STRUCTURAL GEOMETRY AND KINEMATICS OF DEFORMATION IN THE
STANLEY HEAD REGION, WESTERN CORNWALLIS ISLAND, NUNAVUT.

by

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Abstract

The Stanley Head region of western Cornwallis Island, Nunavut is located southeast and along strike from the Late Devonian Zn-Pb Polaris mine. The area incorporates Middle Ordovician - Lower Devonian strata that is folded and faulted and constitutes a part of the Lower Devonian Cornwallis fold belt.

1:15000 scale mapping of the Stanley Head region led to the identification of previously unrecognized shallow level folds and thrust faults that collectively define the Stanley Head anticline. These structures exhibit some of the characteristic elements of "evaporite detachment thin-skinned contractional deformation" (Davis and Engelder, 1985), and are interpreted to represent an evaporite based fold and thrust system in which there are multiple levels of detachment.

Three phases of deformation have been documented in the study area: 1) northeast and southwest verging folds and thrust faults associated with an east-west compressional D1 event, 2) strike slip faulting and folding associated with a north-south compressional D2 event, and 3) normal faulting associated with an east-west extensional D3 event. The D1, D2, and D3 events are interpreted to correspond with the Boothia Uplift, Ellesmerian Orogeny, and Eurekan deformational event respectively. Pervasive calcite veins related to the D2 and D3 events were precipitated from meteoric fluids based on oxygen and carbons stable isotope analysis.

To characterize the structural geology of the Stanley Head region six structural cross sections that transect the study area have been produced. "Balanced" cross sections were constructed using the integration of surficial data and a kinematic model of deformation for the multiple detachment, evaporite based Parry Islands fold belt
(Harrison and Bally, 1988). Cross sections from the study area provide a useful analogue to construct a regional cross section across Cornwallis Island. Both the regional and Stanley Head cross sections are characterized by 1) Southwest and northeast verging folds and thrust faults, 2) a variety of fold-thrust fault interactions, and 3) a proposed structural disharmony between the surficial fold belt and an underlying fold and thrust system.
# TABLE OF CONTENTS

Abstract ................................................. ii

Table of contents ....................................... iii

List of figures .......................................... vii

List of tables .......................................... viii

Acknowledgements ..................................... xiii

1 INTRODUCTION

1.1 Context of study..................................... 1

1.2 Organization of thesis.............................. 4

2 TECTONIC SETTING AND REGIONAL GEOLOGY

2.1 Tectonic setting of the Canadian Arctic Islands............................... 6
   2.1.1 Boothia Uplift...................................... 9
   2.1.2 Ellesmerian Orogeny................................. 13
   2.1.3 Eurekan Deformation................................ 17

2.2 Regional Geology: Cambrian to Lower Devonian Basin Development......... 18
   2.2.1 Cambrian - Lower Ordovician..................... 19
   2.2.2 Lower - Upper Ordovician......................... 19
   2.2.3 Silurian - Lower Devonian........................ 20

3 GEOLOGY OF THE STANLEY HEAD REGION, WESTERN CORNWALLIS ISLAND

3.1 Geomorphologic features of the study area................................. 22

3.2 Stratigraphy: Map and cross section rock formation descriptions........ 25

3.3 Structural Geology.................................... 31
   3.3.1 Introduction......................................... 31
   3.3.2 Physical conditions of deformation.................. 33
3.3.3 Stanley Head anticline and associated D1 structures..........................33
3.3.4 D2 Strike-slip faulting and superposed folding.................................49
3.3.5 D3 Normal Faulting...........................................................................62
3.3.6 Jointing and fracturing within the limestone formations of the
    Stanley Head anticline........................................................................64

4 DISCUSSION AND INTERPRETATION

4.1 Mechanics of evaporite fold and thrust belts.........................................73
   4.1.1 Rock strength..................................................................................74
   4.1.2 Critical taper of an evaporite based fold and thrust belt.................76
   4.1.3 Stress orientations and vergence of structures within fold and
       thrust belts.......................................................................................77

4.2 Review of the Parry Islands fold belt, Melville Island..............................78
   4.2.1 Characterization of structures of the Parry Islands fold belt,
       Melville Island.................................................................................81
   4.2.2 Kinematics of Deformation..............................................................84

4.3 Structural cross sections of the Stanley Head region, western Cornwallis
    Island..................................................................................................88
   4.3.1 Application of the Harrison and Bally model to the Stanley Head
       region, western Cornwallis Island..................................................89
   4.3.2 Methods of cross section construction.............................................92
   4.3.3 The balancing problem: cross sections A-A', B-B', C-C', D-D'.........101
   4.3.4 Results from cross sections A-A', B-B', C-C', D-D'.........................107
   4.3.5 Characterization of cross sections..................................................108

4.4 Fixed versus migrating hinge folding.....................................................109

4.5 Summary of the Stanley Head cross sections..........................................111

4.6 Regional cross section...........................................................................112
   4.6.1 Regional cross section discussion..................................................113
   4.6.2 Conclusion......................................................................................117
5 STABLE ISOTOPES STUDY

5.1 Introduction ........................................................................................................ 119

5.1.1 Spatial distribution of fracturing and veining within the Stanley Head anticline ........................................................................................................ 120

5.2 Stable isotope analysis

5.2.1 Analytical methods ......................................................................................... 124

5.3 δ¹⁸O and δ¹³C results ......................................................................................... 125

5.4 Discussion and Interpretation ........................................................................... 128

5.4.1 Discussion of δ¹⁸O values within the study area ........................................... 128

5.4.2 Interpretation of ¹⁸O depletion within wall rocks and calcite veins .......... 130

5.4.3 δ¹³C values .................................................................................................... 134

5.4.4 δ³⁴S value ..................................................................................................... 134

5.5 Comparison of O and C isotopic signature from the Stanley Head region to the Polaris mine site ................................................................................. 135

5.6 Conclusion .......................................................................................................... 137

6 CONCLUSIONS ......................................................................................................... 140

References .................................................................................................................. 143
## LIST OF TABLES

<table>
<thead>
<tr>
<th>Table</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.1</td>
<td>Peak burial temperatures and $R_0$ values from Little Cornwallis Island</td>
<td>92</td>
</tr>
<tr>
<td>4.2</td>
<td>Volume of Lower Bay Fiord Formation gypsum loss, Hills A-D</td>
<td>103</td>
</tr>
<tr>
<td>4.3</td>
<td>Volume of the upper compressional welt</td>
<td>107</td>
</tr>
<tr>
<td>4.4</td>
<td>Results from cross sections A-A', B-B', C-C', D-D''</td>
<td>108</td>
</tr>
<tr>
<td>4.5</td>
<td>Fixed versus active hinge folding</td>
<td>109</td>
</tr>
<tr>
<td>5.1</td>
<td>$^{18}$O and $^{13}$C values from calcite veins and corresponding wall rocks, Stanley Head region</td>
<td>127</td>
</tr>
</tbody>
</table>
# LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1</td>
<td>Location of the Polaris Zn-Pb-Cu district, central Arctic Islands</td>
<td>2</td>
</tr>
<tr>
<td>2.1</td>
<td>Geological provinces of the Canadian Arctic Islands</td>
<td>7</td>
</tr>
<tr>
<td>2.2</td>
<td>Generalized stratigraphic cross section, Prince of Wales Island to Melville Island</td>
<td>8</td>
</tr>
<tr>
<td>2.3</td>
<td>Tectonostratigraphic elements, cover structures, and major geographic features of the Boothia Uplift and Cornwallis fold belt</td>
<td>10</td>
</tr>
<tr>
<td>2.4</td>
<td>Gravity profiles and interpretive cross sections of the Boothia Uplift</td>
<td>12</td>
</tr>
<tr>
<td>2.5a</td>
<td>Surface geology of Melville Island</td>
<td>15</td>
</tr>
<tr>
<td>2.5b</td>
<td>Simplified cross sections of the Parry Islands fold belt, Melville, Island</td>
<td>16</td>
</tr>
<tr>
<td>2.6a</td>
<td>Upper Ordovician basin development, central Arctic Islands</td>
<td>21</td>
</tr>
<tr>
<td>2.6b</td>
<td>Distribution of Upper Ordovician to Lower Devonian shales and coeval carbonate rocks</td>
<td>21</td>
</tr>
<tr>
<td>3.1</td>
<td>Location of study area and defined limits of mapping</td>
<td>23</td>
</tr>
<tr>
<td>3.2</td>
<td>Major geomorphological features of the study area</td>
<td>24</td>
</tr>
<tr>
<td>3.3</td>
<td>Stratigraphic column of the Stanley Head region, western Cornwallis Island</td>
<td>26</td>
</tr>
<tr>
<td>3.4a</td>
<td>Bedding data: Southern Block</td>
<td>34</td>
</tr>
<tr>
<td>3.4b</td>
<td>Bedding data: Hill A</td>
<td>34</td>
</tr>
<tr>
<td>3.4c</td>
<td>Bedding data: Hill B</td>
<td>34</td>
</tr>
<tr>
<td>3.4d</td>
<td>Bedding data: Hill C</td>
<td>34</td>
</tr>
<tr>
<td>3.4e</td>
<td>Bedding data: Hill D</td>
<td>35</td>
</tr>
<tr>
<td>3.4f</td>
<td>Bedding data: Hill E</td>
<td>35</td>
</tr>
<tr>
<td>3.4g</td>
<td>Bedding data: Hill F</td>
<td>35</td>
</tr>
<tr>
<td>3.5a</td>
<td>Northeast verging 3rd order D1 fold to the east of Hill D</td>
<td>38</td>
</tr>
<tr>
<td>3.5b</td>
<td>Stereographic projection of fold in figure 3.5a</td>
<td>38</td>
</tr>
<tr>
<td>3.5c</td>
<td>Interpretive cross section of fold in figure 3.5a</td>
<td>38</td>
</tr>
</tbody>
</table>
3.6  a. Small scale third order fold located in the hinge zone of the Stanley Head anticline, Hill B. 39
b. Stereographic projection of the fold shown in figure 3.6a. 39

3.7  a. Small scale northeast verging thrust ramp within the Cape Phillips Formation to the west of Hill D. 41
b. Small scale fold associated with slip on a near horizontal detachment within the Cape Phillips Formation to the west of Hill D. 41
c. Stereonet projection of the fold shown in figure 3.7b. 41
d. Stereonet projection of small scale thrusts within the Cape Phillips Formation to the west of Hill D. 41

3.8  a. Small scale thrust within the Lower Thumb Mountain Formation, Hill C. 42
b. Stereonet projection of thrust plane, slickenlines, and bedding. 42
c. Stereonet projection of thrust plane and slickenlines when bedding is rotated to horizontal. 42

3.9  a. Fault breccia immediately adjacent to small scale northeast verging thrust fault within the Upper Bay Fiord Formation, Hill A. 43
b. Slickenlines associated with the small scale northeast verging thrust fault. 43
c. Stereonet projection of fault plane and associated slickenlines. 43

3.10 a. Overturned fold with the Cape Phillips Formation to the west of Hill F. 45
b. Hinge zone of an overturned syncline within the Cape Phillips Formation... 45
c. Steronet projection of poles to bedding within footwall structures west of Hill F. 45
d. Steronet projection of calculated and measured fold axes within the footwall structures to the west of Hill F. 45

3.11 Thrust fault truncating fold train within the Cape Phillips Formation, west of Hill F. 46

3.12 a. Eastern limb of the second order syncline to the east of Hill E. 48
b. Small scale z fold within the eastern limb of the second order syncline. 48
c. Stereonet projection of z fold observed within the eastern limb of the second order syncline. 48

3.13 a. Strike slip fault zone of the Southern Block. 50
b. Stereonet projection of the strike slip fault and associated slickenlines.......50

3.14 a. Extensional structures associated with strike slip faulting..........................52
    b. Contractional structures associated with strike slip faulting.....................52

3.15 Slickenline from D2 normal fault plane within the Southern Block ............54

3.16 a. Superposed fold axis data from the Upper Thumb Mountain Formation,
    Hill D........................................................................................................54
    b. Superposed fold axis data from the Cape Phillips Formation, to the west of
    Hill F........................................................................................................54

3.17 Well defined dome and basin structures within the superposed folded area,
    Hill D........................................................................................................56

3.18 Bedding orientations of superposed folded area within a fault bounded block
    of Upper Thumb Mountain Formation, Hill D...........................................57

3.19 a. Map view of footwall structures west of Hill F.................................58
    b. View of an overturned northerly verging fold, in part defining the area of
    superposed folding within the Cape Phillips Formation to the east of Hill F...58

3.20 Regional Type 1 interference pattern observed on Cornwallis Island.............61

3.21 Stereonet projection of normal faults and associated slickenlines from Hill C...63

3.22 Highly fractured Upper Thumb Mountain Formation, Hill E..........................63

3.23 Joint data from the Stanley Head anticline...............................................67-70

3.24 Maximum depth of tensile fracturing as a function of tensile strength and
    fluid pressure ratio...................................................................................71

4.1 Strength of evaporites and other common rock types as a function of depth,
    temperature, and strain rate........................................................................75

4.2 An illustration of why, in the presence of significant decollement strength,
    forward vergent slip planes dip more shallowly than do backward vergent
    planes...........................................................................................................79

4.3 Mohr-Coulomb diagram showing contrast in the plunge of the axis of maximum
    Compressive stress with respect to the basal decollement for
    a. A strong decollement.............................................................................79
    b. A weak evaporitic decollement..............................................................79
4.4  a. The orientations of candidate slip planes in a strong basal wedge .......... 80
    b. The orientations of candidate slip planes in a wedge with a weak evaporitic
detachment ............................................................................................................. 80
4.5  a. Vergence reversals along strike of a fold are characterized with an associated
    structural saddle, where fold axes exhibit a concave up plunge reversal ...... 83
    b. Relaying displacement transfer of subsurface thrusts and the formation of
    pop-up structures .................................................................................................. 83
4.6  Temporal evolution of an anticline within the multiple detachment Parry Islands
    fold belt, Melville Island ..................................................................................... 85
4.7  A comparison between the mechanical stratigraphy on Cornwallis and Melville
    Islands .................................................................................................................... 97
    a. Stratigraphic column of the Stanley Head region illustrating the mechanical
    stratigraphic units ................................................................................................. 97
    b. Stratigraphic column of the Parry Islands fold belt illustrating the mechanical
    stratigraphic units ................................................................................................. 97
4.8  Contrasting spatial distribution of deformation associated with fixed versus
    migrating hinge folding ....................................................................................... 110
4.9  Simplified geologic map of Cornwallis Island .................................................. 114

5.1  Representative vein types from the Stanley Head anticline ....................... 122
5.2  a. Simplified geologic map illustrating spatial distribution of samples for carbon
    and oxygen isotope study ................................................................................... 123
    b. Stratigraphic column illustrating formations sampled for isotope study ...... 123
5.3  $\delta^{13}$C - $\delta^{18}$O diagram showing the isotopic compositions of calcite veins and
    wall rocks in the Stanley Head anticline ............................................................. 126
5.4  Paleogeographic reconstruction of North America for the
    a. Late Devonian ............................................................................................... 132
    b. Late Cretaceous ............................................................................................ 132
5.5  Sulphur isotope values from sulphides and sulphates in the Polaris district .... 136
5.6  General paragenetic sequence of the various mineral and hydrothermal phases
    In the Polaris ore deposit ................................................................................... 136
5.7 $\delta^{13}$C - $\delta^{18}$O diagram showing the isotopic compositions of calcite veins and wall rocks from the Stanley Head region and the Polaris mine site.............. 138
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1 Introduction

1.1 Context of study

This study is a component of the joint Geological Survey of Canada and Canada-Nunavut Geoscience Office study of "The Regional Dynamics of the Polaris Zn-Pb System" undertaken during the summers of 2000 - 2002. The Polaris Zn-Pb-Cu district lies in the central Arctic Islands, Nunavut, and spans an area that is roughly 450 km north-south by 130 km east-west (Fig. 1.1). The objectives of this regional study were to establish a metallogenic model that describes and explains the regional aquifer systems, the structural controls that focused fluid flow, the nature of the mineralizing fluids, and the link between ore-forming process and tectonic events.

This thesis focuses on the structural geometry and kinematics of deformation within the Stanley Head region of western Cornwallis Island, in an attempt to gain an understanding of the structural controls that led to the creation of the plumbing system of the Polaris Zn-Pb mine.

The study area is located in the Stanley Head (68H/1) map area of western Cornwallis Island (Fig 1.1), southeast and along strike from the Polaris mine, a late Devonian Mississippi Valley type lead-zinc deposit on Little Cornwallis Island. This area incorporates Middle Ordovician - Lower Devonian strata that were folded and faulted to form the Lower Devonian aged Cornwallis fold belt (De Freitas & Mayr, 1993; Okulitch et al., 1986, 1991). Detailed 1:15000 scale mapping was undertaken over six weeks during the summer of 2001 to document the stratigraphy, structural geology, and possible structural controls on fluid migration within the study area.
Figure 1.1
Location of the Polaris Zn-Pb-Cu district in the central Arctic Islands, and the Stanley Head region of western Cornwallis Island. The rectangle defines the boundaries of the Polaris Zn-Pb-Cu district.
The first geological mapping of the region was completed in the 1950's (Thorsteinsson, 1958) by the Geological Survey of Canada. This work was a part of the GSC mapping program, that established the regional stratigraphic framework of the Canadian Arctic islands. A regional 1:250,000 scale map of Cornwallis Island was produced in the 1980's (Thorsteinsson, 1988), that documented the stratigraphy and provided a preliminary interpretation of the structural geology of the island. The structures of Cornwallis Island were described as broad open synclines separated by tighter intervening anticlines that were faulted by normal, and less commonly, steeply dipping reverse faults. Folding of the strata was interpreted to be kinematically linked with this faulting. This interpretation of the structural geology of Cornwallis Island has persisted until the present round of GSC mapping, of which this study was a part.

Mapping by the author has determined that the structural geology of the Stanley Head region of western Cornwallis Island is characterized by thin-skinned fold and thrust structures that have resulted due to slip on low strength evaporitic detachments. The structures found in the area of study have some of the characteristic elements of evaporite based thin-skinned contractional deformation described by Davis and Engelder (1985), and are best explained and interpreted using a kinematic model of deformation developed by Harrison and Bally (1988) for the Parry Islands fold belt on eastern Melville Island. The Harrison and Bally model was used because of the similar mechanical properties of the stratigraphic columns on Melville Island and western Cornwallis Island, in which there are two potential weak detachment layers.

To characterize the structural geology of the Stanley Head region, a 1:15000 scale geologic map, and six structural cross sections that transect the study area have been
prepared. Stereometric analysis of a detailed structural data set, which included planar and linear data from first, second, and third order structures provide insight into the structural geometry across the study area.

1.2 Organization of thesis

The main of objectives of this research are to 1) produce a 1:15000 geologic map that documents the geometry of the structures and the stratigraphy of the study area, 2) produce balanced cross sections across the study area that kinematically model the temporal evolution of the fold and thrust structures, and 3) delineate the origin of fluids found in the study area by means of isotopic studies.

Chapter 2 provides a brief review of the tectonic setting and regional geology as they relate to the area of study.

Chapter 3 presents the structural geology and stratigraphy of the Stanley Head region of western Cornwallis Island. A 1:15000 geologic map with detailed legend and accompanying cross sections have been accepted for publication as a GSC open file.

Chapter 4 presents balanced and schematic cross sections from across the area of study. The objectives of this chapter are to present the structural geometry of the fold and thrust structures that characterize the study area, and to present balanced cross sections that employ a kinematic model of deformation.

Chapter 5 presents the results of a preliminary isotopic study that uses carbon, oxygen, and to a lesser degree sulphur isotopes to delineate the origins of the fluids that precipitated as veins within the study area.
Chapter 6 presents the conclusions of the thesis and how they relate to the goals of the GSC - Canada-Nunavut Geoscience Office study "The Regional Dynamics of the Polaris Zn-Pb system".
CHAPTER 2

TECTONIC SETTING AND REGIONAL GEOLOGY

2.1 Tectonic setting of the Canadian Arctic Islands

The tectonic provinces of the Canadian Arctic islands were first developed and described by Fortier et al. (1954) and then by Thorsteinsson and Tozer (1970), whose interpretation has served as the basis for all subsequent syntheses. The major elements of the present day tectonic framework are the Canadian Shield, Arctic Platform, and Innuitian Tectonic Province (Fig. 2.1).

The Canadian Shield borders the Arctic Platform on the southeast and extends into it as salients and promontories. It consists of a metamorphic-plutonic basement and unconformably overlying sedimentary and volcanic successions. The Arctic Platform is comprised of Phanerozoic sedimentary successions that have not been folded or thrust faulted on a regional scale (Thorsteinsson and Tozer, 1970). The Innuitian Tectonic Province lies along the northern margin of the Arctic Platform and is characterized by sinuous belts of folded sedimentary rocks. Its southern limit corresponds with the limit of Phanerozoic folding and thrust faulting and its northwestern margin with the transition between continental and Mesozoic - Tertiary oceanic crust (Trettin, H.P., 1991). The following discussion will focus on the Innuitian Tectonic Province and the tectonic events that affected the central Arctic Islands.

