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MECHANICS OF THRUST FAULT FORMATION:
DEVELOPMENT OF THE ROCKY MOUNTAIN FORELAND
BELT

by

© Robin Carl Mann, B.Sc.

A thesis submitted to the Faculty of
Graduate Studies in partial fulfilment
of the requirements for the degree of
Masters of Science

Carleton University

Ottawa, Ontario,

May, 1979

The undersigned hereby recommend to the Faculty of
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ABSTRACT

The Rocky Mountain Thrust and Fold Belt is a typical thin-skinned foreland belt. To determine the mechanics involved in the deformation of the Belt, a mechanical model is developed based on plastic rheology, and treating the Belt as homogeneous and isotropic. This model accounts for horizontal compressive forces causing supracrustal shortening and for gravity-induced spreading forces resulting from the build-up of a surface slope.

The results obtained show a close similarity between theoretical thrust faults (slip-lines) and the actual faults, and between the computed velocity vectors and the observed movement of the material in the Belt. The proposed model can be operative only if the material strength of the Belt is not much larger than 100 bars. This value is in accord with values obtained from the mechanical analysis of other foreland belts by means of different methods.

Along individual thrust planes, deformation probably occurs due to pressure-solution slip or frictional sliding unless a concentrated shear zone is present. In the latter case deformation is probably due to pressure solution or superplasticity, which can achieve high strain rates under supracrustal conditions.

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TABLE OF CONTENTS

ABSTRACT	111
ACKNOWLEDGEMENTS	iv
TABLE OF CONTENTS	v
LIST OF FIGURES	vii
LIST OF TABLES	ix
LIST OF SYMBOLS	x
 Chapter 1	
Statement of Problem	
1.1 Previous Work and Purpose of This Thesis	1
1.2 Outline of Contents	3
 Chapter 2	
Geological Evolution of the Canadian Cordillera	
2.1 The Canadian Cordillera: Geology	6
2.2 The Canadian Cordillera: Tectonic Evolution	14
2.3 The Canadian Cordillera: Geophysics and Crustal Structure	19
2.4 The Southern Rocky Mountains: Geology	23
2.5 The Southern Rocky Mountains: Structure	27
2.6 The Southern Rocky Mountains: Tectonic Evolution	30
 Chapter 3	
Theoretical Models for Thrust Fault Formation	
3.1 Fundamental Definitions and Equations of Elasticity Theory	36
3.2 Applications of Elasticity Theory to Thrust Fault Formation	40
3.3 Fundamentals of Plasticity Theory	44
The Rigid Plastic Body	44
Plane Plastic Strain	45
Velocity Fields and Stress and Velocity Discontinuities	47
Compression of a Block Between Rough Plates	52
3.4 Applications of Plasticity Theory to Thrust Fault Formation	55
Glacier Flow Theory	55
Flow Theory Related to the Foreland Belt	59
 Chapter 4	
A Model for the Mechanics of Thin-Skinned Thrusting in the Rocky Mountain Foreland Belt	
4.1 The Compressive-Spreading Wedge Model	65
4.2 Interpretation of the Model	73
4.3 Calculation Methods	77
4.4 Application of the 'Compressive-Spreading Wedge' Model	83
 Chapter 5	
The Upper Limit of Stresses and Deformation Mechanisms Along Thrust Faults	
5.1 General Remarks	100
5.2 Upper Limit on Stress Along Thrusts	101
5.3 Deformation Mechanisms Along Thrusts	102
 Chapter 6	
Conclusions	111
 References	113

Appendix I	List of C's, α 's and ΔH 's for Values of χ	118
Appendix II	Computer Program Used in the Calculation of the 'Compressive-Spreading Wedge' Model	122

LIST OF FIGURES

Figure 2:1	Physiographic Belts of the Canadian Cordillera.	7
Figure 2:2	Lithological Assemblages in the Southern Canadian Cordillera.	8
Figure 2:3	Times of Emplacement of Allochthonous Belts.	17
Figure 2:4	Gravity Models for the Cordillera.	20
Figure 2:5	Crustal Structure Beneath the Rocky Mountains.	21
Figure 2:6	Stratigraphy of the Rocky Mountains.	24
Figure 2:7	Cross-section of the Southern Rocky Mountains.	28
Figure 2:8	Price and Mountjoy Model.	31
Figure 2:9	Campbell Model.	33
Figure 3:1	Stress Elements Used in Elasticity.	38
Figure 3:2	Hafner's Deformed Block.	43
Figure 3:3	Slip Lines.	48
Figure 3:4	Velocity Fields in Plasticity.	50
Figure 3:5	Discontinuities in Plasticity.	51
Figure 3:6	Prandtl's Solution.	53
Figure 3:7	Glacier Model.	56
Figure 3:8	Extending and Compressive Flow.	58
Figure 3:9	Elliott's Model.	61
Figure 3:10	Ramberg's Gravity Collapse.	63
Figure 4:1	Evolution of the Southeastern Canadian Cordillera.	66
Figure 4:2	Chapple's Wedge and Stress Element.	68
Figure 4:3	Flow Through a Channel.	70
Figure 4:4	Chapple's Wedge Showing Thrust Element.	76
Figure 4:5	Wedges Used in the Compressive-Spreading Wedge Model.	79
Figure 4:6	Isostatic Adjustments of the Crust Beneath the Rocky Mountains.	81
Figure 4:7	Predeformational Data and Cross-section.	86

Figure 4:8	Half-time Deformational Data and Cross-section.	86
Figure 4:9	Postdeformational Data and Cross-section.	87
Figure 4:10	Orientation of Thrust Faults in Predeformational Wedge.	93
Figure 4:11	Orientation of Thrust Faults in Half-time Deformational Wedge.	94
Figure 4:12	Orientation of Thrust Faults in Postdeformational Wedge.	95
Figure 4:13	Velocity Vectors for Predeformational Wedge.	96
Figure 4:14	Velocity Vectors for Half-time Deformational Wedge.	97
Figure 4:15	Velocity Vectors for Postdeformational Wedge.	98
Figure 5:1	Plastic Flow of Yule Marble and Solenhofen Limestone.	105
Figure 5:2	Flow Regime Map at Constant Temperature.	107
Figure 5:3	Deformation Mechanism Maps for Quartz and Calcite.	109

LIST OF TABLES

Table 4:1	Results of Calculations for Predeformational Case.	88
Table 4:2	Results for Half-time Case.	89
Table 4:3	Results for Postdeformational Case.	90

LIST OF SYMBOLS

A, C	Constants of integration in 'compressive-spreading wedge' model
C_0	Concentration of saturated solution in equilibrium with unstressed solid
B	Rate of tectonic shortening
a, b, c	Constants in elastic solutions, $a = \rho g$
d	Average grain size
D_p	Grain boundary diffusivity
$f^{(4)}$	Fourth derivative of function f
g	Acceleration of gravity
H, h	Height of body, or thickness
h_0	Average thickness of a segment within the modelled thrust belt
K, κ	Yield strength of material, yield stress
L	Line of discontinuity
L_t	Length of thrust belt
L_1	Thickness of thrust belt at eastern edge
L_2	Thickness of thrust belt at western edge
p	Pressure
s	Speed of material
t	Time
T	Absolute temperature
u_i	Displacement vector
V	Potential
v_i	Velocity vector
w	Effective grain boundary width
x_i, s_i	Cartesian coordinate systems
r, θ	Polar coordinates
α -line, β -line	Slip lines
α	Surface slope
γ	Basal and surface slopes of a body when they are parallel
δ_{ij}	Kronecker delta
ΔH	Increase in elevation from horizontal datum
$\epsilon_{ij}, \dot{\epsilon}_{ij}$	Infinitesimal strain tensor, strain rate
ξ, η	Constants of slip line theory
θ	Clockwise angle between x axis and the trajectory of the algebraically largest principal stress
θ_0	Basal slope of wedge
Θ	$3k\theta$, where k is the bulk modulus and $\theta = \epsilon_{11}$
λ, μ	Lamé constants
ν	Poisson's ratio
V	Molar volume of a solid
ρ	Density
σ_i	Principal stress
$\sigma_{12}, \sigma_{12max}$	Shear stress, maximum shear stress
σ_{ij}	Stress
σ_y	Yield stress in the Hall-Petch relation
σ_0, κ	Constants in Hall-Petch relation
Φ	Airy stress function
X	Yield stress ratio
X_1	Body force per unit mass
ψ	Angle between maximum compressive stress axis and a radius vector

CHAPTER 1

STATEMENT OF THE PROBLEM

1.1 Previous Work and Purpose of This Thesis

The tectonic evolution of the southern Canadian Rocky Mountains within the framework of the entire Canadian Cordillera has received much attention in recent years. The Rocky Mountain Belt is a thin-skinned foreland fold and thrust belt which is defined by its typical unmetamorphosed sedimentary sequences of miogeoclinal type. These rocks, deformed (dominantly by folding and imbricate thrusting) at a shallow level in the crust, are examples of what is referred to as the suprastructure of mountain belts, in contrast to the once-deeper, more metamorphosed, and differently deformed rocks of the interior parts, which compose the infrastructure (Hobbs et al., 1976).

Models have been proposed by various authors, using geological and geophysical evidence in the Rocky Mountains and the Omineca Crystalline Belt, to account for the observed large-scale structure. However, the persistence of differences of opinion on the evolution of the Rocky Mountains (compare for instance Price and Mountjoy, 1970, Campbell, 1973, and Elliott, 1976a) shows that a satisfactory tectonic model, reconciling field evidence with a consistent mechanical analysis, has not yet been found. The main focus of interest here is in the evolution of the Foreland Belt as a whole, and not deformation along single thrusts.

Models for the tectonic evolution of foreland belts have been proposed using both elasticity and plasticity theory. Classical models have treated deformation in a thrust belt on the basis of elastic rheology; however, from the simple fact that rocks yield and show permanent deformation, it is clear that a plastic rheology is more realistic, although of course it does not describe accurately the actual creep behaviour of the rocks involved.

Therefore a treatment of deformation in a thrust belt should assume that deformation occurs mainly due to plastic flow.

There are three major hypotheses as to how thrusts in a foreland belt can form. They may be termed: 1) Gravity gliding; 2) Gravity spreading; and 3) Push from behind.

In the first mechanism (gravity gliding) it is suggested that thrust sheets move entirely under the influence of gravity, and that at the time of movement the basement (or any thrust plane) dipped in the direction of thrust motion. In many cases, however, including the Rocky Mountain Foreland Belt, both the basement and individual thrust planes dipped in a direction opposite to the motion of the thrust sheets; therefore gravity gliding does not appear to be a generally viable mechanism.

The second proposal (gravity spreading) also suggests that thrust sheets move due to gravity; but in this case the driving force for movement is the difference in surface elevation between the interior of the orogen and the craton, and not the slope of the basal surface. A problem with this proposal, however, is that no easily identifiable mechanism exists to produce an initial surface slope, at least in the Rocky Mountains.

Finally, the 'push from behind' model does not involve any gravitationally directed forces other than lithostatic load: thrust sheets are thought to move due to a horizontally directed compressional load acting on the rear edge of the thrust sheet. This mechanism, as a first approximation, does account for the large overall thin-skinned shortening that is observed in foreland belts such as the Rocky Mountains; however, a possible difficulty lies in the strength of the material affected and the magnitude of the stress required.

Starting from considerations similar to those reported above, which are also relevant to other orogens, a model was developed by Chapelle (1978), combining the 'gravity spreading' and the 'push from behind' mechanisms.

In this model (which in this thesis is termed the 'compressive-spreading wedge' model) a wedge-shaped foreland belt is assumed to be acted upon by a horizontally directed compressive load: the resulting overall shortening of the belt originates a surface slope which in turn causes the belt to deform by gravitational forces analogous to the 'spreading' type. However, the belt is in compression throughout. In the model as originally proposed by Chapple, it was suggested that the belt moves along a weak basal layer as is the case in parts of the Appalachians and Alps, and in the Jura Mountains. However, there is no evidence of the existence of such a weak layer in the Foreland Thrust Belt; consequently, any application of Chapple's model to the Rocky Mountains has to take this fact into account, and modifications have to be introduced where necessary.

This thesis is an extension of Chapple's model and an application of it to the tectonic evolution of the Rocky Mountains. Our purpose is to apply a reasonably realistic mechanical model of foreland belt deformation to the Rocky Mountains, taking into account the geology and structure of the belt and, as far as possible, the proposed hypotheses on its tectonic evolution.

1.2 Outline of Contents

This thesis consists of four parts, i.e. (1) an overview of the geology and tectonics of the entire southern Canadian Cordillera with emphasis on the geology and evolution of the southern Rocky Mountains; (2) a survey of relevant concepts and equations of basic elasticity and plasticity theory with applications of both to the tectonics of foreland belts; (3) the presentation, modification and application of Chapple's model to the thin-skinned deformation of a foreland belt without a weak basal layer, such as the southern Rocky Mountains; and (4) a brief investigation of deformational mechanisms along individual thrust faults.

The geology, structure and tectonic evolution of the Canadian Cordillera and more particularly of the southern Rocky Mountains is reviewed in order to elucidate the relevant geological boundary conditions which must be included in any mechanical model. The basic definitions and equations of elasticity theory and its application (due mainly to Hafner, 1951) to the formation of thrust faults in a rectangular block are briefly surveyed since they represent a first attempt at explaining the formation of thrust faults. Basic plasticity theory is considered next, since a plastic material represents a geologically more realistic approximation to the rheology of foreland belts. Applications of plasticity theory such as the development of slip lines in a medium compressed between two parallel plates are discussed and serve as a first-order plastic model for thrust fault formation. Also, gravitational flow of material under 'gravity spreading' conditions is examined, by discussing glacier flow theory and its possible applications to foreland thrust belt deformation.

Following this, in what is the kernel of this thesis, the mechanical model proposed by Chapple (1978) for thin-skinned thrust belts is rederived from first principles and modified to fit the boundary conditions determined from the geology of the Rocky Mountains. The resulting model is then applied to the formation of the Foreland Thrust and Fold Belt by computing various parameters at three times during the tectonic evolution of the Belt. Values determined in each case include the tangential and radial stress (polar coordinates were used), the tangential and radial velocity vectors, and the strength of the material in the wedge, which is related to the required surface slope. Also, lines of maximum shear stress (slip lines) are determined in order to check the degree to which the model reproduces the attitudes of actual thrust planes.

Finally, a comparison is made between studies carried out in the Himalayas and the Zagros Mountains by Bird (1978), criteria developed by Elliott (1976a),

and the conclusions reached on the basis of the compressive-spreading wedge model. This was done in order to determine how the average strength of supracrustal rocks which gives geologically realistic results for the Rocky Mountains compares with strengths obtained by different methods. Also, reducing the scale of observation, deformation mechanisms such as frictional sliding, pressure solution slip, pressure solution and superplasticity along individual thrust planes are reviewed on the basis of studies carried out in the Swiss Alps and experimental results on rock deformation.

CHAPTER 2

GEOLOGICAL EVOLUTION OF THE CANADIAN CORDILLERA

2.1 The Canadian Cordillera: Geology

The Canadian Cordillera (between latitudes 49°N and 60°N) consists of five physiographic and geologic belts (Figure 2:1). They are from east to west: 1) sedimentary strata within the Rocky Mountain Belt, 2) granitic and metamorphic rocks of the Omineca Crystalline Belt, 3) unmetamorphosed and low grade sedimentary and volcanic strata in the Intermontane Belt, 4) granitic rocks of the Coast Range Plutonic Complex, and 5) unmetamorphosed and low grade sedimentary and volcanic strata in the Insular Belt (Monger et al, 1972). These belts are generally regarded as having developed differently in time and space, with the Rocky Mountain and Omineca Crystalline Belts being the only two autochthonous belts (on a gross regional scale) with respect to the North American craton. The other three are generally thought to be allochthonous, with the present geographic configuration being attained in the Late Mesozoic as suggested in figure 2:2 (Monger et al, 1972).

The geology and tectonic evolution of the Cordillera is a very voluminous topic, with data being acquired ever since the time of McConnell (1887). For a comprehensive review of Cordilleran geology and evolution, the reader is referred to Bally et al (1966), Douglas (1970), Wheeler and Gabrielse (1972), Monger et al (1972), Monger (1977), and Monger and Price (1979). Here, we will discuss the rock types, paleoenvironments and tectonic history of the entire Cordillera, emphasizing significant points with respect to the formation of the Rocky Mountain Fold and Thrust Belt.

The rocks of the Cordillera can be described in terms of two distinct orogenic belts, the Pacific and the Columbian Orogens. The Pacific Orogen

Figure 2:1 - Physiographic belts of the Canadian Cordillera, showing the eastern and western exposed limits of Proterozoic sediment. Area of study is the southern 500 km of the Rocky Mountains (after Monger, 1972).

Figure 2:2 - Space and time distribution of lithological assemblages in the southern Canadian Cordillera. From this it can be seen that the only two belts autochthonous with respect to the North American Craton on a regional scale are the Omineca and the Rocky Mountain Belts. The arrows show direction of sediment transportation (after Monger *et al.*, 1972).

is an elongated zone where two massive volcanic piles, the western Insular Belt and the eastern Intermontane Belt, are separated by a plutonic core (the Coast Plutonic Complex). Here, high grade metamorphic rocks are preserved as roof pendants only. The deformation within this zone has migrated away from the core complex and occurred from Early Cretaceous to Middle Tertiary.

To the east of the Pacific Orogen, within the Columbian Orogen, plutonic and volcanic rocks play a subordinate role to sediments in the flanking zones. Continental slope:rise and shelf deposits are present within these zones with the slope:rise deposits being present in the Interior Plateau (western flank of the belt) and the shelf deposits predominating in the Foreland Belt (eastern flank of the belt). Deformation within the Columbian Orogen was earlier than in the Pacific Orogen, with the earliest stages starting in the Precambrian.

The first sediments that were deposited which would later form the Cordillera were the siliceous clastics and carbonates of the Belt-Purcell Supergroup. These deposits are believed to have been deposited, on a miogeoclinal basement (Figure 2:1), which has an age of approximately 1600 Ma (Burwash et al, 1964), in two units separated by minor volcanic activity, over a time span of approximately 600 Ma (sedimentation beginning about 1350 Ma and continuing until 750 Ma). The Purcell strata are regarded as representing a continental terrace wedge, with the mode of deposition as a simple infilling of a basin with deep water conditions in the early stages changing to shallow water conditions at a later time. An alternative hypothesis proposed by Stewart (1976) is that the Belt-Purcell assemblage is a remnant left behind on the North American continent when one or more areas containing Middle Proterozoic sediment were fragmented by the rifting of a new continental margin in the Late Proterozoic.

At approximately 750 Ma there was a period of folding, faulting and regional metamorphism which is referred to as the Kootenay Orogeny in south-east British Columbia and the Racklan Orogeny in the Mackenzie Mountains. This deformation was a mild uplift in the south, being recorded by a minor unconformity, whereas to the north the unconformity was more pronounced. The termination of the orogeny is marked by the presence of a second miogeoclinal platform on the western edge of the craton. On this second miogeocline the Windermere assemblage was deposited in two units separated by a thick sequence of basaltic and andesitic volcanics. The lower unit is an impure clastic sequence which is predominantly grits, sandstones, and conglomeratic mudstones, whereas the upper unit is largely a pelitic and carbonate deposit.

