STRUCTURAL GEOLOGY AND PROCESSES OF DEFORMATION IN THE MESOZOIC AND CENOZOIC EVOLUTION OF THE QUEEN CHARLOTTE ISLANDS

by

Peter DuBois Lewis

B.S., Stanford University, 1984
M.Sc., University of British Columbia, 1987

A THESIS SUBMITTED IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF DOCTOR OF PHILOSOPHY

in

THE FACULTY OF GRADUATE STUDIES
Geological Sciences

We accept this thesis as conforming to the required standard

THE UNIVERSITY OF BRITISH COLUMBIA

June, 1991
© Peter DuBois Lewis, 1991
In presenting this thesis in partial fulfilment of the requirements for an advanced degree at the University of British Columbia, I agree that the Library shall make it freely available for reference and study. I further agree that permission for extensive copying of this thesis for scholarly purposes may be granted by the head of my department or by his or her representatives. It is understood that copying or publication of this thesis for financial gain shall not be allowed without my written permission.

Department of Geological Sciences

The University of British Columbia
Vancouver, Canada

Date June 28, 1991
Abstract

The Queen Charlotte Islands, located within the Insular Belt off the central British Columbia coast, preserve a lithologically diverse group of volcanic, sedimentary, and intrusive rocks ranging in age from Late Paleozoic to Late Tertiary. This study analyzes the deformation history recorded in these rocks through regional and detailed structural analyses of five map areas comprising 20% of the islands' landmass.

The structural model constrained by these studies involves at least four distinct deformation events. Earliest formed structures are Middle Jurassic northwest-trending, southwest-verging folds and contractional faults which accommodated approximately 50% regional structural shortening. During the late Middle Jurassic to Early Cretaceous, dip-slip motion along steep faults tectonically elevated northwest-trending fault blocks, exposing uplifted areas to preferential erosion of Middle Jurassic and older strata. In the Late Cretaceous to Early Tertiary, a second episode of northeast-southwest shortening added an additional 10% regional shortening, concentrated in high-strain zones in Cretaceous strata overlying re-activated block faults. Tertiary structural styles were dominated by strike-slip and extensional faulting, coinciding with the onset of syntectonic sedimentation in the adjacent Queen Charlotte Basin.

Structural styles associated with these events reflect the variable mechanical properties of the deforming units. During periods of crustal shortening, most regional strain was accommodated by movement along steeply-dipping reverse faults in homogeneous, poorly-layered rocks, while well-bedded rocks shortened by mesoscopic to macroscopic folding and movement along shallowly-dipping thrust faults. Strike-slip and extensional faulting episodes were characterized by displacements along steeply-dipping fault surfaces in all units. Penetrative structural fabrics preserved in Mesozoic rocks include axial planar slaty cleavage, mylonitic foliation, and wall-rock foliations
adjacent to major fault surfaces, and all can be tied to some aspect of the documented structural history. Spaced cleavage fabrics in Cretaceous mudstones are interpreted to have formed as dilatent fractures during sediment dewatering and lithification.

Throughout deformation, strains were accommodated by semi-brittle processes of crystalline plasticity (dislocation glide and dislocation creep, twin gliding), pressure solution creep, and brittle failure (extensional and shear fracturing, microcracking, and cataclastic flow). Regional metamorphic conditions were sub-greenschist grade, except for localized areas affected by pluton heating.

This structural history includes substantial refinements of existing models for the structural evolution of the Queen Charlotte Islands, the most significant of which include:

1.) the characterization of Mesozoic contractional and extensional deformation events, and the recognition of their significance to the present stratigraphic distribution and structural configuration,

2.) the delineation of a concentrated strain zone in the central Queen Charlotte Islands, and the recognition that its formation was linked to Mesozoic contraction and extensional events, rather than Tertiary strike-slip faulting as previously interpreted,

3.) the evolution of a general structural model for the Queen Charlotte Islands region in Tertiary time integrating structural, magmatic, sedimentological, and geophysical constraints.
Table of Contents

Abstract ................................................................................................................................. ii
Table of Contents ................................................................................................................ iv
List of Map Plates ................................................................................................................. ix
List of Figures ....................................................................................................................... x
List of Tables ....................................................................................................................... xiii
Acknowledgements ............................................................................................................. xiv
Preface .................................................................................................................................... xv

Part 1: Introduction, Tectonic Setting, Previous Geologic Explorations, and Stratigraphy of the Queen Charlotte Islands ......................................................................................... 1

1. Previous Geologic Investigations .................................................................................. 3
2. Tectonic Setting of the Queen Charlotte Islands ........................................................... 4
3. Project outline ............................................................................................................... 8
4. Stratigraphy .................................................................................................................... 10
   4.1 Assemblage 1: Permian ....................................................................................... 13
   4.2 Assemblage 2: Upper Triassic to lower Middle Jurassic ........................................ 14
      4.2.1 Karmutsen Formation ............................................................................. 14
      4.2.2 Kunga Group ......................................................................................... 17
         Sadler Limestone ......................................................................................... 17
         Peril Formation ......................................................................................... 19
         Sandilands Formation ............................................................................... 20
      4.2.3 Maude Group ........................................................................................ 21
         Ghost Creek Formation ........................................................................... 21
         Fannin Formation .................................................................................... 22
         Whiteaves Formation ............................................................................. 23
         Phantom Creek Formation ................................................................... 23
   4.3 Assemblage 3: Middle Jurassic Volcanic and Sedimentary Rocks ......................... 25
      4.3.1 Yakoun Group ....................................................................................... 25
      4.3.2 Moresby Group .................................................................................... 26
4.4 Assemblage 4: Cretaceous marine sedimentary rocks ........................................ 29
  4.4.1 Longarm Formation .................................................................................. 29
  4.4.2 Queen Charlotte Group ........................................................................... 30
    Haida Formation ............................................................................................ 32
    Skidegate Formation ...................................................................................... 33
    Honna Formation ........................................................................................... 37
    Unnamed volcanic rocks ................................................................................. 39
    Unnamed sedimentary rocks ......................................................................... 39

4.5 Assemblage 5: Tertiary volcanic and sedimentary rocks ................................. 41
  4.5.1 Unnamed (Paleogene) sedimentary rocks .............................................. 41
  4.5.3 Masset Formation .................................................................................... 44

4.7. Metamorphic Geology .................................................................................. 48
  Discussion .......................................................................................................... 49

Part 2: Mesozoic and Cenozoic Structural Geology of the Queen Charlotte Islands ........................................ 52

1. Introduction ...................................................................................................... 52

2. Map Descriptions ............................................................................................ 57

  2.1 Northwest Graham Island ............................................................................ 57
    2.1.1 Southern Langara Island ........................................................................ 57
    2.1.2 Caswell Point to Newcombe Hill ............................................................ 58
    2.1.3 South Sialun Bay .................................................................................... 61
    2.1.4 Fleurieu Point ......................................................................................... 62
    Microscopic Structures ..................................................................................... 62
    2.1.5 Kennecott Point ..................................................................................... 66
    2.1.6 Discussion .............................................................................................. 68

  2.2 Central Queen Charlotte Islands ................................................................. 70
    2.2.1 Rennell Sound/Shields Bay .................................................................... 70
      Stratigraphic Distribution .............................................................................. 70
      Structural geology ......................................................................................... 71
      Discussion ..................................................................................................... 78
    2.2.2 Long Inlet/Kagan Bay Map Area .............................................................. 81
      Stratigraphic Distribution .............................................................................. 81
      Structural Geology ......................................................................................... 83
      Discussion ..................................................................................................... 91
    2.2.3 Southeastern Louise Island Map Area .................................................... 99
      Stratigraphic distribution .............................................................................. 99
2.2.5 Sandspit Fault (Copper Bay) .................................................. 123
2.3 Southern Queen Charlotte Islands ............................................. 132
  2.3.1 Burnaby Island/Juan Perez Sound Map Area ....................... 132
    Stratigraphic Distribution .................................................. 135
    Structural Geology ....................................................... 138
      Western Domain ......................................................... 138
      Central (fault zone) Domain ........................................... 139
      Slim Inlet and Smithe Point shear zones: ......................... 140
      Eastern Domain ....................................................... 147
    Discussion ............................................................. 150
    Relationships to structures outside the map area ................. 152
    A structural model for Burnaby Island/Juan Perez Sound and adjacent areas ......................... 155
    Amounts of fault offset ............................................... 156
3. Strain measurement studies .................................................. 161
  3.1 Introduction .................................................................. 161
  3.2 Results and Discussion .................................................. 164
4. Structural Synthesis: A model for the Mesozoic and Cenozoic evolution of the Queen Charlotte Islands .................................................. 167
  4.1 Pre-Middle Jurassic Deformation ....................................... 168
  4.3 Late Jurassic - Early Cretaceous Block Faulting .................... 173
  4.5 Tertiary Fault-Dominated Deformation .................................. 181
    4.5.1 Middle Eocene to Early Oligocene (Approximately 50 Ma-35 Ma) .................................................. 182
    4.5.2 Early Oligocene to Early Miocene (Approximately 35Ma-20Ma) .................. 186
    4.5.3 Early Miocene to Recent ( < 20 Ma) ............................ 191
  4.6 Implication to Tertiary Queen Charlotte Basin evolution ......... 193
Part 3: Processes of Deformation in Mesozoic and Cenozoic rocks of the Queen Charlotte Islands

1. General Statement ........................................................................................................195
2. Mesoscopic to megascopic structures .........................................................................197
   2.1 Folding ..................................................................................................................197
   2.2 Faulting and fracturing .........................................................................................200
3. Penetrative fabrics ........................................................................................................205
   3.1 Mylonitic, Fault fabrics .......................................................................................205
      3.1.1 Buck Channel Pluton ....................................................................................205
   3.2.1 Louise Island, Rennell Sound, Louscoone Inlet fault fabrics .........................211
      3.2. Slaty cleavage ..................................................................................................213
         3.2.1 Mesoscopic, microscopic descriptions .......................................................213
         3.2.2 X-ray phyllosilicate texture analysis .........................................................216
         3.2.3 Discussion ..................................................................................................223
      3.3. Spaced cleavage ..............................................................................................225
         3.3.1 Mesoscopic, microscopic descriptions .......................................................225
         3.3.2 X-ray phyllosilicate texture analysis .........................................................230
         3.3.3 Discussion ..................................................................................................233
4. Summary ......................................................................................................................238

Part 4: Regional synthesis: Triassic to Late Tertiary Geologic Evolution of the Central Insular Belt and Adjacent Parts of the Canadian Cordillera, from a Queen Charlotte Islands Perspective ........................................................................................................................239

1. Introduction ..................................................................................................................239
2. Pre-Middle Jurassic ......................................................................................................241
   2.1 Terrane Affinities .................................................................................................241
   2.2 Magmatic and stratigraphic setting .......................................................................242
3. Middle and Late Jurassic ............................................................................................245
   3.1 Middle Jurassic Deformation ...............................................................................245
   3.2 Arc magmatism on the Queen Charlotte Islands and adjacent areas ..................247
4. Cretaceous marine sedimentation and tectonic relations ........................................250
5. Tertiary Basin formation ............................................................................................252
6. Summary ......................................................................................................................255
Appendix A: Chronologic Review of Recent Publications on the Structural Evolution of the Queen Charlotte Islands Region .......................................................................................................................... 276

Appendix B: Strain analysis techniques .................................................................................................................. 279

Appendix C: X-ray Texture Analysis of Phyllosilicates, and March Strain Analysis ............................................. 285
  C.1 Introduction .................................................................................................................................................. 285
  C.2 Analytical Procedure .................................................................................................................................. 285
  C.4 FACTOR.PAS Program Listing: .................................................................................................................. 312

Appendix D: Fault Slip Analysis (Section Cove Shear Zone) ................................................................................. 322

Appendix E: Field Station Compilation .................................................................................................................. 324
List of Map Plates
(all map plates included in back pocket)

Plate I: Shoreline geology of northwest Graham Island and parts of Langara Island
Plate II: Detailed structural geology of a) South Sialun Bay; b) Fleurieu Point; c) Kennecott Point
Plate III: Structural geology of the Rennell Sound/Shields Bay map area
Plate IV: Structural geology of the Long Inlet/Kagan Bay map area
Plate V: Structural geology of the southeast Louise Island map area
Plate VI: Structural transect through Skidegate Inlet and Skidegate Channel
Plate VII: Structural geology of the Burnaby Island/Juan Perez Sound map area (2 sheets)
List of Figures

Part 1:

Figure 1.1: Location map for the Queen Charlotte Islands region ........................................... 2
Figure 1.2: Principal tectonic features of the Queen Charlotte Islands region .......................... 7
Figure 1.3: Chronostratigraphic chart showing major lithologic units found in the Queen Charlotte Islands ...................................................................................................................... 11
Figure 1.4: General geologic map of the Queen Charlotte Islands ........................................... 12
Figure 1.5: Karmutsen Formation lithotypes a) pillowed flows; b) stratified amygduloidal flows. 16
Figure 1.6: Sadler Limestone lithotypes a) thickly bedded to massive grey limestone; b) cherty limestone ......................................................................................................................... 18
Figure 1.7: Thinly-bedded Sandilands Formation lithology ......................................................... 20
Figure 1.8: Maude Group lithotypes: a) fissile grey shale (Ghost Creek Formation); b) calcereous siltstone (Fannin Formation) ..................................................................................................... 24
Figure 1.9: Yakoun Group lithotypes: a) lapilli tuff beds; b) lithic wacke sandstone ................. 27
Figure 1.10: Moresby Group cross-bedded conglomerate ....................................................... 28
Figure 1.11: Longarm Formation lithotypes: a) pebble conglomerate; b) black/white granular sandstone ................................................................................................................................. 31
Figure 1.12: Haida and Skidegate Formation lithotypes: a) cross stratified sandstone; b) concretionary shale c) turbiditic sandstone; d) slumped turbidite beds ................................................ 35
Figure 1.13: Honna Formation cobble conglomerate overlying Skidegate Formation mudstones. 38
Figure 1.14: Unnamed Cretaceous volcanic unit: a) basal debris flows conformably overlying Honna Formation; b) individual flow units ............................................................... 40
Figure 1.15: Unnamed Paleogene black shale unit ................................................................. 42
Figure 1.16: Bouldery lahar deposits of unnamed Paleogene volcanic unit ......................... 43
Figure 1.17: Cross-cutting dykes, Louise Island ........................................................................ 47

Part 2:

Figure 2.1: Map study areas and major structural features of the Queen Charlotte Islands ........ 54
Figure 2.2: Stereographic projections of structural data, Northwest Graham Island .............. 60
Figure 2.3: Fault duplex in Peril Formation, Fleurieu Point .................................................... 63
Figure 2.4: Recumbent isocline with axial-planar cleavage, Fleurieu Point ............................ 65
Figure 2.5: Photomicrographs showing elements defining axial-planar cleavage .................. 65
Figure 2.6: Schematic diagram showing bed rotation during faulting, Kennecott Point ........... 67
Figure 2.7: Recumbent isocline in Peril Formation, Rennell Sound ......................................... 73
Figure 2.8: Stereographic projections of structural elements, Rennell Sound ......................... 74
Figure 2.9: Folded calcite veins in foliated Sadler Limestone, Rennell Sound ......................... 75
Figure 2.10: Photomicrograph of faulted porphyroblast shear indicator, Rennell Sound .......................... 76
Figure 2.11: Detailed sketch map showing fault zone geometry at Rennell Sound................................. 77
Figure 2.12: Palinspastic restoration of southwest-verging reverse faults, Rennell Sound Map area ............................. 80
Figure 2.13: Map showing extents of Long Inlet deformation zone............................................................. 84
Figure 2.14: Stereographic projections of structural elements, Long Inlet/Kagan Bay ........................................ 86
Figure 2.15: Axial-planar cleavage in Skidegate Formation, Long Inlet....................................................... 90
Figure 2.16: Pencil lineation parallel to fold axes, Skidegate Formation, Long Inlet ........................................ 90
Figure 2.17: Sequential palinspastic restorations of northeast-southwest cross section through LIDZ, Long Inlet map area ........................................................................................................... 95
Figure 2.18: Stereographic projections of structural elements, Louise Island......................................... 101
Figure 2.19: Photomicrographs of deformed fecal pellets used in fabric analysis, Louise Island map area ............................................................................................................................................... 103
Figure 2.20: Characteristic southwest-verging fold style displayed at Louise Island, Sandilands Formation ........................................................................................................................................ 105
Figure 2.21: Palinspastic restoration of northeast-southwest cross section through Louise Island map area ................................................................................................................................. 106
Figure 2.22: Angular unconformity between Yakoun Group and Sandilands Formation, Cumshewa Inlet ........................................................................................................................................... 107
Figure 2.23: Diagram showing stratigraphic distribution along Skidegate transect ........................................ 110
Figure 2.24: Ductile/Brittle mylonite zone in Karmutsen Formation, Chaatl Island ........................................ 113
Figure 2.25: Examples of mesoscopic fabrics in Buck Channel pluton ....................................................... 115
Figure 2.26: Photomicrographs of microscopic kinematic indicators, Buck Channel pluton ...................... 116
Figure 2.27: Fold profile section of southwest-verging mesoscopic folds, Maude Island ................................ 118
Figure 2.28: Detailed structural map of Copper Bay brittle fault zone (Sandspit Fault) .................................. 125
Figure 2.29: Typical fracture types at Copper Bay fault zone ........................................................................ 127
Figure 2.30: Arcuate fracture systems, Copper Bay fault zone ...................................................................... 128
Figure 2.31: Schematic diagrams showing possible structural evolution of Copper Bay fault zone .......... 130
Figure 2.32: Regional extent of Louscoone Inlet fault system (LIFS) and location of Burnaby Island/Juan Perez Sound map area .................................................................................................................... 133
Figure 2.33: Pillowed Karmutsen Formation flows, west of LIFS .................................................................. 139
Figure 2.34: Marble tectonite, Smithe Point shear zone ................................................................................. 141
Figure 2.35: Termination of asymmetric limestone boudin, Smithe Point shear zone ................................. 141
Figure 2.36: Sketch map showing folded mylonitic fabric, Slim Inlet shear zone ......................................... 143
Figure 2.37: Photomicrographs of kinematic indicators, Slim Inlet shear zone ........................................... 143
Figure 2.38: Field sketch of cross-cutting brittle features, Section Cove shear zone .................................... 146
Figure 2.39: Shear vein systems, Section Cove shear zone ........................................................................... 146
Figure 2.40: Open fold in Skidegate Formation showing axial plane cleavage, Burnaby Island .................. 149
Figure 2.41: Stereographic projections of structural elements, Louscoone Inlet fault system .................. 149
Figure 2.42: Stereographic projections of structural elements east of Louscoone Inlet fault system... 150
Figure 2.43: Mesoscopic fabrics in mylonites, Shuttle Island .......................................................... 154
Figure 2.44: Schematic diagram showing structural evolution of Burnaby Island map area .......... 156
Figure 2.45: Examples of deformed fossils used in strain quantification study ............................... 163
Figure 2.46: Compilation diagram showing strains measured from deformed fossils ..................... 165
Figure 2.47: Middle Jurassic structural elements, Queen Charlotte Islands region ....................... 171
Figure 2.48: Late Jurassic/Early Cretaceous structural elements, Queen Charlotte Islands region .... 175
Figure 2.49: Late Cretaceous/Early Tertiary structural elements, Queen Charlotte Islands region .... 178
Figure 2.50: Schematic section showing Late Cretaceous/Early Tertiary structural configuration .... 180
Figure 2.51: Eocene to Early Oligocene structural elements, Queen Charlotte Islands region ....... 183
Figure 2.52: Sediment thickness map for Queen Charlotte Basin .................................................. 185
Figure 2.53: Oligocene to Early Miocene structural elements, Queen Charlotte Islands region ....... 187
Figure 2.54: Early Miocene and younger structural elements, Queen Charlotte Islands region ....... 190
Figure 2.55: Seismic reflection profile, Queen Charlotte Sound, showing two-part basin fill ......... 192
Figure 2.56: Seismic reflection profile, Hecate Strait, showing structural inversion of normal fault .. 192

Part 3:

Figure 3.1: Mesoscopic features related to folds in Sandilands Formation .................................. 199
Figure 3.2: Bedding-perpendicular pre-tectonic calcite vein, Ghost Creek Formation ...................... 203
Figure 3.3: Association between stylolites, extension veins, Peril Formation .............................. 204
Figure 3.4: Microscopic deformation textures, Buck Channel pluton ......................................... 208
Figure 3.5: Microscopic deformation textures, Sadler Limestone .............................................. 212
Figure 3.6: Photomicrographs, slaty cleavage fabrics ..................................................................... 215
Figure 3.7: X-ray texture analysis results for slaty cleavage samples ........................................... 219
Figure 3.8: Spaced cleavage fabrics ......................................................................................... 227
Figure 3.9: Spaced cleavage/carbonate concretion relationships .................................................. 228
Figure 3.10: Photomicrographs, spaced cleavage fabrics ............................................................. 229
Figure 3.11: X-ray texture analysis results for spaced cleavage samples ...................................... 231
Figure 3.12: Proposed mechanism for spaced cleavage formation ............................................. 236

Part 4:

Figure 4.1: Stratigraphic columns for Wrangellia and Alexander terrane .................................... 243
Figure 4.2: Time correlation diagram for magmatic history of western Cordillera ....................... 249
Figure 4.3: Regional cross sections showing Mesozoic and Cenozoic tectonic elements for the
            Queen Charlotte Islands region ......................................................................................... 256
Appendices:

Figure A.1: Chart showing progression of publications................................................... 278
Figure B.1: Logarithmic spiral measurements used in strain analysis studies .................. 279
Figure B.2: Fry analysis plots.......................................................................................... 284
Figure D.1: Stereographic projection of fault slip data, Section Cove shear zone ........... 322
Figure D.2: Fault slip solution, Section Cove shear zone ................................................ 323

List of Tables

Table 3.1: X-ray texture analysis strain data, slaty cleavage samples............................... 232
Table 3.2: X-ray texture analysis strain data, spaced cleavage samples........................... 232
Table B.1: Deformed fossil strain data............................................................................... 282
Table D.1: Fault slip data, section Cove shear zone ......................................................... 322
Acknowledgements

My principal research supervisor, John Ross, first introduced me to the complexities of Queen Charlotte Islands geology, and I gratefully acknowledge his support and guidance throughout the duration of my studies at U.B.C. Other members of my supervisory "team", William Barnes, Marc Bustin, Kelly Russell, and Bob Thompson, helped keep my research on course and on schedule.

Superb technical, logistic, and scientific support from the Geological Survey of Canada, coordinated by Bob Thompson and Jim Haggart, kept my enthusiasm at a high level. Many fruitful hours of discussion, formal and informal, with my friends and colleagues at the GSC were instrumental to the development of ideas presented in this study. Drs. Anderson, Haggart, Hickson, Souther, Thompson, Tipper, and Woodsworth of the Queen Charlotte Islands Frontier Geoscience Program have had a particularly strong interest and involvement in my studies, and I thank them for their support. Fossil collection identifications by GSC research staff were essential to the success of the field mapping portion of my project, for which I thank Bruce Cameron, Beth Carter, Jim Haggart, Mike Orchard, and Howard Tipper.

Fellow students involved with the Queen Charlotte Islands project provided the appropriate mix of discussion, shared frustration, and attempted one-upmanship, and for this I thank Tony Fogarassy, Charle Gamba, Jonny Hesthammer, Jarand Indrelid, Henry Lyatsky, and Susan Taite. I thank Shane Dennison, Tanya Hale, Mark Hamilton, Alison Huntley, Dave Mercer, and Jason Miller for their enthusiastic field assistance, and Ella Ferland and Audrey Putterill of Sandspit for superlative expediting services. Permission to work in parts of the Queen Charlotte Islands was kindly granted by Parks Canada and the Haida Nation, whom I hope will find my work useful.

Jeff Fillipone, Richard Chen, and Gerhard Oertel introduced me to the word of X-ray texture analysis, and their coaching allowed me to collect meaningful results in a relatively short time.

Financial support was provided by research contracts from the Geological Survey of Canada and Chevron Canada, and NSERC grants. University of British Columbia graduate fellowships kept me from becoming overly impecunious.

Finally, I wish to thank my loosely-defined "family" for their support: my parents and sister and brother, for maintaining our sometimes unconventional and often impractical family traditions; my grandparents, Carlton and Irene Van de Water, for starting my geological education over fifteen years ago with my first rock collection; Ted and Norma Taite, for the "garage" and free use of the dogs; Steve Garwin, for his continued interest from far away; and finally, Susan Taite, who mercilessly edited a final thesis draft, and helped keep my life in perspective for the last five years.
Preface

During the period from 1987 to 1990, over 30 earth scientists from academic, industrial, and government institutions focussed their research efforts on the Queen Charlotte Islands region, off the west coast of Canada. This multidisciplinary study, spearheaded by the Geological Survey of Canada under the Frontier Geoscience Program (FGP), was formulated to evaluate the geologic evolution of the region, as a guide to addressing the hydrocarbon potential of the Queen Charlotte Basin. This thesis is in large part a contribution to the regional and structural geology component of the Queen Charlotte Basin FGP study. It outlines the deformational history of Upper Paleozoic through Cenozoic rocks exposed in the islands, and its conclusions represent significant changes from existing models for the structural/tectonic history of the Insular Belt.

Results appear in four parts, each intended as a stand-alone contribution. Therefore, some repetition between sections is unavoidable, but has been minimized where possible. Chapter 1 serves as an introduction to the geology of the Queen Charlotte Islands region: it comprises a brief review of previous geologic explorations, a discussion of the tectonic setting of the area, an outline of project goals, and descriptions of the rock units present in the islands. The stratigraphic descriptions emphasize those units exposed in the areas mapped in the present study, and are based on new field and petrographic observations and, in some instances, on published descriptions from elsewhere in the islands.

Readers who are familiar with the stratigraphy and general geology of the Queen Charlotte Islands are encouraged to skip ahead to chapter 2, which is a description and discussion of the structural geology of the region. Ideally, a study of this intended scope should incorporate structural mapping of the entire Queen Charlotte Islands- an undertaking clearly beyond the realm of a thesis study. Fortunately, other workers concurrently involved with the FGP contributed map studies of many parts of the islands.
not examined by this project, and the mapping component of this study was able to focus on several well-exposed areas at various locations throughout the islands. Data presented here include regional (1:25,000 and 1:50,000 scale) and detailed (1:1,000 scale) maps appearing on seven separate map sheets, as well as several detailed maps incorporated as text figures. Chapter 2 culminates with a structural synthesis which draws on these studies and, to a limited extent, on those of other FGP workers, to develop a new structural model for the Queen Charlotte Islands.

Chapter 3 explores processes of deformation on a more detailed scale: It examines incipient to poorly developed cleavage in some of the low-grade sedimentary rocks described in chapters 1 and 2, using optical and x-ray diffraction techniques, and discusses some of the mechanistic implications of structural styles and fabrics documented in the Queen Charlotte Islands.

Chapter 4 "steps back" and places the observations and conclusions of the previous three sections in a regional tectonic and stratigraphic framework. This brief synthesis combines recently published geological and geophysical data for the Queen Charlotte region and surrounding areas with the present studies in a discussion which traces the evolution of the central Insular Belt from the Late Paleozoic to the present.

Five appendices provide compilations of field and laboratory data, descriptions of analytical procedures, and clarification of potentially ambiguous ideas presented in published reports. Appendix A is a review of progress reports and formal publications on the structural geology of the Queen Charlotte region which have been completed during the FGP study, most of which are cited in this thesis. Due to the time delay inherent in presenting results in outside publications, many of the more recently dated publications cited are superceded by progress reports with publication dates up to two years previous. To prevent possible confusion among workers using these references in compilations, they are reviewed in a chronologic progression which reflects the timing of data collection and interpretation, rather than publication date. Appendix B is a
description of the mesoscopic and microscopic strain analysis techniques applied in chapter 2, and includes a listing of the data generated. Appendix C is a description of sample preparation and analytical techniques involved in the X-ray texture goniometer study of cleavage fabrics. Appendix D includes a short listing of structural data from a brittle fault zone described in the text, and outlines the basis for the kinematic interpretation of these data. Finally, Appendix E is a compilation of the field locations referred to in the text, and is provided as an aid for possible compilation work.

A study outlining the geologic evolution of the Queen Charlotte Islands would be incomplete without elements addressing magmatic, stratigraphic, and isotopic systems in addition to structural history. Fortunately, these aspects were examined in detail by other FGP participants, allowing the present study to concentrate strictly on deformation studies. Nearly all of the FGP Queen Charlotte Basin studies provided valuable constraints to the regional synthesis presented in chapter 4. Naturally, some of the ideas presented here evolved from discussions with other workers, and some have appeared in publications elsewhere. Where multi-authored previously published data or ideas appear in this thesis, the relative contributions of the thesis author and the co-authors are clearly identified.

Parts of this thesis have been published as progress reports (Lewis and Ross, 1988a, 1989, 1990; Lewis 1990, 1991) and as oral presentations at meetings (Lewis and Ross 1988b; Thompson and Lewis, 1989; Lewis et al., 1990). A condensed version of parts of chapter 2 appeared in the Queen Charlotte Basin FGP interim report (Lewis and Ross, 1990) and parts of Chapter 4 are included in a summary paper in Canadian Journal of Earth Sciences (Lewis et al., 1991). Map plate 7 is available as a Geological Survey of Canada Open File publication (Lewis, 1991), and geologic data from the other map sheets were compiled with those of other workers in maps released in the initial FGP report (Lewis and Hickson, 1990; Hickson and Lewis, 1990; Lewis et al. 1990; Thompson and Lewis, 1990a, 1990b).
The data presented in chapters 1 and 2, along with the geologic maps, represent a permanent original contribution to cordilleran geology, and will prove useful to future geological studies of the Queen Charlotte Islands and western Cordillera. Some of the ideas presented in Chapter 3 represent advances in our understanding of process of fabric development under low pressure-temperature conditions, and provide a basis for further studies. Interpretations and regional tectonic hypotheses presented in chapter 4 are by nature transitory and will be revised as more geological and geophysical data become available from the Queen Charlotte Islands and surrounding regions.

Throughout this thesis, orientation data, when specifically given, are presented in azimuth format, with planar elements following right-hand-rule convention. All contoured stereographic projections use the Kamb (1959) contouring method.
Part 1:

Introduction, Tectonic Setting, Previous Geologic Explorations, and Stratigraphy of
the Queen Charlotte Islands Region

The Queen Charlotte Islands lie off the west coast of North America, mostly
between the latitudes of 52° N and 54° N (Fig. 1.1). Although over 200 separate islands
are present, most of the land mass is represented by Graham Island to the north and
Moresby Island to the south. The islands are separated from the British Columbia
mainland by a shallow Neogene depression (Hecate Strait) forming most of the Tertiary
Queen Charlotte Basin (Shouldice, 1971). This depression varies in width from 50 km to
over 100 km. On Graham Island the Queen Charlotte Basin extends onland, and forms
the low-lying areas present on the northeast half of the island. The western margin of
Graham Island and most of the southern islands are occupied by the northwest-trending
Queen Charlotte Ranges, with peaks rising to above 1200 m. On northwestern Graham
Island this range is subdued and reaches only 200–300 m elevation.

Outcrop is poor over most of the islands. The San Christoval Range on Moresby
Island has good exposure in the alpine, but the massive plutonic and volcanic rocks
present here are of limited interest in deformation studies. Many shorelines, with the
exception of the north and east sides of Graham Island, have nearly continuous bedrock
exposures, and more sheltered waters are easily traversed by small boat. An extensive
logging road system on much of Graham Island and northern Moresby Island provides
motor vehicle access to numerous outcrops in road-metal excavations. Timberline in the
islands ranges from 500 to 1000 m, and alpine regions are accessible either by helicopter
or by hiking from sea level. In heavily forested regions, steep creekbeds offer the best
exposure.
**Figure 1.1:** Map showing the major islands and settlements of the Queen Charlotte Islands, the western part of the Queen Charlotte Basin, and Hecate Strait. Inset map shows the position of the islands relative to five tectonic belts of the Canadian Cordillera.
1. Previous Geologic Investigations

The earliest published geologic account of the Queen Charlotte Islands was that of Richardson (1873), whose chief interest was in coal deposits on Graham Island. Dawson (1880) followed with a treatise on the geology, flora, fauna, and ethnology of the islands, which was remarkably complete considering that it was based on only a two and one half month visit. Succeeding work in the early twentieth century includes surveys by Ells (1906), Clapp (1914), and MacKenzie (1916). Little additional work was contributed until the comprehensive study by Sutherland Brown (1968) which included 1:125,000 scale geologic maps for the whole of the islands. Between 1968 and 1987, most geologic investigations in the islands consisted of focussed studies directed toward individual aspects of islands geology. The possibility of petroleum reserves in the Queen Charlotte Basin spurred offshore exploration in the late 1960's, which included detailed seismic coverage and several wildcat wells. Sutherland Brown's (1968) work formed the basis for tectonic models for the Queen Charlotte region proposed by Yorath and Chase (1981) and Yorath and Hyndman (1983). Another study of note during this period was the stratigraphic analysis of Cameron and Tipper (1985), who largely revised the Jurassic biostratigraphy and lithostratigraphy of the islands. Starting in 1987, the Geological Survey of Canada embarked on the multidisciplinary Queen Charlotte Basin Frontier Geoscience Program, of which this study is one element. This program is still in progress, and has already resulted in numerous papers on geological and geophysical aspects of the Queen Charlotte region (see Thompson, 1988b, 1989; Appendix A for overviews). Woodsworth (1991) presents a detailed annotated bibliography of geologic literature on the Queen Charlotte Islands and Queen Charlotte Basin published prior to 1991. Woodsworth and Tercier (1991) trace the often confusing stratigraphic nomenclature for the islands as it evolved over this same time period. Readers interested
in a complete listing of previous geological studies in the area will find these references useful.

2. Tectonic Setting of the Queen Charlotte Islands

The Queen Charlotte Islands region lies within the Insular Belt of the Canadian Cordillera on the westernmost edge of the North American continent. The Insular Belt is generally considered to be an amalgamation of several distinct tectonostratigraphic terranes, each containing a unique structural and stratigraphic record, and separated from adjacent terranes by a major fault or tectonic suture (Monger et al., 1982; Monger and Berg, 1987). In many instances, the exact terrane boundaries and the timing of terrane assembly and accretion to western North America remain controversial. Early Middle Jurassic and older strata on the Queen Charlotte Islands are considered part of the Wrangell terrane, or Wrangellia (Jones et al., 1977). The Wrangell terrane also underlies most of Vancouver Island and parts of southeastern Alaska, and may occur in northeastern Oregon (Monger and Berg, 1987). Adjacent to the Queen Charlotte Islands, the Wrangell terrane presently abuts the Pacific plate along the Queen Charlotte Fault (Fig. 1.2). Global plate-motion studies (Minster and Jordan, 1978; Riddiough and Hyndman, 1989) indicate that the present relative plate motion along the Queen Charlotte Fault at the latitude of the Queen Charlotte Islands is dominantly right-lateral with a small component of convergence. The Wrangell terrane at the latitude of the Queen Charlotte Islands is bounded on the east by the Alexander terrane. Previous models, which placed the suture between these two terranes coincident with the Sandspit Fault on Graham Island and in Hecate Strait (Yorath and Chase, 1981), are discounted by the recognition of Wrangellian strata on the east side of Hecate Strait (Woodsworth, 1988). In southeastern Alaska, the Pennsylvanian Barnard Glacier pluton crosscuts the
Alexander terrane-Wrangellia boundary (Gardner et al., 1988); this recent discovery necessitates re-examination of earlier assertions that the two terranes evolved separately until at least Jurassic time (Jones et al., 1977). van der Heyden (1989) suggests that on the basis of field mapping and geochronometric studies in the Coast Plutonic Complex, both terranes are elements of a single Alexander/Wrangellia/Stikinia megaterrane, which represents a coherent crustal block that evolved through Mesozoic time with no internal sutures. His argument is based largely on the absence of oceanic floor material within the area of the supposed suture zone, and is so far circumstantial.

The stratigraphic succession preserved in the Queen Charlotte Island records deposition in a number of superimposed basins throughout late Paleozoic, Mesozoic and Cenozoic time. Oldest strata present are unnamed Permian strata and unconformably overlying volcanic rocks of the Triassic Karmutsen Formation. These "basement" units are overlain conformably by Upper Triassic and Lower Jurassic sedimentary rocks. The Permian to Lower Jurassic package is lithologically similar to equivalent age rocks on Vancouver Island and other parts of Wrangellia. Succeeding these are Middle Jurassic through Cretaceous sedimentary and volcanic rocks which likely accumulated in more localized sedimentary basins. Superimposed on these Mesozoic strata are Tertiary sedimentary and volcanic components of the Queen Charlotte Basin. Definitions of the Queen Charlotte Basin vary among authors; in this report the term is used to include only the Tertiary sedimentary basin(s), which occur mostly in offshore regions. This choice of a rather restricted definition stems from the recognition that basin-forming mechanisms are different in the Tertiary than in earlier times, and that deposition was continuous through much of the Tertiary but is punctuated by several major unconformities in the Mesozoic. Others define the basin more broadly in time, and include rocks as old as Middle Jurassic in the basin fill sequence (Thompson et al., 1991).
Figure 1.2 is a regional map showing many of the principal tectonic features in the Queen Charlotte region and adjacent areas on the British Columbia mainland. Mesozoic sedimentary rocks on the Queen Charlotte Islands are separated from similar age strata in the Intermontane Belt by crystalline rocks of the Coast Plutonic Complex, hindering stratigraphic correlations between the areas and paleogeographic reconstructions. The western edge of the Coast Plutonic Complex is dissected by a number of northwest-trending, steeply-dipping faults. The most significant faults at this latitude are the Work Channel Lineament, the Grenville Channel Fault, the Kitkatla Fault, and the Principe-Laredo Fault. Offset histories for these faults are still largely conjectural. At least the Work Channel Lineament and the Grenville Channel Fault have some strike-slip motion, perhaps as old as Late Jurassic (van der Heyden, 1989). Woodsworth et al. (1983) suggested that movement direction along the Work Channel Lineament is predominantly steep with a possible right-lateral component.

Prior to 1988, it was generally accepted that large transverse offsets had occurred along several major faults through the Queen Charlotte Islands (Sutherland Brown, 1968; Yorath and Chase, 1981; Yorath and Hyndman, 1983). Recent work has revised the displacement history and tectonic significance of some of these faults (Thompson and Thorkelson, 1989; Lewis and Ross, 1988a, 1988b; Thompson, 1988a). Figure 1.2 shows only features considered to be of regional significance by these more recent studies.
Figure 1.2: Principal tectonic features of the Queen Charlotte Islands region and adjacent British Columbia mainland.
3. Project outline

The original objectives of this study were:

1. to constrain the timing, geometry, and kinematics of deformation in the Queen Charlotte Islands region,

2. to document transitions in deformational style with structural and stratigraphic levels through field studies at several locations in the islands,

3. to evaluate the mechanisms of deformation on megascopic, mesoscopic, and microscopic scales, and to explore the processes by which strain is accommodated in low grade rocks, and

4. to use the structural constraints obtained above to evaluate existing models or to develop new models for the structural evolution of the Queen Charlotte Islands and Queen Charlotte Basin.

As originally formulated, this study was to rely heavily on the geologic maps and interpretations of Sutherland Brown (1968) as a means of identifying structurally complex areas for detailed map studies. Although these maps proved useful in identifying the overall distribution of lithologies and areas of structural complexity, Sutherland Brown's (1968) structural interpretations were not supported by initial studies (Lewis and Ross, 1988a). As a result, regional mapping was incorporated into the project to provide a more comprehensive database on which to base detailed studies. Areas chosen for the map studies were northwest Graham Island, Rennell Sound, Long Inlet/Kagan Bay, Louise Island, and Burnaby Island/Juan Perez Sound, and include a diverse group of structural and stratigraphic levels (Fig. 2.1). In addition, a geologic reconnaissance of outcrops through Skidegate Inlet and Skidegate Channel provides an east-west transect through some of the more complex structures in the islands. Map studies are augmented by regional strain analyses studies and petrographic analyses of structural fabrics along fault zones, which help constrain deformation kinematics.
A second segment of this study explores processes leading to fabric development in low-grade rocks, using samples collected in the mapping studies as examples. These samples preserve fabrics at various stages of development, and are examined using X-ray and optical analytical techniques.
4. Stratigraphy

Stratified rocks of the Queen Charlotte Islands can be simplistically viewed as comprising five distinct tectono-stratigraphic assemblages. Major unconformities bound each assemblage, and each is divided into several mappable formations. Two of the five assemblages contain internal unconformities; these latter unconformities are of less significance than the major unconformities and do not warrant further stratigraphic subdivision. The five part division proposed here is based both on observed lithologic similarities and on inferred similarities in tectonic setting of the component units, a theme discussed in more detail in chapter 4. The assemblages thus defined are: 1) A Permian sequence of marine carbonates and cherts, only known in a few exposures in the islands, 2) Upper Triassic to lower Middle Jurassic succession comprising characteristic "Wrangellian" strata, 3) Middle Jurassic volcanic and volcaniclastic rocks, 4) Cretaceous marine sedimentary strata, and 5) Tertiary volcanic and sedimentary rocks. Plutons of three distinct suites intrude these stratified rocks and are most abundant on the southern islands. Figure 1.3 provides a condensed reference diagram showing map unit names, age ranges, and approximate thicknesses, and figure 1.4 shows the approximate map distribution of the major units. The stratigraphic nomenclature for the Queen Charlotte Islands is constantly being revised as a result of this and other studies; the nomenclature shown on figure 1.3 is used throughout this report (see Woodsworth and Tercier, 1991 for review).

The following stratigraphic descriptions are based on observations and measured sections from the map studies, and where noted, on previously published descriptions. No attempt is made to provide exhaustive descriptions of each unit as it occurs throughout the islands. Instead, the descriptions are limited to general lithologic characteristics, and to those details pertinent to succeeding discussions. References to more complete published descriptions are made where appropriate.
**Introduction**

Stratigraphy

**Figure 1.3:** Stratigraphic chart outlining ages, lithologies, thicknesses, and contact relationships of the major mappable units of the Queen Charlotte Islands region, modified from Lewis and Ross (1991).
**Figure 1.4:** General geologic map of the Queen Charlotte Islands, showing the distribution of the major lithologic units. Compilation based on mapping in the present study, other FGP mapping projects, and Sutherland Brown (1968)
The local distribution of lithologic units within each of the map study areas is outlined in Part 2, and shown on the accompanying field maps.

4.1 Assemblage 1: Permian

Until recently, the oldest known exposed rocks in the Queen Charlotte Islands were Upper Triassic volcanic strata of the Karmutsen Formation. However, samples collected from several localities during the 1990 field season have yielded Permian conodonts, indicating the presence of strata older than the Karmutsen Formation (Hesthammer et al., 1991). Sections yielding these microfauna are lithologically distinct from all younger units of the Queen Charlotte Islands, and occur on northwest Moresby Island and possibly in Juan Perez Sound. Contacts with younger strata are obscured by intrusions at these locations, but an unconformity at the base of the Karmutsen Formation is indicated by a lack of Early and Middle Triassic fossils.

Hesthammer et al. (1991) provide a detailed description of two successive stratigraphic sections exposed on northwest Moresby Island, totalling over 110 m. The lithologies present there are medium-beded limestone, dolomite, and chert which occur in varying proportions. Grey to pale-green chert is dominant low in section, but is subordinate to carbonate at higher levels. In Juan Perez Sound at Hutton Point, lithologically similar carbonates with thin chert layers are interbedded with volcaniclastic sands and gravels. These rocks occur stratigraphically below a several kilometre-thick section of the Karmutsen Formation, but fossil control is lacking.

Pre-Triassic strata of the Queen Charlotte Islands bear lithologic similarities to parts of the Late Paleozoic Buttle Lake Group, which underlies the Karmutsen Formation on Vancouver Island (Massey and Friday, 1989). The Buttle Lake Group is interpreted as part of a volcanic arc assemblage (Massey and Friday, 1989); continuity of this pre-Karmutsen volcanic arc into the Queen Charlotte Islands is likely.
4.2 Assemblage 2: Upper Triassic to lower Middle Jurassic

Unconformably overlying the Permian strata are Late Triassic through lower Middle Jurassic volcanic and sedimentary rocks. The Triassic/Jurassic succession is a continuous sequence comprising the Karmutsen Formation and the Kunga and Maude Groups.

4.2.1 Karmutsen Formation

The Karmutsen Formation is a thick succession of basic volcanic rocks with minor interbedded sedimentary rocks. It forms extensive outcrops on the Queen Charlotte Islands and on Vancouver Island, and is considered equivalent to the Nikolai Greenstone of southeastern Alaska (Monger and Berg, 1987). Exposures of the contact between the unnamed Permian rocks and the Karmutsen Formation are not known in the Queen Charlotte Islands. The top of the Karmutsen Formation is conformably overlain by and locally interfingers with the Sadler Limestone, and thin limestone layers within the uppermost volcanic rocks yield Late Triassic (Carnian) conodont assemblages similar to those in the overlying limestone (Orchard, 1991). No pre-Carnian fauna have been found within the unit on the Queen Charlotte Islands. On Vancouver Island, the Karmutsen Formation conformably overlies sedimentary rocks of Ladinian age, and is there accordingly assigned to the Upper Ladinian and Carnian stages (Muller et al., 1974).

Sutherland Brown (1968) suggests a minimum thickness of 4,600 metres for the Karmutsen Formation on the Queen Charlotte Islands. However, all sections examined in the present study, including Sutherland Brown's (1968) type section are disrupted by brittle and semi-brittle fault zones, and the degree of stratigraphic omission or repetition cannot be evaluated accurately. His thickness estimate should therefore be treated as an
approximate measure of present structural thickness; extensive outcrop belts on Moresby
and adjacent islands imply that several thousand metres of section are likely present.

The dominant lithologies in the Karmutsen Formation are pillowed flows (Fig.
1.5a), volcanic breccia, broken pillow breccia, and massive green volcanic flows.
Primary textures, including flow boundaries, are commonly completely obscured by
tectonic and metamorphic overprints. Stratification is often defined by intervals of
sedimentary rocks, amygdule density and size grading (Fig. 1.5b), and elongate pillows.
In some pillowed intervals, 5–15 cm thick planar quartz + epidote, chlorite bands are
consistently oriented parallel to long and intermediate pillow axes, and provide a useful
mapping aid. The sedimentary components of the Karmutsen Formation include grey
limestone layers 0.5–5 m thick, thin ( < 2 m ) argillaceous horizons, and bedded chert
intervals up to 10 m thick. At Rennell Sound and on Louise and northwest Graham
islands, a single 3–4 m-thick carbonate layer occurs consistently at approximately 50–
100 m below the top of the volcanic succession; this carbonate layer may prove to be a
useful stratigraphic marker in other locations on the islands. In addition, thinly bedded
siltstone and mudstone intervals occur in two locations on Chaat Island.

Massive flows of the Karmutsen Formation are commonly highly amygdaloidal
and feldspar-phyric. The uppermost 3–4 m of section directly underlying the Sadler
Limestone weather to a pale yellow colour and are enriched in visible sulfide minerals
(chalcopyrite, pyrite, arsenopyrite).

Karmutsen Formation volcanic rocks have been interpreted as oceanic basalts
(Sutherland Brown, 1968), as intra-arc or back-arc volcanic rocks derived from rifting
(Muller et al., 1974; Barker et al., 1989; Andrew and Godwin, 1989), and as island-arc
tholeiites (Souther, 1977). Geochemistry, overall stratigraphy, and field relations all
support an arc association.
Figure 1.5a: Pillowed lava flows overlying a massive limestone layer in the Karmutsen Formation, near Vertical Point, Louise Island. Limestone layers are typical in the uppermost part of the Karmutsen Formation, and range in thickness from 0.5–5 m. (Location 88-410)

Figure 1.5b: Layering in volcanic flows of the Karmutsen formation, near Vertical Point, Louise Island. Layering is defined by amygdule concentration and by colour banding. (Location 88-392)
4.2.2 **Kunga Group**

The Upper Triassic and Lower Jurassic Kunga Group comprises the Sadler Limestone, the Peril Formation, and the Sandilands Formation. These three formations were originally defined as the grey limestone, black limestone, and black argillite members of the Kunga Formation by Sutherland Brown and Jeffrey (1960). Cameron and Tipper (1985) have since elevated the Kunga Formation to group status and designated the name Sandilands Formation for the black argillite member. Subsequently, Desrochers and Orchard (1991) assigned the names Peril Formation and Sadler Limestone to the black limestone and grey limestone members.

**Sadler Limestone**

The Sadler Limestone is a thickly-bedded to massive, grey to dark grey limestone which conformably overlies the Karmutsen Formation. Its basal contact is always sharp and commonly has several metres of relief. Complete sections were observed in the present study throughout the Queen Charlotte Islands, and range from 5 m (southern Burnaby Island) to 50 m (Sadler Point type section) in thickness. Thicknesses of 250 m, as suggested by Sutherland Brown (1968), could not be confirmed in the present study. The age of the Sadler Limestone is well constrained as Upper Carnian by macrofossil and conodont collections (Carter et al., 1989).

The Sadler Limestone is a thickly-bedded to massive limestone locally containing bioclastic intervals (Figs. 1.6a, b) Sections in the southern parts of the islands contain abundant chert nodules which parallel stratification. Both fresh surfaces and weathered surfaces are light grey to dark grey, and calcite-filled veins (thickness 1–10 mm) and stylolites are abundant in all outcrops. Desrochers (1988) has defined two petrographic facies within the formation; these were not retained as field mapping units. Desrochers (1988) and Desrochers and Orchard (1991) interpret the Sadler Limestone as a shallow
Figure 1.6a: Typical shoreline exposure of the Sadler Limestone showing thickly bedded to massive character, Skincuttle Inlet (Location 90-433).

Figure 1.6b: Sadler Limestone containing lozenge-shaped chert nodules outlining bedding, Vertical Point, Louise Island (Location 90-686).
marine limestone deposited on a stable carbonate platform, under variable conditions ranging from low energy to high energy.

**Peril Formation**

The Peril Formation is a thinly- to medium-bedded black limestone and calcareous sandstone. Its contact with the underlying Sadler Limestone is conformable and gradational over one to three metres. The total thickness of the formation is difficult to assess due to structural disruption of all exposed sections, but the aggregate thickness of successive microfossil zones from several locations is approximately 250 m (Orchard, 1991). Abundant macrofossils and microfossils constrain the age of the formation to Upper Carnian to Upper Norian (Carter et al., 1989).

Typical lithologies of the Peril Formation include medium-bedded calcareous siltstone to mudstone, calcarenite, silty micritic limestone, and diagnostic *Monotis subcircularis* coquinas. The *Monotis*-bearing beds occur over several tens of metres near the top of the section, and crop out in all areas examined. On northwest Graham Island these strata contain thin (1–5 mm) silty layers consisting of over 80% euhedral plagioclase feldspar grains. *Monotis* beds commonly also contain abundant calcareous or siliceous spherules that probably were originally radiolarians.

The Peril Formation is interpreted as a predominantly deep-water pelagic fallout by Desrochers (1988). However, it abruptly overlies the shallow-water Sadler Limestone, and sandy intervals on Burnaby Island contain cross-stratification and ripples. More likely, the lower part of the Peril Formation was deposited in a gradually deepening basin during a marine transgression, which drowned the earlier carbonate platform. An abundance of euhedral plagioclase is indicative of nearby pyroclastic volcanic activity accompanying or shortly following the marine transgression.
Sandilands Formation

The Sandilands Formation is a laminated to medium-bedded sequence of varicolored siltstone, mudstone, calcareous shale, sandstone, and tuff. It conformably overlies the Peril Formation along a gradational contact. Similar to the underlying Peril Formation, most outcrops are intensely faulted, but a formation thickness of over 300 metres is likely. Extensive fossil collections constrain the age of the Sandilands Formation to Upper Norian to Sinemurian (Tipper, 1989; Cameron and Tipper, 1985).

The most common lithology consists of thinly-bedded to laminated siltstone, calcareous shale, sandstone, and tuff. Discontinuous authigenic pyrite layers occur with these lithologies and in places form nodules up to 1-2 cm thick. The varicoloured nature and differential weathering of these layers give outcrops of this facies a distinctive banded appearance (Fig. 1.7). Sedimentary load-structures, graded bedding, and

Figure 1.7: Typical varicoloured, thinly-bedded siltstone, sandstone, and tuff of the Sandilands Formation, Huxley Island (Location 90-278).
feeding traces are common; current structures are rare. Many of the siltstone and sandstone beds contain abundant subhedral to euhedral plagioclase grains. Calcareous shale layers contain abundant small (0.2–0.5 mm) calcispheres.

Thickly-bedded to massive sandstone intervals occur in several locations. A 50 m-thick interval of thickly-bedded sandstone was treated as a separate map unit at Rennell Sound, and intervals of massive sandstone over 5 m thick are documented at central Graham Island (Indrelid et al., 1991b) and Tasu Sound (Taite, 1990a).