Three major tectonic events have affected the Paleozoic sedimentary succession (Fig 2.2) of the central Arctic Island Archipelago from the Cambrian to the present. They are the Boothia Uplift, the Ellesmerian Orogeny, and the Eurekan deformational event (Fig. 2.1).
Figure 2.1 Geological provinces of the Canadian Arctic Islands (Trettin, 1991).
Figure 2.2
Generalized stratigraphic cross-section, Prince of Wales Island to Melville Island (Trettin et al., 1991).
2.1.1. Boothia Uplift

The Boothia Uplift is interpreted as a "Laramide-type" basement cored uplift that affected Precambrian crystalline basement and the lower Paleozoic carbonate platform sequence of the Central Arctic Islands (Kerr, 1977; De Freitas & Mayr, 1993). It extends north from the Boothia Peninsula, approximately 1000 km to the Grinell Peninsula on Devon Island (Fig. 2.3). The Boothia Uplift is divided into southern and northern segments that are separated by Barrow Strait (Fig 2.3). The southern two-thirds of the uplift was activated in Late Silurian time (De Freitas & Mayr, 1993) and contains the lower structural level, that is characterized by faulted poly-deformed basement rocks of Proterozoic age. The northern third of the uplift contains the upper structural level, that is characterized by strata of Cambrian - Lower Devonian age. This sedimentary succession was folded and faulted in the Early Devonian to form the Cornwallis Fold Belt (De Freitas & Mayr, 1993; Okulitch, Packard, & Zolnai, 1986).

The Boothia Uplift is interpreted to be a deep-seated eastern dipping thrust block, that may be comparable to the uplifts in the Wyoming province (Kerr, 1977; Miall, 1986; Okulitch et al., 1986), in which uplift of basement blocks was accommodated through thrust faulting (Smithson et al., 1978, 1979). Miall (1986) and Okulitch et al. (1986) suggest that this "thick skinned" fold and thrust model may be kinematically linked to horizontal compression that reactivated preexisting basement weaknesses during the closing pulses of the Ordovician to Early Devonian Caledonian orogeny, more than 1000 km to the east. Sanford et al. (1985) suggested that compressional stresses generated at the margins of Laurentia during the Caledonian orogenesis could have reactivated pre-
Figure 2.3
Tectonostratigraphic elements, cover structures and major geographic features of the Boothia Uplift and Cornwallis fold belt. (From Okulitch et al., 1986 & 1991)
existing fault bounded basement blocks within the cratonic interior. This interpretation is supported by gravity data (Berkhout, 1973; Sobczak et al, 1986; Sobczak & Halpenny, 1990) and structural mapping (Mortenson and Jones, 1986; Stewart and Kerr, 1984).

The west side of the Boothia Uplift on Prince of Wales Island is characterized by reverse faults and an approximately 2 km wide belt in which Paleozoic strata are vertical to overturned in orientation (Miall, 1986). Mortenson and Jones (1986) recognized east dipping reverse faults that thrust Precambrian basement over Paleozoic sedimentary rocks on Prince of Wales Island. Gravity data from the southern segment (Sobczak and Halpenny, 1990), although inconclusive, suggest the presence of low-density sediments below basement rocks in Peel Sound, indicating basement thrusting over Paleozoic rocks (Fig 2.4a - 2.4c). This interpretation is supported by the presence of west verging folds and east dipping reverse faults on Prince of Wales Island, the presence of thrust faults west of the uplift, and possible basement wedges along its eastern margin on the Boothia Peninsula (Okulich et al, 1991). Gravity data from the northern segment of the Boothia uplift from Bathurst and Devon Islands also suggest east dipping basement thrust faults (Fig 2.4d, 2.4e) (De Freitas and Mayr, 1993). Furthermore, in the northern third of the uplift where the basement is not exposed, asymmetrical, thrust related structures are observed on eastern Bathurst Island (Okulitch, 1991, Harrison, J.C, pers. comm., 2002) and western Cornwallis Island within the Paleozoic sedimentary succession (Henrichsen & Kennedy, 2002).

The interaction between the basement structures of the Boothia Uplift and the Paleozoic sedimentary succession are at present poorly understood, as structural studies of the Cornwallis fold belt have been undertaken only at the reconnaissance level, leaving
Figure 2.4
their geometries and kinematics of deformation poorly understood. However, structures within the fold belt are characterized by broad open synclines with tighter intervening anticlines with northwest-northeast trends and numerous steep normal, and less commonly reverse faults (Okulitch, et al, 1991).

### 2.1.2 Ellesmerian Orogeny

The Ellesmerian Orogeny formed during the Late Devonian to Early Carboniferous and is characterized by the Canrobert and Parry Islands fold belts (Fig 2.1). The Canrobert Fold Belt is exposed on northwestern Melville Island and continues to the west and north underneath the younger rocks of the Eglinton Graben and Sverdrup Basin, respectively (Harrison, et al 1991). The Parry Islands Fold Belt extends from northwestern Melville Island to eastern Bathurst Island, where it abuts against the north-trending Cornwallis Fold Belt. To the west the belt continues to southeastern Prince Patrick Island, with southern limits being gradational as the belt passes into relatively undeformed Cambrian to Devonian strata of the Arctic Platform (Harrison et al., 1991).

The Parry Islands Fold Belt can be divided into two segments. The western segment, exposed on central and western Melville Island, involves most if not all of the lower and middle Paleozoic succession, with the existence at depth of a basal detachment surface remaining uncertain. In contrast, the eastern segment, exposed on eastern Bathurst and Melville islands, is characterized by a thin-skinned detachment style of deformation (Harrison et al., 1991).
Structures within the eastern segment of the Parry Islands Fold Belt are characterized by a southerly directed salt based detachment style fold and thrust system (Harrison & Bally, 1988; Harrison, 1995). The deformation style is typical of a weak salt-based decollement system for which a near horizontal compressive stress is responsible for slip along a near horizontal detachment horizon (Harrison, 1995). Thrust faults verge both to the north and south and are concentrated below surface anticlines that are cored by Middle Ordovician Bay Fiord Formation salt (Harrison, 1985).

Regional scale surface folds are typically characterized by symmetrical narrow anticlines with strike lengths up to 150 km in length, wavelengths of 12-15 km, amplitudes of up to 3 km, and limb dips ranging from 30-45°. These anticlines are separated by broad synclines that are 2.5 - 3 times wider than the anticlines (Fox, 1985). Figures 2.5a and 2.5b depict a representative geological map and accompanying cross sections from southeastern Melville Island and illustrates the surface and subsurface geological features described above.

The easternmost margin of the Parry Islands Fold Belt located on eastern Bathurst Island intersects and overlaps the north-trending Cornwallis Fold Belt. This area is characterized by dramatic Type 1 interference structures (Temple, 1965; Kerr 1974), and by a pronounced structurally disharmonic relationship between folds above and below the Bay Fiord Formation salt (see fig.2.2) (Fox, 1985). From seismic data it is observed that structures present at the Thumb Mountain Formation level belong to the east striking Parry Islands Fold Belt, whereas, structures present at the base of the Bay Fiord Formation salt belong to the north striking Cornwallis Fold Belt (Fox, 1985). This disharmonic relationship supports the inference that the two fold belts are of different
Integration of seismic, well, and surficial mapping data (Harrison, 1995).

Figure 2.5b Simplified structural cross section of the Peary Islands fold belt, Melville Island. Cross sections were produced from the
ages. The mobility of the Bay Fiord evaporite enabled structural deformation of Ellesmerian age to be limited to strata above the Bay Fiord evaporite unit, leaving the older structures beneath undistorted (Fox, 1985).

In the area of interference between the Cornwallis and Parry Island fold belts some Cornwallis fold belt structures remain undistorted at the surface in the Lower to Middle Devonian strata despite the subsequent deformation associated with the Ellesmerian orogeny. Kerr (1974) reasoned that the southward translation of rocks characterizing the Ellesmerian fold and thrust deformation was accomplished by strike slip faulting within the Cornwallis Fold Belt, that allowed the most westerly Cornwallis Fold Belt anticlines to be moved southward, with only minor distortion from the growing Ellesmerian folds.

2.1.3 Eurekan Deformation

The Eurekan deformational event is an orogenic and extensional event that occurred from the Late Cretaceous - Early Tertiary and is characterized by compressive deformation in the northeastern Arctic islands and extensional deformation in the central Arctic Islands (Fig 2.1) (Okulitch & Trettin, 1991). Eurekan deformation has been attributed to the counterclockwise rotation and northwestward translation of Greenland relative to North America, as a result of seafloor spreading in the Labrador Sea and adjacent seaways (Kerr, 1967; Peirce, 1982; Miall, 1984). Throughout the Arctic Islands Eurekan structures are generally parallel in trend with Ellesmerian and Boothia structures consistent with reactivation of older structures (Okulitch and Trettin, 1991).
The compressive orogenic deformation is generally characterized by fold and thrust style deformation that is divided into four provinces (see Okulitch and Trettin, 1991, for discussion). However, the orogenic phase of the Eurekan deformation will not be discussed further as it does not effect the central Arctic Islands and thus the structural geology of the area of study.

To the south of the Eurekan Orogen, the central Arctic Islands are characterized by normal faults of Early Cretaceous - Early Oligocene age. Although their specific ages and modes of origin are difficult to determine, it is postulated that they developed in response to the development of the Labrador Sea and Baffin Bay (Okulitch and Trettin, 1991). Structures of the Boothia Uplift and Cornwallis Fold Belt were affected by the Eurekan deformational event through graben formation. In particular, on Cornwallis and western Devon islands, several north-northwest and north-northeast trending grabens containing Tertiary strata were formed parallel in trend to the older Boothia structures (Okulitch and Trettin, 1991). The western extent of the extension is postulated to be on southeastern Bathurst Island where rifting occurred with associated magmatism. Isotopically dated alkaline rocks indicate that the rifting occurred in early Eocene time (Okulitch and Trettin, 1991)

2.2 Regional Geology: Cambrian to Lower Devonian Basin Development

The following discussion focuses on the basinal development of the present day Arctic Platform and the Inuitian Tectonic provinces in relation to the central Arctic Islands. The lower Paleozoic sedimentary succession of these provinces is presented in figure 2.2.
2.2.1 Cambrian to Lower Ordovician

Flexural subsidence of the passive continental margin, that followed Neoproterozoic rifting of Rodinia resulted in a prolonged period of tectonic stability in the Arctic Islands. Transgression over Proterozoic rocks began in Early Cambrian time, with three dominantly clastic, transgressive-regressive packages of rock that have been recognized on Ellesmere Island. These packages of rock correspond roughly with the Early, Middle, and Late Cambrian times. Approximately 1.5 km of Cambrian strata have been intersected in drill core on Cornwallis Island, although the total stratigraphic thickness on the island remain unknown.

2.2.2 Lower - Upper Ordovician

During Ordovician time there were three major depositional belts with somewhat unstable boundaries present in the Arctic Islands. A northwestern rim, an intrashelf basin, and a southern shelf (Fig 2.6a). The intrashelf basin is not interpreted as a structural feature, but as a transient paleobathymetric depression caused by differential sedimentation rates (Trettin et al, 1991). At this time Cornwallis Island was contained entirely within this intrashelf margin and saw the Baumann Fiord, Eleanor River, Bay Fiord, Thumb Mountain, and Irene Bay formations deposited in a conformable sequence (Fig 2.2). The evaporites of the Baumann Fiord, Bay Fiord formations were deposited within restricted circulation, shallow water, saline conditions. Breaching of the carbonate rim to the north and northwest as a result of a rise in sea level, promoted full marine circulation conditions and the limestones/shales of the Eleanor River, Bay Fiord, Thumb Mountain, and Irene Bay formations were deposited.
2.2.3 Silurian - Lower Devonian

During the latest Ordovician time, two major processes brought about a fundamental change in the sedimentary regime. The first is a rise in seal level that caused a major transgression during which eventually most of the continent was flooded. The second process was the rapid subsidence of the outer parts of the shelf, exhibited by a retreat of the shelf margin in eastern Melville, Bathurst, central Cornwallis, and central western Ellesmere Islands, where deep water sediments were deposited (Fig 2.6b). This created a facies front that runs east-west across Cornwallis Island, that separated shallow shelf carbonates to the south from organic rich restricted circulation shales and minor carbonates to the northwest. Carbonate buildups along the old shelf margin to the northwest kept pace with basal subsidence, forming a restricted basin between the shelf and outer reef. South of the facies front, shallow shelf, fossiliferous limestone and dolostone of the Allen Bay and overlying Cape Storm, Douro, and Barlow Inlet formations span latest Ordovician to earliest Devonian time. The restricted basin facies to the north of the front contains the Cape Phillips Formation, that is comprised of mainly of calcareous shale and minor limestone units.
Upper Ordovician basin development central Arctic Islands: Distribution of Upper Ordovician (Caradocian) evaporites and laterally equivalent carbonate rocks. Redrafted from Fox (1985) and Trettin et al. (1991).

CHAPTER 3
GEOLOGY OF THE STANLEY HEAD REGION, WESTERN CORNWALLIS ISLAND

The study area defined within this investigation (Fig. 3.1) will be characterized by the following: 1) Brief description of prominent geomorphological features, 2) Detailed descriptions of the lithological formations, and 3) Detailed description of the exposed structural geology.

3.1 Geomorphic features of the study area

The Stanley Head region of western Cornwallis Island is characterized by two distinct geomorphological regions. The first region is defined as a linear series of northwest trending hills, that have a length and width of 11 km and 2-3 km respectively. These hills exhibit from 50-150 meters topographic relief with respect to the surrounding topography of 25-50 meters above sea-level. Exposure of rock formations on the hills is fair to good, allowing for detailed geologic mapping. The areas surrounding the northwest trending hills are composed of flat to undulating glacial cover, with little to no exposure of rock formations. Streams in these areas have eroded 10-30 meter deep channels that produce variable degrees of outcrop exposure. These hills collectively constitute the Stanley Head anticline (Fig. 3.2) (Henrichsen & Kennedy, 2002) and provide the focus of this thesis.

The second geomorphological region is defined as a northwest trending topographic depression, that has a length of 11 km and a width of 2-3 km. The depression is on average 20 - 40 m above sea level and is bounded by 15 -30 m cliffs that lead up to the surrounding peneplain. Streams within the depression have eroded 2 -10 m
Figure 3.1 Location of study area and defined limits of mapping.
Figure 3.2
Major geomorphological features of the study area. The red lines outline the north north-west trending series of hills that, in part, comprise the Stanley Head anticline. The yellow lines outline an area of topographic low relief comprised of gypsum, this feature is a part of the Stanley Head anticline. The light blue lines outline the Intrepid Bay graben, a northwest trending valley.
deep channels that produce variable degrees of outcrop exposure. This depression comprises the Intrepid Bay Graben (Fig 3.2) (Thorsteinsson, 1988).

3.2 Stratigraphy: Map and cross section rock formation descriptions

The formations found within the study area were first formally defined by Thorsteinson (1958) and Kerr (1967b). Informal subdivisions employed by exploration companies are described by Héroux et al. (2000). For this research, I have subdivided the formations and modified their descriptions to reflect the rocks exposed in the area of study. Formations exposed within the study area are the Bay Fiord, Thumb Mountain, Irene Bay, Allen Bay, Cape Storm, Cape Phillips, Barlow, Bathurst Island, and Eureka Sound formations. Formations not exposed at the surface of the study area, but inferred to occur at depth are the Baumann Fiord and Eleanor River formations. Figure 3.3 illustrates the stratigraphic section found within the area of study.

Unless otherwise noted, unit thicknesses for strata in the Stanley Head area have been calculated, based on areas of more or less continuously exposed stratigraphy, using the standard structure contour three-point problem. Calculated thickness for the Eureka Sound Formation represents minimum thickness of this formation in the Intrepid Bay Graben.

*Baumann Fiord Formation: Lower Ordovician*

The Baumann Fiord Formation on Cornwallis Island is composed predominantly of gypsum and anhydrite, with minor limestone and intraformational conglomerate (750 m, Thorsteinsson, 1988).
Figure 3.3. Stratigraphic column of the Stanley Head Region, western Cornwallis Island.
Eleanor River Formation: Lower - Middle Ordovician

The Eleanor River Formation on Cornwallis Island is composed of a variety of limestone types with smaller amounts of dolostone and very minor amounts of anhydrite. Typical subtidal limestones are medium grey, bioturbated and dolomitic, and contain varying amounts of skeletal material and fecal pellets. Shallow subtidal and intertidal limestones are characterized by stromatolitic limestones and minor intraformational conglomerates (Trettin et al. 1991). The Eleanor River Formation has been subdivided into five members based on their resistance to erosion. From the bottom moving upward they are: member A resistant, member B recessive, member C resistant, member D recessive, and member E resistant (700 m, Thorsteinsson 1988).

Bay Fiord Formation: Middle Ordovician

The Bay Fiord formation is here divided into two units. In ascending order of age these are:

Lower Bay Fiord Formation:

The lower Bay Fiord Formation is composed entirely of foliated gypsum that is characterized by alternating white and dark grey-brown layers that are 2 - 15 mm in thickness. White bands consist of fine-grained crystalline gypsum, whereas dark bands are composed of fine-grained crystalline gypsum with greyish-brown argillaceous material. These rocks are typically extremely weathered and incohesive. Areas underlain by the Lower Bay Fiord Formation rocks are topographically low and wet (200 m).

Upper Bay Fiord Formation:

The Upper Bay Fiord Formation is composed of lime mudstone and skeletal wackestone in beds typically 5-50 cm thick. Lime mudstone and wackestone contains 10-
30% brachipod, gastropod, and crinoid fragments 5-40 mm in size. Rough, weathered surfaces are tan to light grey with variable red-brown and exhibit variable red-brown, irregular burrow mottling. Mottles are predominantly calcitic, 10 - 30 mm in diameter, and form 20-35% of rock volume. Fresh surfaces are medium to dark grey. Common shaley partings are 1.5 - 3 mm thick and laterally continuous (230 m).

**Thumb Mountain Formation:** Middle - Upper Ordovician

**Lower Thumb Mountain Formation:**

The Lower Thumb Mountain consists of fossiliferous lime mudstone in beds 5 - 30 cm thick. The rock is composed of a dark to medium grey micritic lime mud and 2-10% trilobite, crinoid, and coral fragments from 1-3 mm in size. Weathered surfaces are rough and light gray in colour with variable light brown irregular burrow mottles. Mottles are predominantly calcitic, 1-5 mm in diameter and form 5-15% of rock volume. They are rarely replaced by chert in laterally discontinuous horizons. Fresh surfaces are medium to dark gray in colour and reveal a locally present poorly developed fenestral fabric. Common laterally discontinuous argillaceous partings parallel to bedding are 1-2 mm in thickness (130 m).

**Upper Thumb Mountain Formation:**

This Upper Thumb Mountain Formation is informally subdivided into subunits A and B, based on the modal abundance and size of fossils found within the massively bedded (metre-scale) chert-bearing, fossiliferous wackestone or lime mudstone (180 m).

Subunit A consists of 95% dark to medium grey lime mudstone with 5% fossil fragments. Fossils include crinoid and coral fragments that are calcitic, white, and typically 1-3 mm in size.
Subunit B consists of 60-85% dark to medium grey lime mudstone and 15-40% fossil fragments. Fossil fragments include solitary and colonial corals, brachiopods, gastropods, cephalopods, and rare stromatoporoids, which are 5 mm to 15 cm in size and locally silicified. Fossil fragments replaced by chert are light brown-cream colour and display pronounced relief on exposed bedding planes. Calcitic fossil fragments are white. Weathered surfaces of the Upper Thumb Mountain Formation are typically very rough, pitted, and uneven, with light brown, irregular mottling. Mottles are predominantly calcitic, with a minor dolomitic component. Mottles are 1-20 mm in diameter and form 10-25% of rock volume. Fresh surfaces are medium to dark grey.

Irene Bay Formation: Upper Ordovician

Very recessively weathered green shale, with no outcrop in the map area. Its identification is based on its stratigraphic relationship with the underlying Upper Thumb Mountain Formation and the overlying Cliff member of the Cape Phillips Formation (60 m).

Cape Phillips Formation: Upper Ordovician - Upper Silurian

Predominantly calcareous shale and siltstone with minor limestone units, this formation is here informally divided into two units, the Cliff member and the Cape Phillips shale.

Cape Phillips shale: Calcareous, grey to black siltstone and shale. These rocks are commonly millimetrically laminated and contain concretions 1 cm to 1 m in diameter. Rare intraclasts and scour surfaces are present. A trilobite-rich horizon that directly overlies the Cliff unit is used as a marker bed (1000 m).
**Cliff member:** Massively bedded fossiliferous wackestone. 10 - 20% of the rock volume consists of 5 – 25 mm crinoid, brachiopod and trilobite fragments. Weathered surfaces are orange-tan and typically very rough, pitted, and uneven; fresh surfaces are dark to medium grey. Mottles are predominantly calcitic, 5 – 25 mm in diameter, and form 15-25% of the rock volume. Minor calcite- and dolomite-filled vugs locally form up to 5% of the rock (10 m).

**Allen Bay Formation:** Upper Ordovician - Upper Silurian

Coarsely to very coarsely crystalline, massively bedded dolostone. Bed thickness ranges from 50 cm to 10 m. The dolostone is generally characterized by vugs that constitute 15-20% of the rock volume; 80% of the vugs are 1-3 cm in diameter and 20% of the vugs are 5-15 cm in diameter. Weathered surfaces are buff to very light brown; fresh surfaces are almost white. The dolostone contains minor fossil fragments including gastropods and stromatoporoids, and is locally stained dark brown by hydrocarbons (1200-1375 m; Thorsteinsson, 1988).