The bulk of the Windermere assemblage has been considered to represent a continental-terrace wedge which was deposited west of the Purcell sedimentary sequence. The Purcell sequence is thought to have acted as a pseudocraton during this time. The change from fine, mature Purcell sediments to coarse poorly sorted and poorly bedded basal Windermere sediments (Toby Formation) has been suggested by Reesor (1970) to be due to the fact that during this period of the Hadrynian (750-570 Ma) the source areas were of considerable relief which may have been caused by uplift or doming at the edge of the continent. Alternatively, Aalto (1971) proposed that the Toby Formation is diamictite because it is doubtful whether a high coastline and river system could provide the great variety of material present. The two hypotheses are not, however, mutually exclusive, but it should be noted that deposition of the Toby Formation appears to correspond to a world-wide glacial event.

Sedimentation continued on the westward prograding miogeoclinal platform with intermittent volcanic activity until the Middle Cambrian (540 Ma). During the Early Cambrian, a blanket of orthoquartzite covered the terrace wedge.

The maturity of these sediments relative to the underlying Windermere suggests a reduction of local relief. In strata of Middle Cambrian age, conformable massive carbonate units prevail, grading westward into thick shales and argillaceous limestones. During Late Cambrian time there was minor folding and tilting of strata in the north (north of 56°N), and a relatively shallow water platform remained to the south.

The Lower and Middle Ordovician rocks form a conformable sequence overlying the Cambrian to the south. The Upper Ordovician overlies all older strata unconformably, and is conformably overlain by the Silurian. In the north, deposition appears to have been continuous from the Late Cambrian into the Devonian. In Late Ordovician time thick carbonates were deposited while the craton was generally depressed, causing transgression of the sea. This transgression overlapped the truncated margins of the earlier Paleozoic rocks and the Precambrian Shield.

To the north, the Lower and Middle Ordovician strata are similar to those in the south, and Upper Ordovician and Silurian carbonates, shales, and minor pyroclastics form a conformable sequence. In the south, the earlier source areas for clastic carbonate sedimentation, the Purcell and Alberta Arches, were covered by Upper Ordovician carbonates. These in turn were covered by Lower Silurian graptolitic shales, siltstones, basic igneous flows and pyroclastics.

During the Early Devonian, the craton was generally high and actively being eroded. Sedimentary cover was stripped back from the Precambrian lowlands and a considerable topographic relief was established. Sedimentation during this time was mostly fluvial sands and related deposits. In the Late Devonian, extensive limestones blanketed the northern craton, with local barrier reef complexes developing. This sedimentation ended with the formation of an extensive unconformity as the area was uplifted during the Early

Mississippian.

Rocks of the Early Mississippian are mostly plutonic and related to orogeny. Even though this Early Mississippian event is not accurately dated it is thought that uplift was greatest in the north. During this time rocks such as basalt, ultramafics and chert which are believed to be characteristic of deep ocean basins were deposited upon the carbonates of the depressed continental margin. The simplest explanation for the origin of this assemblage has been proposed by Monger and Price (1979) to be an abrupt subsidence followed by deep water sedimentation. These sediments were eventually covered by mafic volcanics, with possible overthrusting occurring during the Early Mesozoic. It has been suggested that this downwarping of the basin occurred in a back-arc or marginal basin behind a volcanic arc; however, this is still not clear as this event has been identified only in southern British Columbia.

The only sediments deposited during this time of orogeny (Late Devonian to Middle Triassic) which can be linked to the North American craton are carbonate-dominated successions which grade westwards and northwards into shale. Westernmost exposures of this unit contain chert-pebble conglomerate and chert quartz-arenite which Monger and Price (1979) suggest is indicative of a westerly source terrane.

Sedimentation at the beginning of the Mesozoic in the eastern Cordillera was mainly fine-grained clastic rocks which were entirely derived from cratonic sources. To the west of this, the environment has been interpreted to be a series of magmatic arcs and associated subduction complexes. Davies et al (1978) have suggested that these Early Mesozoic arcs which formed offshore were separated by deep water troughs, into which clastics were deposited. Also, from work done by Tempelman-Kluit (in press, as described by Monger and Price, 1979), Triassic and Jurassic arcs have been suggested to have formed over west-dipping subduction zones. This would mean that the Whitehorse trough

(Northwestern British Columbia) was a forearc basin, and both the arc and the subduction complex overrode the western North American margin on great thrust faults during the Early Mesozoic. At the end of the Triassic volcanism briefly ceased and a thick carbonate unit was deposited in the Insular Belt.

The Jurassic system which in part lies conformably on top of the Triassic, was deposited during convergence of the Cordillera into some semblance of its present configuration. It has been suggested that profound changes between position and configuration of Early and Middle Jurassic arcs and subduction complexes and those of the Triassic were due to the accretion of the large exotic Stikine block to the western North American margin (Monger and Price, 1979). This accretion resulted in granitic intrusion and regional uplift of the western Cordillera and the formation of three successor basins.

Sediments deposited within the Cordillera during this time were mainly coarse clastic detritus being shed eastwards from the Rocky Mountain Belt, eastward and westward transported detritus from the Omineca Crystalline Belt (thus suggesting uplift), and detritus within the Bowser Basin (west-central British Columbia) of the Intermontane Belt. To the west the Plutonic Complex was emerging as granitic rocks were emplaced.

Sedimentation within the successor basins continued into the Cretaceous contemporaneous with the early stages of the Columbian Orogeny (Douglas, 1970). To the south, large granitic plutons were emplaced during the early stages of the Columbian Orogeny, and extensive metamorphism occurred throughout the Omineca Belt.

In the Late Cretaceous most of the Cordillera was land. East of the Columbian Orogen, most of the sedimentation to the south was alluvial and to the north, transgression of the sea caused extensive sedimentation as detritus poured into the forming Foreland Basin. With the western Cordillera, granite was locally emplaced, and extensive intermediate to acid volcanism occurred

in the Intermontane Belt, and to the west within the Insular Belt, accretion of oceanic rocks was represented by volcanism on Vancouver Island.

Sedimentation into negative tectonic elements continued in the eastern Cordillera during the latest Cretaceous and into the Eocene, while volcanics were extruded due to faulting and granitic intrusions in the Insular Belt and the Coast Plutonic Complex. During Middle Tertiary there was volcanic activity throughout large parts of the Interior Plateau, and plutonism in the core of the Cascade fold belt. In the Pliocene uplift of the Coast Mountains shed alluvial sediment into restricted fault-bounded basins. To a minor extent, volcanism continued until the Pleistocene.

The final episode of sedimentation within the Cordillera, which has continued until the present, resulted in deposition of fluvial deposits. It has been suggested by Wheeler and Gabrielse (1972) that the drainage system operative at present has been functioning since the Miocene.

2.2 The Canadian Cordillera: Tectonic Evolution

Numerous tectonic models have been proposed for the evolution of the Canadian Cordillera. The overall tectonic evolution given here is an overview of recent interpretations on the Canadian Cordillera with respect to plate tectonic concepts (see Price and Mountjoy, 1970; Campbell, 1973; Brown, 1978; and Monger and Price, 1979).

It is generally inferred from the presence of miogeoclinal deposits that the western margin of North America began as an Atlantic-type continental margin. If this is true there must have been rifting of an older Precambrian continental mass and the opening of a Proto-Pacific ocean basin. Therefore, it should be possible to locate the counterpart which separated from the North American craton and formed the opposite side of the ocean. There

is no documentation of this block ever existing; however, it has been proposed by Sears and Price (1978) that the Siberian platform is a likely candidate.

The Atlantic-type continental margin is thought to have persisted along the western edge of the craton from about 1500 Ma until approximately 750 Ma when a minor orogeny is thought to have occurred (see Section 2.1). It has been suggested by Monger et al (1972) that this episode of uplift and orogeny was the result of a short-lived subduction complex operating along the craton's western edge; however, the evidence for this is very inconclusive. From this time of orogeny until the Upper Devonian and Lower Mississippian (350-340 Ma), it seems reasonable to assume from the sedimentary record that the western edge of the North American craton was a steadily filling wedge with local areas of minor deformation. There have been various hypotheses suggesting that the Windermere represented the initial sedimentation of a new ocean (Monger et al, 1972); however, there is no evidence for a western sedimentary source prior to the Middle Devonian (380 Ma). If we therefore accept the initial rifting hypothesis, we cannot accept this latter suggestion, as there would have to be some record for oceanic closing between initial rifting and Windermere deposition which is not seen in the stratigraphic record. Thus, the earliest recognizable stage in the Cordillera's evolution is the development of the northeasterly tapering, westerly prograding miogeocline which occupies most of the core of the Rocky Mountain Belt, and underlies part of the Omineca Crystalline Belt in the south.

As stated previously (Section 2.1), the first significant contribution to the stratigraphic record from a westerly source was during the Late Devonian. Hence it has been suggested that a subduction complex was operative along the western limit of the craton with some complex system of island arcs and back arc basins being present. This analysis has been substantiated by Monger (1977) who grouped the volcanic strata of this time into six assemblages:

a) the eastern assemblage, which could represent a marginal basin, b) an arc subduction complex in southern British Columbia, c) the Cache Creek and related groups which are thought to be remnants of an ancient Pacific Ocean, d) the Stikine assemblage of northwestern British Columbia which is thought to also be an arc terrane which accreted during the Early Mesozoic, e) the Chilliwack Group which may correlate with the subduction complex of southern British Columbia but was later transported along strike slip faults and f) the Sicker-Skolai arc assemblage. Therefore from this we can conclude that from Late Devonian until Middle Triassic there existed a complex system of volcano-genic environments which converged on each other at the onset of the Columbian Orogeny.

During the Middle Jurassic, a profound change in subduction pattern is thought to be due to the accretion of the Stikine block. This also corresponds to the opening of the present North Atlantic Ocean (Wheeler, 1970); therefore, it could be envisaged that the relative westward drift of the North American craton closed the back arc basins which were present before the Middle Jurassic.

From Middle Jurassic to Mid-Cretaceous the Canadian Cordillera gained its present configuration. The five distinct belts converged on each other as a result of compression in the eastern Cordillera and accretion and magmatism related to subduction along the western Cordillera (Figure 2:3). During this time there was the onset of thrusting within the foreland belt, uplift of terranes marginal to the Omineca Crystalline Belt and magmatism along the Coast Plutonic Complex of the western Cordillera. It has also been postulated that during this time the Cache Creek-Anvil Groups were overthrust onto the craton. Monger and Price (1979) suggested that this event was recorded by the intrusion of an anatectic melt into the Omineca Crystalline Belt during the Early Cretaceous. This melt was the result of the cratonic margin being depressed during overthrusting.

Figure 2;3 - Geographic location of the five physiographic belts with time of emplacement in the Canadian Cordillera (after Monger and Price, 1979).

From recent work done by Monger (1977) and Monger and Price (1979) in the southern Cordillera and Tempelman-Kluit (in press, as described by Monger and Price, 1979) in the northern Cordillera, it can be shown that the major strike slip faults within the Cordillera reflect an anticlockwise rotation with the Kula Plate (Grow and Atwater, 1970) in the south and the Pacific Plate in the north. From this it can be suggested that the allochthonous belts did not just 'drift' in from the west, but were obliquely moved along a system of strike slip faults into their present position. It has been demonstrated to be the case for Vancouver Island by Irving and Yule (1972), who concluded from paleomagnetic data that during the Triassic, Vancouver Island was at latitude 35°S while the presently adjacent part of the Cordillera was at latitude 27°N . The latitude gap gradually closed until the Late Mesozoic when Vancouver Island collided obliquely with the forming Cordillera.

During the interval from the Late Cretaceous to the Middle Tertiary, the Omineca Belt of the eastern Cordillera continued to be uplifted (which is thought by various authors, such as Brown and Tippet, 1977, to have begun in the Middle Jurassic) and the Rocky Mountain Foreland Fold and Thrust Belt was developed. Also during this time in the western Cordillera, volcanism was predominant in the Intermontane Belt and intrusions were emplaced in the Coast Plutonic Complex.

During the Tertiary, thrusting continued in the Rocky Mountain Belt until the Oligocene. On the western margin of the Cordillera, differential uplift occurred within the Coast Plutonic Complex which may be the result of subduction ceasing as the North American and Pacific Plates came into contact and a transform fault boundary was established (Monger and Price, 1979). To the south off Vancouver Island, Riddihough and Hyndman (1976) have postulated that this subduction is still operative, with the subduction of the Explorer and Juan de Fuca Plates. The evidence given for this is that the

Neogene sediments on the western continental shelf, which represent fore-arc deep deposits, show traces of deformation related to a subduction complex. Heat flow measurements (Hyndman, 1976), gravity, and seismic results may also be interpreted in this way (Riddihough, 1977).

2.3 The Canadian Cordillera: Geophysics and Crustal Structure

From seismological and gravity data it can be concluded that the crust and upper mantle beneath most of the Cordillera are different from those beneath the Precambrian craton. Wickens (1977) concluded from surface wave studies that beneath most of the southern Canadian Cordillera the lithosphere-asthenosphere boundary is at an average depth of 30-40 km. Locally, however, this boundary changes; to the east of the Omineca Belt the boundary drops below the top of the upper mantle with this zone of thick lithosphere extending along the Kootenay Arc and northwest along the Rocky Mountain Trench. To the west of this zone is an arc of thinner lithosphere corresponding to a zone of high heat flow (the Cordilleran Thermal Anomaly). This anomaly is flanked by the Rocky Mountain Trench on the east and the Fraser River on the west (Berry et al., 1971).

Geomagnetic data also suggest the presence of these different crustal zones within the Cordillera. From east to west the first zone is the pre-Cordilleran craton which extends beneath the Rocky Mountains to the Rocky Mountain Trench in the north (north of 50°N) and 80 km west of the trench beneath the Omineca Crystalline Belt in the south (Wheeler and Gabrielse, 1972). To the west of this is a thinner crustal zone which underlies the Columbia Mountains, Interior Plateau and Coast Mountains. This zone is thought to consist of a crust which is less dense than the flanking zones (Figure 2:4) and to have an underlying conductive, hydrated lower crust (Figure 2:5) which is probably partially molten. Caner (1970) suggested that this lower zone is

Figure 2:4 - Gravity models for the Cordillera. The top three graphs show the gravity model when the crustal density is varied laterally across the strike of the Cordillera. Here the data does not fit the central region of the Interior Plateau. The second set of three graphs shows the gravity model when the mantle density is laterally varied; however, only the central region of the Cordillera fits the data. The third set of graphs shows the model when both the crustal and mantle densities are varied across the Cordillera. The thinning of the crust in the central region of the Belt is thought to be due to the Cordillera Thermal Anomaly (after Stacey, 1973).

A-RESIDUAL ANOMALY

B-CRUSTAL DENSITY

C-MANTLE DENSITY

D-DEPTH TO CRUST-MANTLE BOUNDARY

Figure 2:5 - Crustal structure beneath the eastern section of the southern Canadian Cordillera, showing suggested conductive, hydrated lower crustal layer beneath the Omineca Belt (after Caner, 1970).

related to the Cordilleran Thermal Anomaly (as, like the anomaly, the lower zone tapers out northward).

The westernmost crustal zone underlying the Cordillera's western margin (Vancouver Island) seems to be anomalous with respect to the other zones. The main crustal layer is approximately 30 km thick and is underlain by a layer at least 10 km thick (Berry and Forsyth, 1975). This higher velocity layer has been interpreted by Hyndman (1976) to be oceanic crust of the Juan de Fuca Plate descending into the mantle along a northeasterly dipping subduction zone.

As seen from figure 2:4, variations in both crustal and mantle densities across the Cordillera fit the seismic results rather well. The decrease in the upper mantle density beneath the Cordillera relative to the Plains is in good agreement with the decrease in P_n wave velocity and with the surface wave velocity in the upper 100 km of the mantle (Stacey, 1973). It has been suggested by Stacey (1973) that the crustal density decrease may be related to the apparent absence of a lower layer in the crust beneath the Cordillera. Hence from Stacey's results and data obtained by Berry and Forsyth (1975), indicating a crustal thickness of 60 km below the Rocky Mountain Trench, it can be hypothesized that a decrease in crustal and upper mantle densities occurs west of the Rocky Mountain Trench. Therefore, much of the southern Canadian Cordillera can be interpreted as sitting over a bulge in the top of the asthenosphere. This surface slopes down into the mantle on the eastern edge of the Cordillera and slopes into an eastward dipping subduction complex to the west.

2.4 The Southern Rocky Mountains: Geology

The Rocky Mountain Fold and Thrust Belt covers an approximate area of 31,000 square kilometres and extends in section from the Foothills of Alberta in the east to the Rocky Mountain Trench in the west. The geology within the southern section of the belt (south of 55°N) consists of two main sequences of rocks (Figure 2:6): the miogeoclinal shelf deposits which range in age from Windermere to Late Jurassic, and the clastic wedge sequence which ranges in age from Late Jurassic to Early Tertiary (Price and Mountjoy, 1970).

The miogeoclinal rocks are mainly marine deposits dominated by shallow water carbonates and to a smaller extent terrigenous clastic rocks (as described in Section 2.1). The basal sequence is a succession of sandstones and other clastic rocks which overlap the crystalline basement rocks of the Precambrian craton (Churchill Province). These rocks are only exposed in the westernmost section of the Rocky Mountains as the Miette and Lower Gog Groups.

The early Paleozoic deposits reflect an abrupt change in facies. Here, clastics of the Gog Group give way to a carbonate shelf environment which is thought to be a result of downwarping of the cratonic edge (Wheeler et al., 1972). These carbonates overlapped the cratonic interior and thicken to the southwest. In the vicinity of Field, British Columbia, the carbonates grade abruptly into calcareous shales and slates, which may represent the edge of the proto-continental shelf environment.

The rocks of the Middle and Late Paleozoic are separated from the underlying carbonate units by an extensive unconformity, which may be related to the previously discussed Mississippian Event. Above the unconformity, to the southwest, a second shallow carbonate shelf sequence formed, with the Upper Devonian and Carboniferous sequences forming porous reservoir rocks. Reef complexes also developed during the early Upper Devonian and these can be seen to modify the local structure (Wheeler et al., 1972).

Figure 2:6 - The wedge of supracrustal rocks lying on the Hudsonian basement between the Interior Plains east of Calgary, and the Selkirk Mountains at Rogers Pass. The horizontal datum separates the caltic wedge assemblages (above) from the miogeocline-platform assemblage (after Wheeler et al, 1972).

The final sequence consists of Early and Middle Mesozoic rocks in the miogeoclinal assemblage. These rocks are mostly clastics derived from the craton and form a westward thickening wedge which was deposited on a shallow shelf.

The thickness of the miogeoclinal assemblage before deformation increases from about 1,830 metres at the eastern margin to approximately 12,200 metres in the region of the western Main Ranges. This southwesterly increase in thickness is a combination of thickening of individual units toward the southwest and the emergence of units which are truncated or overlapped northeastward along a series of regional unconformities (Price and Mountjoy, 1970).