Cameron and Tipper (1985) suggest that the Sandilands Formation is a deep-water deposit, which accumulated in an euxinic back-arc basin. However, abundant feeding traces in many outcrops indicate oxygenated conditions in parts of the basin.

4.2.3 Maude Group

The Maude Group is a sequence of shale, siltstone, calcareous siltstone, and sandstone conformably overlying the Kunga Group. Originally mapped as the Maude Formation by MacKenzie (1916), it has since been elevated to group status by Cameron and Tipper (1985). They recognized five mappable formations within the group; in ascending order these were designated the Ghost Creek Formation, the Rennell Junction Formation, the Fannin Formation, the Whiteaves Formation, and the Phantom Creek Formation. More recently, Tipper et al. (1991) have advocated combining the Rennell Junction and the Fannin formations, and using the name Fannin Formation for both units. In many parts of the Queen Charlotte Islands the Maude Group is thin to nonexistent, and only rarely are all four formations preserved in a conformable sequence.

Ghost Creek Formation

The Ghost Creek Formation consists of shale, siltstone, and subordinate limestone interbeds. Its lower contact with the Sandilands Formation is gradational over several
metres. The main difference between the two formations is the lack of volcanic tuffs and
the greater proportion of fissile shale in the Ghost Creek Formation. The thickness of the
Ghost Creek Formation is well established on central Graham Island, where Cameron
and Tipper (1985) describe measured sections from 45 m to 70 m thick. The age range
of the unit, as determined from macrofossil collections, is Late Sinemurian to Early
Pliensbachian.

The dominant lithology of the Ghost Creek Formation is dark-grey shale and
calcareous shale (Fig 1.8a). Bedding is often indistinct in these shales, owing to little
variation in grain size or composition between beds. Near the top of the formation,
calcareous siltstone beds up to 20 cm thick mark the transition to Fannin Formation
lithologies.

Cameron and Tipper (1985) suggest the Ghost Creek Formation accumulated
under conditions similar to those inferred for the underlying Sandilands Formation, but
with waning volcanism. An increase in abundance of coarser material near the top of the
unit suggests a slight shallowing of the depositional basin.

Fannin Formation

The Fannin Formation is a sequence of interbedded siltstone, sandstone,
calcareous sandstone and siltstone, and shale. The lower contact with the Ghost Creek
Formation grades over several metres; the Fannin Formation has a greater sandstone and
siltstone content than the underlying rocks. Cameron and Tipper (1985) assigned the
formation to Lower Pliensbachian to Lower Toarcian.

The most characteristic lithology of the Fannin Formation is medium-bedded
calcareous siltstone (Fig. 1.8b). Lesser amounts of calcareous sandstone and shale occur
and Cameron and Tipper (1985) describe both tuff and tuffaceous sandstone within the
formation. The maximum thickness of the formation in measured sections on Graham Island is approximately 120 m.

Cameron and Tipper (1985) suggest that the transition to coarser-grained lithologies of the Fannin Formation represents a marine regression beginning in Pliensbachian time.

**Whiteaves Formation**

The Whiteaves Formation is a recessive shale or mudstone unit which is exposed only rarely in the Queen Charlotte Islands. It is lithologically distinct from coarser-grained underlying and overlying formations, and both contacts are sharp. Cameron and Tipper (1985) document a thickness of approximately 100 m for this formation. Jakobs (1989) assigns an age of Middle Toarcian based on ammonoid zonations.

In most locations, the Whiteaves Formation is a fissile, greenish-grey shale. Abundant ammonite fragments occur within the shale, and scattered septarian carbonate nodules are characteristic of sections on Maude and Graham islands.

The finer-grained sediments of the Whiteaves Formation may represent a change to deeper marine conditions in Middle Toarcian time.

**Phantom Creek Formation**

The Phantom Creek Formation consists of fine- to medium-grained sandstones which conformably overlie shales of the Whiteaves Formation. The upper contact of the unit is an unconformity, and the maximum thickness observed in the Queen Charlotte Islands is only 15 m (Jakobs, 1989). The Phantom Creek Formation ranges in age from uppermost Middle Toarcian to Aalenian (Jakobs, 1989).
INTRODUCTION / Stratigraphy

Figure 1.8a: Dark grey shale of the Ghost Creek Formation, Whiteaves Bay, northern Moresby Island.

Figure 1.8b: Medium-bedded calcareous siltstone of the Fannin Formation, central Graham Island.
Several distinct lithologies occur within the Phantom Creek Formation. At Rennell Sound, it is a sandstone with a 3–5 m thick belemnite coquina. Cameron and Tipper (1985) note the abundance of chamosite ooliths in some sections. Jakobs (1989) describes limestone concretions and minor amounts of shale interbedded with the sandstone.

The coarse lithologies of the Phantom Creek Formation likely mark the culmination of a second marine regression during Toarcian and Aalenian time.

4.3 Assemblage 3: Middle Jurassic Volcanic and Sedimentary Rocks

An angular unconformity at the top of the Maude Group coincides with an abrupt change in style of sedimentation. Middle Jurassic rocks of the Yakoun and Moresby groups lying on the unconformity are dominated by volcanic lithologies, and are less widely distributed than older rocks. The unconformity surface cuts into older strata as deeply as the Karmutsen Formation, and varies from a disconformity to a sharp angular truncation. The hiatus between the youngest preserved Maude Group sedimentary rocks and the oldest unconformably overlying strata is only approximately 5 million years, but was accompanied by a major regional deformation event and a marked change in tectonic setting.

4.3.1 Yakoun Group

The Yakoun Group comprises volcanic flows, volcanic breccias, and volcanogenic sedimentary rocks which unconformably overlie all older units. These rocks cover large areas on both Graham and Moresby islands. Mackenzie (1916) originally defined the Yakoun "Formation" as including all Jurassic volcanogenic rocks found on Graham Island. Cameron and Tipper (1985) raised the formation to group status and divided it into the Richardson Bay and Graham Island formations. Both of
these formations contain a wide range of lithotypes; mapping in this study differentiated several mappable rock types but did not retain the two formations.

The thickness of the Yakoun Group is highly variable, and ranges up to at least 700 m. Sutherland Brown (1968), Cameron and Tipper (1985), and Hesthammer (1991b) provide detailed descriptions of Yakoun Group lithologies; this study adds little to these descriptions. Typical lithologies include volcanic sandstone, shale, tuff, volcanic conglomerate, and andesitic volcanic flows (Figs 1.9a-b). Sedimentary components consist almost entirely of reworked volcanic fragments. Rarely they contain lithic fragments of sedimentary and plutonic material. Clastic sedimentary rocks of the Yakoun Group are both marine and subaerial (Cameron and Tipper, 1985) reflecting a varied Middle Jurassic topography and local sub-basin development. Volcanic rocks have abundant plagioclase phenocrysts, and are moderately to strongly altered. Hesthammer (1991b) notes that the relative abundance of volcanic flows increases dramatically in a northeast direction on central Graham Islands, and postulates an igneous source in that direction. Fossil collections from the Yakoun Group are entirely Bajocian in age.

4.3.2 Moresby Group

The Moresby Group, comprising the Robber Point, Newcombe, and Alliford formations, was first defined by Cameron and Tipper (1985). It is found only in the Skidegate Inlet area between Sandspit and Queen Charlotte City; previously Sutherland Brown (1968) mapped these rocks as belonging to the Yakoun "Formation" (Sutherland Brown, 1968). Lithologies of the Moresby Group include conglomerate, sandstone, siltstone, shale (Fig. 1.10). The general sedimentary character of these rocks allows them to be differentiated from underlying volcanic Yakoun Group lithologies, which contain primary volcanic rocks. Cameron and Tipper (1985) record up to a 600 m thickness for
**Figure 1.9a:** Yakoun Group lapilli tuff, Moresby Island near Burnaby Strait (Location 90-374)

**Figure 1.9b:** Yakoun Group quartz-feldspar lithic wacke sandstone, central Graham Island.
these rocks. The Moresby Group is Upper Bathonian and Callovian in age, and the base disconformably overlies youngest Yakoun Group strata (Cameron and Tipper, 1985).

Moresby Group strata were deposited in shallow marine conditions, probably in small sub-basins adjacent to uplifted Yakoun Group volcanic centres. The material forming the sedimentary rocks is likely nearly all recycled Yakoun Group volcanic fragments.

Figure 1.10: Cross-bedded volcanic conglomerate of the Moresby Group, exposed at Kwuna Point, northern Moresby Island.
4.4 Assemblage 4: Cretaceous marine sedimentary rocks

The Cretaceous stratigraphic record in the Queen Charlotte Islands is characterized by nearly continuous marine sedimentation and only limited volcanism. Cretaceous strata define two gross fining-upward sequences which lie unconformably on all older units. Cretaceous stratigraphic nomenclature has undergone several revisions during the course of this study, and Haggart (in preparation) is, at the time of this writing, devising a system which will supercede previously used nomenclature schemes. During much of the field work completed for this study, the nomenclature of Sutherland Brown (1968) was used, along with the age assignments of Haggart (1987). Sutherland Brown (1968) divided the Cretaceous system into two lithologic successions, the Longarm Formation and the Queen Charlotte Group, separated by a regional unconformity. Subsequent field mapping by FGP participants, and biostratigraphic investigations summarized in Haggart (1991), showed these two successions are conformable, and that many of the lithofacies present are diachronous. However, the map units as defined by Sutherland Brown (1968) proved mappable in the areas examined in this study and are retained in the following discussion, pending finalization of the new nomenclature scheme.

4.4.1 Longarm Formation

The Longarm Formation is an Upper Jurassic and Lower Cretaceous sandstone, conglomerate, and shale succession which unconformably overlies all older units. It forms a northwest-trending outcrop belt across much of the Queen Charlotte Islands. Sutherland Brown (1968) defined the Longarm Formation as a dominantly sedimentary succession of Late Valanginian to Barremian age which attains thicknesses to 1200 m. Based on examination of outcrop sections at three locations, Haggart (1989) suggested a
more conservative thickness of approximately 450 m, and extended the age limits to Tithonian to Aptian. Haggart (1991) showed that sedimentation is continuous into the overlying Albian and younger Queen Charlotte Group and that the Longarm Formation thus represents a Lower Cretaceous phase of Queen Charlotte Group deposition.

The stratigraphically lowest exposures of Longarm Formation rocks comprise coarse granular sandstones and pebble to boulder conglomerates (Fig. 1.11a) (Haggart, 1989). Interbedded with and overlying these coarse lower facies in most sections are well indurated, massive green fine- to medium-grained sandstones. In the Rennell Sound and Long Inlet map areas, distinctive black and white granule sandstones to pebble conglomerates form layers up to 0.5 m thick in this part of the section, and are diagnostic of the Longarm Formation (Fig. 1.11b). These coarse clastic rocks are succeeded in most areas by fine- to medium-grained grey sandstone with abundant Inoceramid molds and shell fragments. Stratigraphically highest exposures consist of locally concretionary, dark grey, rusty-weathering shale.

Haggart (1989) interprets the Longarm Formation as a single transgressive sequence deposited in water depths rarely exceeding 200 m.

4.4.2 Queen Charlotte Group

The Queen Charlotte Group comprises mid- to Late Cretaceous sedimentary strata present in many locations throughout the Queen Charlotte Islands, including all of the map areas of this study. The stratigraphic nomenclature of the Queen Charlotte Group has a complicated history, both in its original formulation (Whiteaves, 1883; Dawson, 1880; Clapp, 1914; Mackenzie, 1916) and in its more recent usage (Sutherland Brown, 1968; Haggart, 1987; Cameron and Hamilton, 1988; Haggart, 1991). Much of the recent confusion has involved the lithologic definitions and relative ages of the Haida, Honna, and Skidegate formations, which collectively make up the Queen Charlotte Group.
Figure 1.11a: Longarm Formation sandstone with pebble conglomerate layers, Arichika Island. (Location 90-267)

Figure 1.11b: Black and white granular sandstone, characteristic of the lower part of the Longarm Formation, Long Inlet. (Location 87-163)
Sutherland Brown (1968) suggested these three formations represent a continuous sedimentary sequence, with the Haida Formation at the base and the Skidegate Formation at the top. Haggart (1987) showed that rocks previously assigned to the Skidegate Formation are of equivalent age to the upper part of the Haida Formation, but are lithologically distinct and should therefore be kept as a separate formation. Cameron and Hamilton (1988) suggested including the upper fine-grained portion of the Haida Formation in the Skidegate Formation and restricting the term Haida Formation to the coarser sediments lower in section. This designation has fallen into disuse by field geologists, and the present study uses the stratigraphic nomenclature of Haggart (1987) for the three formations.

Recent work at Long Inlet and north of Kagan Bay (Skidegate Inlet area) has led to the recognition of two new stratigraphic units which conformably overlie the Honna Formation and should therefore be included as part of the Queen Charlotte Group. These new units are volcanic rocks which crop out at Saltspring Bay and Gosset Bay (Haggart et al., 1989) and a Santonian to Campanian shale sequence on Slatechuck Mountain (Haggart and Higgs, 1989).

**Haida Formation**

The Haida Formation is a sandstone/shale succession which conformably overlies the Longarm Formation. Sutherland Brown (1968) recognized two mappable units within the Haida Formation, which he informally designated the lower sandstone member and the upper shale member. Fogarassy (1989) further divided the lower sandstone member into three lithofacies, but assigned the upper shale member as a distinct lithofacies of the Skidegate Formation. In the presently used nomenclature, this totals four recognizable Haida Formation lithofacies. In ascending order, these are: 1) a basal cross-stratified granule conglomerate and pebbly sandstone, 2) a fine- to medium-
grained sandstone, 3) an interbedded sandstone and siltstone, and 4) a silty concretionary shale (Fig. 1.12a, b). Haggart (1991) has found that in at least one location Fogarassy's basal Haida Formation lithofacies contains Hauterivian bivalve fossils, which fit into the age range of the Longarm Formation. This apparent discrepancy suggests that some of the Cretaceous lithofacies are diachronous, and lends support to Haggart's (1989) contention that sedimentation was continuous through Early and mid-Cretaceous time. Fogarassy gives a detailed description of the internal stratigraphy and petrology of each of the Haida Formation lithofacies which will not be expanded on here. The total thickness of the four lithofacies is over 1200 m in the type section at Queen Charlotte City (Sutherland Brown, 1968).

Haggart (1986, 1991) summarizes paleontological control on the age of the Haida Formation. The lower sandstones yield Albian fossil collections, while the overlying shales contain fossils of uppermost Albian, Cenomanian, and Lower Turonian age.

Fogarassy (1989) suggests the Haida Formation rocks record a marine transgression in mid-Cretaceous time, with fluvial indicators in the basal sandstones giving way to finer-grained, marine turbidites higher in section. Haggart (1991) and Gamba et al. (1990) differ slightly from this interpretation, in suggesting that all levels of the Haida Formation were deposited in a transgressive marine shelf setting.

**Skidegate Formation**

The Skidegate Formation consists of thinly interbedded sandstone, siltstone, and mudstone and is laterally continuous and coeval with the Haida Formation upper shale member. It forms extensive large shoreline outcrop extents in Long Inlet and Kagan Bay, and also crops out on Moresby Island near Sewell Inlet (Taite, 1990a) and on central Graham Island. It is not certain if the Skidegate Formation is anywhere laterally equivalent to the Haida sandstone member, although faunal collections from Meyer
Island in Kagan Bay suggest a local age equivalence (Haggart, personal communication, 1989).

The dominant lithology of the Skidegate Formation consists of thin (2–10 cm), laterally continuous, graded fine-grained sandstone to mudstone beds (Fig. 1.12c). These beds commonly have scoured bases and abundant primary sedimentary structures. The beds exhibit well-developed Bouma sequences ranging in completeness from $T_{abcd}$ intervals to less complete $T_{bcde}$ intervals, indicating deposition from turbidity currents. Abundant slump structures occur in zones up to 3 m thick (Fig. 1.12d). These characteristics make the Skidegate Formation lithologically distinct from the upper shale member of the Haida Formation, which lacks the well-defined graded bedding and abundant current structures, and commonly contains abundant concretions. The total thickness of the Skidegate Formation is uncertain, and is likely variable. Several hundred metres of section are present at Sewell Inlet and Long Inlet, although outcrop is discontinuous and these sections may be structurally disrupted. Fossil collections from the Skidegate Formation range from Albian (Meyer Island) to Lower Turonian (Haggart, 1991).

The Skidegate Formation roughly forms an outcrop belt to the west of most exposures of the Haida Formation shale member. Some sections show a vertical interfingering of the two lithotypes, and the contact between them is expected to be gradational and of a complex geometry.

The Skidegate Formation was deposited at moderate depths along submarine levees or in interchannel areas. Based on sedimentological evidence and local interfingering with Honna Formation conglomerate, Gamba et al. (1990) interpreted the Skidegate Formation as levee deposits adjacent to submarine fan channels. Haggart (1991) interpreted the Skidegate Formation more generally as a submarine fan deposit which accumulated in outer shelf or upper slope water depths.
Figure 1.12a: Hummocky cross-stratification in Honna Formation sandstone, Onward Point area, northern Moresby Island.

Figure 1.12b: Silty concretionary shale from the upper part of the Haida Formation, Kagan Bay, southern Graham Island. Concretions at this location are less than 30 cm in the longest dimension, but elsewhere exceed 1 m.
**Figure 1.12c:** Turbidites of the Skidegate formation, Sewell Inlet, Moresby Island. (Location 90-ST1)

**Figure 1.12d:** Slump beds in Skidegate Formation turbidites, Kagan Bay, Graham Island. Slump fold axes trend northwesterly, and folds are slightly overturned to the northeast.
**Honna Formation**

The Honna Formation comprises a thick succession of conglomerate and sandstone. Its basal contact displays varying relationships with underlying units. Most commonly, it disconformably or unconformably overlies older Cretaceous units, or unconformably overlies older pre-Cretaceous rocks. However, in several locations, it interfingers with Skidegate Formation strata (Gamba et al., 1990; Indrelid, 1991). Mapping by Thompson (1990) and Thompson and Lewis (1990b) showed that its greatest accumulation forms a thick sheet-like deposit over higher elevations on northeast Moresby Island.

Fogarassy (1989) recognized two mappable lithofacies within the Honna Formation: a clast supported pebble to cobble conglomerate which occurs both at the top and at the bottom of the formation (Fig. 1.13), and a medial siltstone-sandstone-shale unit. New geologic maps for the type section of the Honna Formation at Lina Narrows (Lewis et al., 1990b) showed that Fogarassy’s middle finer-grained facies is a fault sliver within Haida formation rocks, indicating uncertain lateral continuity for this member. Mapping in the present study documents both a fining and overall thinning of the Honna Formation to the west at Long Inlet. The overall thickness of the formation is highly variable, reflecting both original depositional thicknesses and subsequent erosion. The maximum thickness preserved in any one location is the Deena River section of Yagishita (1985a), where a total of 2,500 m of conglomerate and sandstone were measured. This section contains numerous covered intervals which may conceal faults repeating or omitting parts of the section. A minimum thickness estimate of 300 m represents the elevation difference between the flat-lying base of the formation and the highest exposures on northeast Moresby Island.

The age of the Honna Formation is poorly constrained. Inoceramid collections and a single ammonite from the Skidegate Channel area suggest a Coniacian to Santonian
age (Riccardi, 1981; Haggart, 1986). In several locations, Early Turonian shales directly underlie the basal Honna Formation conglomerates along an apparently conformable contact. Marine shales which overlie the Honna Formation at Slatechuck Mountain have Late Santonian fossils (Haggart and Higgs, 1989) although the possibility of yet younger overlying conglomerates cannot be ruled out. A further age constraint is provided by several locations where the Honna Formation apparently interfingers with the top of the Skidegate Formation (Haggart et al., 1990; Gamba et al., 1990). These lines of evidence bracket the age of the formation to Turonian and Coniacian and possibly younger.

The Honna Formation has been interpreted as a marine deep-water fan-delta deposit (Higgs, 1990), or as channel deposits associated with a submarine fan (Yagishita, 1985a; Fogarassy, 1989). Both Higgs (1990) and Yagishita (1985a) describe paleocurrent indicators which overall suggest southwesterly flow directions, although paleocurrent directions range considerably even between closely spaced outcrops, and certainly vary with stratigraphic level.

**Figure 1.13:** Characteristic Honna Formation conglomerate overlying Skidegate Formation mudstones along erosional scour, Legace island, Skidegate Inlet. (Location 87-431).
Unnamed volcanic rocks

Haggart et al. (1989) and Lewis (1990) described a previously unrecognized Upper Cretaceous volcanic unit which is areally restricted to the Long Inlet area. The base of this volcanic unit locally interfingers with sandstones and conglomerates occurring at the top of the Honna Formation (Fig. 1.14a), and it is therefore included in the Queen Charlotte Group. Sutherland Brown (1968) originally interpreted outcrops of this unit as a feeder dyke system associated with Tertiary Masset Formation volcanic rocks.

The unnamed volcanic rocks comprise intermediate to mafic volcanic debris flows, scoria deposits, and massive flows which together attain a maximum thickness of 700–800 m. Stratigraphically lowest rocks in the section are volcanic debris flows. These flows are poorly sorted and are composed dominantly of feldspar-phyric volcanic clasts. Interclast material includes sand and silt stringers with disrupted laminae, and are interpreted to be derived from the underlying Honna Formation. These entrained sediments and the interfingering contact with underlying sandstones provide evidence that the debris flows were deposited in a subaqueous environment. The debris flows and interbedded sandstones are overlain by a succession of scoria, flow breccia, and massive flows (Fig. 1.14b). Oxidation features and a lack of included sedimentary strata suggest a subaerial environment of deposition for this upper part of the unit.

Unnamed sedimentary rocks

Haggart and Higgs (1989) described a previously unrecognized Upper Cretaceous (Santonian) shale unit on Slatechuck Mountain north of Kagan Bay. This unit is not recognized in any of the areas examined in this study, but presumably the unit at one time may have overlain Honna Formation conglomerates in the Long Inlet area. The total thickness of the shales is not known, and the best exposures have only 30 metres of
Figure 1.14a: Conformable contact between Honna Formation conglomerate (left) and unnamed Cretaceous volcanic debris flows (right), Sandstone Islands, Long Inlet. (Location 87-171)

Figure 1.14b: Flow units in unnamed Cretaceous volcanic rocks, Long Inlet. Red bands are interflow breccias. (Location 87-520)
section (Haggart and Higgs (1989). The base of the shales is nowhere exposed, but may be conformable with the underlying Honna Formation. The dominant lithology present is dark grey concretionary shale with minor sandstone lenses. On the basis of faunal collections, Haggart and Higgs (1989) interpret the sediments as marine deposits laid down at shelf water depths.

The presence of Santonian age shales in the Queen Charlotte Group invites speculation that shale deposition was continuous from Cenomanian to Santonian time, and that the Honna Formation is laterally enclosed by shale. Positive identification of an Upper Turonian and Coniacian age shale sequence would help confirm this hypothesis, but is as yet lacking.

4.5 Assemblage 5: Tertiary volcanic and sedimentary rocks

Previous workers on the Queen Charlotte Islands have assigned all Tertiary strata to either the dominantly volcanic Masset Formation or the sedimentary Skonun Formation. Recognition of a new Paleogene sedimentary unit (White, 1990) and the division of the Masset Formation into an older unnamed sequence and several younger map units (Hickson, 1991) has greatly increased the mapping precision and our geologic understanding of the Tertiary geology of the Queen Charlotte Islands.

4.5.1 Unnamed (Paleogene) sedimentary rocks

A recently discovered Paleogene shale and sandstone sequence crops out at Long Inlet just north of Gosset Bay (White, 1990; Lewis, 1990). Haggart et al. (1990) also recognize this unit along Slatechuck Creek and just south of Yakoun Lake on southern Graham Island. Other possible occurrences of Paleogene sedimentary rocks include strata intersected in the Port Louis well (west coast of Graham Island, Fig. 1.4) and conglomerates on Hippa Island, both of which have yielded Eocene-Early Oligocene
palynomorph assemblages (White, 1990). In the Long Inlet area, the Paleogene strata rest with angular unconformity on Upper Cretaceous sedimentary and volcanic rocks. The dominant lithology here is black shale; coal and black sandstone layers are subordinate (Fig. 1.15). Siltstone layers containing iron carbonate cement form rusty weathering, resistant ribs in many outcrops. Non-marine bivalves are abundant in several locations (Haggart, personal communication, 1989). The total thickness of the entire sedimentary succession is indeterminate due to structural complications and incomplete exposure, but is probably several hundred metres. The Paleogene strata probably accumulated in small non-marine basins which may have been fault-controlled. The Hippa Island and Port Louis occurrences may represent deposition in separate but coeval sub-basins, which may have undiscovered analogues elsewhere on the islands.

Figure 1.15: Unnamed Paleogene black shale, north of Mud Bay, Long Inlet.
4.5.2 *Unnamed volcanic rocks*

Tertiary volcanic rocks of the Queen Charlotte Islands were originally all included in the Masset Formation (MacKenzie, 1916; Sutherland Brown, 1968). Hickson (1991) recognizes two petrologically distinct Tertiary volcanic suites, and limits the term Masset Formation to younger (Oligocene to Miocene) feldspar-phyric to aphyric volcanic rocks found mostly on Graham Island. Older (Paleogene) volcanic rocks on southern Graham Island and to the south are assigned to a separate unnamed suite, characterized by abundant hornblende- and feldspar-phyric lava flows and pyroclastic flows. The Dana Facies of the Masset Formation (Sutherland Brown, 1968) is probably part of the unnamed volcanic suite, and attains a thickness of approximately 1500 m in the southern Queen Charlotte Islands.

*Figure 1.16:* Lahar deposit typical of parts of the unnamed Paleogene volcanic succession, east coast Ramsay Island. (Location 90-1181).
The Paleogene volcanic rocks range in composition from intermediate to felsic, and comprise massive flows, flow breccias, agglomerates, and volcanic debris flows (Fig. 1.16). Many sections contain red oxidized intervals, and interbedded sedimentary strata are common. They contain rare to abundant feldspar and/or hornblende phenocrysts and locally abundant pyrite. No evidence is found for submarine volcanism in any of the observed sections.

4.5.3 Masset Formation

Sutherland Brown's (1968) original maps show the Masset Formation covering large portions of the Queen Charlotte Islands. Using Hickson's (1991) restricted definition of Masset Formation limits most of its known occurrence to parts of Graham Island. The Masset Formation was not observed in any of the map areas of the present study, and the following brief description is derived largely from published work by Hickson (1989, 1991).

The Masset Formation is a tholeiitic to calc-alkaline suite of massive flows and pyroclastic rocks which vary in composition from basalt to rhyolite. In all rock types, feldspar is the principal phenocryst phase and hornblende is absent. Present dip directions and rock distribution suggest flows emanated from three low-profile shield volcanoes on Graham Island. Hickson (1991) documents a total of 4000 m of Masset Formation volcanic rocks on Graham Island, somewhat reduced from Sutherland Brown's (1968) estimate of 5500 m. K-Ar dates range from Late Oligocene to Late Miocene, with the largest number of samples in the Early Miocene.
4.5.4 Skonun Formation

The Skonun Formation is laterally extensive under Hecate Strait, but exposures are limited to a few outcrops on eastern and central Graham Island. The formation was not observed in any of the areas mapped in this study.

Oil company boreholes in Hecate Strait and Queen Charlotte Sound encountered up to 5000 m of sedimentary rocks consisting of sandstone, shale, conglomerate, and coal of shallow marine and non-marine origin (Shouldice, 1971). Microfossil collections from the more southerly Queen Charlotte Sound wells contain Early to mid-Miocene foraminifera. The northern (Hecate Strait) wells yielded more diverse collections, comprising undifferentiated Cenozoic forms from shallow levels and mixed Cenozoic and Late Cretaceous forms from deeper levels (J. White, personal communication, 1989). At least the Cenozoic rocks are assigned to the Skonun Formation. Late Cretaceous foraminifera may indicate a continuation of Queen Charlotte Group strata into Hecate Strait.
4.6 Intrusive Rocks

Intrusive rocks of the Queen Charlotte Islands include three distinctive plutonic suites and abundant dykes and sills related to plutons and volcanic units. Anderson and Greig (1989) assign all large intrusive bodies in the islands to either a Tertiary (Oligocene) plutonic suite (Kano plutonic suite) or to one of two Late Jurassic plutonic suites (Burnaby Island, San Christoval plutonic suites). These suites all represent calc-alkaline, I-type plutonism and roughly define northwest-trending magmatic belts. Large intrusions are volumetrically most significant on the southern portion of the Queen Charlotte Islands.

Anderson and Reichenbach (1989) summarize existing radiometric age data for the three plutonic suites and present some new K-Ar and U-Pb dates. The Kano plutonic suite has a K-Ar age of 43.7 ± 11 Ma (from associated dykes) and zircon U-Pb ages of 39-56 Ma, 32-36 Ma, and 27-28 Ma. The geographic distribution of these U-Pb ages documents time-transgressive, northerly younging of the Tertiary suite. Tertiary plutons are typically small (< 20 km²) bodies composed of quartz monzonite and monzonite.

Jurassic intrusions comprise two suites. The San Christoval plutonic suite (U-Pb = 171-172 Ma) occurs as a linear, northwest-trending plutonic welt on the west coasts of the islands, is composed of homogeneous diorite and quartz diorite, and is characteristically foliated. The Burnaby Island plutonic suite (U-Pb = 158-168 Ma) occurs as a more easterly belt of altered, pervasively veined plutons, and is more compositionally heterogeneous, comprising bodies of diorite, quartz monzodiorite, quartz monzonite, trondjhemite, and leucodiorite (Anderson and Greig, 1989).

Dykes and sills are common throughout the Queen Charlotte Islands. Composition ranges from basaltic to rhyolitic, and textures are generally aphanitic to fine grained, aphyric or plagioclase ± hornblende phyric. Dykes generally form tabular bodies several metres wide, but large dykes tens of metres wide with irregular margins.
are common. Concurrent with this study, an examination of dyke orientation and composition was completed by Souther (1988, 1989) and Souther and Jessop (1991). They defined four major dyke swarms in the Rennell Sound, Louise Island, Selwyn Inlet, and Carpenter Bay regions, and suggested that most dykes are related to Tertiary volcanism. Souther (1988) also noted distinct changes in orientation between the four swarms, which he interpreted to be related to either local changes in stress orientations or post-emplacement rotation.

*Figure 1.17:* Cross-cutting dyke contacts on southern Louise Island. Dykes in this area constitute up to 60% of rock volumes. (Location 88-466).
4.7. Metamorphic Geology

The metamorphic geology of the Queen Charlotte Islands has not been extensively re-evaluated since the general analysis of Sutherland Brown (1968), and was not examined in detail in this study. However, several lines of evidence now exist which help to characterize the general thermal history of the islands. Metamorphic mineral assemblages identified by Sutherland Brown (1968), and as part of the present structural studies help constrain the regional patterns of metamorphism. Organic maturation studies (Vellutini, 1989; Vellutini and Bustin, 1991) provide an integrated time-temperature history for exposed sedimentary rocks throughout the islands. Conodont colour alteration indices (CAI) give an indication of maximum paleotemperatures reached in Triassic Kunga Group strata (Orchard and Forster, 1991).

Highest-grade metamorphic rocks of the Queen Charlotte Islands occur in contact aureoles around known plutons, and consist of strongly hornfelsed sedimentary rocks, and metavolcanic schists of the Karmutsen Formation. Distal to these heat sources, regional metamorphism is much less intense, and most rocks have well-preserved primary textures. Metamorphic mineral assemblages in metavolcanic schists include amphibole (both actinolite and bluish-green to green hornblende), andesine, chlorite, epidote, and clinozoisite. Syn-metamorphic veins in these rocks contain chlorite, epidote, and fibrous to blocky quartz. Highest grade, hornblende-bearing assemblages occur on Chaatl Island (west Skidegate Channel), and coincide with a highly-deformed shear zone and the newly-defined Buck Channel pluton, which are described in Part 2 of this study.

Hornfelsic metasedimentary rocks form contact aureoles around plutons up to several kilometres wide (in map view). It is common for the altered sedimentary rocks to have a strongly bleached appearance, especially in calcareous strata of the Kunga Group. Most hornfelsic rocks are too fine-grained for optical identification of metamorphic
mineral assemblages, however, Sutherland Brown (1968) identified diopside, epidote, actinolite, biotite, chlorite, albite, and sphene as primary metamorphic components of some of the coarser grained hornfelses.

More distal to plutonic heat sources, Karmutsen Formation rocks often lack the strong planar fabrics, and never contain metamorphic amphibole. These less metamorphosed rocks retain many of their primary igneous fabrics, and are altered partially to chlorite and other fine phyllosilicates. Abundant vesicles are filled with radiating albite, chlorite, and cryptocrystalline quartz.

In areas more distal to plutons, mineral assemblages are poorly developed. Prehnite and pumpellyite occur in vesicles in the Karmutsen Formation, and chlorite alteration is common in the adjacent ground mass. Visible metamorphic minerals are absent from sedimentary units, although Fogarassy (1989) identified laumontite and diagenetic chlorite in X-ray diffractograms.

CAI values and levels of organic maturation both show a systematic increase to the south in the islands, where plutons are most abundant, and support the inference based on mineral assemblages that most metamorphism is related to plutonic heating. Although levels of organic maturation do not provide absolute constraints, their values provide some general guidelines useful in interpreting the thermal and burial history of the islands. Most of northern Moresby and Graham islands are within or above the oil window (mature to undermature). Time-temperature modelling of vitrinite reflectance values by Vellutini (1989) provide rough control for the thermal and burial history of Mesozoic strata in these areas. Calculated thicknesses of eroded strata above present exposures, assuming a constant geotherm, are less than two kilometres for all units. Regional geothermal gradients are predicted to have been between 45°C/km and 90°C/km assuming a constant geothermal gradient, or up to 150°C/km for short time intervals associated with volcanism, assuming a constant background gradient of
30°C/km. Mesozoic strata on southern Moresby Island have moderately greater values, and most are below the oil window (overmature). Highest values are co-spatial with known plutonic bodies.

Conodont CAI values, based on samples obtained from Kunga Group outcrops, have a complete range from 1–8 (Orchard and Forster, 1991). Although highest values occur primarily in the southern Moresby Island region and are co-spatial with greatest volumes of intrusive rocks, values below 3.5 were measured throughout the islands, indicating that thermal maturation related to regional (burial) metamorphism was limited. Because CAI values are a function of both the time and temperature history, a direct correlation with maximum paleotemperatures is not possible, but approximate maximum temperatures of 200°C are reasonable for Triassic rocks with CAI values of 3.5 (Orchard and Forster, 1991).

Discussion:

Metamorphic mineral assemblages and levels of organic maturation indicate that regional metamorphism in the Queen Charlotte Islands is generally sub-greenschist grade, and local thermal anomalies are associated with intrusive rocks. Highest grade mineral assemblages in metavolcanic rocks of the Karmutsen Formation, limited to areas proximal to Jurassic intrusive bodies, are typical of albite-epidote hornfels to hornblende hornfels levels of metamorphism (Winkler, 1972), generally associated with localized thermal sources and shallow burial depths. Insufficient analyses of mineral assemblage distribution and metamorphic textures have been completed to characterize specific reactions occurring during metamorphism, and consequently it is difficult to constrain temperature and pressure conditions. Winkler (1972) shows that reactions marking the transition from albite-epidote to hornblende-hornfels levels of metamorphism occur at 500°C to 550°C, and have a small pressure dependence. Reasonable depth estimates are
probably best constrained by the stratigraphic and structural record for the islands. The maximum stratigraphic thickness above the base of the Karmutsen Formation was probably around 6–7 km during Late Jurassic plutonism (Fig. 1.3); during Tertiary plutonism depths are harder to estimate, due to variable amounts of erosion associated with Late Jurassic and Late Cretaceous unconformities. Middle Jurassic structural thickening, discussed in detail in Part 2, may have added considerably to stratigraphic thicknesses.

Sutherland Brown (1968) considered mineral assemblages in metamorphosed Karmutsen Formation strata to indicate Abukuma-type facies series metamorphism. However, the key minerals (cordierite, aluminosilicates) which would serve to distinguish this facies series from typical hornfels assemblages are lacking in both the volcanic rocks and the hornfelsic metasedimentary rocks.

Regional metamorphism away from plutonic heat sources was sub-greenschist grade, and the local occurrence of zeolites suggests zeolite facies metamorphism. Based on the known stratigraphic succession in the islands, even the lowest Tertiary strata were not likely to have been buried to depths greater than ten kilometres, and organic maturation studies suggest maximum depths of a few kilometres and maximum paleotemperatures below 300° for most Mesozoic sedimentary strata. The southward increase in levels of organic maturation is probably related to the greater abundance of plutons in that area, because stratigraphic levels exposed are similar to those in the central and northern islands.
Part 2:

Mesozoic and Cenozoic Structural Geology of the Queen Charlotte Islands

1. Introduction

Prior to the commencement of this study in 1987, structural models for the Queen Charlotte region were based on the assumption that most structures were related to several major fault systems: the Rennell Sound/Louscoone Inlet fault system and the Sandspit Fault which transect the Queen Charlotte Islands in a roughly southeast to northwest direction, and the offshore, plate-bounding Queen Charlotte Fault. Sutherland Brown (1968), who originally mapped the onshore faults, hypothesized that several tens of kilometres of right-lateral offset had occurred along them. Although these estimates were based on poorly constrained geologic features (offset of structural trends, offset of volcanic outcrop centres), they nonetheless have been the basis for tectonic interpretations and have appeared in regional compilation maps ever since. For example, Yorath and Chase (1981) and Yorath and Hyndman (1983) presented tectonic models for the evolution of the Queen Charlotte Basin which rely on tens of kilometres of right-lateral offset along the Rennell Sound/Louscoone Inlet and Sandspit fault systems. They interpreted the southern (Louscoone Inlet) portion of the Rennell Sound/Louscoone Inlet fault system as a separate entity from the northern (Rennell Sound) portion, and interpreted the Louscoone Inlet fault as the southern extension of the Sandspit Fault, offset dextrally along the Rennell Sound fault system. They also postulated that the Alexander terrane/Wrangellia boundary is a tectonic suture coinciding with the Sandspit Fault and the offshore extension of the Rennell Sound fault system. Young (1981) showed that nearly all structures mapped by Sutherland Brown (1968) could be interpreted as geometric elements of a regional wrench fault system, driven by movement
along the North America/Pacific plate boundary (Queen Charlotte Fault). He also presented high-resolution seismic reflection and aeromagnetic data which indicate continuity of the Sandspit Fault into Hecate Strait.

Primary objectives of this study were to examine in detail the structural history of the Rennell Sound/Louscoone Inlet and Sandspit fault systems and surrounding areas, to constrain the timing, style, and processes of deformation associated with them, and to evaluate their role in the tectonic evolution of the Queen Charlotte Islands region. Rather than attempt this through regional mapping of a large area, the project combined regional mapping with detailed structural analyses of areas outlined by Sutherland Brown (1968) as being structurally complex, well exposed, and of potential importance to evaluating the regional structural history. The areas chosen to fulfill these requirements provided exposures of a variety of stratigraphic and structural levels at geographically separated locations. These are, from north to south, northeast Graham Island, Rennell Sound, Long Inlet/Kagan Bay, southeast Louise Island, and Burnaby Island/Juan Perez Sound (Fig. 2.1). The Rennell Sound, Long Inlet/Kagan Bay, and Louise Island map areas are all located in the central Queen Charlotte Islands. The Rennell Sound and Long Inlet/Kagan Bay map areas span the northern (Rennell Sound) portion of Sutherland Brown's Rennell Sound/Louscoone Inlet fault system, and the Louise Island map area is located where the northern and southern (Louscoone Inlet) portions intersect. The northwest Graham Island area was examined to assess regional structural continuity and provide further constraints on the Mesozoic structural history of the northern Queen Charlotte Islands, an important consideration in that much of the Mesozoic geology of central Graham Island is obscured by a thick cover of Tertiary volcanic rocks. Mapping at Burnaby Island/Juan Perez Sound provides structural control for the southern Queen Charlotte Islands, and helps outline the structural history of the Louscoone Inlet fault system in that area. Continuity between these principal map areas was provided by concurrent reconnaissance
Figure 2.1: Major fault systems of the Queen Charlotte Islands, as understood prior to 1987; compiled from Sutherland Brown (1968), Yorath and Chase (1981), and Yorath and Hyndman (1983). Shaded areas outline map areas in this study.
mapping of shoreline outcrops along an east-west transect through Skidegate Inlet and Skidegate Channel, and by 1:50,000 scale regional mapping of parts of the central Queen Charlotte Islands by the Geological Survey of Canada (Thompson, 1990; Thompson and Lewis, 1990a, 1990b; Lewis et al., 1990b). The Skidegate Channel transect crosses both the Rennell Sound fault system and the Sandspit Fault, and the western terminus is adjacent to the offshore, plate-bounding Queen Charlotte Fault. It thus provides a representative cross section through most of the major structural elements of the Queen Charlotte Islands. Finally, detailed mapping of a small fault zone, believed to be a splay of the Sandspit Fault, was completed at Copper Bay, just south of the town of Sandspit. This fault zone is the only known surficial exposure where a splay of the Sandspit Fault cuts Mesozoic strata.

To complement the mapping studies, regional strain analyses of deformed ammonites and belemnites were completed. The samples analyzed were collected mostly from Jurassic strata exposed at the map areas shown in figure 2.1, and represent widely distributed localities throughout Moresby, Graham, and adjacent islands.

This chapter outlines the results of the field studies portion of this thesis. It describes the geology of each of the principal map areas, and where appropriate, invokes more detailed descriptions of mesoscopic and microscopic structural fabrics to aid in analysis of macroscopic structures. The information contained is most easily digested if each section is read with the map plates (in back pocket) available for ready examination of key map relations. Each map area description is followed by a discussion of the structural evolution for that area. The chapter concludes with a structural synthesis, which presents a generalized structural model for the Triassic to recent structural history of the Queen Charlotte Islands. This model presents many new ideas arising from the present map studies, and will be useful for future regional compilation work. In addition to the thesis map studies, it utilizes map constraints from work by Thompson (1990),
Hesthammer et al. (1991a), Indrelid et al. (1991a), Taite (1991b), and Sutherland Brown (1968). Because of the island-wide scope of the proposed model, the interpretations I present for these data represent in many cases new ideas, and may be at odds with previous interpretations of the original authors.
2. Map Descriptions

2.1 Northwest Graham Island

On northwest Graham Island and adjacent offshore islands, wave-cut benches provide continuous exposure of Upper Triassic to Tertiary strata. The adjacent inland areas are densely forested, and lack outcrop even in stream valleys. Strip maps of shoreline areas therefore provide the only well-constrained geological information with which to assess the structural history of the region. 1:50,000 scale strip maps were completed for shoreline outcrop in the Hazardous Cove area on Langara Island, and from Caswell Point to Newcombe Hill on Graham Island (Map Plate I). At Kennecott Point, Sialun Bay, and Fleurieu Point on Graham Island, spectacular structures are exposed along wide wave cut benches of Triassic and Lower Jurassic strata. Due to the complexity of structures present, 1:1,000 scale structural maps were completed for these outcrops (Map Plate II).

Subsequent to these map studies, completed geologic map coverage for shorelines of the northwest corner of Graham Island have become available (Hickson and Lewis, 1990; Lewis and Hickson, 1990).

2.1.1 Southern Langara Island

(Map plate I)

Cretaceous strata of the Honna and Haida formations crop out along shorelines of southern and eastern Langara Island. These rocks were mapped at Hazardous Cove, and along the southeast shoreline of the island (Map Plate I). At Hazardous Cove, stratification dips gently or moderately in southerly directions, and is folded by a single, open, north-northwest-trending anticline/syncline pair. Several 1–3 m thick, felsic to intermediate composition aphanitic dykes cut the sedimentary rocks, and in the northern exposures, the rocks are strongly hornfelsed close to the Langara pluton. The
stratigraphic contact between the Honna and Haida formations provides an easily recognizable marker, and is offset by east- and southeast striking, steeply-dipping faults with tens of metres of strike offset.

Minor structures are best developed in siltstones of the Haida Formation, and include spaced cleavage, abundant calcite veins, and intersection pencil lineations. Spaced cleavage is limited to outcrops along the east side of Hazardous Cove, where cleavage surfaces are oriented approximately perpendicular to bedding and have an average spacing of 3–5 mm. They strike east-west, and the intersection of cleavage and bedding-parallel fissility results in a well-developed pencil lineation. The cleavage refracts sharply across bedding surfaces, and is morphologically similar to cleavages in strata of comparable age at Skidegate Channel, described in detail in section 2.2.4 and in Part 3.

2.1.2 Caswell Point to Newcombe Hill

(Map plate I)

Rocks of the Karmutsen Formation, the Kunga Group, and the Haida Formation are exposed along the shoreline between Caswell Point and Newcombe Hill. The Maude, Yakoun, and Moresby groups, and the Longarm Formation are absent from the area and, in one location, the Haida Formation unconformably overlies the Peril Formation. However, 5 km southwest of Caswell Point at White Point, both the Longarm Formation and the Yakoun Group occur in shoreline outcrop. The dominant macroscopic structures between Newcombe Hill and Caswell Point are steeply-dipping, roughly east-striking faults, and broad, west-northwest-trending folds. Bedding is variably oriented, but generally dips steeply to moderately to the north or south. Exposed macroscopic faults occur as narrow (< 0.5 m) zones of weathered fault breccia, and many coincide with dykes which obscure any offset indicators that might have been present. Most faults
have older strata on northern blocks and tens to hundreds of metres of stratigraphic omission, but absolute movement direction is indeterminate. Rare slickensides indicate that latest movement is sub-horizontal. Detailed mapping of mesoscopic faults at Sialun Bay and Fleurieu Point provided much more conclusive constraints on fault offset, and is described below.

Sutherland Brown's (1968) map shows an inferred south-side-down normal fault striking 060° through the north shore of Sialun Bay, where a pronounced topographic lineament has this orientation (Fig. 2.1). Yorath and Hyndman (1983) named this structure the Beresford Bay Fault (not to be confused with Sutherland Brown's (1968) Beresford splay of the Rennell Sound/Louscoone Inlet fault system), and inferred 10-20 km of sinistral strike-slip displacement along it, coincident with Miocene clockwise rotation of the northern Queen Charlotte Islands. In the present mapping, the fault was recognized only as a 5-10 m wide zone of chaotic brittle fracturing. No kinematic fabrics were preserved, however, the Monotis-bearing interval of the Peril Formation, which rarely exceeds 50 m in thickness (Orchard, 1991), is exposed on both sides of this fault. Dip-slip displacement along the fault is therefore not likely to exceed a few hundred metres, unless multiple periods of movement with opposing senses have cancelled apparent offsets.

Continuous, macroscopic folds are difficult to define due to the abundant faults and fault-related changes in bedding attitude. Stereographic projections of poles to bedding planes in Triassic and Jurassic units (Fig. 2.2a) show a diffuse π-girdle defining an east-northeast trending axis, while similar plots for Cretaceous strata (Fig. 2.2b) indicate no preferred orientation. A broad, northwest-trending anticline through Sialun Bay is defined by both outcrop distribution and average dip directions in older units.

Mesoscopic structures are most highly developed in bedded limestones and siltstones of the Peril Formation, and include shallowly to steeply-dipping minor faults
and open to tight folds. These structures were mapped in detail at locations along the south shore of Sialun Bay and just north of Fleurieu Point, as well as south of the study area at Kennecott Point. These smaller map areas are shown on Map Plate II and are discussed separately below.

\[ \text{Figure 2.2: } \] Equal area stereographic projections (lower hemisphere) of orientation data, northwest Graham Island: a) poles to bedding for Caswell Point-Newcombe Hill shoreline transect, Triassic and Jurassic units; b) poles to bedding for Caswell Point-Newcombe Hill transect, Cretaceous units; c) fold axes and poles to fault planes and cleavage at Fleurieu Point map area.
2.1.3 South Sialun Bay

(Map plate II)

Shoreline outcrops of the Triassic Peril Formation along the south shore of Sialun Bay contain a complex array of folds and faults. 1:1,000 scale mapping of the outcrops helped determine the structural history of the area. In general, bedding in the area strikes northwesterly and dips moderately to the southwest. Faults belong to three sets of different orientations: a set striking 055°–065°, one striking 150°–160°, and one striking 090°–095°. Faults of all sets have subvertical dips. Cross-cutting relationships show that the oldest faults are those striking 055°–065°, and the two younger sets formed synchronously. Slickensides indicate a latest sub-horizontal slip direction on all sets. Fault-drag folds and offset marker beds show that faults striking 055°–065° typically have several metres of sinistral offset, those striking 090°–095° have tens of metres of sinistral offset, and those striking 150°–160° have metres of dextral offset.

Mesoscopic folds are common within fault-bound blocks. These folds have wavelengths of 10–30 m, and open to tight, subangular to subrounded hinges. Minor faulting within fold hinges is common. Axial surfaces are subvertical and strike east-west; fold axes plunge gently to the east or west.

Throughout the south Sialun Bay map area intermediate composition feldspar ± hornblende-phyric dykes intrude the stratified rocks. These intrusions measure metres to tens of metres in width and up to 200 m in length, and are cut and offset by all faults. Hickson (1989) suggests dykes in this region are petrologically distinct from Masset Formation volcanic rocks and associated feeders, and may be of similar age to sedimentary strata of the Queen Charlotte Group (mid- to Late Cretaceous). Alternatively, the dykes may be related to Middle Jurassic volcanism or Late Jurassic plutonism.
2.1.4 Fleurieu Point

(Map plate II)

Rocks of the Peril Formation exposed on a small promontory 600 m north of Fleurieu Point contain imbricate faults with structural styles significantly different from other exposures on northwest Graham Island (Lewis and Ross, 1989). Bedding throughout this area dips gently to moderately to the southwest. The total exposed structural thickness perpendicular to bedding is about 100 m; the total stratigraphic thickness represented is considerably less due to repetition of strata across faults and folds. The dominant structural features present are imbricate faults that are subparallel to bedding, and are continuous for strike lengths of several hundred metres. These are marked by thin, recessive layers of clay fault gouge, which can only be distinguished from stratified clay layers because they locally ramp across bedding. The faults dip gently southwest, and have a regularly spacing of 5–10 m. In one location, recognition of distinct bedding sequences within three successive fault bound packages documents a tripling of structural thickness across faults. Duplex structures commonly occur where faults ramp across layering (Fig. 2.3).

Tight to isoclinal recumbent folds occur in strata immediately adjacent to fault surfaces. These folds commonly contain axial-planar cleavage parallel to regional compositional layering. Fold axes lie within the overall fault plane orientation, and trend approximately east-west. The fold axes are roughly parallel to intersections between fault ramps and flats. This similar orientation, together with the spatial association of folds and fault surfaces (Fig. 2.2c), suggests the folds formed as fault-drag features. The sense of asymmetry of folds, and the geometry of duplexes and fault ramps document an overall east-northeast-directed transport direction along the major bedding-parallel faults.
Figure 2.3: Fault duplex formed in thickly-bedded Peril Formation limestone at Fleurieu Point map area, viewed looking northwest. Fault flat/ramp intersections trend west-northwesterly, and help define northeast-vergent thrusting in the map area. (Location 87-361)

The northwest-striking faults are cut and offset by two sets of steeply-dipping faults, striking 055° – 065° and 115° – 125°. The movement history of these younger faults has been reconstructed from offset planar markers, drag folds, and slickensides. Faults striking 055° – 065° commonly display metres of sinistral offset, while those striking 115° – 125° have metres to tens of metres of dextral offset.

Microscopic Structures

Samples of the axial planar cleavage were collected from several fault-drug folds to more closely examine axial-planar cleavage formation. Thin section examinations show that the cleavages are penetrative surfaces defined by parallel preferred orientation of elongate rock constituents. In micritic limestones, elliptical calcite grains, probably representing calcified radiolaria (E. Carter, personal communication 1989), have long
axes contained in the cleavage plane. Particulate organic material commonly has a highly elongate form, also contained in the cleavage plane. Euhedral detrital plagioclase grains in silty layers have only a weak preferred orientation, and show no evidence of grain breakage or internal strain. Thin silty layers may be completely dissaggregated in fold hinges, and occur as isolated concentrations in the dominant micritic matrix.