**Cape Storm Formation:** Upper Silurian

Highly fossiliferous lime wackestone and packstone that is thinly bedded with cyclic resistant and recessive beds. Fossils include abundant crinoid, brachiopod, and coral fragments. Resistant beds range from 30-50 cm thick, whereas recessive beds are 1-2 m thick. Weathered surfaces are rough, light grey to greenish-brown, commonly with greenish-brown mottles. Fresh surfaces are dull grey. Mottles are predominantly calcitic and constitute 15-30% of the rock volume. Minor calcite-filled vugs constitute up to 3% of the rock (520 -670 m; Thorsteinsson, 1988).
Barlow Inlet Formation: Upper Silurian - Lower Devonian

Highly fossiliferous, thinly bedded skeletal grainstone with abundant crinoids, brachiopods, and corals. Beds are 20-100 cm thick. Weathered surfaces are light to medium grey; fresh surfaces are dark grey (50 m; Dewing, pers. com. 2002).

Bathurst Island Formation: Lower Devonian

Silty litharenite composed of 70% calcareous grains and 30% silt. Grains are angular to subangular, moderately sorted, and moderately cemented with calcite. Porosity is 10-15%. Weathered surfaces are a light brown; fresh surfaces are greyish brown.

(450 m; Thorsteinsson, 1988)

Eureka Sound Formation: Cretaceous - Tertiary

Recessive, unlithified quartz arenite interbedded with recessive lignite beds with sharp contacts. Quartz arenite beds are 2-15 m thick; coal beds are 15 cm to 1.5 m thick. The arenite consists of fine- very coarse-grained, poorly sorted, sub-rounded to rounded grains (95% quartz, 5% coal fragments) and has approximately 30% porosity. Sparse lenses of quartz-pebble conglomerate 2-50 cm thick consist of arenite-supported quartz clasts 0.5 - 4 cm in diameter. Sandstone is locally stained and/or cemented by hematite. Lignite beds are characterized by recognizable woody plant remains and minor sulphur content. Locally, both arenite and coal beds have been replaced by hematite (720 m).

3.3 Structural Geology

3.3.1 Introduction

Previous published interpretations that focus on the structural geology for Cornwallis Island are limited to Thorsteinsson's (1988) 1:250,000 scale map of
Cornwallis Island that provided a basis for the previous structural interpretation of the study area. The Stanley Head structure was shown to be a northwest trending block of Thumb Mountain Formation surrounded by down faulted blocks of Cape Phillips Formation to the west and an area of continuous stratigraphy to the east. The uplifted block of Thumb Mountain formation was segmented by numerous normal faults. Tilted strata within the structure could be best explained as drag folds kinematically linked to the normal faulting.

Mapping undertaken for this thesis has significantly changed the structural interpretation of the Stanley Head region of Cornwallis Island. The geology of the Stanley Head map area is dominated by previously unrecognized shallow-level fold and thrust faults that create the distinctive topography of the area. Three phases of deformation have been documented in the study area: 1) northeast and southwest verging folds and thrust faults associated with an east-west compressional D1 event, 2) strike slip faulting and folding associated with a north-south compressional D2 event. The D2 folding results in localized refolding of some northeast and southwest verging folds, and 3) normal faulting associated with east-west directed extensional D3 event.

The outstanding structural feature of the Stanley Head region is the Stanley Head anticline. The anticline has a strike length of 11 km on Cornwallis Island, and is interpreted to extend across Pullen Straight to re-emerge on Little Cornwallis Island, at the Polaris mine site. The anticline has a wavelength of 11 km and is typically offset by numerous fold-axis parallel normal faults and fold-axis orthogonal strike-slip faults that divide it into several fault-bounded structural domains, labeled as Hills A-F and the Southern block (see map). Reference to the geologic map "Geology of the Stanley Head
region, western Cornwallis Island, Nunavut" is essential for a full understanding of the following overview of the exposed structure.

3.3.2 Physical conditions of deformation

The predominance of kink style folding, the presence of extensional and shear fractures, small scale faults with associated fault breccias, and the absence of a penetrative cleavage are consistent with deformation having occurred at shallow depths of less than 6 km by brittle processes. This is supported by the absence of metamorphic minerals identified in the area. Vitrinite reflectance data from Little Cornwallis Island suggests peak temperatures of deformation were 170 °C and 111 °C at the base of the Baumann Fiord and Cape Phillips Formations respectively. These temperatures are consistent with depths of less than 5 km depth for the Baumann Fiord Formation when applying a geothermal gradient of 30 °C/km.

3.3.3 Stanley Head anticline and associated D1 structures

The Stanley Head anticline and associated D1 structures are in part characterized by a change in vergence along strike. Hills E and F in the northernmost regions of the map area are characterized by southwest verging folds and thrust faults; whereas the remainder of the anticline including Hills A-D and the Southern Block, are characterized by northeast verging folds and thrust faults. The Stanley Head anticline is offset by numerous strike-slip tear faults defining individual thrust panels. These distinct, fault bounded structural domains each exhibit a unique structural geometry as manifested by distinct bedding orientations and vergence of structures (Fig. 3.4, see map). The
Figure 3.4 (a-d) Stanley Head anticline bedding data. Dashed lines indicate calculated pi girdle.

- Pole to bedding  • Calculated fold axis.

a) Southern Block poles to bedding and calculated average bedding orientation, 114/38.

b) Hill A poles to bedding and calculated average bedding orientations, 353/45, 180/24. Plunge and trend of fold axis 03/355.

c) Hill B poles to bedding and calculated average bedding orientations, 319/70, 174/34. Plunge and trend of fold axis 17/325.

d) Hill C poles to bedding and calculated average bedding orientations, 326/76, 189/28, 164/52. Plunge and trend of fold axis 17/331.
Figure 3.4 (e-g) Stanley Head anticline bedding data. Dashed lines indicate calculated pi girdle. 
* Pole to bedding.  • Calculated fold axis.
e) Hill D poles to bedding and calculated average bedding orientations: 295/77, 143/16, 19/44. Plunge and trend of fold axis 17/309. f) Hill E poles to bedding and calculated average bedding orientations: 003/43, 320/64. g) Hill F poles to bedding and calculated average bedding orientation, 300/40.
following discussion will focus on northeast and southwest verging structural domains separately.

**North-east verging structural domains**

In the northeast verging structural domains of Hills A-D the Stanley Head anticline is an open, cylindrical, subhorizontal - gently plunging, moderately inclined, northeast-verging fold with a sharp to angular hinge zone. In these domains the forelimb of the anticline strikes northwest and dips steeply to the northeast; the backlimb of the anticline strikes southeast and dips moderately to the southwest (Fig. 3.4). Overall the fold axis of the Stanley Head anticline plunges shallowly to the northwest. The anticline is cored by the weak lower Bay Fiord Formation (gypsum/anhydrite) that changes thickness along strike of the anticline. In outcrop the gypsum is foliated sub-vertically parallel to the trend of the anticline.

The Southern Block is characterized in part by an easterly verging thrust panel that places rocks of the upper Thumb Mountain Formation over the Cape Phillips Formation. Due to the present level of the peneplain only the backlimb of a fold is observed (Fig. 3.4). The present level of erosion and the observed structural geometry of the Southern Block make it impossible to characterize what the possible geometry of the thrust panel may have been immediately after the D1 event.

To the east of Hills A - D several smaller scale 2\text{nd} and 3\text{rd} order folds associated with the Stanley Head anticline are present within the Cape Phillips Formation. A second order syncline anticline pair are present to the east of Hill B, where bedding plane measurements were taken from small stream channels. It appears that the trace of the axial surfaces of these folds trends to the northwest, consistent with the overall trend of
the Stanley Head anticline. The folds appear to have a wavelength of 2-3 km. The scarcity of data available from this area precludes a detailed analysis of their fold geometries.

Where 3rd order folds are exposed, the folds exhibit kink to subangular style fold geometries and typically have wavelengths of 10 - 100 meters and 2-15 meter amplitudes. They are cylindrical, subhorizontal, gently inclined, open folds, that verge to the northeast, plunge gently to the north-northwest, and overall display similar orientations to the Stanley Head anticline (Fig. 3.5). The fold depicted in figure 3.5a provides an excellent opportunity to characterize fold and thrust interaction within the Cape Phillips Formation. Figure 3.5c illustrates a viable interpretation for the fold shown in figure 3.5a. This cross section illustrates that smaller scale intermediate detachments are likely present within the Cape Phillips Formation and that this fold can be interpreted as a fault propagation fold.

One syncline anticline pair is observed within the Upper Bay Fiord Formation on Hill B just to the east of the hinge line of the Stanley Head anticline. These folds are gently inclined, open, and have wavelengths and amplitudes of 10 - 15 m and 2 -5 m respectively (Fig 3.6a). These folds are unique in the study area, as they are the only higher order folds observed within the hinge zone of the Stanley Head anticline. These folds exhibit an m symmetry with respect to the larger hinge zone and plunge moderately to the northwest (Fig 3.6b).

Small-scale east verging thrust faults are observed within the Cape Phillips Formation to the west of Hill D. These thrust faults have on the order of 2 -10 meters displacement and display characteristic flat - ramp geometries and associated folding in
Figure 3.5
a) Northeast verging 3rd order D1 fold to the east of Hill D. Scale of photograph is 50 meters across. View looking north. b) Stereographic projection of the fold in figure 2.6a. Dashed blue great circle is the calculated pi girdle. Red star is the calculated fold axis exhibiting a plunge and trend of: 04/343. c) Interpretive cross section of fold in figure 2.6a. Cross section is viable and exhibits a fault propagation fold mechanism.
Figure 3.6
a) Small scale third order fold located in the hinge zone of the Stanley Head anticline, Hill B. Note the m symmetry to the folding with respect to the main hinge zone. View is to the northwest, beds are dipping away from the viewer. b) Stereographic projection of the folds shown in figure 2.7a. Dashed blue great circle is the calculated pi girdle and the red star represents the fold axis. Plunge and trend of fold axis: 33/305.
the hanging wall (Fig 3.7a - 3.7c). Ramp orientations of these small-scale thrusts are somewhat variable, but typically strike to the south and dip between 15-25° (Fig 3.7d).

One small-scale thrust fault with associated slickenlines is observed within the Lower Thumb Mountain Formation (Fig 3.8a), on the backlimb of the Stanley Head anticline (Hill C). The orientation of the bedding, fault plane, and slickenlines are 178/35, 150/71, and 33/165 respectively (Fig 3.8b). When the bedding is rotated to horizontal, the fault plane and slickenline orientations are 136/42 and 34/189 respectively (Fig 3.8c). These orientations are consistent with the orientations of small-scale thrust faults observed to the west of Hill D within the Cape Phillips Formation. This minor thrust fault is interpreted to have originated prior to the main D1 folding event.

To the west of the exposed gypsum on Hill A, a well developed fault surface and associated fault breccia (Fig 3.9a, 3.9b) occurs within the Lower Bay Fiord Formation. The fault plane strikes southeast, and exhibits slickensides that plunge moderately southwest (Fig 3.9c). Calcite fibers on the fault surface indicate top to the northeast sense of shear, consistent with the overall sense of vergence of Hill A. This thrust fault is interpreted to have originated after the main D1 folding event. This is because the thrust fault dips approximately 35°, an angle that is consistent with a typical ramp orientation within a competent unit.

South-west verging structural domains

Hills E and F represent the southwest verging structural domains of the Stanley Head anticline. They are characterized by southwest verging thrust faults that place rocks of the Bay Fiord Formation on top of rocks of the Cape Phillips Formation. Due to the present level of the peneplain only the backlimb of a fold is observed on Hills E and F.
Figure 3.7
a) Small scale northeast verging thrust ramp within the Cape Phillips Formation to the west of Hill D, displacement on the thrust fault is on the order of 2-3 m. View is to the north. b) Small scale fold associated with slip on a near horizontal detachment within Cape Phillips Formation to the west of Hill D. The fold verges to the northeast. View is to the south c) Stereonet projection of the fold shown in figure 3.7b. Dashed blue great circle is the calculated pi girdle, the red star is the calculated fold axis. Plunge and trend of fold axis: 02/337. d) Stereonet projection of small scale thrust faults within the Cape Phillips Formation to the west of Hill D. Average thrust fault orientation (190/18) is depicted by the great circle.
Figure 3.8

a) Small scale thrust within the Lower Thumb Mountain Formation, located on Hill C. Displacement on the fault is approximately 50 cm. Yellow lines delineate equivalent beds across the fault plane, red lines delineate the fault plane, and green lines outline a small scale footwall anticline.

b) Stereonet projection of thrust plane (164/70), slickenlines (33/165), and bedding (178/35). Note: the solid great circle represents the fault plane, ◆ slickenlines, and • poles to bedding.

c) Stereonet projection of fault plane ( and slickenlines when bedding is rotated to horizontal. Fault plane 136/42, slickenlines 34/189. Note for figures 3.8
Figure 3.9
a) Fault breccia immediately adjacent to small scale northeast verging thrust fault within the Upper Bay Fiord Formation, Hill A. b) Slickenlines associated with the small scale northeast verging thrust fault. c) Stereonet projection of fault plane and associated slickenlines. Note these thrust faults are interpreted to have originated after the D1 folding event. Note great circles represent fault planes, and • represents the slickenlines.
The observed structural geometry and present level of erosion make it difficult to characterize what the original geometry of the thrust panel would have been immediately after the D1 event.

Outcrop exposures along the creek immediately to the west of Hill F provide an excellent opportunity to view footwall structures found within the Cape Phillips Formation. These structures as a whole constitute a contorted fold train that exhibits southwest verging tight to isoclinal overturned anticline and syncline pairs (Fig 3.10a, 3.10b). These sharp to angular folds exhibit amplitudes between 5-30 meters, wavelengths of 5-50 m, and are characterized by gently plunging northwest/southeast trending fold axes (Fig 3.10c, 3.10d). The presence of smaller scale parasitic s, m and z folds aided in determining the overall symmetry of the larger scale folds, where outcrop exposure was limited due to talus. The top of this fold train is truncated by a shallowly dipping southwest trending thrust fault, that places strata of a near horizontal orientation on top of the fold train (Fig 3.11). Locally, within the fold train there are overturned folds that exhibit east-west trending fold axes and appear to verge to the north (Fig 3.10c, 3.10d). These folds are of uncertain origin but may be kinematically linked to the D2 north-south compressional event and will be discussed in section 3.3.4.

The truncation of the footwall structures by the shallowly dipping southwest verging thrust fault indicates the presence of an intermediate detachment horizon within the Cape Phillips Formation on Hill F. This geometry may indicate that the main thrust fault that juxtaposed the Bay Fiord Formation over the Cape Phillips Formation, flattened out within the shales of the Cape Phillips Formation, establishing an overall fault-bend fold mechanism for the thrust panel that defines Hill F.
Figure 3.10

a) Overturned fold within Cape Phillips Formation to the west of Hill F. View looking northwest. b) Hinge zone of an overturned syncline within the Cape Phillips Formation. The illustrated geometry is typical of the geometry of the footwall structures observed within the fold train to the west of Hill F. View is to the north, field of view is approximately 3 m across. c) Stereonet projection of poles to bedding within footwall structures to the west of Hill F. Note that yellow circles indicate deviation from bedding norms and the blue great circle is the calculated π-girdle, and • poles to bedding.
d) Stereonet projection of calculated and measured fold axes within the footwall structures to the west of Hill F. Note that yellow circles indicate deviation from the fold axes norm and • indicate fold axes.
Figure 3.11
Thrust fault truncating fold train to the west of Hill F within the Cape Phillips Formation. The red line delineates the trace of the thrust fault on the cliff, the yellow lines delineate the trace of bedding on the cliff in both the hangingwall and footwall. Approximate orientation of the bedding above the thrust fault is 320/05. Height of cliff is approximately 25 meters. View is to the north.
To the east of Hills E and F a second order fold is exposed within the Cape Phillips Formation as the northern continuation of the syncline to the east of Hills B - D (Fig. 3.12a). Third and fourth order parasitic folds observed on the exposed limb of the syncline were used to establish the overall symmetry of the larger second order fold. These folds when viewed down plunge exhibit a z asymmetry (Fig 3.12a, 3.12b) indicating the outcrop exposure is located within the eastern limb of the second order syncline. These parasitic folds exhibit a kink style geometry, and plunge gently to the northwest (Fig 3.12c).

Gypsum Body

Between Hill A and the Southern Block lies a large ellipsoid shaped body of the lower Bay Fiord Formation gypsum/anhydrite, which is poorly exposed and extremely weathered in outcrop. This body of gypsum is continuous from Hill B to the southern margin of Hill A as a thin narrow neck of gypsum/anhydrite that defines the hinge zone of the Stanley Head anticline. At the southern margin of Hill A, this body rapidly expands into a large ellipsoid shape that terminates along the northern margins of the Southern Block. The shape of this body on the map is delineated based on a few outcrop exposures and a topographically low area observed between Hill A and the Southern Block. Based on the map pattern, it is interpreted that this body of gypsum is kinematically linked to the formation of the Stanley Head anticline. This will be discussed in detail in chapter 4.

Interpretation of the D1 event

Collectively, the northeast and southwest verging structures discussed above are interpreted to be kinematically linked to, and represent a part of, the Boothia aged
Figure 3.12
a) Eastern limb of second order syncline to the east of Hill E, view is to the north looking down plunge. Note the third order z fold in the western margins of the photo. Yellow lines represent overall bedding orientation. b) Small scale z fold observed in the eastern limb of the second order syncline. View is to the north, field of view is approximately 3 meters. c) Stereonet projection of z folds observed within the eastern limb of the second order syncline. Dashed blue great circle is the calculated pi girdle, red star is the calculated fold axis. Plunge and trend of fold axis: 02/317.
Cornwallis fold belt. The D1 structures are interpreted to be of Boothia age for the following reasons: 1) These structures are parallel in trend to the overall trend of the trend of the Cornwallis fold belt. 2) Given the tectonic framework of the central Arctic Islands, if the D1 structures are of Boothian age, then they should be subsequently affected by a north-south compressional D2 event, and an extensional D3 event, which is the case in the study area. Locally these structures represent a northwest trending, northeast and southwest verging, fold and thrust system. The fold style and geometry of the Stanley Head anticline and the associated smaller scale folds are typical of structures expected in a shallow-level evaporite based fold and thrust system.

3.3.4 D2 Strike-slip faulting and superposed folding

Strike-slip faulting

An isolated and prominent fault bounded block defines the southern most expression of the Stanley Head anticline. The major structural feature within the Southern Block is a north-northwest striking strike-slip fault zone approximately 20 m in width (Fig 3.13a), that has an arcuate trace across the structural domain. Where exposed, the fault plane dips steeply to the northeast with slickenlines plunging shallowly to the southeast (Fig 3.13b). This strike slip fault offsets the D1 thrust panel and juxtaposes Upper Thumb Mountain Formation against the Lower Thumb Mountain Formation. The full length of the fault zone is unknown because it has been cross cut by a D3 normal fault at is northern limit. However, it is interpreted to continue northwards through the Cape Phillips Formation. The southern limit of the fault zone appears to terminate within
Figure 3.13

a) Yellow lines delineate the location of the topographic depression representing the strike slip fault zone across the southern block that juxtaposes thinly bedded Lower Thumb Mountain Formation in the foreground with the massively bedded Upper Thumb Mountain Formation in the background. View is to the northeast. b) Stereonet projection of the strike slip fault and associated slickenlines.
the southern margins of the Southern Block, but this is speculative due to poor exposure in the southern margins of the block.

Contractional and extensional deformation associated with strike slip faulting can occur in two settings, either at terminating zones of a strike slip fault or at bends within a strike slip fault zone (Fig 3.14a, 3.14b). Deformation along terminating zones of strike slip faults occurs as a complex zone of imbricate faulting that may be either compressional or extensional (Fig. 3.14a 3.14b). Displacement along strike slip faults with bends typically produce a complex zone of deformation that may be either compressional or extensional depending on the nature of the bend with respect to the sense of displacement across the fault. These bends are known as either restraining bends or releasing bends and typically produce strike slip duplexes in which a series of horizontally stacked horses are bounded on both sides by segments of the main strike slip fault. (Fig. 3.14a, 3.14b) (Twiss and Moore, 1992).

In the case of the strike slip fault that offsets the Southern Block, the deformation associated with the strike slip faulting is interpreted as extensional. This is due to the presence of normal faults within the Upper Thumb Mountain Formation that occur at oblique angles to the main strike slip fault and have a component of strike slip as well as dip slip across the fault plane (Fig. 3.15). These normal faults cross cut the D1 thrust fault and terminate against the strike slip fault, indicating that they are associated with the D2 event. Two possibilities exist to explain the origin of these normal faults with respect to the strike slip fault, they can either represent an imbricate fault array associated with terminal regions of the strike slip fault or they could be part of a extensional strike slip duplex system.
Figure 3.14
a) Extensional structures associated with strike slip faulting. i) Releasing bend on a dextral strike slip fault. ii) Extensional duplex on a dextral strike slip fault. iii) Block diagram of a normal flower structure. iv) Extensional deformation in the termination zones a dextral strike slip fault by the formation of imbricate fans.
b) Contractional structures associated with strike slip faulting. i) Restraining bend on a dextral strike slip fault. ii) Contractional duplex on a dextral strike slip fault. iii) Block diagram of a reverse flower structure. iv) Contractional deformation in the termination zones of a strike slip fault by the formation of imbricate fans. (Figures from Twiss and Moore, 1992)
It is proposed that the D2 normal faults represent an extensional imbricate fault array associated with the termination of the strike slip fault for the following reasons:

1) There is no evidence of these normal faults being bounded on their eastern margin by a strike slip fault, negating the possibility of an extensional duplex structure. 