The second major sequence within the Rocky Mountains is the Late Jurassic to Paleocene clastic wedge sequence. These units are exposed along the eastern Front Ranges, Foothills and Interior Plains and reach a maximum aggregate thickness of about 6,100 metres. The clastic wedge, consisting mainly of terrigenous detritus, represents a complete change in the character of sedimentation due to the onset of orogeny within the foreland belt. The sediments which were shed from the deforming Foreland Thrust Belt were deposited on the flank of the craton in successive pulses, progressively reaching farther northeastward.

The Lower Cretaceous and latest Upper Cretaceous to Paleocene rocks within the sequence are mostly nonmarine. However, early Upper Cretaceous sediments are characteristically marine as subsidence of the foredeep caused minor transgressions of the sea from the north. The nonmarine sediments of the Paleocene eventually spread northeastward over the older shelf deposits to where the nonmarine deposits intertongued with the marine sediments (Price and Mountjoy, 1970).

To the west of the Rocky Mountain Fold and Thrust Belt lie the Omineca Crystalline Belt and the Purcell Mountains. Both of these areas play an

integral role in the formation of the Fold and Thrust Belt. Hence it is necessary to understand their general geology and structure, which has a significance for the formation of thrust faults to the east.

The Omineca Crystalline Belt and the Purcell Mountains can also be generally separated into two assemblages. The first is the westward equivalent of the Windermere sequence found in the Rocky Mountains; however, here the rocks were later metamorphosed. Since the sediments were deposited farther to the west, the craton derived terrigenous sediments are finer grained and intertongue with overlying carbonates and shales of Early Cambrian age.

In the Cambrian, a change in sedimentation occurred when quartzites, pelites and limestones were deposited over the Windermere pelites, grits and conglomerates (Reesor, 1970). A second change in the sedimentary column occurs where the Milford Group limestones (Mississippian-Pennsylvanian), pelites and carbonate-quartzites unconformably overlie the Lardeau Group of the Cambro-Ordovician.

The second assemblage within the area (which has a regional upper limit of 111 Ma, Price and Mountjoy, 1970), a set of highly metamorphosed and deformed units, are the rocks associated with the Shuswap metamorphic core complex. Here, the assemblage is characterized by large domal structures which develop at approximately 80 km intervals along the belt from south to north. Although each dome exhibits its own individual variations, Reesor (1970) and Wheeler *et al* (1972) characterized the domes to be a core zone of migmatite and granitic gneiss and a surrounding mantling gneiss consisting of heterogeneous quartzite, marble and schist. It is generally considered that these domes are the result of remobilized basement, and possibly to some extent migmatized Proterozoic Windermere and younger rocks (Brown, 1978). The emplacement of the gneissic domes (which have been estimated to be Late Jurassic-Early Cretaceous, Brown, 1978) is generally accepted to be the result

of diapirism; however, the contentious issue is whether the domes rose vertically with little to no lateral movement as inferred by Brown and Tippet (1977) and Brown (1978) or whether the domes are the result of a lateral eastward spreading of the infrastructure as shown by Price and Mountjoy (1970).

2.5 The Southern Rocky Mountains: Structure

The structure of the Rocky Mountains is dominated by thrust faults which dip toward the southwest and are concave upward in profile (Figure 2:7). As described by various authors (such as Douglas, 1970; Price and Mountjoy, 1970; Campbell, 1973; and others) the thrust faults, which flatten with depth, gradually cut up through the section but commonly follow the layering over great distances. Many of the faults bifurcate upwards into numerous splays and the total displacement of the thrust is taken up along these splays. Folds are also common within the belt and have developed in conjunction with thrusting, as many of the thrust faults themselves are folded along with the strata. The folds generally verge to the northeast and many thrusts die out into folds along their length. However, since the faults are a mechanically interleaved system and overlap and interfinger with one another along strike, the deformation can be considered to be continuous both across and along the belt on a megascopic scale.

The Rocky Mountains can be subdivided into five structural subprovinces from east to west: the Foothills; Front Ranges; eastern Main Ranges; western Main Ranges; and the Western Ranges. This subdivision is based on changes in structural style, which reflect both southwesterly changes in supracrustal thickness and changes in the structural level that is exposed.

The Foothills consist of deformed Mesozoic clastic rocks and a few large flat-lying thrust sheets involving Paleozoic carbonates (Gordy, 1972). These

EASTERN CORDILLERAN FOLD BELT

Figure 2:7 - Cross-section of the southern Canadian Rocky Mountains and the Omineca Crystalline Belt. (Data used for the Omineca Belt is after Brown, 1978, and for the Rocky Mountains after Price and Mountjoy, 1970.) Section at 51°N.

large thrust bifurcate into numerous splays and are also commonly folded.

The structure within the Front Ranges is characterized by a few imbricate thrust sheets, which expose Upper Paleozoic carbonates and Precambrian sediments. The thrusts within this subprovince dip moderately to steeply toward the southwest, forming linear ranges of Paleozoic carbonate.

A change occurs in structural style at the eastern Main Ranges boundary. Here, there are thick and relatively flat thrust sheets in conjunction with broad open folds and southwesterly dipping gravity faults. Price and Mountjoy (1970) suggest that the character of these structures reflects the fact that they have developed in a thick competent sequence (dominantly carbonate) which is the northeasterly part of the second miogeoclinal wedge (Section 2.1).

The western Main Ranges also show a pronounced change in structural style. This is thought to be due to an incompetent calcareous shale and slate unit within the massive Middle and Upper Cambrian carbonate sequence. Large thrusts within this region are rare, and deformation is commonly in the form of tight folds and anticlinoria.

Southwesterly overturned folds and fault slices are characteristic of the Western Ranges. The fault slices are on the southwestern side of a fan axis that follows the crest of the Porcupine Creek Anticlinorium. These structures continue across the floor of the Rocky Mountain Trench and are overlapped along the Purcell fault by Lower Cambrian and Windermere rocks (Price and Mountjoy, 1970). The northern part of the Western Ranges show tight southwesterly overfolds that are cut up by a complex system of longitudinal and transverse faults.

The Rocky Mountains are bounded on the west by the Purcell thrust, which brings into contact high grade metamorphic rocks on the western side of the trench with low grade rocks of the western Main Ranges and the Western Range.

The overall structural style within the Rocky Mountains reflects a type

of plastic flow or deformation involving large scale strain and translation. The effect of this deformation was the northeastward and upward flow of the sedimentary wedge along a set of thrust faults causing shortening and thickening of the wedge. Also, it has been concluded from seismic studies (Bally et al, 1966), later confirmed from recent seismic work (P. Gordy, personal communication, 1979) and detailed mapping that the deformation within the southern Rocky Mountains did not involve the basement. Thus, two structural levels are evident within the Foreland: the passive undeformed crystalline basement and the layered anisotropic sedimentary suprastructure which moves and thickens northeastward along an array of discrete, discontinuous, interleaved shear surfaces.

2.6 The Southern Rocky Mountains: Tectonic Evolution

The tectonic evolution of the Rocky Mountain Foreland Fold and Thrust Belt has not yet been completely unravelled. Two proposals that have been suggested differ mainly with respect to the possibility of basement involvement beneath the Main Ranges. The first model, proposed by Price and Mountjoy (1970), envisages the thrust faults of the foreland basin forming as a result of supracrustal shortening, which occurred as a response to upwelling and lateral spreading of hot mobile infrastructure (the Shuswap Metamorphic Core Complex) within the Omineca Belt to the west (Figure 2:8).

This hypothesis suggests that large scale (in excess of 200 km) supracrustal shortening occurred in the Rocky Mountains over a passive uninvolved basement. This shortening is associated with upward and lateral displacements within the infrastructure of the Omineca Belt. The upwelling and lateral flow of the infrastructure (hot gneiss and granitic rocks which form the previously described domes of the Shuswaps) occurred beneath a colder, passive, less intensely deformed suprastructure of the sediments and volcanics (which have

COLUMBIA MOUNTAINS   ROCKY MOUNTAINS   PLAINS

Figure 2:8 - The structure of the Rocky Mountain Thrust and Fold Belt according to the Price and Mountjoy (1970) model. It may be noted that the core zone of the Selkirks is envisaged as a mobile mass spreading laterally.

subsequently been eroded). It is therefore suggested that the upwelling of the infrastructure is primarily a buoyancy phenomenon reflecting an unstable density distribution and contrasts in ductility, and that the lateral spreading which resulted in the thrusting of the suprastructure to the east (due to a push from behind) was primarily a gravitational effect.

The second model, proposed by Campbell (1973), hypothesized that shortening within the supracrustal rocks of the Foreland Belt was not as great as proposed by Price and Mountjoy (1970), but only about 35 km. It is suggested in this model that the core zone of the metamorphic complex, together with the Main Ranges of the Foreland Belt, did not move eastward by an amount equivalent to the supracrustal shortening in the eastern Rocky Mountains, but rather that vertical movements predominated in these regions. From this it is implied that passive basement only exists below the Foothills and Front Ranges while involved basement is present farther west, in the Main Ranges (Figure 2:9). These conclusions have been based on a change in structural style across the Main Ranges-Front Ranges boundary and from the absence of flat-lying faults within the Main Ranges. This change, however, can be simply explained on the basis of stratigraphic contrasts. Also as stated earlier (Section 2.3), it has been demonstrated from recent mapping and seismic work within the western Foreland that basement is uninvolved in deformation up to 20 km from the Trench (P. Gordy and R. Price, personal communication, 1979). Therefore, Campbell's hypothesis of minimum crustal shortening must be disregarded at least for the southern section of the Foreland Fold and Thrust Belt.

The above arguments still leave the problem of a driving mechanism unresolved if one does not accept the laterally spreading tongue hypothesis of Price and Mountjoy (1970). In the model proposed by Campbell (1973) it was suggested that westward movement of the craton, which corresponded to the opening of the Atlantic, rafted sediments of the Foreland and eastern Omineca

COLUMBIAN MOUNTAINS

ROCKY MOUNTAINS

Figure 2:9 - Structure of the Rocky Mountain Thrust and Fold Belt and related Shuswap Metamorphic Core Complex according to the Campbell (1973) model. Note that movements west of the Front Ranges are predominantly vertical.

(Selkirk and Purcell Mountains) Belts up against the Shuswap metamorphic complex. This buttressing of the sediments has been suggested to be the result of underthrusting the Precambrian craton beneath the eastern Omineca Belt, causing predominantly folding in the Selkirk and Purcell Mountains with thrusting occurring in the supracrustal sediments of the Foreland. Hence, as the sedimentary wedge moved westward with respect to the Shuswap Core Complex, thrust slices progressively detached from the basement in the Foreland (and to some extent in the eastern Omineca Belt), which caused deformation to migrate from west to east. To the west of the Rocky Mountain Trench (north of 51°N) it has been suggested by some authors (such as Brown, 1978) that underthrust craton was remobilized to form the Selkirk gneiss domes. This was thought to have been accomplished by the uplift and northeastward translation of the basement mass along the Purcell fault as underthrusting continued.

If the basement is not involved in the foreland belt, as the previously discussed evidence seems to imply, at least 200 km of shortening is needed to account for the supracrustal evidence. It seems unrealistic, however, to assume that this entire 200 km was taken up by underthrusting the craton beneath the Omineca Belt and Purcell Mountains as there is no evidence for this amount of cratonic crust to be present west of the Rocky Mountain Trench.

As described earlier, there is evidence of an 80 km width of cratonic crust beneath the Purcell Mountains from geophysical observations. It is also evident that farther to the north (Selkirk Mountains) basement has been involved beneath the core complex. From these observations it is possible to conclude that the cratonic basement has moved westward with respect to the Foreland Belt a minimum of 80 km. However the problem remains of where the additional 120 km of shortening is being taken up. In a recent discussion by Monger and Price (1979) it was suggested that shortening has been accommodated

at a deep level and farther west, beyond the Precambrian cratonic margin, in what is thought to be mainly oceanic rocks, presumably by subduction. Another possibility is to 'subduct' directly the excess cratonic crust beneath the western Cordillera. Subduction of continental crust, up to a few tens of kilometres, has been demonstrated by Molnar and Gray (1979) to be plausible and may possibly serve as a working hypothesis for the solution of a problem that obviously needs more work.

CHAPTER 3

THEORETICAL MODELS FOR THRUST FAULT FORMATION

3.1 Fundamental Definitions and Equations of Elasticity Theory

The formation of faults has been investigated theoretically and experimentally for quite some time. A comprehensive treatment was given by Hubbert (1951), where the famous sand-box experiment was used to show that normal and reversed faults occur in a predictable, regular manner, related to the stress system. In these experiments it was shown that if the principal compressive stress (σ_1) was in the horizontal direction, a reverse fault would form at approximately 28° from the direction of maximum compression, provided the yield strength of the rocks was exceeded. The same result can be arrived at analytically by using the Coulomb-Mohr equation.

This simple theory can be expanded in order to explain thrust faulting within the foreland of an orogenic belt. It has been postulated by various authors (such as Hubbert, 1951; and Hafner, 1951) that σ_1 lies in an approximately horizontal plane as a result of a tectonic force acting at one end of the rock mass within a foreland belt. This is balanced by the bottom shearing stress (σ_{12}). The magnitude of σ_{12} at any given distance along the bottom of the block must be large enough to balance the force due to the stresses difference in the horizontal direction (Hubbert, 1951). Therefore, we now can approximate the boundary stresses and derive complete solutions for geologically significant stress distributions for a rectangular block. The pertinent elasticity theory, which as will be evident in the following section mainly involves the treatment of the Airy stress function, is briefly reviewed here (using the index notation, with the summation convention). For a more comprehensive discussion the reader is referred to any one of numerous texts on elasticity theory (such as Landau and Lifshitz, 1975; and Ranalli and

Gale, 1976).

If we define stress as force acting on a surface element contained within a volume and strain as a kinematic quantity which describes the deformation of a body, it can be shown that conditions of compatibility must be satisfied in order to have the displacements single valued and continuous. Here, the infinitesimal strain tensor (ϵ_{ik}) that corresponds to internal deformations of the material and given by

$$\epsilon_{ik} = \frac{1}{2} \left[\frac{\partial u_i}{\partial x_k} + \frac{\partial u_k}{\partial x_i} \right] \quad (3:1)$$

must satisfy

$$\frac{\partial^2 \epsilon_{ij}}{\partial x_m \partial x_n} + \frac{\partial^2 \epsilon_{mn}}{\partial x_i \partial x_j} - \frac{\partial^2 \epsilon_{im}}{\partial x_j \partial x_n} - \frac{\partial^2 \epsilon_{jn}}{\partial x_i \partial x_m} = 0 \quad (3:2)$$

which consists of eighty-one equations of which only six are essential (Ranalli and Gale, 1976).

It can be shown that the equilibrium conditions are met if the resultant body and surface forces, and the resultant moment about any point within the body, are zero. Also from the derivation of the equilibrium equation we can conclude that the stress tensor is symmetric. If we consider the resultant body and surface forces within a body, at equilibrium they must be zero, hence

$$\frac{\partial \sigma_{ij}}{\partial x_j} + \rho X_i = 0 \quad (3:3)$$

If we assume that the shearing stress acting on the faces normal to the x_1 direction yield positive moments, the shearing stresses acting on faces normal to the x_2 direction give rise to negative moments (Figure 3:1); thus it can be shown that $\sigma_{ij} = \sigma_{ji}$.

Now that stress, strain, the strain compatibility equations and the equations of equilibrium have been briefly discussed we can define 1) the rheological equations of state (Hooke's Law), 2) the stress compatibility

Figure 3:1 - Stress element showing the orientation of the normal and shear stresses. Compression is taken to be negative.

equations, and 3) the Airy stress functions.

Hooke's Law states that stress is linearly proportional to strain, thus

$$\sigma_{ij} = \lambda \theta \delta_{ij} + 2\mu \epsilon_{ij} \quad (3:4)$$

where $\theta = \epsilon_{ii}$, ϵ_{ij} is the strain tensor, and λ and μ are two elastic parameters called Lamé constants. The strain compatibility equations can now be defined in terms of stresses by considering the relations between the Lamé constants, Young modulus, Poisson's ratio and the bulk modulus, and the equilibrium equations. Therefore, using Hooke's Law (equation 3:4) and differentiating the equilibrium equations with respect to x_i we can obtain the compatibility conditions in terms of stress

$$\frac{\partial^2 \sigma_{ij}}{\partial x_k \partial x_k} + \left[\frac{1}{1-\nu} \right] \frac{\partial^2 \theta}{\partial x_i \partial x_j} = -\rho \left[\frac{\nu}{1-\nu} \right] \partial_{ij} \frac{\partial x_k}{\partial x_k} - \left[\rho \frac{\partial x_i}{\partial x_j} \frac{\partial x_j}{\partial x_i} \right] \quad (3:5)$$

Assuming that the body force is a constant vector (i.e. gravity), equation 3:5 becomes

$$\frac{\partial^2 \sigma_{ij}}{\partial x_k \partial x_k} + \left[\frac{1}{1+\nu} \right] \frac{\partial^2 \theta}{\partial x_i \partial x_j} = 0 \quad (3:6)$$

where θ is a harmonic function as it satisfies Laplace's equation (Ranalli and Gale, 1976). Taking the Laplacian of equation 3:6, assuming that σ_{ij} is continuous and differentiable up to the fourth order, and that the body force vector is constant, it can be concluded that σ_{ij} is a biharmonic function. It also follows that ϵ_{ij} is biharmonic from the linear correspondence between the stress and strain components; hence in two dimensions and for constant body forces the stress compatibility equations become,

$$\frac{\partial^2 \theta}{\partial x_i \partial x_i} = \frac{\partial^2}{\partial x_j \partial x_j} (\sigma_{11} + \sigma_{22}) = 0 \quad (3:7)$$

Also, since the equilibrium equations must be satisfied, we can choose a function ϕ , called the Airy stress function, such that

$$\sigma_{11} = \frac{\partial^2 \phi}{\partial x_2^2} + \rho V \quad (3:8)$$

$$\sigma_{22} = \frac{\partial^2 \phi}{\partial x_1^2} + \rho V \quad (3:9)$$

$$\sigma_{12} = \frac{\partial^2 \phi}{\partial x_1 \partial x_2} \quad (3:10)$$

which will satisfy equation 3:3. Therefore from substitution into equation 3:7 and from the fact that $\partial^2 V / \partial x_j \partial x_j = 0$ we get

$$\frac{\partial^4 \phi}{\partial x_1^4} + 2 \frac{\partial^4 \phi}{\partial x_1^2 \partial x_2^2} + \frac{\partial^4 \phi}{\partial x_2^4} = 0 \quad (3:11)$$

Thus ϕ (the Airy stress function) is biharmonic and any solution of equation 3:11 will satisfy both the equation of equilibrium and the compatibility conditions.

