silurian age is indicated by a minimum K-Ar apparent age on Chichagof Island (Lanphere and others, 1965) and preliminary U/Pb (zircon) apparent ages of rocks from Annette, Gravina, and Prince of Wales Islands and the mainland to the southeast (G.E. Gehrels and J.B. Saleeby, unpub. data).

SOum ULTRAMAFIC ROCKS (SILURIAN AND ORDOVICIAN?) -- Pyroxenite, hornblendite, and related ultramafic rocks on southern Prince of Wales Island (MacKevett, 1963; G.E. Gehrels, unpub. mapping) and on southern Dall Island (G.E. Gehrels, unpub. mapping, 1984). Ultramafic rocks on southern Prince of Wales Island are interpreted to be Silurian based on gradational relations with syenitic rocks of Silurian age (Sst), and the intrusive body on Dall Island has yielded a Late Ordovician K-Ar apparent age from hornblende (M.A. Lanphere, written commun., 1984).

SOq QUARTZ DIORITE (EARLY SILURIAN TO MIDDLE ORDOVICIAN) -- Late and Middle Ordovician diorite and quartz diorite, and Early Silurian to Late Ordovician biotite-quartz monzonite on Prince of Wales Island (Gehrels and others, 1983; Eberlein and others, 1983; Saleeby and others, 1984; Lanphere and others, 1964; J.B. Saleeby and G.E. Gehrels, unpub. data); diorite and quartz diorite of Early Silurian to Late Ordovician age on Gravina, Annette, and Duke Islands (Gehrels and others, 1984; G.E. Gehrels and J.B. Saleeby, unpub. data); and layered and foliated quartz diorite on southern Dall Island. These rocks are interpreted to be genetically related to Early Silurian and Ordovician volcanic rocks (SOv and SOsv) (Eberlein and others, 1983).

Ogb GABBRO (MIDDLE? ORDOVICIAN AND ORDOVICIAN?) -- Hornblende gabbro and subordinate hornblende pyroxenite and hornblendite on Sukkwan Island that has yielded a Middle Ordovician K-Ar apparent age (Eberlein and others, 1983), and an undated metagabbro on southern Dall Island that is interpreted to be correlative (G.E. Gehrels, unpub. mapping, 1984).

O€dg DIORITE AND GRANODIORITE (EARLY ORDOVICIAN AND CAMBRIAN) -- Foliated and metamorphosed hornblende diorite and biotite ± hornblende granodiorite on Prince of Wales Island that has yielded preliminary U/Pb (zircon) apparent ages of Early Ordovician and Cambrian (J.B. Saleeby and G.E. Gehrels, unpub. data).

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