2) The southernmost identifiable trace of the strike slip fault occurs at the limits of the D2 normal faulting, indicating that the strike slip fault likely terminated within the confines of the Southern Block. The strike slip fault is cross cut in the northern margin of the Southern Block by a north trending D3 normal fault indicating its trace likely extended northward prior to the D3 event.

3) The geometry of the normal faults is similar to that predicted for an imbricate fault splay associated with the terminating zone of a strike slip fault zone.

Associated with the strike slip fault zone within the Southern Block is the emplacement of barite as veins of variable continuity and thickness. Typically barite is present as 0.5 - 3 cm thick veins in close proximity to the main strike slip fault zone, and is locally cross-cut by calcite veins. One poorly exposed barite vein is approximately 20-30 cm in thickness, dips near vertically and has a trend parallel to the strike slip fault. The only barite found within the study area is in close proximity to the strike slip fault zone within the Southern Block.

Superposed folding

Two areas of superposed folding occur in the northwest quadrant of the study area. The most well defined region occurs in a fault-bounded block within the margins of Hill D where the Upper Thumb Mountain Formation is juxtaposed to the west of the Upper Bay Fiord Formation (see map). The other region occurs immediately to the west
Figure 3.15
Slickenline from a D2 normal fault plane within the Southern block. Note both the dip slip and strike slip component of movement across the fault plane.

Figure 3.16
a) Superposed fold axis data from the Upper Thumb Mountain Formation, Hill D. b) Superposed fold axis data from the Cape Phillips Formation, to the west of Hill F. Note that these fold axes from both superposed fold domains deviate from the regional shallowly plunging northwest trending fold axis.
of Hill F within the footwall structures of the Cape Phillips Formation (see map). In both regions the superposed folding is characterized by fold axes orientations that deviate from the regional shallow plunging north-northwest trending fold axis (3.16a, 3.16b). In addition bedding orientations that define the second order syncline and anticline pair within the Cape Phillips Formation are not consistent with a gently northwest plunging fold axis. It is believed that this area has undergone superposed folding, but a lack of outcrop does not allow for a detailed analysis of this area.

Superposed folding located on Hill D is characterized by shallowly dipping strata of the Upper Thumb Mountain Formation that creates a dome and basin, Type -1 interference pattern. These domes are 15 - 75 m in diameter and are typically separated by tighter intervening basins (Fig 3.17). Figure 3.18 illustrates the most well defined domes in the area where the strike of the bedding rotates around the perimeter of the domes. Fold axis data from this region was calculated from individual domes or basins within the area. These fold axes deviate from the regional north-northwest gently plunging fold axis by typically plunging gently to the south and the southeast.

The area of superposed folding within the Cape Phillips Formation to the west of Hill F is characterized by one well defined region of overturned, tight to isoclinal, northerly verging folds (Fig 3.19). These folds are exposed along an approximately 50 m wide area and exhibit shallowly plunging east-west trending fold axes that deviate from the shallowly plunging northwest trending fold axis that characterizes the majority of the footwall structures within the southwest verging fold train.
Figure 3.17
Well defined dome and basin structures within the superposed folded area, Hill D. These three domes illustrate typical structures found within area. Note the dip of bedding in the two domes in the foreground of the photo. The dome in the foreground illustrates beds dipping away from the hill. The dome in the middle of the photo illustrates beds that dip into the hill as well as being gently folded as depicted by the yellow line. View is to the south southwest.
Figure 3.18
Bedding orientations of superposed folded area within a fault bounded block of Upper Thumb Mountain Formation, Hill D. This region is characterized by a Type 1 interference pattern, where domes are defined as yellow lines that are separated by intervening basins.
Figure 3.19
a) Map view of footwall structures west of Hill F. These structures are characterized by overturned syncline anticline pairs, that are typically characterized by gently plunging northwest trending fold axes. The E-W trending fold axes represent an area of superposed folding. b) View of overturned northerly verging fold, in part, defining the area of superposed folding within the Cape Phillips Formation to the east of Hill F. The yellow line delineates the trace of bedding, the red arrow the fold axis lineation observed within the hinge of the fold.
Plunge and trend of lineation: 06/265
Interpretation of D2 event

Collectively, the strike slip fault zone in the Southern Block and the areas of superposed folding within the Stanley Head anticline are interpreted as a D2 event that is associated with the north-south compressional event representing the Ellesmerian Orogeny. These structures modify the pre-existing D1 fold and thrust deformation and are cross cut by the subsequent D3 deformational event. Given the tectonic framework of the central Arctic Islands, and the fact that the D2 structures are consistent with a north-south compressional stress regime, it is reasonable to interpret these structures as Ellesmerian in origin.

The strike slip fault zone found in the southern block is consistent with this interpretation for the following reasons: 1) The north-south trend of the strike slip fault is consistent with a north-south directed compressional stress regime. 2) The relative age of the strike slip fault is bracketed between the D1 and D3 events due to observed cross cutting relationships. The strike slip fault cross cuts the D1 thrust panel that defines the Southern Block and the normal faults that are kinematically linked to the strike slip fault cross cut the D1 thrust fault that juxtaposes Upper Thumb Mountain Formation with the Cape Phillips Formation (see map). In addition, the strike slip fault zone is cross cut by a north-northeast trending D3 normal fault in the northern margins of the Southern block (see map).

Under the assumption that regions of superposed folding represent the D2 event, stereographic analysis of fold axes from these regions illustrate that the regional north-northeast plunging fold axis has been refolded to plunge either along an east-west trend
or a south-southeast trend. These areas of superposed folding are interpreted to represent localized areas that experienced compressional stress during the D2 event.

The areas of superposed folding are isolated within the Stanley Head anticline to two regions. This poses the question: Are these regions of superposed folding part of a larger regional north-south compressional D2 event or are they localized phenomena? Examining Thorsteinsson's (1988) 1:250,000 scale map of Cornwallis Island one of the most striking features is the dome and basin map pattern that dominates the center portions of the map (Fig 3.20). The dome is defined by the Centre anticline, which is cored by the Lower Ordovician Baumann Fiord Formation. The basin is defined by the Lady Hamilton syncline, which is cored by the Upper Devonian Hecla Bay Formation. The trace of both the anticline and the syncline is curved suggesting that originally north-northeast trending Cornwallis Fold Belt structures have undergone refolding, furthermore the strike of beds rotates around the dome and basin respectively. This regional map pattern is interpreted to represent a Type-1 interference pattern. In addition, new data from the Stuart Bay and Caribou River areas of northeastern Cornwallis Island (Jober p. comm, 2003) demonstrate that the D1 event has been overprinted by east-west trending folds associated with a north-south compressional event. With evidence for superposed folding found across Cornwallis Island, it is reasonable to interpret that superposed folded areas from the study area are part of a regional D2 event in which originally northwest trending D1 folds have been refolded due to north-south directed compressional folding.
Figure 3.20
Regional Type 1 interference pattern observed on Cornwallis Island. The southern regions of the map are dominated by a dome structure defined by the Centre anticline. The northern regions of the map are dominated by a basin structure defined by the Lady Hamilton Syncline. Note that the traces of both the syncline and anticline are curved. These structures are interpreted to represent a refolding of the Boothia aged Cornwallis Fold Belt.
3.3.5 D3 Normal faulting

There are numerous steeply dipping north-south trending normal faults, that in part define several fault blocks that offset the Stanley Head Anticline, and define the Intrepid Bay graben (see map). Normal faults bound the majority of the uplifted, eastern and western, margins of Stanley Head Anticline. These faults typically juxtapose the Upper Thumb Mountain Formation against the Cape Phillips Formation, and have varying amounts of displacement ranging from a few tens of meters up to nearly 400 meters (see cross section sheets A & B). The faults that bound the Intrepid Bay graben have a vertical displacement of at least 720 meters as the thickness of the Eureka Sound Formation within the Intrepid Bay graben is a minimum of 720 meters as calculated from the standard structure contour three point problem method. The geometry of the D3 normal faults at depth is unknown.

It is proposed that the east and west bounding normal faults of the Stanley Head anticline are part of the same normal fault system that form the graben for the following reasons: 1) They are directly along strike and are almost continuous from the graben to the anticline. 2) The only D3 fault plane measurements taken on the Stanley Head anticline are found on Hill C and have orientations of 130/80 and 135/75, with an associated slickenline orientation of 75/250, indicating dip slip faulting (Fig. 3.21). The displacement on these faults is interpreted as normal, based on the stratigraphic relationship that typically juxtaposes Upper Thumb Mountain Formation with the Cape Phillips Formation. Note that D3 fault planes were rarely observed due to fault zones being preferentially weathered and eroded.
Figure 3.21
Stereonet projection of normal faults and associated slickenlines from Hill C, Stanley Head anticline.

Figure 3.22
Highly fractured Upper Thumb Mountain Formation, Hill E, Stanley Head anticline. This outcrop illustrates stockwork fractures common to the study area. Field of view is 7 m.
Interpretation of the D3 event

The D3 event is interpreted to be extensional in nature and associated with the Eurekan deformation in which the structures are characterized by steeply dipping northwest-southeast trending normal faults. This is due to the following reasons: 1) These faults cross cut both D1 and D2 structures. 2) Assuming that the north-south trending normal faults of the Stanley Head anticline and the Intrepid Bay graben are part of the same fault system, the Tertiary aged Eurekan Sound Formation found within the Intrepid Bay graben constrains the age of the faulting to coincide with the Eurekan deformational event.

3.3.6 Jointing and fracturing within the limestone formations of the Stanley Head anticline

Limestone formations throughout the study area are characterized by brittle deformation in which the rocks are highly fractured and veined (Fig. 3.22). Irregular stockwork fractures are pervasive throughout the area, but appear to be concentrated on Hills E, F and the Southern Block, as well as in the hinge zones of Hills A-D. The frequency of these irregular fractures also tends to increase with increasing proximity to normal faults. Well developed planar joints are observed throughout the study area and exhibit a spacing ranging from a few meters to several tens of meters. These joints typically have an aperture of a few millimeters and rarely up to a maximum of a few centimeters. Joint apertures are ubiquitously filled with white calcite.

The study of jointing within the Stanley Head anticline was undertaken to ascertain what, if any, control these joints exerted on the creation of a plumbing system.
and the migration of fluids throughout the area of study. Stereometric analysis was undertaken to elucidate the timing and origin of jointing with respect to D1 folding event. The relative density of stockwork fractures in relation to folding will be discussed in chapter 4.

Joint orientations from each structural domain are plotted on equal area stereonet projections as poles to the fracture plane with great circles representing the average bedding orientations of the fold (Fig. 3.23). Each structural domain is subdivided according to dip domains observed within the Stanley Head anticline, such that each stereonet contains joint orientations found within each separate dip domain. Joint orientations were discarded when they were in close proximity and parallel in trend to either a D1 tear fault or a D3 normal fault, as these joints are interpreted to be kinematically linked to the faulting.

From figures 3.23 (a, c, e, g, i, k, m, o, q, s, u, and w) one feature that is apparent is that the poles to jointing are generally spread around the average bedding plane great circle. This illustrates that the joints are typically (sub)perpendicular to the inclined bedding. Two explanations exist to explain the geometric pattern: 1) the joints predate the fold and were rotated during folding or; 2) the joint patterns were formed at a late stage during fold formation by stress systems symmetrically inclined to the bedding surfaces. The second possibility seems improbable because the stress system would have to systematically change across each structural domain of the Stanley Head anticline to accommodate the different limb orientations across each structural domain of the Stanley Head anticline.
Working under the assumption that the jointing predated folding, the joint data sets have been reoriented by rotation of the bedding planes around the fold axis such that bedding was made horizontal. The reoriented sets are shown as the stereonets on the right hand side of figure 3.23. The rotated poles to the jointing are grouped around the periphery of the net illustrating that the original orientations of the joints were near vertical prior to the D1 event.

It is here proposed that the origin of the jointing observed within the Stanley Head anticline is related to burial of the sedimentary column. There are no documented deformational events prior to the Boothia Uplift recorded on Cornwallis Island, which indicates that the state of stress in the rocks was probably governed by the weight of the overlying sedimentary column according to the relationship:

$$\sigma_z = \rho g z$$

$\sigma_z$ is the vertical stress, $\rho$ is the density of the overburden, $g$ is gravity, and $z$ is the depth of the buried rock. The maximum depth of tensile fracturing as a function of the tensile strength of the rock and fluid pressure ratio is presented in figure 3.24. (For a complete derivation of the equations which produce figure 3.24 refer to Suppe, 1985 p. 190-193).

In tectonically stable sedimentary basins, at depths up to 3 km, measured fluid pressures are typically not greater than hydrostatic (Twiss and Moore, 1992), this suggests that the flow of fluids from the Upper Bay Fiord Formation to the surface was unrestricted. The pore fluid pressure ratio, $\lambda$, is given by the relation:

$$\lambda = \frac{P_f}{\rho g z}$$

(Where $P_f$ is the pore fluid pressure)

In the case of the Stanley Head anticline, the stratigraphically lowest limestone formation exposed is the Upper Bay Fiord Formation, that was buried to a depth of 1.6 km beneath
Figure 3.23
a) Southern Block joint and bedding data (bedding 114/38). b) Southern Block, joint data rotated such that bedding is horizontal. c) Hill A eastern limb of Stanley Head anticline, joint and bedding data (bedding 353/45). d) Hill A eastern limb of Stanley Head anticline, joint data rotated such that bedding is horizontal. e) Hill A western limb of the Stanley Head anticline, joint and bedding data (bedding 180/24). f) Hill A western limb of the Stanley Head anticline, joint data rotated such that bedding is horizontal.
Note: great circles indicate average bedding orientation, • indicates poles to jointing.
Figure 3.23

- **g)** Hill B eastern limb Stanley Head anticline, joint and bedding data (bedding 319/70).
- **h)** Hill B eastern limb of Stanley Head anticline, joint data rotated such that bedding is horizontal.
- **i)** Hill C eastern limb of the Stanley Head anticline, joint and bedding data (bedding 326/76).
- **j)** Hill C eastern limb of the Stanley Head anticline, joint data rotated such that bedding is horizontal.
- **k)** Hill C western limb of Stanley Head anticline, joint and bedding data (bedding 164/52).
- **l)** Hill C western limb of Stanley Head anticline, joint data rotated such that bedding is horizontal.

Note: great circles indicate average bedding orientation, • indicates poles to jointing
Figure 3.23
m) Hill C western limb of the Stanley Head anticline, joint and bedding data (bedding 205/22).
 n) Hill C western limb of the Stanley Head anticline, joint data rotated such that bedding is horizontal.
o) Hill D eastern limb of the Stanley Head anticline, joint and bedding data (bedding 300/75).
p) Hill D eastern limb of the Stanley Head anticline, joint data rotated such that bedding is horizontal.
q) Hill D western limb of the Stanley Head anticline joint and bedding data (bedding 159/44).
r) Hill D western limb of the Stanley Head anticline, joint data rotated such that bedding is horizontal.
Note: great circles indicate average bedding orientation, • indicates poles to jointing.
Figure 3.23
s) Hill E joint and bedding data (bedding 003/43). t) Hill E rotated joint data such that bedding is horizontal. u) Hill E joint and bedding data (bedding 320/64). v) Hill E rotated joint data such that bedding is horizontal. w) Hill F joint and bedding data (bedding 300/40). x) Hill F rotated joint data such that bedding is horizontal.
Note: great circles indicate average bedding orientation, indicates poles to jointing.
Figure 3.24
Maximum depth of tensile fracturing as a function of tensile strength and fluid pressure ratio $I = P_f / r g z$. Where $s_i$ is vertical, and $r = 2700 \text{ kg/m}$. The dashed lines indicate the parameters used in the Stanley Head anticline, fluid pressure is assumed to be hydrostatic $I = 0.4$. Diagram after Suppe, 1985.
the top of the Cape Phillips Formation prior to the D1 event (see map & cross sections). The estimated density of the sedimentary column is 2700 kg/m$^3$. The tensile strength of the mudstones and wackstones of the Bay Fiord and Thumb Mountain formations is estimated to be comparable to the fine grained Carrara marble which has a tensile strength of -8 Mpa (Jaeger and Cook, 1979). If this assumption is valid, from figure 2.24 it is then possible to produce jointing in the limestones of the Bay Fiord and Thumb Mountain formations due to burial.
CHAPTER 4
DISCUSSION AND INTERPRETATION

In an attempt to characterize the geometry and kinematics of deformation within the Stanley Head region of western Cornwallis Island balanced cross sections across the study area have been produced. These cross sections are meant to portray a geologically reasonable and possible solution to the subsurface geometry and have been constructed using the integration of surficial data and a kinematic model of deformation for the multiple detachment evaporite based Parry Islands Fold Belt, on eastern Melville Island. A brief review of the mechanics of fold and thrust belts and of the kinematics of deformation within the Parry Islands Fold Belt is presented to demonstrate the reasoning behind the construction of the cross sections provided in this thesis.

4.1 Mechanics of evaporite based fold and thrust belts

The style of deformation in thin skinned fold and thrust belts is extremely dependent upon the resistance to sliding along the detachment surface (Davis & Engelder, 1985). Evaporites, because of their low coefficient of friction, provide an exceptionally weak horizon in which a basal detachment can form, with the resistance to sliding determined by the yield strength of the evaporite rather than by the considerably higher frictional strength of a stronger basal detachment layer. Mechanical models for fold and thrust belts predict an exceptionally narrow cross sectional taper for a fold and thrust belt if the basal detachment is within an evaporitic horizon (Davis et al. 1983; Dahlen et al. 1984).
The narrow taper observed in evaporite based fold and thrust belts predicts that the orientation of the principal stress axes will be within a few degrees of horizontal and result in a lack of a consistently dominant vergence direction of folds and faults (Davis and Engelder, 1987). The following discussion will compare and contrast the geometries of weak basal detachment fold and thrust belts with strong basal detachment fold and thrust belts.

4.1.1 Rock strength

For typical geologic strain rates of $10^{-14} (±2) s^{-1}$ and temperatures less than 250°C, almost all common crustal rocks deform in a brittle manner that can be modeled by the Coulomb failure criterion (Davis and Engelder, 1987). Byerlee (1978) demonstrated that for most rock types the coefficient of friction is pressure dependent, but essentially independent of rock type. Important exceptions are certain clay minerals, which are relatively weak, and evaporites, which are vastly weaker than any other common rock type. Strengths of common rock types are plotted versus temperature and depth in figure 4.1.

For basal detachment depths of 2-8 km (observed in many fold and thrust belts) it is expected that deformation will be dominated by brittle processes. However, in the case of halite and anhydrite, under these conditions, experimental deformation demonstrates that they deform in the ductile regime (Fig 4.1) (Carter & Hansen, 1983; Muller et al. 1981). Evaporites are weaker than any other common rock type, including shales. For example, at temperatures over 100°C and at typical strain rates of $10^{-14} (±2) s^{-1}$ anhydrite
Figure 4.1. Strengths of evaporites and other common rock types as a function of depth, assuming average continental geothermal gradients and average crustal strain rates. Based on these assumptions at a depth of 3-5 km anhydrite deforms in the ductile field. (From Harrison, 1995. Redrafted from Davis and Engelder, 1987, and Dahlen, 1990.)
has a shear strength of less than 10 MPa (Fig 4.1) (Muller et al. 1981), and halite has a shear strength of less than 1 MPa (Carter & Hansen, 1983).

Because of the expected ductile behaviour of evaporites under pressures and temperatures associated with typical detachment depths the Coulomb failure criterion is not applicable and the resistance to failure along a decollement surface is determined by the yield strength of the evaporite. The shear strength of a basal detachment situated within an anhydrite horizon can be expressed as a constant, $\tau_0$, whose value is likely less than 10 Mpa. Furthermore, the yield strength of evaporites may be lowered further in the presence of excess pore pressure (Spiers et al. 1986). This scenario is geologically reasonable to expect as evaporites can form an impermeable barrier thereby promoting the generation of excess pore pressure.

### 4.1.2 Critical taper of an evaporite based fold and thrust belt

The extreme weakness of evaporites results in a low critical taper angle as illustrated in equation #1. If the basal decollement occurs in an evaporitic horizon the frictional strength term $\mu$ is replaced by the constant $\tau_0$. The critical taper of a the wedge is given by the relation:

$$\alpha + \beta = \beta \rho gz + (1-\lambda) \tau_0 - [2S_0 \cot(\alpha + \beta)/\csc(\phi - 1)]$$

$$\rho gz[1 + 2(1-\lambda)/\csc(\phi - 1)]$$

(Equation #1)

Where, $\alpha$ is the average slope of topography, $\beta$ is the dip of the basal decollement, $\phi$ is internal friction angle of the wedge, and $\tau_0$ is the yield strength of the evaporite. The Hubbert-Rubey pore fluid pressure coefficient, $\lambda$, is given by the relation:

$$\lambda = P_p/\rho gz$$
Here $P_p$ is the pore fluid pressure, $\rho$ is the mean density of the sedimentary overburden, $g$ is the gravitational acceleration, and $z$ is the depth.