Several folds occur in layers containing abundant *Monotis* shell fragments. In these folds, parallel shell fragments define a strong cleavage (Figs. 2.4, 2.5). Wherever compositional layering is at high angles to this cleavage, the shells have been folded or broken and rotated into the cleavage plane. A rough estimate of the amount of shortening across fold hinges within a given volume may be obtained by unfolding individual shells (assuming simple rotation into the present configuration); values obtained by this method indicate up to 70% shortening. Shell fragments are often concentrated in layers which separate horizons of differing composition in the surrounding rock. In one such example, a micritic layer with elongate calcispheres oriented parallel to fold axial planes is juxtaposed across a folded shell against a silty layer containing abundant plagioclase crystals (Fig. 2.5b). Although the silty layer and the micritic layer experienced similar amounts of shortening, no internal penetrative fabrics are present in the silty layer. In places, small dykes filled with feldspar grains and small amounts of matrix were injected into the adjacent micritic layers, and are parallel to the axial planes of the mesoscopic folds. These fabrics all indicate that cleavage formation occurred by a combination of mechanical rotation of shell fragments, dissaggregation of silty layers, and flattening of "soft" rock components, mechanisms consistent with deformation at low temperatures and elevated pore pressures in poorly-lithified sediments (Knipe, 1986b; Groshong, 1988). If deformation had occurred in lithified sediments, brittle structures would be expected in the silty layers, and the
**Figure 2.4:** Axial-planar cleavage in fault-drag folds of the Peril Formation at Fleurieu Point map area. Tight recumbent fold trends west-northwest, axial surface dips southwesterly. (Location 87-409)

**Figure 2.5:** Photomicrographs of samples collected from hinge region of fold in figure 2.4: a) axial planar cleavage defined by: i) parallel preferred orientation of *Monotis* shell fragments, and ii) flattened calcispheres, probably representing calcified radiolaria; b) *Monotis* shell fragments separate layer consisting of micritic organic-rich limestone (i) from silty layer with abundant euhedral plagioclase crystals (ii). Note the lack of penetrative flattening fabrics in silty layer relative to micritic layer. Field of view in both photographs is 2.5 mm, left to right.
rotation of large shell segments into the cleavage plane would have been prohibited by the cohesiveness of the enclosing sediments. Thus, cleavage formation, and by implication, local thrust faulting probably occurred not long after deposition of the Peril Formation, before significant loss of connate water through compaction.

2.1.5 Kennecott Point

At Kennecott Point, highly faulted rocks of the Peril and Sandilands formations are exposed for several hundred metres along wave-cut benches (Map Plate II). The stratigraphic section here spans the Triassic-Jurassic boundary, and consequently has been the focus of intense biostratigraphic analyses, completed concurrently with this study (Desrochers and Orchard, 1991; Orchard, 1991; Tipper et al., 1991). Kennecott Point is located approximately 20 km south of the area mapped in the Newcombe Hill-Caswell Point transect, but the well-displayed structures present, and the utility of biostratigraphic markers for constraining fault offsets make it a valuable contribution to the Graham Island mapping. In addition, a detailed mapping base was provided by a 1:660 scale aerial photograph of the outcrop, prepared as a base for biostratigraphic studies in the area.

Bedding at the Kennecott Point map area strikes easterly to northeasterly, and dips moderately to the south and southeast. Macroscopic structures comprise two nearly orthogonal sets of sub-vertical faults, striking 030°-040° and 130°-140°. The northeast-striking faults are the most obvious set, and they consistently cut and offset northwest-striking faults. Faults of both sets occur as either discrete slip surfaces, or narrow (< 1 m) brittle shear zones with cataclastic features. Northeast-striking faults span the width of exposed outcrop (commonly greater than 200 m), whereas northwest-striking faults extend for only tens of metres. Amounts and sense of offset were determined through examination of calcite slickenfibres on fault surfaces, fault-drag folds, and offsets of
marker beds and biostratigraphic horizons. Both sets of faults record subhorizontal, left-lateral offset. The older, northwest-striking faults commonly have metres to tens of metres of offset. Northeast-striking faults have strike separations greater than their exposed strike length, or at least 200 m. Biostratigraphic studies indicate stratal omissions across faults consistent with offsets only slightly greater than strike-length exposure (H.W. Tipper, pers. comm., 1989).

Major northeast-striking faults divide the outcrop into several northeast-trending structural blocks. Within these blocks, changes in bedding orientation outline broad northwest-trending folds. Tight folds are rare, and when present, are limited to strata immediately adjacent to major faults. Both the open warps and the tight folds are probably related to movement on the bounding faults. In the most elongate fault blocks bedding strikes 050°–080°, slightly more easterly than regional bedding trends. This different bedding orientation may be related to clockwise rotation of the strata within the block during faulting, and is consistent with the left-lateral offsets along the bounding faults (Fig. 2.6).

![Diagram of bedding orientation and faulting](image)

**Figure 2.6:** Schematic model for Kennecott Point map area, showing proposed relationship between bedding in northeast-trending fault blocks and bounding faults. Left-lateral slip along faults occurs in response to north-northwest-directed shortening; bedding attitudes within fault blocks rotate clockwise to accommodate this shortening. Rotation is likely accompanied by slip along bedding surfaces, which in the Sandilands and Peril formations form parting surfaces with little cohesion.
Most fault blocks contain internal spaced fractures with no apparent offset (pervasive joints). The most common joint trends are northwest (140°–150°) and northeast (030°–040°); joints with both orientations are subvertical and spaced 5–30 cm apart.

2.1.6 Discussion

Structures on northwest Graham Island formed during several folding and faulting episodes. The incomplete stratigraphic succession in the area impedes placing absolute timing constraints on individual deformation events, but a relative chronology of structures can be constructed from field relationships. The oldest structures documented in the detailed map studies are north-northeast-verging thrust faults and folds at Fleurieu Point, and strike-slip faults in the other two map areas. Although timing constraints indicating synchronous formation of these structures is lacking, all are consistent with shortening directions in northeast and southwest quadrants. A macroscopic anticline in Triassic strata trending northwesterly through Sialun Bay may be a larger scale manifestation of this northeast-southwest shortening. The south Sialun Bay and Fleurieu Point map areas are both on the southwest flank of this structure, and changes in structural style between the two areas may be related to their relative proximity to the macroscopic fold hinge. Thrust faulting at Fleurieu Point is northeast-directed (toward the hinge area), whereas strike-slip faults at Sialun Bay record no asymmetry, consistent with formation in the fold hinge region. Cleavage fabrics associated with fault-drag folds at Fleurieu Point indicate that earliest deformation took place at low temperatures in poorly consolidated rocks. The thickness of the deformed sedimentary layers here, the lateral continuity of thrust faults and thrust sheets, and the absence of erosional scours, local unconformities, and drapes all rule out a possible slump-fold interpretation for these features. An Early to Middle Jurassic age for deformation is most likely, and is
compatible with timing constraints for earliest deformation elsewhere in the Queen Charlotte Islands (Lewis and Ross, 1991; Thompson et al., 1991).

Younger macroscopic structures on northwest Graham Island include easterly striking steeply-dipping faults with minor offsets, and less commonly, north and northwest-trending broad folds. These features occur in Upper Cretaceous rocks, but no further timing constraints are available. Major strike-slip faults at Sialun Bay, such as proposed by Yorath and Hyndman (1983), are not recognized.
2.2 Central Queen Charlotte Islands

Most of the thesis mapping for this study was concentrated within the central Queen Charlotte Islands, where relatively good exposure combines with easy accessibility and a rich history of biostratigraphic investigations. Studies in the central islands included 1:25,000 scale mapping at Rennell Sound, Long Inlet/Kagan Bay, and southeast Louise Island; a regional transect of shoreline outcrops in Skidegate Channel and Skidegate Inlet, and detailed mapping along a strand of the Sandspit Fault on northwest Moresby Island.

2.2.1 Rennell Sound/Shields Bay
(Map Plate III)

The southern part of Rennell Sound (Shields Bay) provides some of the most continuous exposures of Upper Triassic and Lower Jurassic strata in the central Queen Charlotte Islands. This area occurs along the northwest end of Sutherland Brown's (1968) Rennell Sound/Louscoone Inlet fault system, and thus is an excellent location for examining the structural history of the fault system. 1:25,000 scale mapping was completed for most of the Shields Bay area and surrounding hillsides. Detailed sketch maps of key areas and petrographic analyses of fault fabrics helped to better constrain the timing and kinematics of deformation.

Stratigraphic Distribution

Stratified rocks in the Rennell Sound/Shields Bay map area range in age from Late Triassic (Karmutsen Formation) to Early Cretaceous (Longarm Formation). The Karmutsen Formation and Kunga Group are exposed principally along shorelines at Shields Bay. These strata are lithologically similar to exposures elsewhere in the Queen
Charlotte Islands, except for the Sandilands Formation, which at Rennell Sound contains thick (50–150 m) intervals of thickly-bedded to massive sandstone. Although sandstone intervals are documented in this unit elsewhere in the islands (Taite, 1991a; Tipper et al., 1991), none approaching this thickness is recognized. Sutherland Brown (1968) originally mapped these sandstone intervals as belonging to the Yakoun Group; however, the present, more detailed mapping shows them to be interstratified with characteristic thinly-bedded Sandilands Formation siltstones and argillites, and overlain by rocks of the Maude Group. Three separate sections of the sandstone occur at Shields Bay, but it is unclear whether they represent three different stratigraphic horizons, or are the same, structurally repeated horizon. For them to be separate intervals requires an unreasonably large total thickness for the Sandilands Formation, and they are consequently interpreted as structurally repeated on map plate III. The Maude Group is represented in the map area by exposures of the Phantom Creek Formation along the west shore of Shields Bay and along the road to Rennell Junction (Cameron and Tipper, 1985). Yakoun Group strata occur in only one isolated location at Shields Bay, but crop out extensively along the surrounding hillsides. The youngest rocks in the map area belong to the Longarm Formation, and their occurrence is limited to higher elevations on the the hills south of Shields Bay.

Mesozoic stratified rocks are intruded by the Kano Pluton, which underlies the south end of Shields Bay and has a broad contact aureole. A pluton of the Late Jurassic San Christoval plutonic suite (Anderson and Greig, 1989; Anderson and Reichenbach, 1989) forms the most westerly exposures of the map area.

**Structural geology**

Macroscopic structures in the Rennell Sound/Shields Bay area comprise west and northwest-striking faults and rare northwest trending folds. Cross-cutting
relationships between these features and associated mesoscopic structures support a history of at least three deformation events.

Oldest structures in the map area are recumbent isoclinal folds, recognized in only two outcrops, both of medium-bedded Peril Formation limestone. These folds have northwest-trending axes, minor layer thickening in hinges, and axial surfaces parallel to regional bedding (Fig. 2.7). Because of their limited occurrence, the regional significance and structural thickening represented is probably minor and no large-scale stratigraphic inversion is documented.

Dominant structures in the map area are imbricate, steeply northeast-dipping faults, best exposed along shorelines of islands in Shields Bay. These faults strike from 135° to 160°, roughly parallel to regional bedding, and occur with a spacing of hundreds of metres. Regional fault dips are uncertain due to limited vertical exposure, but both fault surfaces and fault fabrics in adjacent strata are subvertical or steeply northeast dipping at the present erosional surface. Most faults show stratigraphic repetition, exposing older strata on the northeast sides. This fault geometry, northeast-side-up slip direction indicators (described below), and regional northeast bedding dips in adjacent rocks, are consistent with southwest-directed shortening. Stereographic projections of poles to bedding (Fig. 2.8a) which show significant clustering of northeast dips are in accord with this structural vergence. Individual faults are only traceable for, at most, a kilometre before they are cut and offset by younger structures. Faults occur as either discrete slip surfaces, or as brittle fault zones up to 20 m wide. Rocks exposed adjacent to discrete fault surfaces commonly have penetrative tectonic foliations defined by grain-shape fabrics. These foliations are best developed in massive rocks of the Sadler Limestone, and are limited to rocks within 5–15 m of the fault surface. The foliation is approximately parallel to the exposed fault surface, and usually contains a faint downdip mineral elongation lineation. Calcite veins cut the foliation at small to large angles and
Figure 2.7: Recumbent isocline in thickly-bedded Peril Formation limestone at Shields Bay. Features similar to this are limited to the Peril Formation, and represent the oldest structures in the Rennell Sound map area. The fold axis trends northwest, away from the viewer; axial surface is sub-horizontal. (Location 88-122)

are variably folded. Many are isoclinally folded, containing axial surfaces parallel to the foliation and fold axes parallel to the downdip lineation (Fig. 2.9).

Thin section examination of a suite of foliated limestone samples collected at graduated distances of up to 25 m from a major fault on southern Clapp Island help elucidate the deformation history associated with fabric formation. All samples consist of a very fine-grained recrystallized calcite matrix cut by variably deformed calcite veins. Veins occur in numerous orientations with respect to foliation; rare veins perpendicular to foliation are least deformed, sub-planar, and contain original fibrous calcite infillings, while veins sub-parallel to foliation are commonly isoclinally folded, and either recrystallized or filled with blocky, strongly twinned calcite. Several trends in fabric development are apparently related to distance from the major fault surface: Firstly, the
**Figure 2.8:** Equal area stereographic projections (lower hemisphere) of structural elements in the Rennell Sound/Shields Bay map area: a) poles to bedding, Karmutsen Formation and Kunga and Maude groups. Strong clustering in southwest quadrant reflects southwest structural vergence; b) fold axes and axial planes, all units. Axes consistently trend northwest, with steeply-dipping axial surfaces; c) poles to fault planes, all units.
Figure 2.9: Variably folded calcite veins demonstrate high strains contained in foliated limestone wall rocks adjacent to northwest-striking fault. Fold axial surfaces are parallel to northwest-striking, northeast-dipping foliation; fold axis plunges down dip. (Location 88-207)

Volumetric abundance of veins decreases away from the fault surface. More noticeably, the degree of recrystallization of the calcite matrix decreases. Samples closest (less than 5 m) to the fault surface have a matrix consisting of extremely fine-grained (10—15 \( \mu \text{m} \)) equigranular, equant calcite grains. Samples collected at farther distances (up to 25 m) from the fault consist of both slightly coarser (15—30 \( \mu \text{m} \)) matrix and larger (100—300 \( \mu \text{m} \)) elongate grains, with long axes contained in the foliation plane. Sense of offset indicators are rare in all samples owing to the degree of recrystallization, and only samples collected at distances of over 15 m from the fault plane contain usable indicators. Indicators in these sections include calcite porphyroclasts cut by extensional microcracks and synthetic shear bands that fault porphyroclasts (Fig. 2.10). Both types suggest northeast-side-up movement along the steeply-dipping foliation, consistent with reverse movement along the major fault surface.
Subvertical faults striking $80^\circ$ to $90^\circ$ cut and offset the northwest-striking faults in the map area. These faults occur with a spacing of 0.4 to 0.7 km, and most expose slightly deeper structural levels on their north sides. They typically form brittle fault zones up to 20 m wide similar to those found along some of the northwest-striking faults. Within these zones, anastomosing faults juxtapose slivers of well-bedded strata (Peril and Sandilands formations) between blocks of older, more massive volcanic or sedimentary rocks (Karmutsen Formation and Sadler Limestone). These fault slivers of bedded rocks are themselves internally faulted and strongly folded. The folds are disharmonic, tight, and have subhorizontal axes which are parallel to or slightly oblique to fault zone boundaries. One such brittle fault zone, striking approximately $085^\circ$, is shown in figure 2.11. In this location, major structural elements present are northwest-trending, subhorizontal mesoscopic folds, northwest trending minor faults, and minor faults.
parallel to the overall fault zone boundaries. The fold geometry is consistent with formation during left-lateral sub-horizontal motion along the fault zone, but no independent offset indicators for the minor faults are present. The distribution of lithologies and regional dips adjacent to these faults indicates either north-side-up dip-slip, or right-lateral strike-slip movement, suggesting that more than one sense of movement has occurred.

Broad, macroscopic folds are inferred from regional dip direction variations in the Longarm Formation, exposed at higher elevations northwest of Mt. Matlock. These folds trend northwesterly and have rounded hinges and wavelengths of several kilometres. They are not apparent in older rocks along trend at Shields Bay; the imbricate fault nature and steeper dips present in this region might obscure such structures.

**Figure 2.11:** Sketch map of east-west striking brittle fault zone at Rennell Sound. Fold orientations are oblique to fault zone boundaries, and indicate northeast shortening direction, consistent with sinistral, sub-horizontal slip on fault system.
Discussion

Early, recumbent isoclinal folds are rarely observed, are limited to thickly-bedded intervals in the Peril Formation, and only locally invert bedding. Their limited occurrence and the lack of associated structural fabrics suggest they do not represent a regional deformation event, but more likely formed during slumping of cohesive, partially lithified sediments. This movement would have occurred not long after deposition of the sedimentary succession, probably in latest Triassic time. Features considered diagnostic of soft-sediment deformation at the sediment-water interface (bedding truncations, drapes, fluid escape structures) are not observed in outcrops containing these folds, suggesting that they may have formed within the sediment pile.

In contrast, southwest-directed reverse faulting is ubiquitous throughout the map area and represents a regionally significant deformation event. Minimum amounts of shortening during this deformation were determined from palinspastic restoration of fault offsets, using the cross sections on map plate III. These sections are drawn approximately parallel to movement direction, but unfortunately, the subsurface geometry can only be inferred from discontinuous outcrop data. The restorations, shown in figure 2.12; yield shortening values of 26% for the northern section, and 49% for the southern section. This apparent discrepancy between the two sections may be accounted for in several ways: Firstly, the northern section has poor stratigraphic control on its southwestern end, and intraformational faults which accommodate further shortening may occur within the Karmutsen Formation. Secondly, the restored section lengths are highly sensitive to fault geometry and inferred positions of hangingwall and footwall cutoffs, and the interpreted geometry shown in the sections may contain significant errors. Alternatively, amounts of shortening may be greater in the southern part of the map area; steeper average bedding dips here may reflect this concentrated shortening.
The absolute timing of southwest-directed shortening is poorly constrained in the map area. Northwest-striking reverse faults cut rocks as young as early Middle Jurassic (Phantom Creek Formation), but in surrounding areas where younger rocks occur, poor exposure leaves the faulting history uncertain. East of Rennell Sound on southern Graham Island, Hesthammer et al. (1991a) show northeast-dipping, southwest-directed thrust faults cutting Middle Jurassic Yakoun Group and Cretaceous Longarm Formation strata. In the northern part of the Rennell Sound map area, northeast stratal dips in the Yakoun Group are concordant with dips in the underlying Sandilands Formation, suggesting the two share, at least in part, a similar structural history. Best timing constraints are provided by Anderson and Reichenbach (1991), who have identified a ca. 168 ± 4 Ma (U-Pb zircon date) satellitic intrusion of the Burnaby Island plutonic suite which crosscuts south-verging folds of the Sandilands Formation at Shields Bay. If these folds are genetically related to the southwest-directed contractional faulting, the intrusive relationship constrains the timing of initial shortening to Middle or latest Early Jurassic.

Movement indicators on west-striking faults are ambiguous, and may reflect more than one period of displacement. West-striking faults offset and are therefore younger than northwest-striking contractional faults, but their relationship to Cretaceous strata is unclear. On central Graham Island, faults with similar orientations offset Lower Cretaceous stratigraphic belts and have an inferred north-side-down displacement (Hesthammer et al., 1991a). Relative timing of east-striking faults and northwest-trending folds of Cretaceous rocks at Mt. Matlock, in the southeastern part of the study area, is unclear.
Figure 2.12: Palinspastic reconstruction of northwest-trending dip-slip faults at Shields Bay, drawn from east-west cross sections shown on map plate III. Removal of southwest-directed fault offset gives estimates of 26% shortening for section A, 49% for section B.
2.2.2 Long Inlet/Kagan Bay Map Area

(Map plate IV)

The most extensive exposures of Cretaceous strata in the Queen Charlotte Islands occur along shorelines of eastern Skidegate Inlet at Long Inlet and Kagan Bay. This area has been the focus of numerous geologic studies, including the pioneering work of Dawson (1880) and Whiteaves (1883), owing largely to the easy access, well-displayed structure, and good fossil control. Sutherland Brown (1968) mapped three strands of the Rennell Sound/Louscoone Inlet fault system striking northwest through Long Inlet, where steeply dipping, strongly-folded rocks are common. Present map studies concentrated on elucidating structural relationships in this area, and documenting the progression to more gently dipping, less deformed strata to the east.

Stratigraphic Distribution

Most rocks exposed at Long Inlet and Kagan Bay are Cretaceous sedimentary strata of the Longarm, Haida, Skidegate, and Honna formations. Older volcaniclastic sedimentary rocks of the Yakoun Group are limited to the southernmost map area along Skidegate Channel. The Longarm Formation crops out along Skidegate Channel, along the west shore of Long Inlet, and in ravines north of Long Inlet. Lithologies present in these areas are dominantly sandstones rich in inoceramid valves and shell debris, and lesser amounts of conglomerate and shale. Black and white granule pebble conglomerate horizons, characteristic of the lower part of the Longarm Formation (Sutherland Brown, 1968; Haggart, 1989) are interstratified with massive sandstone at Young Point (Long Inlet).

Rocks of the Queen Charlotte Group form a majority of the shoreline outcrops in the northern and eastern parts of the map area. The Skidegate Formation forms extensive outcrop of turbiditic mudstone, shale, siltstone, and fine-grained sandstone along the
north shore of Kagan Bay and along the east shore of Long Inlet. Small isolated outcrops of Cretaceous concretionary shale in western Kagan Bay are more typical of the Haida Formation; rocks in these areas were mapped as included in the Skidegate Formation, due to the difficulty in tracing formational boundaries at the scale of mapping conducted. Conglomerate and sandstone of the Honna Formation overlaps rocks of the Longarm, Haida, and Skidegate formations over much of the map area. Where the lower contact is exposed, it is either conformable or it truncates bedding in underlying units at a slight angle (<10°). Westernmost exposures of the Honna Formation, on the east shore of Long Inlet, are dominantly sandstone with discontinuous conglomerate horizons. The relative proportion of sandstone in the unit systematically decreases to the east through Kagan Bay, and at Lina Island, conglomerate is the major component.

Upper Cretaceous volcanic rocks conformably overlie the Honna Formation at the mouth of Long Inlet. The contact between the units is well exposed on Gust Island, where Honna Formation conglomerates interfinger with volcanic debris flows of the overlying succession over a 20 m interval. These Upper Cretaceous volcanic rocks are unknown outside of the Long Inlet/Kagan Bay map area (see also Haggart et al., 1989).

Tertiary strata in the Long Inlet/Kagan Bay area include a newly-described Paleogene shale unit (Lewis, 1990; White, 1990) which crops out north of Gosset Bay, and gently-dipping volcanic rocks which form the higher peaks south of Long Inlet. These volcanic rocks contain abundant feldspar and hornblende phenocrysts, distinct from typical Masset Formation lithologies (Hickson, 1991), and are accordingly assigned to the unnamed Paleogene volcanic unit.
Structural Geology

Geologists have long recognized the structural complexity of the Long Inlet area. Despite the relatively young age of strata present, map relationships indicate several distinct episodes of deformation in Late Cretaceous and Tertiary time. Much of this deformation is concentrated in a 4 to 5 km-wide zone which trends northwesterly through the central part of the map area. This zone contains both unique structural styles and a unique stratigraphic record, and recent studies within it have been central to many evolving ideas for the structural and stratigraphic history of the Queen Charlotte Islands (Thompson, 1988a; Haggart et al., 1989; Lewis, 1990). Thompson and Lewis (1990a, 1990b) and Thompson (1990) have traced folds and faults which characterize this zone of deformation southeast of the Long Inlet/Kagan Bay area, into Cumshewa Inlet and across Louise Island. In general, these features are co-spatial with the northern (Rennell Sound) portion of Sutherland Brown's (1968) Rennell Sound/Louscoone Inlet fault system. However, evidence for the large-scale structural dislocations inferred by Sutherland Brown (1968) is lacking, which prompted Thompson (1988a) and Lewis and Ross (1988a) to abandon the term "fault system" in their descriptions. Thompson (1988a), after recognizing regionally continuous folds in Cretaceous units, suggested the descriptor "Rennell Sound fold belt" for the zone of deformation. However, these folds do not continue into Rennell Sound, and because much of the deformation in the zone occurs through faulting, this term is misleading. Nonetheless, the feature has considerable significance to the structural and stratigraphic evolution of the Queen Charlotte Islands and should be named for ease of future reference. This study proposes the term "Long Inlet deformation zone" (LIDZ) for the northwest-trending belt of concentrated deformation. This term is appropriate because many of the structural features characteristic of the zone are best displayed at Long Inlet, and because it carries no implications of formative mechanisms. Figure 2.13 shows the approximate
boundaries of the LIDZ on northern Moresby and southern Graham islands, based on the present map studies and maps by Thompson and Lewis (1990a, 1990b) and Thompson (1990). The following structural descriptions are divided between features occurring within the LIDZ, and those in peripheral areas, a division which underscores the importance of the deformation zone to regional structural evolution.

Figure 2.13: Location of the Long Inlet deformation zone in the central Queen Charlotte Islands. For the purpose of comparisons, faults mapped by Sutherland Brown (1968) as components of the Rennell Sound/Louscoone Inlet fault system are shown (most of these faults are not recognized in more recent map studies).
Structures within the LIDZ

The LIDZ is a structurally complex zone which has accommodated anomalously large amounts of strain. Macroscopic structures within it are dominated by northwest-trending folds and faults. Much of the LIDZ forms a structural low, where Cretaceous and Tertiary strata are flanked by belts of older strata outside the high-strain zone. Bedding strikes within the LIDZ parallel the major structures, and dips are moderate to steep to both the northeast and southwest (Fig. 2.14a).

Macroscopic folds are dominated by a northwest-trending anticline/syncline pair parallel to the axis of Long Inlet. These folds extend through the map area across Skidegate Channel, and are inferred from bedding dips to extend to the north shore of Louise Island, over 30 km to the southeast (Thompson and Lewis, 1990a, 1990b). Lewis and Ross (1991) refer to these regional folds as the Long Inlet anticline and the Long Inlet syncline, and they are the features which originally prompted Thompson (1988a) to propose the designation "Rennell Sound fold belt".

At Long Inlet and Skidegate Channel, the folds are easily mapped through changes in dip and facing direction in Cretaceous sedimentary strata, although they are obscured by younger faults at Long Inlet. The anticline trace lies approximately 2 km east of the syncline trace, and the fold pair has an amplitude of 1.5 km. The shared limb is east-facing, steeply east dipping to overturned, and defines a northeast-directed sense of vergence. The anticline has a subangular closure and planar limbs, and the syncline has a subrounded form. The east facing, steeply east-dipping to overturned limb is unconformably overlapped by Paleogene sedimentary strata and by Paleogene to Neogene volcanic rocks both north and south of Long Inlet. Stereographic projections of poles to bedding in the LIDZ (Fig. 2.14a) clearly show the steeply-dipping, northeast vergent structural style which characterizes the area.
Figure 2.14: Equal area stereographic projections (lower hemisphere) of structural features of the Long Inlet map area: a) poles to bedding, all units within the LIDZ; b) poles to bedding, all units outside the LIDZ; c) poles to cleavage planes, most from within the LIDZ; d) fold elements and intersection pencil lineations, LIDZ.
Although Paleogene strata unconformably overlie the steeply dipping, east-facing limb of the Long Inlet anticline/syncline, they are themselves disrupted by northwest-trending folds and contractional faults (Lewis, 1990). The Paleogene strata are juxtaposed against Cretaceous volcanic rocks to the west along a northwest-striking fault through Gosset Bay (the Gosset Bay fault, Lewis, 1990). Steepest stratal dips in the Paleogene unit occur adjacent to this fault. Just north of the Long Inlet/Kagan Bay map area on Mt. Matlock, Haggart et al. (1990) mapped unfolded Miocene sedimentary and volcanic strata unconformably overlying these deformed Paleogene strata.

Major faults within the LIDZ strike northwesterly, parallel to macroscopic fold traces, and dip moderately to steeply to the northeast or southwest. These faults cut the limbs of the Long Inlet anticline and syncline, and some were mapped by Sutherland Brown (1968) as components of the Rennell Sound/Louscoone Inlet fault system. They are rarely directly observed in outcrop, and structural fabrics in adjacent rocks could not be conclusively tied to fault movement. However, several lines of evidence suggest the faults have dominantly dip-slip offset: Firstly, faults of similar orientation at the Rennell Sound and Louise Island map areas (in older rocks) preserve kinematic fabrics indicating dip-slip offsets. More compelling, the faults die out within 10 km along strike, as demonstrated by both the present study and by regional mapping of Thompson and Lewis (1990a, 1990b). Dip-slip restoration of offset contacts would require only hundreds to at most a thousand metres of offset, while strike-slip restoration would require several kilometres of offset, an unlikely scenario given the short strike lengths of the faults. Similarly oriented faults on central Graham Island (Hesthammer et al., 1991a) and on northeast Moresby Island (Thompson, 1990) are interpreted as dip-slip faults. Fault dips cannot be determined at present levels of exposure, but most appear to be steep.
Some of the northwest-striking faults in the LIDZ are unconformably overlapped by the Paleogene sedimentary succession just east of Mt. Seymour (Lewis, 1990). However one fault, the Gosset Bay fault, cuts these same sedimentary rocks, juxtaposing them against Cretaceous volcanic rocks to the west. The Paleogene strata are themselves cut by two northwest-striking faults which dip moderately to steeply westward. One such fault is located parallel to and approximately 400 m east of the Gosset Bay fault. Strata between the two faults dip steeply to vertically. East of this section, bedding dips outline a northwest-trending antiform above the second, west-dipping fault; this fold may be related to thrust movement along the underlying fault.

Minor east-northeast (055°–070°) striking faults cut and offset northwest-trending folds and faults, and extend outside of the LIDZ. Slip direction indicators are lacking, but simplest reconstructions indicate a few hundred metres of left-lateral offset. These faults are similar in orientation and sense of offset to faults at Rennell Sound, and have an orientation similar to dip-slip faults described by Hesthammer et al. (1991a) on central Graham Island. Higgs (1990) proposed, on the basis of his interpretation of paleocurrent data, that the most northerly of these faults bounds two tectonic blocks which have undergone differential rotations of up to 90° in the Tertiary. This interpretation is not supported by mapping in this study, which shows that Late Cretaceous/Early Tertiary structures have similar orientations on both sides of this fault (Lewis et al., 1991b), and alternative interpretations of the paleocurrent data are suggested below, and by Lewis et al. (1991b).

Mesoscopic structures in the LIDZ include open to tight folds in the Skidegate Formation, penetrative cleavage in these same rocks and in the Paleogene sedimentary strata, and minor faults and joints in all map units. Cleavage and mesoscopic folds are best developed in the core of the Long Inlet anticline near Gosset Bay and in Skidegate Channel (Fig 2.15). These folds have upright axial surfaces and subhorizontal to
moderately plunging, northwesterly-trending axes, parallel to macroscopic folds. Penetrative axial planar cleavage trends northwesterly and dips steeply to the southwest, parallel to axial surfaces of the Long Inlet anticline/syncline pair. Pencil lineations, formed by the intersection of bedding-plane fissility and cleavage, are parallel to regional and mesoscopic fold axes (Fig. 2.16).

Minor faults in the folded rocks strike northwesterly, parallel to major fold axial traces, and cut bedding at shallow angles. Fault-drag folds in adjacent strata indicate reverse or thrust offset along them.

Mesoscopic structures in the coarse clastic and volcanic lithologies (Honna Formation, unnamed Cretaceous and Tertiary volcanic successions) are limited to fractures with minor offset (generally < 1.0 m) and ubiquitous calcite- and quartz-filled veins. These planar features occur in numerous orientations and cannot be conclusively tied to specific macroscopic structures.

The Paleogene shales north of Gossett Bay contain numerous minor faults, and a strong cleavage which is commonly sub-parallel to bedding. This cleavage trends northwesterly and dips steeply to the southwest. Minor north-striking faults (offset < 10 cm) dip steeply westward, contain subhorizontal slickensides, and offset strata dextrally. A subordinate set of variably dipping, north to northwest-striking minor faults with slickensides indicating normal offset occurs in many outcrops. Relative timing of the two sets of minor faults could not be determined.
**Figure 2.15:** Limb area of a mesoscopic tight fold with axial-planar cleavage in the Skidegate Formation at Long Inlet. Folds trend northwesterly, parallel to boundaries of the LIDZ, and have sub-vertical axial surfaces. Within the Long Inlet/Kagan Bay map area, this intensity of deformation in Cretaceous rocks is only found in exposures in the LIDZ. (Location 87-151)

**Figure 2.16:** Pencil intersection lineation in the Skidegate Formation, defined by intersection of spaced cleavage and bedding-plane fissility. Pencil lineations are most intensely developed within mudstones in the LIDZ, where they parallel regional and mesoscopic fold trends. (Location 87-502)
Structures outside the LIDZ

Most of the Long Inlet/Kagan Bay map area outside the LIDZ contains outcrops of weakly to moderately deformed Cretaceous rocks. Major structures in these units consist of broad, open, northwest-trending folds and rare northwest-striking faults, and strata in the area are generally flat lying (Fig. 2.14b). Macroscopic folds, best displayed along the north shore of Kagan Bay, have steeply-dipping axial surfaces and wavelengths of 2-3 km. Fold hinges are typically rounded, limbs are curviplanar, and fold axes are sub-horizontal.

Macroscopic northwest striking faults are rare outside the LIDZ, except in the northeastern region near Lina Island. Faults here lack unequivocal slip indicators, but their similarity to structures elsewhere in the map area and the distribution of adjacent strata suggest they are east-side-down normal faults. East-northeast striking faults, described above as sinistral strike-slip features within the LIDZ, extend across deformation zone boundaries and are mapped near Lina Island.

Fine-grained rocks of the Skidegate and Haida formations contain penetrative cleavages similar to those within the LIDZ, but these fabrics are generally less strongly developed and have a less regular orientation than in the highly deformed rocks. Mesoscopic folds in these units are absent outside the LIDZ.

Discussion

Map studies in the Long Inlet/Kagan Bay area support a structural history involving alternating periods of shortening and extension, with much of the deformation concentrated in the LIDZ. This deformation began in the Late Cretaceous or Early Tertiary, with northeast-directed folding of Cretaceous units. Structural relationships at Cumshewa Inlet, outside the map area, suggest that the geometry and strain localization associated with Late Cretaceous/Early Tertiary folding may have been controlled by a
pre-existing regional block fault, the Dawson Cove fault (Thompson et al., 1991). The Dawson Cove fault is not exposed directly at Long Inlet, but a fault juxtaposing thick successions of Longarm and Skidegate formation strata across the inlet is along trend with the Dawson Cove fault on northern Moresby Island (Thompson and Lewis, 1990a), and is inferred to be its northern extension. Late Jurassic movement along the Dawson Cove fault is believed to have uplifted a tectonic block northeast of the fault, resulting in the erosion of Middle Jurassic strata from this area (Lewis et al., 1991a; Thompson et al., 1991). Cretaceous sedimentation blanketed both sides of the fault, forming the thick successions which cover most of the Long Inlet/Kagan Bay map area. Latest Cretaceous to Early Tertiary shortening, resulting in the northwest-trending folds common in the map area, may also have re-activated the Dawson Cove fault, leading to more intense deformation in overlying strata. The northeasterly fold vergence in the LIDZ is consistent with southwest-side-up movement on the fault, which fits reverse-slip reactivation of a feature which formed initially as a southwest-side-down normal fault. Unfortunately, the regional dip direction of the Dawson Cove fault, a crucial test to this hypothesis, cannot be conclusively determined from the surficial mapping.

The coincidence of strongly-developed cleavage and mesoscopic folds with the tightest macroscopic folds suggests that structures on both scales formed during the same shortening event. Minor contractional faults in the core of the Long Inlet anticline at Skidegate Channel probably formed during hinge zone tightening in the Long Inlet anticline. Mesoscopic folds similar to those in the Long Inlet/Kagan Bay map area occur southeast along structural trend at Cumshewa Inlet, also within the LIDZ. These localities are noteworthy, in that mesoscopic tight folds of the Skidegate Formation are rare elsewhere in the central Queen Charlotte Islands.

Northwest-striking dip-slip faults in the LIDZ formed during a Tertiary episode of northeast-southwest extension, which post-dated the above shortening. The
localization of most of these faults in the LIDZ suggests that again, reactivation of older structures controlled the positions and styles of younger structures. Sedimentary overlap of these faults by Paleogene strata north of Gosset Bay constrains the timing of extensional faulting to pre-Oligocene (Lewis, 1990). However, the formation of the Paleogene nonmarine basin at Long Inlet may be tied to an earlier pulse of this extensional faulting. All known occurrences of Paleogene sedimentary rocks in the Queen Charlotte Islands lie either within the LIDZ, or along trend with the LIDZ to the north—perhaps the most compelling evidence for structural control on sedimentation found in the islands.

Minor folds and reverse faults in the Paleogene sedimentary strata indicate a second episode of localized shortening. Faults active at this time include the Gosset Bay fault, and minor faults in the Paleogene section. Because shortening is approximately coaxial to the earlier shortening episode, structures in Cretaceous strata related to the two events cannot be differentiated.

Many of the complex temporal and geometric relationships between structural and stratigraphic elements in the LIDZ can be elucidated by sequentially restoring fault offsets in a cross-section drawn parallel to inferred movement direction (Figs. 2.17a–d). Figure 2.17a shows the present fault and fold geometry in an northeast-southwest section through the LIDZ (section line shown on map plate IV). In this section northwest-striking faults are shown as normal faults with dips of 60° to 80°, except for the Gosset Bay fault which is constrained by field relationships to be a west-dipping reverse fault. Although this geometry is not well constrained by field mapping, it is supported by several independent lines of evidence: Firstly, steeply dipping normal faults are the dominant structures in Cretaceous strata elsewhere in the islands (Thompson and Thorkelson, 1989) and minor extensional faults are abundant in outcrop. Secondly, a geometry involving vertical or reverse faults leads to geometrically unreasonable
configurations when fault offsets are palinspastically restored, while normal faults satisfy geometric considerations.

Figures 2.17b-d are interpretive cross sections showing successive palinspastic restorations of the original section (Fig. 2.17a), which serve to illustrate the geometric evolution of the LIDZ as well as reveal some noteworthy structural/stratigraphic relationships. Figure 2.17b shows the LIDZ with latest reverse offset on the Gosset Bay fault restored. Although the total shortening across the length of the section measures less than 5%, local deformation effects can be intense. The following figures illustrate the structural geometry following restoration of offsets along east-dipping normal faults (Fig. 2.17c) and west-dipping normal faults (Fig. 2.17d). West dipping normal faults include earliest movement on the Gosset Bay fault, and similarly oriented faults to the east. Figure 2.17c shows the Gosset Bay fault rooting into the Dawson Cove fault at depth, and infers that the two faults were linked during west-side-down extension. The combined extension amounts for both east and west dipping normal faults approach 40%, but are highly sensitive to poorly constrained fault dips.

Figure 2.17d shows the configuration of the Long Inlet anticline and Long Inlet syncline immediately following folding and inferred reverse sense re-activation of the Dawson Cove fault. Restored fold geometry suggests that early folding accounts for approximately 23% shortening, which roughly balances later extension.

The palinspastic restorations shown in figure 2.17 are not uniquely constrained by field data, but they do represent the most reasonable, geometrically feasible reconstruction for a structurally complex area. Reconstructions attempted assuming other starting configurations invariably led to prohibitive space problems, and were discarded.
Figure 2.17a: Northeast-southwest cross section through the Long Inlet deformation zone at Long Inlet. (section line shown on Map Plate IV)

NW-side-up reverse fault removed

Figure 2.17b: Figure 2.17a redrawn with reverse-sense offset removed along Gosset Bay fault. Bed-length changes indicate that reverse faulting accommodated approximately 5%, northeast-southwest shortening. Dotted line shows restored position of present topographic profile.
NE-side-down extension removed (Eocene)

Figure 2.17c: Figure 2.17b redrawn with offset removed along east-dipping normal fault, representing probably Paleocene to Eocene configuration.

SW-side-down extension removed (Late Cretaceous)

Figure 2.17d: Cross section in Figure 2.17c redrawn with west-dipping normal fault offsets restored. Both episodes of extensional faulting combine to accommodate roughly 40% extension. Restored section shows greatest concentration of Cretaceous volcanic rocks coincident with the Dawson Cove fault.
The inferred structural history for the Long Inlet/Kagan Bay area, together with compositional and distribution trends in several stratigraphic units, suggest that the Cretaceous and Tertiary stratigraphic evolution may have been locally influenced by structural events in the LIDZ. Structural control on original distribution trends is suggested for three units: the Honna Formation, the unnamed Cretaceous volcanic unit, and the unnamed Paleogene shale unit. As noted above, the Honna Formation undergoes a westward stratigraphic thinning and fining approaching the LIDZ, and it is conformably succeeded within the LIDZ by shallow marine volcanic rocks, which grade upward into nonmarine volcanic strata. Paleocurrent indicators in the Upper Cretaceous Honna Formation indicate a dominant sediment transport direction from the east (Gamba et al., 1990; Higgs, 1990; Yagishita, 1985a). Indicators north of Lina Narrows and in eastern Kagan Bay display a swing to more northerly current directions. Although these paleocurrent indicators likely represent flow directions at different stratigraphic levels, and therefore, different ages of deposition, this northward swing may be related to a transition to axial flow directions in regions more distal from the source. The abrupt transition to nonmarine volcanic rocks overlying the most westerly exposures of the Honna Formation suggest that these distal areas were also located in shallow water depths, possibly near a western basin edge. This basin edge coincides with the trace of the Dawson Cove fault, suggesting structural control. However, the lack of coarse, westerly derived material in the Honna Formation in this area indicates that regions west of the Dawson Cove fault were not sufficiently elevated to serve as a sediment source region during Late Cretaceous sedimentation.

Thickest accumulations of Cretaceous volcanic rocks, when observed in the restored cross sections (Fig. 2.17d) also coincide with the trace of the Dawson Cove fault. This suggests that Cretaceous volcanic rocks were extruded at or near sea level along the margin of a Late Cretaceous basin, with the Dawson Cove fault serving as a
conduit. The volcanic rocks were deposited as debris flows, viscous lava flows, and scoria, resulting in a high-relief volcanic accumulation which thinned rapidly into the basin. This volcanic buildup may also have affected distribution patterns in the uppermost Honna conglomerates, and may be responsible for some of the changes in paleocurrent direction. The original extent of the volcanic rocks may not have encompassed a much larger area than is presently exposed. These interpretations differ slightly from the proposed Cretaceous paleogeography of Haggart (1991), who hypothesizes that a wide, westerly-deepening Cretaceous basin persisted across the central Queen Charlotte Islands and gradually transgressed eastwards through the islands. The uplifted regions west of the Dawson Cove fault may represent a local emergent area in this larger-scale basin, and may have been a divide between local sub-basins, one of which is preserved in the Kagan Bay region.

The limited extent of Paleogene strata in the LIDZ suggests that the region had a complex topography in the early Tertiary, with localized nonmarine deposition, perhaps in fault-bound basins. Palinspastic reconstructions (Fig. 2.17b) suggest that these basins, at least at Long Inlet, formed along east-side-down normal faults. However, the strata are insufficiently preserved to assess expected facies changes approaching the basin edge.
2.2.3 Southeastern Louise Island Map Area

(Map Plate V)

Shorelines of southeastern Louise Island provide nearly continuous exposure of deformed Triassic and Jurassic strata of the Karmutsen Formation and the Kunga Group. Sutherland Brown's (1968) map shows the northern (Rennell Sound) and southern (Louscoone Inlet) portions of the Rennell Sound/Louscoone Inlet fault zone intersecting on Louise Island. His structural interpretations imply that strike-slip movement is transferred from the west-northwest-striking northern portion to the north-northwest-striking southern portion of the fault system. This transfer of motion should generate secondary structures related to the bend in the fault system, and these structures should be easily recognizable in the field. Specifically, extensional structures should form on the outside (northeast side) of the fault bend, and shortening should occur on the inside. Yorath and Chase (1981) suggest as an alternative that the southern (Louscoone Inlet) portion of the fault system was once continuous with the Sandspit Fault, and has since been offset from it by dextral movement along the northern (Rennell Sound) portion. They propose that the northern fault system extends through southeast Louise Island area into the offshore, where it forms the Alexander terrane/Wrangellia boundary.

Present field studies on southeast Louise Island were directed toward determining whether either of these structural models could be supported by field data, or whether the structural evolution of the area was more closely tied to other parts of the Rennell Sound "fault system."

Stratigraphic distribution

The oldest strata exposed in the map area are pillowed and massive volcanic flows of the Karmutsen Formation, which are most extensively exposed in the vicinity of Vertical and Dass points. Stratification within the flows is defined by limestone layers
up to 3 m thick, by layered concentrations of amygdules, and by weathering contrasts between successive flow units. Kunga Group rocks cover most of the remaining map area, except near Skedans Bay where the Yakoun and Maude groups and the Longarm Formation crop out. Intermediate composition dykes of the Selwyn Inlet dyke swarm (Souther, 1989) are abundant in shoreline exposures between Vertical Point and Dass Point, and locally constitute up to 80% of the rock volume. Souther and Jessop (1991) describe the orientation, geochemistry, and probable age of these dykes in detail, and their effort is not duplicated in the present study. Both Souther and Jessop (1991) and Thompson and Lewis (1990a) find that dykes in the Louise Island region are related to Tertiary volcanism.

**Structural Geology**

Dominant structures in the southeast Louise Island map area are macroscopic, steeply-dipping, west- to northwest-striking faults. These faults are best displayed in Triassic and Lower Jurassic units, and are similar in geometry and style to those mapped at Rennell Sound/Shields Bay. Less common macroscopic features are northwest-trending folds. In contrast to the Rennell Sound/Shields Bay map area, distinct fault sets with consistent cross-cutting relationships cannot be delineated. Instead, a complete range of fault strike orientations between 090° and 140° is present (Figure 2.18b). Most faults are subvertical or steeply northeast dipping, and have older strata on their northeast sides. This fault geometry, when combined with northeast bedding dips prevalent in much of the map area (Fig. 2.18a), is consistent with moderate south-southwest-directed shortening and structural thickening, as displayed in cross-section in Map Plate V.
Figure 2.18: Equal area stereographic projections (lower hemisphere) of structural elements at Louise Island field area: a) poles to bedding define northeast-striking π-plane, consistent with southwest-northeast shortening. Strong pole clustering in southwest quadrant corresponds with southwest-directed structural asymmetry determined from map studies. b) Poles to mesoscopic fault planes and fault-plane foliation. Both features have widely strike distribution in northwest quadrant, and steep dips. c) Mesoscopic fold axes and poles to fold axial surfaces.
Macroscopic faults form either discrete slip surfaces or brittle fault zones 10–30 m wide. In many locations where faults cut the Sadler Limestone, the limestone wall rock has a well-developed foliation parallel to the fault surface and a strong downdip mineral lineation. This foliation persists for up to 15 m from the main fault surfaces, and it is most common along west-northwest-striking, steeply-dipping faults. It forms a strong parting surface in outcrop, which is heightened by preferential weathering and fracture.

Thin section analyses of fault fabrics help constrain the strain field associated with foliation development, and by inference, the approximate stress-field orientation and the kinematics of faulting. Samples of the foliated rocks are composed almost exclusively of coarsely crystalline to very fine-grained calcite. Most calcite grains are strongly twinned, and the greatest twin densities occur in the coarser grains. Microstylolites are abundant and are oriented at low to moderate angles to the foliation. Most contain black opaque films (organic material?) and some, rare fine quartz grains. In the most highly foliated samples, calcite grains are highly elongate (up to 8:5:1) parallel to the down-dip mesoscopic lineation. In one sample, deformed fecal pellets allow strains to be more accurately quantified (Fig. 2.19). Pelloid axial ratios of $R_{xz} = 6.1:1$ and $R_{xy} = 3.3:1$ were obtained by centre-to-centre analysis of sections cut perpendicular to foliation (see Appendix 2). Although the pellets may have originally been non-spherical, similar elliptical shapes and orientations for all fecal pellets implies an originally nearly spherical shape, and the axial ratios can at least be used qualitatively. In the down-dip (XZ) sections, a weak secondary fabric defined by discrete shear bands cuts and offsets the elongation fabric at small angles ($10^\circ–20^\circ$). The sense of offset along the shear bands is consistently northeast-side-up.
Figure 2.19: Thin section photomicrograph of foliated Sadler Limestone, showing ellipsoidal deformed fecal pellets a) XY principal section, $R_{xy} = 3.3:1$, assuming original spherical shapes. b) XZ section, $R_{xz} = 6.1:1$. XZ section oriented parallel to down dip mesoscopic lineation, XY section oriented parallel to fault strike, perpendicular to mesoscopic foliation. Field of view in both sections is 5.0 mm, left to right. (Location 88-430).
Fault-related foliations are unique to the Sadler Limestone and rarely, the Karmutsen Formation, but do not occur along all faults in these units. At higher stratigraphic levels, faults form discrete surfaces or anastomosing brittle fault zones and adjacent wall rocks lack penetrative fabrics. Anastomosing fault zones have a complex internal geometry characterized by fault slivers of tightly-folded, thinly-bedded limestone (Peril Formation) wedged between blocks of massive Karmutsen Formation or Sadler Limestone strata. Folds in these fault zones are strongly disharmonic, and have curvilinear axes subparallel to fault zone boundaries.

Other folds in the map area occur as macroscopic, broad, open structures and as open to tight, angular to subangular mesoscopic folds. On both scales, folds have subhorizontal, northwest-striking axes (Fig. 2.18c). Macroscopic folds form southwest-verging antiform/synform pairs, with antiformal traces lying approximately 500 m northeast of synformal traces. These macroscopic folds are most common in Kunga Group rocks, but broad folds also occur in Yakoun Group strata at Skedans Bay. Mesoscopic folds are most common in well-bedded strata of the Peril and Sandilands formations, and have wavelengths of several metres and sub-vertical axial surfaces (Fig. 2.20). Bed-length shortening associated with these folds typically measures 10–20%, but may locally exceed 40%. Most mesoscopic folds have southwest-directed asymmetry.
**Figure 2.20:** Mesoscopic southwest-verging fold in Sandilands Formation, southeast Louise Island (viewed looking down-plunge to northwest). Total shortening accommodated along this type of feature typically measures between 10% and 50%, estimated from bed length shortening. (Location 88-478)

**Discussion**

Most mesoscopic and macroscopic folds and faults mapped in the southeast Louise Island area can be attributed to a single episode of southwest-directed shortening. Down-dip mineral elongation fabrics in wall rocks adjacent to several of the west-northwest striking faults indicate a dominant dip-direction slip component. Dip-direction separation of stratigraphic markers is commonly only a few hundreds of metres in a north-side-up sense. Although the total shortening accommodated by any one fault is relatively minor, the close spacing (less than 1 km) of mappable faults allows larger regional strains to be accommodated. Estimates of shortening can be made from palinspastic restorations of cross sections drawn parallel to movement direction (Fig.
2.21a,b). With reasonable approximations of fault dip, fault spacing, bedding dip, and fault throw, shortening values of up to 25% are estimated.

**Figure 2.21:** Palinspastic restoration of northeast-southwest cross section through Louise Island map area (cross section same as shown on Map Plate V.): a) interpretive cross section showing present configuration, based on surficial geology. b) same cross section with fault offset removed. Total shortening calculated from change in section length is 25%.
Only approximate timing constraints on deformation exist in the map area. Both the Triassic-Lower Jurassic (Karmutsen Formation and Kunga Group) and the Middle Jurassic-Cretaceous (Yakoun Group, Longarm Formation) packages contain northwest-striking contractional structures. The greater strain intensity of these structures in the older rocks suggests initial shortening pre-dated Yakoun Group deposition. This timing is supported by the observation of a sharp angular unconformity between the Yakoun and Kunga groups at Cumshewa Inlet (Fig. 2.22; Thompson and Thorkelson, 1989), along trend with and approximately 20 km northwest of the map area.

Figure 2.22: Sharp angular unconformity between Sandilands Formation and Yakoun Group, Cumshewa Inlet. Beds below unconformity are subvertical and contain tight to isoclinal folds, strata above unconformity dip gently northward and contain only broad folds. Unconformity surface has up to 2 m of relief at this outcrop, and is not a structural detachment surface.
2.2.4 Skidegate Inlet/Skidegate Channel Geologic Transect

(Map plate VI)

The waterway formed by Skidegate Inlet and Skidegate Channel cuts east-west through the central Queen Charlotte Islands, separating Graham Island to the north from Moresby and adjacent islands to the south. The passage trends perpendicular to the structural grain of the islands, and provides an easily accessible cross section through several of the major structural features of the Queen Charlotte Islands. A strip map of the geology exposed along shorelines of the transect was completed as a means of providing structural continuity between the map areas examined in this study. This transect crosses the main strands of the Rennell Sound "fault zone" of Sutherland Brown (1968); his fault strands coincide with the approximate location of the newly-defined Long Inlet deformation zone (see section 2.2.2 above) at Skidegate Channel. The western terminus of the transect is just inboard of the Queen Charlotte Fault, and the eastern end approaches the Sandspit Fault. The transect thus provides an excellent section illustrating structural transitions between these major tectonic features.