3.2 Applications of Elasticity Theory to Thrust Fault Formation

In order to represent stresses in a thrust block we must first have a working hypothesis on standard stress state. According to the definition given by Anderson (1942), the standard (lithostatic) state represents a two-part stress system: 1) the effect of gravity and 2) the superposed horizontal stress which is constant in any particular horizontal plane, but increases vertically; thus, if x_1 is the horizontal, and x_2 the vertical axis,

$$\begin{aligned} \sigma_{11} &= \sigma_{22} = -\rho g x_2 \\ \sigma_{12} &= 0 \end{aligned} \quad (3:12)$$

The elastic stresses in a laterally confined self-gravitating body can be obtained from

$$\sigma_{22} = \rho g x_2, \quad \sigma_{11} = \sigma_{22} \nu / (1 - \nu) \quad (3:13)$$

where the horizontal components of stress due to the weight of the body are smaller than the vertical components ($\nu < \frac{1}{2}$). However, one standard state

may lead to the second if we assume that equations 3:13 relax to equations 3:12 over geologic time by plastic flow. Then, with initial state of $\sigma_{11} = \sigma_{22}$, we can still assume superposition of a new stress system over elastic time limits.

Two major assumptions have to be made in methods of calculating stress systems in rocks. The stressed block will be treated as if it were: 1) homogeneous and, 2) isotropic; furthermore, we will consider only a two dimensional case. Since within any foreland belt there are at least two or more lithologies which will differ in composition along their length and width, our two assumptions are geologically not totally valid. However, for the use of a 'first approximation' mechanical model our assumptions are justified.

The internal state of stress within a block can be represented by using a set of two curves: 1) stress trajectories, being the tangents to the principal stress directions at all points; and 2) lines of maximum shear stress directions.

The method used by Hafner (1951) for the determination of the internal stress distributions is based on the previously discussed Airy stress function. The problem is to find a suitable function which satisfies the equation (equation 3:11) and also more importantly gives a system of geologically realistic boundary stress conditions. On the basis of the previously discussed derivation of the Airy stress function and the working hypothesis of standard state, Hafner (1951) proposed various examples of two-dimensional stress systems. One example is to assume the presence of a supplementary horizontal stress but absence of an associated vertical stress component, hence

$$\sigma_{22} = \frac{\partial^2 \phi}{\partial x_1^2} = 0 \text{ for all values of } x_1 \quad (3:14)$$

This can then be integrated to give

$$\phi = cf_1(x_2) x_1 + ax_1 + bf_2(x_2) + d \quad (3:15)$$

which satisfies equation 3:11. Thus from equation 3:11 we obtain

$$cx_1 f_1^{iv}(x_2) + bf_2^{iv}(x_2) = 0 \quad (3:16)$$

where f_1^{iv} and f_2^{iv} must be zero. At the surface the boundary conditions require that $f_1^i(x_2) = 0$ for $x_2 = 0$; therefore Hafner (1951) set up three subgroups

$$1) f_1^i(x_2) = 0, f_2^{ii}(x_2) = x_2 + d \quad (3:17)$$

thus $\sigma_{11} = bx_2 + d$, $\sigma_{22} = 0$ and $\sigma_{12} = 0$;

$$2) f_1^i(x_2) = x_2, f_2^{ii}(x_2) = 0 \quad (3:18)$$

thus $\sigma_{11} = cx_1$, $\sigma_{22} = 0$ and $\sigma_{12} = -cx_2$;

$$3) f_1^i(x_2) = \frac{1}{2}x_2, f_2^{ii}(x_2) = 0 \quad (3:19)$$

thus $\sigma_{11} = cx_1x_2$, $\sigma_{22} = 0$ and $\sigma_{12} = -\frac{1}{2}cx_2^2$.

The first case is simple, because if b is zero the superposed stress is restricted to a constant horizontal component $\sigma_{11} = d$ and thus the maximum shearing stress is also constant. In the second case or a combination of equations 3:18 and 3:19 the constant d is zero. Hence, these equations together with equations 3:12 (standard state equations) will give

$$\begin{aligned} \sigma_{11} &= cx_1 + bx_2 - ax_2 \\ \sigma_{22} &= -ax_2 \\ \sigma_{12} &= -cx_2 \end{aligned} \quad (3:20)$$

where $a = \rho g$ and $-ax_2$ represents hydrostatic pressure. σ_{11} has constant gradients in both the vertical and horizontal directions; also, the shear stress is constant in any horizontal plane and it increases at a constant rate vertically equal to the lateral gradient of σ_{11} . Therefore, using equations 3:20 and assuming the presence of a supplementary horizontal stress but absence of an associated vertical stress, it can be determined that if the supplementary horizontal pressure has only a small lateral gradient (half the order of the vertical gradient) the potential area of thrusting is confined to a shallow, gently dipping wedge (Figure 3:2). The deformation will consist

Figure 3:2 - Supplementary stress system of superimposed horizontal pressure; here the horizontal stress has a vertical gradient equal to the horizontal one (after Hafner, 1951).

of a series of thrust slices covering a broad belt but only extending to a shallow depth. The presence of the vertical gradient will steepen the wedge and the boundary becomes nearly vertical only if the horizontal gradient approaches the magnitude of the vertical gradient. In this event Hafner (1951) suggests that thrusting can take place throughout a thick, but narrow zone of the crust. These elastic models based on Airy's stress function are a first step in the mechanical analysis of foreland thrusting.

3.3 Fundamentals of Plasticity Theory

The Rigid Plastic Body

As discussed in the previous section, in elasticity theory we are dealing with an elastic body which will recover its original form if the stress is removed. An elastic body is also characterized by a one-to-one correspondence between stress and strain. In plasticity theory we will be dealing with a rigid-perfectly plastic body, which does not show the above stress-strain relationships. Here, an infinity of stresses are compatible with the same strain, and conversely the same strain is compatible with an infinite number of different stresses (Prager, 1959). Hence, while elastic stress trajectories still give an idea of the stress state within a foreland belt, plasticity theory is a better tool for modelling, because rocks do flow (i.e. they show permanent deformation).

The following characteristics of the rigid-perfectly plastic body are used in plasticity theory:

- 1) the body is isotropic and homogeneous;
- 2) the body shows no strain hardening;
- 3) the elastic strain is negligible with respect to the plastic strain;

- 4) the yield criterion is independent of hydrostatic pressure or tension;
- 5) the yield criterion that must be satisfied is that of Von Mises; i.e., the body becomes plastic when the deviatoric strain energy throughout the body reaches a value κ , which is characteristic of the material;
- 6) the flow condition is that of Levy and Von Mises, i.e.

$$t_{ij} = \frac{1}{2} \left[\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right] = \lambda \sigma'_{ij} \quad (3:21)$$

where $\sigma'_{ij} = \sigma_{ij} + \delta_{ij}p$, $p = -1/3 (\sigma_{ii})$ and $\lambda \geq 0$.

While the body is undergoing deformation, the velocity and stress fields within it must satisfy both the equations of motion and the continuity equation. Because the body is incompressible, the density (ρ) is constant and $D\rho/Dt$ is zero (where D/Dt is the material derivative), and hence the equations for motion are

$$\frac{\partial \sigma_{ij}}{\partial x_i} + \rho X_j = \rho \frac{Dv_j}{Dt} \quad (3:22)$$

where X_j is the body force per unit mass. Also, neglecting body forces and for steady-state flow, the equations of motion reduce to those of equilibrium

$$\frac{\partial \sigma_{ij}}{\partial x_j} = 0 \quad (3:23)$$

which must be satisfied by the stress system of the deforming body. Therefore equations 3:21 and 3:23, the continuity equations from incompressibility $\partial v_i / \partial x_i = 0$ and the yield criterion ($\sigma'_{ij} \sigma'_{ij} = 2\kappa^2$) are the basic equations of the theory of rigid perfectly plastic bodies.

Plane Plastic Strain

The state of plane strain has been defined by various authors (such as Hill,

1950) as 1) the flow is everywhere parallel to a given plane, and 2) the motion is independent of x_3 (coordinate axis normal to the plane of flow). Because the volume of an element of plastic-rigid material is constant, each incremental distortion in a state of plane strain consists of pure shear. Hence for an isotropic material the state of stress at each point is a pure shear stress superimposed to a hydrostatic pressure (p). The stress normal to the flow planes (σ_{33}) is therefore equal to $-p$ (where compression is negative) and the other principal stresses to $-p \pm \sigma_{12}$. Thus the yield criterion is of the form

$$\frac{\sigma_1 - \sigma_2}{2} = \sigma_{12\max} = \frac{1}{2} \left[(\sigma_{11} - \sigma_{22})^2 + 4\sigma_{12}^2 \right]^{1/2} = \kappa \quad (3:24)$$

The clockwise angle (θ) between the x_1 axis and the trajectory of the algebraically largest principal stress is given by

$$\tan 2\theta = \frac{2\sigma_{12}}{(\sigma_{11} - \sigma_{22})} \quad (3:25)$$

which shows that the surfaces on which the maximum tangential stresses act make angles $\pm \pi/4$ radians with the trajectory of the principal stress directions.

From the transformation equations for the stress tensor (see Ranalli and Gale, 1976, p. 14) and making the change $\theta = \phi + \pi/4$, it follows that the equilibrium equations may be written, in plane strain, as

$$\begin{aligned} -\frac{\partial p}{\partial x_1} - 2\kappa \cos 2\phi \frac{\partial \phi}{\partial x_1} - 2\kappa \sin 2\phi \frac{\partial \phi}{\partial x_2} &= 0 \\ -\frac{\partial p}{\partial x_2} + 2\kappa \cos 2\phi \frac{\partial \phi}{\partial x_2} - 2\kappa \sin 2\phi \frac{\partial \phi}{\partial x_1} &= 0 \end{aligned} \quad (3:26)$$

which must be of hyperbolic type (see Korn and Korn, 1968, section 10.3) in order for solutions to be constructed. It can be proven that for the hyperbolicity of the system to be established it is necessary to show the existence of two distinct real families of curves called characteristics. The existence

of these curves can be shown using Cauchy's problem (Kachanov, 1971, p. 152), and from this proof it follows that characteristics coincide with slip lines.

Kachanov (1971) defines a slip line as "a line which is tangent at every point to the surface of maximum tangential stress"; consequently, there are two orthogonal families of slip lines characterized by $x_1 = x_1(\alpha, \beta)$ and $x_2 = x_2(\alpha, \beta)$ where the parameters of α and β are used to label the two families of slip lines, β being constant along α -lines and vice-versa.

Taking as coordinate axes s_1 and s_2 , which coincide in direction with the tangents to the slip lines, transforms equations 3:26 into

$$\frac{\partial}{\partial s_\alpha} (\sigma - 2k\theta) = 0; \quad \frac{\partial}{\partial s_\beta} (\sigma + 2k\theta) = 0 \quad (3:27)$$

and where $\partial/\partial s_\alpha$ and $\partial/\partial s_\beta$ are derivatives along the α and β lines. If we now take P (Figure 3:3) to be an arbitrary point on the slip line it follows that along slip lines of the α and β families we will respectively have (Kachanov, 1971)

$$\begin{aligned} \frac{\partial x_2}{\partial x_1} &= \tan \theta & \frac{\partial x_2}{\partial x_1} &= -\cot \theta \\ \frac{\sigma}{2k} - \theta &= \text{constant} = \xi & \frac{\sigma}{2k} + \theta &= \text{constant} = \eta \end{aligned} \quad (3:28)$$

Thus, in passing from one slip line to another within the α family ξ generally will change; similarly η changes within the β family.

If both the slip line field and the parameters ξ and η are known σ and θ can be determined at each point; hence σ_{11} , σ_{22} and σ_{12} will be known.

Velocity Fields and Stress and Velocity Discontinuities

From equation 3:25 we can obtain the axes of principal stress; hence the slope of the axes of principal strain rate is given by

Figure 3:3 - Slip lines; α and β slip lines are tangent at every point to the surface of maximum tangential stress. Along α lines β is constant and along β lines α is constant. The α line is inclined to the right of the first principal direction at 45° and the β line is inclined to the left of the first principal direction at the same angle (after Kachanov, 1971).

$$\frac{\partial v_1}{\partial x_2} - \frac{\partial v_2}{\partial x_1} \tan 2\phi + \frac{\partial v_1}{\partial x_1} - \frac{\partial v_2}{\partial x_2} = 0 \quad (3:29)$$

since the material is isotropic and $\theta = \phi + \pi/4$. Thus any valid velocity field must satisfy the incompressibility condition ($\partial v_1/\partial x_1 = 0$) and equation 3:29. It can be proven that characteristics of stresses and velocities coincide with slip lines. If we assume the velocity components to be known along a closed curve c , and since mass must be conserved, v_1 just inside c must be equal to v_1 just outside c at any point P (see Figure 3:4). Thus, transferring to a local coordinate system (s_1, s_2) the derivatives $\partial v_1/\partial s_1$ and $\partial v_2/\partial s_2$ are continuous across c and $\partial v_2/\partial s_2$ is continuous provided ϕ does not equal zero or $\pi/2$ radians. If it does, the tangent to c at P coincides with a characteristic direction; hence the stress and the velocity characteristics coincide, with their trajectories being those of maximum shear stress.

It is evident from equation 3:29 and the equation of continuity that the rate of elongation along slip lines is zero (Ranalli and Gale, 1976). Thus, if we consider an infinitesimal segment ds_α of an α -line and neglect second order quantities, the rate of elongation in the α direction is $\{(v_1 + dv_1) - (v_2 + dv_2)\}$, hence along slip lines

$$\begin{aligned} dv_\alpha - v_\beta d\phi &= 0 \text{ on an } \alpha\text{-line} \\ dv_\beta - v_\alpha d\phi &= 0 \text{ on a } \beta\text{-line} \end{aligned} \quad (3:30)$$

where v_α and v_β are velocity components in the α and β directions respectively.

We shall now consider L to be a line of discontinuity (Figure 3:5) and s to be an infinitesimal element lying on the line. Due to the equilibrium considerations, the normal and shear stress acting along the element must be continuous. The only stress which may be discontinuous (in local coordinates)

Figure 3:4 - Velocity components in plane strain (after Ranalli and Gale, 1976).

Figure 3:5 - L is a line of discontinuity with an infinitesimal element lying on it. The normal and tangential stresses act on the element as shown (after Kachanov, 1971).

is the tangential normal stress (σ_{22}). From the plasticity condition (which is assumed to be valid on both sides of L) the magnitude of the discontinuity is constrained, thus

$$\sigma_{22} = \sigma_{11} + 2\sqrt{\kappa^2 - \sigma_{12}^2} \quad (3:31)$$

Since $\sigma_{12} = \kappa$ along a slip line, a line of stress discontinuity cannot coincide with a slip line. Hence, in crossing a line of stress discontinuity, the slip lines undergo reflection about the tangent to the discontinuity at the intersection point.

It is also possible (Hill, 1950) to have a discontinuity in the tangential velocity component across a slip line. This discontinuity must be regarded as the limit of an infinitely great shear strain rate; hence v_β can be discontinuous across a β line, $(\partial v_\beta / \partial s_\alpha) \rightarrow \infty$, and v_α can be discontinuous across an α line, $(\partial v_\alpha / \partial s_\beta) \rightarrow \infty$, with the jump in v_α or v_β being constant along their respective slip lines. These discontinuities are frequent in a non-hardening plastic body, and correspond in a real material to a more or less narrow transition region where the shear strain rate is very large (thrust faults).

Compression of a Block Between Rough Plates

If we now consider the problem of a plastic layer between two rigid plates, the layer will flow from the central region towards the edges if it is compressed (Figure 3:6). Even though this example on the application of plasticity theory is not very realistic with respect to foreland belts, it offers a good first approximation to the slip line patterns that are developed when a plastic material is subjected to compressive plastic flow.

It is evident that large differential stresses arise at the contact surfaces and in order to develop plastic deformation these stresses must attain the yield value κ (Hill, 1950). Following Prandtl's solution, we assume that the thickness ($2h$) of the layer is substantially less than its breadth ($2L$),

Figure 3:6 - Slip line field in a block compressed between perfectly rough plates, together with a comparison of the actual pressure distribution with Prandtl's. In our case we would only consider the bottom half of Prandtl's cell (after Hill, 1950).

so that we can disregard exact boundary conditions at either end of the cell. In the case of a thrust belt we only have to consider half the thickness of Prandtl's cell or (h) .

From the equations of equilibrium, the yield criterion and the stresses which satisfy these conditions we can find the velocity components when each of the plates moves on the layer with a speed s

$$\begin{aligned} v_1 &= V + s (x_1/h) - 2 (1 - x_2^2/h^2)^{1/2} \\ v_2 &= -sx_2/h \end{aligned} \quad (3:32)$$

These equations satisfy the incompressibility condition and the inverse of equation 3:29 which states that the direction of the surface of ~~maximum~~ tangential stress coincides with the direction of the surface which experiences the maximum rate of shear strain, for arbitrary values of the constants s and V . With the use of equation 3:28 the differential equations of slip lines can be found which when separated and integrated give the parametric equations of the slip line families

$$\begin{aligned} x_1 &= -h (2\phi + \sin 2\phi) + c_1 & (\alpha) \\ x_2 &= h \cos 2\phi \\ \bar{x}_1 &= h (2\phi - \sin 2\phi) + c_2 & (\beta) \\ x_2 &= h \cos 2\phi \end{aligned} \quad (3:33)$$

Thus two orthogonal families of cycloids generate the slip line field (h is the radius of the generating circle). $x_2 = \pm h$ are straight lines representing the envelopes of the cycloid families, hence are lines of discontinuity (Kachanov, 1971, p. 225).

The Prandtl cell, although it is not valid along the edges, still gives a good first order approximation of the slip line distribution for a thin layer.

3.4 Applications of Plasticity Theory to Thrust Fault Formation

Glacier Flow Theory

If we treat the foreland thrust belt as a rigid plastic body and hypothesize that deformation occurs as plane plastic strain, we will in the case of some models be using arguments similar to those used in the study of the deformation and movement of glacier ice. Since rocks within a foreland belt and glacier ice have similar material properties (both are polycrystalline solids which deform plastically), it has been suggested that the same rheological equations can be used in describing their deformation. Therefore, it is of some interest to briefly discuss plasticity theory related to glacier ice, since it is pertinent to proposed models of thrust fault formation.

The first approximation to an ice sheet is that of a parallel-sided slab of thickness h resting on a rough surface with a slope γ (Figure 3:7A). It is assumed that the lateral extent of the mass is large compared to h , and that the mass does not slide on its base; hence deformation is solely due to gravity. It can then be shown that the weight component of a column of ice which is perpendicular to the basement ($\rho g h \sin \gamma$) is balanced by the shear stress across the base of the column (Paterson, 1969). Thus

$$\sigma_{12} = \rho g h \sin \gamma \quad (3:34)$$

which implies that $h \sin \gamma$ must be constant along the base.

The above model assumes that the surface slope is parallel to the base; however, in the case of most glaciers and in foreland belts this is definitely not the case. Thus we now let the basement slope (θ_0) be different from the surface slope (α). We consider two boundaries separated by ∂x_1 which enclose an element (ABCD) of unit width (Figure 3:7B). Assuming that hydrostatic pressure everywhere equals the pressure of the overlying mass we can approximate the normal pressure on AB as $\frac{1}{2} \rho g h$ (where $AB = h$) and on CD by $\frac{1}{2} \rho g h - d/dx(\frac{1}{2} \rho g h)$

Figure 3:7 - (A) coordinate system for a parallel-sided slab; (B) equilibrium of wedge-shaped block of ice. Here the basal angle is dipping in the opposite sense to that of the surface slope; this is the case within most foreland thrust belts (after Paterson, 1969).