Given a basal decollement at a typical depth of 2-8 km, and typical geological strain rates halite will have a shear strength, $\tau_0$, of less than 1MPa. When this value of $\tau_0$ is put into the critical taper relation for a wedge riding over evaporites, a critical taper of only a few tenths of 1° is required for a thrust sheet to ride over a salt based decollement. This very narrow critical taper implies that it is possible to obtain a broader fold belt. The same is true for anhydrite, although to a lesser extent than halite (Davis and Engelder, 1987).

4.1.3 Stress orientations and vergence of structures within fold and thrust belts

A second important parameter controlled by the presence of a very weak basal detachment is the angle $\Psi_b$, which is the angle at which $\sigma_1$ plunges toward the foreland with respect the basal detachment. The magnitude of $\Psi_b$ is very strongly dependent upon the shear stress that can be supported across the basal decollement (Davis and Engelder, 1985, 1987). The orientation of $\sigma_1$ is important with respect to deformation in fold and thrust belts because at every point in a critically tapered wedge there will be two planes in the thrust panel oriented symmetrically about the $\sigma_1$ axis at an angle, $\theta$, which will satisfy the Coulomb failure criterion. These two candidate slip planes represent forward and backward verging thrust faults. $\theta$ is defined by the relation:

$$\theta = \pm (45^0 - \phi/2)$$

In the case of a fold and thrust belt with a strong basal detachment, forward verging slip planes, $\delta_f$, dip at an angle of $\delta_f = \theta - \Psi_b$, and backward vergent slip planes
will dip at an angle of $\delta_b = \theta + \Psi_b$. Where $\theta$ is the angle from the $\sigma_1$ axis to the thrust plane (Fig. 4.2). Forward verging slip planes dip more shallowly by a factor of $2\Psi_b$, than backward verging slip planes. The difference in dip partially explains why forward verging thrusts are more common than back thrusts in most thrust belts (Davis and Engelder, 1985). This is because the forward verging slip plane permits a greater amount of horizontal shortening for the same increase in gravitational potential energy (Davis and Engelder, 1985). Strength anisotropy within the stratigraphic column will also favour the formation of forward verging thrusts closer to the orientation of bedding. In addition, the section of the thrust sheet behind the frontal thrust zone has typically been thickened and thus strengthened by earlier thrusting, making back thrusts less likely (Davis and Engelder, 1987).

In the case of evaporitic basal detachments $\Psi_b$ will be very small, implying that the two candidate slip planes that will form in the thrust sheet will have essentially equal dips, and therefore an equal number of forward and backward verging thrust faults should form. Figure 4.3 illustrates the difference between the dip $\Psi_b$ for a strong and weak basal detachment and figure 4.4, illustrates the difference between the two candidate slip planes in a thrust sheet over-riding a strong basal detachment and a weak basal detachment. In the case of blind thrust faults, fault related folds should have a relatively symmetrical form (Davis and Engelder, 1985, 1987).

4.2 REVIEW OF THE PARRY ISLANDS FOLD BELT, MELVILLE ISLAND

A review of the salt based Parry Islands Fold Belt on Melville Island is necessary in order to put into context the structural geology of the Stanley Head region of western
Figure 4.2 An illustration of why, in the presence of significant decollement strength, forward vergent slip planes dip more shallowly than do the backward vergent planes. This dip difference is a direct result of the fact that $y_b$ is positive. For an evaporite based fold belt $y_b$ is very small ($< 1-3$ degrees), so thrusts and folds are relatively symmetric. (From Davis and Engelder, 1985.)

Figure 4.3. Mohr-Coulomb diagram showing the contrast in $y_b$, the plunge of the axis of maximum compressive stress with respect to the basal decollement, for A) a strong and B) a weak evaporitic decollement. (From Davis and Engelder, 1985.)
Figure 4.4. The orientations of candidate slip planes in (A) a strong basal wedge and (B) a wedge with a weak evaporitic detachment. (From Davis and Engelder, 1985.)
Cornwallis Island. The structural geometry and kinematics of deformation of the fold and thrust belt on Melville Island is well constrained through seismic imaging, well data, and geological mapping (Harrison & Bally, 1988). Melville Island provides a comparable stratigraphic and tectonic setting to that found on Cornwallis Island and is used as an analogue for the possible geometry and kinematics of deformation for the structures found in the Stanley Head region.

It is here proposed that the structures found at the Stanley Head region are comparable to the structures that characterize Melville Island's salt-based fold belt due to the following observations: 1) both regions exhibit a stratigraphic sequence that have similar mechanical properties, and 2) both regions exhibit a similar lack of a preferred vergence direction in folds and thrust faults. Melville Island's salt based fold belt is in part characterized by two weak detachment layers within the evaporitic unit of the Bay Fiord Formation and the basin-fill mudrock of the Cape De Bray Formation (Harrison & Bally, 1988). The stratigraphic sequence on Cornwallis Island is similar in that it contains two possible weak detachment layers in the evaporites of the Baumann Fiord and lower Bay Fiord formations. For a complete review of Mellville Island's salt-based fold belt refer to Geological Survey of Canada Bulletin 472 (Harrison 1995).

4.2.1 Characterization of structures of the Parry Islands Fold Belt, Melville Island

The structural geology of Melville Island is characterized by a southerly directed thin-skinned, salt-based fold belt (Harrison & Bally, 1988). The fold belt has in common many of the characteristics of evaporite based fold and thrust belts described by Davis and Engelder (1985, 1987). A more unique feature of this fold belt is a pronounced
structural disharmony that exists between exposed surface structures and an underlying fold and thrust belt, (Fox 1983, 1985; Harrison & Bally, 1988; Harrison 1995). The disharmony between the two structural levels can be explained in terms of multiple detachment levels (Harrison and Bally, 1988; Harrison, 1995). (Refer to figures 2.5a and 2.5b for a representative view of the surface and sub-surface geology of the Parry Islands fold belt on eastern Melville Island.)

A significant feature to the Parry Islands fold belt is the diversity of observed fold - thrust fault interaction and the lack of a consistently dominant vergence direction. There is an almost equal number south vergent and north vergent thrust faults, and variations in the asymmetry of surface folds are almost equally distributed between north and south verging anticlines (Harrison, 1995). Furthermore, vergence is frequently observed to change along strike of individual anticlines. Vergence reversals along strike of a fold are typically associated with a structural saddle, where fold axes exhibit a concave up plunge reversal (Fig 4.5a). The transfer of vergence along strike in a fold is accomplished through the development of a "pop-up" structure in the lower fold (Fig. 4.5b) (Harrison and Bally, 1988). In addition, detachment folds, fault bend folds, and fault propagation folds are all recognized along strike of individual anticlines (Harrison, 1995).

The stratigraphic column of the fold belt can be divided based on the basis of the mechanical competency of the rock units. These tectonostratigraphic units are described from top to bottom as follows: 1) An upper rigid layer (URL) composed of the Middle and Upper Devonian clastic wedge. 2) An upper ductile layer (UDL) composed of Middle Devonian basin-filled mudrock of the Cape De Bray Formation. 3) A medial
Figure 4.5a Vergence reversals along strike of a fold are characterized with an associated structural saddle, where fold axes exhibit a concave up plunge reversal (Harrison and Bally, 1988).

Figure 4.5b Relaying displacement transfer of subsurface thrusts and the formation of pop up structures. The "pop-ups" are created beneath structural saddles on surface anticlines and mark the location vertically above the point where displacement on a forward verging thrust is almost equal to the displacement on the backward verging thrust (Harrison and Bally, 1988).
rigid beam (MRB) composed of Ordovician through basal Middle Devonian shelf carbonates and coeval basin facies mudrock of the Cape Phillips, Irene Bay, Thumb Mountain, and the upper Bay Fiord formations. 4) A lower ductile layer (LDL) composed of Middle Ordovician evaporites of the lower Bay Fiord formation. 5) A lower rigid layer (LRL) composed of Lower Ordovician carbonates of the Eleanor River Formation and underlying Cambrian strata (Harrison, 1995).

4.2.2 Kinematics of Deformation

A typical characteristic of the folds on Melville Island is a structural disharmony that exists between the surficial expression of the upper fold and an underlying fold and thrust system at depth (Fig. 2.5b). The medial rigid beam is characterized by broad low amplitude folding that is offset by thrust faults; whereas, the upper rigid layer is characterized by open higher amplitude folding that is typically not offset by thrust faults. The style of deformation in these folds is dominated by: 1) Buckling and thrust faulting of the medial rigid beam. 2) Accommodation of the room problems caused by concentric folding being resolved by the mobility of the lower and upper ductile units. 3) Open folding with some minor faulting of the upper rigid layer (Harrison, 1995).

The following model for the kinematics of deformation represents the temporal evolution of salt based fold and thrust structures within the Parry Island's fold belt on Melville Island. This model is based on the integration of surficial, and sub-surface well and geophysical data. Each phase of fold evolution is illustrated in the schematic representation in figure 4.6.
Figure 4.6. Temporal evolution of an anticline within the multiple detachment Parry Islands fold belt, Melville Island. (From Harrison and Bally, 1988.)
Phase 1 Ductile Salt Welt (Fig. 4.6a)

The earliest phase of deformation is characterized by 0-1% shortening and is manifested by the buckling of the competent medial rigid beam into a low amplitude fold. Coevally salt in the lower ductile unit migrates into the axial region of the anticline to form simple short-wavelength welts that resolve the initial hinge area room problems. The term welt refers to a cuspatel shaped body of ductile rock formed in a compressional setting (Harrison & Bally, 1988), and are distinguished from diapiric structures that generally form from gravitational instabilities due to stratigraphic density inversions. Shortening in the upper ductile layer is redistributed through differential compaction over the low amplitude anticline. Therefore, there is little or no apparent shortening in the upper rigid layer.

Phase 2 Breached Salt Welt (Fig. 4.6b)

Upon reaching 3-4% shortening the medial rigid beam experiences irreversible brittle failure, by forward and/or backward verging thrusts faults that typically nucleate in the hinge areas of the anticlines. At this stage the thrust tip may not propagate through the entire section of the medial rigid beam and shortening will then be accommodated by folding. Migration of the lower ductile layer from the synclinal regions of the low amplitude fold continues and the upper ductile layer continues to compact differentially over the hinge area of the anticline. Low amplitude folding begins in the upper rigid layer.

Phase 3: Completed Ramps (Fig. 4.6c and 4.6d)

Upon reaching 5-6% shortening, compressive deformation is characterized by the full vertical linking of the lower and upper ductile layers through a single thrust or
several thrust imbricates. Shortening may also be accomplished through an intermediate detachment level within the medial rigid beam, where a thrust will propagate with an opposite sense of vergence through the overlying section (Fig 4.6d). The formation of an upper welt is accomplished through the thickening of the upper ductile layer in front of the flattening principal thrust. The upper anticline continues to grow, but at this point maintains a low amplitude and long wavelength profile. Migration of the lower ductile unit from the synclinal regions of the lower anticline may at this stage be completed.

*Phase 4 Upper Ductile Layer Welt (Fig. 4.6 e)*

Upon reaching 6-8% shortening there is a rapid increase in the amplitude of anticline in the upper rigid layer, that is accompanied by the migration of the upper ductile layer into the anticlinal hinge to resolve room problems created by folding in the upper rigid layer. In the advanced stages of flowage of the upper ductile layer, the limbs of the surface anticline steepen dramatically and the ductile rock in the hinge region of the welt becomes pinched forming an upward pointing cusp. This results in the creation of an upper welt or cusp stacked vertically above the lower ductile layer welt. At this stage in the fold development the migration of the lower ductile unit from the synclinal regions of the lower anticline is completed, resulting in a greatly increased basal shear stress required for failure. If there is a preferred vergence exhibited by major thrusts in the medial rigid beam, then the lower ductile layer welt will grow asymmetrically.

Convergent thrust faults propagating above the lower ductile layer welt can cause partial or complete encapsulation by evaporites of fault bounded tectonic fragments of the medial rigid beam.
Phase 5 Breaching of the Upper Ductile Layer Welt. (Fig. 4.6f)

Beyond 8% shortening, the upper rigid layer may deform by brittle failure. At this point the upper welt is breached and thrusts propagate into the upper rigid layer. At this stage the overall view of the fold is one in which brittle deformation is concentrated within a panel beneath each surface anticline. The disharmony between the surface folds and subsurface structure is accomplished through a complete linkage of upper and lower detachment levels through both forward and backward verging thrust faults each of which propagate through their respective overlying rigid layers.

4.3 Structural cross sections of the Stanley Head region, western Cornwallis Island

I use the structural geometry and kinematics of deformation of the Stanley Head anticline as a useful analogue to characterize the structural geology of the Cornwallis fold belt. Recent G.S.C. mapping, of which this project is a part, has also identified fold and thrust deformation in the central and northwestern portions of Cornwallis Island respectively (Dewing & Turner, in prep; Jober, in prep). The structures found in these areas are similar to those found in the Stanley Head region, and together these areas define a fold and thrust belt across the island. The Stanley Head region of western Cornwallis Island provides an excellent opportunity to characterize the structural geometry and infer the kinematics of deformation for the Cornwallis fold belt as outcrop exposure was good and allowed for detailed 1:10000 scale mapping, which is the most detailed mapping to date.

Six cross section lines (A, B, C, D, F, and W) have been utilized to construct six D3 deformed state cross sections and five D1 deformed state cross-sections. Cross
sections have been constructed from the integration of surface geology and the application of the kinematic model of deformation for the multiple detachment evaporite based Parry Islands fold and thrust belt on eastern Melville Island (Harrison & Bally, 1988). The cross sections presented are permissible but are not unique solutions.

4.3.1 Application of the Harrison and Bally model to the Stanley Head region, western Cornwallis Island

The application of Harrison and Bally's (1988) model to the area of study is merited for the following reasons: 1) Both the Parry Islands and Cornwallis fold belts are associated with compressive deformation, and 2) The stratigraphic columns of both Melville and Cornwallis islands are characterized by the presence of two possible weak detachment layers. In addition, attempts to balance cross sections across each structural domain in the Stanley Head anticline proved problematic using a single detachment model within either the Baumann Fiord Formation or the lower Bay Fiord Formation when using either a fault propagation, fault bend, or detachment folding mechanism model. Therefore, two detachment horizons provide the most geologically reasonable cross section.

To justify the use of multiple detachment layers in the sub-surface geometries of the Stanley Head anticline it is necessary to examine the thrust origins of the Boothia uplift. Interpretive cross sections of the southern and northern segments of the Boothia uplift based on the gravity data and models of Berkout (1973) and Sobczak and Halpenny (1990) illustrate thrust faults originating in the crystalline basement that propagate through the Paleozoic sedimentary succession. As these thrust faults propagate through
the Paleozoic sedimentary succession it is reasonable to assume there is the potential for numerous detachment surfaces. In the case of the Stanley Head region of Cornwallis Island there are two known potential detachment surfaces in the weak evaporitic Baumann Fiord and lower Bay Fiord formations. Other potential detachment surfaces may be present within the Cambrian strata overlying the crystalline basement; however, because little is known about the Cambrian sediments the cross sections presented do not involve these strata.

Application of Harrison and Bally's (1988) model to the area of study requires the evaporites of the Baumann Fiord and lower Bay Fiord formations deform ductily to form compressional welts. The large gypsum body exposed between Hill B and the Southern Block thickens from north to south, and is interpreted to be a compressional welt structure. From Hill B to the southern margin of Hill A the gypsum is present as a narrow neck that defines the hinge zone of the Stanley Head anticline. South and along strike of the Stanley Head anticline, at the southern margin of Hill A, the gypsum swells outward into a large ellipsoid shape that terminates along the northern margins of the Southern Block. From this rapid change in thickness it is inferred that the lower Bay Fiord Formation flowed and deformed in a ductile manner.

At the onset of the Boothia Uplift the lower Bay Fiord and Baumann Fiord formations would have been buried at depths of 1.8 km and 3.25 km respectively. Using a surface temperature of $30^\circ$C and a geothermal gradient of $30^\circ$C/km the temperature at the base of the lower Bay Fiord and Baumann Fiord formations would have been $84^\circ$C and $127.5^\circ$C respectively. The Baumann Fiord Formation would have deformed ductily at the onset of the Boothia Uplift as the critical temperature of $100^\circ$C was reached. In the case
of the lower Bay Fiord Formation a burial temperature of $84^\circ$C may have been insufficient to promote ductile deformation according to the laboratory experiments of Muller (1981). However, the presence of intraformational fluids and the probability of excess pore pressure being developed in the formation as deformation progressed may make the gypsum/anhydrite even weaker at geologic strain rates than the low strength suggested by laboratory experiments (Davis and Engelder, 1987).

In an attempt to quantify the temperatures of deformation during the Boothia Uplift, thermal maturity data in the form of vitrinite equivalent graptolite reflectance is analyzed from Little Cornwallis Island. The vitrinite reflectance data presented by Hureoux et al. (2000) can be used to establish a peak temperature in the burial history of the rocks and thus constrain the temperature of deformation during the Boothia uplift. This data demonstrates that the temperature exceeded $100^\circ$C at base of the Cape Phillips Formation, indicating that both the lower Bay Fiord and Baumann Fiord formations would have deformed ductily during the Boothia Uplift.

Graptolite reflectance data was collected from the Cliff member of the Cape Phillips Formation, the contact between the lower and upper Thumb Mountain formations, and from the base of the Thumb Mountain Formation. Rocks were collected from locations that were sufficient distance from the Polaris ore body so that they would not be influenced by elevated temperatures associated with the mineralizing fluids (Hureoux et al., 2000). Therefore the $R_0$ (mean random reflectance) values obtained reflect background values associated with burial.

The vitrinite reflectance geothermometer presented by Barker and Pawlewicz (1994), is calibrated against peak temperatures based on re-equilibrated fluid inclusion
homogenization temperatures, and gives peak burial temperatures. \(R_o\) values (Hureoux et al. 2000) and associated peak temperatures are presented below in table 4.1

Table 4.1

<table>
<thead>
<tr>
<th></th>
<th>(R_o) %</th>
<th>Burial Temp. °C</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cliff member: Cape Phillips Formation</td>
<td>0.73</td>
<td>111°</td>
</tr>
<tr>
<td>Lower/upper Thumb Mountain Formation contact</td>
<td>0.77</td>
<td>115°</td>
</tr>
<tr>
<td>Base of the Thumb Mountain Formation</td>
<td>0.88</td>
<td>126°</td>
</tr>
</tbody>
</table>

[Peak temperatures are calculated using the Barker and Pawlewicz (1994) vitrinite reflectance geothermometer]

These peak temperatures indicate an elevated geothermal gradient of 38.5° per km, and burial temperatures that are inconsistent with the known stratigraphy of Cornwallis Island. Essentially, the burial temperature of 111° for the Cliff member of the Cape Phillips Formation would have required a corresponding burial depth of 2.8 km applying a geothermal gradient of 38.5° and a surface temperature of 30°. This would require an extra 1.8 km of strata to be placed on top the Cape Phillips Formation, a scenario that cannot be resolved through sedimentary deposition. I propose that the elevated burial temperatures associated with vitrinite reflectance reflect tectonic thickening of the sedimentary package due to folding and thrusting during the Boothia Uplift.

4.3.2 Methods of cross section construction

The construction of each cross section involved the projection of data (eg. topographic profile, stratigraphic contacts, and fault and bedding plane measurements)
onto the line of section. No subsurface data exists. All formations above the Eleanor River (OE) Formation were projected onto the sections based on structural field measurements. Extrapolation of surficial data and the application of the Harrison and Bally (1988) model were used to elucidate a viable structural geometry at depth with respect to the Eleanor River (OE) and Baumann Fiord (OB) formations.

Cross section lines are drawn such that they are perpendicular to both the individual trend of each structural domain (Hills A-D, F) and to the regional plunge direction of the Stanley Head region. This requires that the section lines have an orientation of 250° outside of the defined structural domains so that they are perpendicular to the overall regional trend of 340° for the Stanley Head anticline. Each structural domain within the Stanley Head anticline is separated by two tear faults and is characterized by unique bedding orientations, fold axis orientation, and the overall sense of vergence that define the structural geometry of each structurally bounded block. Across each structural domain the section lines are perpendicular to the observed structural trend within the core of the anticline. Cross section lines were drawn in this way for the following reasons: 1) the cross section lines maintain a constant projection so that they do not cut up or down section, and 2) the geometry of the Stanley Head anticline is accurately portrayed within each structural domain.

Fundamental fold and thrust belt rules were applied in the construction of all sections, these rules include: 1) Thrusts cut up-section in the direction of transport and tend to die out in folds; 2) Competent bed thickness remains constant; 3) Cut off angles of thrusts are steepest in the strata of greatest competence; 4) Older rocks are thrust over younger rocks; 5) The volume of rock is conserved during deformation; 6) The deformed
state cross section should be palinspastically restorable. Other general rules applying to thrust belts concerned with vergence, kinematics and relative timing of fault movement will probably not apply to the evaporite based fold and thrust system of western Cornwallis Island.