*Stratigraphic distribution*

Rocks representative of most stratigraphic units present in the Queen Charlotte Islands crop out along the Skidegate Channel transect. Exposures of pre-Middle Jurassic rocks are concentrated at the western end of the transect around Chaatl Island, and in a central area encompassing parts of Maude, Sandilands, and northern Moresby islands. A foliated granodiorite pluton crops out south of Buck Channel, and may be a previously unrecognized element of the Middle Jurassic San Christoval plutonic suite. Between these two areas exposing older rocks, a wide variety of Middle Jurassic through Tertiary rock units crops out. Both this study, and concurrent map studies summarized by Thompson et al. (1991) show that the completeness of the Mesozoic stratigraphic
succession on northern Moresby and southern Graham islands has great lateral variability. Figure 2.23 summarizes this variability, and illustrates the following trends from west to east: At the westernmost end of the transect, the pre-Middle Jurassic "basement" succession (Karmutsen Formation and Kunga Group) is directly overlain by volcanic and sedimentary rocks of the Yakoun Group, which in turn are overlain by Tertiary volcanic strata. Moving east to Trounce Inlet, an intervening section of Longarm Formation sedimentary rocks is preserved below the Tertiary succession. Near east Skidegate Narrows, the Upper Cretaceous Honna Formation overlies the Longarm Formation and is succeeded by Tertiary volcanic rocks. East of the narrows, the Honna Formation directly overlies the Haida Formation; in this area it is unclear whether the Longarm Formation occurs below the Haida Formation. At Sandilands Island, all strata of the Maude, Yakoun, and Moresby groups and the Longarm Formation are missing, and both the Haida Formation and the Honna Formation directly overlie deformed Kunga Group strata. However, just 2 km east on Maude Island, the Maude, Yakoun, and Moresby groups are all present, and just to the south on northern Moresby Island the Longarm Formation occurs as well (Thompson 1990). Finally, between Kwuna Point and Onward Point, the Haida Formation is directly overlain by the Honna Formation, and both units overlie Yakoun Group strata.

This large and locally abrupt variation in stratigraphic preservation is not unique to the Skidegate Channel region. Thompson et al. (1991) describe similar variations throughout the central Queen Charlotte Islands, and were able to map northwest-trending belts defined by similar preserved successions. Of the areas mapped in this study, however, only the Skidegate Channel transect displays this variation so clearly. Thompson et al. (1991) ascribe the large variation in stratigraphic preservation to multiple episodes of Mesozoic and Cenozoic block faulting. They postulate that northwest-trending structural blocks were first differentially uplifted in the Late Jurassic,
Figure 2.23: Schematic diagram showing variation in stratigraphic preservation along Skidegate transect. See text for discussion of significance of distribution patterns. Geographic locations given on map plate VI.
leading to stripping of Yakoun Group strata from elevated areas. In their models, these
blocks were tectonically inverted in the Cretaceous and Tertiary, leading to complex
structural associations and different amounts of preserved Cretaceous strata. Taite
(1991a) documents similar patterns of stratigraphic distribution on Moresby Island at
Sewell Inlet and Tasu Sound, and also relates them to block faulting. The stratigraphic
distribution along the Skidegate transect argues for similar structural control: for
example, the rapid westward thinning of Haida and Skidegate formation strata at east
Skidegate Narrows may be related to Late Cretaceous (pre-Honna Formation) faulting
and erosion in that area. Elsewhere, gradual changes in stratigraphic thickness suggest
sedimentary onlap onto an irregular basement topography; this is best illustrated on
Maude and Skidegate islands, where both the Honna and Haida formations overlie Kunga
Group basement rocks. The role of structural control on stratigraphic distribution is
further discussed in following sections.

Structural Geology

Because of the wide variety of rock types and stratigraphic levels exposed and
the presence of several major tectonic features, the Skidegate transect provides a
sampling of most of the structural styles found in the central Queen Charlotte Islands.
Contact relationships along the transect provide additional constraints on the timing and
nature of deformation events inferred from the other map areas. In general, rocks
exposed on the western end of the transect represent lowest structural and stratigraphic
levels, and therefore, the following descriptions progress from west to east.

ChaatI Island, at the western end of the transect, consists mostly of moderately-
to highly-deformed, Karmutsen Formation volcanic rocks. The westernmost outcrops
examined comprise actinolite-epidote-chlorite schist with minor marble tectonite layers.
A sub-vertical foliation in these rocks trends 090° to 130° and contains a weak
subhorizontal lineation. This foliation locally forms a mylonitic fabric, usually limited to zones up to 10 m wide. Microscopic movement indicators are obscured in these rocks by post-tectonic recrystallization and alteration; however, brittle features preserved in one of the mylonite zones allow possible interpretation of movement direction. This locality contains en-echelon, sigmoidal quartz-filled veins which cross-cut the mylonitic foliation at moderate angles (Fig. 2.24). The veins are up to 2 m in length, and have an average orientation of 075°/90°. If it is assumed that they formed as a late-stage brittle response to the same deformation episode responsible for the mylonitic fabrics, their orientation and sigmoidal nature implies sinistral movement along the zone.

On Moresby Island south of Buck Channel, a moderately to strongly foliated granodiorite pluton crops out over an area of approximately 75 km². The large extent of this pluton was not recognized in earlier mapping (Sutherland Brown, 1968), nor was the structural significance of the foliation discussed. Petrographic characteristics of the pluton are similar to those of the Middle Jurassic San Christoval plutonic suite (Anderson and Greig 1989; Anderson and Reichenbach 1991) with which it is tentatively correlated, but, isotopic confirmation of a Jurassic age is pending. For purposes of discussion, this pluton is informally designated the Buck Channel pluton.

The intensity of foliation development in the Buck Channel pluton is unparalleled in any other plutons in the Queen Charlotte Islands. Foliation ranges from faint preferred orientation of phyllosilicate minerals to a highly mylonitic fabric defined by elongate quartz ribbon grains and very fine-grained recrystallized quartz fabrics. Feldspars generally form rigid, variably-fractured porphyroclasts. Mylonitic fabrics are most intense along and just south of Buck Channel; here, the foliation is subvertical and trends 085°–110°. Farther south the foliation is less well developed and less regularly oriented, although local mylonitization occurs. In general, the strike of the foliation changes to more northerly attitudes in more southerly locations.
Figure 2.24: Field sketch (a) and photograph (b) of ductile/brittle shear zone in Karmutsen Formation metavolcanic rocks, Chaat Island. Shear zone boundaries and submylonitic foliation strike northwesterly; crosscutting quartz-filled en-echelon sigmoidal veins strike northeasterly, consistent with late-stage brittle failure in sinistral shear zone. (Location 87-725)
A weakly- to strongly-defined, shallowly-plunging mineral elongation lineation is present in most locations. Kink bands are locally developed in the mylonitic fabrics. Kink band boundaries are steep, occur in a wide range of orientations, and have no consistent sense of asymmetry (Fig. 2.25a).

Fabric indicators in many of the submylonitic to strongly mylonitic rocks help constrain the kinematics associated with fabric development. These features are most obvious in rocks exposed just south of Buck Channel, where they consistently show sinistral shear senses. Mesoscopically visible indicators include composite S-C fabrics and asymmetric porphyroclast pressure shadows (Figs. 2-25b,c). Microscopic indicators include smaller scale examples of these, as well as low-angle synthetic shear bands and dynamically recrystallized grain shape orientation (Figs. 2.26a,b). All fabrics indicate a sinistral simple-shear component associated with fabric development, and some samples contain as many as three mutually consistent indicators.

Well-stratified rocks of the Kunga Group are exposed between Demariscove Point and West Skidegate Narrows, to the northeast of the Karmutsen Formation and Buck Channel pluton. The Sadler Limestone conformably overlies the Karmutsen Formation at Demariscove Point; the latter unit contains abundant variably oriented stylolites and calcite-filled extension veins. The overlying Sandilands and Peril formations host abundant outcrop-scale, northwest-plunging, open to tight chevron folds. These folds have steep axial surfaces and neutral to southwest-vergent asymmetry. Southwest-directed, outcrop-scale thrust faults cut fold limbs but do not contribute to substantial structural thickening.

The Yakoun Group is sparsely exposed between West and East Skidegate Narrows, and is inferred to unconformably overlie the units described above. Outcrops consist of both sedimentary and volcanic components, and dip directions are variable.
Figure 2.25: Mesoscopic kinematic features associated with the foliated Buck Channel pluton: a) kink bands of mylonitic foliation occur in a wide range of orientations and have no consistent sense of asymmetry (Location 90-840); b) composite S-C fabrics; S surfaces are vertical and strike 110°, C surfaces are subvertical and strike 100°; combination of fabrics defines sinistral sub-horizontal shear direction and sense (Location 90-854)
Figure 2.26: Thin section photomicrographs of sinistral shear indicators in submylonitic to mylonitic Buck Channel pluton: a) low-angle synthetic shear bands (C'); b) dynamically recrystallized quartz fabric inclined 20° clockwise to shear fabric. Both sections are cut from subhorizontal planes containing mesoscopic lineation and perpendicular to foliation, Location 90-769. Field of view for both photomicrographs is 2 mm, left to right.
Outcrop data are insufficient to infer any major structures. The contact between the Yakoun Group and the unconformably overlying Longarm Formation follows Skidegate Channel east for several kilometres from East Skidegate Narrows. Bedding attitudes in this area strike roughly east-west, and moderate to steep dips prevail. Several open, north-trending folds are inferred from gradual changes in bedding orientation.

The Skidegate transect crosses the Long Inlet deformation zone just east of Skidegate Narrows. Rocks present within the LIDZ here belong to the Longarm, Haida, Skidegate, and Honna formations. Structures are dominated by a major east-verging fold pair, the Long Inlet anticline and Long Inlet syncline; these folds are along trend with their counterparts at Long Inlet. The folds have a geometry similar to that in Long Inlet, but they are much less disrupted by younger faults. Skidegate Formation mudstones forming the anticline core contain a well-developed northwest-striking slaty cleavage parallel to macroscopic fold axial surfaces, pencil lineations parallel to major fold axes, and sparse mesoscopic folds. A small island exposing the west limb of the anticline near the fold hinge contains numerous easterly-verging (hingeward) minor thrust faults and associated drag folds. Map plate IV shows additional detail of this area.

Flat-lying Honna Formation strata occupy a several kilometre-wide outcrop belt between the LIDZ and Sandilands Island. East of this belt, on Sandilands, Maude, and northern Moresby islands, a window exposes basement rocks (Kunga and Maude groups) below the Cretaceous unconformity. These older rocks are folded by open to tight mesoscopic chevron folds, which are parasitic to a macroscopic, northwest-trending, southwesterly-overturned antiform. Mesoscopic folds are best exposed on the south shore of Maude Island, where they have consistent southwest vergence and moderately northwest-plunging axes. Total bed length shortening estimated from fold profiles here is approximately 20% (Figure 2.27), but additional uncertain amounts of shortening are accommodated by minor thrust faults along the fold limbs.
**Figure 2.27:** Schematic profile section through Middle Jurassic folds in the Sandilands Formation, Maude Island. Northwest-plunging folds accommodate 20% southwest-direct bed-length shortening.
Additional exposures of the Yakoun Group, and the only known exposures of the Moresby Group in the Queen Charlotte Islands, occur on eastern Maude Island and on northern Moresby Island near Kwuna Point. These areas were mapped by Thompson (1990) and were not re-examined in the present study. Both Thompson (1990) and Sutherland Brown (1968) show a northwest-trending, macroscopic syncline just west of Kwuna Point. Thompson (1990) mapped several steeply dipping, northwest-striking faults in this area (Mt. Poole, Sachs Creek faults), and suggests that multiple episodes of movement along them is responsible for the complex distribution of strata in adjacent areas (Thompson et al. 1991).

The easterly end of the Skidegate transect crosses shoreline exposures of shallowly to moderately west-dipping Haida Formation mudstone and siltstone. Broad open folds are mappable from changes in bedding attitude, but the rocks are otherwise largely undeformed. Concretionary mudstone outcrops west of Onward Point contain spaced cleavage which is consistently oriented sub-perpendicular to bedding. This cleavage bears no clear geometric relationship to the broad folds in this area, and its occurrence within calcareous concretions suggests it formed early in the lithification process. Spaced cleavage in these rocks is the subject of more extensive examination in a succeeding section.

Discussion

Structures exposed along the Skidegate Transect formed during several distinct deformation episodes which correlate well with deformation events inferred from map studies in the other areas. A Middle Jurassic deformation event, similar in geometry to that documented at the Rennell Sound and Louise Island map areas, is implied by the inferred angular unconformity at the base of the Yakoun Group. Folds and contractional faults exposed below this unconformity record moderate amounts of southwest-directed
shortening; the presence of these structures everywhere along the transect where older rocks are exposed attests to the regional nature of the Middle Jurassic event. Estimates of 25% shortening based on fold geometry represents a minimum for regional strain because they neglect components accommodated by contractional faulting and penetrative shortening. Middle Jurassic deformation is not easily documented in structurally deeper rocks of the Karmutsen Formation, where massive lithologies inhibited fold formation, and lack of mappable stratigraphy makes identification of throughgoing faults impossible.

The Buck Channel pluton almost certainly post-dates Middle Jurassic deformation, because the oldest dated plutonic rocks in the Queen Charlotte Islands are slightly younger than the post-tectonic Yakoun Group (Anderson and Reichenbach, 1989, 1991). The tectonic foliation in the pluton, and by inference mylonitic fabrics in the Karmutsen Formation occurring along trend on Chaatl Island, are latest Middle Jurassic or younger features. The fabrics are also probably pre-Tertiary, because in the early Tertiary the Chaatl Island area was unroofed prior to being covered by volcanic rocks (Indrelid et al., 1991a). It is unlikely that the area was at great enough depths in the Tertiary to form the greenschist-grade mineral assemblages associated with mylonitic fabrics in the Karmutsen Formation.

The LIDZ is one of the most structurally complex regions in the Queen Charlotte Islands, and several relationships exposed along the Skidegate transect provide important constraints to its evolution. Cretaceous rocks exposed in the LIDZ core contain mesoscopic folds, contractional faults, and flattening fabrics, all of which indicate southwest-northeast oriented shortening, perpendicular to deformation zone boundaries. These highly-deformed strata are unconformably overlain by gently-dipping volcanic strata. Although no good radiometric dates exist for these volcanic rocks, their
petrologic character is consistent with including them in the unnamed, Paleogene volcanic unit.

Stratigraphic distribution patterns adjacent to the LIDZ provide hints of older structures which might have controlled and localized deformation in the Cretaceous sequence. The apparent lack of Yakoun Group strata several kilometres to the east of the LIDZ at Sandilands Island suggests that the area was uplifted and susceptible to erosion in the Late Jurassic. West of the LIDZ, the Yakoun Group has a considerable stratigraphic thickness, both along the transect and in adjacent areas mapped by Thompson and Lewis (1990b). This same stratigraphic distribution is observed to the southeast along trend at Cumshewa Inlet (Thompson, 1990; Thompson and Lewis, 1990b), where it is attributed to Late Jurassic, dip-slip movement along the Dawson Cove fault (Thompson and Thorkelson, 1989; Thompson et al., 1991). The Dawson Cove fault cannot be clearly recognized in the Skidegate transect. However, if it is extended north from its most northerly mapped position on Moresby Island, it crosses the transect in the core of the LIDZ, where the most highly-deformed Cretaceous strata are found. It is likely that Late Cretaceous to Early Tertiary structural inversion of the fault led to strain localization in Cretaceous strata within the LIDZ, a relationship identical to that proposed above for the Long Inlet/Kagan Bay map area.

Steeply-dipping, northwest-striking faults mapped by Thompson (1990) near the eastern end of the Skidegate Transect are described by Thompson et al. (1991), and the present study provides no further constraints on their evolution. Based on stratigraphic and structural evidence, these workers outline a multiple movement history dominated by dip-slip offsets on these faults, spanning Late Jurassic to Tertiary time.

No significant changes in deformation style occur within Cretaceous rocks approaching the Sandspit Fault on the eastern end of the transect. Models proposing that the Sandspit Fault formed as a Late Cretaceous west-vergent thrust fault which provided
an easterly uplifted source for Honna Formation detritus (Higgs, 1988, 1990) are not supported by the present mapping (see also Lewis et al., 1991b).
2.2.5 Sandspit Fault (Copper Bay)

The Sandspit Fault was first identified by Sutherland Brown (1968) as a major northwest-trending lineament defined by surficial topography, offset drainage patterns, and Quaternary scarps and sag ponds. Sutherland Brown (1968) was able to trace the lineament across Graham and Moresby islands for 60 km, and noted its continuation in seafloor expression for an additional 80 km to the southeast in Hecate Strait. Magnetic anomalies along trend with the fault to the north suggest its continuation into Dixon Entrance. Despite this long strike length, the Sandspit Fault is very poorly exposed and previous estimates of the amount and timing of slip have been based on inferences from regional geology, geophysical studies, and Quaternary surface features. Sutherland Brown (1968) suggested a movement history with a large east-side-down component on a steeply-inclined surface, but also noted that strike-slip offsets are likely for a linear feature of such strike length. Yorath and Chase (1981) proposed that the fault represents the tectonic boundary between the Alexander terrane and Wrangellia and is the northern portion of a proposed composite Louscoone Inlet-Sandspit fault system. Young (1981) used potential-field data to propose at least 1500 m of east-side-down offset on a steeply east-dipping surface, based on Tertiary sediment thickness estimates. Well-data also show an abrupt eastward thickening of Tertiary sedimentary strata across the fault (Shouldice, 1971, 1973). Higgs (1988, 1990) postulates that the Sandspit Fault originated as a Late Cretaceous thrust fault, based on sedimentological studies of the Upper Cretaceous Honna Formation. As pointed out in the previous section, and more completely by Lewis et al. (1991b), his model is inconsistent with regional map studies in areas adjacent to the fault.

Surficial geology constraints on fault movement derive from two main sources: Firstly, the distribution of outcrops and structural styles in areas adjacent to the fault are documented by regional map studies. Secondly, a brittle fault zone on northeast
Moresby Island (Copper Bay), identified by Sutherland Brown (1968) as a splay of the Sandspit Fault, is easily accessed for detailed structural analysis. Regional mapping constraints are provided by Thompson (1990) for northern Moresby Island, by Hesthammer et al. (1991a) for southern Graham Island, and by Hickson (1990a, 1990b) for central and northern Graham Island. Thompson (1990) shows that areas west of the Sandspit Fault on northern Moresby Island are cut by Late Cretaceous or younger, east-side-down normal faults striking parallel to the Sandspit Fault, but recognizes no outcrops of Mesozoic or Cenozoic rocks east of the fault. Hesthammer et al. (1991a) find that Cretaceous strata crop out on both sides of the Sandspit Fault on southern Graham Island, placing rough limits on the likely magnitude of Cretaceous and younger vertical fault motion. Hickson (1990a) does not recognize a northwest continuation of the Sandspit Fault through central Graham Island, but suggests that offsets occurred along more northerly striking features on northern Graham Island. An east-dipping normal fault (the Spegonia Fault) explored at the Cinola mining camp on central Graham Island is interpreted to be a splay of the Sandspit Fault (Hickson, 1991). Also in this area, Richards et al. (1979) note both horizontal slickensides and offsets of geochemical anomalies as possible evidence for strike-slip displacements.

The present study attempted to add to these regional constraints through detailed mapping of the splay of the Sandspit Fault exposed at Copper Bay on northeast Moresby Island. The area mapped comprises a northeast-striking brittle fault zone characterized by intense fracturing and veining, and is exposed in beach outcrops approximately 8 km south of the town of Sandspit along the north shore of Copper Bay, immediately west of the Sandspit Fault (Fig. 2.28). Deformation is localized within an approximately 300 m long, northeast-trending zone of continuous outcrop; the width of the deformed zone varies from a few metres to several tens of metres. A 1:100 scale sketch map of part of
Figure 2.28: 1:100 sketch map of Copper Bay brittle fault zone. (Location shown in inset.)
this area was completed as a basis for structural analysis of the fault zone (Fig. 2.28).

Structural Geology

Bedrock exposures on both sides of the fault zone at Copper Bay comprise volcanic and epiclastic sedimentary rocks of the Bajocian Yakoun Group (Sutherland Brown, 1968; Thompson, 1990). These rocks are strongly chloritized in the most intensely deformed rocks, with lesser amounts of alteration in adjacent, less deformed areas.

The fault zone is characterized by abundant steeply-dipping calcite-filled fractures of variable width and continuity, and less common unfilled planar fractures and spaced joints. Calcite within the fractures occurs both as fibrous crystals oriented at high angles to fracture walls, and as fine-grained gouge which forms a matrix surrounding angular wall rock fragments (Figs. 2.29a, b). Traces of sulfide mineralization (pyrite, chalcopyrite) are common in many fractures.

Most fractures can be grouped by orientation into two sets: a steeply-dipping set striking 020° to 040°, and a steeply-dipping set striking 140° to 160°. Cross-cutting relations between these two sets are equivocal, suggesting latest movement involved offsets of both sets. Near the south end of the fault zone, an arcuate fracture system connects fractures of the northeast-striking set with those of the northwest-striking set (Fig. 2.30).

Northeast-striking fractures contain both gouge and fibrous calcite infilling, with fibrous infilling more common. This fibrous calcite is often brecciated and preserved as angular vein segments in a finer calcite gouge matrix. In contrast, northwest-striking fractures contain both gouge and fibrous calcite fabrics, but the brecciated vein segments described above are rare.
Figure 2.29: Types of calcite-filled fractures typical of Copper Bay brittle fault zone: a) Fibrous veins, with fibres roughly perpendicular to fracture walls and internal zoning; b) calcite gouge-filled fracture; note angular segments of fibrous calcite vein surrounded by fine calcite gouge matrix.
Fractures of both sets cut and offset various types of markers (dykes, other fractures), indicating a component of translation parallel to fracture walls. Slip directions of fractures were determined in the field using slickensides, conjugate fracture geometry, and geometric relationships to syn-tectonic extension veins. Fractures of both sets show evidence of primarily subhorizontal slip. Northwest-striking fractures usually have dextral offsets, while northeast-striking fractures have sinistral offsets. Where measurable, apparent offsets are up to 5 m, but greater offsets are likely on the most continuous fractures, which lack clear offset markers. Offsets are recorded on both gouge-filled and fibre-filled fractures.

Figure 2.30: Arcuate calcite-filled fracture systems, joining northeast-striking fractures with northwest-striking fractures.
The geometric form of the steeply-dipping curviplanar fractures suggests that they accommodated mostly down-dip displacements. These arcuate fractures either cross-cut or connect the planar fractures, indicating a component of dip-slip movement likely postdates the horizontal offsets.

Discussion

The fracture patterns described above contain elements which could not have formed within a fixed stress field, indicating their structural history involved either non-coaxial progressive deformation, or fracture re-activation due to changing stress configurations. If one assumes that fine-grained calcite gouge fillings formed as a result of displacement parallel to fracture walls, and fibrous vein fillings formed during extensional fracturing (Ramsay, 1980), a rough chronology of fracture movement events is evident. Brecciated fibrous vein segments set in fine-grained gouge likely indicate shear displacements superimposed on extensional fractures, and by similar reasoning, fibrous veins with measurable wall-parallel offsets might indicate extensional re-activation of shear fractures. Using these guidelines, most northwest-striking fractures could be best interpreted as early shear fractures with later extensional re-activation, and most northeast-striking fractures are interpreted as re-activated extensional fractures. The displacement directions on both sets are consistent with an overall sinistral sense of shear across the fault zone, and northwest- and northeast-striking fractures would respectively represent antithetic and synthetic shears related to this movement. The multiple-stage formation of the fractures, however, suggests a rotation of stress axes relative to fracture orientation through time, from a more northeasterly-directed to a more northwesterly-directed maximum compression. This apparent rotation may either be a regional phenomenon indicating two separate pulses of deformation, or it may be related to internal rotation of shear zone elements within a constantly oriented stress field.
Figure 2.31: Schematic diagram showing one possible model for the evolution of the Copper Bay fault zone. A) Fault zone formed originally as antithetic sinistral brittle shear zone related to dextral strike-slip movement on the Sandspit Fault. Northwest-striking antithetic (R$_2$) shears formed synchronously with northerly-striking extensional fractures. B) Dextral offset along Sandspit Fault, and sinistral offset along Copper Bay fault zone and proposed similarly oriented faults results in small clockwise tectonic rotation of areas directly west of the Sandspit Fault. Extension fractures formed during A) rotate out of principal stress plane, and re-activate as R$_1$ shear planes; R$_2$ fractures rotate toward $\sigma_1$ direction and re-activate with mostly dilational component of movement. Alternatively, a counter-clockwise rotation of the external stress field during separate tectonic events could result in similar re-activation sequences to that of the proposed model.
during progressive deformation (Fig. 2.31). The latter would be favoured if the fault zone is one of a family of similarly oriented sinistral fault zones: synchronous movement along these fault zones could lead to clockwise rotation of both the faults and the intervening fault-bound blocks, or an apparent counterclockwise rotation of stress axes if one uses the fault blocks as a fixed reference frame. The dip-slip offsets inferred from arcuate fracture trends appear to post-date the horizontal offsets, and are not related to development of the fracture zone.

Although it is impossible to unequivocally show that the fault zone exposed at Copper Bay is genetically linked to movement on the Sandspit Fault the proximity to the Sandspit Fault, the lack of deformation in surrounding rocks, and the fracture trends present are all consistent with the fault zone being an antithetic structure to the larger structure. Such linkage would suggest that initial movement on the Sandspit Fault involved dextral strike-slip displacement, followed by unknown amounts of dip slip displacement. East-side-down dip-slip movement on the Sandspit Fault is supported by the present distribution of Tertiary sediments, which indicates substantial Tertiary movement. This interpretation suggests that the fault is a steeply-dipping surface, which is supported by the linear surface expression, Quaternary scarps, and gravity anomaly patterns.
2.3 Southern Queen Charlotte Islands

Sutherland Brown (1968) shows that the Queen Charlotte Islands south of Louise Island are in several ways geologically distinct from the more northerly parts of the islands. The main difference he shows is the more northerly structural trends found in southern areas, dominated by the north-northwest-striking southern (Louscoone Inlet) portion of the "Rennell Sound/Louscoone Inlet fault system." Plutonic rocks are shown as more volumetrically significant in surficial exposures, and stratified rocks are dominated by Triassic and Lower Jurassic units, rather than the Cretaceous and younger units present in the central and northern islands. Differences are also reflected in more recent FGP studies: a more intense thermal history for the southern islands has been established through examinations of conodont color alteration indices (Orchard and Forester, 1991) and levels of organic maturation (Vellutini, 1989; Vellutini and Bustin, 1991). These differences suggest that the geologic history for southern parts of the islands contains elements unique to these areas, and that a structural model for the complete Queen Charlotte Islands region must utilize field constraints from all parts of the islands. This need was fulfilled in the present study through mapping at Burnaby Island and Juan Perez Sound, an area containing a wide range of rocks and structural styles.

2.3.1 Burnaby Island/Juan Perez Sound Map Area

(Map plate VII)

Mapping at the Burnaby Island/Juan Perez Sound area was originally directed toward establishing the structural history of the Louscoone Inlet fault system (LIFS), which transects the area in a roughly north to south direction. The LIFS comprises a group of subparallel, north- to northwest-striking faults which extend over 120 km through the southern Queen Charlotte Islands (Fig. 2.32).
Originally, the fault system was considered by Sutherland Brown (1968) to be the southern portion of his Rennell Sound/Louscoone Inlet fault system. Based on structural style, mesoscopic tectonic fabrics, and offset fold trends, Sutherland Brown (1968) hypothesized Late Mesozoic and Cenozoic dextral strike-slip displacements of between 20 km and 100 km along the LIFS. However, these offset estimates were not well constrained, being based mostly on possible offsets of major map features with little detailed structural analysis. Sutherland Brown's (1968) interpretations strongly influenced Yorath and Chase (1981), who proposed a tectonic model linking...
displacement on the fault system to Neogene rifting in the southern Queen Charlotte Basin. They considered the LIFS and the Rennell Sound fault zone (RSFZ) to be separate entities, with the RSFZ offsetting the LIFS from its postulated northern extension, the Sandspit Fault.

Mapping at Long Inlet and Rennell Sound as part of this study, and new regional mapping of the central Queen Charlotte Islands (Lewis et al. 1990b; Thompson and Lewis 1990a, 1990b; Thompson 1990) has greatly expanded our understanding of the evolution of the northern parts of the Rennell Sound/Louscoone Inlet fault system. This segment, now recognized as the Long Inlet deformation zone, is now interpreted to have formed as a result of episodic shortening and block faulting with little or no strike-slip displacement along it. Recent mapping in the Louise Island area recognizes some of the northern strands of the LIFS shown by Sutherland Brown (1968), but demonstrates that they do not offset Late Cretaceous/Early Tertiary folds and faults of the LIDZ. At Selwyn Inlet, the base of the Tertiary volcanic section projects across the most continuous strand of the LIFS with little or no apparent offset (Thompson and Lewis, 1990a; Thompson et al., 1991). These relationships effectively "lock" the northern part of the LIFS beginning in the Early Tertiary. However, reconnaissance examinations along the southern portions of the LIFS reveal sub-horizontal mineral fabrics (Sutherland Brown, 1968; Anderson, 1988; Woodsworth, 1988) which may have formed in response to strike-slip displacements. The anastomosing fault splays shown by Sutherland Brown (1968) also suggest a strike-slip fault geometry. These observations present an apparent contradiction: reasonable evidence for strike-slip displacements in the southern islands must be balanced against mapping constraints in the central islands which lock the fault since the Tertiary. Several possible solutions are obvious: Strike-slip displacements may be Mesozoic or older, and still satisfy most map constraints. Alternatively, reconnaissance studies in the southern islands may have missed important structural
constraints. A third solution may be that strike-slip displacement decreases to the north along the fault system, and is transferred into accommodation structures in one of the fault blocks.

Present mapping in the Burnaby Island/Juan Perez Sound area was aimed at evaluating these possible solutions and reconciling the apparently disparate observations of previous workers, as well as adding new detailed map coverage for the southern Queen Charlotte Islands. Shoreline exposures in this area provide a good transect through the fault system, and the high density of islands gives some of the best exposure found in the Queen Charlotte Islands. The new mapping described below shows that structures present in the area are compatible with several tens of kilometres of strike-slip movement along the LIFS. However, structures adjacent to the main fault strands implicate a structural model satisfying mapping constraints outlined by Thompson and Lewis (1990a) and Thompson et al. (1991) for the northern part of the fault system.

**Stratigraphic Distribution**

Rocks in the map area range from Late Triassic to Tertiary in age. Oldest rocks, represented by the Upper Triassic Karmutsen Formation, comprise basaltic composition pillowowed flows, volcanic breccia, and massive flows, and form extensive outcrops on Moresby Island. They are conformably overlain by Upper Triassic and Lower Jurassic carbonate and siliciclastic rocks of the Kunga Group (Sadler Limestone, Peril Formation, and Sandilands Formation), which are best exposed at Huxley Island, Skincuttle Inlet, and along shorelines of Burnaby Island. The lowermost unit exposed, the Sadler Limestone, locally interfingers with pillowowed flows of the underlying Karmutsen Formation. A Middle Jurassic unconformity separates this sequence from sparsely distributed pyroclastic and volcanogenic sedimentary rocks of the Bajocian Yakoun Group. The Yakoun Group may have covered extensive parts of the map area when
deposited, but it is presently limited to three isolated outcrop areas. No stratigraphic record exists for the remainder of the Middle Jurassic and all of the Upper Jurassic. A second major unconformity occurs at the base of the Lower Cretaceous Longarm Formation, and is spectacularly exposed on Arichika Island and at numerous locations in Skincuttle Inlet. At these localities a basal pebble to boulder conglomerate transgressive lag deposit (Haggart and Gamba, 1990) grades upward into thickly-bedded to massive sandstone. Turbiditic sandstone and siltstone of the Skidegate Formation, and mudstone and siltstone of the Haida Formation form the remainder of Cretaceous strata in the map area. The Skidegate Formation defines a northerly-trending outcrop belt on western Burnaby Island; the Haida Formation crops out on Ramsay and adjacent islands, and on eastern Moresby Island south of Goodwin Point. Elsewhere in the Queen Charlotte Islands these units conformably succeed sandstones of the Longarm Formation (Haggart, 1991); this relationship breaks down along the south shore of Ramsay Island, where Longarm Formation conglomerate and sandstone overlies Lower Cretaceous turbidite beds lithologically identical to those in the Skidegate Formation. Youngest stratified rocks in the map area are Tertiary volcanic flows, lahars, and interstratified volcanogenic sedimentary rocks, all of which unconformably overlie the Mesozoic rocks. Volcanic components of these units are intermediate to mafic in composition and commonly feldspar-phyric. Exposures of these Tertiary volcanic strata are limited to the northern map area on Huxley, Ramsay, and adjacent islands. On Ramsay Island, Oligocene ages for the flows are constrained by K-Ar whole rock dates of 35.9 ± 1.4 Ma and 41.1 ± 1.4 Ma (compiled by Hickson, 1991).

Intrusive igneous rocks in the map area include Jurassic and Tertiary plutons and stocks and extensive Tertiary dykes. The Middle and Late Jurassic Burnaby Island plutonic suite (U-Pb = 158–168 Ma, Anderson and Reichenbach, 1991) is represented by the Poole Point, Jedway, and Collison Bay plutons (Anderson and Greig, 1989). These
intrusions are dominantly medium-crystalline quartz monzonite, quartz monzodiorite, and quartz diorite, and are characterized by pervasive alteration and veining along fracture surfaces (Anderson and Greig, 1989). Skarn deposits are co-spatial with these plutons where they intrude carbonate lithologies of the Kunga Group. These deposits have been extensively mined at Jedway and Ikeda Cove, as well as explored by drilling and tunnelling throughout the eastern Skincuttle Inlet area. The Middle Jurassic (U-Pb = 171-172, Anderson and Reichenbach, 1991) San Christoval plutonic suite was not examined in the present mapping, but is known from mapping by Sutherland Brown (1968) to form extensive outcrop in the southwest corner of the Burnaby Island map sheet, just outside the present map-area limits.

A north-trending belt of northerly-elongate Tertiary(?) plutons intrude Cretaceous and older units on western Burnaby Island. Most of these intrusions are fine to very fine-grained, highly-altered diorite to quartz diorite, and some have foliated margins.

Dykes are concentrated in several swarms at Skincuttle Inlet (Burnaby Island and Carpenter Bay dyke swarms, Souther, 1989), and locally compose up to 80% of outcrops. Souther and Jessop (1991) provide a detailed description of dyke orientation, composition, and probable age, including a map showing the dyke distribution at Skincuttle Inlet. Most dykes shown strike northerly, and are interpreted to be genetically related to Early Tertiary volcanism. Eocene K-Ar whole rock dates from dykes at Ikeda Point (41.1 ± 4.6 Ma, Souther and Jessop, 1991) and Carpenter Bay (43.7 ± 1.1, Anderson and Reichenbach, 1989) overlap dates from volcanic rocks to the north within uncertainty limits, and are in close agreement with U-Pb dates from the Carpenter Bay pluton (Anderson and Reichenbach, 1991).
Structural Geology

The geology of the Burnaby Island/Juan Perez Sound map area is best described as three distinct structural domains separated by north-northwest trending boundaries (Map plate VII). The western domain, lying mostly on Moresby Island west of Burnaby Strait, is characterized by moderate southwest stratal dips in internally undeformed rocks. The central domain coincides with the major strands of the LIFS as mapped by Sutherland Brown (1968). It forms a 3 km-wide zone of abundant north-northwest-striking faults and intrusive contacts, and contains locally intensely folded and faulted rocks which often have penetrative structural fabrics. The eastern domain contains moderately north-dipping beds, cut by orthogonal east-west and north-south fault systems.

Western Domain

All rocks in western domain belong to the Triassic Karmutsen Formation, except at George Bay, where fault-bound Sandilands Formation and Yakoun Group strata crop out. Regional dips within the Karmutsen Formation average 30° to 50° to the southwest, except west of Tangle Cove, where thickly-stratified volcanic breccias dip to the south-southeast. Rocks in the western domain are mostly undeformed, and many primary igneous features are well preserved (Fig. 2.33). The only mappable structures present are narrow (<4 m wide) steeply-dipping brittle fault zones with variable trends. These fault zones commonly contain concentrations of carbonate gouge and breccia, and are most abundant in the eastern part of the domain along Burnaby Strait.
Central (fault zone) Domain

The boundary between western and central structural domains is obscured by water in northern Burnaby Strait, and occurs within the Karmutsen Formation in southern Burnaby Strait and Skincuttle Inlet. Here, the boundary is gradational over several hundred metres and divides rocks with well-preserved original textures (western domain) from those containing superimposed penetrative fabrics (central domain). Rocks in the central domain range in age from Late Triassic (Karmutsen Formation) to Tertiary (unnamed volcanic rocks and intrusions). These map units outline northerly-elongate belts bounded by faults and intrusive contacts. Steeply-dipping macroscopic faults occur in two main orientations: a set striking 160° to 180°, and a set striking 110° to 130°. Most commonly, the northwest-striking faults cut the more northerly-striking faults, but in places the opposite relationship occurs. Where faults are exposed, they generally form brittle or ductile-brittle shear zones with a mix of mylonitic, submylonitic, and cataclastic
fabrics. Unequivocal offset markers are lacking, leaving these fabrics the primary means
of inferring kinematic histories. Fault fabrics are present to varying degrees in all map
units, but are best displayed in the Karmutsen Formation and Kunga Group in shear
zones at Slim Inlet, Smithe Point, and Section Cove.

\textit{Slim Inlet and Smithe Point shear zones:}

Slim Inlet and Smithe Point are located along north-northwest striking faults, and
are along structural trend from one another. Both locations are characterized by
subvertical penetrative foliations containing weakly- to strongly-developed subhorizontal
elongation lineations. At Smithe Point, these foliations occur in faulted slivers of the
Karmutsen Formation and Sadler Limestone, and strike 160° to 175° (Fig. 2.34). Steep,
outcrop-scale dextral faults which cut the foliation strike 005° to 015°. Foliated
limestone pods bounded by these faults have asymmetric sigmoidal shapes in plan view
which are consistent with formation in a dextral shear zone (Fig. 2.36).

At Slim Inlet, penetrative schistosity and mylonitic foliation in the Karmutsen
Formation generally strikes northerly, but is locally re-oriented by mesoscopic Z-folds
and crenulations (Fig. 2.37). Both folds and crenulations have clockwise asymmetry,
variably plunging axes contained in the regional foliation, and vertical axial surfaces
oriented roughly 100°-110°/90°. This geometry is consistent with their origin as late-
stage flattening features superimposed on a dextral shear zone. Asymmetric boudins
formed from pulled-apart calcite veins are also consistent with dextral shear.

Thin sections of fabrics from Smithe Point and Slim Inlet confirm the kinematic
interpretations made from mesoscopic features. Sections of tectonized Karmutsen
Formation from Slim Inlet comprise a very fine-grained matrix of calcite and chlorite
surrounding euhedral plagioclase porphyroclasts. Pressure shadows adjacent to the
Figure 2.35: Foliated fault sliver of Sadler Limestone at Smithe Point shear zone, Burnaby Island. Foliation strikes north-northwest (160°-175°) and contains subhorizontal mineral elongation lineation. (Location 90-615)

Figure 2.36: Termination of asymmetric limestone boudin, Smithe Point shear zone, Burnaby Island. Asymmetry defines sinistral shear sense, consistent with that interpreted from microscopic indicators (Figs. 2.37a-b).
Porphyroclasts consist of fibrous chlorite and calcite, and have delta-shapes defining dextral shear (Fig. 2.37a). Tectonized limestone samples from Smithe Point consist of a very fine-grained (10–20 µm) dynamically recrystallized matrix surrounding scattered calcite porphyroclasts. Dextral shear is confirmed by three separate features, all visible in a single thin section: dextral synthetic faults inclined to the foliation at small angles (<15°) cut calcite porphyroclasts; these calcite porphyroclasts contain internal deformation twin lamellae inclined with the opposite sense to the dominant fabric; and finally, a weak preferred orientation is visible in the recrystallized matrix which is parallel to the deformation twin lamellae (Fig. 2.37b).
Figure 2.36: Field sketch showing configuration of kinked and folded mylonitic foliation in Slim Inlet shear zone, Moresby Island. Fold orientation is consistent with late stage re-orientation of shear fabric during progressive, sinistral shear deformation.

Figure 2.37: Thin section photomicrographs showing dextral shear sense indicators at Slim Inlet and Smith Point shear zones: a), pulled-apart calcite porphyroclast; both low-angle shear plane and deformation twin plane orientation consistent with dextral shear (Location 90-615). b), plagioclase porphyroclast with asymmetric pressure shadows. Field of view is 150 mm left to right in both figures. (Location 90-542).
Section Cove shear zone (SCSZ)

At Section Cove on Burnaby Island, a west-northwest-striking brittle shear zone approximately 200 m wide separates strata of the Kunga Group and Karmutsen Formation from stratigraphically higher mudstones and siltstones of the Skidegate Formation. This shear zone is characterized by intense brecciation and pervasive calcite veining along fractures. Throughgoing shear surfaces strike 085° to 110° and dip steeply to both the north and south. Due to poor exposure, the shear zone cannot be traced inland from the shoreline, so its absolute orientation is uncertain. Regional stratigraphic distribution indicates a trend of 100° to 120°, approximately parallel to the throughgoing fractures within the zone.

The Section Cove shear zone consists of relatively undeformed blocks of Karmutsen Formation, Sadler Limestone, and feldspar-phyric intrusive rocks, surrounded by a highly-tectonized, fractured, and veined matrix of dark grey mudstone and calcareous siltstone. This matrix is probably derived from rocks of the Skidegate, Sandilands, and Peril formations, but its highly tectonized nature has obscured all primary features, making positive identification impossible.

The geometry of the massive blocks is chaotic, but a host of brittle mesoscopic indicators in the surrounding matrix allows sense and direction of net slip in the shear zone to be determined. Most useful brittle features are shear veins, elongation fabrics, and extension veins. The relative orientations of, and cross cutting relationships between, these structures are illustrated schematically in figure 2.38. Shear veins form calcite-filled fractures within which calcite fibres are oriented subparallel to fracture walls. They indicate a relative displacement of fracture walls parallel to fibre elongation direction, with a small dilational component allowing calcite precipitation within the fracture (Ramsay and Huber, 1983). In the SCSZ, these shear veins are common along the throughgoing, west-northwest-striking fracture surfaces. Calcite slickenfibres plunge
shallowly to the east and west, indicating a large component of sub-horizontal motion (Fig. 2.39a). A sinistral sense of slip is indicated by the asymmetric form of the shear veins (Fig. 2.39b); this asymmetry results from irregularities or steps in the fracture surfaces, which create openings during fault movement.

Independent confirmation of this sense of movement along shear veins was obtained through systematic orientation analysis of fibres in a host of differently oriented fractures. Ideally, because fibres in shear veins join points which were once in contact, they indicate relative displacement direction of fracture walls, and therefore the direction of maximum shear stress on a given fracture surface during faulting. This shear stress direction is dependent on four parameters of the local stress tensor: the orientations of the three principal stresses, and a constant $\Phi$ expressing the relative magnitudes of these stresses:

$$\Phi = \frac{\sigma_2 - \sigma_3}{\sigma_1 - \sigma_3}$$  \hspace{.5cm} (2.1)

Slip directions for any four fault planes which were active in a homogeneous stress field will therefore uniquely determine these four components. Several workers have presented algorithms which converge on a best-fit solution for redundant (i.e., more than four fault planes) data sets (e.g., Angelier, 1979, 1984; Etchecopar et al., 1981). Allmendinger et al. (1989) presented a computer program which was applied to fault slip data from the SCSZ. The slip data, when inverted following their method, result in a stress tensor with maximum principal stress oriented approximately 055/05 and a minimum principal stress oriented 145/00 (Appendix D). These values correspond to a stress field favouring sinistral strike-slip movement along west-northwest striking vertical surfaces, which is consistent with mesoscopic fabric indicators.
Figure 2.38: Field sketch showing types and relative orientations of brittle features, Section Cove brittle shear zone.

Figure 2.39: Photographs of shear veins at Section Cove brittle shear zone: a) sub-horizontal calcite fibres are subparallel to fracture walls, indicating sub-horizontal slip; b) asymmetric form of calcite filled pull-aparts indicate dextral movement along fracture (individual fibres join points on fracture walls which were original adjacent). (Location 90-648)
Planar calcite-filled extension veins are also common within the shear zone, and have an orientation roughly consistent with the stress tensor calculated above. Most extension veins are subvertical, strike 020° to 040°, and contain straight fibres approximately perpendicular to vein walls. In places, the east-northeast-striking shear fractures offset the veins several centimetres in a left-lateral sense; elsewhere the extension veins crosscut fractures of this orientation.

Weakly-developed elongation fabrics are visible in the cataclastic matrix of the shear zone. This fabric is defined by the parallel preferred orientation of angular rock fragments ranging in length from 1 cm to 10 cm. Largest clast faces are subvertical and oriented 110°-140°, shallowly inclined to the throughgoing shear veins, an orientation again consistent with an inferred sinistral slip sense in the shear zone.

In addition to the fault-related structures, mesoscopic folds are common in the central zone, especially in the thinly- to medium-bedded strata of the Peril, Sandilands, and Skidegate formations. Most folds are open to tight and contain axial-planar penetrative cleavage (Fig. 2.40). Fold axes plunge shallowly to steeply to the northwest and southeast, and subvertical axial surfaces strike from 130° to 160° (Fig. 2.41). Deformed bedding plane ammonite prints show elongation directions parallel to cleavage traces (Appendix B).

Eastern Domain

The eastern structural domain is characterized by orthogonal north-south and east-west fault sets and consistent northerly bedding dips in stratified rocks. The boundary between the central and eastern domains is gradational over 1 to 2 km. Strata present range from Triassic to Tertiary, and are intruded and thermally altered by Jurassic and Tertiary plutons (Anderson and Greig, 1989).
Rocks in the eastern domain lack penetrative structural fabrics and are largely unfolded, even in locations where well-bedded strata (Sandilands and Peril formations) are exposed continuously for several kilometres. Strata of all ages throughout the domain exhibit northerly to northwesterly dips of 20° to 40° (Fig. 2.42a). Dominant structural features are two nearly orthogonal sets of major faults (Fig. 2.42b). Faults of both sets form either discrete slip surfaces or narrow (<0.5 m) brecciated zones. Faults striking 070° to 100° dip gently to steeply southward, and most have tens to hundreds of metres of normal offset (south-side-down). In areas with good exposure and stratigraphic control, for example the islands in Skincuttle Inlet, these normal faults occur with a spacing of a few hundreds of metres or less. In less well exposed areas, only faults with significant offset and strike length can be mapped; these larger faults have a 1–2 km spacing and offsets of hundreds of metres.

The normal faults are cut and offset by subvertical to steeply east-dipping faults striking 150° to 180°. Fault surfaces of the latter set contain both subhorizontal and subvertical slickensides, and apparent slip directions are right-lateral or east-side-down.

The Early Tertiary Skincuttle Inlet dyke swarm is located completely within the eastern structural domain (Souther, 1989; Souther and Jessop, 1991). Dykes here locally form up to 80% of outcrop volume. In general, individual dykes are subvertical, tabular, and strike northerly; dyke orientation is independent of bedding orientation in country rocks. Paleomagnetic analyses of several dykes sampled at Skincuttle Inlet reveal magnetic vectors 10° to 20° steeper than Eocene cratonic paleopoles (T. Irving, personal communication, 1991). This deviation could either result from equivalent amounts of post-emplacement northerly tilting of the dykes, or alternatively, could signify southward latitudinal displacement of the islands since dyke emplacement. The former interpretation is favoured because it is consistent with the northerly stratal dips in volcanic units of similar age on Ramsay Island.
**Figure 2.40:** Fold hinge area of folded Cretaceous turbiditic sandstone (Skidegate Formation) with axial-planar cleavage at Section Cove, Burnaby Island. Axial planar cleavages are rare in Cretaceous rocks not in the Long Inlet deformation zone or adjacent to the Louscoone Inlet fault system. (Location 90-221)

**Figure 2.41:** Equal area stereographic projections (lower hemisphere) showing structural elements of central (fault zone) domain, Burnaby Island/Juan Perez Sound map area: a) poles to bedding broadly distributed along girdle defining northwest-trending \( \pi \)-pole; b) northwest-trending fold axes lie on steep axial surfaces.
Figure 2.42: Equal area stereographic projections (lower hemisphere) showing structural elements of eastern domain, Burnaby Island/Juan Perez Sound map area: a) Poles to bedding show strong clustering on north-northwesterly dips; b) Fault plane data cluster in two groups, a north-striking subvertical set, and a moderately to steeply south dipping set.

Discussion

Features in the Burnaby Island/Juan Perez Sound area support a structural history dominated by dextral strike-slip faulting along the LIFS and north-south extension east of the fault system. Most dextral fault displacements occurred on north- to northwest-striking faults distributed across the width of the central domain. The most compelling evidence supporting a strike-slip interpretation comes from mesoscopic and microscopic kinematic indicators preserved in the Slim Inlet and Smithe Point shear zones. The orientations of flattening fabrics and fold axial planes within the fault zone, although not diagnostic if taken by themselves, are consistent with dextral offset along a north- to northwest-striking fault system. West-northwest-trending folds of mylonitic fabrics at
Slim Inlet (Fig. 2.37) likely record late stage folding superimposed on the older, more northerly-striking shear fabric. Microscopic fabrics indicate that shear zones formed at lower greenschist grades of metamorphism, accompanied by fibrous calcite and chlorite growth in pressure shadows.

Macroscopic fault orientations also support a strike-slip interpretation. The anastomosing fault geometry of the central domain is typical of wrench fault systems (e.g. Moody and Hill, 1956). Sinistral, west-northwest striking brittle fabrics of the Section Cove shear zone are antithetic to the dominant north-northwest striking fault system, although timing constraints on movement are poor. The age of rocks affected (Triassic to Tertiary), variable cross-cutting relationships, and similarities in structural style all suggest contemporaneous formation.

Structures in the eastern domain support a history of south-directed, asymmetric extension, roughly coeval with transcurrent faulting in the central domain. Well-bedded strata of the Sandilands and Peril formations in eastern domain are virtually non-folded on the outcrop scale, and show no evidence of the Mesozoic shortening event documented in the central Queen Charlotte Islands (this study; Lewis and Ross, 1991; Thompson et al., 1991). The combination of gentle to moderate northerly stratal dips and moderate to steeply-dipping, south-side-down normal faults supports a structural model of asymmetric extension accompanied by fault block rotation about horizontal axes ("bookshelf" or "domino style" faulting). Tertiary tilting is independently confirmed by paleomagnetic studies (T. Irving, personal communication, 1991).

Normal faults mapped in the eastern domain do not extend into the central and western domains. This suggests that normal faulting was either antecedent to or synchronous with transcurrent faulting in the central domain. If normal faulting preceded transcurrent faulting, structural styles similar to those in the eastern domain should occur offset to the north on the west side of central domain. Neither present map
studies (Thompson et al., 1990; Taite, 1991b) nor Sutherland Brown's (1968) original mapping recognize such structures, although Taite (1991b) recognizes some evidence for limited northward tilting in Tertiary and Cretaceous strata on central Moresby Island.

Major faults were not mapped in the western domain, which behaved as a rigid block during deformation. The timing of southwest tilting of this block is unconstrained.

The best controls on timing of strike-slip and extensional faulting are provided by stratigraphic-structural relationships in the central and eastern domains. In the central domain, Skidegate Formation strata yielding Early Cretaceous inoceramids are cut by both sets of faults, and contain northwest-trending folds and flattening fabrics. Volcanic and intrusive rocks cut by these same faults are undated, but, on the basis of lithologic character and contact relations, are interpreted to be of Tertiary age. More conclusive timing relationships are provided in the eastern domain on Ramsay Island. Oligocene volcanic strata here have northward tilts of similar magnitude to those in older strata to the south at Burnaby Island and Skincuttle Inlet, indicating that tilting and extension in the eastern domain, and by inference, strike-slip faulting in the central domain, were active since the Early Oligocene. No upper age limit for faulting is constrained.

Relationships to structures outside the map area

The dextral strike-slip faulting and associated extension in the Burnaby Island/Juan Perez Sound map area represents a structural geometry that has not been documented elsewhere in the Queen Charlotte Islands. Sutherland Brown (1968) proposed that strike-slip offset along the LIFS was transferred to the more westerly-striking Rennell Sound portion of the Rennell Sound/Louscoone Inlet fault zone at Louise Island. However, recent field mapping in this area shows that Cretaceous/Tertiary fold trends of the Long Inlet deformation zone are continuous across the most northerly extension of the LIFS, even though their formation must pre-date any
Tertiary movement on the LIFS (Thompson and Lewis, 1990). Faults mapped by Sutherland Brown (1968) as elements of the northern LIFS are either not recognized in this recent mapping, or have only minor offset. For example, the Carmichael Passage fault, which has hundreds to a few thousand metres apparent dip-slip offset at Louise Narrows, does not offset contacts along the north shore of Cumshewa Inlet (Thompson and Lewis, 1990a).

