THE EFFECT OF DEFORMATIONAL MECHANISMS
ON THE PERMEABILITY OF
UPPER PALEozoIC LIMESTONE, DOLOSTONE AND
SANDSTONE NEAR OVERFOLD MOUNTAIN,
55 KILOMETRES SOUTHEAST OF FERNIE, BRITISH COLUMBIA

by
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Date **2 June, 1989**
ABSTRACT

Overfold Mountain is located approximately 55 km southeast of Fernie British Columbia in the MacDonald Range of the Rocky Mountains. In the area, limestones, dolostones and sandstones of the Upper Carboniferous Rundle Group and Rocky Mountain Group have been thrown into a series of northwesterly trending folds on the hanging wall of the Lewis Thrust sheet. This study focuses on the deformational mechanisms which have led to the development of one of these megascopic structures, with an emphasis on the role of permeability before, during and after deformation.

Deformational mechanisms which have been active near Overfold Mountain include solution processes (pressure solution and hydraulic fracturing), shear fracturing, and intragranular mechanisms (mechanical twinning, dislocation glide, and microfracturing). How strain is partitioned between these mechanisms is largely governed by the permeability of the unit. Permeability is of primary importance in the determination of how a rock will respond in a nonhydrostatic stress field at low temperatures (<0.5 Tm). In the study area, carbonate rocks with a high initial permeability have accommodated strain by pressure solution. Carbonates and sandstones of low initial permeability have accommodated strain by shear fracturing and intragranular mechanisms.
Finite permeability in the carbonates and sandstones of the study area, has been altered as a result of these deformational mechanisms. Units which had a high permeability prior to deformation have had their permeability blocked by pressure solution. Units which had a low permeability prior to deformation, have developed microfractures which have increased the finite permeability. This latter phenomenon is well illustrated in the dolostone units studied, both of which have a very well developed fracture porosity.
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1. INTRODUCTION

The Canadian Rocky Mountains represent the easternmost part of the Canadian Cordillera. This mountain belt has a structural style which is unique within the Canadian Cordillera, consisting of large, imbricate thrust faults and associated folds which have shortened the rocks of the belt by more than 200 km. This study focuses on the deformational mechanisms which have led to the development of one of these megascopic structures.

Several intimately related factors have contributed to the style of deformation found in the Canadian Rocky Mountains. Of foremost importance, however, is the anisotropic nature of the sedimentary package. Throughout the region the stratigraphic sequence commonly includes interbedded limestones, dolostones, shales and sandstones, each of which behaves very differently in a deviatoric stress field. The viscosity contrast and abrupt contacts between each of these lithologies has facilitated the development of flexural-slip folds and bedding parallel shear.

Within the study area, the deformational mechanisms of primary importance are solution processes (hydraulic fracturing and pressure solution), shear fracturing, and intragranular mechanisms (mechanical twinning and dislocation glide). Some of these mechanisms have increased the permeability of the rock during deformation, enhancing
fluid flow and possibly hydrocarbon migration. Remnant structures created by these mechanisms, in particular hydraulic fractures and stylolite seams, act as kinematic indicators which depict not only a strain history of the rock, but also a history of the permeability over the deformation interval.

An effort is made in this study to unravel the deformational and permeability histories of distinct lithologies within a megascopic fold pair in southeastern British Columbia. This has been carried out by the detailed analysis of the size, density, and cross-cutting relationships of the remnant structures on both the microscopic and mesoscopic scales.

1.1. LOCATION AND ACCESS

Overfold Mountain is located between Lodgepole Creek and Bighorn Creek, approximately 55 km southeast of Fernie British Columbia in the MacDonald Range of the Canadian Rocky Mountains (Lat. 49° 13' N, Long. 115° 47' W; Figure 1). The study area is accessible by two-wheel-drive vehicle via logging roads to the head of a well maintained horse trail which may be followed for the remaining 5 km to the base camp. Field work was conducted between July 2 and August 30, 1986.
Figure 1 - Location map for study area.
1.2. **Previous Studies**

Geologic observations of the Canadian Rocky Mountains were first published by James Hector (1863), a member of Captain Palliser's expedition. Soon after, pioneering work by Dawson (1875, 1886), Dawson and McConnell (1885), Willis (1902) and Daly (1912) began to reveal the fold and thrust nature of the belt as well as a basic stratigraphic sequence.

Pioneering spirit gave way to monetary aspirations with the discovery of gas and oil in the Turner Valley in 1913. At this time, the search for structural traps and productive beds began, bringing a deluge of petroleum geologists into the area, each adding to the wealth of knowledge of both the structure and the stratigraphy of the fold and thrust belt.

Several major regional studies have centred on the Rocky Mountains of southeastern British Columbia. Bally and others (1966) presented a landmark study which provided an understanding of the extent and nature of the fault systems there. This study also showed the important variations in structural style, from the Rocky Mountain Trench to the west, through to the Front Ranges.

Important insight into the tectonic significance of the structures of the fold and thrust belt has been given by Price (1959, 1962, 1965) who has mapped the Rocky Mountains of southeastern British Columbia in significant detail. From this mapping, inferences have been made as to
the structural controls on the deformation seen there

Only a minor effort has been made to relate mesoscopic
and microscopic structures to the megascopic structures in
the Rocky Mountains. In a notable exception, Price (1967)
found a good correlation between the mesoscopic and
megascopic fabrics in the southern Rocky Mountains, an area
which includes Overfold Mountain. There have, however, been
many important studies of mesoscopic and microscopic
structures in other, similar, structural provinces. Some
researchers have related mesoscopic fractures and stylolites
to megascopic folds (Groshong, 1975) and to faults
(Friedman, 1969; Rispoli, 1981; Hancock, 1985). Others have
used these structures, as well as calcite twin lamellae, to
estimate the finite strain within major structures (Friedman
and Heard, 1974).

A relationship between the development of these
mesoscopic and microscopic structures and fluids within the
rock during deformation has been suggested by Phillips
(1972, 1974), Alvarez and others (1976), Beach (1977),
Kerrich (1978), Geiser and Sansone (1981), and Etheridge and
Vernon (1983) as well as others. Many of these researchers
agree that pressure solution and hydraulic fracturing not
only require fluids for development, but may also enhance
the bulk fluid flow through the rock.
2. GEOLOGIC SETTING

The Canadian Rocky Mountain fold and thrust belt has long been recognized as a major source of petroleum, natural gas and coal. Because of the economic significance of the area, extensive geological and geophysical data have been accumulated, making this one of the most studied and best understood structural provinces of its kind in the world.