Geometries predicted for evaporite based fold and thrust belts (Davis & Engelder, 1985) developed using mechanical models for fold and thrust belts (Davis et al. 1983; Dahlen et al. 1984) were applied to the cross sections. These geometries include an exceptionally narrow cross sectional taper and an observed lack of a consistently dominant vergence direction of model thrust faults within lower fold and thrust structure. In addition, kink style folding is pervasive throughout the map area and is applied to all cross sections.

**Bed length, Area, and Volume Balancing**

The construction and completion of the structural cross sections was accomplished through bed length balancing of rigid units and area/volume balancing of ductile units. The equal bed length method assumes that all bed lengths of rigid units in a cross section are conserved during deformation so that bed lengths are identical in both the deformed state and restored cross sections. Ductile units do not maintain a constant deformed state thickness, for these units cross sectional areas are measured and restored to an identical area that also corresponds to the restored length of the competent units.

Bed length balanced horizons on the cross sections include the Eleanor River Formation, the upper Bay Fiord Formation, the Thumb Mountain Formation (upper and lower members), the Irene Bay Formation, and the Cape Phillips Formation. Area balanced horizons include the Baumann Fiord Formation and volume balanced horizons...
include the lower Bay Fiord Formation. The area balancing method of restoration assumes that the magnitude of horizontal shortening was identical amongst all competent and ductile units, and that no material is lost or gained into the line of section. The volume balancing method of restoration assumes that the volume of rock is identical in three dimensions parallel and perpendicular to the structural trend of the Stanley Head anticline in both restored and deformed state cross sections.

Construction of D1 and D3 cross-sections

Construction of the D1 and D3 cross-sections began with the field data that provided the basis for the surficial and, in part, underlying folds in the cross-sections. The D3 cross-sections represent the true deformed state cross-sections. However, in an attempt to characterize the structural style and geometry of a Cornwallis fold belt structure the D1 deformed state cross sections were constructed. These cross sections removed the structural complexity associated with normal faulting, giving a view of the Stanley Head anticline as it would have been following the Boothia Uplift.

Construction of the D1 and D3 cross sections was an iterative process that used a combination of forward and then inverse modeling. The starting point for all cross sections was the surface expression of the geology, and in particular the determination of a structural "horst" which was not displaced by normal faulting. These horsts were invariably located in the hinge zone of the anticline (see map), and coincided with the location of the branch line for the axial planes of the upper fold. From field observations, it was determined that the branch line of the axial planes was located at, or just above, the contact between the lower and upper Bay Fiord Formation. These axial planes were then projected outward as undeformed by the D3 event, providing the basis for the shape of
the D1 fold. The model stratigraphic section presented in figure 3.3 was then applied atop the horst with beds folding at the axial planes. At this stage the D1 fold’s geometry is established.

The next step in the construction of the cross-sections involved estimating dips of the normal faults, as field measurements were typically not available. The dips of these faults are constrained from field observations and fault plane orientation measurements to range from 75-90°. This was an iterative process that involved ensuring that the surface expression of the geology was accurate once displacement on the normal faults occurred. At this stage the D3 deformed state cross section was complete.

Cross-section parameters

1) Division of Stratigraphic column into mechanical stratigraphic units

The stratigraphic column was divided into the following units based on their mechanical competency (Fig. 4.7a). These are described top to bottom and include: 1) An upper rigid layer (URL) composed of the Cape Phillips, Irene Bay, Thumb Mountain, and the upper Bay Fiord formations. 2) An upper ductile layer (UDL) composed of the lower Bay Fiord formation. 3) A medial rigid beam (MRB) composed of the Eleanor River Formation. 4) A lower ductile layer (LDL) composed of the Baumann Fiord Formation. 5) A lower rigid layer (LRL) composed of competent Cambrian sediments and the underlying crystalline basement. The ordering of these tectonostratigraphic units, are in essence, the same as those outlined by Harrison (1995) for Melville Island’s salt based fold belt, although they are composed of different formations. Refer to figure 4.7 for a comparison between the mechanical stratigraphy on Melville and Cornwallis islands.
Figure 4.7
a) Stratigraphic column of the Stanley Head region illustrating the mechanical stratigraphic units.
b) Stratigraphic column of the Parry Islands fold belt illustrating the mechanical stratigraphic units. Note: Thicknesses of formations from Melville Island (Harrison, 1995).
2) Critical Taper

The critical taper, that would have supported the development of the Cornwallis Fold Belt, is at present unknown because; 1) there is no subsurface data to elucidate the dip of the basal decollement, \( \beta \), and 2) the paleotopographic slope, \( \alpha \), of the ancient mountain belt has been eroded. Because the Cornwallis Fold Belt is interpreted to be an evaporite based fold and thrust belt, it is reasonable to assume that the critical taper of the wedge is on the order of a few degrees.

For the purpose of constructing the cross sections a horizontal dip was given to \( \beta \), the dip of the basal decollement. This is a reasonable estimation if the Parry Islands and Cornwallis fold belts are comparable. The seismically imaged dip of \( \beta \), on Melville Island is given as 0.1° from 0-50 km distance from the foreland (Harrison, 1995). If the interpretation that the Boothia Uplift is a west verging deep-seated thrust block is correct, then the Stanley Head Region of western Cornwallis Island is approximately 50 km from the from the undeformed foreland on eastern Bathurst Island, the western termination of the Cornwallis Fold Belt. Therefore, the assumption of a near horizontal dip for the basal decollement for the Stanley Head region is justified since it is within 50 km of the undeformed foreland.

3) Anticlinal welts with the Baumann and Bay Fiord Formations

Compressional welts are proposed to have developed during the D1 event within the Baumann Fiord and lower Bay Fiord formations, in accordance with the Harrison and
Bally (1988) model. The evaporites of these welts migrate from synclinal regions into the hinge zones of the anticlines.

4) Geometry of the proposed lower fold situated beneath the lower Bay Fiord Formation

This fold is modeled as symmetric with limbs exhibiting two dip domains of 4° and 15°. These limb dips reflect extrapolated bedding attitudes measured to the west of Stanley Head anticline within the Cape Phillips and Bathurst Island formations. Bedding attitudes measured within the Cape Phillips Formation on the western margin of the Stanley Head remain constant from Hill A -F (south to north), providing a good basis for the modeling of the lower fold. The wavelength of the fold is interpreted to be 11 km as determined from the presence of horizontal bedding observed within the Bathurst Island Formation located approximately 5.5 km from the observed hinge zone of the upper fold of Stanley Head Anticline.

5) Conjugate thrust faults within the medial rigid beam

The cross sections were modeled using conjugate thrust systems within the Eleanor River Formation. The proposed narrow taper of the panel of rock dictates that the hypothetical orientation of the maximum principle stress is likely within a few degrees to being parallel to the basal decollement. In this case, the two candidate thrust faults that will form according to Coulomb’s fracture criterion will have nearly equal dips and therefore a nearly equal chance of forming.

6) Characterization of thrust fault dips within the Stanley Head anticline:

In the case of brittle failure, the relationship between the angle of the thrust fault and the competence of the rock through which it propagates is expressed by the friction coefficient $\mu$, such that $\mu = \tan\phi$. From $\mu = \tan\phi$, the two candidate slip planes that will
form during brittle failure according to the Coulomb failure criterion will be symmetrical about the $\sigma_1$ axis at an angle, $\theta$, defined by the relation $\theta = (45^\circ - \phi/2)$. For horizontal attitudes of $\sigma_1$, higher fault angles imply higher friction coefficients and greater bed competence. In the case of thrust faults propagating through the very weak lower and upper ductile units the angle of the thrust faults is dictated by the yield strength of the evaporite. Therefore the fault angles within the medial and upper rigid beams are greater than those observed within the lower and upper ductile layers.

Thrust fault orientations within the Thumb Mountain Formation obtained from the Parry Islands fold belt thrust fault demonstrate that the north vergent thrusts had a mean dip of $34^\circ$ and south vergent thrusts had a mean dip of $27^\circ$. Comparing the bed competency of the Thumb Mountain Formation with the Eleanor River Formation is then necessary to estimate probable thrust fault angles within the Eleanor River Formation. The Eleanor River Formation is characterized, in part, by two recessive units (Thorsteinsson, 1988) and is not as competent as the Thumb Mountain Formation which has no recessive units. Therefore an estimate of a $25^\circ$ dip for thrust faults within the Eleanor River Formation seems reasonable.

The hinge zone and inflection points of the lower fold serve as common locations for the locus of thrust ramps (Harrison, 1995). Thrust faults were positioned such that they would begin to propagate through Eleanor River formation at the inflection points of the lower fold. These points represent zones of inherent weakness in the medial rigid beam caused by initial buckling.

The magnitude of the displacement on the thrust faults satisfied the requirement that bed length of the medial rigid beam would be equal to the bed length observed in the
upper rigid layer to give a "balanced" cross section. Displacement on the thrust faults was distributed equally between convergent thrust faults.

7) Normal faults

Normal faults are given a vertical to sub-vertical orientation based on fault plane measurements taken from the Stanley Head anticline and the Intrepid Bay graben. The normal faults within the Stanley Head anticline are interpreted to be kinematically linked to the graben bounding faults because they are directly along strike and almost continuous in outcrop from the graben to the anticline. Normal faults are interpreted to extend below the Baumann Fiord Formation for the following reason: The calculated thickness of the Eureka Sound Formation in the Intrepid Bay graben is 720 meters, indicating that the graben bounding faults had at least that much vertical displacement. The displacement on these faults likely did not dissipate within the evaporites of the Baumann Fiord Formation as it would require the migration of nearly the entire 750 m of gypsum from underneath the graben. These faults likely terminated below the base of the Baumann Fiord Formation, presumably within the underlying Cambrian sediments. If the normal faults within the Stanley Head anticline are linked to the graben bounding faults it is reasonable to assume that they also extend beyond the base of the Baumann Fiord Formation. The geometry of the normal faults at depth within the Stanley Head anticline and Intrepid Bay graben is unknown.

4.3.3 The balancing problem: cross sections A-A', B-B', C-C', D-D'

The observed development of the compressional welt within the lower Bay Fiord Formation and the proposed development of a compressional welt within the Baumann
Fiord Formation illustrate a change in thickness of these evaporites. The evaporitic horizons must then be area balanced, with the principle assumption that no material enters or leaves the line of section, otherwise known as the plane strain constraint.

In the case of the anticlinal welt developed within the Baumann Fiord Formation, there is no field evidence to suggest that the gypsum/anhydrite flowed perpendicular to the lines of section, so it is assumed that the plane strain constraint is valid. This allows the Baumann Fiord Formation to be successfully area balanced.

In the case of the anticlinal welt developed within the lower Bay Fiord Formation, the plane strain constraint is likely not valid. From map patterns, the lower Bay Fiord Formation can be seen to change thickness along strike of the Stanley Head anticline. From the D1 deformed state cross sections A-A', B-B', C-C', and D-D' it is observed that the lower Bay Fiord Formation would not area balance when restored. In fact, the area of Bay Fiord Formation gypsum in each of the D1 deformed state cross sections, when restored, would be insufficient to account for a 200 m thickness of the unit. This leads to two possibilities, either the 200 meter estimate for the lower Bay Fiord Formation is too great, or the unaccounted for Bay Fiord Formation gypsum flowed parallel to the trend of the anticline and is present within the gypsum body observed between Hill A and the Southern Block. In the first case it is possible to area balance the Bay Fiord gypsum in each cross section, simply by changing its estimated thickness. However, this does not address the origins or volume of the compressive welt observed between Hill A and the Southern Block. This leads to an attempt of a three-dimensional volume balancing of the lower Bay Fiord Formation.
Volume Balancing the lower Bay Fiord Formation

The fundamental premise behind the volume balancing of gypsum/anhydrite of the Bay Fiord Formation is that the formation of the large compressive welt observed between Hill A and the Southern block is kinematically linked to the formation of the Stanley Head anticline on Hills A-D. In these structural domains the Stanley Head anticline is characterized by a detachment style of folding in which the fold tightens from south to north across each structural domain. It is here proposed that during the formation of the upper detachment fold observed on Hills A-D, gypsum from Bay Fiord Formation flowed north to south, along the hinge zone of the Stanley Head anticline, into the compressive welt found between Hill A and the Southern Block.

To volume balance the gypsum of the lower Bay Fiord Formation it is necessary to calculate the amount of area lost on each D1 deformed state cross section in comparison to an undeformed restored package of rock. Area loss of the Bay Fiord gypsum was calculated for the cross sections A-A', B-B', C-C', and D-D' by measuring the area of gypsum in the D1 deformed state cross sections between pin lines #2 and #3 and subtracting this area from a corresponding restored state cross sections, in which the lower Bay Fiord Formation has a thickness of 200 m. The area loss of each D1 deformed state cross section was then converted into a volume loss by multiplying the area deficiency of the Bay Fiord gypsum for each of the cross sections by the corresponding length of the structural domain. The results are presented in table 4.2.

Table 4.2. Volume of gypsum loss from Hills A-D

<table>
<thead>
<tr>
<th>Cross section</th>
<th>Area loss of Bay Fiord gypsum (m²)</th>
<th>Length of structural domain (m)</th>
<th>Volume loss (m³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A-A'</td>
<td>5.40 x 10⁶</td>
<td>1005</td>
<td>5.44 x 10⁸</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-------</td>
<td>-------</td>
<td>-------</td>
<td>-------</td>
</tr>
<tr>
<td>B-B'</td>
<td>1.12 x 10^6</td>
<td>1100</td>
<td>1.23 x 10^9</td>
</tr>
<tr>
<td>C-C'</td>
<td>1.70 x 10^6</td>
<td>390</td>
<td>6.63 x 10^8</td>
</tr>
<tr>
<td>D-D'</td>
<td>1.55 x 10^6</td>
<td>800</td>
<td>1.24 x 10^9</td>
</tr>
<tr>
<td>Totals</td>
<td>4.91 x 10^6</td>
<td></td>
<td>3.67 x 10^9</td>
</tr>
</tbody>
</table>

**Development of the upper compressional welt in three dimensions**

The upper compressional welt may be separated into the following two physical regions: 1) The area of deposition that occupies the present day ellipsoidal region of the upper compressional welt. 2) The source area of the gypsum/anhydrite, which is comprised of the structural domains Hills A-D. The boundary between the two areas is interpreted to coincide with the southern termination of the Bay Fiord limestone on the west limb of Hill A (see map).

The following discussion of the development of the upper compressional welt will follow the Harrison and Bally (1988) model for the kinematics of deformation as applied to the Stanley Head anticline. During the formation of the upper detachment fold the gypsum/anhydrite of the Bay Fiord Formation will flow into the axial region of the anticline to resolve initial room problems created by folding the upper rigid layer. In the advanced stages of gypsum/anhydrite migration the limbs of the upper detachment fold have steepened dramatically, as the fold becomes tighter with progressive tectonic shortening. At some point the fold becomes sufficiently tight that the original volume of the upper ductile welt can no longer be accommodated within the core of the upper detachment anticline. At this point the evaporitic rocks of the upper ductile layer will flow horizontally to the area of deposition, an assumed region of lower stress, where the surficial anticline would have a more open geometry.
Evidence for horizontal flow within the Bay Fiord gypsum is based on the following: The map pattern demonstrates that the gypsum of the lower Bay Fiord Formation is continuous from Hill B to the northern margin of the Southern Block. The thickness of the gypsum changes rapidly from north to south along Hill A. The plunge of the Stanley Head anticline is near horizontal along Hill A, indicating that the gypsum changes thickness horizontally along the same structural level. This change in thickness can be accomplished either through the horizontal flow of gypsum, or by upward flow of gypsum as a diapir. The second possibility is unlikely, as the gypsum defines the hinge zone of the Stanley Head anticline along Hills A and B and is thus kinematically linked to the evolution of the fold which has been shown to be a part of a compressive fold and thrust structural setting.

In addition, foliation measurements taken from the lower Bay Fiord Formation on the southern margin of Hill A are 135/70 and 323/77, respectively. Due to the weathered nature of these rocks no lineations were observed. These foliations are sub-vertical, have a strike parallel to the axial surface of the Stanley Head anticline, and are disharmonious with the bedding orientations observed on Hill A. It is proposed that these foliation planes formed due to $\sigma_1$ having a near horizontal plunge oriented perpendicular to the fold axis of the Stanley Head anticline.

The largest calculated volume loss of gypsum/anhydrite within the Stanley Head anticline is from within Hill D and diminishes in magnitude moving south to Hill A (Table 4.2). In each of the structural domains the upper detachment fold has become tight enough that it can no longer accommodate the original volume of the upper ductile welt. The gypsum then flows horizontally from the tightest parts of the Stanley Head
anticline to more open parts of the fold, until it is finally accommodated (deposited) in the ellipsoid region of the compressional welt structure. This requires that the dimensions of the ellipsoid region of the compressional welt are able to accommodate 3.73 x 10^9 m^3, in addition to the volume of gypsum that would have already been present within the structure prior to the influx of gypsum from Hills A-D.

The D1 deformed state cross section W-W' represents a reasonable proposed geometry of the upper compressional welt, immediately after the D1 event. The following assumptions are made with respect to the construction of the D1 cross section W-W': 1) Prior to the deposition of gypsum from Hills A-D, the structure may be approximated by phase 3 of Harrison & Bally's (1988) model, where the surficial anticline is harmonious with the lower anticline. At this stage there is an unthickened package of the upper ductile layer present. 2) During the deposition of excess evaporite from Hills A-D the beds comprising the upper rigid layer deform in a manner such that the bed length of all rigid beds within the cross-section balance. The upper rigid layer is postulated to deform through a complex process of folding and faulting that are analogous to those observed in strata immediately adjacent to gravitationally based diapirs. This condition is required in order for the cross section to balance. The geometry of these proposed structures adjacent to the compressional welt in cross section W-W' are unexposed and therefore their geometries are unknown and are not represented on the cross section.

The cross sectional area of the upper ductile layer in the D1 cross-section W-W' is representative of the entire depositional region of the compressional welt structure, and for the purposes of modeling is kept constant along the length of the structure. The area
The calculated volume of the compressional welt demonstrates, as a first order approximation, that it is possible to accommodate the volume of gypsum expelled from Hills A-D. This allows for the gypsum/anhydrite of the Bay Fiord Formation to balance in three dimensions.

4.3.4 Results from cross sections A-A', B-B', C-C', D-D'

See map and cross section sheets A and B.

Results from these balanced cross sections give valuable insights into the geometry, amount of uplift, and percentage of tectonic shortening of a well constrained Cornwallis Fold Belt structure. The total amount of uplift of each cross section is based on the amplitude of the Stanley Head anticline measured from the D1 cross sections at the Irene Bay Formation - Cape Phillips Formation contact. The amount of tectonic shortening is based on the bed length measurements at the contact between the Irene Bay
and Cape Phillips formations between pin lines 2 and 3. These pin lines are 11 km apart and represent the wavelength of the Stanley Head anticline. Post orogenic extension is measured across the entire section line.

Table 4.4

<table>
<thead>
<tr>
<th>Cross section</th>
<th>Section Length (km)</th>
<th>Post Orogenic extension (m)</th>
<th>Pre-extension section length (km)</th>
<th>Bed length of OSCP/OCI (km)</th>
<th>Shortening of OSCP/OCI (m) &amp; (%)</th>
<th>Total Uplift (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A-A'</td>
<td>12.42</td>
<td>220</td>
<td>12.20</td>
<td>11.43</td>
<td>432 (3.9%)</td>
<td>960</td>
</tr>
<tr>
<td>B-B'</td>
<td>12.72</td>
<td>160</td>
<td>12.56</td>
<td>11.85</td>
<td>850 (7.7%)</td>
<td>1150</td>
</tr>
<tr>
<td>C-C'</td>
<td>12.35</td>
<td>120</td>
<td>12.23</td>
<td>11.98</td>
<td>980 (8.9%)</td>
<td>1230</td>
</tr>
<tr>
<td>D-D'</td>
<td>12.25</td>
<td>115</td>
<td>12.4</td>
<td>11.97</td>
<td>970 (8.8%)</td>
<td>1200</td>
</tr>
</tbody>
</table>

4.3.5 Characterization of cross sections

The cross sections are separated into two groups based on the nature of the observed fold-thrust fault interaction. Cross sections A-A', B-B', C-C', and D-D' are balanced interpretations of the structural geometry of the central portion of the Stanley Head anticline, and are characterized by a detachment style of folding for the surficial anticline. Cross section F-F' is considered to be representative of the structural style observed on Hills E, F, and the Southern block that are characterized by a thrust panel that is interpreted to display only a backlimb of an anticline due to the present level of erosion. Cross section F-F' is not balanced since the fold geometry of the proposed anticline is unknown. The exact nature of the fold-thrust fault interaction of these domains remains uncertain and therefore any speculation regarding what the geometries of these structures would have been following the D1 event will be avoided. However,
cross section F-F' illustrates a geometry in which a thrust ramp propagates through the upper Bay Fiord, Thumb Mountain, and Cape Phillips formations, indicating either a fault-bend or fault-propagation mechanism for folding is probable.