Reconnaissance studies at several locations outside of the Burnaby Island/Juan Perez Sound map area where Sutherland Brown (1968) mapped the LIFS recognize structural styles similar to those in the central domain. The most spectacular examples are at Shuttle Island in Darwin Sound. There, mylonitic fabrics are extensively developed over a 300 m-wide zone in altered rocks of the Karmutsen Formation and a fine-grained intrusion. The mylonitic foliation consistently strikes north-northwest (150° to 170°) and dips moderately to steeply to the southwest and northeast. A well-developed subhorizontal mineral lineation is ubiquitous, and asymmetric fabrics and kinematic indicators consistently show dextral offset. Best examples of these indicators include dextral kink bands and microfolds (Fig. 2.43a), and asymmetric boudins formed from segmented dykes (Fig. 2.43b).

Sutherland Brown (1968), Anderson (1988), and Woodsworth (1988) described mylonitic fabrics in the Karmutsen Formation and Jurassic intrusive rocks on southern Kunghit Island at Luxana and Howe bays. These locations were not examined in the present study, but the fabrics they describe strike north-northwest, have subhorizontal mineral lineations, and are along trend with the LIFS at Skincuttle Inlet. East of the Luxana Bay and Howe Bay mylonite zones, Sutherland Brown's (1968) map shows subhorizontal dips in Triassic and Jurassic Kunga Group strata. This implies that the extension-related block tilting documented in the eastern domain at Skincuttle Inlet may not extend far south.
Figure 2.43: Mesoscopic fabric elements in Karmutsen Formation rocks at Shuttle Island: a) asymmetric kink bands of mylonitic foliation; kink band boundaries of major set are subvertical and strike 100° to 110° and mylonitic foliation strikes 150° to 160°; b.) asymmetric boudin formed from segmented dyke. Both features are consistent with sub-horizontal dextral shear along the mylonitic foliation. (Location 90-1195)
A structural model for Burnaby Island/Juan Perez Sound and adjacent areas

The structural history described above for the Burnaby Island/Juan Perez Sound area, when combined with constraints provided by mapping in the central Queen Charlotte Islands (this study; Thompson et al., 1991), defines a structural model for the southern Queen Charlotte Islands centered largely on Tertiary strike-slip movement on the LIFS. The Mesozoic structural history outlined above, and by Thompson et al. (1991), for the central Queen Charlotte Islands involves three regional structural events: (1) Middle Jurassic southwest-directed contractional faulting and folding, (2) Middle and Upper Jurassic block faulting, and (3) Late Cretaceous/Early Tertiary folding. The two major shortening events are concentrated in the central islands, cospatial with the trace of the northern part of the RSFZ outlined by Sutherland Brown (1968), and they contributed to moderate structural thickening of the supracrustal assemblage in this region. Mesozoic structural thickening is not recognized in the Burnaby Island/Juan Perez Sound area, nor is there clear evidence for Late Jurassic block faulting, although recognition of the latter would be hindered by the paucity of Middle Jurassic strata in the area. Beginning in Tertiary time with faulting on the LIFS, structural links between the central and southern islands are apparent. Strike-slip motion on the LIFS did not extend into the central Queen Charlotte Islands, but it dies out at Cumshewa Inlet, where the fault intersects Mesozoic structures of the Long Inlet deformation zone. Geometric considerations dictate that the displacement documented in the southern islands be transferred into one of the blocks bound by the fault. In this case, it is accommodated by extension in the eastern domain, and the extension direction is roughly parallel to displacement direction on the LIFS. This model, schematically illustrated in Figure 2.44, allows for several tens of kilometres of transcurrent faulting in the southern Queen
Charlotte Islands, without offsetting structural trends in the central islands. The coincidence of transcurrent faulting with areas not significantly affected by Mesozoic deformation invites speculation that the two may be related; one possible connection is that the LIFS was mechanically restricted to weakened crust in the southern Queen Charlotte islands. This weakening may be an artifact of enhanced thermal gradients accompanying Tertiary plutonism in the area, or may be related to the lack of Mesozoic thickening.

Amounts of fault offset

The above model suggests that the LIFS should be considered a transfer fault separating extended rocks (eastern domain) from a relatively rigid block (western domain) rather than a regional strike-slip fault. One geometric consequence of such a fault is that amounts of displacement will vary along its length, from zero at the north end, to a maximum at southernmost exposures. Unfortunately, no clear markers exist on
which to base estimates of fault offset. Sutherland Brown (1968) suggested up to 100 km of movement, but he based his interpretations on offsets of poorly constrained structures and outcrop belts. Thompson et al. (1991), working near Louise Island, were unable to document any horizontally offset features along the northern extensions of the fault system.

The best estimate of fault offset in the Burnaby Island/Juan Perez Sound map area can be determined from the present mapping by measuring the amount of extension in the eastern domain, and assuming that all of this extension is transferred into dextral faulting in the central domain. Values obtained by this method will represent the difference in offset between the north and south boundaries of the map area, and can be treated as a minimum amounts of slip for the southern part of the fault system. Extension amounts would be best calculated from palinspastic restorations of cross sections drawn parallel to movement direction on the normal faults. However, these restorations are highly sensitive to structural geometry at depth; for example, the depth at which faults might sole into a sub-horizontal detachment surface. Given the high degree of geometric uncertainty in the present example, it is more useful to base extension estimates on simple calculations based on surficial geometry and inferred geometry at depth. These calculations can more appropriately show the range of likely amounts of extension than could restoration of a single interpretive cross section.

Thompson (1960), and more recently, Wernicke and Burchfiel (1982) show how amounts of extension accommodated by a set of rotating parallel normal faults will vary with original fault dip ($\phi$) and amount of tilting ($\theta$):

\[
\text{% extension} = 100\% \times \frac{\sin \phi}{\sin (\theta - \phi)} - 1 \quad (2.2)
\]
This formulation only holds if the following rigorous conditions are met:

1) Faults are planar,
2) Rotating blocks between the faults are undeformed,
3) All faults dip in the same direction and no antithetic faults are present, and
4) All extension occurs along a single set of faults.

With large amounts of extension, more than one generation of normal faults is commonly present. Equation 2.2 can be modified under these conditions:

\[
\text{% extension} = 100\% \times \left[ \frac{\sin(\theta)}{\sin\left(\frac{\phi}{n}\right)} \right]^n - 1 \tag{2.3}
\]

where \( n \) is the number of fault generations, each generation contributing equally to tilting of fault blocks. The failure to meet other assumed conditions is less easily quantified, but in general, will lead to underestimation of amounts of extension.

*amount of tilting (\( \theta \)):

If strata are flat-lying at the initiation of extension, bedding dips are the most reliable indicator of tilting due to block rotation. Rocks in the eastern domain of the Burnaby Island/Juan Perez Sound map area contain no significant structures older than the east-west-striking faults, and north-northwest-striking faults are interpreted as synchronous with or post-dating extension. Average bedding dips for all ages of strata in the domain are 20° to 40° to the north and northwest. Lesser amounts of tilting are indicated by paleomagnetic studies of dykes in the area, which suggest approximately 15° of northward tilting since Eocene emplacement.

The difference between amounts of tilting inferred from bedding dips and from paleomagnetic studies can be explained in several ways. Most likely, dyke emplacement occurred after some initial tilting, and only record the some of the total rotation.
Alternatively, if dykes were emplaced when the Queen Charlotte Islands were at a lower latitude relative to the craton, they will record apparent tilts less than actual tilts; but evidence for post-Eocene northward translation of the islands is lacking.

**original fault dip (Φ):**

Field studies in extended terranes show that normal faults commonly form with surface dips of 50° to 90°, and may shallow at depth. In the eastern domain, where rocks have undergone 30° to 40° of northward tilting, faults initiated at the onset of extension should have present-day dips of 20° to 60°. However, fault surfaces measured in the field dip 40° to 90°, with most dipping around 60°. Several factors may be responsible for this apparent discrepancy. 1) Faults may be steeper at the present levels of exposure than at depth. 2) Some of the observed faults may be antithetic, and have rotated to steeper dips during tilting. Several subvertical, north-side-down faults mapped in the eastern domain are probably antithetic. 3) Faults may have formed continuously throughout extension, and steeper faults therefore represent structures formed more recently.

Two end-member geometries illustrate minimum and maximum reasonable values for extension. The first case considers minimum likely extension amounts: faults are assumed to have formed nearly vertically (dipping 80° south) and to have tilted 15° to present dips of 65°. Equation 2.2 yields an extension value of just under 10% for these parameters. A second calculation treats maximum likely extensions: it assumes extension occurred on two generations of faults, each forming with a dip of 60° and accommodating 20° of tilting (more than two generations of faults would yield smaller amounts of extension, but are unlikely where total tilting is only 40°). For this geometry, equation (2) gives an extension of 82%! In both cases, the calculated values represent minimums, as the assumptions made in using the equations favour smaller values of
extension. In particular, listric fault geometries can accommodate considerably more extension than is apparent from surficial fault throw. White (1987) calculates the effects of hanging wall deformation above a listric sole fault, and finds large discrepancies between actual extension and extension estimated from fault throw. On the other hand, localized extension of 80% or more would cause significant crustal thinning, unless linked to more widely-distributed extension at lower crustal levels along a low-angle detachment (Royden and Keen, 1980). Low-level and upward-continued gravity data contain no anomalies in offshore areas along trend east of the map area which might indicate crustal thinning (Lyatsky, 1991a), which would appear to favour the more conservative values.

Rocks in the eastern domain affected by extension cover an area from at least southern Lyell Island in the north, to Carpenter Bay in the south, a cross-strike distance of 50 km. Extension estimates of 10% and 80% would yield elongations over this distance of 4.5 km and 22 km respectively. The geometric model presented above dictates that this extension is transformed into transcurrent faulting in the central domain. These values therefore are the difference in strike-slip offset between the northern and southern ends of the map area, and should be considered a minimum amount for total regional offset on the LIFS. Recent mapping (Thompson and Lewis, 1990a), reconnaissance studies, and Sutherland Brown's (1968) original mapping all show that the fault system extends north of the study area, probably as far as Cumshewa Inlet, and some strike-slip motion must occur north of the study area. Therefore, displacements for the southern part of the fault system will be greater than calculated above.
3. Strain measurement studies

3.1 Introduction

A key element of any structural analysis is the characterization of finite strain associated with deformation events. Previous sections made some initial attempts to estimate regional strain on the basis of mappable structural geometry or outcrop scale features, but these estimates only consider the strain components associated with observable, easily interpreted structures. In one case, more rigorous treatment of microscopic strain indicators (deformed fecal pellets at Louise Island) provided a more accurate determination of localized strains; this approach was useful in interpreting specific structural features but does not reflect values or directions of regional strain. Geologic strain in the Queen Charlotte Islands is highly heterogeneous, and includes components on macroscopic (map scale faults and folds), mesoscopic (outcrop-scale faults and folds) and microscopic (deformation through grain-scale mechanisms) scales, and these components must all be considered when interpreting regional deformation. Although the islands lack the levels of exposure required for rigorous strain analysis studies, they contain many strain indicators of use to these interpretations. Specifically, many of the Mesozoic units contain fossils which have been variably distorted, and give a measurable indication of the strain component accommodated through grain-scale mechanisms. Triassic and Jurassic ammonites of the Kunga, Maude, and Yakoun groups are the most useful in this respect; none of the fossils examined in younger units showed measurable distortion. In rare locations, segmented belemnites inn the Yakoun Group can be used.

Most Triassic and Jurassic ammonites have keel lines which outline logarithmic spirals when undeformed. The deformation of this logarithmic spiral can be quantified using several methods; the method described by Blake (1878) was used in this study. Blake's method requires visual estimation of elongation direction, and therefore loses
accuracy at low values of strain. Strains determined by this method are two-dimensional, reflecting only the deformation in the plane containing the keel spiral. In all samples measured, the ammonites were preserved as prints on bedding surfaces (Fig. 2.45a), and the values obtained are thus a measure of the two-dimensional bedding-plane finite strain. Ammonites preserved within concretions are undeformed and were not analyzed, as the competency contrast between concretions and surrounding rocks shields the fossils from deformation.

Belemnites are preserved as either empty molds or as calcite shells within sandstone and siltstone lithologies. The calcite shells were competent than surrounding rocks, and as a result, often failed brittlely during deformation (Fig. 2.45b). Strain magnitudes and directions were determined by measuring the amounts of elongation for belemnites of several different orientations, and applying a graphical Mohr circle solution (Ramsay and Huber, 1983). Ideally, a randomly oriented distribution of fossils will allow three-dimensional strains to be analyzed; in this case, belemnites are preserved along bedding planes, and only two-dimensional bedding plane strains were obtained.

Ammonites were collected and/or photographed at 22 locations, representing a wide geographic range across the islands. The fossils are from intervals within the Peril, Sandilands, Ghost Creek, and Phantom Creek formations, and the Yakoun Group. None of the fossils observed in Cretaceous strata in the course of the mapping studies displayed measurable distortion. For map presentation purposes, measured elongation directions were restored to horizontal by a single rotation about bedding strike. The strain analysis procedures, sample locations and results are tabulated in Appendix B.
**Figure 2.45:** Examples of deformed fossils used in strain analysis studies: a) distorted ammonite print: distortion of spiral can be measured to determine two-dimensional strain state; b) broken and stretched belemnite fragments: elongation of belemnites in several different orientations can be measured and used to graphically determine the strain state.
3.2 Results and Discussion

Samples from all locations showed restored elongation directions in the northwest quadrant (Fig. 2.46). In the central Queen Charlotte Islands, this elongation direction is approximately parallel to regional structural trends and perpendicular to shortening directions inferred from mapping studies. On central Graham Island, the two most northerly locations show a more westerly elongation directions than localities to the south. Structural trends in this area do not show a similar change in orientation (Hesthammer et al., 1991a), and the sampling density is too sparse to determine whether the deviation in strain directions is significant. Samples collected from within the LIFS had northwesterly elongation directions, inclined to the fault zone boundaries in a direction consistent with that expected in a dextral fault system.

Magnitudes of measured strain ratios $R \left(1+\varepsilon_1/1+\varepsilon_3\right)$ range from 1.0 (unstrained) to 1.72. These ratios correspond to shortening values of up to 24% assuming constant volume plane strain, or up to 70% if all shortening is accommodated through volume reduction. Strain ratios vary systematically between map units, and appear to reflect lithologic characteristics of the enclosing strata. For example, lowest strain values were obtained from well-bedded rocks with strongly-defined mechanical stratification (Sandilands and Peril formations). Lithologically homogeneous massive siltstones and sandstones of the Phantom Creek and Ghost Creek formations usually contain more highly strained fossils. Highest values of $R$ were obtained from fossils at the boundary between the Sadler Limestone and the Peril Formation within the LIFS. These high values reflect both the massive character of the older unit and the strain localization proximal to the fault system. Predictably, the trends in strain magnitude are the opposite to trends in intensity of mesoscopic structural development: well-bedded strata of the
Figure 2.46: Compilation map of bedding-plane strains in the Queen Charlotte Islands, as determined from analyses of distorted fossils in Jurassic strata. Most locations display northwesterly elongation directions; in northern areas these are related to Middle Jurassic deformation, on southern Moresby Island they are related to Tertiary deformation associated with movement on the LIFS. Strain ratios correlate reasonably well with lithologic unit: massive, relatively homogeneous rock types (Ghost Creek, Fannin, Phantom Creek formations) record greater strain ratios than well-stratified, lithologically variable units (Sandilands Formation).
Sandilands Formation typically contain the most significant outcrop-scale folds and faults, and the more lithologically homogeneous units lack these abundant features.

Finite strains measured from deformed fossils are a measure of local cumulative bedding-plane strain since the time of original deposition, and are not indicative of regional strain values. If regionally consistent, however, measured fossil strains give an indication of minimum values for regional strain, and can be considered in conjunction with heterogeneous strain accommodated by fault offsets and folding. Because the bedding-plane strains may be cumulative over several deformation events, it is useful to determine the likely contribution of these different events to the measured strain. In the central Queen Charlotte Islands, Jurassic and Triassic strata contain structures formed during two co-axial periods of shortening and several possible episodes of extensional or block faulting. Most mesoscopic structures can be related to either Middle Jurassic or Late Cretaceous/Early Tertiary shortening on the basis of style and orientation, but it is difficult to differentiate between the two events in older rocks. Deformation during Late Cretaceous/Early Tertiary shortening is highly heterogeneous on the map scale, and outside of zones of concentrated deformation (e.g., Long Inlet deformation zone, Yakoun Lake area) only broadly folds Cretaceous rocks. Most samples analyzed from the central Queen Charlotte Islands are from outside of these zones of concentrated deformation, and measured strains therefore probably reflect mostly the earlier event.

On the southern Queen Charlotte Islands, evidence for Mesozoic shortening is missing, and most macroscopic and mesoscopic structures are related to movement on the Louscoone Inlet fault system. Strains measured from rocks adjacent to the LIFS therefore are probably related to this same Tertiary episode of faulting.
4. Structural Synthesis: A model for the Mesozoic and Cenozoic evolution of the Queen Charlotte Islands

Many of the conclusions derived from the present studies contradict structural models for the Queen Charlotte Islands region which existed prior to 1987, and a complete re-evaluation of the area using the new constraints is timely. This section presents a new structural model for the Queen Charlotte Islands, based mainly on the mapping studies and structural analyses described in preceding sections. The model also draws on other mapping components of the Frontier Geoscience Program (Hickson, 1990a, 1990b; Hickson and Lewis, 1990; Lewis and Hickson, 1990; Lewis et al., 1990a; Thompson, 1990; Thompson and Lewis, 1990a, 1990b; Hesthammer et al., 1991a; Indrelid et al., 1991b; Taite, 1991b) and affiliated research projects which focussed on single elements of Queen Charlotte Islands geology (Anderson and Reichenbach, 1991; Hickson, 1991; Souther and Jessop, 1991). The discussion would be incomplete without addressing the history of the offshore basin areas; important data sources for this area include seismic reflection profiles (Rohr and Dietrich, 1990, 1991), potential-field data (Geological Survey of Canada, 1987a, b), and drill-hole data from offshore wells (Shouldice, 1971, 1973). Some elements of this model have appeared in recently published progress reports (Lewis and Ross, 1988, 1989, 1991; Thompson, 1988; Thompson and Thorkelson, 1989; Lewis, 1990, 1991; Lewis et al., 1990a, 1991a; Thompson et al., 1991), however, this is a first attempt to synthesize all existing field constraints into a coherent structural model for the Queen Charlotte Islands, and it contains many ideas not previously presented. The synthesis is presented as a chronologic progression, beginning in the Triassic and ending at the present. As an aid to the reader and a method of illustrating key geometric relationships, structural features associated with particular time periods are illustrated in a series of maps of the Queen Charlotte Islands (Figs. 2.47, 2.48, 2.49, 2.51, 2.53, 2.54). These maps show the islands
in their present configuration with no attempt to palinspastically restore horizontal fault movements; such restoration would only obfuscate the model presented. The model treats the Queen Charlotte Islands region in isolation from adjacent parts of the Cordillera. A more comprehensive discussion of relationships between geologic events in the islands and Cordilleran evolution follows in Part 4.

4.1 Pre-Middle Jurassic Deformation

With the recent discovery of Permian strata in the Queen Charlotte Islands (Hesthammer et al., 1991b), the opportunity to constrain Late Paleozoic and early Triassic deformation now exists. Initial examinations of the Permian strata at Englefield Bay (northwestern Moresby Island), and of possible correlative units at Hutton Point, just west of the Burnaby Island map area, reveals that these strata are moderately folded and faulted, and commonly contain a strong tectonic foliation. The timing of this deformation is not well constrained, and contact relationships with younger units are unclear. The lack of Lower and Middle Triassic strata on the islands suggests that the area was emergent in the early Mesozoic, but the structural significance of this unconformity is uncertain. This Late Paleozoic/Early Triassic unconformity is a feature common to all known Cordilleran occurrences of Wrangellia (Jones et al., 1977; Vallier, 1977; MacKevett, 1978; Massey and Friday, 1989; Hesthammer et al., 1991b).

The Upper Triassic through lower Middle Jurassic stratigraphic succession in the Queen Charlotte Islands contains no regional unconformities (Poulton et al., 1991; Tipper et al., 1991), and tectonism during this time period is unlikely.
4.2 Middle Jurassic shortening

In Middle Jurassic time, southwest-directed shortening created the earliest recognized structures in the Queen Charlotte Islands. The regional significance of this deformation is clearly demonstrated by the unconformity at the base of the Bajocian Yakoun Group, which truncates contractual faults and folds in underlying units. This unconformity is exposed at the Rennell Sound, Skidegate transect, and Louise Island map areas, and other workers have mapped it on central Graham Island (Hesthammer et al., 1991a), at Tasu Sound (Taite, 1991a, 1991b), and at Cumshewa Inlet (Thompson, 1990; Thompson and Lewis, 1990a).

Timing constraints on initial deformation include structural, stratigraphic, and intrusive relationships at several locations. The most spectacular example is the basal Yakoun Group unconformity exposed in Cumshewa Inlet near Dawson Cove (Fig. 2.22; Thompson and Thorkelson, 1989) This unconformity separates folded, steeply-dipping Sandilands Formation strata from overlying flat-lying rocks, but only limits the timing of deformation to post-Sinemurian and pre-Bajocian. Better timing constraints are available farther north—on Maude Island in Skidegate Inlet, the youngest unit below the unconformity is the Toarcian Whiteaves Formation (Jakobs, 1989), but on Graham Island, the Toarcian and Aalenian Phantom Creek Formation conformably succeeds the Whiteaves Formation. Therefore, the unconformity really only spans parts of the Aalenian and Bajocian stages (Fig. 1.3). Further confirmation of a Middle Jurassic age for deformation comes from a 168 ± 2 Ma U-Pb zircon age from a pluton which crosscuts southwest-verging folds in the Sandilands Formation at Rennell Sound (Anderson and Reichenbach, 1991). Thrust faults mapped cutting the Yakoun Group on central Graham Island (Lewis et al., 1990a) were suggested by Indrelid et al. (1991a) to indicate a continuation of contractual deformation during Yakoun Group time. However, more recent compilations (Hesthammer et al., 1991a) show that similarly
oriented thrust faults in that area cut Cretaceous strata. Therefore, the faults must either be re-activated following Cretaceous deposition, or have formed solely during the Late Cretaceous or Tertiary.

Middle Jurassic deformation strongly affected the central and northern Queen Charlotte Islands, but had little effect on the southern islands. The pre-Yakoun Group unconformity varies in character from a disconformity with little or no angular discordance in the southern Queen Charlotte Islands, to a sharp angular unconformity with near 90° truncations, as at Cumshewa Inlet. Rocks exposed below the unconformity contain strongly developed folds and faults in the northern and central islands, but to the south at the Burnaby Island map area are virtually undeformed. Figure 2.47 is a regional compilation map showing the distribution of Middle Jurassic macroscopic features. Southwest-directed thrusting is most common on northern Moresby and southern Graham islands. Pre-Middle Jurassic rocks are not exposed over most of Graham Island, but contractional structures on the northwest corner of the island suggest thrusting is regionally extensive. Structural trends swing from west-northwest on northern Moresby Island, to northwest on Graham Island. It is uncertain whether this swing reflects original structural trends or later modification. Structural trends are relatively chaotic on central Moresby Island, and early structures are limited to mesoscopic folds and faults, and rare northwest-trending folds (Taite, 1991b). Mesoscopic folds truncated by the basal Yakoun Group unconformity are evidence that Middle Jurassic deformation reached this area.
MIDDLE JURASSIC

- No pre-Middle Jurassic rocks exposed
- Map-scale thrust fault (teeth on upper plate)
- Map-scale fold axial surface trace (antiform, synform)

Southwest-directed shortening concentrated in central Queen Charlotte Islands region

Figure 2.47: Middle Jurassic structural elements of the Queen Charlotte Islands. Southwest-vergent folds and thrust faults in the central islands region are missing on southern Moresby Island, suggesting that a Middle Jurassic deformation front extends northwesterly across Moresby Island.

Dips equivalent in Triassic/Jurassic and Cretaceous units (Sutherland Brown, 1968)
The lessening of structural intensity to the south, together with the consistent southwesterly structural vergence, implies that northern Moresby Island was near the leading edge of a southwesterly migrating deformation front in Middle Jurassic time. This front likely extended northwesterly across the island, parallel to regional structural trends (Fig. 2.47). Future mapping of new areas on southern Moresby Island will help assess this proposal; present levels of mapping outside of the areas examined in recent studies (Sutherland Brown, 1968) are inconclusive.

Dominant structures formed during Middle Jurassic deformation are mesoscopic to macroscopic folds and contractional faults. Nearly all macroscopic features verge southwesterly, and compilations of bedding attitudes in pre-Middle Jurassic units show consistent preferred northeast tilts (Figs. 2.8a, 2.18a). Outcrop-scale structures are also mostly southwest verging, and the few northeast-vergent structures can usually be shown to be parasitic to larger scale features. Structural styles vary with both stratigraphic level and location. The Karmutsen Formation accommodated shortening by displacements along steeply-dipping reverse faults, which are usually only mappable where they offset the contact with the overlying Sadler Limestone. The Kunga Group deformed by broad folding and reverse faulting in the relatively massive Sadler Limestone, and by chevron folding and thrust faulting in the well-stratified Peril and Sandilands formations. At Rennell Sound, folding in these units is less prevalent than in other map areas, and shortening is accommodated by imbricate thrust faults. The lithologically variable Maude Group displays less strongly developed folds and faults, but deformed fossil prints indicate larger strains accommodated by grain-scale mechanisms.

Total amounts of shortening during Middle Jurassic deformation are difficult to estimate due to the incomplete exposure in the islands and the variety of structural styles associated with the same deformation episode. Three different approaches were discussed in previous sections: palinspastic restorations of fault offset (26% to 49% at
Rennell Sound, 25% at Louise Island); estimates from mesoscopic fold geometry (17% at Maude Island, Skidegate transect); and deformed fossil measurements (0% to 70% shortening depending on location, lithology, and volume change component). On all of these scales, shortening is oriented in a northeast-southwest direction. Because these different estimates consider strains accommodated on different scales, true regional strains will contain components of all three. However, shortening accommodated by mesoscopic structures is not additive with that accommodated by grain-scale mechanisms (and determined through measurements of deformed fossils) unless measured in the same outcrop, as the two mechanisms balance one another to some extent. A very approximate estimate for Middle Jurassic shortening in the central Queen Charlotte Islands would therefore be 50%, while the southern islands, in a foreland position, underwent negligible deformation.

4.3 Late Jurassic - Early Cretaceous Block Faulting

Late Jurassic to Early Cretaceous structural development of the Queen Charlotte Islands was dominated by vertical movement of large, fault-bound crustal blocks. This block faulting structural style (as defined by Hobbs et al., 1976) was first proposed for the islands by Thompson and Thorkelson (1989) on the basis of stratigraphic distribution trends, and has since been significantly revised as new field data have become available (Lewis and Ross, 1991; Taite, 1991a; Thompson et al., 1991). Some of these authors have suggested that block faulting occurred episodically through the Late Jurassic and Cretaceous; however, recent clarifications of several key biostratigraphic relations alleviate the need for all but a single period of block faulting.

Evidence for block faulting is based largely on patterns of stratigraphic distribution and inferences which can be drawn from structural styles, and only rarely are the actual block-bounding faults exposed. No Upper Jurassic strata occur in the Queen
Charlotte Islands, except for a single occurrence of Tithonian clastic rocks on northwest Graham Island (Jeletzky, 1984; Haggart, 1989; Gamba, 1991). The lack of strata of this age is consistent with much of the present islands region having been emergent prior to the onset of Cretaceous deposition. Rocks exposed below the Cretaceous unconformity range in age from the Triassic Karmutsen Formation to a Late Jurassic pluton of the Burnaby Island plutonic suite (Anderson and Greig, 1991). However, it is the distribution of Yakoun Group strata below the unconformity that provides the most compelling evidence for Late Jurassic block faulting. On the central Queen Charlotte Islands, northwest-trending belts with thick accumulations of Yakoun Group rocks alternate with belts with little or no age-equivalent strata (Fig. 2.48). Thompson and Lewis (1990a, 1990b) and Thompson (1990) show northwest-striking faults (e.g., Dawson Cove fault, Maude Island fault) separating these outcrop belts, and Thompson et al. (1991) suggest that dip-slip offset along these faults elevated those areas presently lacking Yakoun Group rocks, preferentially exposing them to erosion. The bounding faults are likely steeply-dipping planar features, but their dip direction, and therefore their type of motion (reverse or normal) is unknown. They cannot be clearly tied to any independently mapped extensional or contractional deformation, but may alternatively be related to regional uplift of the islands.

An alternative explanation for the present Yakoun Group distribution is that the outcrop belts reflect original depositional patterns and are unrelated to later block faulting. Several lines of evidence favour the tectonic explanation: Firstly, the outcrop belts are relatively linear and are parallel to regional and mesoscopic structural trends. Secondly, the thickness changes across faults is very rapid (up to 1000 m thickness differential over a few hundred metres), yet no facies changes are observed approaching the faults, such as one would expect for strata onlapping an escarpment. Thirdly, strain
LATE JURASSIC/EARLY CRETACEOUS BLOCK FAULTING

- Yakoun Group overlies basement rocks
- Cretaceous/Tertiary strata overlie basement rocks (no Yakoun Group preserved)
- Insufficient data (unconformity buried below Tertiary cover, or region of pre-Middle Jurassic exposure)

Possible Late Jurassic block faults (unmapped)

Late Jurassic (U-Pb = 158 Ma)
Poole Point pluton unroofed, non-conformably overlain by Hauterivian Longarm Formation strata.

Figure 2.48: Late Jurassic/Early Cretaceous structural elements of the Queen Charlotte Islands. Major features are northwest-striking block faults, separating elevated tectonic blocks (U) from depressed areas (D).
is localized in later deformation along the outcrop belt boundaries, reflecting possible re-activation of a crustal weakness zone formed by the bounding block fault.

Gamba (1991) notes that the single known occurrence of Late Jurassic strata, located on northwest Graham Island, contains sedimentary features consistent with rapid deposition in a fault-bound basin. This anomalous section may be representative of the type of deposition which must have existed adjacent to uplifted fault blocks in the Late Jurassic, but equivalent strata have yet to be documented elsewhere in the islands.

Thompson et al. (1991) propose that Cretaceous structural inversion of several Jurassic block faults on northern Moresby Island may have controlled stratigraphic distribution of Cretaceous units. They also note that the base of the Honna Formation, which often interfingers with the underlying Skidegate Formation, also locally lies unconformable on pre-Cretaceous units, suggesting block faulting prior to Honna Formation deposition. However, more recent biostratigraphic investigations in the vicinity of these faults on northern Moresby Island finds this evidence less compelling (Haggart, personal communication, 1991). Furthermore, it is unclear whether the local unconformity at the base of the Honna Formation represents areas where underlying Cretaceous rocks were stripped from the basement, or whether they represent local topographic highs within a shallow Cretaceous basin which were not submerged prior to deposition of the Honna Formation. In summary, the most convincing evidence for block faulting is restricted to the Late Jurassic, and a Cretaceous block faulting episode is not required to explain the present map distribution or structural features seen in the islands.
4.4 Late Cretaceous - Early Tertiary Shortening

A second major shortening event affected the Queen Charlotte Islands in Late Cretaceous to Tertiary time. In most areas, structures formed during this deformation event have northwest trends parallel to those related to Middle Jurassic deformation, and the two events cannot be differentiated in pre-Middle Jurassic rocks.

Structures formed during Late Cretaceous/Early Tertiary shortening include northwest-trending, mesoscopic and macroscopic folds and thrust faults (compiled in Fig. 2.49), and mesoscopic structural fabrics. In contrast to Middle Jurassic shortening, deformation is strongly localized. Most intense shortening is concentrated within the Long Inlet deformation zone (LIDZ), and is well documented within the Long Inlet map area, and to the southeast at Cumshewa Inlet and northern Louise Island (Thompson, 1990; Thompson and Lewis, 1990b). The LIDZ is characterized throughout its length by mesoscopic to macroscopic, open to tight folds of Cretaceous strata, and well developed minor features (axial planar cleavage, pencil lineation) related to the folding. Major macroscopic folds (Long Inlet anticline, Long Inlet syncline) are overturned, northeast-verging structures which, although cut by numerous faults, can be traced for over 30 km along trend. On northern Louise Island, at Kitson Point, these folds are broken by northeast-verging thrust faults which place Yakoun Group rocks over Longarm Formation strata.

A second area of concentrated strain occurs on Graham Island in the Yakoun Lake area (Hesthammer et al., 1991a; Indrelid, 1991). There, a northwest-trending belt of moderately to steeply dipping Cretaceous strata overlie similarly deformed Yakoun Group rocks, which are in turn thrust southwesterly over a second sequence of Cretaceous rocks (Fig. 2.49). Near Rennell Junction (north of Yakoun Lake), the upper plate rocks contain tight, northeast-verging mesoscopic folds with weak axial planar
Figure 2.49: Late Cretaceous/Early Tertiary structural elements of the Queen Charlotte Islands. Most deformation is localized in open to tight, northwest-trending folds along the Long Inlet deformation zone and near Yakoun Lake. No significant structures are documented on southern Moresby Island. Northeast-verging folds and thrust faults in the LIDZ may indicate structural inversion of west-dipping basement faults. In the central islands outside of concentrated strain zones, broad folds in Cretaceous strata are the only manifestations of shortening.
cleavage. However, the consistent northeast bedding and fault dips elsewhere in the region imply the bulk of the deformation is westerly directed.

Outside of these two areas of concentrated strain, Cretaceous strata contain broad, northwest-trending folds and few related mesoscopic structures. This less-deformed style is common both east and west of the LIDZ (Long Inlet and Skidegate transect map areas; Sewell Inlet area, Taite, 1991b; northeast Moresby Island, Thompson, 1990). Faults on central Graham Island which thrust Sandilands Formation strata southwesterly over Yakoun Group rocks may be deeper-level features associated with Late Cretaceous/Early Tertiary shortening, and may have formed as blind thrusts which cored the broad folds in overlying Cretaceous rocks. Cretaceous strata in the southern Queen Charlotte Islands contain no structures which can be shown to have formed during Cretaceous/Tertiary shortening, and similar to the Middle Jurassic deformation, this younger contraction is insignificant in that area.

Timing of deformation is bracketed by map relations within the LIDZ at Long Inlet, as outlined in previous sections. Northeast-verging folds in the Coniacian Honna Formation are overlapped by Early Eocene to Early Oligocene (White, 1990) shales, implying that folding occurred sometime between deposition of the two units. Amounts of shortening during folding are difficult to estimate, but fold geometry, as shown in cross section on map plate IV (Long Inlet) gives an approximation of regional strain. Outside of the high-strain area associated with the LIDZ fold profiles suggest less than 10% shortening normal to axial trends, while fold geometry within the LIDZ suggests over 20% localized shortening (Fig. 2.17). This latter estimate ignores strains accommodated through penetrative deformation, which may be considerable, given the strong mesoscopic fabric development in parts of the LIDZ.

The strain localization which characterizes Late Cretaceous/early Tertiary deformation, together with the structural styles exhibited in the high strain zones suggest
that deformation is strongly influenced by re-activation of older crustal features. A comparison of figures 2.48 (Jurassic block fault distribution) and 2.49 (Cretaceous structures) illustrates that areas of concentrated deformation are co-spatial with pre-existing block faults. Figure 2.50 is a schematic east-west section through the central Queen Charlotte Islands, showing how older faults might have controlled deformation patterns in the Cretaceous strata. It suggests that reactivation of basement faults, together with the sharp contrast in rheological properties between Cretaceous strata and their underlying basement, led to the complex structures present today. During structural inversion of basement faults, blocks of relatively competent basement material are forced upward into less competent Cretaceous strata, forming force folds, secondary faults, and localized shortening above the basement faults. The resulting geometry bears similarities to the well documented cuspate-lobate structural styles typically associated with contractional deformation of basement-cover interfaces (Gratier and Vialon, 1980; Ross et al., 1985).

*Figure 2.50:* Schematic east-west cross section through central Queen Charlotte Islands, showing how reactivation of Jurassic faults might control location and style of structures in Cretaceous cover.
Structural vergence within the high strain zones may be related to dip direction of underlying basement faults, although this proposition is difficult to test with the limited field exposure available. For example, the northeast-directed asymmetry of the LIDZ may imply a contractional reactivation of a southwest-dipping Dawson Cove Fault. This would be consistent with the fault forming originally as a normal fault, uplifting the eastern block to preferential erosion in the Late Jurassic. Similarly, a progressive thickening of Yakoun Group strata moving west to east across central Graham Island (Hesthammer et al., 1991a) may be related to Late Jurassic downdrop of fault blocks along northeast-dipping normal faults which, when re-activated, led to southwest verging structures in overlying strata.

The timing of tectonic foliation development in the Buck Channel pluton is poorly constrained, but it may be related to Late Cretaceous/Early Tertiary deformation. Sinistral offsets along subvertical, easterly-striking mylonitic zones are consistent with the northeast-southwest shortening at this time.

4.5 Tertiary Fault-Dominated Deformation

The Tertiary structural development of the Queen Charlotte Islands involves a complex history of faulting and related folding and interactions between structures of several different trends. Compressional, extensional, and strike-slip styles are all common to the Tertiary, and in places such as the Burnaby Island/Juan Perez Sound map area, interplays between these different styles create complex structural patterns. Several independent data sets help constrain Tertiary structural evolution, including seismic reflection (Rohr and Dietrich, 1990), potential-field (Geological Survey of Canada, 1987a,b) and well core (Shouldice, 1971, 1973) data in the offshore, and dyke orientation (Souther and Jessop, 1991), magmatic history (Anderson and Reichenbach, 1991; Hickson, 1991) and paleomagnetic (T. Irving personal communication, 1991) onshore.
In general, the offshore data suffer from either wide spacing (seismic reflection and drill-hole data) or lack of resolution (seismic reflection and potential-field data), but they are useful for testing hypothesized extensions of onland structural styles into the offshore. The following discussion outlines a structural model which attempts to accommodate all these data sets, but is biased towards and relies most heavily on constraints from surficial geology. The wealth of data available for the Tertiary Period dictates a relatively complex model, which for purposes of clarity is broken into three time intervals, beginning in the Eocene.

4.5.1 Middle Eocene to Early Oligocene (Approximately 50 Ma-35 Ma)

Structural, magmatic, and stratigraphic elements of the Middle Eocene to Early Oligocene Queen Charlotte Islands region are compiled in figure 2.51. Several independent lines of evident suggest moderate amounts of east-west extension during this time period. Radiometric dates of volcanic flows, dykes, and plutons all show that magmatism was concentrated in the southern islands region. Northerly striking Eocene dykes of the Carpenter Bay and Burnaby Island dyke swarms indicate up to 10% localized east-west crustal dilation, but normal faults related to this extension are difficult to document in the southern islands. Possibly the southern portion of the LIFS was active as an east-side-down normal fault, leading to the significant southwest tilt still preserved in the Karmutsen Formation west of the fault.

In the central Queen Charlotte Islands, localized nonmarine deposition commenced in early Eocene to early Oligocene time. The concentration of these deposits along the LIDZ, and the contemporaneous extension in the southern islands, suggests that the basins may have been fault controlled. Faults controlling sedimentation within the LIDZ are probably reactivated pre-existing structures. Northwest-striking extensional faults on central Graham Island (Hesthammer et al., 1991a), and north-striking faults on
Figure 2.51: Eocene to Early Oligocene structural elements of the Queen Charlotte Islands. Limited east-west extension in the southern and central islands is inferred from dyke trends, localized sedimentary basin formation, and mappable normal faults.
central Moresby Island (Taite, 1991b) may have also formed at this time, but they have no record of syn-tectonic deposition and their absolute timing of offset history is conjectural.

Seismic reflection surveys in offshore areas show that significant parts of the Tertiary sedimentary section accumulated in grabens and half-grabens, and sediment thicknesses are highly variable. Local unconformities and well core data (Higgs, 1991) suggest sedimentation in two stages, a Miocene and older syn-tectonic stage, followed by Miocene and Pliocene post-tectonic deposition. The existence of Eocene or Oligocene strata in the syn-tectonic sequence has not been confirmed, although hints of Paleocene palynomorphs have been noted in drill cores from Hecate Strait (J. White, personal communication, 1989). Sediment thickness maps of the offshore basin, based on detailed seismic surveys from the late 1960's (Fig. 2.52; Lewis et al., 1991a) show a prevalence of north-south striking faults in the central and southern Queen Charlotte Basin. These faults may have formed initially during the Eocene, in response to the same extension episode documented onshore.

Tertiary extension onshore, and possibly in the offshore basin, are roughly synchronous with the 43 Ma global plate reorganization, when relative plate motion along the Pacific/North America plate boundary changed from dominantly convergent to dominantly dextral strike-slip (Engebretson et al., 1985). Possible relationships between this global event, structures in the Queen Charlotte Islands, and other Cordilleran features are explored more fully in Part 4.
Basin, based on marine seismic reflection data (Modified from Lewis et al., 1991b). Dominant structural features present are northerly-striking faults in southern parts of the basin, and northwest- and east-northeast-striking faults and depocentres in central and northern parts of the basin.

Figure 2.52: Sediment thickness map for offshore parts of the Queen Charlotte Basin, based on marine seismic reflection data (Modified from Lewis et al., 1991b). Dominant structural features present are northerly-striking faults in southern parts of the basin, and northwest- and east-northeast-striking faults and depocentres in central and northern parts of the basin.
4.5.2 Early Oligocene to Early Miocene (Approximately 35Ma-20Ma)

The extension-dominated structural styles of the Eocene and early Oligocene were superseded in the Middle Oligocene by a more complex structural setting involving continued extension, strike-slip faulting, and localized shortening (Fig. 2.53). Magmatic activity shifted northward during this time period in two steps (Anderson and Reichenbach, 1989, 1991)—plutonism was concentrated in the central islands by around 35 Ma, and on Graham Island by 25 Ma. Dominant dyke trends in these areas are east-northeast for both the Rennell Sound and Selwyn Inlet swarms (Souther, 1989; Souther and Jessop, 1991); although no radiometric ages yet exist for these dykes, they may track the northerly younging plutonism. If so, they suggest a change to a more northerly extension direction in the Oligocene. Strike-slip faulting along the Louscoone Inlet fault system on Moresby Island also began during this same time interval, and was accompanied by north-south extension and northward tilting of rocks east of the fault. This structural transition is independently confirmed by paleomagnetic studies of the dyke swarms: dykes sampled at Carpenter Bay, Burnaby Island, Selwyn Inlet, and Louise Island all show a systematic deviation of 11° to 17° from the expected pole position (T. Irving, personal communication, 1991). This deviation can be accounted for by large southward translation of the Queen Charlotte Islands, or by northward tilting of the dykes following emplacement. Dykes from the Tasu Sound swarm, west of the the northern extension of the LIFS, show similar northward tilts, and Taite (1991a) argues for Tertiary tilting there on the basis of Tertiary contact trends. This area is northwest of the most significant portion of the LIFS, and its structural history is probably not closely related to offset on the fault system.
Figure 2.53: Major Oligocene to Early Miocene structural elements, Queen Charlotte Islands. Dextral strike-slip movement on the Louscoone Inlet fault system dominated deformation in the southern islands, and may have been linked to postulated dextral movement on the Sandspit Fault. See text for more complete discussion.
On southern Kunghit Island, east of the LIFS, Sutherland Brown's (1968) mapping shows a change to southerly dips in Triassic Karmutsen Formation strata. This area has not been re-examined since this initial mapping, but the geometry shown suggests a change in asymmetry of extension in this area, to extension on north-dipping faults accompanied by southward block tilting.

Earliest movement on the Sandspit Fault may have occurred during this same time period. The present position and orientation of the fault suggest that initial offset may have been linked to that on the LIFS, as an en-echelon, dextral slip feature. This connection is difficult to test due to limited field exposure, but several field relationships are compelling: Firstly, the Sandspit Fault loses definition to the south at about the same latitude that offset on the LIFS dies out to the north. Secondly, the two faults have a similar orientation, and both have thick accumulations of Tertiary sediments to the east. Structural features at Cinola gold camp (Richards et al., 1978) and Copper Bay suggest a component of dextral strike-slip movement along the Sandspit Fault (section 2.2.5), despite the thick accumulation of strata east of the fault which seems to imply east-side-down movement. The Sandspit Fault may have a structural style analogous to that of the LIFS: both systems form transfer zones dividing regions with differential amounts of north-south extension. The en-echelon geometry suggests that the area just south of Louise Island would be the linkage zone between the two fault systems, and should contain extensional structures related to the right-stepping transfer in offset. One geologic peculiarity to this area, consistent with localized extension, is the great thickness of Tertiary volcanic rocks exposed on Talunkwan Island, and the large volume of subvolcanic feeders on southern Louise Island.

The geometry of youngest faults on Graham Island is consistent with formation during this same period of deformation. East-northeast striking, south-side-down normal faults at central Graham Island suggest similar, but lesser amounts of extension and
northward tilting as that documented on Moresby Island east of the LIFS. At Rennell Sound, youngest structures are easterly-striking sinistral strike-slip or south-side-down (oblique slip?) normal faults. Northwest-trending folds and reverse faults in the Paleogene strata at Long Inlet are transverse to the extensional features, must pre-date Miocene volcanism, and therefore may also have formed during this time interval.

In the offshore, thick accumulations of syn-tectonic sediments signify extension-related subsidence during the Miocene. This offshore extension may be tied to northward translation of much of the present-day islands along the combined Sandspit-Louscoone Inlet transform fault system. Although most extension would be localized east of this fault system, lesser amounts west of the system, on Graham Island, could have led to east-northeast striking extensional features noted above. Minor crustal thinning beneath Graham Island may have contributed to crustal melting, leading to Masset Formation calc-alkaline volcanism in the area (Hickson, personal communication, 1991). The proposed mechanism for generating extension in the offshore differs from conventional strike-slip pull-apart basin models in lacking an eastern, basin-bounding transform fault. Instead, extension is considered a product of distributed shear across the Queen Charlotte Islands region, with movement concentrated along transfer faults such as the Louscoone Inlet and Sandspit systems, and possible analogues offshore.

Sediment thickness maps of offshore areas show a variety of fault and sub-basin trends, and cannot be used as conclusive support for this or other models. One can speculate that northeast-trending (e.g., Moresby Ridge), and northwest-trending structures are extension and transform features respectively, but a complete analysis of basin structure is necessary to evaluate this possibility and is beyond the scope of the present study.
Figure 2.54: Major Early Miocene to Recent structural elements of the Queen Charlotte Islands.
4.5.3 Early Miocene to Recent (< 20 Ma)

The post-20 Ma structural history for the Queen Charlotte region is difficult to constrain onland because of a lack of young strata. Youngest major features are northwest-striking normal faults mapped within the LIDZ and on northeast Moresby Island (Thompson, 1990; Thompson and Lewis, 1990b) (Fig. 2.54). North-south striking dip-slip faults at Sewell Inlet and Tasu Sound (Taite, 1991a) may also have been re-activated during this period of deformation. Hickson (1991) notes that little deformation has affected the Masset Formation on Graham Island, suggesting that onland areas were relatively stable during the Pliocene.

Offshore, a seismically-imaged Miocene and Pliocene sedimentary record provides valuable constraints on deformation. Several general features are notable in the seismic data. A Late Miocene unconformity dividing syn-tectonic and post-tectonic sedimentary successions (Fig. 2.55) indicates that by the Late Miocene, much of the faulting associated with basin formation had ceased. Youngest features imaged offshore are structural inversions of northwest-striking normal faults, most common in Hecate Strait (Fig. 2.56; Rohr and Dietrich, 1991). Rohr and Dietrich (1991) attribute these features to Pliocene plate convergence, which began at roughly 3.0—4.0 Ma (Harbert, 1991; Harbert and Cox, 1989).
Figure 2.55: Seismic reflection profile in Queen Charlotte Sound, showing unconformity between syn-tectonic (a) and post-tectonic (b) basin fill (line position indicated on figure 2.54).

Figure 2.56: Seismic reflection profile in Hecate Strait showing Plio-Pleistocene structural inversion of graben-bounding normal fault (line position indicated on figure 2.54).
4.6 Implications to Tertiary Queen Charlotte Basin Evolution

The structural model presented above for the Tertiary evolution of the Queen Charlotte Islands region represents a first step toward re-evaluating Queen Charlotte Basin evolution using all available constraints. It is more complete than other recent geologic syntheses (Lewis et al., 1991a; Thompson et al., 1991) in that it recognizes the significance of strike-slip and extensional faulting styles on land, and proposes a scenario in which similar structural styles may be important to offshore basin evolution. Likewise, because it is based primarily on observable, mapped structural features onland, it is more tightly constrained than largely geophysically-based summaries (Hyndman and Hamilton, 1991; Lyatsky, 1991a; Dietrich and Rohr, 1991).

The three-part model detailed above underscores the structural complexity of the Tertiary history of the Queen Charlotte region. This same complexity is mirrored in seismic reflection records of offshore areas, which are characterized by laterally discontinuous unconformities, structural inversions of basin-boundary features, and a general lack of continuous reflectors in all but the youngest basin fill. Some of the structural features discussed in this synthesis, if taken by themselves, offer inconclusive evidence for the timing or mechanism of formation. This is especially true of the offshore seismic reflection data, where limited well control, wide line spacing, almost no penetration below Tertiary basement, and in places, probable acoustic basement within the Tertiary basin fill (Lyatsky, 1991b) all hamper structural interpretation. Nonetheless, the model satisfies all well-constrained data sets, and at least provides an internally consistent mode of formation for the less constrained features.

The Tertiary Queen Charlotte Basin formed during all three time divisions of the Tertiary outlined above. The general model above proposes that earliest basin formation was related to small amounts of east-west extension, which may have been localized in southern (Queen Charlotte Sound) areas. However, most syn-tectonic sedimentation
probably accompanied northwest-southeast basin wide extension, which was in turn linked to distributed shear along a series of north-northwest striking transform faults, analogous to the Louscoone Inlet fault system of the southern Queen Charlotte Islands. A quantitative basin analysis tying this tectonic history with subsidence rates and thermal history was not attempted in this study, nor was a detailed structural analysis of features imaged in the offshore seismic reflection data. However, both of these future studies will profit greatly from having the above model available as a starting point for calibration with onland geologic history.
Part 3:

Processes of Deformation in Mesozoic and Cenozoic rocks
of the Queen Charlotte Islands

1. General Statement

This section discusses in more detail the processes of strain accommodation active during the structural history outlined in Part 2, and examines some of the relationships between lithology, inferred pressure/temperature conditions, and rock behaviour.

Several summary statements on the nature of deformation in the Queen Charlotte Islands can be made from the discussion in Part 2:

1) Strain is heterogeneous on the mesoscopic to macroscopic scales throughout the history of Mesozoic and Cenozoic deformation.

2) Largest regional strains were accommodated through mesoscopic to map-scale folding and faulting.

3) Many massive lithologic units experienced small to moderate amounts of penetrative deformation accommodated through grain-scale mechanisms during regional contractional deformation. Evidence for distributed deformation in these units includes distorted fossils, cleavage development, and pressure solution fabrics.

4) Most penetrative tectonic fabrics are spatially associated with mappable structural features, and microscopic fabric geometry provides a valuable aid in interpreting larger scale features.
5) Ubiquitous, multiple generations of calcite and quartz-filled extension veins in nearly all rocks attest to the importance of fluid-aided extensional crack growth during deformation.

The following discussion uses the above summary as a starting point for a more detailed examination of deformation processes. It briefly addresses the topics of mesoscopic and macroscopic folding and faulting, and discusses more fully the processes leading to penetrative fabric development. Low metamorphic grade rocks in the Queen Charlotte Islands contain a variety of such fabrics, and the determination of their formative mechanisms may improve our understanding of the early stages of fabric development.
2. Mesoscopic to macroscopic structures

2.1 Folding

The greatest component of regional tectonic strain in the Queen Charlotte Islands was accommodated throughout deformation by folding and faulting, the products of which are widely visible on both mesoscopic and macroscopic scales. Most lithologic units contain folds of variable geometry and style, the rare exceptions being the massive lithologies of the Karmutsen Formation and Tertiary volcanic rocks. Most folds formed during two periods of contractional deformation: In upper Middle Jurassic and Cretaceous rocks, most folding occurred during Late Cretaceous/Early Tertiary, northeast-southwest shortening. In older strata, earliest folds formed during Middle Jurassic southwest-directed shortening, and were subsequently modified during the younger, nearly co-axial event. The relative contribution of each deformation event to the final structural form is not usually factorable in these older rocks, due to the parallelism of shortening directions. Localized folds in strata adjacent to major faults are interpreted to have formed as a result of movement along the faults during regional strike-slip or extensional faulting events.

Fold styles are in general agreement with styles predicted by lithologic characteristics of the deformed units. Most mesoscopic folds have parallel profiles with little or no variation in bed thickness between hinges and limbs. Exceptions include mesoscopic folds in Cretaceous mudstones (Haida and Skidegate formations), and some of the less competent layers in the thinly-bedded Sandilands and Peril formations; folds in these rocks have moderately greater bed thicknesses in hinges (class 1c profiles; Ramsay, 1967) and weakly to moderately well developed axial planar cleavage. The parallel fold styles are consistent with formation through layer buckling and flexural slip and flexural flow processes, with superimposed homogeneous shortening in those folds containing thickened hinges. Minor fold-related features include bedding-plane
slickensides perpendicular to axial trends, hinge collapse zones, and faulted limbs (Fig. 3.1a,b); the latter features typically form in response to "room problems" in flexural folds.