The keys to the style of deformation within the fold and thrust belt lie firstly in the strongly anisotropic nature of the sedimentary layers involved, and secondly in the contrast in competency between the crystalline basement and this overlying sedimentary cover. Variations in thickness and lithology across the belt have also had a substantial effect on the style of deformation, a factor which has prompted the division of the fold and thrust belt into four zones: the Foothills, Front Ranges, Main Ranges and Western Ranges (Figure 2; Bally and others, 1966).

2.1. STRATIGRAPHIC FRAMEWORK

The sedimentary sequence of the Rocky Mountain fold and thrust belt may be divided into two distinct packages. Oldest of these is a succession of distinctly bedded miogeoclinal sediments which range in age from Late Precambrian (Helikian) to Late Jurassic. This package is overlain by a Latest Jurassic to Early Tertiary clastic wedge sequence. The miogeoclinal sediments onlap the metamorphic and intrusive rocks of the earlier Precambrian
Figure 2 - Structural provinces of the southern Canadian Rocky Mountains. After Bally and others (1966).
basement, which were consolidated during the Hudsonian orogeny (1,700 Ma), and represent a westward extension of the Canadian Shield.

Sedimentary rocks found at Overfold Mountain are Upper Paleozoic in age and belong to the widespread miogeoclinal succession. Regionally, the miogeoclinal succession is characterized by interbedded limestones, dolostones, sandstones and shales which were deposited into a shallow, warm, marine environment. The clastics in this succession were provided by the emergent cratonic platform to the east. The thickness of the miogeoclinal package varies from about 2,000 m, east of the Front Ranges to more than 12,000 m in the western Rocky Mountains (Price and Mountjoy, 1970).

In the Late Jurassic, tectonic processes in the western Cordillera caused the development of a foredeep basin in the area and the subsequent deposition of a clastic wedge sequence (Bally and others, 1966). This sequence consists of marine sediments and non-marine clastics which were shed from the newly emergent rocks to the west. Fossil evidence has shown that the earliest influx of clastics into the basin occurred in the Latest Jurassic and continued in pulses throughout the Cretaceous and Paleocene (Wheeler, 1966). During this time, ongoing tectonic processes caused the migration of the foredeep basin toward the northeast, where the clastics are seen to interfinger with, and overlap, marine sediments. Remnants of the wedge are best preserved in the Rocky Mountain Foothills, where a
maximum thickness of 6,500 m is reached (Price and Mountjoy, 1970).

2.1.1. **LOCAL STRATIGRAPHY**

Sedimentary rocks exposed in an overturned, asymmetric fold pair found near Overfold Mountain belong to the Lower Carboniferous Mount Head and Etherington Formations (Rundle Group) and the overlying Upper Carboniferous Rocky Mountain Group. In the study area, limestones, dolostones, and shales of the Mount Head Formation are very well exposed along a northwesterly trending ridge which follows the hinge of the anticline (Figure 3). Dolostones and arenaceous dolostones of the overlying Etherington Formation and sandstones and arenaceous dolostones of the Rocky Mountain Group are exposed at the hinge and limbs of the syncline.

The Mount Head Formation is widespread along the fold and thrust belt and has been divided into seven members on the basis of lithologic and faunal observations (Macqueen and Bamber, 1968). The fauna within the Mount Head Formation of the study area have been placed within the zonal assemblage of the uppermost three members of this formation, deposited in the Late Visean (Maddison, 1987; Figure 4). The oldest of these units is the Loomis Member which consists of thick-bedded to massive pelmatozoan and oolitic grainstones. These rocks are overlain by the thin to medium-bedded limestone, finely crystalline dolostone and shale of the Marston Member, which in turn is overlain by
Figure 3 - Photograph illustrating the outcrop pattern of three Members of the Mount Head Formation in the anticline within the study area. The Etherington Formation outcrops in the syncline, along the valley floor.
ROCKY MOUNTAIN GROUP
Calcereous quartz arenite

ETHERINGTON FORMATION
Interbedded dolostone and sandy dolostone

MOUNT HEAD FORMATION

CARNARVON MEMBER
Interbedded limestone and dolostone

MARSTON MEMBER
Interbedded limestone, dolostone and shale

LOOMIS MEMBER
Massive limestone

Figure 4 - Simple stratigraphic column of the Upper Carboniferous units studied near Overfold Mountain. Tic marks measure 50 m.
thick to thin-bedded skeletal and oolitic grainstone and lime mudstone of the Carnarvon Member.

The contact of the Mount Head Formation with the overlying Etherington Formation is a regional disconformity (Douglas, 1958). No obvious surface of unconformity was seen in the study area and the contact was taken as the top of the uppermost limestone unit. In the study area, the Etherington Formation is composed of medium-bedded dolostones, arenaceous dolostones, calcareous quartz arenites and shales. Brachiopods are common in some units, and the arenaceous units are locally distinctively cross-laminated. This formation is exposed on the overturned limb of the folds where it reaches a thickness of approximately 60 m.

The contact of the Etherington Formation with the overlying Rocky Mountain Group is also believed to be a regional disconformity (Douglas, 1958). Since no unconformable surface was observed in the study area, the contact was taken as the uppermost dolomitic unit. The Rocky Mountain Group consists of very clean light-grey massive quartz arenite with calcareous cement. This unit is well exposed at the hinge and limbs of the syncline and has a minimum thickness of 100 m.
2.2 STRUCTURAL FRAMEWORK

On the regional scale, the Rocky Mountains are seen to comprise a wedge of sediments which tapers toward the northeast. These sediments have been shortened by nearly 200 km by the development of large, listric thrust faults (Price, 1965). Displacement varies between individual faults, but the largest thrust sheets are known to have been transported as much as tens of kilometres (Bally and others, 1966). These faults generally have a southwesterly dip, and have thrust older rock onto younger as the hanging wall was displaced toward the northeast or east (Price, 1965).

The well developed sedimentary layering has imparted an anisotropy which has controlled the style of deformation throughout the Canadian Rocky Mountains. Because of the abrupt changes in lithology between individual beds and the correspondingly abrupt changes in mechanical properties, tectonic shortening has principally been taken up by layer parallel slip (Price, 1973). This condition has resulted in the development of flexural slip folds and thrust faults. The thrust fault surfaces are characterized by broad intervals of slip parallel to bedding, typically along or within less competent layers such as shales, connected by short intervals of slip (ramps) across the more competent layers (Douglas, 1958). These thrust faults do not extend down into the crystalline basement rocks; instead the
basement surface represents a regional decollement along which the thin sedimentary cover was literally "scraped" off. The basement rocks were left relatively undisturbed by this process (Bally and others, 1966).