4.4 Fixed versus migrating hinge folding

In fault related folds the distribution of deformation is mainly controlled by fault-fold kinematics (Salvini & Storti, 20001). Fault related folding mechanisms can generally be divided into two categories: 1) Fixed hinge folding, where folding is produced by limb rotation about axial surfaces that are fixed with respect to the panel of rock being folded, and 2) Active hinge folding, where folding involves the lateral migration the rock panel through an axial surface (Fig. 4.8). In fixed hinge folding the anticlinal crests are the preferred site for concentrated deformation, whereas, in active hinge folding deformation is concentrated in the limbs of the folds leaving the fold crests relatively undeformed (Salvini & Storti, 2001). The main characteristics of these mechanisms are presented below in Table 4.5

Table 4.5 Fixed hinge versus active hinge folding (Salvini & Storti, 2001).

<table>
<thead>
<tr>
<th>Fixed hinge folding</th>
<th>Active hinge folding</th>
</tr>
</thead>
<tbody>
<tr>
<td>- Deformation is concentrated in the axial surface areas and, as the fold develops, the dimensions of the deformed zone do not significantly increase.</td>
<td>- Deformation develops in the hanging wall rocks as they pass through an active axial surface.</td>
</tr>
<tr>
<td>- Deformation intensity is concentrated in the hinge zone and decreases towards the limbs.</td>
<td>- Deformation is concentrated in fold limbs, leaving the crest of the fold unaffected.</td>
</tr>
<tr>
<td>- A correspondence between layer curvature and deformation intensity is expected, with the maximum deformation intensity at the fold hinges.</td>
<td>- Continued folding and fault slip produces an increase in the dimensions of these deformed rock panels.</td>
</tr>
<tr>
<td></td>
<td>- Deformation intensity may be assumed to depend on the angle, number and sense of rotation of the deformation panels.</td>
</tr>
</tbody>
</table>
Figure 4.8
Contrasting spatial distribution of deformation associated to fixed hinge folding (a) and active hinge folding (b) respectively Salvini & Storti (2001).
Within the region of study Hills A-D of the Stanley Head anticline are characterized by a detachment style of folding. Within these structural domains the intensity of fractures and veins increases as the fold hinge is approached. This suggests that the detachment style of folding that characterizes Hills A-D was produced by fixed hinge fold development. Hills E, F, and the Southern Block are characterized by thrust ramps that propagate through the Bay Fiord, Thumb Mountain, Irene Bay, and Cape Phillips formations, indicating either a fault bend or fault propagation mechanism for folding. Within these structural domains intense fracturing is evenly spread throughout the exposed fold limb. This suggests that the thrust panel migrated through a fixed axial plane and that an active hinge folding mechanism would have produced the folds.

4.5 Summary of the Stanley Head cross sections

The geologic map and associated structural cross sections of the Stanley Head anticline on western Cornwallis Island provides the basis to begin characterizing the structural style of the Cornwallis Fold Belt. The “balanced” cross sections, constructed with the use of surficial structural data and an application of a previously published model for the kinematics of deformation, illustrate that the Stanley Head anticline can be interpreted and explained as an evaporitic based fold and thrust system, in which there are multiple levels of detachment. Extensive work done on the Parry Island’s Fold Belt (Fox, 1983 & 1985; Harrison and Bally, 1988; Harrison, 1995), based on geologic mapping and seismic data provide a model which can readily be applied to Cornwallis Island due to the similar nature of the mechanical properties of the stratigraphic columns.
The cross sections presented in this thesis were constructed such that they balanced and obeyed the characteristics of a multiple detachment, evaporitic based, fold and thrust belt. This was manifested in the cross-section parameters as: 1) A pronounced structural disharmony between the surficial expression of the Stanley Head anticline, and an underlying fold and thrust system. 2) A horizontal orientation for the dip of the basal decollement, $\beta$, a feature that would be reasonable to expect in a narrowly tapered evaporitic based fold belt. 3) Compressional evaporitic welts within the Bay Fiord and Baumann Fiord formations and associated synclinal thinning of the ductile layers. 4) Conjugate thrusts displacing the lower fold within the Eleanor River Formation.

Balancing of the deformed state cross sections was achieved through bed length, area, and volume balancing techniques. The deformed and restored state cross sections demonstrate that the proposed structural style of the Stanley Head fold and thrust system is a structurally viable and admissible interpretation.

The results from these cross sections are significant with respect to the Cornwallis fold belt because it has been shown that the Stanley Head region is characterized by shallow level evaporite based fold and thrust structures. The structural geometry and kinematics of deformation portrayed in the Stanley Head cross sections provide an excellent basis with which to construct a regional cross section across the Cornwallis fold belt.

4.6 Regional cross section

To characterize the structural style of the Cornwallis fold belt the regional cross section R-R' (see cross section sheet B in pocket) has been constructed across the
northern half of Cornwallis Island (Fig. 4.9). This cross section has been constructed from the integration of surface geology and the application of the Harrison and Bally (1988) kinematic model of deformation for the Parry Islands fold belt. No attempts were made to balance the cross section; however, it was constructed using the reasoning, methods, and cross section parameters outlined in detail for the cross sections of the Stanley Head region of western Cornwallis Island. Maps from the Stanley Head region, Rookery Creek region, and Thorsteinsson's regional map provide the basis for cross section construction. The geology of the regional map (Thorsteinsson, 1988) and Rookery Creek map (Turner & Dewing, 2002) were reinterpreted by the author such that the structures presented in the regional cross section would be characteristic of an evaporitic based, multiple detachment fold and thrust belt.

4.6.1 Regional cross section discussion

The Cornwallis fold belt displays structures representing a range of fold and thrust development and various thrust fault -fold interactions. The main structural feature that characterizes this cross section is the pronounced structural disharmony between the surficial expression of the fold belt, and an underlying fold and thrust system. This disharmony is explicable in terms of detachment levels within the ductile evaporites of the Baumann Fiord and lower Bay Fiord formations. The following discussion will focus on the observed and proposed structures across the cross section moving from southwest to northeast.
Regional geologic map of Cornwallis Island (Thorsteinsson, 1986). This map portrays the distribution of rock types and faults across the island. Note for the purposes of reinterpreting the structural geology of the island, faults are portrayed as black lines without any indication of fault type. Note that only rock formations present within the regional cross section are described on the accompanying legend.
Stanley Head region

The western margins of the cross section transect the Stanley Head region of Cornwallis Island. The structures observed within the Stanley Head region of western Cornwallis Island provide the basis for the structural styles that are applied to the rest of the cross section. The outstanding feature of this portion of the cross section is the Stanley Head anticline that is characterized by both northeast and southwest verging folds and thrust faults. Fold-thrust fault interactions above the lower Bay Fiord Formation are characterized by both detachment style folding and thrust faults that propagate through the upper Bay Fiord, Thumb Mountain, and Cape Phillips formations, in which either a fault propagation or fault bend fold geometry may have been established.

Between pin lines #1 and #2 the cross section is balanced (see discussion in sections 4.3.2, 4.3.3, 4.3.4) and is characterized a full vertical link-up between the lower and upper ductile layers through thrust faulting. The geometry of the underlying fold and thrust belt is governed by: 1) structural measurements that are extrapolated to depth, 2) the Harrison and Bally (1988) kinematic model for deformation of the multiple detachment Parry Islands fold belt, and 3) the necessity that the cross section balance such that the line length of the medial rigid beam equals that of the upper rigid layer.

The surficial anticline is characterized by a detachment style of folding where the upper rigid layer is folded above the anticlinal welt of the upper ductile layer, which in turn is stacked vertically above the anticlinal welt of the lower ductile layer. Synclines adjacent to the Stanley Head anticline exhibit evacuation of evaporites from both the Baumann Fiord and lower Bay Fiord Formations to resolve room problems created by the buckling of the lower and upper rigid layers.
Midshipman Anticline

The Midshipman anticline is characterized by its low amplitude in comparison to structures found at the Stanley Head and Rookery Creek regions of Cornwallis Island. The low amplitude of this fold is interpreted to represent movement predominantly on the lower detachment level within the Baumann Fiord Formation. It is also interpreted that the upper detachment layer is beginning to be activated to account for steeper limb dips observed within the core of the anticline. Between pin lines #2 and #3 the bed lengths of the medial rigid beam and upper rigid layer balance.

Between pin lines #2 and #3 the cross section line is offset by 6.5 km to the northwest, perpendicular to the line of section. A regional north-northwest trending plunge of $5^\circ$ for the Cornwallis fold belt requires 565 m of vertical offset, as illustrated in the cross section.

Rookery Creek region

The Rookery Creek region of the cross section is characterized by comparatively wide surface exposures of the upper Bay Fiord and Thumb Mountain formations. Within an evaporite-based fold and thrust structural setting, convergent imbricate thrust systems can easily account for the observed repeated nature of the stratigraphic section. The northeast verging thrust faults are portrayed as a leading imbricate thrust system; whereas the southwest verging structures are portrayed as a trailing imbricate thrust system.

No attempts were made to balance the cross section between pin lines #3 and #4, as the geometries of the surface structures are poorly constrained. However, the overall width of the proposed imbricate thrust systems at Rookery Creek implies a wide zone of anticlinal welt development in the lower and upper ductile layers, resulting in extreme
thickening in the lower and upper ductile layers. The portrayed geometry of the underlying fold and thrust system within the medial rigid beam is extremely simplified and does not reflect the complexity required to balance the bed lengths of the medial rigid beam with the upper rigid layer.

**Normal faulting**

I propose that the style of normal faulting within the Rookery Creek region differs from that in the Stanley Head region. Normal faults within the Rookery Creek region are interpreted to terminate within the extremely thickened package of the lower Bay Fiord Formation, whereas normal faults within the Stanley Head region are interpreted to terminate below the base of the Baumann Fiord Formation. Normal faults within the Rookery Creek region are interpreted to terminate within the lower Bay Fiord Formation because the displacement on the normal fault can be accommodated by flow within the thickened upper ductile layer. Normal faults at Stanley Head are interpreted to extend below the base of the Baumann Fiord for following reasons: 1) The presence of 750 m of Eurkea Sound formation within the Intrepid Bay graben indicates that normal faults must terminate below the 750 m thick Baumann Fiord Formation, and 2) Rift related volcanics are observed on eastern Bathurst Island indicating deep seated extension.

**4.6.2 Conclusion**

In summary, the schematic regional cross-section provides insight into the various structural styles and the range of strain accommodated by structures across the Cornwallis fold belt. These structures are characterized by: 1) Southwest and northeast verging folds and thrust faults, 2) a variety of fold - thrust fault interactions, including
detachment style folding and thrust imbricate systems, and 3) a proposed structural disharmony between the surficial fold belt and an underlying fold and thrust system.
Chapter 5

STABLE ISOTOPE STUDY

5.1 Introduction

The purpose of this study is to delineate the origin of the paleo-fluids responsible for veining within the Stanley Head region of western Cornwallis Island. Calcite veining is pervasive throughout the area of study and typically increases in intensity with increasing proximity to normal fault zones and the hinge zone of the Stanley Head anticline. Barite veining is spatially limited to the strike slip fault zone within the Southern Block. To address the origins of the fluids in the study area oxygen and carbon isotopic data was obtained from calcite veins and their corresponding wall rocks. Sulphur isotope data was collected from the largest barite vein in the Southern Block. The results from the stable isotope data are compared to similar studies done at the Polaris mine site in an attempt to establish whether or not the isotopic signatures are comparable, and therefore possibly related, as part of a regional fluid system.

To put into context the regional fluid regime with respect to the area of study, a brief discussion of the Polaris mine and structural geology of the Stanley Head region is merited. The study area is located south and along strike from the Late Devonian, Zn-Pb Mississippi Valley type Polaris mine. The deposit is interpreted to have formed from a hydrothermal system that first produced dolostone through replacement of limestone, then deposited Zn, Pb, and Fe sulphides through replacement of the dolomite in dissolution pores, and finally deposited calcite in fractures and pores that post-date all mineralization (Savard et al., 2000). Three phases of deformation have been documented in the study area: 1) northeast and southwest verging folds and thrust faults associated
with an east-west compressional D1 event, 2) strike slip faulting (with associated normal faulting) and folding associated with a north-south compressional D2 event, and 3) normal faulting associated with east-west extensional D3 event. The D1, D2, and D3 events are interpreted to correspond to the late Silurian - early Devonian Boothia Uplift, the late Devonian - Early Carboniferous Ellesmerian Orogeny, and the late Cretaceous - early Tertiary Eurekan event, respectively.

5.1.1 Spatial distribution of fracturing and veining within the Stanley Head anticline

The intensity and spatial distribution of fractures and calcite veins in study area is interpreted to be controlled by predominantly two factors: 1) the folding mechanism that produced the Stanley Head anticline, and 2) normal and strike slip faulting. The structural domains within the Stanley Head anticline may be separated into two groups based on the folding mechanisms. Hills A-D are characterized by detachment style folding, where the fold is interpreted to be produced through a fixed hinge fold mechanism. In these domains fractures and veins are concentrated in the hinge zone of the anticline. Hills E, F, and the Southern Block are characterized by thrust ramps that propagate through the Bay Fiord, Thumb Mountain, Irene Bay, and Cape Phillips formations, suggesting either a fault bend or fault propagation fold mechanism. In these structural domains intense fracturing and veining is spread throughout the exposed thrust panel, suggesting an active hinge folding mechanism. Of the structural domains characterized by an active hinge folding mechanism, the Southern Block has the greatest intensity of fractures and veins. In fact, the intensity of the fracturing and veining does
not allow for reliable formation identification in the southern margins of the domain (see map).

**Calcite veins**

Calcite veins are typically present as an irregular stockwork (Fig. 5.1a) and increase in frequency with increasing proximity to normal fault zones and the hinge zone of the Stanley Head anticline. Veins are also present as planar features that have filled extensional and shear fractures (Fig. 5.1b). Locally both vein types comprise up to 20% of the rock volume. Calcite veins are interpreted to be associated with a specific deformation event based on their spatial proximity to either D1, D2, or D3 structures.

**Barite veins**

Barite veins and fault breccia in the Stanley Head anticline are found only within the Southern Block in close proximity to the strike slip fault zone and are locally cross-cut by calcite veins. These veins exhibit variable continuity and thickness, but are typically 0.5 - 3 cm thick. One poorly exposed barite vein is approximately 20-30 cm in thickness, dips near vertically and has a trend parallel to the strike slip fault.

**Sample selection**

Figure 5.2 illustrates the spatial distribution and stratigraphic horizon where each sample originates from within the Stanley Head anticline. Samples for carbon and oxygen isotope analysis were chosen such that they were in close proximity to either D2 normal faults associated with the strike slip fault zone in the Southern block, or D3 normal faults on Hills C, D, and E. This was done in an attempt to recognize if there would be different isotopic signatures from the D2 and D3 calcite veins. Sample #137 was collected from a small-scale normal fault that was parallel in trend to the Stanley
Figure 5.1
Representative vein types from the Stanley Head anticline. a) Irregular stockwork fractures and veins. Photo is of sample # 82 located 10 meters from a D3 normal fault on Hill D within the Upper Thumb Mountain Formation. b) Planar calcite veins filling extensional and shear fractures. Photo is of sample #39, located within the Lower Thumb Formation from the Southern Block.
Figure 5.2
a) Simplified geologic map of the Stanley Head anticline illustrating spatial distribution of samples. Sample locations are delineated by yellow stars. b) Stratigraphic column to accompany figure 5.2a, yellow stars delineate formations sampled for carbon and oxygen isotope analysis.
Head anticline. Veining within all samples is characterized by a stockwork morphology due to their close proximity to the fault zones, although sample #39 also exhibits planar veins. One sample was collected from western fold limb of Hill A in an area of continuous stratigraphy that was not associated with any faulting. This sample was selected in an attempt to identify an isotopic signature that may reflect the origin of fluids during the D1 event.

Seven of the eight samples were collected from the lower and upper Thumb Mountain Formations. The eighth sample was collected from the Cliff member of the Cape Phillips Formation. The limestone formations from which these samples originate are very similar in lithological characteristics and age. The limestones are clean mudstones or wackestones composed predominantly of micritic lime mud and are Middle - Upper Ordovician in age.

The sample for sulphur isotope analysis was chosen from one of the thickest barite veins in close proximity to the strike slip fault in the Southern Block. The isotopic signature from this sample is compared to sulphur isotopic signatures from barite at the Polaris mine site and at the Truro Pb-Zn showing on Little Cornwallis Island.

5.2 Stable Isotope analysis

5.2.1 Analytical Methods

Carbonate powders were drilled from slabs using diamond impregnated drill bits in a Dremel tool, with powders being derived from less than 5 mm³ of sample material.
Powders were analyzed for their stable isotope values using the techniques outlined in McCrea (1950) for liberation of CO$_2$ by reaction with H$_3$PO$_4$. The extracted gas was analyzed on a gas source, isotope ratio mass spectrometer at Washington University (St. Louis, Missouri). Oxygen and carbon isotopic values are expressed in the conventional notation and given in per mil ($^\circ$/oo) relative to SMOW and VPDB standards respectively. The precision of the data is better than $\pm 0.2^\circ$/oo.

Sulphur isotope analysis was carried out at the University of Calgary using Continuous Flow-Isotope Ratio Mass Spectrometry. Between 100 - 300 $\mu$g of pure homogenized sample is combusted to analyze SO$_2$ gas. Sulphur isotopic values are expressed in conventional notation and given in per mil ($^\circ$/oo) relative to V-CDT standard. The precision of the data is better than $\pm 0.7^\circ$/oo.

5.3 $\delta^{18}$O and $\delta^{13}$C Results

Results of the stable isotope analysis are presented in table 5.1 and figure 5.3. The following trends are apparent from the stable isotope analysis: 1) All samples, except #37, exhibit a depletion of $^{18}$O in the calcite veins with respect to the wall rock. The range of $^{18}$O depletion in the veins is from 4.8 $^\circ$/oo in sample #103 to 13.7 $^\circ$/oo in sample #85. 2) Wall rock $\delta^{18}$O values from the Southern Block are systematically lower than wall rock values from the rest of the Stanley Head anticline. $\delta^{18}$O values of wall rocks from the Southern Block range from 16.8 - 19.9 $^\circ$/oo, whereas $\delta^{18}$O values of wall rocks from Hills A, C, D, and E range from 20.0 - 25 $^\circ$/oo. 3) There is a difference in $^{18}$O depletion in veins associated with D2 normal faults from the Southern Block and veins associated with D3 normal faults from Hills C, D, and E. Three of the four samples from
Figure 5.3. $\delta^{13}\text{C} - \delta^{18}\text{O}$ diagram showing the isotopic compositions of calcite veins and wall rocks in the Stanley Head anticline. Box Mc indicates the isotopic composition of Middle to Upper Ordovician marine limestones (after Qing and Veizer, 1994).
the Southern Block exhibit \(^{18}\)O depletions that range from 4.8 \(0\%/00\) to 5.4 \(0\%/00\), the exception is sample # 180 which exhibits an \(^{18}\)O depletion of 9.3 \(0\%/00\). Calcite veins associated with D3 normal faulting from Hills C, D, and E exhibit \(^{18}\)O depletions that range from 8.8 \(0\%/00\) to 13.7 \(0\%/00\). 4) There is a depletion of \(^{13}\)C in the calcite veins with respect to the wall rock except sample #37. The range of \(^{13}\)C depletion in veins with respect to wall rock is 0.9 - 1.9 \(0\%/00\).

**Table 5.1** \(^{18}\)O and \(^{13}\)C values from calcite veins and their corresponding wall rocks.

<table>
<thead>
<tr>
<th>Sample #</th>
<th>(^{18})O</th>
<th>(^{13})C</th>
<th>Vein type</th>
</tr>
</thead>
<tbody>
<tr>
<td>WR = wall rock</td>
<td>18O (0%/00)</td>
<td>13C (0%/00)</td>
<td>Stockwork = S</td>
</tr>
<tr>
<td>V = vein</td>
<td></td>
<td></td>
<td>Planar = P</td>
</tr>
<tr>
<td>U. Th = Upper Thumb Mountain Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cliff = Cliff member Cape Phillips Formation</td>
<td>137 WR (U. Th Hill E)</td>
<td>20.0</td>
<td>-0.7</td>
</tr>
<tr>
<td>137 V (U. Th Hill E)</td>
<td>10.3</td>
<td>-2.0</td>
<td></td>
</tr>
<tr>
<td>82 WR (U. Th Hill D)</td>
<td>21.9</td>
<td>-0.4</td>
<td></td>
</tr>
<tr>
<td>82 V (U. Th Hill D)</td>
<td>13.1</td>
<td>-2.3</td>
<td></td>
</tr>
<tr>
<td>85 WR (L. Th Hill C)</td>
<td>24.0</td>
<td>-2.0</td>
<td></td>
</tr>
<tr>
<td>85 V (L. Th Hill C)</td>
<td>10.3</td>
<td>-3.6</td>
<td></td>
</tr>
<tr>
<td>37 WR (Cliff Hill A)</td>
<td>22.4</td>
<td>-1.1</td>
<td></td>
</tr>
<tr>
<td>37 V (Cliff Hill A)</td>
<td>25.0</td>
<td>-0.5</td>
<td></td>
</tr>
<tr>
<td>103 WR (U. Th Southern Block)</td>
<td>16.8</td>
<td>-2.8</td>
<td></td>
</tr>
<tr>
<td>103 V (U. Th Southern Block)</td>
<td>12.1</td>
<td>-4.3</td>
<td></td>
</tr>
<tr>
<td>180 WR (U. Th Southern Block)</td>
<td>19.9</td>
<td>-2.5</td>
<td></td>
</tr>
<tr>
<td>180 V (U. Th Southern Block)</td>
<td>10.5</td>
<td>-3.4</td>
<td></td>
</tr>
<tr>
<td>97 WR (U. Th Southern Block)</td>
<td>18.4</td>
<td>-2.5</td>
<td></td>
</tr>
<tr>
<td>97 V (U. Th Southern Block)</td>
<td>13.0</td>
<td>-4.3</td>
<td></td>
</tr>
<tr>
<td>39 WR (U. Th Southern Block)</td>
<td>17.4</td>
<td>-2.9</td>
<td></td>
</tr>
<tr>
<td>39 V (U. Th Southern Block)</td>
<td>12.0</td>
<td>-4.4</td>
<td></td>
</tr>
</tbody>
</table>

Sulphur isotope results collected from the barite vein within the southern block yielded a \(\delta^{34}\)S value of 36.6 \(0\%/00\).
5.4 Discussion and Interpretation

5.4.1 Discussion of $\delta^{18}O$ values within the study area

The following discussion will focus on $\delta^{18}O$ values for wall rocks, and then a comparison between the $\delta^{18}O$ values of veins with respect to the wall rocks.