The geometry and scale of folds in different lithologic units correlate well with general material properties of the host rocks. Mechanical theory predicts that in a multilayered sequence, layer-parallel shortening will lead to buckle instabilities, the initial spacing of which will be governed by layer thickness and by viscosity contrast between layers (Biot, 1961; Ramberg, 1960; Ramsay and Huber, 1987). Fold wavelengths observed in the Queen Charlotte Islands are in general agreement with these qualitative predictions: mesoscopic folds are limited to the thinly-bedded Sandilands, Peril, and Skidegate formations and parts of the Maude Group, while the thickly-bedded to massive units (e.g., Honna Formation, Sadler Limestone, parts of the Yakoun Group) are typified by broad macroscopic folds. Several workers have presented theoretical quantitative expressions relating material properties to fold geometry (e.g., Ramberg, 1964, 1968, 1970); these expressions are useful for analyzing fold geometry under rigidly controlled laboratory conditions with rock analogues, but are not easily applied to the more variable material properties and environmental conditions associated with natural deformation. However, some of the general trends predicted can be qualitatively applied to mesoscopic folds in the Sandilands and Peril formations. Based on Ramberg's (1968, 1970) analyses, Ramsay and Huber (1987) show how fold geometry varies in regularly alternating multilayered sequences as a function of layer thickness and viscosity contrast between layers. The Sandilands and Peril formations provide an illustrative example of these relationships: both units are characterized by competent sandstone, siltstone, or limestone layers 2–20 cm thick, separated by either bedding-plane parting surfaces or by thin (generally < 2 cm thick) less competent mudstone or tuffaceous
Figure 3.1: Typical secondary features associated with mesoscopic folds in the Sandilands and Peril formations: a) hinge-directed thrust fault cuts layering along fold limb; b) collapsed hinge zone creates substantial hinge thickening in less competent layer (below lens cap).
layers. These material characteristics—high competency contrast and moderate to high ratios of competent to incompetent layer thickness—generally lead to the formation of chevron folds with little or no layer-parallel shortening of the competent layers. Layer-parallel slip is concentrated along layer boundaries and within less competent layers, which as a result occasionally contain local cleavage or hinge-zone thickening (class 2-3 profiles). Forces required for buckle initiation do not vary greatly for different fold wavelengths and, as a result, a wide range of fold wavelength to layer thickness ratios is common. Variations in competent layer thickness lead to room problems as shortening progresses, which are accommodated by limb thrusting or hinge collapse.

Descriptions in Part 2 noted that major folds in Cretaceous units are concentrated along re-activated, pre-existing structural features, and interpreted their geometry to be largely controlled by the older structures. This strain localization, together with the generally more homogeneous material properties of the Cretaceous mudstone units, contributed to the formation of flattening fabrics (axial planar cleavage) in these units. The strain state and formative mechanisms associated with this cleavage are analyzed in more detail in following sections.

2.2 Faulting and fracturing

Faulting occurred on all scales during all deformation events. During periods dominated by extensional or strike-slip deformation styles, steeply-dipping faults were active in all map units. Middle Jurassic and Late Cretaceous/Early Tertiary episodes of shortening contributed to more complex fault patterns; the massive Karmutsen Formation, because of its great thickness and lack of internal stratification, resisted buckling and deformed largely through loss of cohesion and translation (brittle response) along steeply-dipping fault surfaces. These faults cut upward into overlying bedded strata, where displacement gradually dissipated along splays which root low-angle slip
surfaces and folds in the Peril and Sandilands formations. Thus, the lower parts of the Kunga Group formed a broad zone of decoupling between fault-dominated styles in the underlying Karmutsen Formation, and mesoscopic to macroscopic folding in younger rocks. The nature of these faults at depth (below the Karmutsen Formation) is uncertain; they may flatten with depth into the well-bedded rocks of the Late Paleozoic sedimentary successions, where shortening could have been taken up by displacement along bedding-parallel detachments and through folding.

Classical fault mechanics models (Anderson, 1942) predict that at shallow depths in mechanically homogeneous upper crust, faults will form inclined to the maximum principal stress direction by roughly 30°. This model dictates that at the earth's surface, which is a free surface and a principal plane of stress, normal faults should dip 60° and strike-slip faults should dip 90°. Variations in fault orientations will result from rock inhomogeneity and changes in stress orientation with depth (e.g., Yin, 1989). Allowing for these uncertainties, and for post-faulting rotation of structures, most faults analyzed in the present study are well within the range of orientations predicted by classical fault theory. A more detailed analysis of fault formation processes is not considered appropriate, given the poor exposure and resultant uncertainty of fault geometry in the Queen Charlotte Islands.

A second type of mesoscopic brittle response is represented by calcite-filled and quartz-filled extensional veins, common in all lithologic units in the Queen Charlotte Islands, and less common composite veins with epidote, chlorite, and calcite fillings, limited to the Karmutsen Formation. In most locations, veins occur in a broad range of orientations and show contradictory cross-cutting relationships, hampering the utility of veins for kinematic interpretations (notable exceptions are the Sandspit Fault splay and the Section Cove shear zone). Variable cross-cutting relationships with folds and faults indicate that extensional vein formation was active during all stages of deformation.
Both syntaxial and antitaxial veins occur in most rocks, with antitaxial veins being most common. Typical antitaxial vein geometry consists of fibrous calcite and quartz crystals oriented perpendicular to vein walls. Internal banding parallel to fracture walls is defined by wall rock fragments or compositional changes within the vein-filling material. These characteristics are consistent with vein formation during cyclic hydraulic fracturing, through the crack-seal mechanism (Durney and Ramsay, 1975; Ramsay, 1980). The crack-seal model dictates that veins form in extensional fractures as values of effective stress became equal to or greater than rock tensile strength. Laboratory measurements give a rough approximation of stress states favouring extensional crack growth: typical tensile rock strengths fall below 30 MPa in ambient conditions, and may be lower at shallow crustal levels (Lama and Vutukuri, 1978). Furthermore, sub-critical crack growth may proceed at stress values of only 50% of the tensile rock strength, resulting in fractures indistinguishable from those associated with catastrophic failure (Atkinson, 1980). Mechanical considerations require low values of differential stress to form extensional fractures (generally less than four times tensile rock strength; Etheridge, 1983); high values will favour propagation of conjugate shear fractures. The ubiquitous occurrence of syntectonic extensional veins in the Queen Charlotte Islands therefore implies that deformation occurred at high values of pore fluid pressure and low to moderate differential stresses. Pore fluid pressures must have been sufficient to overcome the least principal stress in addition to tensile rock strength. Classical interpretations for developing such pressures rely on some combination of rapid burial rates, deformation-induced pore volume and permeability reduction, aquathermal pressuring (Gretener, 1983), and diagenetic or metamorphic dewatering, all of which are feasible in Queen Charlotte Islands examples.

Veins worthy of special consideration are those oriented perpendicular to bedding, most common in the Sandilands and Ghost Creek formations (Fig. 3.2). Offset
of these veins by slip along bedding surfaces is interpreted to have occurred during Middle Jurassic flexural slip folding, which implies that the veins formed either during compaction, or during an undocumented pre-Middle Jurassic deformation. An unrecognized tectonic event is unlikely, because even in small areas, bedding-perpendicular veins occur with a wide variety of strike orientations.

![Image](image.png)

**Figure 3.2** Bedding-perpendicular, calcite-filled extensional fracture in Ghost Creek Formation, Maude Island. Vein is offset along bedding surface by slip interpreted to have taken place during Middle Jurassic folding.

The presence of large amounts of vein-filling material implies that large volumes of minerals were transported through the surrounding rocks, by a combination of diffusion and advective transport. Without completing isotopic analyses of vein material, it is difficult to speculate on the origins of circulating fluids, but contributions from connate water, circulating meteoric waters, and diagenetic and metamorphic dewatering are all likely. Similarly, sources for the vein-filling minerals are uncertain, but the close association of veins with pressure solution fabrics (Fig. 3.3) suggests that
dissolution played a strong role. The widespread abundance of antitaxial veins is consistent with advective transport of fluids in an open system.

Figure 3.3: Photograph of bedding surface showing stylolites and perpendicular veins, Peril Formation, northern Moresby Island. The nearly perpendicular orientation of the features, and variable cross-cutting relationships implies that the veins act as "sinks" for material removed along solution surfaces.
3. Penetrative fabrics

Although deformation is largely heterogeneous on the map scale, penetrative fabrics attest to local areas of homogeneous deformation. These penetrative fabrics take many forms, but nearly all developed through semi-brittle mechanisms of crystal plasticity, microfracturing, mechanical rotation, fluid-aided diffusion, and limited dynamic recrystallization and dynamic recovery. This section largely discuss how strain is partitioned between and accommodated by these processes. It focusses on three geometrically and genetically distinct types of penetrative fabrics found in Mesozoic rocks of the Queen Charlotte Islands: a) mylonitic and fault fabrics; b) slaty cleavage in Cretaceous rocks, and c) spaced cleavage in Cretaceous mudstones. Field relations (structural style, orientation, and location) unequivocally demonstrate that the first two fabric types are associated with regional deformation events. The spaced cleavage fabrics are of less certain origin, and occur only in relatively undeformed regions.

3.1 Mylonitic, Fault fabrics

3.1.1 Buck Channel Pluton

The Late Jurassic(?) Buck Channel pluton, located on northwest Moresby Island just south of Chaatl Island, contains the only known example in the Queen Charlotte Islands of a well-developed tectonic foliation within a pluton. Penetrative fabrics are common in other Middle and Late Jurassic plutons, especially those of the San Christoval plutonic suite, (Anderson and Greig, 1989), but their origin has been ascribed to igneous processes. The foliation in the Buck Channel pluton, however, contains features usually considered characteristic of syn- to post-emplacement tectonic fabrics (Paterson et al., 1989): it is continuous across intrusive contacts into adjacent country rocks (Karmutsen Formation), it is defined by microfabrics formed during solid-state deformation, and it is in part defined by secondary minerals (epidote).
The geometric description and the kinematic interpretation of foliation in the Buck Channel pluton are discussed in Part 2, and will not be repeated here. For review, the fabrics are interpreted to have formed as part of a west-northwest trending, sub-vertical sinistral shear zone, active in Late Cretaceous or Early Tertiary time. More exact timing constraints on formation will require isotopic analysis of the pluton.

A broad range of microscopic textures is preserved in the pluton fabrics. Most of the Buck Channel pluton is compositionally heterogeneous, and consists of foliated quartz diorite to granodiorite. The dominant mineral constituents are abundant feldspar, quartz, mica, and hornblende, with lesser sphene and secondary epidote and chlorite. Fabrics present range from a weak foliation defined by parallel preferred orientation of quartz ribbon grains, to highly-banded tectonites, to quartz mylonites. The banded rocks are volumetrically most abundant, and are characterized by segregation of most of the quartz component into bands 0.1 mm to 2 mm thick (Fig. 3.4a). Kink bands are common, with the kink-band boundaries oriented at high angles to the foliation (Fig. 2.27a). Within many of the quartz-rich domains, highly-elongate (up to 15:1 aspect ratios) ribbon grains are internally divided into subgrains ranging in size from 50 µm to 100 µm. Two types of subgrain boundaries occur: planar subgrain walls divide subgrains with mismatches in crystallographic orientation typically less than 5° (measured in plane section), and irregular, sinuous, boundaries separate grains with considerably different crystallographic orientations (Fig. 3.4b). Many quartz ribbon grains are transected by planar arrays of sub-micrometre sized bubbles. These bubble trails are oriented at high angles to the mesoscopic foliation, and cross subgrain boundaries without apparent offset or change in orientation (Fig. 3.4c).

Quartz-poor layers in these same rocks comprise mainly plagioclase and alkali feldspar, and less common accessory phases. Internal textural features in the feldspar are almost completely obscured by the high degree of alteration. Feldspar crystals have slightly elongate outlines (generally < 2:1) with long dimensions parallel to the
mesoscopic foliation. In addition, quartz + epidote, chlorite filled cracks 10 \( \mu \text{m} \)-20 \( \mu \text{m} \)
wide cut the feldspar-rich domains at moderate to high angles to foliation (Fig. 3.4a). These cracks are often parallel to bubble trails in the quartz-rich domains.

Weakly-foliated rocks lack well-defined mineral segregation layering, but do contain distributed quartz-rich lenses. These lenses have many of the same features found in the quartz-rich layers described above, but the degree of subgrain development within is less complete. Many of the lenses contain internally strained grains divided into elongate subgrains, and deformation lamellae at high angles to subgrain boundaries. In some grains, migration of subgrain boundaries has pinched off new recrystallized grain areas.

Highly-banded quartz mylonites probably represent the most highly strained rocks, and form narrow zones (< 5 m wide) within less deformed plutonic rocks. These mylonites are composed of 95% subequant to moderately elongate (<3:1 aspect ratios) recrystallized quartz grains with longest dimensions ranging from 15 \( \mu \text{m} \) to 100 \( \mu \text{m} \) (Fig. 2.28b). Quartz grains display a preferred dimensional orientation inclined 20° to 30° to the mesoscopic foliation plane, which is defined by chloritized micaceous folia. Recrystallized quartz grains lack deformation lamellae, undulose extinction, and internal microcracks, and are therefore largely strain free. Rare, highly-elongate quartz ribbon grains up to 2 mm in length contain an internal subgrain structure with subgrain walls parallel to the external dimensional quartz fabric (Fig. 3.4d). Weak deformation lamellae defined by birefringent banding are visible in the ribbon grains, and are oriented at high angles to subgrain boundaries.

Equant feldspar porphyroblasts up to 2 mm in diameter are scattered within the quartz fabric. Micaceous folia wrap around the porphyroblasts, which are moderately to
Figure 3.4: Microscopic deformation features characteristic of foliated parts of the Buck Channel pluton: a.) layering defined by mineralogical segregations, alternating between quartz-rich layers and quartz-poor layers; epidote-quartz-filled extensional fracture cuts all layers; b) sutured subgrain boundaries undergoing grain-boundary migration; c.) linear bubble trails crossing subgrain boundaries in quartz ribbon grain; d.) recrystallized quartz fabric, showing quartz subgrain formation (polygonization) in non-recrystallized ribbon grains. Field of view in all photographs is 1 mm, left to right.
DEFORMATION PROCESSES \\ *penetrative fabrics* 209

highly altered to very fine grained phyllosilicates. Feldspar porphyroblasts contain no microscopically visible internal deformation features.

Discussion:

Microfabrics in the Buck Channel pluton are consistent with post-emplacement solid-state deformation, in accord with map relationships. Textures preserved in most parts of the pluton indicate that deformation was accommodated by semi-brittle creep (e.g., Kirby, 1983; Ross and Lewis, 1989)—the combination of dislocation creep, solid state and possibly fluid-aided diffusion (grain-boundary diffusion and pressure solution creep), and microcracking. Subgrain formation, and recrystallization by grain boundary migration (bulge nucleation along sutured grain boundaries) indicate that both dynamic recovery and dynamic recrystallization were active as recovery mechanisms in quartz. Products of dynamic recrystallization occur to only a limited extent in those rocks with weak mineralogical segregation, but are ubiquitous in the more highly banded rocks, while subgrain development occurs in all rocks. This suggests that recovery is initially dominated by polygonization, but undergoes a transition to dynamic recrystallization. This transition may be related to either lower temperatures during waning stages of deformation, or larger amounts of strain concentrated in the more quartz-rich rocks.

Feldspar crystals in the quartz-poor domains contain none of the typical dislocation creep microstructures. The dimensional orientation in these domains is difficult to interpret, due to the alteration obscuring crystal boundaries. Tullis and Yund (1987) show that in laboratory deformation experiments at high strain rates and confining pressure, feldspar fabrics with a strong parallel preferred orientation can form during both cataclastic flow and dislocation creep, and point out that the two mechanisms can be virtually impossible to distinguish between using conventional optical methods.

Brittle deformation processes are implicated by both the planar bubble arrays in quartz, and the mineral-filled extensional cracks in feldspar domains. Planar bubble
arrays are usually interpreted as partially healed dilatent microcracks (Wanamaker and Evans, 1985) which form as fluid-filled inclusions in response to lattice mismatch during diffusion-aided crack healing. Similarly, quartz ± chlorite and epidote-filled extensional cracks in feldspar are further examples of dilatency during deformation. Partially healed microcrack arrays which transect subgrain boundaries, as well as the preservation of bubbles within strain-free grains, imply that brittle failure occurred late in, or postdated, diffusion-biased creep processes.

Semi-brittle creep occurs over a wide range of environmental conditions (Ross and Lewis, 1989), and, by itself, places little constraint on pressures and temperatures during deformation. Evans (1988) found that deformation in granitic rocks at low temperatures and high strain rates is dominated by cataclasism in feldspar and fracture in quartz, and plasticity in quartz is inferred to be active at mid-crustal levels (e.g., Carter and Tsenn, 1987). The transition from dominantly brittle behaviour to dislocation creep in rocks is dependent on the ease with which dislocations can migrate freely through the crystal structure, which is closely correlated with temperature. In general, creep by diffusion becomes the rate-controlling mechanism of deformation at approximately half the melting temperature of a particular mineral (e.g., Carter, 1976). However, creep can operate at lower temperatures, though not necessarily be the dominant mechanism, and will certainly be facilitated by the presence of fluids and low strain rates. Recovery-biased, dislocation creep microstructures in the Buck Channel pluton imply that deformation occurred at moderate temperatures, but these thermal conditions may have been related to deformation during cooling, rather than a regional geothermal gradient.
2.1.2 Louise Island, Rennell Sound: Louscoone Inlet fault fabrics

Dip-slip faults at Rennell Sound and Louise Island, and major splays of the Louscoone Inlet strike-slip fault system, are all characterized by varying degrees of tectonic foliation development in adjacent wall rocks. These foliations contain microscopic and mesoscopic textures indicating that a variety of deformation mechanisms contributed to fabric development. The geometry and kinematic interpretation of these fabrics are briefly discussed in Part 2, but several important observations of microscopic features with mechanistic implications are added here.

Tectonized limestones along dip-slip faults at Rennell Sound and Louise Island consist of strongly-twinned calcite porphyroblasts, often set in a very fine-grained matrix of recrystallized calcite. With aspect ratios of up to 5:1, these porphyroblasts define a strong mesoscopic and microscopic rock fabric (Fig. 3.5a). Recrystallized grain size varies from 10 μm–50 μm in most samples. Evidence for brittle deformation acting synchronously with recrystallization and twinning is seen in several features: Variably folded calcite-filled veins cut the foliation at moderate to high angles and, in some samples, are offset along the foliation. Siliceous layers in some of the limestones have been brittlely segmented, and fibrous calcite infills gaps between the segmented sections (Fig. 3.5b).

Similar features, indicating a mix of crystal-plastic and brittle deformation mechanisms, are preserved in fabrics associated with the Louscoone Inlet fault system. For example, figure 2.39b illustrates the interaction between three different, synchronously active deformation mechanisms: a twinned calcite porphyroblast is cut by a brittle dextral microfault; the surrounding matrix consists of very fine-grained, recrystallized calcite. These textural features all demonstrate that, similar to fabric development in the Buck Channel pluton, tectonic strains during faulting were accommodated by several different semi-brittle deformation processes.
Figure 3.5: Microscopic deformation textures associated with fabrics along Middle Jurassic reverse fault, Rennell Sound. a.) twinned elongate calcite porphyroblasts; b.) Fibrous calcite growth filling in between segmented portions of silicious layer in carbonate. Field of view in both photographs is 5 mm, left to right.
3.2 Slaty Cleavage (Cretaceous rocks)

3.2.1 Mesoscopic, microscopic descriptions

Sedimentary rocks of the Haida and Skidegate formations contain two morphologically distinct types of cleavage. The first occurs only in mudstones located at least a kilometre outside the Long Inlet deformation zone (LIDZ), and represents spaced cleavage surfaces oriented at high angles to bedding. This spaced cleavage is always sub-perpendicular to bedding, but bears no consistent geometric relationship to any mappable or local structures; it is addressed separately below (section 2.3). The second type is slaty cleavage, and is found in mudstones and siltstones within and adjacent to the LIDZ and along the Louscoone Inlet fault system (LIFS). The two cleavage types are never found in the same outcrop. Geometric descriptions in Part 2 note that slaty cleavage is parallel to steeply-dipping axial surfaces of Late Cretaceous/Early Tertiary folds. Within the centre of the LIDZ, the cleavage consistently strikes northwesterly and has varying angular relationships to bedding, depending on position on folds. In less strongly folded areas, slaty cleavage is less well defined and more variable in orientation. In some of these areas, the dominant mesoscopic fabric is a strong fissility shallowly inclined to bedding, and has no consistent orientation with respect to regional or local structural trends.

Slaty cleavage is also well developed in folded Cretaceous rocks along the Louscoone Inlet fault system. Here, northwest-striking cleavage surfaces are shallowly to moderately inclined to fault zone boundaries, and are interpreted to have formed during Tertiary dextral strike-slip faulting (Part 2).

Slaty cleavage surfaces impart a strong rock fissility which often makes sample collection difficult. Cleavage surfaces are locally planar, but refract gently across lithologic boundaries. Pencil intersection structures form where weakly anastomosing cleavage surfaces are inclined steeply to bedding plane fissility.
Thin sections of samples containing slaty cleavage show variable degrees of fabric development. In samples collected from the LIFS, slaty cleavage forms the dominant microscopic fabric, and is defined by phyllosilicate-rich films which anastomose around feldspar and quartz grains. Only a weak fabric parallel to lithologic layering is preserved (Fig. 3.6a). Phyllosilicate films have an average spacing of similar magnitude to mean grain size, which ranges from 20 \( \mu m \) to 100 \( \mu m \). Cleavage surfaces, in addition to decreasing spacing dramatically, refract to lower bedding/cleavage angles in the finer grained layers (Fig 3.6b). Feldspar and quartz framework grains between the phyllosilicate layers have a weak preferred dimensional orientation (aspect ratios up to 2:1) parallel to slaty cleavage traces. Overgrowths on some quartz grains, and microstylolitic grain boundaries parallel to cleavage, indicate that solution/reprecipitation processes contributed to this weak preferred orientation.

Thin sections of rocks with slaty cleavage collected from the LIDZ have strong microscopic bedding-parallel fabrics. Although cleavage is well developed in outcrop, only a weak microscopic fabric is visible. This fabric is similar to those described above: weakly aligned framework grains are interposed between phyllosilicate-rich films.
Figure 3.7: Photomicrographs showing features associated with slaty cleavage in Cretaceous siltstones. a.) Cleavage is defined by domainal phyllosilicate distribution separating quartz-feldspar-rich microlithons. Bedding-parallel compaction fabrics are almost non-existent. b.) Refraction of slaty cleavage across lithologic layering defined by changes in grain size. Field of view in both photographs is 2 mm, left to right.
3.2.2 X-ray phyllosilicate texture analysis

The degree of preferred orientation of phyllosilicates was analyzed in four samples, using the automated X-ray texture goniometer laboratory at the University of California, Los Angeles. Appendix C reviews the analytical techniques and contains a listing of the data generated. The X-ray texture goniometer measures the relative intensity of reflections for all minerals with a set d-spacing at various orientations within a rock sample, giving an indirect quantitative measure of the degree of preferred orientation for those minerals. In this study X-ray diffractograms showed strong chlorite [002] peaks in all samples, no significant muscovite or biotite peaks, and moderate peaks for clay minerals; chlorite was therefore examined in the texture analyses.

X-ray texture analysis has often been used as a strain measurement technique in low-grade metamorphic and sedimentary rocks by integrating the preferred orientation results with the March (1932) model for reorientation of passive markers. Although several workers have raised various objections to this application, most of which argue that the March model does not reflect the actual orienting mechanisms by which cleavage forms, results compare remarkably well with strains determined using other methods (Oertel, 1970). Oertel (1984) discusses many of these objections, and convincingly argues that in many low-grade sedimentary rocks, the March model strain estimation represents a valid and potentially valuable analytical technique.

The March method measures the total accumulated strain since initial deposition, which generally comprises an early compaction component and superimposed tectonic strain(s). To take full advantage of the information afforded by this type of analysis, the measured strain must be factored into these two components. No unique factoring solution exists, several assumptions regarding the geometry of each strain event are made. Oertel (1970) outlines these assumptions, and presents a strain factoring technique which was used in modified form in the present study. His technique assumes that mesoscopic cleavage is parallel to the principal tectonic strain axes; that tectonic
deformation is constant volume; that strain is homogeneous throughout the sample; and that compaction strains comprise a uniaxial shortening perpendicular to bedding. If these criteria are met, the finite plane of flattening determined from the March analysis should have an orientation lying between cleavage and bedding, and this plane will rotate towards the cleavage plane during successive increments of tectonic strain. If the first assumption stated above is erroneous, and mesoscopic cleavage is parallel to the finite (total) strain ellipse rather than the tectonic strain ellipse, the March strain axes should lie close to the cleavage orientation. In this case, if strains are to be factored, tectonic strain axes must be chosen which lie outside the cleavage orientation, and can only be constrained when other field observations indicate tectonic strain directions.

This strain factoring procedure provides an iterative solution in three steps: i) an initial estimate of the tectonic principal strain magnitudes is made, ii) the reciprocal of this strain estimate is applied to both the total (March) strain and the bedding orientation, and iii) the resultant factored strain is analyzed to determine whether it represents a uniaxial shortening perpendicular to restored bedding orientation. Successive iterations which modify the original strain estimates will converge on a solution satisfying the geometric requirements for compaction strains, provided that a good estimate of tectonic strain axis orientations was made in the first step. A seemingly simpler approach, to estimate the magnitude of the compaction strain and subtract it from the total strain, cannot be used because strain superimposition is a non-commutative operation. To ease the cumbersome tensor transformations necessary for each iteration, a PASCAL computer program FACTOR was written for this study; the program algorithm and source code are included in appendix C.
Results:

Calculated strains are shown graphically in figure 3.9, and tabulated in Table 3.1 and appendix C. Most slaty cleavage samples produced "noisy" records, both for the original data, and data corrected for absorption, but a good degree of preferred orientation is evident. In samples collected from the LIFS (90-219, 90-233, 90-258) the calculated March flattening plane (i.e., plane with the highest intensity of chlorite basal reflections) is within 10° of mesoscopic cleavage. The calculated elongation direction is down-dip in one sample (233), and parallel to strike in the others. Samples 90-219 and 90-258 showed a fair fit to the idealized March model, and were analyzed further by factoring the March strain into compaction and tectonic components.

When the March flattening plane and mesoscopic cleavage are nearly parallel, as in these samples, factoring tectonic strains is difficult because the choice of tectonic strain axis strongly influences the outcome. For a homogeneous, coaxial strain, unless cleavage remains perpendicular to bedding throughout tectonic deformation, one would expect the finite flattening plane to lie within the acute angle between bedding and the tectonic flattening plane. This geometry occurs in both samples, if it is assumed that cleavage is parallel to the tectonic flattening plane (Oertel, 1970). Therefore, strain factoring was attempted by using cleavage as a guide to tectonic strain directions. Elongation direction within the cleavage plane was assumed to be parallel to dip direction, which is nearly vertical in both samples. This is most reasonable because it represents the unconfined direction for material to flow during deformation; in any case, if the orientation of the strain directions is incorrectly chosen, the strain factoring will not converge on a uniaxial compaction strain.
Figure 3.7: Stereographic projections (lower hemisphere) showing results from X-ray goniometer texture analysis for chlorite, and March strain analysis in slaty cleavage samples. All samples except 90-1246 show March flattening plane sub-parallel to cleavage. In sample 90-219, the March flattening plane lies between cleavage and bedding; if cleavage orientation represents tectonic strain orientation, small arcs show direction of rotation of March finite strain towards tectonic strain during final incremental strain. $X_m$, $Y_m$, $Z_m$ represent maximum, intermediate, and minimum March elongations on all diagrams.
For illustrative purposes, the final iteration for sample 90-219 using FACTOR.PAS are reproduced below:

Estimated principal tectonic strains, based on the nine previous iterations, are

\[
\begin{bmatrix}
1.24 & 0.96 & 0.84
\end{bmatrix}
\] (principal elongations);

\[
1 + e_1 = 142^\circ/00^\circ \text{ (parallel to local fold axis)}
\]

\[
1 + e_2 = 232^\circ/80^\circ \text{ (parallel to cleavage, normal to fold axis)}
\]

\[
1 + e_3 = 052^\circ/10^\circ \text{ (normal to cleavage)}
\]

The reciprocal strain tensor, using the same coordinate axes is:

\[
\begin{bmatrix}
0.8064 & 1.0416 & 1.1905
\end{bmatrix}
\]

March strains determined from phyllosilicate orientation are:

\[
\begin{bmatrix}
1.23 & 1.00 & 0.81
\end{bmatrix}
\]

\[
1 + e_1 = 330^\circ/01^\circ
\]

\[
1 + e_2 = 237^\circ/73^\circ
\]

\[
1 + e_3 = 060^\circ/17^\circ
\]

The March strains, transformed into the principal strain axis coordinate system, are:

\[
\begin{bmatrix}
1.2218 & -0.0014 & -0.0563
-0.0014 & 1.0016 & -0.0232
-0.0563 & -0.0232 & 0.8203
\end{bmatrix}
\]

Combination of the reciprocal strain tensor and the March strain tensor yields a factored strain of:

\[
\begin{bmatrix}
0.9853 & -0.0014 & -0.0563
-0.0014 & 1.0432 & -0.0232
-0.0563 & -0.0232 & 0.9766
\end{bmatrix}
\]

which transforms into compass coordinates as:

\[
\begin{bmatrix}
1.0392 & 0.0139 & -0.0116
0.0139 & 0.9325 & -0.0330
-0.0116 & -0.0330 & 1.0333
\end{bmatrix}
\]
The reciprocal strain must also act upon the bedding plane orientation, which changes from 189°/64° to 190°/62°. Finally, the transformation of the factored strain into a coordinate system defined by bedding strike and dip direction yields a bedding-referenced strain tensor:

\[
\begin{pmatrix}
1.0408 & 0.0128 & 0.0123 \\
0.0128 & 1.0361 & 0.0253 \\
0.0123 & 0.0253 & 0.9281
\end{pmatrix}
\]

where R1 is parallel to bedding strike, R2 is parallel to dip direction, and R3 is normal to bedding plane.

Ideally, for a uniaxial compaction strain, the upper left and central components of the tensor will be equal, and the non-diagonal cross shear terms will be zero. The above deviation from this idealized behaviour likely reflects the noisy quality of the original phyllosilicate texture data (appendix C), the uncertainties in choosing the orientations of the original strain axes, and the low strain intensity in the sample. Slightly better results shown by Oertel (1970) were obtained from samples with greater strain intensity.

Sample 90-258 yielded a tectonic strain of:

\[
\begin{pmatrix}
1.8444 & 1.130 & 0.480
\end{pmatrix}
\]

(principal elongations);

and a pre-tectonic compaction strain tensor of:

\[
\begin{pmatrix}
1.4513 & -0.0952 & 0.0008 \\
-0.0952 & 1.4203 & -0.0119 \\
0.0008 & -0.0119 & 0.4804
\end{pmatrix}
\]

It is interesting to note the differences in both strain components in these two samples. Sample 90-258 shows a greater strain intensity for both components, even though both samples have similar lithologies and were obtained from similar structural settings. The calculated pre-tectonic shortening perpendicular to bedding in for sample 90-219 represents a compaction strain of only -0.11, or 11% volume reduction. This is an
unreasonably low value for fine-grained sedimentary rocks. One possible explanation is that tectonic strains were superimposed on sediments which had undergone only partial compaction. However, it is unlikely that such large pore volumes could be maintained for the significant time period between original deposition (Early Cretaceous) and initial deformation (latest Early Tertiary). More likely, the unreasonably low value stems from analytical uncertainties. Any further interpretation of these results would be over-interpreting the limited data available.

Sample 1246, collected from the core of a Late Cretaceous-Early Tertiary mesoscopic fold in the Long Inlet deformation zone, shows a March flattening plane nearly parallel to bedding, and elongation parallel to the bedding/cleavage intersection direction. Strain factoring in this sample, using the same procedure outlined above, yielded an estimated tectonic strain of:

\[
\begin{bmatrix}
0.960 & 1.578 & 0.660 \\
1 + e_1 &= 150^\circ/00^\circ \\
1 + e_2 &= 230^\circ/89^\circ \\
1 + e_3 &= 050^\circ/09^\circ
\end{bmatrix}
\]

and a pre-tectonic compaction of:

\[
\begin{bmatrix}
1.3986 & -0.0935 & 0.0385 \\
-0.0935 & 1.4603 & -0.0000 \\
0.0385 & -0.0000 & 0.4817
\end{bmatrix}
\]

(bedding plane coordinate system).

One sample (90-MA1) lacking slaty cleavage, but containing deformed ammonites was analyzed as a means of assessing the validity of the March strain model. Bedding plane strain ratios for this sample, derived from the distorted fossils, are \(R_{xy} = 1.34\) (Appendix B), while March strains, when resolved to bedding plane coordinates, are
R_{xy} = 1.37. This close correspondence is encouraging, especially because it is based on only one section due to irregularities in the second section (see Appendix C for notes on irregularities in individual samples). If both sections had been possible to use, an even closer fit to the March model may have been achieved, such as results reported by Oertel (1970) and Chen (1991b).

3.2.3 Discussion

Most workers accept that slaty cleavage forms through a combination of mechanisms, including:

1.) internal grain-shape changes (crystal plasticity),
2.) solution/redeposition along grain boundaries,
3.) mechanical reorientation of minerals, and
4.) stress-controlled growth of minerals (e.g., Sorby, 1953; Tullis, 1976; Williams, 1977; White and Knipe, 1978). Cretaceous rocks in the Queen Charlotte Islands reveal how these different processes might contribute to fabric development at low (sub-greenschist grade) temperature and pressure conditions.

In most samples, the chlorite orientation fits the March model for passive rotation well, and the predicted March strains, when factored, yielded reasonable values for pre-tectonic compaction. This fit does not strictly imply that preferred phyllosilicate orientations formed purely through mechanical rotation processes; Tullis (1976) found that preferred orientations of syntectonically recrystallized phyllosilicates in experimentally deformed rocks also match the predictions of rotation models.

The occurrence of chlorite as a major phyllosilicate phase can be attributed to diagenetic growth. Scanning electron microscopy of Cretaceous rocks of the Queen Charlotte Islands (Fogarassy, 1989) shows that chlorite occurs with a euhedral pore-filling or pore-lining habit, consistent with early diagenetic growth at the expense of
ferro-magnesian detrital minerals. Much of this growth probably occurred prior to
defacement, and its subsequent reorientation led to the strong preferred orientation
fabrics documented. Syn-tectonic chlorite growth may have replaced existing clay
minerals, and inherited their preferred orientation.

Framework grains (mostly quartz and feldspar) are largely equant, and rarely
display sutured grain boundaries or overgrowths. This implies that the role of crystalline
plasticity and pressure solution in framework grains is relatively minor in the rocks
examined, relative to the phyllosilicate-orienting processes. The preference for re-
orientation processes, which probably comprise both particulate flow and diagenetic
growth, over pressure solution and plasticity is consistent with deformation in poorly-
lithified rocks at low confining pressures and high fluid pressures (Knipe, 1986b). Such
conditions are reasonable for Cretaceous rocks in the Queen Charlotte Islands, which
were not likely to have been buried to depths greater than 1-2 km prior to deformation.

It is noteworthy that the March analysis determines a flattening plane oriented
between cleavage and bedding surfaces in some samples, because it implies that the
preferred orientation determined by the texture analysis does not define the mesoscopic
cleavage. One possible explanation for this relationship is that mesoscopic cleavage is
largely defined by syn-tectonically recrystallized grains, while the phyllosilicate
distribution includes recrystallized grains, grains of diagenetic origin, and detrital grains.
In addition, the weak preferred orientation of framework grains through dissolution and
rotation will contribute to the mesoscopic fabric, but is not quantified by the March
analysis.

It should also be noted that due to sample defects in sample 90-233 (see Appendix C) and
the near parallelism of bedding and March flattening plane in sample 90-1246, this
geometric relationship can only be clearly demonstrated in two samples, and further
analyses are required to determine if it represents a general example.
3.3 Spaced Cleavage (Cretaceous mudstones)

3.3.1 Mesoscopic, microscopic descriptions

A second, morphologically distinct type of cleavage limited to Cretaceous sedimentary rocks consists of discrete parting surfaces spaced 2 mm to 10 mm apart. This cleavage type is common in concretionary mudstone and siltstone lithotypes in the upper part of the Haida Formation, and to a lesser degree, in finer grained parts of the Skidegate Formation. It is best seen in shoreline outcrops at eastern Skidegate Inlet and Cumshewa Inlet, and is absent from the more highly deformed rocks of the LIDZ which contain slaty cleavage. Spaced cleavage has two geometric characteristics which make it morphologically distinct from all other fabric types encountered: firstly, it invariably cuts bedding at high angles (45° to 90°), regardless of the structural position of outcrops on mesoscopic or macroscopic folds, and secondly, it refracts sharply across bedding interfaces (Fig. 3.8a). This refraction can be quite pronounced: in the most extreme cases, cleavage surfaces bend up to 60°–70° across bedding interfaces, and in some cases reverse their sense of inclination to layering. This refraction gives cleavage surfaces a kinked or folded appearance, with fold axial surfaces parallel to lithologic boundaries. Cleavage orientation in any given location is variable, and no consistent regional trends are apparent.

Cleavage spacing is directly proportional to grain size, ranging from 2 mm in the finest grained mudstones, to 10 mm in siltstones. Individual cleavage surfaces are continuous for several tens of centimetres in the most homogeneous lithologies, but in more lithologically variable, strongly-layered rocks, they anastomose and bifurcate across lithologic boundaries. Rocks weather preferentially and fracture easily along cleavage surfaces. Freshly-exposed cleavage surfaces in some locations have smooth, shiny coatings with surface striations oriented perpendicular to bedding traces (Fig. 3.8b). This surface coating consists of fine-grained calcium carbonate, and the striations represent variations in thickness of the surface coating. Inter-cleavage microlithons lack
discernible mesoscopic fabrics. Bedding traces are almost always continuous across cleavage surfaces, but in rare examples, apparent offsets of up to 20 cm occur.

Nearly all lithologies containing spaced cleavage also host calcium carbonate-cemented concretions, forming oblate spheroids up to 1 metre across. An increase in bedding thickness within the concretions of up to twice that outside the concretions indicates that they formed early in the compaction history of the sediments. Several relationships between spaced cleavage and concretions exist. Rarely, cleavage is preserved intact within concretions, and there is little change in orientation or style from that in the enclosing mudstone (Fig. 3.9a). This mode of preservation is most common in the smaller (<20 cm) concretions, which also display the smallest changes in bedding thickness relative to the surrounding strata. Most concretions, including all with diameters over 0.5 m, show a more typical configuration: cleavage is preserved only within the outermost few centimetres (Fig. 3.11b). In a few examples, cleavage surfaces bend around massive, uncleaved concretions.

In thin section, cleavage surfaces form irregular parting surfaces across which there is no apparent offset of bedding. Lithons between cleavage surfaces contain no discernible microfabrics other than the bedding-parallel orientation of elongate or platy detrital elements. Cleavage parting surfaces bend around the larger detrital grains and, in a few instances, cut individual detrital grains (Fig. 3.12a, b). Dark brown staining extends into wall rocks for 100 μm to 200 μm from cleavage surfaces.
Figure 3.8: Spaced cleavage fabrics in Cretaceous mudstones: a.) spaced cleavage forms parting surfaces approximately perpendicular to lithologic layering, and refracts sharply across bedding surfaces, b.) polished cleavage surfaces contain striations perpendicular to bedding-cleavage intersection line.
Figure 3.9: Geometric relationships between spaced cleavage and carbonate concretions: a.) in smallest concretions, cleavage is preserved intact, and has similar geometry to cleavage outside of concretions; b.) in most concretions, cleavage fabrics are preserved in the outer parts but are missing from concretion cores.
**Figure 3.10:** Photomicrographs of spaced cleavage fabrics (vertical surface in both photographs): a.) cleavage forms parting surfaces which bend around larger detrital grains. No fabric other than bedding-parallel preferred orientation of grains is visible in microlithons. b.) Rarely, cleavage surfaces cut detrital grains. Field of view in both photographs is 0.5 mm, top to bottom.
3.3.2 X-ray phyllosilicate texture analysis

Three samples of spaced cleavage from Skidegate Channel and Cumshewa Inlet were analyzed for preferred orientation of phyllosilicates using the same techniques applied in the study of slaty cleavage, outlined above and in appendix C. Samples were oriented in the sample holder apparatus such that the rock between cleavage surfaces was irradiated, as a means of evaluating the strain history of inter-cleavage domains. All samples showed peak intensities for chlorite basal planes parallel to bedding, with no measurable deviation toward the cleavage plane direction (Fig. 3.11; Table 3.2). Thus, neither texture analysis, nor conventional optical methods, finds a penetrative strain fabric parallel to the spaced cleavage. Calculated March strains indicate moderate values of shortening perpendicular to bedding, and moderate subhorizontal, north to northwest elongation on bedding surfaces. This elongation direction is subparallel to both the regional structural grain and local bedding strike orientation. Strain factoring in these samples was difficult, in that they lacked mesoscopic fabrics indicating local tectonic strain axis orientations. Factoring was attempted by assuming that the tectonic flattening plane contains the bedding-surface elongation direction determined by the March analysis, and is subvertical, consistent with local and regional structural styles. Results are compiled in Table 3.2 and Appendix C: Two samples restored easily to uniaxial compaction strains perpendicular to bedding, and indicated tectonic flattening values of $1+e_3 = 0.64$ and 0.825, compensated by extension in the vertical direction.

Pre-tectonic compactions in these samples indicate volume losses of 47% and 41%, compatible with pore-volume reductions in fine-grained sedimentary rocks. The third sample failed to converge on a uniaxial compaction strain, suggesting that the local strain history for the sample is not consistent with the above assumptions.
Figure 3.11: Stereographic projections (lower hemisphere) showing results of X-ray goniometer texture analysis of chlorite, and March strain analysis of spaced cleavage samples. All samples show March flattening plane sub-parallel to bedding, with no apparent relationship to cleavage. Northwest elongation directions are parallel to regional and local structural grain.
Table 3.1

<table>
<thead>
<tr>
<th>Sample No. and location</th>
<th>March Strain</th>
<th>Orientation</th>
<th>Bedding orientation</th>
<th>Cleavage orientation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1 + e₁</td>
<td>1 + e₂</td>
<td>1 + e₃</td>
<td>X</td>
</tr>
<tr>
<td></td>
<td>XY plane</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Louscoone Inlet Fault System</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>90-219</td>
<td>1.23</td>
<td>1.00</td>
<td>0.81</td>
<td>330/01</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>150/74</td>
<td>188/64</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>142/80</td>
<td></td>
</tr>
<tr>
<td>90-233</td>
<td>1.46</td>
<td>1.00</td>
<td>0.68</td>
<td>234/54</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>144/64</td>
<td>080/28</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>137/54</td>
<td></td>
</tr>
<tr>
<td>90-258</td>
<td>1.47</td>
<td>1.07</td>
<td>0.63</td>
<td>134/07</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>315/86</td>
<td>006/40</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>139/78</td>
<td></td>
</tr>
<tr>
<td><strong>Long Inlet deformation zone</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>90-1246</td>
<td>1.41</td>
<td>0.94</td>
<td>0.75</td>
<td>314/22</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>232/22</td>
<td>266/15</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>120/90</td>
<td></td>
</tr>
</tbody>
</table>

**Table 3.1:** Strain data for Cretaceous lithologies containing slaty cleavage fabrics, based on X-ray texture goniometer analysis of chlorite basal planes and March strain model. Samples from the Louscoone Inlet fault system show moderate flattening perpendicular to cleavage, while sample from Long Inlet deformation zone has major flattening perpendicular to bedding.

Table 3.2

<table>
<thead>
<tr>
<th>Sample No. and location</th>
<th>March Strain</th>
<th>Orientation</th>
<th>Bedding orientation</th>
<th>Cleavage orientation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1 + e₁</td>
<td>1 + e₂</td>
<td>1 + e₃</td>
<td>X</td>
</tr>
<tr>
<td></td>
<td>XY plane</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Skidegate Inlet</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>87-A108</td>
<td>1.30</td>
<td>0.97</td>
<td>0.79</td>
<td>320/06</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>042/22</td>
<td>340/30</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>133/70</td>
<td></td>
</tr>
<tr>
<td>88-ON2A</td>
<td>1.35</td>
<td>0.92</td>
<td>0.80</td>
<td>186/09</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>178/46</td>
<td>020/42</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>175/30</td>
<td></td>
</tr>
<tr>
<td><strong>Cumshewa Inlet</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>88-379A</td>
<td>1.20</td>
<td>1.04</td>
<td>0.80</td>
<td>140/01</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>139/34</td>
<td>153/30</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>350/20</td>
<td></td>
</tr>
</tbody>
</table>

**Table 3.2:** Strain data for Cretaceous mudstone lithologies containing spaced cleavage mesoscopic fabrics, based on X-ray texture goniometer analysis of orientation of chlorite basal planes and March strain model. All samples show maximum flattening plane close to bedding orientation, and elongation parallel to regional structural grain.
3.3.3 Discussion

Spaced cleavage fabrics are common features in deformed rocks worldwide, but few examples share the morphologic characteristics of the spaced cleavage found in Cretaceous sedimentary strata of the Queen Charlotte Islands. Many early geologists interpreted spaced cleavage as surfaces of brittle failure, which led to general usage of the term "fracture cleavage" for many years, along with its genetic connotations. More recently, most geologists have accepted that spaced cleavage is usually formed in the plane of finite flattening, as a result of pressure solution paired with crystalline plasticity and oriented growth of minerals (Williams, 1972). Rare examples have also been demonstrated to represent extensional fractures, or closely spaced joints (Siddans, 1977; Foster and Hudleston, 1986).

The spaced cleavages described above are not likely to have formed parallel to the finite plane of flattening for several reasons:

1) Cleavage surfaces lack the characteristic features of pressure solution surfaces (concentration of residues, sutured grain contacts), and show no microscopically visible evidence of oriented mineral growth.

2) Cleavages surfaces cut through individual detrital grains, but there is no offset of grain segments or loss of material along the cleavage surfaces.

3) X-ray texture analyses fail to document any tectonic flattening strain parallel to cleavage surfaces in the inter-cleavage domains.

4) Varying relationships with carbonate concretions suggest formation during gravitationally-induced compaction.

5) The degree of cleavage refraction is probably too great to be accounted for by the slight rheological contrasts across layering.

However, most of these observations favour formation of cleavage surfaces as extensional microfractures, coincident with compaction and dewatering of pore fluids.
Such extensional features could form in the following manner: following sedimentation, continued burial by sediment accumulation at higher levels would lead to an increase in pore fluid pressure levels in the low permeability mudstones and siltstones of the Haida and Skidegate formations, provided that the rate of expulsion of pore fluids could not compete with pore collapse related to overburden stress. This is a reasonable mechanism for pore-fluid pressure increase in the central Queen Charlotte Islands, in that mid-Cretaceous mudstones are succeeded by thick sequences of Honna Formation conglomerate, which were likely deposited rapidly. Fracture spacing would be controlled by local permeability: individual fractures will only be capable of lowering pore-fluid pressure levels for small distances laterally. This fracture spacing/permeability relationship could easily account for the observed dependence of fracture spacing on grain size.

This model suggests that spaced "fracture" cleavage surfaces formed while sediments were still undergoing compaction strains, a requirement also demonstrated by geometric relationships to carbonate concretions. Clearly, the compacting sediments must have had a cohesive strength sufficient to support propagation of brittle fractures, in some instances through individual grains. This implies that limited cementation and authigenic pore-filling mineral growth had occurred by the time fractures began forming. This cementation would also have contributed to lowering permeability, enhancing the eventual brittle failure.

Original fracture orientation would have been nearly perpendicular to bedding, in response to a vertically directed overburden stress. Minimum and intermediate principal stress axes were likely nearly equal in magnitude, which is reflected in the wide local and regional variation fracture orientation. Fracture orientation was probably most strongly controlled by local variations in the lesser principal stress orientations, stemming from irregularity in basement surfaces, down-slope sediment creep, or lateral lithologic variation within the compacting sediments.
This extensional model for cleavage formation does not explain the large magnitude of refraction observed, which must therefore be related to post-failure modification of fracture orientation. Because refraction occurs across bedding interfaces, rheologic contrasts between adjacent beds probably play the dominant role in determining fracture orientation. Although the rheological layering may slightly alter the principal stress orientation across interfaces if compaction strains are locally non-coaxial, the magnitude of the angular change is not likely to produce the sharp angular discordances observed in the field. Instead, an explanation involving amplification of small original angular discordances through later compaction is proposed: Local variations in basin topography may have resulted in slight original dips in bedding, and further burial may have induced minor down-slope creep in the compacting sediments. This creep will be preferentially concentrated in less competent layers, which will undergo small increments of non-coaxial strain, most likely accommodated by internal grain contact sliding and particulate flow. As a result, stress trajectories will vary slightly between the non-deforming competent layers, and the weaker, deforming layers (Fig. 3.12a). If extensional fractures form in orientations reflecting these local stress directions, the enhanced permeability afforded by the fracture network will lead to more rapid dewatering and sediment compaction. This further compaction will amplify the small angular variation in fracture surfaces in layers of differing competence (Fig. 3.12b). Any further increments of non-coaxial strain along the weaker layers during this compaction will cause additional rotation of fractures in those layers. As fractures rotate during this compaction strain, fractures in favourable orientations may undergo small shear displacements parallel to fracture walls; calcite precipitation along fractures during this shearing will lead to the striated fracture surfaces found in many samples.
Figure 3.12: Schematic diagram showing proposed mechanism for the formation of spaced cleavage fabrics during compaction: a.) slight original bedding dips may lead to downslope creep of sediments, accommodated by non-coaxial strain in weaker layers. Spaced cleavage forms as extensional fabric, and varies slightly in orientation in different layers, as a result of changes in orientation of principal stress direction due to localized non-coaxial strain; b.) this small angular difference is amplified during subsequent compact (60% compaction for "weak" layers, 40% compaction for all others shown above), either through passive rotation in a coaxial strain field, or through further increments of non-coaxial strain. This diagram shows a paleoslope angle of 3°; greater or lesser slope angles are possible, provided resolved shear stresses on bedding surfaces are sufficient to cause downslope creep.

One aspect of this qualitative model can easily be tested, by determining whether the amount of angular variation presently observed along cleavage surfaces can be accounted for by orientation modification during compaction. Cleavage/bedding angles within the most sharply refracted layers typically measure between 50° and 60°. If a uniaxial compaction strain of 50% is subtracted from this (50% is the approxiamte value predicted from factored March strains, and is consistent with expected pore-volume reductions in siltstone lithologies), an original cleavage/bedding angle of 67° to 74° is
predicted. This value seems too small for the suggested model, and it may imply that some of the compaction in the more sharply refracted layers is non-coaxial. Alternatively, some of the angular modification may have occurred post-compaction, during the formation of broad folds.

Possible analogues to the refracted cleavage of the Haida Formation are "vein structures" or spaced foliations described in partially consolidated muds from Deep Seas Drilling Project (DSDP) cores collected in forearc regions (Cowan, 1982; Lundberg and Moore, 1986). These features have been interpreted to have formed as extensional features (Arthur et al., 1980; Cowan, 1982; Knipe, 1986a; Lundberg and Moore, 1986).
4. Summary

Processes of deformation during the Mesozoic and Cenozoic evolution of the Queen Charlotte Islands are typical of those associated with low pressure and temperature (sub-greenschist grade) deformation. The bulk of tectonic strain is accommodated through brittle processes; i.e., those involving loss of cohesion and translation along either pre-existing surfaces (stratification, foliation), or along newly-formed fracture surfaces. Examples of penetrative strain accommodation include shear fabric development within some of the less stratified units and along some faults, and the slaty cleavage formation in folded Cretaceous sedimentary rocks. Spaced cleavage surfaces, limited in occurrence to the less deformed, fine-grained Cretaceous rocks, are interpreted to have formed during compaction, and pre-date tectonic deformation of the host units.

Penetrative (shear) fabrics in the Buck Channel pluton and along some of the major faults formed largely through semi-brittle creep, the combination of crystalline plasticity (both dislocation creep and twinning), pressure solution, and microcracking. The role of crystalline plasticity is diminished in the formation of slaty cleavage, which developed largely through grain rotation, oriented growth/replacement of phyllosilicates, and minor pressure solution. Spaced cleavage fabrics are interpreted to have formed solely through brittle extensional failure.