2.2.1. **LOCAL STRUCTURE**

Overfold Mountain exists as a surface expression of one of a series of northwesterly trending folds which have formed on the hanging wall of the Lewis thrust sheet. In the area, the resistant Paleozoic limestones and dolostones, exposed in the hinges of anticlines, form sharp ridges and peaks. Less resistant Upper Paleozoic and younger sandstones and shales are exposed in the hinges of synclines, the axes of which typically fall along valley floors.

The anticline and syncline in the study area each have one limb which dips gently to moderately toward the southwest and have a shared limb which is vertical to overturned (Figure 5). Fold axes (plunge 10° toward 155°) of the two folds are parallel. The axial plane of the syncline (148/55 SW) dips somewhat more steeply than that of the anticline (145/45 SW). The orientation of these fold structures follows the northwesterly regional trend.
Figure 5 - Poles to bedding in the syncline and anticline studied near Overfold Mountain. Great circle represents the axial plane. Solid dot represents the fold axis.
3. **DEFORMATIONAL MECHANISMS - A BACKGROUND**

Deformational processes which have taken place at Overfold Mountain are evidenced on both the mesoscopic and microscopic scale. On the mesoscopic scale, strain is documented by extension fractures, shear fractures and stylolite seams. These features arise from intergranular processes, that is, the deformational mechanisms have operated between individual grains. On the microscopic scale, smaller versions of the mesoscopic structures are seen, as well as structures resulting from intragranular processes. Intragranular processes operate within grains and cause distortions, or slip, within the crystal lattice.

Intragranular mechanisms:

1. Polysynthetic twinning.
2. Translation gliding (slip).
3. Deformation lamellae development.
4. Extension fracturing.

Intergranular mechanisms:

1. Rock-fluid interactions.
   a. Hydraulic fracturing.
   b. Pressure solution.
      1) Stylolite seams in carbonates.
      2) Flat to concavo-convex contacts between quartz grains in sandstones.
2. Shear fracturing.
3. Faults.

All of the aforementioned deformational mechanisms have been utilized to some degree throughout the study area. Lithology, however, plays a dominant role in the determination of how strain is partitioned between these mechanisms. Factors which influence the way in which a rock
deforms include: framework and matrix mineralogy, abundance of matrix, grain size, permeability and degree of cementation. For example, the difference in the chemical composition and lattice structures of calcite, dolomite and quartz cause profound differences in their strength and solubility. Two rocks composed of unequal amounts of these minerals will thus differ in their response to non-hydrostatic stress. Different lithologic units from the same section of the fold may therefore reveal a similar strain history, but have arrived at their final point of finite strain via different pathways.

3.1 INTRAGRANULAR PROCESSES

Intragranular deformation results in the accommodation of strain by processes that take place within the crystal lattice of the grain. The intragranular deformational mechanisms that have been operative in the rocks of the study area include:

1. Glide mechanisms - Ductile regime.
   a. Polysynthetic twinning.
      1) Calcite.
      2) Dolomite - Minor.
   b. Translation glide.
      1) Dolomite.
      2) Calcite - Minor.
   c. Deformation lamellae development.
      1) Quartz.

2. Extension fracturing - Brittle regime.
   1) Calcite.
   2) Dolomite.
   3) Quartz.

Polysynthetic twins are abundant in the calcite grains and cement within the limestones of the study area. These
twins are particularly well developed in the biosparite units (Units 6 and 8; Figure 6). Dolomite crystals, on the other hand, do not show much deformational twinning, a reflection of the stronger structure of dolomite crystals.

The mechanism of translation glide, or slip, has probably been operative in the calcite and dolomite grains. Since slip results in a perfect lattice structure, the amount of strain accommodated by this mechanism cannot be measured unless the original grain shape is known.

Quartz grains containing deformation lamellae are found in the sandstone and sandy dolostone units. They are not abundant, only about 20 cm$^2$, this indicates that this mechanism has not played a very large role in the accommodation of the strain induced by folding.

Intragranular extension fractures are found in all lithologies. In calcite and dolomite, these fractures often follow twin or cleavage planes but are typically random. Quartz appears to have preferentially accommodated strain by intragranular fracturing. The process by which fractures are initiated and propagated will be covered in more detail in the section on fracturing.

Each mineral and each aggregate of minerals is thus seen to accommodate strain by a different mechanism. How a rock responds to an applied stress is an expression of the crystal chemistry of the minerals within it. Crystals have behaved ductilely where the lattice structure has allowed
Figure 6 - Photomicrograph illustrating the abundant twin lamellae within the grains and cement of the biosparite unit, Unit 8. Long edge of photograph measures 0.3 mm.
it. In the absence of ductile mechanisms, crystals ultimately fail by fracture.

The foundation of ductility lies in the ability of a material to flow, without loss of cohesion or strength, when stressed beyond its elastic limit. Ductile strain in crystals results from gliding along one or more glide planes. Glide planes, (T), are typically planes of closely packed atoms within the crystal lattice. The direction of glide along this plane, <t>, is governed by the arrangement of atoms. The two intracrystalline glide mechanisms operative in the carbonates result from the migration of a dislocation along a glide plane. The first mechanism, twinning, leads to the distortion of the crystal lattice on one side of the plane such that the distorted lattice has a twin relationship with the undistorted lattice (Figure 7). This results in lattice imperfections along the twin plane, a factor which may cause these planes to be the locus of later fracturing. The second glide mechanism, translation glide or slip, occurs by the migration of a dislocation along a glide plane, and will not result in the accumulation of dislocations along a plane.

3.1.1. CALCITE

The structure of calcite is known to consist of layers of Ca$^{2+}$ ions alternating with layers of CO$_3^{2-}$ groups. This layered structure provides many planes of closely packed ions along which glide may occur (Figure 8). Calcite forms polysynthetic twins very easily (low stress and low
Figure 7 - Idealized crystal structure of twinned calcite. Twin plane is \{0012\}. After Higgs and Handin (1959).

Figure 8 - Idealized crystal lattice structure of undeformed calcite. See Figure 7 for legend.
temperature; Nicolas and Poirier, 1976). By far the most common twin is formed parallel to the (0112) planes (the e-planes). Calcite twinning results in simple shear parallel to the twin plane, in an amount proportional to the thickness of the twinned material. Many researchers have used these twins to assign stress configurations to deformed limestones (Turner and Wiess, 1963). In this study the density of the polysynthetic twins have been used as a guide to the relative amount of strain that has been accommodated by this mechanism.