Angles above the horizontal are taken to be negative.

Therefore we obtain

$$\sigma_{12} = \rho g h \alpha \quad (3:35)$$

since $\partial h / \partial x_1 = \theta_0 - \alpha$ and $\sin \theta_0 \approx \theta_0$ for small angles.

The above analysis does not consider any variation of velocity (v) with distance travelled along the glacier (x_1). Thus, a more general solution than the one for laminar flow should be obtained. If we again assume a parallel-sided slab it can be shown using the equation of incompressibility that the solution for the velocity components which satisfy the boundary conditions (no shear stress on the free surface of the slab and a zero velocity component at the base) are (from Paterson, 1969, p. 96)

$$v_1 = v_0 + r x_1 + f(x_2) \quad (3:36)$$

$$v_2 = r(h - x_2) \quad (3:37)$$

where v_0 is the velocity at the origin, r is the longitudinal strain rate and $f(x_2)$ is a function of x_2 which can be determined from the equations for the strain rates.

The stresses within the slab can now be determined, and when substituted into the yield condition (equation 3:24) give

$$\sigma_{11} = -p - \rho g x_2 \cos \gamma \pm 2 \left[\sigma_{12}^2 - (\rho g x_2 \sin \gamma)^2 \right]^{1/2} \quad (3:38)$$

From the determination of the stresses, σ_{22} and σ_{12} increase linearly with depth, with σ_{22} becoming more compressive.

From equation 3:38 two solutions are possible for σ_{11} . The first, where the last term is positive, makes $\sigma_{11} + p$ a positive value at the surface and σ_{11} will be greater than σ_{22} at all depths; this case is termed extending flow. When the square root is negative, $\sigma_{11} + p$ will also be negative at the surface and $\sigma_{11} < \sigma_{22}$ at all depths; this case is termed compressive flow (Figure 3:8). Thus $r > 0$ in extending flow and $r < 0$ in compressive flow, with laminar flow ($r = 0$) being a special case.

If we now assume that the upper and lower boundaries of the slab are not

Extending Flow

Figure 3:8 - Slip line fields under extending and compressive flow.

Compressive Flow

parallel, then the condition that the velocity component perpendicular to the basement must be zero gives

$$-v_1 \sin(\theta_0 - \alpha) + v_2 \cos(\theta_0 - \alpha) = 0 \quad (3:39)$$

Thus, the velocity component v_1 is greatest at the surface and decreases with depth, v_2 does not depend on x_1 and varies linearly with depth (depending on r) and the shear stress will equal the yield stress at the base of the slab.

Flow Theory Related to the Foreland Belt

A mechanical paradox has been present in the models on the movement of thrust slices within a foreland belt for a number of years. The problem is that thrust faults show displacements of up to several tens, or even hundreds, of kilometers, while from classical models on thrust formation the rocks could not withstand the stress required to move them over such distances. This paradox arises in the classic 'push from behind' theory proposed by Hafner (1951). Here, the pushing stress has to be so strong (approximately 6 times stronger than the crushing strength, according to Hubbert and Rubey, 1959) that it would crush the back of the thrust rather than cause it to move en bloc across its base. Modifications have been the presence of high pore fluid pressure along the base of the thrust (Hubbert and Rubey, 1959) or a relatively weak basal layer (or décollement zone) such as salt, anhydrite or shale on which the block slides (Kehle, 1970). However, these conditions are exceptional and do not exist everywhere there is a thrust fault.

A second model for thrust sheet emplacement is the gravity spreading hypothesis proposed by Ramberg (1967, 1977a) and by Elliott (1976a, 1976b). This model is basically the same as that used to describe the spreading of ice sheets (see above). Here, the mechanism of horizontal spreading is similar: the plastic flow is due to the weight of the overlying mass at each point within the spreading mass. The difference between the two mechanisms

is that ice will creep as a result of accumulation within the central region, while thrust faults move due to the presence of rising zones in the core of the orogen. An example of this may possibly be the rise of gneissic domes within the Omineca Belt as described by Price and Mountjoy (1970).

In the gravity spreading model discussed by Elliott (1976a), identical arguments are used in order to determine the basal shear stress (σ_{12}) as were used in the subsection on glacier flow (Paterson, 1969). Since $(\theta_0 + \alpha)$ is not much greater than 5° , Elliott (1976a) has concluded that the following is valid for a thrust belt

$$\sigma_{12} = \rho g H \alpha \quad (3:40)$$

which is the same as equation 3:34 where H is the thickness of the belt. From this analysis the shear stress at the base of the thrust is a function of the surface slope (α). Hence in order to have a geologically realistic surface slope and allow plastic flow, the yield strength of the rock within the thrust must be small: an upper limit of 200 bars has been suggested by Elliott (1976a). This value was arrived at by 1) extrapolation of experiments carried out by Heard and Rubey (1966) which have shown that greywacke has a strength of 1 kbar which is reduced to 10% of this value if minor quantities of weak rock is present; and 2) from the fact that for high-strength rocks, high heat flow should be present, since for strong rocks there would have to be large amounts of mechanical work expended in deformation which would produce a high heat flow. Therefore, from this and the fact that there are not anomalously high heat flow values over most foreland thrust belts, it can be concluded that the strength is no greater than approximately 200 bars.

From equation 3:40 Elliott (1976a) calculated the regional shearing stress acting along the base of the foreland belt, done by averaging H over 60 km intervals, and taking $\rho = 2.70 \times 10^3 \text{ kgm}^{-3}$. A typical order-of-magnitude value for the average basal shearing stress is 20 bars with the maximum being 75 bars

Figure 3:9 - (A) Cross-section of the Rocky Mountains; (B) Variations in average regional surface slope α ; (C) Thickness H from the surface to the basement. Each value of H and α is averaged over lengths of 60 km. (D) The regional basal stress σ_{12} which is equivalent to the yield stress. (E) Gravity spreading of the Rocky Mountains, with the cross denoting the highest point of topographic elevation, which acts as the spreading centre.

(Figure 3:9). If we compensate for isostasy as shown by Price (1973), i.e., decrease the original basal angle from its present 3 to about 1.5, the shearing stress in the basal rocks can be estimated to have been 50 bars. However, two problems arise from this analysis, the first geological and the second mechanical. From the geological viewpoint, it can be seen in figure 3:9 that thrust faults occur in the Front Ranges and Foothills while within the Main Ranges normal faults occur. Following Elliott's analysis these will all have to occur contemporaneously which in the case of the Rocky Mountains is not true. As described in Section 2.5, deformation within the Foreland Belt migrated from west to east, not from the central region outward, nor was it contemporaneous across the belt. The mechanical objection has been stated most clearly by Ramberg (1977b), who argues that equation 3:40 cannot be used for gravity spreading within thrust belts, because in the plastically flowing thrust slice, the true horizontal stress is less than what corresponds to the lithostatic conditions through the bulk of the thrust slice. The reason a thrust slice moves forward is that its weight compresses the rock vertically. Since the rock is incompressible, a vertical compression occurs if it is compensated by simultaneous horizontal extension. Therefore, it is this horizontal extensional strain which constitutes the movement of the thrust slice. Combining this strain with the vertical compressive strain requires that the horizontal normal stress is less than the vertical normal stress; consequently, the shear stress at the base of the thrust is less than that given by equation 3:40 (Ramberg, 1977a). Also, equation 3:40 does not account for the strain which occurs throughout the bulk of the thrusts due to gravitational collapse. The actual horizontal displacement of the thrust is a consequence of this strain which dissipates energy throughout the entire thrust and hence equation 3:40 is generally invalid for estimates of the shear stress.

Figure 3:10 - Foreland belt moving horizontally in response to gravity collapse. The lines at the bottom of the block represents increasing displacement towards the front of the belt (after Ramberg, 1977).

Ramberg (1977a) has proposed an alternative gravity spreading model which he termed the 'collapse model', where there is a large shear strain component parallel to the base within the lower section of the thrust block (Figure 3:10). The thrust slice thins as it spreads, which increases the length/thickness ratio (r) and consequently, in time, decreases the spreading rate. However, since both proposals are suggesting thrust slice thinning it would be expected that the faults developed within the belt would be of the "extensive type"; it would also be the case that these faults would form continuously throughout the belt through time. Even though the model proposed by Ramberg (1977a) seems to work successfully for the Scandinavian Caledonides, it requires modification to be a valid hypothesis for individual thrust slice movement within the Rocky Mountains. Thrust faults in the Caledonides show tectonic thinning whereas within the Rocky Mountains we have described tectonic thickening, supracrustal shortening, and eastward migration of deformation (Section 2.6). Therefore, any model for thrust fault movement in the Foreland of the Canadian Cordillera must consider not only gravity collapse as proposed by Ramberg or the 'glacier analogy' proposed by Elliott, but also active horizontal compressive forces.

CHAPTER 4

A MODEL FOR THE MECHANICS OF THIN SKINNED THRUSTING IN
THE ROCKY MOUNTAIN FORELAND BELT4.1 The Compressive-Spreading Wedge Model

As described in Section 2.6, the foreland of the Canadian Cordillera shows thin-skinned tectonics (the basement was not involved in thrusting). It is also evident from palinspastic reconstructions of the foreland that tectonic thickening has occurred due to an overall shortening of the belt by over 200 km (Price and Mountjoy, 1970). Therefore, from the proposed underthrusting model for the evolution of the foreland belt (Section 2.6), it would seem that a viable hypothesis for thrust fault formation would be the 'push from behind' model proposed by Hafner (1951) (Section 3.2). This 'push from behind' would simulate the tectonic force originating from the convergence of the core zone (Omineca Crystalline Belt) and the Foreland. However, in Section 3.4 we have concluded that a purely elastic model is unrealistic, and furthermore that any model proposed for the Rocky Mountains must combine both gravity spreading and horizontal compressive forces.

A 'horizontally compressive-gravity spreading' model (which we will call the compressive-spreading wedge model) has been suggested by Ramberg (1977a) and quantitatively elaborated by Chapple (1978) for purely plastic rheology. The foreland, of triangular shape in cross-section, is envisaged to be acted upon by a horizontal compressive force along its rear edge (in the case of the Rocky Mountains this is due to the convergence of the foreland's supra-structure with the Shuswap metamorphic complex). This compressive force is believed to have caused the tectonic shortening of the sedimentary pile which in turn originated a surface slope (Figure 4:1).

Chapple (1978) proposed a similar model (citing the Appalachians) in

Figure 4:1 - Formation of the Rocky Mountain Foreland Belt;

(A) Miogeoclinal wedge; (B) Relative movement of the craton westward, uplift of the Omineca Belt, and horizontal compressive forces causing thrusting as the miogeoclinal wedge is buttressed against the Shuswap Metamorphic Complex; (C) Continued relative movement of the craton; gravity spreading predominates as the major thrust fault generating mechanism due to formation of a surface slope from horizontal compressive forces; (D) Sketch showing present Foreland Fold and Thrust Belt.

Sections not to scale

which a weak basal layer is present and as a consequence the required surface slope is reduced to geologically realistic values. However, there is no evidence of a weak basal layer such as shale, salt or anhydrite in the Rocky Mountains, so that in order to use this model, modifications have to be made. If we do without the assumption of a weak basal layer, Chapple's (1978) model seems to fit rather well the boundary conditions as shown from the geology and structure.

The basic assumption made in the 'compressive-spreading wedge' model is to treat the belt as a homogeneous rigid-plastic body (see Section 3.3). Consequently, the model can be described in terms of an average stress and strain rate. In the Rocky Mountains, although the displacement along a single thrust may substantially change over a short distance, the cumulative displacement across a group of thrusts appears to remain relatively constant (Price and Mountjoy, 1970), and all the individual thrust faults are small when compared to the total area of the Foreland Belt. They end along strike and surrounding rocks are physically continuous around the ends of all of the thrust faults. Thus there has not been a megascopic loss of cohesion in the mass of rocks during thrusting, and the assumption of average stress and strain rates can be justified.

The model proposed by Chapple (1978) is shown in figure 4:2. A wedge-shaped region with surface slope α and basal slope θ_0 is considered in polar coordinates. At the surface of the wedge the normal stress (σ_θ) and the shear stress (σ_r) vanish, and at the base of the wedge the shear stress is constant and equal to the yield stress of the basal layer. The normal velocity component (v_θ) vanishes at the base. However, the problem is overconstrained. Since the surface slope contributes to the stresses within the wedge, it is impossible to specify simultaneously that the material in the wedge is yielding throughout and that the surface slope has some prescribed value. Therefore, if we choose

Figure 4:2 - Coordinates, parameters, and positive stresses for the 'compressive-spreading wedge' model (from Chapple, 1978). Magnified element shows convention for positive stresses; inequalities show sign of stresses in the Foreland Belt.

a wedge with a horizontal surface which is initially stress free and require it to be yielding in compressive flow, we may interpret the required normal stress along the top of the wedge in terms of the topographic slope needed for flow to occur. We must also note that since there is no shear stress on the horizontal surface, the material above this does not participate in compressive flow; hence this model will result in an overestimation of the required topographic slope.

The solution sought is similar to that of Hill (1950, p. 209), for the problem of flow through a converging channel. In this case (Figure 4:3) it is assumed that the sides of the channel are rough and that the frictional stress at the boundaries is constant. In the case of the Foreland Thrust Belt the channel is long and the flow of the material is assumed to be steady. Hence, the state of the stress in the material which is remote from the ends is effectively independent of conditions at the end of the wedge, and the directions of the slip lines depends only on θ and not on r . The converging channel problem is a variation of the problem of two converging parallel plates squeezing out plastic material, as described in Section 3.4.

Rederiving the solutions of Hill (1950) and Chapple (1978) it can be seen that the yield criterion is satisfied if we introduce the parameter Ψ (which depends on θ , thus $\Psi = \Psi(\theta)$) such that

$$\sigma_{r\theta} = K \sin 2\Psi \quad (4:1)$$

and

$$\sigma_r - \sigma_\theta = -2K \cos 2\Psi \quad (4:2)$$

where K is the yield stress and Ψ (which ranges between $+\frac{1}{2}\pi$ to $-\frac{1}{2}\pi$, and is zero when coincident with the radius vector) is the angle between a radius vector and the direction of the algebraically greater principal stress (since Ψ has the same sign as θ , Ψ is always positive). Compression ($\sigma_r - \sigma_\theta$) is negative, and the magnitude of the radial compression is greater than the

Figure 4:3 - Stress components for analysis of the flow of a plastic mass through a converging wedge-shaped channel (after Hill, 1950).

tangential compression.

From plasticity theory the equilibrium equations (equations 3:23) become, in polar coordinates and including body forces

$$r \frac{\partial \sigma_r}{\partial r} + \frac{\partial \sigma_{r\theta}}{\partial \theta} + \sigma_r - \sigma_\theta + \rho g r \sin \theta = 0 \quad (4:3)$$

$$r \frac{\partial \sigma_{r\theta}}{\partial r} + \frac{\partial \sigma_\theta}{\partial \theta} + 2\sigma_{r\theta} + \rho g r \cos \theta = 0 \quad (4:4)$$

Combining equation 4:1 with 4:3 Hill (1950) obtains

$$\frac{\sigma_\theta}{2K} = - \int \sin 2\psi d\theta + f(r) - \rho g \frac{r \sin \theta}{2K} \quad (4:5)$$

similarly combining 4:2 with 4:4 gives

$$\frac{\sigma_r}{2K} = - \int \sin 2\psi d\theta + f(r) - \cos 2\psi - \rho g \frac{r \sin \theta}{2K} \quad (4:6)$$

Taking the partial derivatives of equation 4:5 with respect to r and of equation 4:1 with respect to θ and then substituting into equation 4:3 we obtain Hill's solution for the functions $f(r)$ and $\psi(\theta)$

$$r f'(r) + \cos 2\psi (\psi'(\theta) - 1) = 0 \quad (4:7)$$

Since the first term of equation 4:7 is a function of r only and the second term is a function of θ only, it follows that

$$r f'(r) = C \quad (4:8)$$

and

$$\psi'(\theta) = C \sec 2\psi + 1 \quad (4:9)$$

where C is an integration constant. Thus from Hill (1950) it follows that

$$f(r) = -C \ln r + A \quad (4:10)$$

and

$$\theta = \int_0^\psi \frac{\cos 2\psi d\psi}{C + \cos 2\psi} \quad (4:11)$$

where A is a second integration constant.

Therefore, integrating equation 4:11 we obtain (see Dwight, 1966, p. 105)

$$\theta = \psi - \frac{C}{(C^2 - 1)^{1/2}} \tan^{-1} \left[\frac{[C - 1]^{1/2}}{[C + 1]^{1/2}} \tan \psi \right], \quad C > 1 \quad (4:12)$$

and

$$\theta = \psi - \frac{C}{(1 - C^2)^{1/2}} \tanh^{-1} \left[\frac{[1 - C]^{1/2}}{[1 + C]^{1/2}} \tan \psi \right], \quad -1 < C < 1 \quad (4:13)$$

To obtain the stresses, equation 4:8 can be used to change the integration variable in equations 4:4 and 4:5 from θ to ψ , which will give (Chapple, 1978)

$$\frac{\sigma_\theta}{2K} = -C \ln r + A + \frac{1}{2}(C + \cos 2\psi) - \frac{1}{2} C \ln(C + \cos 2\psi) - \frac{(\rho g r \sin \theta)}{2K} \quad (4:14)$$

$$\frac{\sigma_r}{2K} = -C \ln r + A + \frac{1}{2}(C - \cos 2\psi) - \frac{1}{2} C \ln(C + \cos 2\psi) - \frac{(\rho g r \sin \theta)}{2K} \quad (4:15)$$

The shear stress ($\sigma_{r\theta}$) at the top of the wedge ($\theta = 0$) is zero and the appropriate boundary conditions at the bottom surface ($\theta = \theta_0$) are a constant shear stress

$$\sigma_{r\theta} = \chi K \quad (4:16)$$

and a zero normal velocity (v_θ), where χ (the yield stress ratio) depends on the contrast in the yield stress of the basal layer to that of the entire wedge. Hence, a basal salt, anhydrite or shale will give a low χ . In the case of the Rocky Mountains (as described in Section 2.4) the basal sequences are limestones and sandstones which appear to be similar to the lithologies of the wedge as a whole. In this case, a yield stress ratio of $\chi = 1.0$ must be used.

Combining equations 4:1 and 4:16 we obtain

$$2\psi_0 = 2\psi(\theta_0) = \sin^{-1} \chi \quad (4:17)$$

along the base of the wedge. Therefore if $\chi = 1.0$ in the foreland ψ_0 must be approximately 45° , which implies that the thrust faults must be tangent to the basement at θ_0 . From seismic sections obtained by Bally *et al* (1966) it can be

seen that this is in fact the case, thus supporting the conclusion that $\chi = 1.0$ for the Rocky Mountains.