Fluids are implicated as deformation-enhancing agents throughout deformation. This is best demonstrated by the ubiquitous occurrence of extensional veins formed through hydraulic fracturing in all deformed rocks, and the common presence of pressure solution features. Elevated pore-fluid pressure levels are required for the formation of compaction-induced, bedding-perpendicular extensional fractures, and probably contributed to re-orientation processes in the formation of slaty cleavage fabrics.
Part 4:

Regional synthesis: Triassic to Late Tertiary Geologic Evolution of the Central Insular Belt and Adjacent Parts of the Canadian Cordillera, from a Queen Charlotte Islands Perspective

1. Introduction

This chapter fits structural, stratigraphic, and magmatic elements of the Queen Charlotte Islands into a more regional geological context, and discusses implications that recent geologic studies in the islands have on Insular Belt and Cordilleran evolution. Because of this broad scope, only some of the data discussed represent contributions from this study, and results from other FGP studies are used freely.

The Insular Belt forms the westernmost part of the Canadian Cordillera, and as such, will provide one of the best geologic records of west coast plate interactions through much of the Mesozoic and Cenozoic. Prior to the commencement of the Queen Charlotte Basin FGP study, our understanding of Insular Belt evolution was based largely on regional geologic surveys of the Queen Charlotte Islands (Sutherland Brown, 1968) and Vancouver Island (e.g. Jeletzky, 1976; Muller et al., 1974; Muller, 1977), as well as numerous smaller, more specialized studies. The plutonic history of the adjacent Coast Plutonic Complex provides many clues to the regional magmatic evolution (e.g., van der Heyden, 1989), but unfortunately, obscures much of the stratigraphic history of the eastern Insular Belt. The geologic framework outlined by this and other FGP studies provides one of the most broad-based, and at the same time, detailed, contributions to our understanding of Insular Belt geology since the importance of accreted terranes to
Cordilleran evolution was first recognized, and therefore, a re-evaluation of the tectonics of the area is appropriate.

The following discussion traces the evolution of the central Insular Belt through the Mesozoic and Cenozoic, focusing on many of the structural, stratigraphic, and magmatic elements of the Queen Charlotte Islands, the offshore Queen Charlotte Basin, and adjacent parts of the Coast Plutonic Complex. One danger in interpreting the tectonic history for such a large area lies in not recognizing major lateral displacements along faults through the area. Movement along such unrecognized faults might have juxtaposed elements within the area which evolved separately over much of their history. This is particularly important to the present discussion, in that biostratigraphic (Tipper 1984) and paleomagnetic (Irving et al., 1985) evidence indicates Late Mesozoic northward latitudinal displacement of many Insular and Intermontane belt terranes. The magnitude and timing of this northward displacement, and the locations of faults along which this displacement occurred are topics of considerable controversy (e.g., May and Butler, 1986; Umhoefer et al., 1989). For two reasons, major latitudinal displacements along faults within the Queen Charlotte Islands, Queen Charlotte Basin, and adjacent Coast Plutonic Complex are unlikely. First, although northwest-trending faults are common along the western flank of the Coast Plutonic Complex, much of their recent history involves primarily vertical uplift of the eastern areas, and major strike-slip displacements are difficult to establish (Crawford et al., 1987; van der Heyden, 1989). Second, paleomagnetic analyses of Insular Belt rocks yield Late Triassic paleopoles similar (within error range) to Intermontane Belt rocks, implying that most northward translation occurred along faults within or east of the Intermontane Belt (May and Butler, 1986; Irving and Wynne, 1991). Therefore, it is inferred that elements of the Insular Belt, Coast Plutonic Complex, and western Intermontane Belt adjacent to the Queen Charlotte Islands have maintained their approximate relative positions throughout the
Mesozoic and Cenozoic, although during this time period they were south of present latitudes, both in global coordinates and relative to cratonic North America.

2. Pre-Middle Jurassic

2.1 Terrane Affinities

It is now generally recognized that much of the western Canadian Cordillera is composed of discrete tectonostratigraphic terranes, each defined by a distinct stratigraphic, structural, and magmatic signature (e.g., Monger et al., 1982; Monger and Berg, 1987). The nature and positions of many specific terrane boundaries, and the history of terrane interactions are still subject to much debate. The Insular Belt, or Insular superterrane, comprises two segments: the Wrangell terrane (Wrangellia), comprising rocks of the Queen Charlotte Islands, Vancouver Island, and westernmost parts of the British Columbia mainland and southeast Alaska, and the Alexander terrane, consisting of rocks found east of Wrangellia and west of the Coast Plutonic Complex. Earliest linkage between these terranes was originally thought to be represented by the Late Jurassic to Early Cretaceous Gravina-Nutzotin overlap assemblage (Monger et al., 1982), but more recently, linkage since the Late Paleozoic by plutonic stitching has been established (Gardner et al., 1988). The boundary between the terranes at the latitude of the Queen Charlotte Islands probably extends through northern Hecate Strait and along the western margin of the Coast Plutonic Complex southwest of Prince Rupert (van der Heyden, 1989). Stratified rocks in this area are largely obscured by Mesozoic plutons and have not been mapped in detail, but no terrane suture incorporating oceanic crustal material is recognized (Hutchison, 1982; van der Heyden, 1989).

The Insular superterrane is separated from more easterly Intermontane Belt terranes by metamorphic and plutonic rocks of the Coast Plutonic Complex. These crystalline rocks obscure the boundary between the two superterranes, which has led to
much controversy on the timing and nature of the proposed suture, and on the origin of
the Coast Plutonic Complex itself. Most recent workers subscribe to one of two theories:
The "collision" model views the Coast Plutonic Complex as the metamorphic/plutonic
welt formed as a result of the Cretaceous accretion of the Insular superterrane onto the
western margin of North America (e.g., Monger, 1982; Monger and Berg, 1987).
Proponents of the Cretaceous collision model point to synchronous Cretaceous
deformation across the Coast Plutonic Complex, and syn-orogenic sedimentation styles
in adjacent areas as evidence. A second group proposes that the Insular Belt and the
western part of the Intermontane Belt (Stikine terrane) have been a common entity since
a least the Middle Jurassic. This model views the Coast Plutonic Complex as the
exhumed root of a long lived, intra-plate, easterly migrating volcanic arc, formed above
an east-dipping subduction zone (e.g., Brew and Ford., 1983; van der Heyden, 1989).
Several elements of Queen Charlotte Islands geology may bear on this controversy, and
will be discussed later.

2.2 Magmatic and stratigraphic setting

The definitive Wrangellian succession comprises upper Triassic tholeiitic basalts
(Karmutsen Formation in the Queen Charlotte Islands and Vancouver Island), succeeded
by thickly-stratified shallow-water carbonates and more thinly bedded, deeper-water
carbonate and siliciclastic rocks (Kunga and Maude Groups) (Jones et al., 1977). The
recent discovery of Upper Paleozoic volcaniclastic and marine sedimentary strata in the
Queen Charlotte Islands (Hesthammer et al., 1991a) points to another stratigraphic
characteristic common to all known Wrangellian occurrences, which should also be
considered definitive of the terrane: in all locations, Upper Paleozoic volcaniclastic,
carbonate, and siliciclastic strata are separated from the Late Triassic tholeiitic volcanic
rocks by a Late Permian to Middle Triassic unconformity (Fig. 4.1).
Figure 4.1: Comparative stratigraphic sections for Wrangellian assemblages in Alaska, Vancouver Island, Oregon, and Queen Charlotte Islands, and for Alexander Terrane in southeastern Alaska.

**ABBREVIATIONS:**

- Arg. = argillite
- Rhy. = rhyolite
- Bas. = basalt
- Br. = breccia
- Clst. = conglomerate
- Aggl. = agglomerate
- Dac. = dacite
- Lst. = limestone
- Dol. = dolomite
The oldest widely-exposed rocks on the Queen Charlotte Islands are Upper Triassic submarine volcanic flows. Geochemical evidence suggests these rocks were generated as products of arc rifting (Andrew and Godwin 1989; Barker et al., 1989). Overlying Upper Triassic carbonates record a transition from shallow to moderate water depths, believed to be related to deposition on a subsiding or drowned shelf (Desrochers and Orchard, 1991). Cameron and Tipper (1985) suggest that Lower Jurassic rocks of the Queen Charlotte Islands accumulated in a back arc setting, on the basis of tuffaceous components within the sedimentary succession. In the present study, this tuffaceous component, represented by euhedral plagioclase-rich silts, was documented in rocks as old as the Norian Peril Formation. Associated arc rocks are not found in the Queen Charlotte region, but possible candidates occur elsewhere in the Insular Belt: both the Talkeetna Arc rocks of the Peninsular terrane of southeast Alaska (Plafker et al., 1989) and the Bonanza Group on Vancouver Island have Late Triassic volcanic arc components (Jeletzky, 1969). Plafker et al. (1989) suggest that the Talkeetna Arc may have been contiguous with the Bonanza Group arc in the Mesozoic, and has since been translated northward along strike-slip faults outboard of the present plate margin. If so, its Late Triassic to Early Jurassic position may have been adjacent to the Queen Charlotte Islands.

Recognition of a Late Paleozoic linkage between the Alexander and Wrangell terranes (Gardner et al., 1988) implies that Mesozoic successions within the two terranes should have common features. Indeed, Alexander terrane successions contain the same Late Paleozoic/Early Triassic stratigraphic hiatus now recognized in all Wrangellian exposures. However, early Mesozoic assemblages on the two terranes contain some notable differences (Fig. 4.1). Late Triassic Alexander terrane strata in southeast Alaska consist of submarine volcanic breccias, overlain by rhyolite flows and tuffs, succeeded by limestone, clastic rocks, and basaltic flows (Gehrels and Saleeby, 1987). Although
the limestones and sedimentary strata in this succession are similar in age to those found in the Kunga Group on the Queen Charlotte Islands, the underlying felsic volcanic component, and the overlying basaltic flows are not recognized in Wrangellian successions. However, Upper Triassic volcanic assemblages in both locations are interpreted as rift-related (Gehrels and Saleeby, 1987; Barker et al., 1989). Additional work is needed to evaluate whether the apparent stratigraphic discrepancies are related to different rift environments within the composite terrane, or whether they indicate lateral displacements of the different components.

3. Middle and Late Jurassic

3.1 Middle Jurassic Deformation

Aalenian and older strata of the Queen Charlotte Islands were involved in a period of contractional deformation not seen in younger rocks. The angular unconformity at the base of the Bajocian Yakoun Group strata tightly brackets the timing of this deformation to Late Aalenian-Early Bajocian. The structural geometry of Middle Jurassic deformation is outlined in detail in Parts 2 and 3; in brief, it comprises southwest-verging faults and folds, which decrease in strain intensity across a northwest-trending deformation front on central Moresby Island (Fig. 2.48). It is interesting to note the remarkable similarities in timing between this and other Cordilleran orogenic events, both within and outside the Insular Belt, some of which are listed below:

A. Insular Belt

1.) North of the Queen Charlotte Islands, McClelland and Gehrels (1990) document Early to Middle Jurassic dextral shearing along the eastern margin of the Alexander terrane.

2.) Monger (1991) recognizes folds in the Lower and (?)Middle Jurassic Bowen Island Group in the southern Insular Belt which are cut by a 155 Ma pluton.
B. Coast Plutonic Complex

1.) van der Heyden (1989) describes the Late Jurassic Banks Island plutonic belt (160 Ma - 150 Ma) as intruding possible Late Triassic, deformed and metamorphosed Wrangellian strata, and suggests the plutons are syn-orogenic.

C. Intermontane Belt

1.) amalgamated terranes of the Intermontane Superterrane were accreted to North America in the Early to Middle Jurassic, coincident with a widespread deformation event in both allochthonous and autochthonous units (Monger et al., 1982; Ross et al., 1985; Brown et al., 1986; Fillipone and Ross, 1990).

2.) The Stikine, Cache Creek, Bridge River, and Cadwallader terranes were assembled and accreted to Quesnellia in Middle Jurassic time (Mortimer, 1986; Rusmore et al., 1988).

3.) Internal deformation of the Cache Creek terrane occurred in Middle to early Late Jurassic time (Mortimer et al., 1990).

4.) Earliest deformation in the Skeena fold belt (Stikine terrane) occurred between Oxfordian and Albian time, but cannot be more closely constrained (Evenchick, 1991).

Clearly, deformation was widespread throughout the western Cordillera in late Early to early Late Jurassic time, on both sides of the Coast Plutonic Complex. This coincidence in timing is noteworthy because it brings into question, but in no way disproves, Cretaceous "collision" models for the origin of the Coast Plutonic Complex. The "collision" model would have Insular and Intermontane superterranes evolving separately until the Cretaceous, and both would have been subject to unrelated Middle Jurassic orogenic events. A more parsimonious model might call for widespread Middle Jurassic assembly of all outboard terranes along the west coast of North America, with all
younger basins and tectonic features superimposed on this landmass. This more fixist view would consider the Coast Plutonic Complex a feature formed within a composite superterrane (van der Heyden, 1989), and would see Cretaceous orogenic events as related to intra-terrane deformation.

3.2 Arc magmatism on the Queen Charlotte Islands and adjacent areas

Youngest volcanic arc activity within the Queen Charlotte Islands is marked by the Bajocian Yakoun Group, although records of older volcanism exist as tuffaceous components in the Kunga and Maude Groups. The polarity of the Yakoun Group arc is uncertain, but the apparent lack of crustal sutures to the east of the islands is consistent with volcanism above an east-dipping subduction zone. The present location of the Yakoun Group arc rocks, directly adjacent to the Pacific/North American plate boundary, implies that the Jurassic plate boundary was farther west (present day coordinates) than at present, and that segments of transitional crust have been tectonically removed from the margin.

Middle and Late Jurassic plutonic rocks of the Queen Charlotte Islands are too young to be arc roots to Yakoun Group volcanic rocks. However, their spatial and compositional characteristics suggest they may be earliest elements of easterly younging and compositionally evolving magmatic belts in the adjacent Coast Plutonic Complex (Anderson and Greig, 1989; van der Heyden, 1989; Anderson and Reichenbach, 1991). U-Pb ages of plutonic rocks, compiled from van der Heyden (1989) and Anderson and Reichenbach (1991) define five magmatic belts (Fig. 4.2). From west to east, these are 1) the Middle Jurassic San Christoval plutonic suite (172-171 Ma) and 2) the Middle to Late Jurassic Burnaby Island plutonic suite (168-158 Ma) in the Queen Charlotte Islands (Anderson and Reichenbach, 1991); and 3) the Late Jurassic Banks Island belt (160-155 Ma), 4) the Early Cretaceous McCauley Island belt (131-123 Ma), and 5) the
mid-Cretaceous Ecstall belt (110-94 Ma) on the mainland (van der Heyden, 1989). West to east modal and geochemical changes of plutons are consistent with proposed magmatism above an east-dipping subduction zone west of the San Cristoval plutonic suite (van der Heyden, 1989; Anderson and Reichenbach, 1991): felsic (locally peraluminous) compositions are most abundant in the easterly suites.

It is tempting to speculate that this trend of easterly-migrating arc volcanism continues back into the earliest Middle Jurassic parts of Wrangellia in Alaska, which may have been displaced from the Queen Charlotte region via 600-1000 km of Jurassic-Cretaceous sinistral faulting along the present Alexander terrane/Wrangellia boundary (Plafker et al., 1989). For example, a possible older and more primitive part of the eastwardly-evolving magmatic front may be the 187-175 Ma Border Ranges ultramafic-mafic assemblage of the Peninsular terrane (Plafker et al., 1989). This suite is considered to have been proximal to and northwest of the southern Wrangell terrane in Alaska throughout their shared geological history (Plafker et al., 1989). The Border Ranges plutonic assemblage is spatially and petrologically linked to Talkeetna Formation volcanic rocks (Burns, 1985; Plafker et al., 1989). Thus, the Border Ranges ultramafic-mafic assemblage may be the displaced, petrologically most primitive and earliest expression of the eastwardly-migrating and evolving magmatic front.

In latest Jurassic time, magmatic activity was concentrated along the Banks Island plutonic belt; a late Jurassic stratigraphic hiatus in the present-day Queen Charlotte Islands suggests much of the area was emergent. This emergence may be related to regional uplift in the wake of the thermal-orogenic front which passed through the islands in the Middle Jurassic.
Figure 4.2: Plutonic history of the Queen Charlotte Islands and adjacent Coast Plutonic Complex, compiled from Anderson and Reichenbach (1991) and van der Heyden (1989). Magmatic trends show consistent easterly younging from Middle Jurassic to Early Tertiary.
4. Cretaceous marine sedimentation and tectonic relations

By the Early Cretaceous, plutonic activity had shifted well into the western Coast Plutonic Complex and the Queen Charlotte Islands region was tectonically and magmatically quiescent. Renewed marine sedimentation began in the Early Cretaceous. Extensive biostratigraphic studies document a gradual eastward transgression of shallow marine facies across the islands, spanning Hauterivian to Albian time. These strata, represented by sandstone facies of the Queen Charlotte Group and Longarm Formation, are largely feldspar-lithic arenites, compatible with derivation from underlying volcanic and sedimentary units, and are primarily easterly-derived (Yagishita, 1985a, 1985b).

The inception of coarse clastic sedimentation (Coniacian Honna Formation) represents a marked change in depositional styles from underlying sandstones and shales. Most clasts in the Honna Formation conglomeratic deposits are lithologies common to the Queen Charlotte Islands, but rare magmatic epidote and muscovite + biotite bearing granitic rocks are more compatible with a Coast Plutonic Complex source (Woodsworth, personal communication, 1990). Again, paleocurrent indicators suggest easterly derivation.

The westward flowing paleocurrents, and the two-part division of Cretaceous strata directs attention to the adjacent Coast Plutonic Complex to examine possible coeval tectonic events which might be related to sedimentation style. Both crustal shortening, and regional uplift in the Coast Plutonic Complex overlap the timing of Cretaceous sedimentation in the Queen Charlotte Islands (Crawford et al., 1987; van der Heyden, 1989; Rubin et al., 1990). Immediately east of Hecate Strait, the west-directed Prince Rupert thrust system is the local expression of a regionally extensive, mid-Cretaceous thrust system (Rubin et al., 1990). Syn-tectonic emplacement of the Ecstall pluton (U=Pb = 98 ± 4 Ma, Woodsworth et al., 1983) indicates that much of the thrusting and associated folding took place in the Late Albian (Crawford et al., 1987).
Deformation of this age is not documented in the Queen Charlotte Islands: here, earliest Cretaceous deformation follows deposition of the Honna Formation (Late Coniacian or younger), when all Cretaceous units are involved in northeast-southwest shortening and folding. Stratigraphic/magmatic relations in the islands do not allow more accurate determination of the timing of this deformation, but it is tempting to speculate that it is the young foreland expression of mid-Cretaceous west-directed thrusting to the east.

Rubin et al. (1990) suggest that the Queen Charlotte Group represents foredeep deposits that "probably represent late orogenic deposits that were derived from the [Prince Rupert] thrust system". However, this interpretation is at odds with recent Cretaceous stratigraphic models by Haggart (1991) which demonstrate that the largely Valanginian to Albian Longarm Formation is an inseparable part of the Queen Charlotte Group, and a lithologic equivalent to the Haida Formation. The inclusion of the Longarm Formation in the Queen Charlotte Group extends the age of the group to at least 20 Ma before the earliest documented shortening in the Prince Rupert thrust system.

The coarse clastic sediments of the Honna Formation are likely products of the short-lived progradation of submarine fan complexes into the basin from the east, which began in the Late Turonian to Coniacian (Haggart, 1986, 1991; Haggart et al., 1989), and ceased by Santonian-Maastrichtian time. Honna Formation rocks are only a few million years younger than deformation in the Prince Rupert thrust system, and are more likely candidates for foredeep deposits than are the older Queen Charlotte Group rocks. This interpretation is favoured by Higgs (1990), who proposes that fans feeding the Honna Formation originated from uplands east of the islands, formed during west-vergent thrust faulting within the composite Wrangellia-Alexander terrane. However, he places the leading edge of the thrust belt at the Sandspit Fault, a proposition not supported by mapping in the Queen Charlotte Islands (Lewis et al., 1991b).
The above Cretaceous tectonostratigraphic evolution of the Queen Charlotte Islands region differs extensively from that proposed by Yorath and Chase (1981) and elaborated by Yorath and Hyndman (1983). Central to these earlier models is the interpretation that the Cretaceous sedimentary succession in the basin is directly related to terrane amalgamation. The Lower Cretaceous Longarm Formation was interpreted by Yorath and Chase (1981) and Yorath and Hyndman (1983) as a "suture assemblage", a chaotic and lithologically variable unit which accumulated in a narrow and deep trough parallel to the suture of the Wrangell and Alexander terranes. The overlying "post-suture assemblage" consisting of the Queen Charlotte Group (the Haida, Honna and Skidegate formations) was interpreted as stable-shelf deposits overlapping the Longarm Formation "trough". These ideas were perpetrated prior to recognition of Paleozoic linkage between the terranes, and we now know that the Longarm Formation consists almost entirely of shelf deposits similar to the overlying Queen Charlotte Group (Haggart, 1989; Haggart and Gamba, 1990).

5. Tertiary Basin formation

The Tertiary Queen Charlotte Islands and adjacent areas developed in a plate tectonic setting significantly different from earlier periods. This difference produced distinct structural, stratigraphic, and magmatic styles in the region: major structures are strike-slip and extensional faults; great thicknesses of syn-tectonic and post-tectonic sediments accumulated in structurally controlled offshore basins (Rohr and Dietrich, 1990); magmatic activity included syn-extensional calc-alkaline plutons (Anderson and Greig, 1989), Paleogene mafic to intermediate calc-alkaline volcanic flows, and largely Neogene, mafic to felsic, calc-alkaline to tholeiitic volcanic flows and pyroclastic rocks (Hickson, 1991). This change coincides closely with Eocene (about 57 Ma and 43 Ma) changes in Pacific/Kula/North America relative plate motion, during which the North
American plate margin shifted from oblique convergence to mainly dextral transcurrent motion (Engebretson et al., 1985; Lonsdale, 1988; Stock and Molnar, 1988; Engebretson, 1989). The absolute position of the plate boundary at the latitude of the Queen Charlotte Islands at this time is uncertain, but many geologic features indicate that it was outboard of its present position. For example, Cretaceous marine sedimentary rocks represent shelf to outer shelf deposits right to the edge of the present North American plate, and no Late Cretaceous subduction assemblages occur (Haggart, 1991). Thick accumulations of Paleogene marine and non-marine clastic rocks occur on the westernmost part of Graham Island (Port Louis well, Shouldice, 1971; White, 1990), just inboard of the present plate margin. Crustal fragments displaced from the margin in the Tertiary may now occur in the Yakutat block in southern Alaska. Tertiary relative plate motions (Engebretson et al., 1985) are capable of transporting such a fragment from an Eocene position off the Queen Charlotte Islands to its present position.

Much of the early Tertiary, pre-Masset volcanism on the southern Queen Charlotte Islands pre-dates the 43 Ma change in plate motions, and may be associated with the earlier oblique convergence (Hickson, personal communication, 1991). In contrast, post-43 Ma volcanism, plutonism, and extension-related sedimentation in the Queen Charlotte Basin developed in the mainly transcurrent plate setting. This period shortly followed extensional faulting and uplift in the Coast Plutonic Complex (Crawford et al., 1987; Friedman and Armstrong, 1988; Heah, 1990; van der Heyden, 1989) and widespread regional extension in south-central British Columbia (Tempelman-Kluit and Parkinson, 1986; Parrish et al., 1988).

Despite the large body of knowledge now available for the Queen Charlotte Region, several disparate views for the formation of the Queen Charlotte Basin exist. Previous models for the Tertiary history of the Queen Charlotte region incorporated strike-slip faulting with offsets of tens of kilometres through the Queen Charlotte Islands
accompanied by rifting in the southern Queen Charlotte Basin and more recent subsidence of the northern basin related to regional crustal flexure above a subducting Pacific plate (Yorath and Chase, 1981; Yorath and Hyndman, 1983). More recent proposals vary in their interpretations, according to the primary source of data used. Dietrich and Rohr (1990) view the Queen Charlotte Basin as the product of distributed shear across the western Pacific Plate Margin, and base their arguments mainly on offshore seismic reflection data. In contrast, Hyndman and Hamilton (personal communication, 1991) propose that the basin developed as a result of crustal stretching, and use refraction data, styles of volcanism, and relative plate motion studies to support their arguments. Lyatsky (1991a), drawing primarily on potential-field data, suggests the basin formed mainly due to vertical block movement, and neglects horizontal displacements as a driving mechanism. Aspects of basin evolution discussed in Part 2 (section 4.5) suggest that components of several of these models might apply, but on-land geological styles are supportive of extension related to distributed shear. Clearly, an integrated effort utilizing these many different data sources is needed, and is the focus of ongoing studies.
6. Summary

1. Upper Paleozoic Wrangellian strata represent deposition in an arc setting, and may overlie a basement assemblage of Lower Paleozoic rocks similar to those found in the Alexander terrane.

2. Late Triassic to Middle Jurassic Wrangellian strata record a period of rift-related volcanism, followed by marine sedimentation on a subsiding or deepening shelf.

3. Middle Jurassic contractional deformation is synchronous with widespread Cordilleran deformation events, and may mark the assembly and accretion of outboard Cordilleran terranes onto North America.

4. Middle Jurassic Yakoun Group volcanism represents the earliest occurrence of Mesozoic subduction-related arc volcanism in the Insular superterrane (Fig. 4.3a). A Mesozoic to Early Cenozoic magmatic front migrated eastward into the Coast Plutonic Complex, and was followed by widespread uplift and periods of non-deposition.

5. Early Cretaceous shallow marine facies transgressed easterly across the Queen Charlotte Islands following Late Jurassic uplift, and were deposited in a tectonically quiescent setting (Fig. 4.3b).

6. Sudden progradation of Honna Formation conglomerate fans into the basin was triggered by either structural thickening or regional uplift in the western Coast Plutonic Complex (Fig. 4.3c).

7. Widespread marine deposition ceased in latest Cretaceous to Paleogene time with the onset of marine and nonmarine sedimentation in extensional sub-basins, and northerly-migrating, Eocene and Oligocene volcanism and associated plutonism.

8. The Tertiary Queen Charlotte Basin is mainly an Eocene to Pliocene extensional basin. Onshore geologic constraints are consistent with basin formation related to limited amounts of strike-slip faulting, distributed across the present-day basin.
A. Middle Jurassic (initiation of Yakoun Gp. arc)

Queen Charlotte Islands
Yakoun Group Arc

Oceanic Lithosphere

B. Mid-Cretaceous (widespread marine sedimentation)

Queen Charlotte Islands

Late Jurassic block faults

Magmatism shifted east into Coast Plutonic Complex

C. Late Cretaceous (easterly-derived coarse clastic deposits)

Queen Charlotte Islands

Uplift in Coast Ranges, west-directed thrust faulting on Prince Rupert Thrust system

D. Early Tertiary (change to strike-slip margin)

Queen Charlotte Islands

Renewed magmatism in Queen Charlotte Islands

early extension and sedimentation in Queen Charlotte Basin

Figure 4.3: Schematic regional cross sections showing relative positions of major Mesozoic and Cenozoic tectonic elements discussed in text.

A. Middle Jurassic: Arc volcanism above east-dipping subduction zone is represented by Yakoun Group strata. Reasonable arc-trench gap values place plate boundary west of present position of margin, requiring intervening crustal material.

B. Mid-Cretaceous: Widespread marine sedimentation (Queen Charlotte Group) overlaps Late Jurassic block faults, focus of magmatism has shifted east into Coast Plutonic Complex.

C. Late Cretaceous: Structural thickening and/or uplift in western Coast Plutonic Complex triggers progradation of coarse clastic deposits across the present-day Queen Charlotte Islands, which interfinger with and overlie earlier marine sediments.

D. Early Tertiary: Plate boundary changes to dominantly transcurrent margin; extensional and strike-slip faulting in the Queen Charlotte Islands and areas to the east herald initial stages of Queen Charlotte Basin deposition.

All sections are drawn only to illustrate relative positions of elements; no implications regarding lithospheric thicknesses, slab dip angle, or structures omitted are intended.
REFERENCES

Allmendinger, R.W., Marret, R.A., and Cladhous, T.

Anderson, E.M.
1942: The dynamics of faulting and dyke formation with application to Britain; Oliver and Boyd, Edinburgh, 192 p.

Anderson, R.G.

Anderson, R.G. and Greig, C.J.

Anderson, R.G. and Reichenbach, I.


Andrew, A. and Godwin, C.I.

Angelier, J.
1979: Determination of the mean principal stresses for a given fault population; Tectonophysics, v. 56, p. t17-t26.


Arthur, M.A., Carson, B., and von Huene, R.

Atkinson, B.I.
REFERENCES


Barker, F., Sutherland Brown, A., Budahn, J.R., and Plafker, G.

Barrett, C.S. and Masalski, T.B.

Biot, M.A.

Blake, T.F.
1878: On the measurement of the curves formed by cephalopods and other mollusks; Philosophical Magazine, v. 5, p. 241-262.

Brew, D.A. and Ford, A.B.

Brown, R.L., Journeay, J.M., Lane, L.S., Murphy, D.C, and Rees, C.J.

Burns, L.E.

Cameron, B.E.B. and Hamilton, T.S.

Cameron, B.E.B. and Tipper, H.W.

Carter, E.S., Orchard, M.J., and Tozer, E.T.

Carter, N.L.

Carter, N.L. and Tsenn, M.C.
REFERENCES

Chen, R.T.


Chen, R.T. and Oertel, G.
1991: Determination of March strain from phyllosilicates preferred orientation: a numerical method; Tectonophysics, in press.

Clapp, C.H.

Cowan, D.S.

Crawford, M.L., Hollister, L.S., and Woodsworth, G.J.
1987: Crustal deformation and regional metamorphism across a terrane boundary, Coast Plutonic Complex, British Columbia; Tectonics, v. 6, p. 343-361.

Crespi, J.M.

Dawson, G.M.

Desrochers, A.

Desrochers, A. and Orchard, M.J.

Dietrich, J.R. and Rohr, K.M.M.
1991: Distributed strike-slip deformation across the Queen Charlotte Basin; in EOS, transactions, American Geophysical Union, v. 71, p. 1581.

Dunnet, D.
Durney, D.W. and Ramsay, J.G.

Ells, R.W.

Engebretson, D.
1989: Northeast Pacific-North America plate kinematics since 70 Ma; Geological Association of Canada, Pacific Section, Annual Symposium, Victoria, B.C., Programme and Abstracts, p. 4-6.

Engebretson, D.C., Cox, A., and Gordon, R.G.
1985: Relative motions between oceanic and continental plates in the Pacific basin; Geological Society of America, Special Paper 206.

Etchecopar, A., Vasseur, G., and Daignieres, M.

Etheridge, M.A.

Evans, J.P.

Evenchick, C.A.
1991: Geometry, evolution, and tectonic framework of the Skeena fold belt, north central British Columbia; Tectonics, in press.

Fillipone, J.A., and Ross, J.V.

Fogarassy, J.A.S.
1989: Stratigraphy, sedimentology, and diagenesis of the Cretaceous Queen Charlotte Group, Queen Charlotte Islands, British Columbia; unpublished MSc. thesis, University of British Columbia, Vancouver B.C.

Foster, M.E. and Hudleston, P.J.

Friedman, R.M. and Armstrong, R.L.
REFERENCES

Fry, N.
1979: Random point distributions and strain measurement in rocks; Tectonophysics, v. 60, p. 89-105.

Gamba, C.A.

Gamba C.A., Indrelid, J., and Taite, S.P.


Gehrels, G.E. and Saleebey, J.B.

Geological Survey of Canada

1987b: Preliminary magnetic anomaly map (residual total field), Prince Rupert, British Columbia; Geological Survey of Canada, Map NN-8-9-M.

Gratier, J.P. and Vialon, P.
1980: Deformation pattern in a heterogeneous material: folded and cleaved sedimentary cover immediately overlying a crystalline basement (Oisans, French Alps); Tectonophysics, v. 65, p. 151-180.

Gretener, P.E.

Groshong, R.H.

Haggart, J.W.

REFERENCES


Haggart, J.W. and Gamba, C.A.

Haggart, J.W. and Higgs, R.

Haggart, J.W., Indrelid, J., Hesthammer, J., Gamba, C.A., and White, J.M.

Haggart, J.W., Lewis, P.D., and Hickson, C.J.

Harbert, W.

Harbert, W., and Cox, A.

Heah, T.S.T.

Hesthammer, J.

1991a: Stratigraphic and structural geology of Jurassic units on central Graham Island, Queen Charlotte Islands, British Columbia; unpublished MSc. thesis, University of British Columbia, Vancouver, British Columbia

Hesthammer, J., Indrelid, J., Lewis, P.D., and Haggart, J.W.

Hesthammer, J., Indrelid, J., Lewis, P.D., and Orchard, M.J.

Hickson, C. J.


Hickson, C.J. and Lewis, P.D.

Higgs, R.

1990: Sedimentology and tectonic implications of Cretaceous fan-delta conglomerates, Queen Charlotte Islands, Canada; Sedimentology, v. 37, pp. 83-103.

Hobbs, B.E., Means, W.D., and Williams, P.F.

Hutchison, W.W.
1982: Geology of the Prince Rupert-Skeena map area; Geological Survey of Canada, Memoir 394.

Hyndman, R., and Hamilton, T.

Indrelid, J.
REFERENCES

Indrelid J., Hesthammer, J., and Lewis, P.D.

Indrelid, J., Hesthammer, J., and Ross, J.V.

Irving, E., and Wynne, P.J.

Irving, E., Woodsworth, G.J., Wynne, P.J., and Morrison, A.

Jakobs, G.K.

Jeletzky, J.

Jeletzky, J.

Jones, D.L., Silberling, N.J., and Hillhouse, J.

Kamb, B.

Kirby, S.
1985: Rheology of the Lithosphere; Reviews of Geophysics and Space Physics, v. 21, p. 1458-1487.

Knipe, R.J.
1986a: Microstructural evolution of vein arrays preserved in Deep Seas Drilling Project cores from the Japan Trench, Leg 57; Geological Society of America Memoir, 166, p. 75-87.

1986b: Deformation mechanism path diagrams for sediments undergoing lithification; Geological Society of America Memoir 166, p. 151-160.
REFERENCES

Lama, R.D. and Vutukuri, V.S.

Lewis, P.D.


1991b: Geology of the Burnaby Island/Ramsay Island map area, Queen Charlotte Islands, British Columbia; Geological Survey of Canada, Open File 2316.

Lewis, P.D. and Hickson, C.J.
1990: Geology, Langara Island (west half), British Columbia; Geological Survey of Canada, Map 9-1990, scale 1:50,000

Lewis, P.D. and Ross, J.V.

1988b: Crustal shortening in a wrench fault tectonic setting, Queen Charlotte Islands, British Columbia; in Geological Society of America Abstracts with Programs, 84th Annual Cordilleran Section Meeting, Las Vegas, Nevada.


Lewis, P.D. and Ross, J.V.

Lewis, P.D., Haggart, J.W., Rohr, K.M.M., Dietrich, J.R. and, Thompson, R.I.,

Lewis, P.D., Hesthammer, J., Indrelid, J., and Hickson, C.J.

1991a: Triassic to Neogene geologic evolution of the Queen Charlotte Basin; Canadian Journal of Earth Sciences, in press.
REFERENCES

Lewis, P.D., Thompson, R.I., Haggart, J.W., and Hickson, C.J.
1991b: Comment on "Sedimentology and tectonic implications of Cretaceous fan-delta conglomerates, Queen Charlotte Islands, Canada"; Sedimentology, in press.

Lonsdale, P.

Lundberg, N. and Moore, J.C.

Lyatsky, H.V.


MacKenzie, J.D.

MacKevett E.M. Jr.

Massey, N.W.D. and Friday, S.J.

March, A.

May, S.R. and Butler, R.F.

McClelland, W.C. and Gehrels, G.E.

Minster J. and Jordan, T.H.
Monger, J.W.H.

Monger, J.W.H. and Berg, H.C.
1987: Lithotectonic terrane map of western Canada and southeastern Alaska; United States Geological Survey Miscellaneous Field Studies Map, MF-1874-B.

1982: Tectonic accretion and the origin of the two major metamorphic and plutonic welts in the Canadian Cordillera; Geology, v. 10, 70-75.

Moody, J.D. and Hill, M.J.

Mortimer, N.

Mortimer, N., van der Heyden, P., Armstrong, R.L., and Harakal, J.
1990: U-Pb and K-Ar dates related to the timing of magmatism and deformation in the Cache Creek terrane and Quesnellia, southern British Columbia; Canadian Journal of Earth Sciences, v. 27, p. 117-123.

Muller, J.E.

Oertel, G.


1980: Strain in ductile rocks on the convex side of a folded competent bed; Tectonophysics, v. 66, p. 15-34.

1981: Strain estimation from scattered observations in an inhomogeneously deformed domain of rocks; Tectonophysics, v. 77, p. 133-150.

1984: The relationship of strain and preferred orientation of phyllosilicate grains in rocks— a review; Tectonophysics, v. 100, p. 413-447.

Oertel, G. and Ernst, W.G.
1978: Strain and rotation in a multilayered fold; Tectonophysics, v. 48, p. 77-106.
Orchard, M.J.

Orchard, M.J. and Forester, P.J.L.

Owens, W.H.

Parrish, R.R., Carr, S.D., and Parkinson, D.L.

Paterson, S.R., Tobisch, O.T., and Bhattacharyya, T.

Plafker, G., Nokleberg, W.J., and Lull, J.S.

Poulton, T.P., Hall, R.L., Tipper, H.W., Cameron, B.E.B., and Carter, E.S.

Ramberg, H.
1960: Conatact strain and folding instability of a multilayered body under compression; Geologische Rundschau, v. 51, p. 405-439.

1964: Selective buckling of composite layers with contrasted rheological properties, a theory for simultaneous formation of several orders of folds; Tectonophysics, v. 1, p. 307-341.


Ramsay, J.G.

REFERENCES

Ramsay, J.G. and Huber, M.I.  


Ratschbacher, L. and Oertel, G.  

Riccardi, A.C.  

Richards, G.G., Cristie, J.S., and Livingstone, K.W.  

Richardson, J  

Riddihough, R. and Hyndman, R.D.  

Rohr, K.M.M. and Dietrich, J.R.  


Ross, J.V. and Lewis, P.D.  

Ross, J.V., Fillipone, J.A., Montgomery, J. F., and Bloodgood, M.A.  

Royden, L. and Keen, C.E.  

Rubin, C.M., Saleeby, J.B., Cowan, D.S., Brandon, M.T., and McGroder, M.F.  
1990: Regionally extensive mid-Cretaceous west-vergent thrust system in the northwestern Cordillera: Implications for continental margin tectonism; Geology, v.18, p. 276-280.
REFERENCES

Russmore, M.E., Potter, C.J., and Umhoefer, P.J.
1988: Middle Jurassic terrane accretion along the western edge of the Intermontane superterrane, southwestern British Columbia; Geology, v. 16, p. 891-894.

Schultz, L.G.

Secor, D.T. Jr.

Shouldice, D.H.

Siddans, A.W.B.

Sorby, H.C.

Souther, J.G.

Souther, J.G. and Jessop, A.

Sutherland Brown, A., and Jeffrey, W.G.
1960: Preliminary geological map, southern Queen Charlotte Islands; British Columbia Department of Mines.
Sutherland Brown, A.

Taite, S.P.


Tempelman-Kluit, D.J., and Parkinson, D.

Thompson G.A.

Thompson, R.I.

Thompson, R.I.

Thompson, R.I.

Thompson, R.I. and Lewis, P.D.


Thompson, R.I. and Thorkelson, D.J.

Thompson, R.I., Haggart, J.W., and Lewis, P.D.

Tipper, H.


Tipper, H.W., Smith, P.L., Cameron, B.E.B., Carter, E.S., Jakobs, G.K., and Johns, M.J.

Tullis, T.

Tullis, J. and Yund, R.A.
1987: Transition from cataclastic flow to dislocation creep of feldspar: mechanisms and microstructures; Geology, v. 15, p. 606-609.

Umhoefer, P.J., Dragovich, J., Cary, J., and Engebretson, D.C.
1989: Refinements of the "Baja British Columbia" plate-tectonic model for northward translation along the margin of western North America; Tectonics, v. 8, p. 101-111.

Vallier, T.L.
van der Heyden, P.

Vellutini, D.

Vellutini, D. and Bustin, R.M.

Wanamaker, B.J. and Evans, B.
1985: Experimental crack healing in olivine; in Mineral Physics, v. 1, American Geophysical Union Monogram Series.

Wenk, H. R.

Wernicke, B. and Burchfiel, B.C.

White, J.
1990: Evidence of Paleogene sedimentation in the Queen Charlotte Islands, west coast, Canada; Canadian Journal of Earth Sciences, v. 27, p. 533-538.

White, N.
1987: Constraints on the measurement of extension in the brittle upper crust; Norsk Geologisk Tidsskrift, v. 67, p. 269-279.

White, S.H. and Knipe, R.J.
1978: Microstructure and cleavage development in selected slates; Contributions to Mineralogy and Petrology, v. 66, p. 165-174.

Whiteaves, J.F.
1883: On the Lower Cretaceous rocks of British Columbia; Royal Society of Canada, Transactions, v. 1, sec. IV, p. 81-86.

Williams, P.F.

REFERENCES

Winkler, H.G.F.

Woodsworth, G.J.


Woodsworth, G.J. and Tercier, P.E.


Yagishita, K.

Yagishita, K.

Yin, A.

Yorath, C.J. and Chase, R.L.

Yorath, C.J. and Hyndman, R.D.

Young, I.F.
Appendix A: Chronologic Review of Recent Publications on the Structural Evolution of the Queen Charlotte Islands Region

Numerous formal publications, progress reports, conference presentations, and geologic maps detailing new geologic discoveries in the Queen Charlotte Islands have been produced since the inception of the Queen Charlotte Basin Frontier Geoscience Program study in 1987. Because of the varied nature of these productions, the lead time between data gathering and publication date varies from several months to several years. As a result, some of the more recently published works are superceded by papers with dates one or two years earlier, a point of possible confusion to those reviewing the literature. As an aid to the compiler, this appendix lists all of the recent works pertaining to the structural geology and tectonic evolution of the Queen Charlotte Islands, by presenting them in a chronologic order consistent with the time period over which the data presented were collected. Figure A.1 shows this progression in diagramatic form. This approach should minimize the apparently contradictory data and interpretations that would otherwise plague compilers who are studying the Queen Charlotte Islands region. No attempt is made to list papers pertaining to non-structural aspects of the geology of the region.

Studies completed prior to the 1987 field season are discussed by Woodsworth (1991) and Woodsworth and Tercier (1991) and are not listed here. The Queen Charlotte Basin FGP began in spring, 1987, with over 30 geoscientists based in the Queen Charlotte Islands. Preliminary results from this summer's work appear in GSC Paper 88-1E (Lewis and Ross, 1988a; Thompson, 1988); a discussion by Hamilton and Cameron (1988) also found in this volume is based on work completed prior to the 1987 field season. Lewis and Ross (1988b) also presented their preliminary analysis of structural styles in the islands at the Geological Society of America's annual Cordilleran Section meeting.
The 1988 field season was far more productive in terms of outlining a workable structural model for the central Queen Charlotte Islands. Results of field studies appear in GSC Paper 89-1H (Hesthammer et al., 1989; Lewis and Ross, 1989; Thompson and Thorkelson, 1989) and conference presentations (Thompson and Lewis, 1989). Two papers on structural geology included in the Queen Charlotte Basin FGP report (Indrelid et al., 1991a; Lewis and Ross, 1991) are based on fieldwork completed up to the end of the 1988 field season. In addition, geologic maps included in the FGP report (Hickson, 1990a, 1990b, Hickson and Lewis, 1990; Lewis and Hickson, 1990; Lewis et al., 1990a; Thompson and Lewis, 1990a, 1990b; Thompson, 1990a, 1990b) reflect mapping completed to this date. Higgs' (1990) interpretation of the tectonic setting accompanying Cretaceous deposition is based on work completed to the end of the 1988 field season.

Geologic studies continued at a reduced scale during the 1989 field season, and several reports in GSC Paper 90-1F outline major new discoveries (Haggart et al., 1990; Hesthammer, 1990; Indrelid, 1990; Lewis, 1990; Taite, 1990a). A structural-stratigraphic analysis appearing in the FGP report (Thompson et al., 1991) also includes work completed during the 1989 field season, but is based primarily on earlier studies. Two conference presentations (Lewis et al., 1990b; Taite, 1990b) and two formal publications (Lewis et al., 1991a; Lewis et al., 1991b) were completed following the 1989 field season, and do not include the most recent (1990) field studies.

Mapping during the 1990 field season greatly increased the geologic map coverage in the Queen Charlotte Islands, and is compiled in four GSC Open File reports (Hesthammer et al., 1991b; Indrelid et al., 1991b; Lewis, 1991b; Taite, 1991b). Two reports appearing in GSC Paper 91-1A summarize new ideas concerning the structural evolution of the islands (Lewis, 1991a; Taite, 1991a), and four graduate theses at the University of British Columbia include work completed to the end of the 1990 field season.
**PUBLICATION CHRONOLOGY**

<table>
<thead>
<tr>
<th>Based on field work up to and including:</th>
<th>Current Research Progress Reports</th>
<th>Frontier Geoscience Program Two-Year summary</th>
<th>Conference Presentations</th>
<th>Geologic Maps</th>
<th>Journal Publications</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Lewis and Ross (1988b)</td>
<td>Lewis and Ross (1988b)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lewis (1990)</td>
<td>Thompson et al. (1991)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Tate (1990a)</td>
<td>Thompson et al. (1991)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1990 Field Season</td>
<td>Lewis (1991a)</td>
<td>Lewis et al. (1995a)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Tate (1991a)</td>
<td>Lewis et al. (1995a)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>U.B.C. Graduate Theses</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Hesthammer (1991a)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Indreid (1991)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Tate (1991)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Appendix B: Strain analysis techniques

B.1 Strain measurement from deformed ammonites and belemnites

Finite strains was analyzed at 21 locations using deformed prints of ammonites, and at one location using elongated belemnite moulds. Strains were determined from ammonite prints following the method proposed by Blake (1878), described below:

Most ammonites define a logarithmic spiral when undeformed. Blake's method allows the distortion of this spiral to be determined using three successive radial measurements at increments of 90°. A logarithmic spiral is described by the equation:

$$ r = ke^{\alpha \cot(\gamma)} \quad (A.1) $$

where $r$ is the spiral radius measured from the origin, $k$ is a constant reflecting the rate of spiral growth, $\alpha$ is the angle through which the spiral has passed, and $\gamma$ is the angle between the tangent to the spiral and the radius at the point of tangency (Fig. B.1).

![Figure B.1: Logarithmic spirals in undeformed (A) and deformed (B) states, showing axes of measurement used in strain analysis studies.](image-url)
For three successive radial measurements at intervals of 90°, calculated radii for a given spiral would be:

\[ r_1 = k e^0 = k \]
\[ r_2 = k e^{(\pi \cot(\gamma))/2} \]
\[ r_3 = k e^{(\pi \cot(\gamma))} \]

and it follows that:

\[ r_2 = \sqrt[2]{r_1 \times r_3} \]

When the spiral is homogeneously deformed, with principal strain axes parallel to these same radii, the strain ratio, \( R \), can be represented by:

\[ R = \frac{2\sqrt{(r_1' \times r_3')}}{r_2'} \]

where \( r_1', r_2', \) and \( r_3' \) are the lengths of the deformed radii. This ratio can be determined for a deformed ammonite by making three measurements along the visually-estimated principal strain axes contained in plane spiral plane.

This method only determines the two-dimensional finite strain ratio for the plane containing the deformed spiral, which is usually parallel to bedding surfaces. To estimate either the three-dimensional strain, or the absolute magnitude of a given bedding plane strain, simplifying assumptions of the three dimensional strain geometry must be made. For example, bedding-plane strain axes cannot be calculated without prior knowledge of the volume change component of strain. To reflect this uncertainty, two different strain values are listed in Table B.1: a value for constant volume deformation,
and a value assuming uniaxial shortening. These two values give the likely bounds for actual shortening amounts.

Strain axes were rotated to horizontal for map projection purposes by a single rotation about bedding strike direction; this simple rotation was considered reasonable given the regional subhorizontal fold axial trends in most areas.

Strain values for each ammonite were given a subjective weighting between 1 and 5, reflecting the spiral preservation and confidence of measurement. The arithmetic mean of the weighted strain values was considered to be the representative strain magnitude for each location, and a vector mean of weighted orientations was used as the representative strain orientation (Table B.1).

Belemnite moulds at one location were extended along micro-extensional faults. A Mohr circle construction (Ramsay and Huber, 1983) was used to determine the strain orientation and magnitude, using three moulds with different orientations and amounts of extension.
<table>
<thead>
<tr>
<th>Location Number</th>
<th>Map unit</th>
<th>Number of Samples</th>
<th>Restored Elongation</th>
<th>Percent shortening&lt;sup&gt;1&lt;/sup&gt;</th>
<th>Percent shortening&lt;sup&gt;2&lt;/sup&gt;</th>
<th>Strain Ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ammonites:</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>88-MA1</td>
<td>Ghost Creek</td>
<td>7</td>
<td>136</td>
<td>15.4</td>
<td>25.4</td>
<td>1.34</td>
</tr>
<tr>
<td>88-279</td>
<td>Phantom Ck.</td>
<td>2</td>
<td>123</td>
<td>18.7</td>
<td>29.1</td>
<td>1.41</td>
</tr>
<tr>
<td>88-311</td>
<td>Yakoun Grp.</td>
<td>2</td>
<td>115</td>
<td>9.1</td>
<td>16</td>
<td>1.19</td>
</tr>
<tr>
<td>88-367</td>
<td>Ghost Creek</td>
<td>1</td>
<td>115</td>
<td>14</td>
<td>23.1</td>
<td>1.30</td>
</tr>
<tr>
<td>88-369</td>
<td>Sandilands</td>
<td>14</td>
<td>128</td>
<td>7.7</td>
<td>13.8</td>
<td>1.16</td>
</tr>
<tr>
<td>88-370</td>
<td>Ghost Creek</td>
<td>7</td>
<td>136</td>
<td>16.2</td>
<td>25.9</td>
<td>1.35</td>
</tr>
<tr>
<td>88-371</td>
<td>Ghost Creek</td>
<td>15</td>
<td>119</td>
<td>14.9</td>
<td>24.2</td>
<td>1.32</td>
</tr>
<tr>
<td>88-379</td>
<td>Phantom Ck.</td>
<td>2</td>
<td>123</td>
<td>18.7</td>
<td>29.1</td>
<td>1.41</td>
</tr>
<tr>
<td>88-386</td>
<td>Sandilands</td>
<td>9</td>
<td>165</td>
<td>3.9</td>
<td>7.4</td>
<td>1.08</td>
</tr>
<tr>
<td>88-387</td>
<td>Ghost Creek</td>
<td>14</td>
<td>155</td>
<td>3.9</td>
<td>7.4</td>
<td>1.08</td>
</tr>
<tr>
<td>88-388</td>
<td>Ghost Creek</td>
<td>2</td>
<td>102</td>
<td>20.8</td>
<td>31.5</td>
<td>1.46</td>
</tr>
<tr>
<td>88-389</td>
<td>Ghost Creek</td>
<td>1</td>
<td>139</td>
<td>12.7</td>
<td>21.3</td>
<td>1.27</td>
</tr>
<tr>
<td>88-391</td>
<td>Ghost Creek</td>
<td>2</td>
<td>100</td>
<td>3.9</td>
<td>7.4</td>
<td>1.08</td>
</tr>
<tr>
<td>88-466</td>
<td>Sandilands</td>
<td>2</td>
<td>123</td>
<td>7.7</td>
<td>13.8</td>
<td>1.16</td>
</tr>
<tr>
<td>88-468</td>
<td>Sandilands</td>
<td>2</td>
<td>185</td>
<td>4.9</td>
<td>9.1</td>
<td>1.10</td>
</tr>
<tr>
<td>88-481</td>
<td>Sandilands</td>
<td>2</td>
<td>164</td>
<td>7.2</td>
<td>13</td>
<td>1.15</td>
</tr>
<tr>
<td>89-13</td>
<td>Ghost Creek</td>
<td>9</td>
<td>98.5</td>
<td>3.9</td>
<td>7.4</td>
<td>1.08</td>
</tr>
<tr>
<td>90-ST1</td>
<td>Sandilands</td>
<td>5</td>
<td>110</td>
<td>15.3</td>
<td>24.8</td>
<td>1.33</td>
</tr>
<tr>
<td>90-246</td>
<td>Sandilands</td>
<td>1</td>
<td>130</td>
<td>6.3</td>
<td>11.5</td>
<td>1.13</td>
</tr>
<tr>
<td>90-272</td>
<td>Sandilands</td>
<td>1</td>
<td>130</td>
<td>8.2</td>
<td>14.5</td>
<td>1.17</td>
</tr>
<tr>
<td>90-275</td>
<td>Peril</td>
<td>15</td>
<td>127</td>
<td>14.5</td>
<td>23.7</td>
<td>1.31</td>
</tr>
<tr>
<td>90-457</td>
<td>Peril</td>
<td>5</td>
<td>145</td>
<td>14.5</td>
<td>23.7</td>
<td>1.31</td>
</tr>
<tr>
<td>Belemnites:</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>88-313</td>
<td>Yakoun</td>
<td>-</td>
<td>090</td>
<td>16.2</td>
<td>25.9</td>
<td>1.35</td>
</tr>
</tbody>
</table>

<sup>1</sup>Assumes constant volume, plane strain
<sup>2</sup>Assumes uniaxial shortening

**Table B.1:** Bedding plane strains, determined from measurements of deformed ammonites and belemnites.
B.2 Fry Strain analysis of pelloidal limestones

The distribution and shape of fecal pellets in a sample of deformed Sadler Limestone was used to quantify strains. Several methods can be used to measure strain in a population of originally spherical or elliptical objects. If strain markers were originally completely spherical, measurements of the ellipsoidal shapes are sufficient to characterize strain. More generally, a deformed aggregate of originally randomly oriented elliptical markers can be analyzed using the $R_f/\phi$ technique (Ramsay, 1967; Dunnet, 1969). Alternatively, if the deformed objects had an originally statistically uniform (anticlustered) distribution, the Fry centre-to-centre method (Fry, 1979) can be used, and has the advantage of being independent of original shape.