Twin lamellae are best developed in the grainstone units such as Unit 8 (Figure 6) and appear to have been less important in the accommodation of strain in the grains of packstone and wackestone units such as Units 7 and 2 (Figure 9 and 34). The inverse relationship between the density of twin lamellae and percentage of matrix results from two processes.

1. In the units with fine-grained matrix, strain has beenaccommodated by intragranular deformation of the matrix grains instead of the framework grains.

2. In the units with fine-grained matrix, strain has been accommodated by pressure solution, a process which is believed to take place at stresses lower than the critical stress for twinning.

The rationale for the latter mechanism is presented in the section on pressure solution. The foundation of the first mechanism lies in the effect of grain size on the ease with which calcite deforms by intra-crystalline glide mechanisms. The concept of a grain size effect relies on the theory that the surface of a grain will be a source of defects within
Figure 9 - Photomicrograph illustrating the low density of twin lamellae within the grains of skeletal packstone (Unit 7). Long edge of photograph measures 0.06 mm.
the grain. These defects will have little effect on the strength of the crystal if the grain is large, but likely is very important when dealing with grains in the micritic size range (less than 2 um).

Experimental evidence shows that at room temperature, calcite is 10 to 20 times stronger for the mechanism of translation glide than it is for e-twinning (Turner and Wiess, 1963). This effect is reduced with increasing temperature though not sufficiently to make translation glide an important deformational mechanism in the calcite from the study area.

Calcite has behaved brittley on the intra-crystalline level by the development of extension fractures within grains. Fracturing in calcite may occur at any point in the crystal lattice that contains a concentration of dislocations. As mentioned, twin planes, grain boundaries and their intersections represent areas in the crystal where this situation may arise, and where fracturing is commonly initiated.

3.1.2. DOLOMITE

Like calcite, dolomite consists of layers of carbonate groups alternating with layers of cations. There is no solid solution between calcite and dolomite, due to the large difference in ionic radii between the two cations, Mg\(^{2+}\) (0.065 nm) and Ca\(^{2+}\) (0.099 nm). Layers of Mg\(^{2+}\) consistently alternate with layers of Ca\(^{2+}\) (Figure 10). The difference in the size of the two cations makes e-twinning
Figure 10 - Idealized crystal structure of undeformed dolomite.
impossible and results in a crystal structure which is much stronger than the calcite structure.

Dolomite does not behave ductiley as easily as calcite. When it does, it is known to deform by translation glide parallel to \(0001\) (the c-plane) and, to a lesser extent, by twin gliding along the f-plane \(0221\). Other researchers have shown that under a confining pressure of 500 MPa, dolomite will undergo translation glide at all temperatures, provided the crystal is suitably oriented (approximately the same requirements for translation in calcite). If not suitably oriented for translation, the dolomite crystals fail brittley if below 400°C. Above 400°C the crystal deforms by twin glide. Another contrast between dolomite and calcite is that dolomite becomes stronger with increasing temperature between 25°C and 400°C, opposite to the response in calcite (Higgs and Handin, 1959).

As for calcite, fracturing in dolomite will most easily be initiated where defects are concentrated. Again, as for calcite, twin planes and grain boundaries represent two areas where this condition is met. Twin planes are rare in the dolomite in the study area. The grain boundaries are thus the main locus for the initiation of fractures. This effect is enhanced by the rhombic shape of the grains. The sharp edges and points of the dolomite rhombs are very effective stress concentrators (Figure 11), a factor which is responsible for the widespread microfracturing in the
Figure 11 - Photomicrograph illustrating the rhombic shape of dolomite crystals within the dolostone units studied near Overfold Mountain. Long edge of photograph measures 0.02 mm.
dolostones of the study area, and is probably also responsible for the high fracture permeability found in the dolostones of some producing hydrocarbon reservoirs.

3.1.3. **QUARTZ**

The quartz structure consists of one Si\(^{4+}\) ion covalently bonded in tetrahedral coordination with four O\(^{2-}\) ions. Each O\(^{2-}\) ion is coordinated with two Si\(^{4+}\) ions. As a result of these covalent bonds, quartz is very strong and does not readily deform by intra-crystalline glide mechanisms. In fact, experimental evidence has shown dry quartz to have a strength almost equal to its theoretical strength at 500°С (Griggs, 1967). This unusual strength is due to the difficult task of breaking the strong covalent Si-O bonds. This task is made much easier by the addition of water to the crystal lattice. Water leads to the hydrolysis of the Si-O bonds and the weakening of the lattice (Griggs, 1967). The thus formed Si-OH HO-Si bridges are easily broken, allowing the migration of dislocations. Hydrolytic weakening of quartz is known to be important in encouraging all deformational mechanisms involving quartz (i.e. dissolution, crystallization, fracturing).

The only direct evidence of intra-crystalline glide mechanisms within the quartz arenite of the study area is the occurrence of deformation lamellae. Deformation lamellae are narrow planer features along which the crystal lattice has been disrupted. These lamellae have been identified by many researchers in experimentally and
naturally deformed quartzites, and have been the subject of intense investigation (Christie et al., 1964; McLaren et al., 1970; Christie and Ardell, 1974). Lamellae in experimentally deformed quartz form along the basal plane of the quartz structure (0001). Lamellae in naturally deformed quartz have a more random orientation and are found to form at angles up to 30° from the basal plane (Heard and Carter, 1968).

3.2. INTERGRANULAR PROCESSES

Intergranular deformation results in the accommodation of strain by processes that take place between grains. The structures resulting from intergranular deformational mechanisms are:

1. Hydraulic fractures.
2. Shear fractures.
3. Pressure solution features.
   a. Stylolite seams in carbonates.
   b. Flattened contacts in sandstones.

The development of each of these structures is closely tied to the permeability of the rock at the time of its formation. The reasons for this will be covered in detail for each structure.

3.2.1. FRACTURES

Excellent reviews of the development of current theories on the brittle deformation of rocks have been given by Nadai (1950), Paterson (1978) and Jaeger and Cook (1976). Some of the earliest experimental work centred on brittle failure was done in 1773 by C. A. Coulomb. This work was
later modified in the late 19th century and early 20th century by researchers such as Tresca (1868) and Mohr (1901, 1914).

One of the most noteworthy advances in fracture theory was made by A. A. Griffith in 1920. Griffith determined that the strength of a material is dependent upon the existence of tiny cracks (Griffith's cracks) within grains and at grain boundaries. He also postulated that a crack will propagate by extension when stress is applied, provided it is in an orientation which will allow it to overcome the tensile strength of the material. His theory further encompassed the effect of pore pressure on crack propagation.