Wall rocks

Wall rocks can be subdivided into two groups based on their $\delta^{18}O$ values: 1) Samples #82, #85, #37, and #137 exhibit $\delta^{18}O$ values that fall largely within the domain of calcite that has precipitated in equilibrium with Middle to Upper Ordovician seawater (Fig 5.3). Sample #137 is slightly depleted in $^{18}O$ in comparison to the other samples in this group. 2) Samples #103, #180, #97, and #39 exhibit $^{18}O$ values that are depleted with respect to calcite that has precipitated in equilibrium with Middle to Upper Ordovician seawater (Fig. 5.3). These results indicate that a whole rock $^{18}O$ depletion occurred within the Southern Block.

Veins

The trend of $^{18}O$ depletion in the calcite veins with respect to the wall rocks is systematic across all the samples, except sample #37, which is slightly enriched in $^{18}O$. The relative amount of $^{18}O$ depletion changes in veins associated with D2 normal faults from the Southern Block and veins associated with D3 normal faults from Hills C, D, and E. Although the data set is small in number the results are interpreted to represent either a different isotopic signature for the fluids that precipitated the calcite veins associated
Mechanisms for $^{18}$O depletion

Anomalously low $\delta^{18}$O values in wall rocks from the Southern Block and the systematic depletions in vein calcite relative to the wall rock suggest several possible mechanisms for isotopic depletion. The two most plausible mechanisms are: 1) exchange with an exotically derived, low-$\delta^{18}$O fluid, and 2) localized, fluid assisted isotopic exchange with coexisting, lower $\delta^{18}$O phases such as quartz or muscovite. The following discussion will cover the merits of each of the proposed mechanisms for isotopic depletion with respect to the region of study.

Exotically derived low $\delta^{18}$O fluids

The infiltration of surface derived meteoric fluids into extensional fault zones is a likely possibility to explain the systematic $^{18}$O depletion in calcite veins and wall rock from the Southern Block. The extensive fracture network surrounding normal fault zones would provide the permeability network required to focus the flow of meteoric fluids down-section. In addition, the results of this study indicate that $^{18}$O depletion in calcite veins is a localized phenomenon that is spatially associated with the normal fault zones. This interpretation is supported by the fact that calcite veins from sample #37, the only sample not associated with normal faulting, did not exhibit $^{18}$O depletion.

Fluid assisted isotopic exchange with coexisting, lower $\delta^{18}$O phases such as quartz or muscovite

An obvious local reservoir for the low $^{18}$O fluid in the Stanley Head region is the shales of the Cape Phillips Formation. If intra-formational fluids in isotopic equilibrium
with the Cape Phillips Formation shales migrated into limestone formations of the
Stanley Head anticline, the calcite veins in these formations would exhibit depleted $\delta^{18}O$
values with respect to the limestone wall rock. This scenario, although possible, does not
take into account that normal faults, during the D2 and D3 events, would have propagated
through the Cape Phillips Formation and therefore presumably established channels
through which meteoric water could be focused downward. In addition, the low
permeability of shales makes it unlikely that large volumes of fluid flowed from the Cape
Phillips Formation into the limestone formations of the Stanley Head anticline.

5.4.2 Interpretation of $^{18}O$ depletion within wall rocks and calcite veins

Meteoric waters are proposed to be responsible for the observed depleted $\delta^{18}O$
values observed in the wall rocks from the Southern Block and the calcite veins from the
region of study. Two questions remain to be addressed: 1) Why is there a difference in
wall rock isotopic signatures from the Southern Block and the rest of the Stanley Head
anticline, and 2) Why are D3 related veins more depleted than D2 related veins?

Wall rock depletion in the Southern Block

The Southern Block exhibits the most intense fracturing and veining of all rocks
within the region of study. This is interpreted to reflect fractures produced during the D1
event due to active hinge folding, and fractures produced during D2 normal faulting. The
extensive fracture network would permit the pervasive infiltration and flow of meteoric
fluids throughout the structural domain, resulting in the whole-scale isotopic depletion of
the wall rock.

Hill E (sample #137) has a similar fracture network to the one in the Southern
Block, although not as well developed. This fracture network is also interpreted to reflect
fractures produced during both the D1 event due to active hinge folding and D3 normal faulting. Sample # 137 exhibits a wall rock $\delta^{18}$O value of 20 $^\circ/o$, which is 2 $^\circ/o$ less than other wall rocks in the Stanley Head anticline that do not possess as well a developed fracture network.

Vein Depletion

The difference in $^{18}$O depletion in D2 and D3 related calcite veins and the difference in $\delta^{18}$O values in whole rocks from the Southern Block and the rest of the Stanley Head anticline may have been influenced by the geographic position of the study area during the D2 and D3 events. The D2 and D3 events are interpreted to be part of the Late Devonian to Early Carboniferous Ellesmerian orogeny and Late Cretaceous to Early Tertiary aged Eurekan deformational event, respectively. In the early Devonian the present day Arctic region of Canada was in the equatorial region of the globe (Fig. 5.4a), in the earliest Tertiary this region had migrated north, to a latitude of approximately 60-70$^\circ$ N (Fig. 5.4b).

The isotopic variations of meteoric waters on earth are extremely systematic and can be closely approximated by the equation which defines the meteoric water line (Craig, 1961):

$$\delta D = \delta^{18}O + 10 \text{ (in per mil).}$$

Modeling of atmospheric precipitation, as a Rayleigh fractionation process, gives an excellent description of the observed H and O isotopic compositions of meteoric waters on a global scale (Sheppard, 1986). Water that condenses from atmospheric vapour in the air mass is richer in $^{18}$O and D than the vapour, resulting in precipitation that is
Figure 5.4
Paleogeographic reconstruction of North America for a) Late Devonian, and b) Late Cretaceous ages. Red ovals outline approximate positions of the present day regions of the Canadian Arctic Islands. (After Blakey, 2003)
continuously depleted in $^{18}$O and D. Therefore, the most $^{18}$O and D depleted meteoric waters are found at high latitudes and elevations.

For ancient meteoric waters to follow the same relationship, the O and H isotopic compositions of the ancient ocean waters must be comparable to present day values. Taylor (1997) states that the $\delta^{18}$O values of ocean water is mainly controlled by a steady state balancing act between low temperature processes such as weathering, diagenesis, and biogeneic or authigenic mineral formation that tend to enrich rocks in $^{18}$O (and by material balance deplete the hydrosphere in $^{18}$O) and high temperature processes such as metamorphism and hydrothermal activity that tend to deplete the rocks in $^{18}$O (thus enriching the hydrosphere in $^{18}$O). Gregory and Taylor (1981) and Gregory (1991) have concluded that due to the enormous amount of water rock interaction taking place at mid-ocean spreading ridges the ancient ocean waters would maintain a $\delta^{18}$O value of near zero, assuming that seafloor spreading occurred at a rate comparable to the present day. Although there is still some controversy regarding the isotopic composition of ancient ocean waters, Taylor (1997) proposes that the isotopic composition of the ocean has remained stable throughout geologic time.

In the Devonian, the arctic regions were in the equatorial regions of the earth and meteoric waters from this time would not have been as depleted in $^{18}$O as the corresponding meteoric waters would have been in the Tertiary when the arctic regions were at a latitude of 60-70 degrees. It is therefore reasonable to expect a greater amount of $^{18}$O depletion in both wall rocks and calcite veins associated with D2 Ellesmerian aged normal faults than with D3 Eurekan aged normal faults.
The observed depletion of the wall rocks in the Southern Block and the relative difference in $^{18}$O depletion between calcite veins associated with D2 and D3 normal faults can be explained as an isotopic system that was rock buffered. The interpretation that the fold mechanism responsible for the creation of a fracture network is linked to the whole scale depletion of the wall rock is supported by data observed from Hill E, sample #137. An extensive fracture network similar to the one observed in the Southern Block, although not as well developed, characterizes Hill E. This fracture network is interpreted to reflect fractures produced during the D1 event due to active hinge folding, and fractures produced during D3 normal faulting. Sample #137 exhibits a wall rock $\delta^{18}$O value of 20.0 $^{0}_{/00}$, a value depleted by approximately 2 $^{0}_{/00}$ with respect to calcite that has precipitated in equilibrium with Middle to Upper Ordovician seawater.

5.4.3 $\delta^{13}$C values

There is a systematic relationship in which $^{13}$C is slightly depleted in the veins with respect to the wall rocks. This trend is observed in all the samples that are located in close proximity to normal faults, the only exception is sample #37 which is slightly enriched in $^{13}$C. $^{13}$C depletion in veins with respect to wall rock ranges from 0.9 $^{0}_{/00}$ in sample #180 to 1.9 $^{0}_{/00}$ in sample #82. The source of carbon in the calcite veins is almost certainly from the dissolution of the wall rocks given the narrow range of $^{13}$C values observed. The trend of $^{13}$C depletion in the calcite veins may reflect a slight fractionation that occurred during the dissolution of limestone wall rocks.

5.4.4 $\delta^{34}$S value

Barite is a common constituent of some Mississippi Valley Zn-Pb districts (Anderson & Macqueen, 1988). Barite is found at the Polaris mine site, the Truro Island
showing and at the Southern Block of the Stanley Head anticline. The $\delta^{34}S$ value from Stanley Head is 36.6 $\%$ and is comparable to a $\delta^{34}S$ values of 24.0 $\%$ and 38.0 $\%$ from barite found at Polaris and Truro Island respectively. $\delta^{34}S$ values for evaporites, sulphides, and sulphates from the Polaris district are presented in figure 5.5.

5.5 Comparison of O and C isotopic signatures from the Stanley Head region to the Polaris mine site.

To compare oxygen and carbon isotope signatures from the calcite veins and wall rock found at the Stanley Head region to similar data from the Polaris mine, it is necessary to give a brief paragenesis of the various carbonate phases found at the mine site with respect to the mineralizing event.

Unaltered limestone, diagenetic calcites, and pore and vein filling calcite, and five types of dolomite are recognized and discerned based on microscopic determinations of colour, crystal size, crystal size distribution, crystal shape, inclusion content, Fe content, and cathodoluminescence (Savard et al., 2000). Unaltered limestone is sedimentary micrite (SC). Diagenetic calcites include fine blocky crystals in dissolution voids of fossils (FC), and neomorphic pseudospar replacing sedimentary and early diagenetic calcites (RC). Pore and vein filling calcite (PC) is coarse and blocky. The paragenesis of these phases with respect to mineralization is given in figure 5.6.

SC is the unaltered limestone at the Polaris mine and exhibits $^{18}O$ and $^{13}C$ values that are consistent with calcite that precipitated in equilibrium with Middle to Upper Ordovician seawater (Fig. 5.7). It is interpreted that the $\delta^{18}O$ values of the unaltered limestones (SC) at Polaris are comparable to the $\delta^{18}O$ values of the limestone wall rocks
Figure 5.5
Sulphur isotope values from sulfides and sulphates in the Polaris district. The $\delta^{34}\text{S}$ value from the barite veins at Stanley Head is given as a red rectangle (Randall and Anderson, 1996).

Figure 5.6
General paragenetic sequence of the various minerals and hydrothermal phases in the Polaris deposit. Grey boxes indicate calcite phases, orange boxes indicate dolomite phases, and black boxes indicate sulphide phases.
at Stanley Head. Vein and pore filling (PC) postdates all the sulphides and dolomites found at the Polaris mine and exhibits a wide range of both $^{18}$O and $^{13}$C values (Fig. 5.7). Both D2 and D3 calcite veins found in the Stanley Head region are interpreted to postdate the mineralizing event at Polaris mine and are therefore comparable to the PC calcite.

The unaltered limestones at the Polaris mine site and the wall rocks of the Stanley Head region have isotopic signatures that overlap one another. The calcite veins from both areas are systematically depleted in $^{18}$O, the cause of which is inferred to be the infiltration of meteoric waters, and therefore are potentially related as part of a regional fluid system. There is considerable spread in the both $\delta^{18}$O and $\delta^{13}$C values of the PC calcite found at Polaris. It is postulated that the spread in $\delta^{18}$O values may be related to calcite precipitating in fractures during different deformational events, and/or different amounts of fluid rock interaction. Severely depleted $\delta^{13}$C values may reflect fluids derived from Tertiary aged volcanics found on eastern Bathurst Island.

5.6 Conclusion

The purpose of this study was to place constraints on the origin of the fluids that precipitated calcite veins within the study area. This preliminary study yielded the following results: 1) The systematic depletion in $^{18}$O in the calcite veins, with respect to the corresponding wall rock, indicates that meteoric waters were focused spatially into extensional fault zones. 2) Wall rock $^{18}$O depletion within the Southern Block, and to a lesser extent Hill E, is likely a function of pervasive fluid flow through fracture networks that were produced during active hinge folding during the D1 event. 3) The degree of
Figure 5.7 $\delta^{13}$C - $\delta^{18}$O diagram showing the isotopic compositions of calcite veins and wall rocks from the Stanley Head region and the Polaris mine site. Polaris mine isotopic data from Savard et al., 2000.
$^{18}$O depletion in calcite veins associated with D2 normal faulting is not as great as $^{18}$O depletion in calcite veins associated with D3 normal faulting. This trend is interpreted to reflect two possibly scenarios. The first scenario is a result of pervasive fluid flow through the well developed fracture network of the Southern Block, in which extensive fluid-rock interactions would have left the D2 meteoric fluids slightly buffered by the wall rock. The second scenario is that the relative geographic position of the present day Arctic regions of Canada during the D2 and D3 events influenced the amount of depletion within the veins, whereby Devonian meteoric waters are not as depleted in $^{18}$O as Tertiary meteoric waters. 3) The carbon source for calcite veins in the Stanley Head region is likely from the wall rock. 4) The sulphur isotope signature of the barite from the Southern Block is comparable to other barite found in the district.
CHAPTER 6

CONCLUSIONS

Mapping undertaken for this thesis has significantly changed the structural interpretation of the Stanley Head region of western Cornwallis Island. The geology of the map area is dominated by previously unrecognized shallow level folds and thrust faults. These structures exhibit some of the characteristic elements of "evaporite detachment thin-skinned contraction deformation"(Davis and Engelder, 1985), and are collectively interpreted to represent an evaporitic based fold and thrust system in which there are multiple levels of detachment.

Three phases of deformation have been documented in the study area: 1) northeast and southwest verging folds and thrust faults associated with an east-west compressional D1 event, 2) strike slip faulting and folding associated with a north-south compressional D2 event. This D2 folding resulting in localized refolding of some northeast and southwest verging folds, and 3) normal faulting associated with an east-west extensional D3 event. The D1, D2, and D3 events are interpreted to correspond with the Boothia uplift, Ellesmerian Orogeny, and Eurekan deformational event respectively.

"Balanced" cross sections across the area of study were produced using the integration of surficial data and a kinematic model of deformation for the multiple detachment, evaporite based Parry Islands fold belt (Harrison & Bally, 1988). These cross sections were constructed such that they obeyed the characteristics of a multiple detachment, evaporite based fold and thrust belt. This was manifested in the cross sections as: 1) A pronounced structural disharmony between the surficial expression of the Stanley Head anticline, and an underlying fold and thrust system. 2) A horizontal
orientation for the dip of the basal decollement. 3) The presence of compressional evaporitic welts within the lower Bay Fiord and Baumann Fiord formations and associated synclinal thinning of the ductile layers. 4) Conjugate thrusts displacing the lower fold within the Eleanor River Formation.

Balancing of the deformed state cross sections was achieved through bed length, area, and volume balancing techniques. The deformed and restored state cross sections demonstrate that the proposed structural style of the Stanley Head fold and thrust system is a structurally viable and admissible interpretation.

The structural geometry and kinematics of deformation portrayed within the Stanley Head cross sections provides an excellent analogue to characterize the structural geometry of the Cornwallis fold belt. The regional cross section displays structures representing a range of strain states, and various structural styles that characterize the structures of the Cornwallis fold belt. Overall structures of the Cornwallis fold belt are thought to be characterized by: 1) Southwest and northeast verging folds and thrust faults, 2) a variety of fold - thrust fault interactions, including detachment style folding and thrust imbricate systems, and 3) a proposed structural disharmony between the surficial fold belt and an underlying fold and thrust system.

The stable isotopes study was conducted to delineate the origin of the paleofluids responsible for the precipitation of calcite in extensive fracture networks within the study area. To address the origins of the fluids in the study area oxygen and carbon isotopic data was obtained from calcite veins and their corresponding wall rocks. This preliminary study yielded the following results: 1) The systematic depletion in $^{18}$O in the calcite veins, with respect to the corresponding wall rock, indicates that meteoric waters
were focused spatially into extensional fault zones. 2) Wall rock $^{18}$O depletion within the Southern Block, and to a lesser extent Hill E, is likely a function of pervasive fluid flow through fracture networks that were produced during active hinge folding during the D1 event. 3) The degree of $^{18}$O depletion in calcite veins associated with D2 normal faulting is not as great as $^{18}$O depletion in calcite veins associated with D3 normal faulting.

The results of this study have important implications with respect to the development of a regional metallogenic model for the Polaris Zn-Pb-Cu district. In particular, this study provides insight into the structural controls that focused fluid flow into the Polaris ore deposit. The Stanley Head anticline is well exposed and interpreted to extend across Pullen straight to re-emerge on Little Cornwallis Island, at the Polaris mine site. The structural style and geometry of the Stanley Head anticline when applied to the Polaris mine site gives valuable insight into what the plumbing network and geometry of the host rocks may have been prior to mineralization.

The Stanley Head anticline is interpreted to be a kinematically linked to the Boothia uplift. This means that the anticline could have acted as a structural trap for the mineralizing fluids of the Polaris ore deposit. The fault related folding mechanisms of the Stanley Head anticline have been interpreted to be divided into two categories: 1) Fixed hinge folding, and 2) Active hinge folding. In both cases fracture networks will be developed as folding progresses, providing the necessary plumbing system to allow mineralizing fluids to migrate into the pre-existing anticlinal structural trap.
REFERENCES:


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Representative cross sections of the Stanley Head region, Western Cornwallis Island, Nunavut (Sheet A)

Discussion of cross section B-B'

- Mobility of evaporites: It is observed that the evaporites of the OCB-I migrate into the axial region of the surficial anticline of the deforming rock panel is on the order of a few degrees. For the purpose of constructing the cross-sections a horizontal unit moved out of the line of section during the formation of the Stanley Head anticline to form the large compressive strain, the horizontal shortening of the unit is the same as expressed as the bed lengths of the OF. OCB-u. OCT-1. OCT-u. OCI, and gained into or out of the line of section when area balancing the ductile units. The OB unit is balanced with the assumption that there are no significant changes in slope.

- This cross section removed the structural complexity associated with normal faulting, giving a view of the Stanley Head anticline without the influence of faulting. The structural geometry of the Stanley Head anticline is characterized by a detachment style of folding above the evaporites, with the surficial anticline displaying a asymmetric geometry.

- The D3 cross-sections B-B' and F-F' represent true deformed state cross-sections. In an attempt to characterize the diversity of structures that characterize the map area, cross sections have been constructed from the integration of surface geology and the application of a kinematic model of deformation for the (seismically imaged) multiple detachment evaporite based Parry Islands fold belt on Melville Island. The application of this model to the map area is appropriate for the following reason: The deformation for the (seismically imaged) multiple detachment evaporite based Parry Islands fold belt on Melville Island is characterized by a detachment style of folding of the surficial anticline. Hills E, K and the Southern block are characterized by thrust panels that display only the backlimb of a proposed anticline.

- The change of vergence, structural style, and fold geometry along strike on the Stanley Head anticline is not unique only to this region, but rather is a reflection of the complex tectonic history of the Parry Islands fold belt. The conjugate thrust faults are kinematically linked to the graben bounding faults as they are directly along strike and almost continuous from the graben to the fold belt.

- The Coulomb's fracture criterion will have nearly equal dips and therefore an equal chance of forming.

References


NOTES TO ACCOMPANY SECTION B-B' AND F-F'

- Pre-extension section length: 12.56 km
- Bed length of OSCP between pin lines #2 and #3 (the wavelength of the Stanley Head anticline) - 11.85 km
- Shortening of OSCP 13445 - 12560 - 884 m (7.0%)