Next, following Chapple's (1978) argument, it can be shown that a velocity distribution which is consistent with the stresses and boundary conditions exists. The velocity distribution must satisfy the incompressibility condition, and give parallel axes of stress and strain rate

$$\frac{\sigma_r - \sigma_\theta}{2\sigma_{r\theta}} = \frac{\dot{\epsilon}_r - \dot{\epsilon}_\theta}{2\dot{\epsilon}_{r\theta}} \quad (4:18)$$

and reduce to zero normal velocity on the bottom boundary. A velocity distribution which satisfies these conditions is

$$v_r = -Br (C + \cos 2\Psi) \quad (4:19)$$

$$v_\theta = -Br (\sin 2\Psi - \chi) \quad (4:20)$$

where B is a positive constant matching the desired rate of tectonic shortening. Equations 4:19 and 4:20 represent a velocity distribution which corresponds to a constant shortening rate ($\dot{\epsilon}_r$) and a vertical velocity component (v_θ) which will increase toward the thick end of the wedge.

4.2 Interpretation of the Model

In order to solve for the above stress distribution (equations 4:14, 4:15) and velocity (equations 4:19, 4:20), the constants of integration C and A must first be determined. Since Ψ has the same sign as θ , C must be positive; it was also determined that C is greater than 1 (by trial and error); therefore it can be determined from equation 4:12. The constant A could in principle be determined using the fact that at the surface (from equation 4:14) the following relation is valid

$$\sigma_\theta = -2KC \ln r + K [2A + C + 1 - C \ln(C + 1)] \quad (4:21)$$

and $\sigma_\theta = 0$ for $r = r_0$ (at the front of the foreland belt). It follows, then, that constant A depends on the horizontal distance (r_0) from the origin of our

coordinate system to the eastern limit of thrusting. Since r_0 varies during deformation, we determine stresses down to a parameter A which does not affect stress trajectories nor the stress-difference.

Equation 4:21 specifies that the normal stress imposed on the horizontal top boundary of the wedge must become more compressive toward the thick end of the wedge. This is interpreted in terms of a topographic surface sloping toward the direction of movement (toward the craton). Hence, equation 4:14 defines the necessary surface topography needed for shortening and thrusting of the Foreland Belt in terms of the yield stress of the material in the wedge (Chapple, 1978).

By taking the derivative of equation 4:21 with respect to r it can be seen that the constant C is related to the surface topography

$$\frac{\partial \sigma_{\theta}}{\partial r} = -\frac{2KC}{r} \quad (4:22)$$

Also, in terms of the topographic slope,

$$\frac{\partial \sigma_{\theta}}{\partial r} = -\rho g \tan \alpha \quad (4:23)$$

where ρ is the density of the material and α is the surface slope. Therefore, C is related to the surface slope by the equation

$$C = \frac{r \rho g \tan \alpha}{2K} \quad (4:24)$$

Rearranging equation 4:24, the required topographic surface slope needed for thrust faults to occur can be determined for a given C from

$$\alpha = \tan^{-1} \left[\frac{2KC}{r \rho g} \right] \quad (4:25)$$

From equation 4:25 it can be seen that α is a function of r , $\alpha(r)$. In order to obtain the average surface slope, which will be of use in our analysis, we must eliminate r from equation 4:25.

Chapple (1978) presents a solution for the case in which both θ_0 and χ are small. We now relax the latter assumption. Consider equation 4:3

$$\frac{\partial \sigma_r}{\partial r} + \frac{1}{r} \frac{\partial \sigma_{r\theta}}{\partial \theta} + \frac{\sigma_r - \sigma_\theta}{r} + \rho g \sin \theta = 0$$

From equations 4:14 and 4:15 it can be shown that both σ_θ and σ_r have the same dependence on r

$$\frac{\partial \sigma_\theta}{\partial r} = -\frac{2KC}{r} - \rho g \sin \theta = \frac{\partial \sigma_r}{\partial r} = -\rho g \tan \alpha - \rho g \sin \theta \quad (4:26)$$

Therefore, with reference to figure 4:4

$$\frac{\Delta \sigma_r}{\Delta r} = \frac{\Delta \sigma_\theta}{\Delta r} = -\rho g \frac{\Delta H}{\Delta r} - \rho g \sin \theta \quad (4:27)$$

where ΔH is the required surface topography at the near end of the thrust belt.

From equations 4:2 and 4:27 we obtain

$$\frac{\sigma_r - \sigma_\theta}{r} = \frac{-2K \cos \psi \tan \theta_0}{h} \quad (4:28)$$

Since $\sigma_{r\theta} = 0$ at $\theta = 0$ and $\sigma_{r\theta} = \chi K$ at $\theta = \theta_0$ then

$$\frac{\Delta \sigma_{r\theta}}{r \Delta \theta} = \frac{\chi K}{r \theta_0} \approx \frac{\chi K}{h_0} \quad (4:29)$$

where h_0 is the average height of the area involved in thrusting. If we now combine equations 4:2 with 4:26, 4:28 and 4:29, multiply through by h_0 , and take into consideration that $\tan \theta_0 \approx \theta_0$ since θ_0 is always small, we obtain

$$\rho g h_0 \tan \alpha + 2K \theta_0 \cos 2\psi = \chi K \quad (4:30)$$

where the right hand side of the equation is the resistance to flow and on the left hand side are the driving force terms. The first term is the 'glacier spreading' force due to the topographic slope (as described in Section 3.4), while the second term is the driving force due to the wedge shape of the thrust belt. This second driving force is due to the fact that a horizontal compressive stress has a larger area to work over along the rearward edge of the

Figure 4:4 - (A) Geometric variables for area within wedge involved in thrust faulting; (B) Stresses acting on element (after Chapple, 1978).

element, hence even a constant stress across the element will give a greater force at the thicker end.

The average surface slope α can be obtained from equation 4:25 using the fact that $r = h_0 / \tan \theta_0 \approx h_0 / \theta_0$ since θ_0 is small; hence

$$\alpha = \tan^{-1} \left[\frac{2KC\theta_0}{\rho gh_0} \right] \quad (4:31)$$

Since in our case χ equals 1.0, ψ_0 will be 45° which will give a large C. This, in turn, means that in order to avoid an unrealistically high average surface slope, the strength of the whole wedge must be rather low. In this case, α will be lowered to within acceptable limits. Once this is accomplished we can then calculate the stress and velocity distributions which would have been present during thrust fault movement. From this we can then show the attitudes of the faults which should correspond to cross-sections of the actual thrusts in the Rocky Mountains.

4.3 Calculation Methods.

As has been described in the previous section, the model proposed by Chapple (1978) is a viable explanation of how deformation occurs in a thin skinned fold and thrust belt. In order for the required surface slope to be geologically realistic, we must be able to vary at least one of the variables on the right hand side of equation 4:31. This then means that K must be a low value since θ_0 , C and h_0 are fairly well known and ρ and g are constants.

The calculations used in this study were performed by transforming Hill's (1950) and Chapple's (1978) equations into a computer program, taking into account our modifications, since $\chi = 1.0$ and ψ is not small. The program can be broken down into five sections, each section using available data or educated guesses from the southern section of the Foreland Belt of the Canadian Cordillera. The data were obtained and fed into the program as follows:

- 1) The thrust belt was subdivided into six approximately equal segments (Figure 4:5); this was done so as to give a better approximation of α , since α will increase from west to east. This increase in α can be attributed to the decreasing h_0 term in equation 4:31.
- 2) The basal angle for the entire belt was next read into the program. This angle represents the average basement dip over the entire thrust belt determined from the depth to basement at both the rear and front edges of the Foreland Belt.
- 3) The values of $\Psi(\theta_0)$, corresponding to $\chi = 1.0$ through $\chi = 0.3$ at increments of 0.1, were calculated. This was done for comparison purposes only, and to provide an overall view of how the surface slope decreases as χ decreases in foreland belts.
- 4) The average thickness (h_0) for each segment within the thrust belt was read in. This value was determined from the basal angle (θ_0) and the horizontal distance to the centre of each section.
- 5) The next value to be read in was the overall length of the thrust belt (L_c). This length (together with all the above parameters) was determined for three times (Figure 4:5), as follows: pre-deformation length (Late Jurassic, 145 Ma); length of the thrust belt half-way through the belt's formation (Late Cretaceous, 95 Ma) and length of the belt at the termination of thrusting (Oligocene, 35 Ma). The predeformational data were obtained from palinspastic reconstructions of the Foreland Belt as determined by Price and Mountjoy (1970). This section has been constructed by using stratigraphic marker sequences and moving them back along the thrust faults. The information within the section is determined relative to a horizontal datum (the top of the Jurassic)

Figure 4:5 - Three different thrust belts used in the calculations;

(A) Predeformational wedge with only the last three segments involved in thrusting; (B) Half-time deformational wedge with only the last three segments involved in thrusting; (C) Postdeformational wedge where all segments were involved in thrusting.

Dashed lines indicate required surface slope for a given strength (100 bars in this case). Dot-dashed line on (C) indicates the present surface slope.

A

100 KM

B

Required Slope

Present Slope

C

Vertical Exaggeration 4 Times

which is effectively a sea level datum as it existed before the deposition of the clastic wedge (Section 4.4) and prior to any thrusting. The postdeformational section was constructed on recent cross-sections drawn through the Rocky Mountains, hence giving the length of the thrust belt as thrusting ended. The 'half time' length was simply determined by taking the length the belt would be by assuming a constant rate of tectonic shortening. As can be seen in figure 4:6, isostasy as prescribed by Price (1973) was taken into account. This was done due to the fact that as sediments were piled up from one thrust slice over-riding the sediments in front of it the basement would be depressed under the increased load, hence as deformation migrated from west to east the average basement angle increased from 1.86° to 3.00° .

- 6) The final data put into the program were the values for the overall strength of the material within the thrust belt. The program was run for different strengths (from 600 bars to 50 bars at 50 bar intervals) in order to determine the strength of the rocks required to give a geologically realistic surface slope.

With the input of the data, the program then determined the seven unknowns (C , α , ψ , σ_r , σ_θ , v_r , and v_θ) in five steps:

- 1) The determination of C was done using equation 4:12 and solving by iteration. A value was determined for each increment of χ ; hence eight values were obtained (see Appendix I); only the value of C corresponding to $\chi = 1$ was used in further calculations.
- 2) Alpha was determined next, using equation 4:31 and solving for each of the six sections within the thrust belt. This was computed for each of the above C 's with the values for $\chi = 1.0$ being

2 2
OF / DE

Figure 4:6 - Tectonic thickening of supracrustal rocks, isostasy, and the origin of a migrating foredeep trough. The relationships which are shown schematically are based on the assumption that, due to the flexural rigidity in the lithosphere, isostatic subsidence extends beyond the actual limits of the supracrustal load (after Price, 1973).

Undeformed Wedge of Supracrustal Rocks

Tectonically Thickened

Amount of Regional Isostatic Subsidence

the only values of C retained for further calculation. An average α was also determined for the entire belt for C corresponding to $\chi = 1.0$. This was done by taking the weighted average of the six sectional alphas.

- 3) The angle $\Psi(\theta)$ was determined using equation 4:12 (keeping C fixed) after the computation of α . A Ψ value was determined for each increment of θ (0.25°) from the surface ($\theta = 0$) to the basal slope (θ_0). This was only calculated for $\chi = 1.0$, hence $\Psi(\theta_0)$ was always 45° .
- 4) Using only the average surface slope for the belt (determined from step 2), with corresponding θ values and Ψ values, the radial and tangential stresses (equations 4:14 and 4:15) were determined. These stresses are given in terms of $\{(\sigma_\theta/2K) - A\}$ and $\{(\sigma_r/2K) - A\}$ as described in the previous section. Both the radial and tangential stresses were calculated along every θ line (see program calculation 3, above) which has a corresponding Ψ -value. This was done at 5 km intervals along each line, hence giving the stresses throughout the wedge.
- 5) The final calculation was that of the radial and tangential velocities (equations 4:19 and 4:20) which were determined using the same grid as was used in the determination of the radial and tangential stresses. The velocities were determined along each θ -line at 5 km intervals, thus giving the velocity distribution throughout the wedge. In this calculation the strain rate constant (B) was taken as unity; hence the velocities are only relative. It is also possible to determine the actual velocities by multiplying our calculated values by the estimated strain rate. This rate can be simply estimated by taking the amount of

horizontal shortening (approximately 60%) and dividing by the time involved in thrusting (110 Ma); hence B is approximately $1.8 \times 10^{-16} \text{ s}^{-1}$ for the Rocky Mountains.

Therefore, from the above program we can select a geologically realistic surface slope which will give the maximum strength of the material within the thrust belt during deformation. The fault planes can then be estimated from the maximum compressive stress trajectories obtained from the determination of Ψ at 5 km intervals (the thrust faults are drawn at 45° from the Ψ value along any one θ -line). We can also show the relative average particle velocities within the wedge from the final calculation of the program.

4.4 Application of the 'Compressive-Spreading Wedge' Model

As discussed in Section 2.4, it is evident from geologic and seismic cross-sections that the thrust faults within the southeastern Canadian Cordillera flatten at depth to become tangent to the basement rocks. This means, on the basis of Chapple's analysis and simple slip line theory (Section 3.3) that the trajectories of maximum compressive stress within the wedge will intersect the basement at an angle approximately equal to 45° . It therefore can be shown from equation 4:17 that χ must equal 1.0, which brings us back to our earlier conclusion that no weak basal layer exists in the Rocky Mountains.

The above conclusion (of no weak basal layer) is of course not valid for every foreland fold and thrust belt. In Chapple's study, for example, the Pennsylvania Plateau, and the Valley and Ridge provinces of the American Appalachians were used to support the presence of a low χ value. Within the Pennsylvania Plateau a weak slate layer is present at the base, and in the Valley and Ridge a weak shale is envisaged as a décollement zone. Also, within the Jura Mountains Laubscher (1971) describes a 'weak basal cushion' which

was also described by Lemoine (1973) and others as a 'basal lubricating horizon' of evaporite. The presence of this zone causes a large contrast in the strength of the materials in the basal layer and the belt as a whole. Also with the structural cross-sections of the Jura it can be seen that thrust faults intersect the basement at high angles (Laubscher, 1971) thus suggesting a low χ value. There are also various other examples of a weak basal sequence in parts of other foreland belts; hence in dealing with the Canadian Rocky Mountains we may not be dealing with the most typical case.

Since a weak basal layer is not present in the Rocky Mountains, we must return to the problem of the strength of the material involved in thrusting, which will enable the surface slope to be geologically realistic. From the data shown in figures 4:7, 4:8 and 4:9, results were obtained for a thrust belt having strengths of 600, 300, 150, 125, 100, and 50 bars. The results are shown in Tables 4:1, 4:2 and 4:3. As can be seen, a geologically realistic surface slope cannot be expected if the overall strength of the material within the Foreland Belt is greater than 150 bars; a value of 100 bars is probably a good estimate. For this value, the average surface slope ranges between 2.1° and 2.6° for the three time periods. However, it must be remembered that the calculated average surface slope is an overestimation; thus, the material strength may be slightly greater than 100 bars.

As can be seen in Tables 4:1, 4:2 and 4:3, the increase in elevation (ΔH) at the western end of the thrust belt is within acceptable limits (a few kilometres), with a material strength of approximately 100 bars. However, it must be noted that the presence of a surface slope also causes our material strength value to be an underestimation. This is due to the fact that in the model the mass beneath the horizontal datum is the only material which is considered to be involved in thrusting; thus, in equation 4:31, h_0 is measured from the datum to the top of the basement. Since the creation of a surface

$$O = 1.86$$

$$L_1 = 1.83 \text{ km}$$

$$L_2 = 14.00 \text{ km}$$

SECTION	WIDTH (A)	h
1	60 km	2.80 km
2	60 km	4.75 km
3	60 km	6.70 km
4	60 km	8.65 km
5	60 km	10.60 km
6	75 km	12.80 km

(B)

AVERAGE	375 km	N/A

Figure 4:7

HALF-TIME DEFORMATION

$$\Theta = 2.20$$

$$L_1 = 3.40 \text{ km}$$

$$L_2 = 13.40 \text{ km}$$

SECTION	WIDTH (A)	h_o
1	40 km	4.17 km
2	40 km	5.70 km
3	40 km	7.24 km
4	40 km	8.78 km
5	40 km	10.30 km
6	60 km	12.20 km

(B)

AVERAGE	260 km	N/A

Figure 4:8

POSTDEFORMATION

$$\Theta = 2.99$$

$$L_1 = 4.96 \text{ km}$$

$$L_2 = 12.79 \text{ km}$$

SECTION	WIDTH (A)	h_0
1	25 km	5.60 km
2	25 km	6.92 km
3	25 km	8.22 km
4	25 km	9.53 km
5	25 km	10.83 km
6	25 km	12.14 km

(B)

AVERAGE	150 km	N/A

Figure 4:9

Table 4:1

PREDEFORATION

SECTION	h_0 (km)	L (km)	K (bars)	α
1	2.80	60	600	39.70
			300	22.54
			150	11.73
			125	9.81
			100	7.88
			50	3.96
2	4.75	60	600	26.08
			300	13.75
			150	6.98
			125	5.82
			100	4.66
			50	2.34
3	6.70	60	600	19.13
			300	9.84
			150	4.96
			125	4.13
			100	3.31
			50	1.66
$\Delta H(m)$				
AVERAGE	—	375	600	12.40
			300	6.28
			150	3.15
			125	2.62
			100	2.10
			50	1.05

SECTION	h_0 (km)	L (km)	K (bars)	α
4	8.65	60	600	15.04
			300	7.65
			150	3.84
			125	3.20
			100	2.56
			50	1.28
5	10.60	60	600	12.37
			300	6.26
			150	3.14
			125	2.62
			100	2.09
			50	1.05
6	12.80	75	600	10.30
			300	5.19
			150	2.60
			125	2.17
			100	1.74
			50	0.87

$C = 14.62$

$\chi = 1.00$

$\Theta_0 = 1.86$

$L_1 = 195 \text{ km}$

$$C = 14.62$$

$$\chi = 1.00$$

$$\Theta_0 = 1.86$$

$$L_1 = 195 \text{ km}$$

Table 4:2

HALF-TIME DEFORMATION

SECTION	h_0 (km)	L (km)	K (bars)	α
1	4.17	40	600	28.90
			300	15.45
			150	7.86
			125	6.56
			100	5.26
2	5.70	40	50	2.63
			600	21.99
			300	11.42
			150	5.77
			125	4.81
3	7.24	40	100	3.85
			50	1.93
			600	17.64
			300	9.03
			150	4.54
4	8.78	40	125	3.79
			100	3.03
			50	1.57
			600	12.36
			300	6.26
5	10.30	40	150	3.14
			125	2.62
			100	2.09
			50	1.05
			600	2559
6	12.20	60	30674	
			15367	
			7675	
			6396	
			5117	

SECTION	h_0 (km)	L (km)	K (bars)	α
4	8.78	40	600	14.69
			300	7.47
			150	3.75
			125	3.13
			100	2.50
5	10.30	40	50	1.25
			600	12.59
			300	6.37
			150	3.17
			125	2.66
6	12.20	60	100	2.13
			50	1.07
			600	10.65
			300	5.37
			150	2.69

SECTION	h_0 (km)	L (km)	K (bars)	α
4	8.78	40	125	2.24
			100	1.80
			50	0.90
			600	
			300	
5	10.30	40	150	
			125	
			100	
			50	
			600	
6	12.20	60	300	
			150	
			125	
			100	
			50	

$C = 12.24$
 $\chi = 1.00$
 $\Theta_0 = 2.20$
 $L_1 = 140$ km

AVERAGE	-	260	

Table 4:3

POSTDEFORMATION

SECTION	h_o (km)	L (km)	K(bars)	α
1	5.60	25	600	21.88
			300	11.36
			150	5.73
			125	4.78
			100	3.83
			50	1.92
2	6.92	25	600	18.01
			300	9.23
			150	4.65
			125	3.87
			100	3.10
			50	1.55
3	8.22	25	600	15.30
			300	7.79
			150	3.91
			125	3.26
			100	2.61
			50	1.31
$\Delta H(m)$				
AVERAGE	-	150	600	15.12
			300	7.72
			150	3.88
			125	3.24
			100	2.59
			50	1.30

SECTION	h_o (km)	L (km)	K(bars)	α
4	9.53	25	600	13.28
			300	6.73
			150	3.38
			125	2.82
			100	2.25
			50	1.13
5	10.83	25	600	11.73
			300	5.93
			150	2.97
			125	2.48
			100	1.98
			50	0.99
6	12.14	25	600	10.50
			300	5.29
			150	2.65
			125	2.21
			100	1.77
			50	0.88

$C = 8.80$
 $\chi = 1.00$
 $\Theta = 2.99$
 $L_i = 150 \text{ km}$

$$C = 8.80$$

$$\chi = 1.00$$

$$\Theta_o = 2.99$$

$$L_i = 150 \text{ km}$$

slope implies that material must be above this horizontal datum and from the fact that this material will be involved in deformation, h_0 should be a larger value. Therefore increasing h_0 in equation 4:31 reduces the required surface slope or alternatively increases the material strength (if a specific slope is chosen). This is another factor which qualifies the surface elevation at the rear of the wedge as an overestimate.