Griffith's theory has been extensively applied and modified since its inception. Of interest to this study is the effect of fluids within the cracks, a component which can increase the rate of crack growth by two mechanisms; first by the increase in fluid pressure (Paterson, 1978; p. 86) and second by the effect of stress corrosion (Atkinson, 1982; Atkinson and Meredith, 1987).

3.2.1.1. HYDRAULIC FRACTURES

Hydraulic fractures are extension fractures which form in response to high pore pressures (as described by Griffith's theory of failure). When fluid within a rock is compressed, it exerts a force which is normal to all surfaces. This results in a reduction of the effective stresses by the pore pressure value.
\[ \sigma_1^{\text{e}} = \sigma_1 - P \\
\sigma_2^{\text{e}} = \sigma_2 - P \\
\sigma_3^{\text{e}} = \sigma_3 - P \]

\( \sigma_1 \) = Minimum Compressive Stress  \\
\( \sigma_2 \) = Intermediate Compressive Stress  \\
\( \sigma_3 \) = Maximum Compressive Stress  \\
\( P \) = Pore Pressure

Note: Compressive stress is negative.

The result of this process is displayed diagrammatically on a Mohr diagram (Figure 12). When the effective stresses are plotted, it is seen that if the pore pressure is high enough, and the deviatoric stress low enough, the circle will become tangent to the failure envelope in the field of tensile failure. Fractures formed by this process will always propagate in a direction parallel to the maximum compressive stress direction, in a plane which contains the intermediate principal stress and is normal to the minimum compressive stress direction. Knowledge of fracture orientation thus gives insight into the stress configuration at the time of its formation.

Hydraulic fractures originate at the microscopic level by the propagation and interconnection of tiny Griffith's cracks. These tiny cracks are believed to develop where dislocations within the crystal lattice "pile up", either within the crystal (along cleavages or twin planes), or at grain boundaries (Atkinson and Meredith, 1987). The sharp tips of these slit-like cracks are efficient stress concentrators. Water within a crack reacts easily with this
Hydraulic Fractures – The Mohr Circle Model

Figure 12 - Mohr circle representation of the effect of pore pressure in the determination of the effective stresses acting on a rock in a deviatoric stress field. Compressive stress is negative.
stressed material causing the weakening of the material at the tip and the propagation of the crack. This process, known as stress corrosion, can cause crack propagation at stresses lower than the critical value for the material in question.

Two conditions will cause a fracture to remain microscopic in size:

1. The rock is depleted in fluids so that pore pressure cannot build-up.
2. The microfracture becomes rotated with respect to the stress axes such that it is no longer in a favorable position to propagate.

If conditions allow, microfractures will connect to form larger extension fractures. Mesoscopic fractures are produced by the cyclic re-opening and filling of smaller cracks, a process termed "crack-seal" deformation (Ramsay, 1980). Repetitions of this process can lead to the development of large mesoscopic fractures.

Two growth geometries, syntaxial and antitaxial, have been described for the crystallization of material within hydraulic fractures (Durney and Ramsay, 1973; Cox and Etheridge, 1983; Ramsay, 1980). Both are recognized in the study area on the mesoscopic and microscopic scale.

Syntaxial fractures are characterized by a fracture filling which has a mineralogy that is common in the host rock. This suggests that the fluid from which the vein material crystallized had not travelled far, and thus has implications as to the permeability of the rock at the time of fracture formation. The crystal growth within the
fracture is seen to be fibrous. Fibre growth is often in optical continuity with a host grain in the wallrock and is perpendicular to the fracture wall (Figure 13). The wall is irregular, and appears to follow grain boundaries, indicating that these areas of weakness have guided the fracture development. Later episodes of cracking have taken place along the centre of the fracture, leading to a break in the optical continuity of the fibre across the vein, as well as the development of a crystalline fill that is free of wallrock fragments. Earliest crystallization in syntaxial fractures occurs along the walls of the fracture, latest crystallization occurs at the centre of the fracture.

The fracture filling in antitaxial fractures is often of a mineral species that is foreign to the host rock such as calcite veins formed in quartz arenite. This indicates a relatively large distance of fluid transport, which in turn suggest that the host rock was relatively permeable at the time of fracture formation. As with syntaxial fractures, mineral growth is fibrous, and fibres (if straight) grow perpendicular to the vein walls. Unlike syntaxial fractures, however, fibres are in optical continuity across the fracture (Figure 14). Later cracking occurs between the wall and the fill due to the mechanical differences between the two materials. This causes the inclusion of wall rock fragments within the crystalline fracture fill. Earliest crystallization in antitaxial fractures occurs along the
fracture centre, latest crystallization occurs along the fracture wall.
Figure 13 - Mode of development of syntaxial fractures. After Durney and Ramsey, 1973.
Figure 14 - Mode of development of antitaxial fractures. After Durney and Ramsey, 1973.

Wallrock fragments are included with the new crystal growth.

New material crystallizes at the fracture boundary.
3.2.2. PRESSURE SOLUTION

Pressure solution was first studied as a deformational mechanism by Sorby (1858, 1863). Sorby (1858) observed the morphology of cleavage structures and was the first to suggest mineral migration along cleavage planes into associated hydraulic fractures. He also attempted one of the first studies dealing with strain estimation from deformed oolites (Sorby, 1908). In this study, Sorby showed that material was preferentially dissolved along some grain contacts and that the orientation of these contacts could be used for the determination of the stress configuration at the time of dissolution.

The much quoted experimental work by Riecke (1894, 1912) involved the effect of non-hydrostatic stress on solubility. He showed that, given two crystals in a solution with which they are in equilibrium, if a stress is applied to one crystal it will dissolve and new mineral growth will occur on the unstressed crystal. Further experimental work by Correns (1949) determined that growth and dissolution of non-hydrostatically stressed crystals is regulated by stress, temperature, type of material, and crystallographic orientation. This principle was first applied to geologic processes by Sorby (1863) and Spring (1888), and has been exercised extensively in the development of current theory on pressure solution.

Fuchs (1894) determined that stylolite seams were formed as a result of pressure solution. In his study,
Fuchs observed two types of stylolite, diagenetic and tectonic. Diagenetic stylolites result from the stresses associated with sediment loading and thus form parallel to bedding. Tectonic stylolites form oblique to bedding and are a result of tectonic loading.

The first half of the 20th century proved to be a time of little innovative work in the area of pressure solution. Most work done during this period merely elaborated that done previously (Wright, 1906; Knopf, 1933). In the 1960's, a new interest in pressure solution phenomena was spawned by investigators such as Flinn (1965), Rast (1965), and Ramsay (1967). Since this time, there has been a great deal of research concentrated on the importance of pressure solution phenomena in both the diagenetic and the tectonic realms.