For the Sadler Limestone sample, the centre-to-centre method was chosen as providing the most accurate solution, because, 1.) The uncertainty of initial shape makes simple ellipsoid shape measurements suspect, and 2.) Only small amounts of elongation axis dispersion were observed in sections, making the $R_f/\phi$ method inaccurate. The centre-to-centre method provides a graphical solution based on a difference in nearest-neighbour spacing with direction in the deformed aggregate. The accuracy of Fry's method is independent of the original shape of the objects and the competency contrast between markers and matrix, and is directly related to the degree of anticlustering and number of objects used (Crespi, 1986). Two perpendicular sections of the same sample were analyzed. One section was oriented perpendicular to foliation and parallel to mineral lineation (XZ section), the other was oriented perpendicular to foliation and lineation (YZ section). Centres of 224 fecal pellets were used in the XZ section, and 381 were used in the YZ section to produce the graphical solutions in Figure B.2. Finite strain ratios of 6.1:1 parallel to lineation and 3.3:1 perpendicular to lineation were measured.
Figure B.2   Fry centre-to-centre analysis plots for Sadler Limestone sample containing deformed fecal pellets: a.) YZ section, $R_{YZ} = 3.3$; b.) XZ section, $R_{XZ} = 6.1$.  

[Diagram a) and b) of Fry centre-to-centre analysis plots]
Appendix C: X-ray Texture Analysis of Phyllosilicates, and March Strain Analysis

C.1 Introduction

The X-ray diffractometer with the texture goniometer attachment provides a convenient method for qualitative assessment of the degree of preferred orientation of phyllosilicate minerals in rocks. X-ray texture analysis techniques were pioneered for metallurgical applications (Schultz, 1949a, 1949b), and were first used by geologists to analyze quartz petrofabrics (Baker et al., 1969) and sulfide minerals. More recently, the techniques have been widely applied to examining phyllosilicate orientation as a measure of cumulative strain in rocks (e.g., Oertel and Ernst, 1978; Oertel, 1980, 1981; Ratschbacher and Oertel, 1987; Paterson et al., 1989; Chen and Oertel, in press). This application utilizes the March theory (1932), which proposes that the degree of preferred orientation in phyllosilicates correlates directly with the magnitude of finite strain in rocks with limited recrystallization. Owens (1973) proposes a generalized equation relating these two parameters, which is used by all later workers.

C.2 Analytical Procedure

Barret and Massalski (1966) and Wenk (1985) outline the general techniques for X-ray analysis of preferred grain orientations. Two sample configurations are commonly used, one employing X-rays transmitted through a thin specimen, the other utilizing X-rays reflected off a planar surface of a thick specimen. Transmission mode runs are preferred by most current workers; the advantages over reflection mode are discussed by Oertel (1978).

The present studies were conducted using the texture goniometer laboratory at the University of California, Los Angeles, which is configured for transmission mode scans. Prior to beginning texture analysis, all samples were scanned in 2θ mode to identify
peaks to be analyzed. All samples showed strong chlorite {002} peaks at angles of approximately 14° (cobalt kα radiation). Other phyllosilicate peaks were weak or non-existent, or shifted position in different samples. For each specimen, two perpendicular sections between 100 μm and 200 μm thick were prepared, each oriented perpendicular to foliation. During a texture scan, the section is rotated about two axes, τ (perpendicular to the plane containing the 2θ angle) and β (perpendicular to the section surface, and rotated about τ). Two stepping motors rotate the samples through 6° intervals on both axes, and a total of 420 sampling points covering a 78° wide strip across each sample are collected. Because of the strong X-ray absorption inherent in the transmission technique, long counting times are required (85 seconds for each point), and each section takes over ten hours for a complete scan.

Raw data were analyzed using computer programs developed by Chen (1991a). This analysis comprises three steps:

1. Intensities for each counting point are corrected for absorption and irradiated volume, both of which vary with sample orientation. This step requires estimation of absorption coefficients, which can vary significantly in different samples, or with direction in a given sample (Chen, 1991b). Although choice of absorption coefficients will partially determine both the shape of the pole figure and the calculated March strain, in a given specimen, mean principal strain values from two perpendicular sections are almost identical for corrected and uncorrected samples (Chen and Oertel, 1991).

2. Corrected intensities are plotted on a stereogram, giving a graphical representation of the degree of preferred orientation in the sample. Principal strain axis directions are visually picked from these pole figures.
3. Relative intensities and predicted strain axis orientations are fit to the March model, using the modified Levenberg-Marquard method (Chen and Oertel, 1991).

Once these three steps are completed for the two perpendicular sections of a given sample, the results are superimposed. Ideally, principal strain axis directions and magnitudes will be similar for both samples. Largest sources of deviation from a perfect match probably result from incorrect choice of absorption coefficients, inhomogeneity of strain or irregularities within samples, or departure from March model predicted orientations. The former may be corrected for by choosing different absorption coefficients in the initial processing step; in the present study, most data were processed using at least four different absorption coefficients to find the best match between sections.
C.3 Compilation of Texture Analysis data

11 samples were analyzed using the techniques outlined above. The results from these analyses are presented in the following manner:

1) Stereographic projections show the contoured intensities, corrected for absorption, for each of the two perpendicular sections. The contoured area covers a 78° wide strip centred about the east-west net equator, which corresponds to the sample plane. Visually picked principal strains are indicated for each sample. Ideally, a 90° rotation of one of the section pole figures, when superimposed on the second, should yield a match in overlapping regions. Few samples show this correspondence, and possible reasons for the mismatch are addressed in notes accompanying each data set.

2.) Linear scan records for each section show the relative intensities (corrected for absorption) plotted on the stereographic projections, and have superimposed on them the predicted intensities based on the March model, as calculated from the strain fitting program. Statistical parameters outlining the goodness of fit, using the Levenberg-Marquardt method (Chen and Oertel, 1991), are given for each scan.

3.) A stereographic projection shows the mean strain axis orientations and magnitudes for both sections, based on the statistical fit to the March model. Great circles outline bedding and cleavage orientation for the sample.

4.) In some samples, calculated (March) strains were factored into tectonic and compaction components, using the program STRFACTOR. The results of this factoring is presented by listing the orientations and magnitudes of the tectonic strain axes, and the compaction strain tensor in bedding-plane coordinates.
5. Short notes elucidate the results presented in 1-4, giving estimates of reliability and possible causes for any apparent discrepancies noted.
SAMPLE NUMBER 90-219 (slaty cleavage)
SAMPLE NUMBER 90-219 (slaty cleavage)

Mean Stretch \[ \begin{bmatrix} 1.23 & 1.00 & 0.81 \end{bmatrix} \]

Stretch Tensor \[
\begin{bmatrix}
1.128 & -0.175 & -0.024 \\
-0.175 & 0.927 & -0.048 \\
-0.024 & -0.048 & 0.988 \\
\end{bmatrix}
\]

(in geographic coordinates, 
R1 = north 
R2 = east 
R3 = down)

\[
1 + e_1 = 330/01 \\
1 + e_2 = 237/73 \\
1 + e_3 = 060/17
\]

Bedding Orientation: 188/64
Cleavage Orientation: 142/80

STRAIN FACTORING SOLUTION:

Estimated tectonic strain: \[ \begin{bmatrix} 1.24 & 0.96 & 0.84 \end{bmatrix} \]

\[
1 + e_1 = 232/80 \\
1 + e_2 = 142/00 \\
1 + e_3 = 052/10
\]

Factored compaction strain tensor, in bedding plane coordinates \[
\begin{bmatrix}
1.0408 & 0.0128 & 0.0123 \\
0.0128 & 1.0361 & 0.0253 \\
0.0123 & 0.0253 & 0.9282 \\
\end{bmatrix}
\]

R1 = strike 
R2 = dip direction 
R3 = bedding normal

NOTES:

1. The two pole figures show an above average match when one is rotated 90°, indicating that both the record obtained and the absorption corrections applied are reliable. Both sections fit the idealized March model well, considering the low strain intensity involved, and the differences in principal values and principal axis orientations between the two sections are small.
2. The March flattening plane lies within the acute angle between cleavage and bedding, slightly closer to cleavage orientation. This is consistent with the inference (see text) that cleavage marks the tectonic strain plane and the March strain represents the total finite strain.

3. Strain factoring was attempted by assuming that mesoscopic cleavage represents the tectonic flattening plane, and elongation direction is up-dip on the cleavage plane. This assumes that shortening is compensated by extension in the unconfined, nearly vertical direction, and that there is little post-cleavage re-orientation of fabric elements. Factoring resulted in a bedding plane-normal uniaxial shortening strain, with only small-cross-shears on bedding surfaces.
APPENDIX C / X-ray Texture, March strain analysis

SAMPLE NUMBER 90-233 (slaty cleavage)
SAMPLE NUMBER 90-233 (slaty cleavage)

Mean Stretch \{ 1.46 1.00 0.68 \}

Stretch Tensor \[
\begin{pmatrix}
0.9828 & -0.0249 & -0.2156 \\
-0.0249 & 0.9635 & -0.2991 \\
-0.2156 & -0.2991 & 1.2004 \\
\end{pmatrix}
\]

(in geographic coordinates, 
R1 = north 
R2 = east 
R3 = down)

1 + e_1 = 054/54.5
1 + e_2 = 144/00
1 + e_3 = 234/35

Bedding Orientation: 080/28
Cleavage Orientation: 137/54

NOTES:

1. The pole figures for the two sections show a very poor match, and section B displays a highly irregular intensity pattern. The concentration of lower than expected intensities in the southwest quadrant indicates a sample irregularity in this region, possibly related to lithologic variation, or interference with the sample holder assembly.

2. The very poor fit to the March model for section B reflects the irregularities in intensity shown in the pole figures. It is notable that despite the poor fit shown, the March flattening plane is close to cleavage orientation, similar to other slaty cleavage samples from the LIFS.

3. Given the irregular pole figures and very poor fit to the March model, strain factoring was not attempted.
SAMPLE NUMBER 90-258 (slaty cleavage)

Mean Stretch \[ \{ 1.57 \ 0.95 \ 0.67 \} \]

Stretch Tensor \[
\begin{pmatrix}
1.3058 & -0.4102 & 0.0248 \\
-0.4102 & 0.9427 & 0.0258 \\
0.0248 & 0.0258 & 0.9428
\end{pmatrix}
\]

(in geographic coordinates,
R1 = north
R2 = east
R3 = down)

\[ 1 + e_1 = 327/01 \]
\[ 1 + e_2 = 062/83 \]
\[ 1 + e_3 = 237/07 \]

Bedding Orientation: 006/54
Cleavage Orientation: 141/79

STRAIN FACTORING SOLUTION:

Estimated tectonic strain:
\[
\begin{pmatrix}
1.844 & 1.130 & 0.480
\end{pmatrix}
\]

\[ 1 + e_1 = 294/66 \]
\[ 1 + e_2 = 139/20 \]
\[ 1 + e_3 = 051/11 \]

Factored compaction strain tensor,
in bedding plane coordinates
\[
\begin{pmatrix}
1.4513 & -0.0952 & 0.0008 \\
-0.0952 & 1.4023 & -0.0119 \\
0.0008 & -0.0119 & 0.4804
\end{pmatrix}
\]

R1 = strike
R2 = dip direction
R3 = bedding normal

NOTES:

1. Pole figures for both sections display very regular symmetry, and the 90° fit between the two sections is very good.
2. Recorded intensities from both sections fit the March model well, although section A shows slightly greater elongation values (this is evident from the greater elongation in the section A pole figure as well). The March flattening plane lies close to the Mesoscopic cleavage orientation, within the acute angle between cleavage and bedding.

3. Strain factoring was attempted by assuming that cleavage orientation represents the tectonic flattening plane, and that tectonic elongation is perpendicular to bedding/cleavage intersection direction. A moderate fit to a uniaxial shortening normal to bedding was reached, and a much greater compaction strain (65% shortening) was measured than in sample 219. This difference may be related to different structural and stratigraphic positions for the two samples, or to the imperfect fits to the March model. A number of different elongation directions within the cleavage plane were attempted, all yielded similar bedding-perpendicular compaction strain results, but cross-shears were lowest for elongation perpendicular to bedding/cleavage intersection.
APPENDIX C / X-ray Texture, March strain analysis

SAMPLE NUMBER 90-1246 (slaty cleavage)
SAMPLE NUMBER 90-1246 (slaty cleavage)

Mean Stretch: \[ \{ 1.41 \ 0.94 \ 0.75 \} \]

Stretch Tensor:
\[
\begin{pmatrix}
1.1181 & -0.1874 & 0.1665 \\
-0.1874 & 1.1379 & -0.1590 \\
0.1665 & -0.1590 & 0.8486
\end{pmatrix}
\]

(in geographic coordinates,
\( R1 = \text{north} \)
\( R2 = \text{east} \)
\( R3 = \text{down} \))

\[ 1 + e_1 = 314/22 \]
\[ 1 + e_2 = 045/03 \]
\[ 1 + e_3 = 142/68 \]

Bedding Orientation: 232/22
Cleavage Orientation: 120/90

STRAIN FACTORING SOLUTION:

Estimated tectonic strain:
\[ \{ 1.58 \ 0.96 \ 0.66 \} \]
\[ 1 + e_1 = \text{vertical} \]
\[ 1 + e_2 = 120/00 \]
\[ 1 + e_3 = 030/00 \]

Factored compaction strain tensor,
in bedding plane coordinates:
\[
\begin{pmatrix}
1.3986 & -0.0935 & 0.0385 \\
-0.0935 & 1.4603 & 0.0000 \\
0.0385 & 0.0000 & 0.4817
\end{pmatrix}
\]

\( R1 = \text{strike} \)
\( R2 = \text{dip direction} \)
\( R3 = \text{bedding normal} \)

NOTES:

1. Pole figures for the two sections show a very poor match. Section A shows a greatest intensity parallel to pole to bedding direction, while section B shows a low value in the centre, and an increase in intensity toward the north and south poles. Experimentation with a wide range of absorption coefficients failed to improve this poor fit. The section A plot confirms that the highest concentration of chlorite basal planes...
corresponds closely to bedding, unlike the previous slaty cleavage samples which had greatest concentrations perpendicular to foliation. As a result, greatest expected intensity would plot outside of the area scanned in section B, which helps explains the poor match between sections.

2. The poor fit to the March model for section B underscores the problem noted above: not scanning the area with highest expected intensity also results in a poor March model fit. Section A shows a reasonably good fit to the March model, despite the noisy appearance of the pole figure.

3. Strain factoring assumed that the cleavage plane marked the tectonic flattening plane, and that elongation direction was parallel to cleavage dip direction. A reasonably close approximation to a uniaxial shortening perpendicular to bedding for the early compaction strain resulted. Amounts of pre-tectonic compaction were similar to those for sample 90-258.
SAMPLE NUMBER 88-379a (spaced cleavage)
SAMPLE NUMBER 88-379a (spaced cleavage)

Mean Stretch \[ \{ 1.20 \ 1.04 \ 0.80 \} \]

Stretch Tensor \[
\begin{bmatrix}
1.1162 & -0.0969 & -0.0581 \\
-0.0969 & 1.0828 & -0.0665 \\
-0.0581 & -0.0665 & 0.8417 \\
\end{bmatrix}
\]

(in geographic coordinates,
R1 = north
R2 = east
R3 = down)

\[ 1 + e_1 = 140/00 \]
\[ 1 + e_2 = 230/24 \]
\[ 1 + e_3 = 049/66 \]

Bedding Orientation: 153/30
Cleavage Orientation: 350/20

STRAIN FACTORING SOLUTION:

Estimated tectonic strain:
\[ \{ 1.212 \ 1.00 \ 0.825 \} \]

\[ 1 + e_1 = \text{vertical} \]
\[ 1 + e_2 = 140/00 \]
\[ 1 + e_3 = 050/00 \]

Factored compaction strain tensor,
in bedding plane coordinates
\[
\begin{bmatrix}
1.1984 & 0.0105 & 0.0178 \\
0.0105 & 1.1981 & -0.1219 \\
0.0178 & -0.1219 & 0.7093 \\
\end{bmatrix}
\]

R1 = strike
R2 = dip direction
R3 = bedding normal

NOTES:

1. This sample gives the best results, and best match between sections for any of the three spaced cleavage samples. There is a fair to good 90° match between the two sections, and both show maximum intensity near the pole to bedding orientation.
2. Both sections fit the March model very well, although the elongation values are slightly different. Mean shortening direction is nearly perpendicular to bedding, and the spaced cleavage orientation appears to be independent of the March strain axes.

4. Strain factoring was attempted by choosing tectonic axes reflecting regional structural trends: the tectonic flattening plane was assumed to be vertical, trending northwesterly, parallel to March elongation direction. A moderate pre-tectonic bedding plane compaction was obtained, although one of the cross-shear components is still fairly large.
APPENDIX C / X-ray Texture, March strain analysis

SAMPLE NUMBER 87-A108 (spaced cleavage)

Contour intervals 1–6:
26.5, 29.2, 31.8, 34.5, 37.2, 39.9

Estimated Strain Aves:
1 + e1 = *
1 + e2 = *
1 + e3 = *

A

B

Dots : 108-1YBCR-log; Crosses: Calc. Intensity: X : Tau, Y : Intensity
March Strch: [ 0.9569 0.7399 1.4123] RmsPoiEq = 6.89 Bline = 23.76 ChiSq = 23.76

March Strch: [ 1.3131 0.9711 0.7842] RmsPoiEq = 3.93 Bline = 19.23 ChiSq = 16.77
SAMPLE NUMBER 87-A108 (spaced cleavage)

Mean Stretch  \[ \{ 1.30 \ 0.97 \ 0.79 \} \]

Stretch Tensor
\[
\begin{pmatrix}
11651 & -0.1593 & 0.0377 \\
-0.1593 & 1.1032 & -0.0327 \\
0.0377 & -0.0327 & 0.7956
\end{pmatrix}
\]

(in geographic coordinates,
R1 = north
R2 = east
R3 = down)

\[ 1 + e_1 = 321/06 \]
\[ 1 + e_2 = 230/00 \]
\[ 1 + e_3 = 137/84 \]

Bedding Orientation: 340/30
Cleavage Orientation: 133/70

STRAIN FACTORING SOLUTION:

Estimated tectonic strain:
\[ \{ 1.35 \ 0.823 \ 0.900 \} \]
\[ 1 + e_1 = 140/00 \]
\[ 1 + e_2 = 050/00 \]
\[ 1 + e_3 = \text{vertical} \]

Factored compaction strain tensor,
in bedding plane coordinates
\[
\begin{pmatrix}
0.9623 & -0.0518 & 0.0011 \\
-0.0518 & 0.9718 & -0.0244 \\
0.0011 & -0.0244 & 1.0727
\end{pmatrix}
\]

R1 = strike
R2 = dip direction
R3 = bedding normal

NOTES:

1. Sections A and B have a poor 90° match. Although section A shows a symmetric pole figure, it is centered about a low intensity point, and the maximum intensity is outside of the area scanned. Section B shows that greatest intensity values are sub-parallel to the pole to bedding.
2. Sections A and B both show a reasonable fit to the March model, but the solution converged on for section A has strain axes switched from those visually chosen from the pole figure. The mean stretch recorded above therefore is taken from section B alone, and is considered only approximate because it is difficult to evaluate absorption corrections for these sections.

3. Strain factoring was attempted, using the same assumptions regarding tectonic strain directions noted for sample 88-379a. Although the factored strain represents a uniaxial strain with low values for cross shears, it suggests a negative compaction component. This inconsistency probably stems from the difficulties noted above — the incompatibility of one section with the March model, and the inability to evaluate absorption corrections.
SAMPLE NUMBER 88-ON2A (spaced cleavage)

Mean Stretch \( \{ 1.35 \; 0.92 \; 0.80 \} \)

Stretch Tensor
\[
\begin{bmatrix}
1.2895 & 0.1626 & 0.0351 \\
0.1626 & 0.9230 & -0.0486 \\
0.0351 & -0.0486 & 0.8634
\end{bmatrix}
\]

(in geographic coordinates, 
R1 = north 
R2 = east 
R3 = down)

\( 1 + e_1 = 021/02 \)
\( 1 + e_2 = 289/45 \)
\( 1 + e_3 = 112/45 \)

Bedding Orientation: 020/42
Cleavage Orientation: 175/30

STRAIN FACTORING SOLUTION:

Estimated tectonic strain:
\[
\begin{bmatrix}
1.4205 & 1.100 & 0.640
\end{bmatrix}
\]

1 + \( e_1 \) = vertical
1 + \( e_2 \) = 020/00
1 + \( e_3 \) = 110/00

Factored compaction strain tensor,
in bedding plane coordinates
\[
\begin{bmatrix}
1.2530 & -0.0593 & -0.0215 \\
-0.0593 & 1.2600 & -0.2043 \\
-0.0215 & -0.2043 & 0.6689
\end{bmatrix}
\]

R1 = strike
R2 = dip direction
R3 = bedding normal

NOTES:

1. This sample suffers from the same problem noted in sample 87-A108: section A is parallel to the maximum concentration of chlorite planes, and therefore the maximum intensity lies outside of the area scanned. The sample is best treated using data from section B alone; the strain values listed above reflect
this bias. Section B shows greatest intensity parallel to the bedding pole, and moderate elongation parallel to bedding strike.

2. Strain factoring for section B assumed a vertical tectonic flattening plane passing through the March elongation direction. This choice of axes resulted in a pre-tectonic strain reasonably close to a uniaxial compaction, except that one of the cross-shear components is large. The difficulty in minimizing the cross shears probably originates from the inability to assess absorption corrections when using only one section.
APPENDIX C / X-ray Texture, March strain analysis

SAMPLE NUMBER 88-MA1 (deformed ammonite)

Contour intervals 1-6:
19.3, 21.6, 24.0, 26.3, 28.6, 31.0

Estimated Strain Axes:
1 = ε1
1 = ε2
1 = ε3

Contour intervals 1-6:
19.9, 21.6, 23.3, 25.0, 26.7, 28.4

A

B

March Stack: [ 1.077 1.355 0.803] R(Sq) = 9.00 R(Q) = 13.59 Chisq = 28.81

A

B

March Stack: [ 1.368 0.919 0.736] R(Sq) = 4.62 R(Q) = 17.94 Chisq = 11.40
SAMPLE NUMBER 88-MA1 (deformed ammonite)

Mean Stretch \[ \begin{bmatrix} 1.37 & 1.00 & 0.73 \end{bmatrix} \]

Stretch Tensor \[ \begin{bmatrix} 1.0140 & -0.2493 & 0.0957 \\ -0.2493 & 1.1952 & 0.0884 \\ 0.0957 & 0.0884 & 0.8898 \end{bmatrix} \]

(in geographic coordinates,
R1 = north
R2 = east
R3 = down)

1 + e_1 = 124/02
1 + e_2 = 032/50
1 + e_3 = 216/40

Bedding Orientation: 309/59

NOTES:

1. Section B shows a typical, elliptical pole-figure, but section A shows a sharp increase in intensity in the northern hemisphere. This increase suggests preferential transmission of X-rays through the sample for some orientations, and probably is the result of hairline cracks through the sample.

2. Section B fits the March strain model quite well, while section A shows a lesser fit, probably due to the postulated sample defects. Elongation direction in section B is consistent with regional structural trends, and the strain magnitude is similar to that determined from distorted fossils in the same outcrop (Appendix B). The strain values reported above are based solely on results from section B.
C.4 STFACT.PAS Program Listing:

To ease the cumbersome strain tensor operations necessary for the strain factoring exercises, a PASCAL language program FACTOR.PAS was designed. The commented source code for this program is included below, as an aide to those workers desiring to analyze the procedure in more detail, or wishing to include any subroutines in other application programs.

```pascal
PROGRAM FACTOR.PAS;

{This program factors two strains, given the principal axis magnitudes and orientations of the second strain. If bedding orientation is known, it finds the bedding orientation at the end of the first strain event, and calculates the strain tensor for an axis system with 1 = bedding strike, 2 = bedding dip direction, and 3 = pole to bedding.}

type
tens = ARRAY[1..3,1..3] of REAL;  {strain tensors, rotation matrices}

VAR
    M,  {March strain tensor}
    CCS,  {Cleavage direction cosines}
    a,  {dummy direction cosines}
    P,  {predicted principal strains, user specified}
    RP,  {reciprocal predicted cleavage strains}
    TRP,  {transformed reciprocal strain tensor, map coords.}
    S,  {dummy transformed strain values}
    COM,  {computed factored strain}
    BSRM,  {Strained bedding rotation matrix}
    FS,  {rotated factored strains}
    MCS,  {March strain, in cleavage coords}
    CCSTR,  {transposed cleavage rotation matrix (transpose of CCS)}
    FMSM  {Factored strain, in map coords.}
    :tens;

    chk1,  {These are terms which check the fit of the iterations}
    chk2,  {when the original strain is known.  chk1 = E11-E22, }
    chk3,  {other values are non-diagonal tensor components }
    chk4
    :ARRAY[1..50] of REAL;
```
iter, \{iteration counter\}
i,j: INTEGER; \{tensor index counters\}

CS,CD, \{Cleavage strike and dip\}
BS,BD, \{Bedding strike and dip\}
BSC,BDC, \{Bedding strike and dip, strained\}
Plunge, \{plunge of principal strain in cleavage plane\}
rake, \{rake of 2nd principal strain axis in cleavage plane\}
doop
:REAL;

PC: ARRAY[1..50,1..3] of real;

ANS: CHAR;

Marchfile,OUTFILE:string[20];
\{input and output file names\}
ofile,mfile:text;

---

{Trigonometric Functions Defined}

FUNCTION ARCCOS(F:REAL):REAL;

BEGIN
  ARCCOS := PI/2 - ARCTAN(F/SQRT(1-SQR(f)));
END;

FUNCTION ARCSIN(F: REAL): REAL;

BEGIN
  ARCSIN := ARCTAN(F/SQRT(1-SQR(f)));
END;

FUNCTION SINE(F: REAL): REAL;

BEGIN
  SINE := SIN(F*(PI/180));
END;

FUNCTION COSINE(F: REAL): REAL;

BEGIN
  COSINE := COS(F*(PI/180));
END;
FUNCTION TAN(F:REAL):REAL;

BEGIN
  TAN := SINE(F)/COSINE(F);
END;

PROCEDURE INPUT;

{Prompts user for file name containing:
line 1: cleavage strike, dip (right hand rule)
line 2: bedding strike, dip (right hand rule)
lines 3-5: strain tensor, in map coordinates
  1 = north
  2 = east
  3 = down

Entries on the same line can be separated by any number of spaces}

BEGIN
  WRITELN('Enter file name containing MARCH strain tensor in map coords');
  READLN(marchfile);
  WRITELN('Enter plunge of principal strain in cleavage plane, measured from strike');
  READLN(plunge);
  ASSIGN(mfile,marchfile);
  RESET(mfile);
  READLN(mfile, CS, CD);
  READLN(mfile, BS, BD);
  FOR i := 1 to 3 do
    BEGIN
      FOR j := 1 to 3 do
        READ(mfile, M[i,j]); {reads combined strain tensor, in map coords}
        READLN(mfile);
    END;
  END;
END;

PROCEDURE directioncos(p,s,d:REAL);

{This procedure finds a 3X3 matrix of direction cosines which relate a new
set of axes to map coordinate axes.
Input parameters:
  s: strike of principal plane in new coord system, right hand rule.
  d: dip of plane in new coord system
  p: plunge of principal direction contained in above principal
  plane, measured from the "strike" end of plane.

New coordinate system has these axes:
VAR
az \quad \{ \text{azimuth of principal strain } "p" \}\quad: \text{REAL};

BEGIN
az := s + \arcsin(\tan(p)/\tan(d))*180/\pi;
\text{a[1,1]} := \cosine(az) * \cosine(p);
\text{a[1,2]} := \cosine(az-90) * \cosine(p);
\text{a[1,3]} := \cosine(90-p);
\text{a[3,1]} := (\cosine(s-90)) * (\cosine(90-d));
\text{a[3,2]} := (\sin(90-s)) * (\cosine(90-d));
\text{a[3,3]} := \cosine(d);
rake := \arctan((\sin(az-s)/\cosine(d)/\cosine(az-s)) \times 180/\pi);
doop := \arctan(\cosine(90 - (\arctan(\sin(doop) * \tan(d)) \times 180/\pi));
\text{a[2,3]} := \cosine(\arctan(\sin(az-s)/\cosine(az-s))) \times 180/\pi);
\text{a[2,2]} := (\text{a[1,1]}*\text{a[2,3]}*\text{a[3,3]} - \text{a[1,3]}*\text{a[3,1]}*\text{a[2,3]})/(\text{a[3,1]}*\text{a[1,2]}-\text{a[3,2]}*\text{a[1,1]});
\text{a[2,1]} := \sqrt{1 - \text{a[2,2]}} - \text{a[2,3]} \times \text{a[3,3]} \times \text{a[2,3]});
IF (s < 180) then
\text{a[2,1]} := -1 * \text{a[2,1]}; \quad \{ \text{changes sign of component according} \}
END; \quad \{ \text{to dip direction} \}

PROCEDURE clvgcosines (str,dip:REAL);

\{This procedure finds the direction cosine matrix for the cleavage or flattening plane of the second strain.\}

Input parameters:
\text{str} = \text{strike of cleavage plane in RHR}
\text{dip} = \text{dip of cleavage plane}

The actual calculations are completed in procedure directioncos\}

BEGIN
\text{directioncos(plunge,str,dip)};
FOR i := 1 to 3 do
FOR j := 1 to 3 do
BEGIN
\text{CCS[i,j]} := \text{a[i,j]};
\text{CCSTR[i,j]} := \text{a[j,i]}; \quad \{ \text{defines transpose of direction cosine matrix} \}
END;
END;
Procedure Transform (T, r: tens);

(This procedure transforms a tensor into a new coordinate system, given a
direction cosine matrix.

Input parameters:
T = tensor to be transformed
r = direction cosine matrix relating two coord. systems.)

BEGIN
FOR i := 1 to 3 do
BEGIN
FOR j := 1 to 3 do
S[i,j] :=

\[
\begin{align*}
& r[i,1] \times r[i,1] \times T[1,1] + r[i,1] \times r[i,2] \times T[1,2] + r[i,1] \times r[i,3] \times T[1,3] + \\
& r[i,2] \times r[i,1] \times T[2,1] + r[i,2] \times r[i,2] \times T[2,2] + r[i,2] \times r[i,3] \times T[2,3] + \\
& r[i,3] \times r[i,1] \times T[3,1] + r[i,3] \times r[i,2] \times T[3,2] + r[i,3] \times r[i,3] \times T[3,3];
\end{align*}
\]
end;
end;

PROCEDURE predictstrain;

(This procedure accepts user input for predicted strain values. It
automatically assumes constant volume strains, and thus only accepts
two strain values as input)

BEGIN
FOR i := 1 to 3 do
FOR j := 1 to 3 do
BEGIN
P[i,j] := 0.0; {Initializes all components as zero}
RP[i,j] := 0.0;
END;
WRITE('Enter predicted strain normal to cleavage plane (1 + e value)');
READLN(P[3,3]);
PC[iter,3] := P[3,3];
WRITE('Enter strain parallel to specified plunge in cleavage plane');
READLN(P[1,1]);
PC[iter,1] := P[1,1];
P[2,2] := (1/(P[1,1]/P[3,3]));
PC[iter,2] := P[2,2];
FOR i := 1 to 3 do
RP[i,i] := 1/P[i,i]; {Finds reciprocal principal strain values}
END;
PROCEDURE Combine;

{This procedure combines the components of two strain tensors.
One tensor must be set up along principal axes. Longitudinal strains
are added algebraically, cross strains are added (but are zero for
one tensor)

BEGIN
FOR i := 1 to 3 do
FOR j := 1 to 3 do
BEGIN
IF i = j then {longitudinal components}
COM[i,j] := RP[i,j] * MCS[i,j]
ELSE
COM[i,j] := MCS[i,j];
END;
END;

PROCEDURE correctbedding;

{This procedure calculates the new bedding orientation after the
second strain is removed. }

VAR
AD, {Apparent dip of bedding in E-W vertical section}
ADC {Apparent dip after strain correction}
:REAL;

BEGIN
BSC := arctan(tan(BS) / TRP[2,2] / TRP[1,1]) * 180 / PI; {Finds corrected strike}
AD := arctan(sine(BS) * tan(bd)) / PI;
ADC := arctan(tan(ad) / TRP[3,3] / TRP[2,2]) * 180 / PI; {corrects apparent dip}
BDC := arctan(tan(adc) / sine(BSC)) * 180 / PI; {finds corrected dip from}
IF BDC < 0 then {corrected apparent dip}
BEGIN
BSC := BSC + 180; {this corrects for non-uniqueness of}
BDC := -1 * BDC; {arctan values. it checks to insure}
END; {that right-hand-rule is maintained}
WRITELN;
WRITELN(' Corrected bedding orientation is ','BSC:4:1','BDC:4:1');
WRITELN;
END;

PROCEDURE listtensor(n:tens);
{This procedure lists the components of a 3X3 matrix passed
as parameter "n" to the screen}

BEGIN
FOR i := 1 to 3 do
BEGIN
FOR j := 1 to 3 do
WRITE(n[i][j]:8:6,');
WRITELN;
END;
END;

PROCEDURE compare;

{This procedure lists the values of the cross-strain components
and the difference between the first two principal strain values
for the factored strain in bedding coordinates, for the last three
iterations}

BEGIN
WRITELN;
WRITELN('STRAIN FIT PARAMETERS');
WRITELN('   E(1,1)-E(2,2)   E(1,2)   E(2,3)   E(1,3)');
IF ITER > 2 then
WRITELN('ITER #',iter-2,'   ',Chk1[iter-2]:5:4,
'   ',Chk2[iter-2]:5:4,'   ',Chk3[iter-2]:5:4,'   ',Chk4[iter-2]:5:4,
'   ',PC[iter-2,1]:4:3,'   ',PC[iter-2,2]:4:3,'   ',PC[iter-2,3]:4:3);
WRITELN('ITER #',iter-1,'   ',Chk1[iter-1]:5:4,
'   ',Chk2[iter-1]:5:4,'   ',Chk3[iter-1]:5:4,'   ',Chk4[iter-1]:5:4,
'   ',PC[iter-1,1]:4:3,'   ',PC[iter-1,2]:4:3,'   ',PC[iter-1,3]:4:3);
WRITELN('ITER #',iter,'   ',Chk1[iter]:5:4,
'   ',Chk2[iter]:5:4,'   ',Chk3[iter]:5:4,'   ',Chk4[iter]:5:4,
'   ',PC[iter,1]:4:3,'   ',PC[iter,2]:4:3,'   ',PC[iter,3]:4:3);
END;
APPENDIX C / X-ray Texture, March strain analysis

{ MAIN PROGRAM }

BEGIN
iter := 0; {initialize}
input; {accept input file name}
directioncos(0,bs,bd); {find direction cosines for bedding in input file}
transform(M,a); {finds strain tensor in bedding plane coordinates}

Writeln('BEDDING PLANE MARCH STRAINS:');
listtensor(S);
{lists to screen the strain tensor in bedding coordinates}
clvgcosines(CS,CD); {finds direction cosine matrix for cleavage (or principal axes), stores as CCS}

transform(M,CCS); {finds March strain tensor for cleavage coord. system}
FOR i := 1 to 3 do
 FOR j := 1 to 3 do
   MCS[i,j] := S[i,j]; {Stores cleavage strain tensor as MCS}

ANS := 'Y'; {initialize}

{The following section consists of program segments which can be repeated in a series of successive iterations, if one knows the early strain values and wants to match them}

REPEAT
ITER := ITER + 1; {counter}
PREDICTSTRAIN; {prompts user for principal strain values}
WRITELN('ITERATION #',ITER);
WRITELN;
WRITELN('RECIPROCAL STRAIN TENSOR');
listtensor(rp);
WRITELN;
transform(RP,CCSTR); {transforms reciprocal strain tensor into map coordinate system}
FOR i := 1 to 3 do
 FOR j := 1 to 3 do
   TRP[i,j] := S[i,j]; {stores transformed reciprocal strains}

Combine; {adds reciprocal strains (RP) to March strains (MCS) maintaining cleavage coordinate system}
WRITELN('FACTORED MARCH STRAIN TENSOR, in CLEAVAGE COORDS:');
listtensor(COM);
TRANSFORM(COM,CCSTR);  \{transforms combined (factored) strain into map coordinates\}

FOR i := 1 to 3 do
  FOR j := 1 to 3 do
    FMSM[i,j] := S[i,j];  \{assigns factored strains (map coords) to FMSM\}

WRITELN(‘FACTORED STRAIN, IN MAP COORDS:’);
  listtensor(FMSM);

Correctbedding;  \{Finds bedding orientation, restored after reciprocal strain added\}

directioncos(0,BSC,BDC);  \{Finds direction cosine matrix for restored bedding orientation\}

FOR i := 1 to 3 do
  FOR j := 1 to 3 do
    BSRM[i,j] := a[i,j];  \{defines matrix BSRM relating restored bedding axes to map directions\}

transform(FMSM,BSRM);  \{finds factored strain in bedding plane coordinate system\}

WRITELN(‘Factored strain in bedding plane coordinates:’);
  listtensor(S);
  FOR i := 1 to 3 do
    BEGIN
      FOR j := 1 to 3 do
        BEGIN
          FS[i,j] := S[i,j];  \{defines Tensor FS, containing factored strains in bedding plane coordinates\}
        END;
    END;
    WRITELN;
  END;

Chk1[iter] := FS[1,1] - FS[2,2];
Chk2[iter] := FS[1,2];
Chk3[iter] := FS[1,3];
Chk4[iter] := FS[2,3];

If Iter > 1 then
  compare;  \{copies a list of strain value checks to the screen for last three iterations\}

WRITELN(‘Do you want to run another iteration?’);
  READLN(ANS);
  UNTIL (ANS = ‘N’) or (ANS = ‘n’);
WRITELN(‘Save results to file?’);
  READLN(ANS);
  If (ANS = ‘Y’) or (ans = ‘y’) then
    BEGIN
      WRITELN(‘OUTPUT FILE NAME?’);
      READLN(OUTFILE);
ASSIGN(OF,OUTF);
REWRITE(OF);
WRTELN(OF,'Input file name = ',Marchfile);
WRTELN(OF);
WRTELN(OF,'Bedding orientation = ',BS:4:1,'/',BD:4:1);
WRTELN(OF,'Cleavage orientation = ',CS:4:1,'/',CD:4:1);
WRTELN(OF);
WRTELN(OF,'Compaction strain tensor, in bedding coordinates:');
FOR i := 1 to 3 do
  BEGIN
    FOR j := 1 to 3 do
      WRITE(OF,FS[i,j]:5:4,' ');
    WRTELN(OF);
  END;
WRTELN(OF,'Principal tectonic strains: [ ','P[1,1]:5:4,' ','P[2,2]:5:4,' ','P[3,3]:5:4,' ']');
END;
IF (ans = 'y') or (ans = 'Y') then
  CLOSE(OF);
end.
Appendix D: Fault Slip Analysis (Section Cove Shear Zone)

At Section Cove on northern Burnaby Island, a brittle shear zone trending 100°–110° cuts Mesozoic sedimentary and volcanic rocks. Mesoscopic shear fractures within the shear zone trend 090°–120°, display left lateral offset, and contain subhorizontal calcite slickenfibres. Fracture and slickenfibre orientation were measured for 13 of the most continuous mesoscopic fracture planes for analysis of stress tensor orientation (Table D.1, Fig. D.1). If all fractures are assumed to have moved in response to the same, homogeneous stress field, the slip direction for any given fracture records the maximum shear stress direction on a plane with that orientation. Measurement of a series of planes therefore gives a reasonable approximation of the overall stress tensor orientation, and can be used as an independent means of confirming the overall sense of movement on the major shear zone.

Fault slips were analyzed using the program FAULT KINEMATICS (Almendinger et al., 1989). This program analyzes each fault separately, determines the possible orientations for the maximum compressive stress direction for each fault slip data point, and statistically averages these for all fault planes.

<table>
<thead>
<tr>
<th>Fracture orientation</th>
<th>Slickenfibre orientation</th>
<th>Slickenfibre rake</th>
</tr>
</thead>
<tbody>
<tr>
<td>273/56</td>
<td>293/26</td>
<td>32</td>
</tr>
<tr>
<td>268/86</td>
<td>087/10</td>
<td>10</td>
</tr>
<tr>
<td>083/64</td>
<td>093/19</td>
<td>22</td>
</tr>
<tr>
<td>086/48</td>
<td>096/16</td>
<td>20</td>
</tr>
<tr>
<td>277/71</td>
<td>284/21</td>
<td>23</td>
</tr>
<tr>
<td>271/71</td>
<td>278/21</td>
<td>22</td>
</tr>
<tr>
<td>263/52</td>
<td>284/26</td>
<td>33</td>
</tr>
<tr>
<td>273/62</td>
<td>290/28</td>
<td>32</td>
</tr>
<tr>
<td>108/71</td>
<td>116/23</td>
<td>25</td>
</tr>
<tr>
<td>093/56</td>
<td>110/22</td>
<td>28</td>
</tr>
<tr>
<td>094/73</td>
<td>104/29</td>
<td>30</td>
</tr>
<tr>
<td>122/81</td>
<td>122/00</td>
<td>0</td>
</tr>
<tr>
<td>081/72</td>
<td>091/29</td>
<td>31</td>
</tr>
</tbody>
</table>

Figure D.1: Stereographic projection of shear fracture and slickenfibre data, Section Cove shear zone.
Results show a subhorizontal, northeast-trending "P" (compressive stress) axis, and a subhorizontal, northwest-trending "T" (tensile stress) axis (Figure D.2). These roughly correspond to maximum and minimum regional principal stress orientations. The inferred regional stress directions are consistent with a strike-slip stress field, favouring sinistral movement along the west-northwest-trending shear zone.

![Figure D.2: Average "P" and "T" dihedra and axes determined from mesoscopic shear fractures in the section cove shear zone.](image-url)
Appendix E: Field Station Compilation

Numerous field and sample locations are listed by number in the main text and appendices. Listed below are the positions of these localities in Universal Transverse Mercator (UTM) coordinates, the NTS map sheet number containing the area, the lithologic unit present, and the map area in which it is included.

Abbreviations used:

Map areas
bi Burnaby Island/Juan Perez Sound
lan Langara Island
lk Long Inlet/Kagan Bay
ls Louise Island
nwg Northwest Graham Island
rs Rennell Sound/Shields Bay
sk Skidegate Inlet/Skidegate Channel transect

Lithologic units
bp Buck Channel pluton
ha Haida Formation
ho Honna Formation
ka Karmutsen Formation
kv Cretaceous volcanic unit
ma Maude Group
ms Masset Formation
pe Peril Formation
sa Sandilands Formation
sd Sadler Limestone
sk Skidegate Formation
ya Yakoun Group

<table>
<thead>
<tr>
<th>Station number</th>
<th>Map area</th>
<th>Lithologic unit</th>
<th>Map sheet</th>
<th>UTM easting</th>
<th>UTM northing</th>
</tr>
</thead>
<tbody>
<tr>
<td>87-A108</td>
<td>lk</td>
<td>ha</td>
<td>103 F/1</td>
<td>685800</td>
<td>5897600</td>
</tr>
<tr>
<td>87-151</td>
<td>lk</td>
<td>ho</td>
<td>103 F/1</td>
<td>683900</td>
<td>5898100</td>
</tr>
<tr>
<td>87-163</td>
<td>lk</td>
<td>ho</td>
<td>103 F/1</td>
<td>681400</td>
<td>5898200</td>
</tr>
<tr>
<td>87-171</td>
<td>lk</td>
<td>ho</td>
<td>103 F/1</td>
<td>683400</td>
<td>5897500</td>
</tr>
<tr>
<td>87-369</td>
<td>nwg</td>
<td>pe</td>
<td>103 K/3</td>
<td>626100</td>
<td>5994400</td>
</tr>
<tr>
<td>87-409</td>
<td>nwg</td>
<td>pe</td>
<td>103 K/3</td>
<td>626100</td>
<td>5994500</td>
</tr>
<tr>
<td>87-431</td>
<td>lan</td>
<td>ha</td>
<td>103 K/3</td>
<td>630100</td>
<td>6008400</td>
</tr>
<tr>
<td>87-498</td>
<td>sk</td>
<td>sk</td>
<td>103 F/1</td>
<td>688300</td>
<td>5893200</td>
</tr>
<tr>
<td>87-502</td>
<td>sk</td>
<td>sk</td>
<td>103 F/1</td>
<td>688200</td>
<td>5892700</td>
</tr>
<tr>
<td>87-520</td>
<td>lk</td>
<td>kv</td>
<td>103 F/1</td>
<td>683100</td>
<td>5898200</td>
</tr>
<tr>
<td>Code</td>
<td>Type</td>
<td>Code</td>
<td>Type</td>
<td>Code</td>
<td>Type</td>
</tr>
<tr>
<td>--------</td>
<td>------</td>
<td>------</td>
<td>------</td>
<td>------</td>
<td>------</td>
</tr>
<tr>
<td>87-643</td>
<td>sk</td>
<td>ma</td>
<td>103 F/1</td>
<td>698900</td>
<td>5897000</td>
</tr>
<tr>
<td>87-665</td>
<td>sk</td>
<td>ma</td>
<td>103 F/1</td>
<td>696600</td>
<td>5898100</td>
</tr>
<tr>
<td>87-725</td>
<td>sk</td>
<td>ka</td>
<td>103 F/1</td>
<td>669200</td>
<td>5888900</td>
</tr>
<tr>
<td>88-MA1</td>
<td>sk</td>
<td>ma</td>
<td>103 F/1</td>
<td>696500</td>
<td>5898100</td>
</tr>
<tr>
<td>88-QN2A</td>
<td>sk</td>
<td>ha</td>
<td>103 G/4</td>
<td>303500</td>
<td>5901050</td>
</tr>
<tr>
<td>88-122</td>
<td>rs</td>
<td>pe</td>
<td>103 F/8</td>
<td>671070</td>
<td>5913820</td>
</tr>
<tr>
<td>88-207</td>
<td>rs</td>
<td>sd</td>
<td>103 F/8</td>
<td>671810</td>
<td>5910080</td>
</tr>
<tr>
<td>88-208</td>
<td>rs</td>
<td>sd</td>
<td>103 F/8</td>
<td>671810</td>
<td>5910210</td>
</tr>
<tr>
<td>88-279</td>
<td>rs</td>
<td>ma</td>
<td>103 F/8</td>
<td>670990</td>
<td>5909750</td>
</tr>
<tr>
<td>88-311</td>
<td>rs</td>
<td>ya</td>
<td>103 F/8</td>
<td>670850</td>
<td>5914700</td>
</tr>
<tr>
<td>88-313</td>
<td>rs</td>
<td>ya</td>
<td>103 F/8</td>
<td>670800</td>
<td>5914650</td>
</tr>
<tr>
<td>88-367</td>
<td>rs</td>
<td>sd</td>
<td>103 F/8</td>
<td>617800</td>
<td>5910080</td>
</tr>
<tr>
<td>88-369</td>
<td>sk</td>
<td>sa</td>
<td>103 F/1</td>
<td>694500</td>
<td>5895200</td>
</tr>
<tr>
<td>88-370</td>
<td>sk</td>
<td>ma</td>
<td>103 F/1</td>
<td>696500</td>
<td>5898100</td>
</tr>
<tr>
<td>89-13</td>
<td>--</td>
<td>ma</td>
<td>103 F/8</td>
<td>677950</td>
<td>5923700</td>
</tr>
<tr>
<td>89-ST1</td>
<td>--</td>
<td>sk</td>
<td>103 B/13</td>
<td>300800</td>
<td>5862550</td>
</tr>
<tr>
<td>90-219</td>
<td>bi</td>
<td>sk</td>
<td>103 B/6</td>
<td>339920</td>
<td>5805040</td>
</tr>
<tr>
<td>90-221</td>
<td>bi</td>
<td>sk</td>
<td>103 B/6</td>
<td>339600</td>
<td>5809870</td>
</tr>
<tr>
<td>90-233</td>
<td>bi</td>
<td>sk</td>
<td>103 B/6</td>
<td>338820</td>
<td>5810130</td>
</tr>
<tr>
<td>90-246</td>
<td>bi</td>
<td>sa</td>
<td>103 B/6</td>
<td>338430</td>
<td>5811660</td>
</tr>
<tr>
<td>90-258</td>
<td>bi</td>
<td>sk</td>
<td>103 B/6</td>
<td>339860</td>
<td>5806380</td>
</tr>
<tr>
<td>90-267</td>
<td>bi</td>
<td>lo</td>
<td>103 B/6</td>
<td>341250</td>
<td>5815980</td>
</tr>
<tr>
<td>Code</td>
<td>Type</td>
<td>Group</td>
<td>Box</td>
<td>Latitude</td>
<td>Longitude</td>
</tr>
<tr>
<td>-------</td>
<td>------</td>
<td>-------</td>
<td>-----</td>
<td>----------</td>
<td>-----------</td>
</tr>
<tr>
<td>90-272</td>
<td>bi</td>
<td>sa</td>
<td>103 B/6</td>
<td>340120</td>
<td>3814660</td>
</tr>
<tr>
<td>90-275</td>
<td>bi</td>
<td>pe</td>
<td>103 B/6</td>
<td>339840</td>
<td>5814210</td>
</tr>
<tr>
<td>90-278</td>
<td>bi</td>
<td>pe</td>
<td>103 B/6</td>
<td>339640</td>
<td>5814030</td>
</tr>
<tr>
<td>90-374</td>
<td>bi</td>
<td>ya</td>
<td>103 B/6</td>
<td>340740</td>
<td>5797410</td>
</tr>
<tr>
<td>90-433</td>
<td>bi</td>
<td>sd</td>
<td>103 B/6</td>
<td>345880</td>
<td>5799090</td>
</tr>
<tr>
<td>90-457</td>
<td>bi</td>
<td>pe</td>
<td>103 B/6</td>
<td>342580</td>
<td>5800390</td>
</tr>
<tr>
<td>90-542</td>
<td>bi</td>
<td>ka</td>
<td>103 B/6</td>
<td>341570</td>
<td>5794370</td>
</tr>
<tr>
<td>90-615</td>
<td>bi</td>
<td>ka</td>
<td>103 B/6</td>
<td>342120</td>
<td>5799530</td>
</tr>
<tr>
<td>90-648</td>
<td>bi</td>
<td>ka</td>
<td>103 B/6</td>
<td>339900</td>
<td>5810760</td>
</tr>
<tr>
<td>90-686</td>
<td>bi</td>
<td>sd</td>
<td>103 B/6</td>
<td>346750</td>
<td>5799250</td>
</tr>
<tr>
<td>90-840</td>
<td>sk</td>
<td>bp</td>
<td>103 F/1</td>
<td>675700</td>
<td>5884780</td>
</tr>
<tr>
<td>90-854</td>
<td>sk</td>
<td>bp</td>
<td>103 F/1</td>
<td>676150</td>
<td>5882320</td>
</tr>
<tr>
<td>90-939</td>
<td>bi</td>
<td>ka</td>
<td>103 B/6</td>
<td>349340</td>
<td>5798110</td>
</tr>
<tr>
<td>90-1049</td>
<td>bi</td>
<td>ka</td>
<td>103 B/6</td>
<td>341470</td>
<td>5795050</td>
</tr>
<tr>
<td>90-1195</td>
<td>--</td>
<td>ka</td>
<td>103 B/12</td>
<td>318200</td>
<td>5836050</td>
</tr>
<tr>
<td>90-1246</td>
<td>lk</td>
<td>sk</td>
<td>103 F/1</td>
<td>287380</td>
<td>5892800</td>
</tr>
</tbody>
</table>