The magnitude of ΔH was obtained for the three different times on the basis of three assumptions: 1) the only section of the Rocky Mountains involved in initial thrusting was the Main Ranges; 2) only the Main Ranges and parts of the Front Ranges were involved in thrusting in the 'half time' calculations; and 3) the Main Ranges, Front Ranges and the Foothills were involved in the final stage of thrusting. Therefore, in the Rocky Mountains the width of the deforming belt involved at any one of the three times was approximately ($\pm$ 25 km), the same as the width of the thrust belt at the present time (150 km). This length was chosen since the original width of the Main Ranges is approximately 150 km in the palinspastic reconstruction of Price and Mountjoy (1970). Similarly, if we take the palinspastic width of the Main Ranges and Front Ranges and the present width of the two belts and average them, we obtain the approximate 'half time' result of 150 km involved in thrusting, assuming a constant rate of supracrustal shortening. The actual values in the program were: an active length of 195 km for initial thrusting (this length also estimates the additional length involved due to deformation in the eastern Omineca Belt, see Section 3.6); 140 km for 'half time' deformation and 150 km for the final stages of thrust fault movement.

It must also be noted that this system is dynamic. Therefore the above values are only estimates of the extent of the deforming zone. From geological and structural evidence (Section 2.5), it has been concluded that deformation has migrated from west to east. If we assume a constant rate of supracrustal

shortening, the area which has been deformed will vary with time, but the actual area being deformed at any given time will only be some portion of this. Since our estimated width involved in thrusting only deals with three specific times, the value of ΔH is an overestimation when we assume that the entire width (≈ 150 km) at that time is involved in deformation. For instance, if we assume an average surface slope of 2.5° (which corresponds to a K of 100 bars), ΔH will have a value of approximately 6500 metres if the whole width of 150 km is simultaneously involved in deformation, but a value of only about 3300 metres if the width involved is 75 km. It must also be noted that erosion will tend to 'level' the rear of the wedge; hence a sufficient surface slope will only be present in a narrow zone at the front of the deforming sections.

From the computation of C , α , and Ψ 's corresponding to lines of equal θ -values, we can calculate the radial and tangential stresses; these are shown qualitatively along the base and the rear of the thrust belt in figures 4:10, 4:11 and 4:12. These figures also show the lines of maximum shear stress (slip lines), which give an indication of where thrust faults would be expected to form. The lines were drawn using the maximum stress trajectories and converting to the maximum shear stress directions (45° from the trajectories). Also, after the determination of Ψ , values for particle velocities were calculated (Figures 4:13, 4:14, and 4:15). Here, the magnitudes of the velocities are relative as the magnitude of the velocity vectors depends on the externally imposed velocity at the rearward boundary of the wedge. The values obtained can be easily converted to actual particle velocities by multiplying through to the rate of tectonic shortening (assumed constant) which in the case of the Rocky Mountains was $1.8 \times 10^{-16} \text{ s}^{-1}$ (see Section 4.2).

As can be seen from figures 4:10, 4:11 and 4:12, the results closely resemble what would be expected and is observed within the Rocky Mountain Foreland Thrust Belt. The thrust faults have the same characteristic trend

LATE JURASSIC [145Ma]

Figure 4:10 - Thrust faults, ΔH and surface slopes predicted from the 'Predeformational' data using the compressive-spreading wedge model with an overall strength of 100 bars. The stresses shown are not drawn to scale.

Figure 4:11 - Thrust faults, ΔH and surface slopes predicted from the 'Half-time deformational' data using compressive-spreading wedge model with an overall strength of 100 bars. The stresses shown are not drawn to scale.

OLIGOCENE (35 Ma)

Figure 4:12 - Thrust faults, ΔH and surface slopes predicted from the 'Postdeformational' data using the compressive-spreading wedge model with an overall strength of 100 bars. The stresses shown are not drawn to scale.

LATE JURASSIC

[145 Ma]

Figure 4:13.- Relative velocity vectors predicted from Predeformational data using the compressive-spreading wedge model with an overall strength of 100 bars. The velocity vectors for the wedge can be determined by taking the approximate rate of tectonic shortening within the Rocky Mountains to be $1.8 \times 10^{-16} \text{ sec}^{-1}$.

UPPER CRETACEOUS

[95Ma]

Figure 4:14 - Relative velocity vectors predicted from the Half-time deformational data using the compressive-spreading wedge model with an overall strength of 100 bars. The velocity vectors for the wedge can be determined by taking the approximate rate of tectonic shortening within the Rocky Mountains to be 1.8×10^{-16} sec⁻¹.

Figure 4:15 - Relative velocity vectors predicted from the Postdeformational data using the compressive-spreading wedge model with an overall strength of 100 bars. The velocity vectors for the wedge can be determined by taking the approximate rate of tectonic shortening within the Rocky Mountains to be $1.8 \times 10^{-16} \text{ sec}^{-1}$.

and slope as the actual thrusts (note the vertical exaggeration). The model obviously cannot generate the folded thrusts (and refolded structures) which are present in the western section of the thrust belt, nor simulate anisotropy (layering effect). The average relative particle velocity distribution also shows that the material was 'pushed' and 'flowed' up within the foreland and away from the core zone, with the greatest velocities being at the upper rear surface and decreasing with depth and distance toward the craton (eastwards). Therefore, since the top of the wedge moves more rapidly than the bottom, in a real situation there will be an appreciable amount of layer-parallel shear and asymmetric structures (folds) near the base of the wedge. These are evident within the Rocky Mountains, as asymmetric folds are dominant beneath the Main Ranges with bedding plane or layer-conforming thrust faults being present in the Front Ranges and Foothills.

In conclusion, the Foreland Thrust Belt could have formed according to the specifications of the 'compressive-spreading wedge' model if the overall strength of the wedge was approximately 100 bars (see Chapter 5). This strength would give an average surface slope of $2.0-2.5^\circ$, which in turn would give an upper estimate of the elevation at the rear of the wedge of about 7km. This value is acceptable, given the various factors that contribute to making it an overestimation.

CHAPTER 5

THE UPPER LIMIT OF STRESSES AND DEFORMATION MECHANISMS ALONG
THRUST FAULTS5.1 General Remarks

Until now we have dealt with the mechanical properties and the tectonic evolution of the southern Rocky Mountain Foreland Belt as a whole. At this scale of observation, the deformation consists of the overall change in the shape of the northeasterly tapering wedge of supracrustal rock. Deformation on a regional scale can best be described as a type of penetrative plastic flow (Price, 1973). If the scale of observation is reduced to the scale of individual thrust faults, it can be seen that deformation is a kind of inhomogeneous ductile shear failure, which is described by Price (1973) as a combination of penetrative plastic flow and relatively brittle faulting. At this scale the displacement along any one thrust fault differs along strike; the strain is taken up by an adjacent thrust and surrounding rocks so that the total deformation distributed over the belt as a whole is relatively continuous. Along any individual thrust, however, deformation is concentrated and discontinuous: essentially unstrained sediments have remained intact while undergoing large-scale translations on a distinct shear surface. We shall now briefly investigate the mechanics of structures at this scale of observation. Obviously, these features cannot be modelled in terms of a rigid-purely plastic rheology, although this has proven to be an adequate model for the Foreland Belt as a whole. We precede this discussion with some additional remarks on the stress levels to be expected during thrusting.

5.2 Upper Limit on Stress Along Thrusts

Elliott (1976a) suggested that the upper limit of the shear strength of the material in the Rocky Mountains is about 200 bars (see Section 3.4). This value was obtained on the basis of work done by Brune et al (1969) and Brune (1974), where heat produced during faulting was used to estimate the strength of the rock involved. In these studies, the lack of an observable heat flow anomaly along the San Andreas fault was used to give an upper limit of approximately 200 bars for the average stress in the upper 20 km of the crust along the fault averaged over geologic time. Elliott (1976a) cites seismic evidence in the eastern Andes (which are believed to be an active foreland belt) which suggests that the time-averaged stress along thrust faults is lower than along transform faults, since seismic activity within the eastern Andes is much less than that along a transform fault. Therefore, the upper limit of 200 bars should be valid for thrust faults.

Work by Bird in the Zagros Mountains (1978a) and in the Himalayas (1978b) also indicates a relatively low upper limit of shear stress along thrust faults. Within both orogenic belts there is an analogous 'type' of mechanism deforming the belt as has been proposed for the Rocky Mountains (however, in the Rocky Mountains deformation occurs in a supracrustal wedge while in the Zagros and Himalayas the entire crust is considered): this is the buttressing of one segment of continental crust (where the thrust faults begin) against a second crustal segment as a result of the lower part of the former crustal slab being underthrust beneath the latter.

Bird (1978a, 1978b) obtained results using both a 'rock mechanics' argument and a 'topographic' argument for the maximum shear stress along the basal thrust. Based on the rock mechanics argument, which states that the horizontal integral of the shear stress is limited by the vertical integral of the stress differences that the trailing edge of a thrust sheet can support, a value of less

than 300 bars was extrapolated from laboratory results for the average shear stress along the basal thrust within the Himalayas. Using the topographic argument, which is similar to the model proposed by Elliott (1976a), a value of 200 bars was obtained for the Himalayas and approximately 75 to 100 bars for the Zagros Mountains. Therefore the value of 100 bars for the overall strength of the wedge in the Rocky Mountains is not an unrealistic estimate when it is considered that for the Himalayas and the Zagros Mountains, the entire crust is involved while in the Rocky Mountains we deal only with a relatively thin supracrustal wedge.

Also, it could be mentioned here that Itô (1978), using granite slabs to simulate the crust in real-time experiments, determined that the strength of the crust can be as low as 10 bars in shear provided the shearing strain rate is less than $10^{-15} \text{ sec}^{-1}$. The strength increases to 100 bars for a shear strain rate of $10^{-14} \text{ sec}^{-1}$.

5.3 Deformation Mechanisms Along Thrust Faults

There are three known mechanisms by which an individual thrust fault can move: 1) Frictional sliding along discrete fault surfaces, with or without the aid of pore fluid pressure; 2) pressure solution slip along a discrete fault surface; and 3) deformation within a narrow shear zone. The first mechanism (frictional sliding) has been proposed by numerous authors (such as Hubbert and Rubey, 1959), and suggests that the slip associated with thrusting obeys the linear law of sliding friction between solids. Therefore, slip cannot occur unless the shearing stress along the sliding surface reaches the value of yielding determined by the cohesive (interlocking) strength, the coefficient of sliding friction, and the normal stress on the slip surface corrected for pore pressure.

The second mechanism, 'pressure solution slip' (proposed by Elliott, 1976b)

is a diffusion-controlled sliding law where material is moved by diffusion around surface asperities along the fault. It is suggested that much of the diffusion along the sliding surfaces occurs within a discrete, hydrous film rather than by grain boundary diffusion. This is based on the observation of fibrous mineral growths which are often arranged into imbricate fibrous shingles or accretion steps (Elliott, 1976b). The process of pressure solution slip is independent of frictional sliding, although they can occur simultaneously.

Both of the above processes require thrust faults to slide along discrete surfaces; this discrete slip may be difficult to achieve, given the magnitude of the stresses in foreland belts and the rheological properties of the rock within the thrust sheet. Elliott (1976b), however, suggests that thrust faults are ductile faults and that local shear stress near the thrust can be higher than the regional value due to the stress concentration caused by failure.

The third process by which a thrust sheet can move is by concentrated deformation in a narrow (< 100 m, down to a thickness of 1m) shear zone (or low grade ductile shear zone). This process can occur by various mechanisms, the two most likely being: 1) superplasticity and 2) pressure solution. From a study done by Olsson (1974), it was determined that for calcitic rocks at temperatures of 25°C to 300°C, deforming by intragranular (dislocation) mechanisms the yield stress increases with decreasing grain size according to the Hall-Petch relation $\sigma_y = \sigma_0 + \kappa d^{-1/2}$, σ_y is the yield stress, d is the average grain size and κ and σ_0 are constants. If, however, the rock deforms superplastically (i.e., if diffusion-controlled grain boundary sliding predominates and accounts for up to 60-90% of the total strain) the situation is reversed, and the strain rate increases for decreasing grain size. This dependence is usually described by the relation $\dot{\epsilon} \propto d^{-b}$ at a constant stress where d is again the grain size and b is a parameter with values usually between 2 and 3 (Schmid et al, 1977). Since material in the narrow ductile shear zones is very fine grained (in the

order of 1-10 μm) it would be expected that superplasticity is a likely mechanism by which shear can be concentrated.

As an example, we consider the Glarus overthrust in the Swiss Alps (unfortunately, no comprehensive microrheological studies have been carried out in the Rocky Mountains). It has been shown by Schmid (1975) that an extremely ductile calcareous mylonite layer (the Lochseitenkalk), 1 to 2 metres thick, has taken up all the differential movement of the Glarus overthrust with respect to the substratum along the interface between two more or less rigid blocks. The amount of displacement along the thrust is estimated to be a minimum of 35 km, which occurred over a maximum time of 100 Ma. Assuming a 1 m thick calc-mylonite zone the local shear strain rate is approximately $10^{-10} \text{ sec}^{-1}$. However, comparison with experimental results on the Solenhofen limestone at relatively low temperature shows that rocks of this grain size ($\approx 10 \mu\text{m}$ for the Solenhofen and $1 \mu\text{m}$ for the Lochseiten limestones) are too strong (the required differential stress would be of the order of 1 kbar) to flow at such a fast strain rate, if the flow mechanism is dislocation creep. Work done by Briegel and Goetze (1978) on the Lochseiten limestone has shown that dislocation densities and other microstructural evidence are indicative of a very high differential stress (2 kbars). However, this microstructural evidence most likely records the final stage, and not necessarily the typical stage of overthrusting. Briegel and Goetze (1978) point out that experimental data for the Solenhofen limestone and Yule marble, extrapolated at $\dot{\epsilon} = 10^{-9} - 10^{-11} \text{ sec}^{-1}$, show that calcitic rocks soften remarkably at temperatures higher than 300°C . Therefore, if the temperature at the time of thrusting was in this range, the Lochseitenkalk may have acted as a 'lubricating layer' with a strength as low as 100 bars (Figure 5:1).

Superplastic flow was observed by Schmid et al (1977) in the Solenhofen limestone. In the superplastic regime grains remained approximately equant even

Figure 5:1 - Plastic flow of Yule marble and Solenhofen limestone shown by hatched bands. The limestones clearly soften rapidly in the range 300°-500°C. (Data and diagram after Briegel and Goetze, 1978.)

after a large deformation, which implies that there must have been a large component of grain boundary sliding. The flow can be described by the equation

$$\dot{\epsilon} = A_1 d^{-b} \exp(-E/RT) \sigma^n \quad (5:1)$$

where σ is in bars, $E = 50.9$ kcal/mole, $\log A_1 = 4.98$, $b = 3$, $n = 1.66$, and d = grain size (μm). As can be seen from figure 5:2 (plots of flow-regime maps, superplasticity in Solenhofen limestone occurs at relatively low stresses and the strain rate increases with decreasing grain size. At $10^{-10} \geq \dot{\epsilon} \geq 10^{-14} \text{ sec}^{-1}$ the superplastic flow regime extends to a grain size much larger than $10 \mu\text{m}$, and the expected flow stresses are $\sigma \leq 1300$ bars at $T = 400^\circ \text{C}$. However, σ is very sensitive to d , such that for a $d < 5 \mu\text{m}$ at 400°C and $\dot{\epsilon} = 10^{-10} \text{ sec}^{-1}$ the $\sigma \leq 200$ bars. Therefore, it can be suggested that superplastic flow may be widespread in mylonites, and that a reduction in grain size by syntectonic recrystallization can be envisaged as a preliminary development leading in the course of large scale deformations to a stable zone of weakness within a coarser grained country rock (Schmid et al, 1977).

Pressure solution (to be distinguished from pressure solution slip that was discussed in Section 5.3) is another process by which shear zones may develop. In this mechanism, which requires the presence of a fluid phase, grains undergo diffusive mass transfer as material is removed along grain boundaries perpendicular to the maximum compressive stress and is transferred through the fluid phase to boundaries perpendicular to the minimum stress. Therefore, grains are flattened as a chemical potential gradient for vacancy flow is set up (McClay, 1977).

The theoretical rate equation for pressure solution, as derived by Rutter (1976) for $\sigma < 300$ bars is

$$\dot{\epsilon} = 32\sigma v C_0 D_b w / RTpd \quad (5:2)$$

where: σ = applied stress
 v = molar volume of a solid
 C_0 = concentration of saturated solution in equilibrium with unstressed

Figure 5:2 - Flow regime maps at constant temperature (400°C) in stress-grain size coordinates. Thick lines are strain rate contours. (after Schmid et al, 1977)

solid
 D_b = grain boundary diffusivity
 w = effective grain boundary width
 T = absolute temperature
 ρ = density of the solid
 d = grain size

Deformation maps constructed by Rutter (1976) show that pressure solution is predominant for quartz and calcite at low temperatures ($T < 450^\circ\text{C}$ for calcite and $< 650^\circ\text{C}$ for quartz) and relatively low stresses ($\log \sigma [\text{bars}] < 2$ for calcite). McClay (1977) concluded that pressure solution in fine-grained quartz and calcite rocks can give rise to geological strain rates (10^{-9} - $10^{-11} \text{ sec}^{-1}$) at temperatures from 200° to 350°C . These results are summarized in figure 5:3 where the rate equations for pressure solution have been evaluated for quartz and calcite (using modified forms and the data of Rutter (1976) as interpreted by McClay, 1977). It is evident from figure 5:3 that for quartz, pressure solution gives strain rates typical of thrust zones (10^{-9} to $10^{-11} \text{ sec}^{-1}$) for grain sizes up to $100 \mu\text{m}$ and for calcite pressure solution is realistic over grain sizes ranging from approximately 1 to $10 \mu\text{m}$ (McClay, 1977).