Pressure solution is seen to occur to some extent in all of the lithologies in the study area. In carbonates, it is one of the most important mechanisms for the accommodation of strain. Typically, sandstones do not undergo pressure solution to the extent that carbonates do and, in the study area, no mesoscopic fabric is developed. Dissolution in the sandstones from the study area is demonstrated only on the microscopic scale as straight or concavo-convex contacts between quartz grains (Figure 15). This contrast between lithologies is mainly due to the difference in solubility between the two materials, carbonates being far more soluble than quartz in the acidic environment of the stylolite seam.
Figure 15 - Photomicrograph illustrating the straight and concavo-convex contacts between quartz grains in Unit 1. Long edge of photograph measures 0.06 mm.
When a granular substance is subjected to compression, a considerable stress concentration occurs at grain contacts. The resultant stresses at the grain contacts may be many times greater than the overall stress placed upon the substance. According to Riecke's principle, when a solvent acts upon an elastic material in a compressive stress field, the material is dissolved most easily at the points of stress concentration. Pressure solution is thus most readily initiated at grain contacts which are subjected to the highest normal stress.

The orientation of the stylolite seam is governed by the direction of the specific permeability. Ideally, in a homogeneous, isotropic medium, stylolite seams form in a plane normal to the maximum compressive stress direction. Sedimentary rocks are rarely isotropic or homogeneous, however, due to variations in grain size, degree of cementation, and mineralogy, as well as the presence of pre-existing fractures and stylolites. These features affect the permeability of the rock and will tend to cause a deviation of the seam orientation from the ideal (Nelson, 1983).

The toothlike appearance of stylolites in cross-section results from the presence of columns or ridges along the seam (Figure 16). These two distinct stylolite morphologies, columns and ridges, may evolve in different environments; the shape of the structure being governed by the path of least resistance for fluid movement. Columnar
stylolites might then be favoured under conditions where the magnitudes of the minimum and intermediate principal stresses are equal, and fluid can move in all directions parallel to the stylolite seam (Figure 17). This stress configuration would have existed during sediment loading. During folding, the magnitudes of the principal stresses are unequal. Under these conditions, if the sediments are relatively homogeneous, fluid movement would be favoured in the direction of minimum strain along the seam (Figure 17). This process could produce stylolites with the ridge morphology commonly seen in the study area. The lineation formed by the crest of the ridges would parallel the direction of minimum strain. Clay flakes within both of these stylolite morphologies parallel the stylolite seam.

"Ridge" stylolites within the study area differ from similar structures described by other authors (Hancock; 1985). Hancock (1985) describes stylolites with a ridge morphology which have formed in continuum with columnar stylolites (Figure 18). The lineation associated with these ridges parallels the maximum compressive stress. Lineations similar to that described by Hancock (1985) are seen on the microscopic scale in a columnar stylolite found east of the study area, within the Altyn Formation. A scanning electron microscope image of this stylolite reveals lineations along the column which parallel the maximum compressive stress direction (Figure 19). This lineation appears to be the direct result of clay catalyzed
Figure 16 - Morphology of columnar and ridge stylolites.

Figure 17 - Direction of fluid movement along a stylolite seam. During compaction, intermediate and minimum principal stresses are equal, fluid moves freely in all directions. During deformation, intermediate and minimum stresses are unequal, fluid moves most easily in the direction of the minimum stress.
Figure 18 - Continuum between columnar stylolites and stylolites with lineations as described by Hancock (1985).
dissolution along the sides of the column. Kaolinite flakes along the sides of the columns are ordered into "rods" (Figure 20) such that the flakes parallel the seam, and the clay rods are normal to the seam (parallel to the maximum compression). Striations parallel these clay rods, and even mimic the hexagonal shape of the kaolinite flakes. This suggests that the striations have been "carved" out of the limestone by the acidic clay edges.

Illite is the common clay mineral found within the stylolites near Overfold Mountain. Unlike kaolinite, illite flakes do not become ordered into rods. Instead, the elasticity of illite allows it to bend around the stylolite columns. No striations normal to the stylolite seams were found in stylolites containing illite, just as no columnar stylolites were found. It is possible then, that clay mineralogy, as well as fluid movement, may have played a major role in the determination of the morphology of the stylolite.

As dissolution progresses, a selvage of insoluble residue is built up. Depending upon the lithology present, this residue may consist of quartz, clays, kerogen, bitumen, pyrite, oxides and/or dolomite. Dolomite appears to be less soluble than calcite under conditions of pressure solution. Limestone units which are slightly dolomitic tend to have a concentration of dolomite rhombs within stylolite seams (Figure 21). Pyrite and iron oxide are seen to accumulate in seams within dolostones as is quartz (Figure 22).
Figure 19 - SEM micrograph illustrating lineations found on columnar stylolites from the Altyn Formation. These lineations parallel the stylolite columns and developed parallel to the maximum compressive stress direction.
Figure 20 - SEM micrograph illustrating the alignment of kaolinite flakes along stylolite columns. The "carving" action of kaolinite is suggested by the hexagonal shape of the striations on the carbonate. K - Kaolinite; C - Carbonate.
Figure 21 - Photomicrograph illustrating the concentration of dolomite rhombs in stylolite seams within a slightly dolomitic packstone (Unit 7). Long edge of photograph measures 0.3 mm.
Figure 22 - Photomicrograph illustrating the concentration of quartz grains, opaque minerals (pyrite and iron oxide), and organics within a stylolite seam in a sandy dolostone from the Etherington Formation. Long edge of photograph measures 0.3 mm.
Bitumen appears to be present in most seams, indicating that the seams were active during or after hydrocarbon migration.

The role of clay minerals in the catalysis of pressure solution has been given much attention (Heald, 1956; Weyl, 1959; De Boer, 1975; Engelder and Marshak, 1985). Reasons for this phenomenon are still not clear, but it is suggested that the oriented clay flakes act as a diffusion pathway along which fluids can migrate (Weyl, 1959; Geiser and Sansone, 1981). The ability of clay minerals to act as Lewis acids may also aid in the dissolution of carbonate minerals (Goldstein, 1982, 1983). An SEM micrograph of clay in a stylolite seam within a sandy dolostone from the Etherington Formation (Figure 23) shows the alignment of platy phyllosilicates surrounding euhedral quartz. It can be seen here that calcite has been preferentially dissolved. The euhedral nature of the quartz suggests that quartz has actually precipitated as an overgrowth on an existing quartz grain within the stylolite seam. This theory is supported by the quartz in the surrounding rock being well rounded and lacking overgrowths.

After removal from the grain surfaces, dissolved ions diffuse away from the point of dissolution and precipitate in areas of lower (hydrostatic) stress. If the system is closed, the material will precipitate in nearby extension fractures or may diffuse into the pore spaces of the rock. If the system is open (Geiser and Sansone, 1981), some of the material may be transported out of the rock.
Figure 23 - SEM micrograph illustrating the alignment of illite flakes (I) parallel to a stylolite seam within sandy dolostone from the Etherington Formation. Calcite (C) has been preferentially dissolved along the seam, whereas quartz (Q) appears to have precipitated within the seam as indicated by the euhedral crystal form.
The ability of clay to catalyze the process of pressure solution is inherent in the crystallographic structure of the phyllosilicate minerals. The most common phyllosilicate minerals found in the stylolite seams at Overfold Mountain are muscovite and chlorite. The muscovite and chlorite have likely formed by the recrystallization of smectite clay, a transition which typically occurs at depths of burial less than 4 km (Weaver, 1959).

The common structure of these layered minerals is a variation on the 2:1 phyllosilicate structure (Figure 24). Each of the layers of the phyllosilicate is made up of one sheet of octahedrally coordinated cations (Al$^{3+}$, Fe$^{3+}$, Fe$^{2+}$ and Mg$^{2+}$), sandwiched between two sheets of tetrahedrally coordinated cations (Si$^{4+}$ and Al$^{3+}$). Substitution between cations of unequal valence state in these sheets creates a charge deficit which results in a negative surface charge on the layers (Pauling, 1930). This surface charge is balanced by the cations within the interlayer. Each of the phyllosilicate minerals differs in their degree of isomorphous substitution and consequently in the amount of surface charge on the interlayer. Each mineral also differs in the manner in which this charge is neutralized.

The unit cell of muscovite has a layer charge of $-2$ e$^-$. This charge is largely due to the substitution of Al$^{3+}$ for Si$^{4+}$ in the tetrahedral sheet and is neutralized by potassium ions (K$^+$) within the interlayer (Figure 24). Potassium is especially well suited to the muscovite
Figure 24 - Crystal structure of muscovite, chlorite, and smectite.
structure in that it fits very snugly into the space created by the six coordinated tetrahedra adjacent to the interlayer. As a result of this snug fit, and the relatively low bond strength between K\(^+\) and H\(_2\)O, muscovite has a relatively low cation exchange capacity and would likely not be as important in aiding diffusion during pressure solution as other, more reactive clay minerals.

The chlorite unit cell has a layer charge similar to that of muscovite (Dixon and Weed, 1977; p.342) again due to the substitution of trivalent cations for Si\(^{4+}\) in the tetrahedral sheet. This negative charge is neutralized by a positively charged interlayer hydroxide sheet (Figure 24). The cation exchange capacity of chlorite is relatively low, indicating that it is probably of little importance in exchange reactions.

Chlorite and muscovite are a common product of the recrystallization of smectites. The smectite structure is well suited to pressure solution theories which infer that clay minerals have provided a pathway for fluid movement and the diffusion of ions through the system (Weyl, 1959; Geiser and Sansone, 1981). The 2:1 layer structure is similar to muscovite and chlorite, but isomorphous substitution in both the octahedral and tetrahedral sheets results in a somewhat lower layer charge (-0.66 e\(^-\) per unit cell). This charge is neutralized by hydrated cations within the interlayer (Figure 24). These hydrated cations are readily exchanged, and diffusion of cations along this interlayer occurs
relatively easily. The low layer charge also allows smectite layers to be pulled apart under small tensile stresses, a factor which may aid bulk fluid flow along the interlayer.

Broken bonds on the edges of the clay structure cause the exposure of OH\(^-\) groups from the tetrahedral and octahedral sheets. These OH\(^-\) groups readily donate their protons, causing the clay to behave as a Lewis acid. This may in turn increase the dissolution rate of the carbonate minerals according to the reaction:

\[
\text{CaCO}_3 + 2\text{H}^+ \rightarrow \text{Ca}^{2+} + \text{CO}_2 + \text{H}_2\text{O}
\]

The negative charge at the broken edges may serve to attract cations toward the interlayer and thus may be effective in catalyzing the cation exchange reactions, a factor which would increase the rate of diffusion and push the above reaction toward calcite dissolution. The edge charge created at the broken edges is known as "pH-dependent charge" due to the effect of pH on the magnitude of charge. Low pH conditions will inhibit the dissociation of the OH\(^-\) groups (Dixon and Weed, 1977; p. 314) and decrease the rate of diffusion of cations. The catalysis of calcite dissolution by clays thus occurs by two processes. First, by the locally acidic conditions created at the broken smectite edges. Second, by the increased rate of diffusion of the products of dissolution away from the pressure solution site.
The presence of euhedral quartz within the stylolite seam may be explained by precipitation of quartz as overgrowths during stylolite formation. The locally acidic conditions within the stylolite seam may be responsible for quartz precipitation, as quartz solubility decreases with a decrease in pH.

A slightly different condition is involved in the development of pressure solution in quartz arenite. Since the solubility of quartz is low below a pH of approximately 9, the acidic nature of clay minerals would be expected to inhibit pressure solution. Investigators have found however, that clay minerals do enhance pressure solution in quartzites (Heald, 1956; Weyl, 1959; De boer, 1977; Wanless, 1979; Marshak and Engelder, 1985; Engelder and Marshak, 1985). This enhancement is likely mechanical in origin, the clays providing a fluid film for the diffusion of ions. Unlike calcite, the dissolution of quartz is known to be a slow process, even under ideal conditions (high pH and high temperature). This factor, as well as the effect of clays, would explain the contrast in the response of the two minerals to pressure solution.
4. **METHOD OF DATA COLLECTION AND ANALYSIS**

A folded surface is defined by three orthogonal axes; a, b, and c; where c is perpendicular to bedding and b parallels the axis of folding (Figure 25; Hancock, 1985; Price, 1967). Stylolites, shear fractures, and semi-brittle shear zones each define their own set of kinematic axes (a, b, and c). These minor axes have a unique relationship to the megascopic reference axes, and are the basis in the determination of the strain history induced in the rock as folding progressed. The method of analysis of each of these structures follows.

Field work included the observation and measurement of mesoscopic structures found in each lithologic unit. Several stations along the limbs and hinge regions of the folds were chosen so that any variations in strain and deformational mechanism might be determined. Oriented samples for thin section analysis of microstructures were also collected at each station. Station locations, outcrop patterns of lithologic units, and structural data have been plotted on a 1:5000 topographic map (Map 1).