From the above analysis it can be concluded that pressure solution and superplasticity are possible deformational mechanisms along thrust faults deforming at temperatures ranging between 200° and 350°C and at strain rates of the order of 10^{-9} to $10^{-10} \text{ sec}^{-1}$, if the relative motion is taken up by narrow shear zones of calcitic or quartzitic material of grain size up to $10 \mu\text{m}$. This result is of general validity and therefore could be applied to individual thrusts within the Foreland Belt where the relative motion has been taken up by narrow shear zones rather than along discrete surfaces. Unfortunately, there is insufficient information at present to reach a definite conclusion. If we take the McConnell thrust as an example, where a displacement of 40 km has taken place (Elliott, 1976b), the presence of a 1 m thick shear zone would give a strain rate of approximately $10^{-10} \text{ sec}^{-1}$. The temperature within

Figure 5:3 - Deformation mechanism plots of strain rate against grain size for quartz and calcite. The data used for the plots were taken from Rutter (1976) (after McClay, 1977).

the Rockies is estimated to have been in the range of 200°-350°C (Elliott, 1976b) during the time of thrusting. Therefore, from the data obtained in the Alps, it would seem that pressure solution and superplasticity are certainly mechanisms which could have been operational in the Rocky Mountains at the time of thrusting. When no finite shear zone is present, this explanation obviously fails; in this case, pressure solution slip seems to be the most viable model for displacement along discrete thrust planes, but more research must be carried out before definite conclusions can be reached.

CHAPTER 6

CONCLUSIONS

The main points of this thesis can be summarized as follows:

- 1) The southern Canadian Rocky Mountain Foreland Thrust Belt represents a typical foreland belt; compressive forces must have been operative along its rear edge, no matter how the supracrustal shortening was accomplished (i.e. by gravity spreading of the core zone or by underthrusting of the craton).
- 2) Classical elastic approaches, using Airy's stress function to analyse the deformation of foreland belts, do not give satisfactory results; plasticity theory and analogies with glacier flow lead to more realistic models.
- 3) A model for the deformation of foreland belts underlain by a weak basal layer has been proposed by Chapple (1978). We have modified this for the case where there is no marked difference in the rheological properties of the basal layer and of the belt as a whole. The model uses perfectly plastic rheology, and postulates that supracrustal shortening and thrusting are a consequence of tectonic forces acting at the rear of the belt and generating a surface slope toward its front.
- 4) Calculations for three specific times (Upper Jurassic (145 Ma), Upper Cretaceous (95 Ma) and the Oligocene (35 Ma)) show that the 'compressive-spreading wedge' model gives results that fit reasonably well with the observed geology and structure. The material strength of the wedge as a whole has to be approximately 100-150 bars in order to have a geologically realistic average surface slope and a reasonable elevation at the western edge of the belt (approximately 5 km or less); this value for the strength compares well with others obtained by different methods in different orogenic belts. The theoretical thrust faults (slip lines) have the same characteristic trend and

slope as the actual faults. The average relative particle velocity distribution shows that the material in the wedge was 'pushed' and 'flowed' up, away from the core zone (Omineca Belt), with the greatest velocities being at the upper rear surface and decreasing with depth and distance toward the craton.

5) Deformation along individual thrust planes (where stress can be locally higher than the regional stress) is probably due to frictional sliding or pressure solution slip if thrusting occurs along discrete surfaces. If the relative movement is taken up along a narrow (on the order of a few metres) zone of concentrated shear, pressure solution and superplasticity are the most likely mechanisms if thrusting occurred at temperatures larger than 200°C. The absence of detailed microrheological work in the Foreland Belt does not allow any more definite statements on this last point.

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APPENDIX I

List of C 's and α 's for the three different times for all values of X
calculated.

Results for the Oligocene (35 Ma)

For a K of 100 bars.

χ	C	Ave α	ΔH (m)
1.0	8.80	2.59	6788
0.9	7.77	2.29	5993
0.8	6.79	2.00	5238
0.7	5.78	1.70	4459
0.6	4.80	1.41	3702
0.5	3.83	1.13	2955
0.4	2.86	0.84	2206
0.3	1.88	0.55	1450

Results for the Cretaceous (95 Ma)

○ For a K of 100 bars.

X	C	Ave α	ΔH (m)
1.0	12.24	2.09	5117
0.9	10.87	1.86	4544
0.8	9.54	1.63	3988
0.7	8.19	1.40	3423
0.6	6.86	1.17	2869
0.5	5.55	0.95	2320
0.4	4.24	0.73	1773
0.3	2.91	0.50	1217

Results for the Jurassic (145 Ma)

For a K of 100 bars

X	C	Ave α	ΔH (m)
1.0	14.62	2.10	7152
0.9	13.01	1.87	6365
0.8	11.45	1.65	5601
0.7	9.85	1.42	4819
0.6	8.28	1.19	4051
0.5	6.75	0.97	3302
0.4	5.19	0.75	2539
0.3	3.62	0.52	1771

APPENDIX II

Computer program used in the calculation of the Compressive-Spreading Wedge model. Program was written from calculations and formulae discussed by Hill (1950) and Chapple (1978).

12

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1. C*****C
2. C C
3. C C
4. C THE PURPOSE OF THIS PROGRAM IS TO CALCULATE THE REQUIRED SURFACE SLOPE. C
5. C WITHIN A FORELAND FOLD AND THRUST BELT, THE METHOD USED IS THE METHOD C
6. C DETERMINED BY CHAPPEL(1978). WITHIN THE PROGRAM THE FORELAND BELT IS C
7. C SUBDIVIDED INTO SIX SECTIONS, A SET OF C'S ARE DETERMINED FOR EACH SECTION C
8. C WITH EACH C CORRESPONDING TO A SPECIFIC CHI. THE CHI'S START AT 1.0 AND C
9. C DECREASE TO 0.5 AT INCREMENTS OF 0.1, ALSO FOR EACH SECTION THERE IS THE C
10. C DETERMINATION OF A SET OF ALPHA'S, EACH ONE CORRESPONDING TO A SPECIFIC C
11. C C. THERE IS ALSO THE CALCULATION OF AN AVERAGE C, ALPHA, AND AVERAGE HEIGHT C
12. C WHICH IS ALWAYS DESIGNATED INTO SECTION 7. THESE AVERAGE VALUES ARE USED C
13. C TO DETERMINE THE TANGENTIAL STRESS, RADIAL STRESS, TANGENTIAL VELOCITY, AND C
14. C RADIAL VELOCITY, THE FINAL FOUR VALUES ARE DETERMINED AT 5 KM INTERVALS C
15. C ALONG THE RADIUS LINES. C
16. C THE STRENGTH OF THE MATERIAL WITHIN THE FORELAND THRUST BELT WAS TAKEN TO C
17. C BE EITHER 600, 300, 200, 150, 125, 100 OR 50 BARS THROUGHOUT THE PROGRAM. C
18. C C
19. C C
20. C*****C
21. C DIMENSION SIGMAT(80,20,8), SIGMAR(80,20,8) C
22. C DIMENSION THETA(7), XPSI(8), C(7,8) C
23. C DIMENSION XH(7), ALPHA(7,8) C
24. C DIMENSION APSI(20,7,8), ATHETA(20,7,8) C
25. C DIMENSION PSIAVE(20,8), ALAVE(8), CAVE(8) C
26. C DIMENSION THAVE(20,8), LOB(1), DR(1) C
27. C DIMENSION MN(1), XD(1), XQ(6), IP(1), RR(1) C
28. C READ 163, (MN(I), I=1,1) C
29. C 163 FORMAT(I10) C
30. C READ 10, (THETA(I), I=1,7) C
31. C 10 FORMAT(7F5.2) C
32. C READ 20, (XPSI(J), J=1,8) C
33. C 20 FORMAT(8F5.3) C
34. C READ 26, (XH(I), I=1,6) C
35. C 26 FORMAT(7F10.1) C
36. C READ 823, (LOB(I), I=1,1) C

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37.      623      FCRMAT(I10)
38.      READ 166,(OR(I),I=1,1)
39.      166      FCRMAT(F3,1)
40.      READ 1000,(XG(I),I=1,6)
41.      1000      FCRMAT(6F10,1)
42.      READ 1002,(IP(I),I=1,1)
43.      1002      FCRMAT(I2)
44.      READ 1001,(RR(I),I=1,1)
45.      1001      FCRMAT(F10,1)
46.      READ 164,(XC(I),I=1,1)
47.      164      FCRMAT(F12,1)
48.      DC 1006,I=1,1
49.      M=RN(I)
50.      X=XD(I)
51.      R=RR(I)
52.      I=IP(I)
53.      1006      CONTINUE
54.      C*****
55.      C
56.      C
57.      C THE DETERMINATION OF C. THE C'S WILL BE SET UP INTO A TWO DIMENSIONAL
58.      C ARRAY WITH C 1,1 BEING THE VALUE OF C CORRESPONDING TO A X VALUE
59.      C OF 1.0 WITHIN THE FIRST SEGMENT OF THE THRUST BELT. C 1,2
60.      C IS THE VALUE OF C CORRESPONDING TO X=0.9 ALSO WITHIN THE FIRST
61.      C SEGMENT, ETC. C 2,1 IS THE VALUE OF C FOR AN X OF 1.0 FOR THE SECOND
62.      C SEGMENT AND SO ON.
63.      C
64.      C
65.      C*****
66.      DC 100,I=1,7
67.      THETA(I)=THETA(I)/57.295779
68.      DC 200,J=1,8
69.      X1=1.5
70.      51      Z1=(X1/SGRT(X1*X1-1.0))
71.      Z2=(SGRT((X1-1.0)/(X1+1.0)))
72.      Y1=XPSI(J)-Z1*ATAN(Z2*ATAN(XPSI(J)))-THETA(I)

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73. IF (Y1, L, 0.0) GO TO 50
74. X1=X1+0.5
75. GC TC 51
76. X2=X1
77. 50 Z3=((X2/√SQRT(X2*X2-1.0)))
78. 53 Z4=(SQRT((X2-1.0)/(X2+1.0)))
79. Y2=XPSI(J)-23*ATAN(Z4*ATAN(XPSI(J)))-THETA(I)
80. IF (Y2.GT.-0.000001) GO TC 52
81. X2=X2-0.01
82. IF (X2.LE.1.0) GO TO 52
83. GC TC 53
84. C(I,J)=X2
85. YPSI=XPSI(J)*57.295779
86. YPETA=THETA(I)*57.295779
87. PRINT 500,I,J,C(I,J),YPSI,YPETA
88. 500 FORMAT('I,20X,THE VALUE OF C',I,1X,11,' IS ',F7.3,' FOR A PSI
89. $ OF ',F6.3,' AND A THETA VALUE OF ',F7.4)
90. IF (I.EC.NP) CAVE(J)=C(I,J)
91. 200 CCNINUE
92. PRINT 529
93. 529 FORMAT('0')
94. 100 CCNINUE
95. C*****
96. C
97. C
98. C DETERMINATION OF ALPHA FOR SIX SECTIONS WITHIN THE FORELAND IMHUST
99. C FELT
100. C
101. C
102. C*****
103. RHC=2.5
104. XC=980.0
105. DC=0.0
106. MA=MM-1
107. XX=1.0
108. DC 300,J=1,8

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109. DC 400, I=1, MA
110. YA=XX/1000000.0
111. AI=XP(I)/100000.0
112. XTHETA=THETA(I)*57.295779
113. Z7=2.0*XX*C(I,J)*THETA(I)
114. Z8=XG*RHMC*XP(I)
115. ALPHA(I,J)=ATAN(Z7/Z8)
116. PRINT 59, XX, YX, I
117. 59 FORMAT(' ', 10X, 'FOR AN X VALUE OF ', F3.1, 5X, ' WITH A K CF ', F5.1,
118. 55X, ' WITHIN SECTION ', I1)
119. BALPHA=ALPHA(I,J)*57.295779
120. PRINT 69, C(I,J), AI, XTHETA, BALPHA
121. 69 FORMAT(' ', 20X, 'C=', F6.2, 10X, 'AVERAGE THICKNESS =', F8.5, ' KM', 10X,
122. 5THETA=', F5.2, ' DEGREES', 10X, 'ALPHA=', F5.2, ' DEGREES')
123. IF (I.LT.IG) GO TO 400
124. DG=DG+BALPHA*XG(I)
125. 400 CCNTINLE
126. DG=DG/RG
127. PRINT 729, XX, YA, C(I,J), DG
128. 729 FORMAT(' ', 10X, 'FOR AN X VALUE OF ', F3.1, 5X, ' WITH A K CF ', F5.1,
129. 5SX, ' AND A C = ', F6.2, 5X, 'ALPHA = ', F5.2)
130. AG=RG*ATAN(DG/57.295779)
131. PRINT 730, AG
132. 730 FORMAT(' ', 10X, 'THEREFORE THE HEIGHT AT THE WESTERN LIMIT OF THE
133. 5BELT IS ', F10.2, 2X, ' METERS')
134. XX=XX+0.1
135. PRINT 530
136. 530 FORMAT('01')
137. 300 CCNTINUE
138. C*****
139. C
140. C
141. C THE DETERMINATION OF PSI FOR GIVEN THETAS AND DIFFERENT VALUES OF C
142. C PSI IS DETERMINED FOR 1/4 DEGREE INTERVALS IN THETA WITHIN EACH
143. C SECTION OF THE THRUST BELT FOR ALL 8 VALUES OF C.
144. C
145. C
146. C*****

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147. DC 800,J=1,I
148. DC 900,I=1,M
149. PRINT 281,I,C(I,J)
150. FCRRMAT(' ',10X,'THE VALUES OF PSI FOR CORRESPONDING THETAS
151. WITHIN SECTION ',11,2X,' WITH A C VALUE OF ',F6.2,' ARE ')
152. PRINT 286
153. FCRRMAT(' ',30X,'THETA',10X,'PSI')
154. N=IFIX(THETA(I)/(0.25/57.295779))+2
155. AN=N-1
156. RXY=0.0
157. PPSI=0.0
158. Z9=(C(I,J)/SGRT(C(I,J)*C(I,J)-1.0))
159. Z10=SGRT((C(I,J)-1.0)/(C(I,J)+1.0))
160. DC 901,K=1,N
161. PTHETA=PPSI-Z9*ATAN(Z10*ATAN(PPSI))
162. IF(PTHETA.GE.RXY) GO TO 651
163. PPSI=PPSI+(0.01/57.295779)
164. GC TO 652
165. AFSL(K,I,J)=PPSI
166. XAPSI=APSI(K,I,J)*57.295779
167. RTHETA=PTHETA*57.295779
168. ATHETA(K,I,J)=PTHETA
169. PRINT 28,RTHETA,XAPSI
170. FCRRMAT(' ',30X,F5.2,10X,F5.2)
171. IF(K.EQ.N) GO TO 861
172. RXY=RXY+(0.25/57.295779)
173. GC TO 109
174. RXY=THETA(I)
175. IF(I.EQ.M)PSIAVE(K,J)=APSI(K,I,J)
176. IF(I.EQ.M)THAVE(K,J)=ATHETA(K,I,J)
177. CCNTINUE
178. PRINT 962
179. FCRRMAT('01')
180. CCNTINUE
181.

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182. C*****
183. C
184. C THE DETERMINATION OF SIGMA THETA AND SIGMA R
185. C (TANGENTIAL STRESS AND RADIAL STRESS )
186. C FOR THE THRUST BELT USING AN AVERAGE C, AN AVERAGE
187. C THETA (BASAL SLOPE), AND A CORRESPONDING AVERAGE
188. C PSI.
189. C
190. C
191. C
192. C*****
193. DO 380, I=1,1
194. L=LUB(I)/5
195. CHIDR(I)
196. 386 CONTINUE
197. DO 920, J=1,1
198. PRINT 993, CHI, CAVE(J)
199. 993 FORMAT(' ', 10X, 'FOR A CHI VALUE OF ', F3.1, ' AND A C VALUE OF ',
200. *F6.2)
201. DO 921, K=1, N
202. HAVE=50000.0
203. XTHAVE=THAVE(K, J)*57.295779
204. XPSIAVE=PSIAVE(K, J)*57.295779
205. PRINT 865, XTHAVE, XPSIAVE
206. 865 FORMAT(' ', 10X, 'AND AN AVERAGE THETA VALUE OF ', F5.2,
207. $ ' AND CORRESPONDING PSI VALUE OF ', F5.2)
208. PRINT 967
209. 967 FORMAT(' ', 20X, 'R', 12X, 'NORMAL STRESS', 7X, 'LONGITUDINAL STRESS',
210. $, 1X, 'NORMAL VELOCITY', 5X, 'LONGITUDINAL VELOCITY')
211. DO 922, M=1, L
212. Z11=(CAVE(J))*ALOG(HAVE))
213. Z12=CAVE(J)+COS(2*PSIAVE(K, J))
214. Z13=(RHO*XG*HAVE*SIN(THAVE(K, J)))/(2*XX)
215. Z14=CAVE(J)-COS(2*PSIAVE(K, J))
216. SIGMAH(M, K, J)=(-Z11+0.5*Z14-0.5*CAVE(J))*ALOG(Z12)-Z13)
217. SIGMAH(M, K, J)=(-Z11+0.5*Z12-0.5*CAVE(J))*ALOG(Z12)-Z13)

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218. XSIGMAR=SIGMAR(M,K,J)
219. XSIGMAT=SIGMAT(M,K,J)
220. XRAVE=RAVE/100000.0
221. C*****
222. C
223. C
224. C DETERMINATION OF THE VELOCITY VECTORS USING AN AVERAGE C, AVERAGE
225. C PSI AND NOT USING THE CONSTANT B (RATE OF SUPRACRUSTAL SHORTENING).
226. C HENCE THE VELOCITIES ARE RELATIVE
227. C
228. C
229. C*****
230. C
231. C B=1.0
232. C SIGMAT(M,K,J)=B*RAVE*(9IN(2.0*PSIAVE(K,J))-CHI)
233. C SIGMAR(M,K,J)=-B*RAVE*(CAVE(J)+COS(2.0*PSIAVE(K,J)))
234. C YSIGMAT=SIGMAT(M,K,J)
235. C YSIGMAR=SIGMAR(M,K,J)
236. C PRINI 923,XRAVE,XSIGMAT,XSIGMAR,YSIGMAT,YSIGMAR
237. C FORMAT(' ',20X,F5.1,6X,F15.2,5X,F15.2,5X,F15.2)
238. C RAVE=RAVE+500000.0
239. C CONTINUE
240. C PRINI 924
241. C FORMAT('10')
242. C CONTINUE
243. C CHI=CHI-0.1
244. C CONTINUE
245. C CALL EXIT
246. C END

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END

1	0	1	0	9	8	4

FIN