4.1. **EXTENSION FRACTURES**

The close relationship between hydraulic fracturing and the permeability of the rock has been discussed in Section 3.2.1. The magnitude of each fracturing episode may be used as an indicator of the relative permeability of the
Figure 25 - Three orthogonal axes described by a folded surface. The c-axis is perpendicular to bedding; the b-axis parallels the axis of folding. After Hancock (1985), and Price (1967).
rock over that interval. In other words, a higher density of fractures suggests greater permeability.

Fractures with different orientations often cross-cut each other, revealing their relative ages. This information can be utilized, along with the kinematic information inherent in each fracture, in the development of a kinematic history of the rock as well as a permeability history.

At the outcrop, fractures were grouped into sets on the basis of parallelism and cross-cutting relationships. The density of the fractures in each set was then noted as well as the thickness, type, and habit of material filling the fractures. Where possible, extension fractures were categorized as syntaxial or antitaxial.

4.2. **SHEAR FRACTURES AND FAULTS**

The classification of shear fractures and faults was based on the presence of slickenside striae or visible offset. The orientation of the fracture surface and, where possible, the slickenside striations were recorded.

A slickensided fracture represents a unique set of kinematic axes (Figure 26).

*a-axis* - The line of the slickenside striae.

*b-axis* - The axis of rotation of the slip, the line which is perpendicular to the slickenside striae in the plane of the fracture. If shearing is concordant with fold development, this axis will parallel the megascopic fold axis.

*c-axis* - The line normal to the fracture surface. Together, the *c*-axis and *a*-axis define the plane of deformation (the *ac* plane of the fold)
Figure 26 - Kinematic axes described by a slickensided shear fracture. For explanation see page 59. After Price (1965).
This data, in conjunction with other kinematic indicators is an invaluable tool in the analysis of the kinematics of the rock.

4.3. **SEMI-BRITTLE SHEAR ZONES**

In this context, shear zones are zones of en echelon, often sigmoidal, extension fractures which form within zones parallel to the theoretical shear direction. The sense of shear on these zones may be determined by the orientation of the fractures with respect to the zone (Figure 27). The kinematic axes defined by shear zones parallel those defined by slickensided shear fractures. The \( c \)-axis is the normal to the zone, the \( a \)-axis lies in the plane of the zone, in the direction of shear. The line of intersection between the shear zone and the associated extension fractures represents an axis of rotation, or \( b \)-axis, for the zone. If shear zone development was concordant with fold development, this axis of rotation should parallel the megascopic fold axis.

4.4. **STYLOLITE SEAMS**

Stylolites describe a plane of dissolution, the normal to which represents a kinematic axis of compression (\( c \)-axis; Figure 28). The stylolites within the study area commonly have a ridge morphology, these ridges may also be used as kinematic indicators. The ridge crests appear to parallel the direction of fluid movement along the stylolite seam, this direction parallels the direction of minimum
Figure 27 - A) An array of shear zones within limestone (Unit 2). B) Kinematic axes described by a semi-brittle shear zone. The c-axis is the normal to the zone; the b-axis is the line of intersection between the zone and the extension fractures within the zone; the a-axis is the direction of shear on the zone.
Figure 28 - Kinematic axes described by columnar and ridge stylolites. The c-axis is normal to the stylolite seam, or parallel to the stylolite column when columns are oblique to the seam; the a-axis parallels the ridge crests in ridge stylolites.
compression (direction of minimum strain) along the seam at the time of stylolite development (kinematic $a$-axis). A third kinematic axis is represented by the line perpendicular to the ridge crests within the plane of the stylolite ($b$-axis). This axis is expected to parallel the megascopic fold axis if concordant with folding.

On the mesoscopic scale, stylolite seams were separated into sets based on their orientation and cross-cutting relationships. Densities of stylolite sets were measured and, where possible, the age relationships between stylolites and fractures were noted. The amplitude of the stylolite ridges and the thickness of the insoluble residue within the seams were recorded to aid in the determination of the amount of material removed by dissolution.
4.5. DATA COLLECTED - MICROSCOPIC SCALE

4.5.1. EXTENSION FRACTURES AND PRESSURE SOLUTION

Structures visible on the mesoscopic scale, in particular extension fractures and stylolite seams, are also found on the microscopic scale. The orientation of each of these microstructures was determined by universal stage analysis of three mutually perpendicular oriented thin sections of each sample. Microstructures were grouped into sets on the basis of orientation and cross-cutting relationships. The linear density, width, and type of fill were noted for each set.

Unit 1, the quartz arenite unit, showed some flattening of quartz grains due to pressure solution. The orientations of the flattened contacts were measured on the universal stage.

Unfilled microfractures were analyzed by fluorescence microscopy. Rock slabs were impregnated under vacuum with a low viscosity epoxy resin (Epotek 301) containing the fluorescent dye Rhodamine B. Thin sections of the slabs were then viewed in blue light. Microfractures were easily recognized due to the fluorescence of the dye within them.

4.5.2. TWIN LAMELLAE IN CARBONATES

Where possible, the density of twin lamellae in calcite and dolomite grains and cement was examined. Although this method is not a quantitative indicator of strain within the
rock, it does give an indication of the importance of this mechanism in the deformation of different lithologic units.

4.5.3. DEFORMATION LAMELLAE IN QUARTZ

In the units containing quartz grains, the percentage of quartz grains containing deformation lamellae was determined. This was done as an estimation of the importance of this mechanism in the accommodation of strain during folding.
5. **UNIT DESCRIPTIONS AND OBSERVATIONS**

Variation in lithology both between and within individual formations has allowed ample opportunity for the selection of distinct lithologic units. The miogeoclinal sediments found near Overfold Mountain include limestone, dolostone, sandstone and shale, each with a unique grain size, bedding thickness, and diagenetic history. Chosen for this study were seven limestone units, two dolostone units, and one sandstone unit.

5.1. **LITHOLOGIC UNITS**

**Rocky Mountain Group**

Unit 1 - Quartz arenite. Tan and pale grey weathering (5Y8/1), light to medium grey (N7), fine to medium grained calcareous quartz arenite; massive to medium bedded with minor thin cherty beds; Unit 1 has a moderate to low resistance to weathering; thickness is approximately 100 m. Two stages of cementation are seen in thin-section (Figure 29). Early quartz overgrowths have been followed by a later, sparry calcite cement which is seen to fill pore spaces and replace quartz grains as well as overgrowths.

**Mount Head Formation - Lower Carnarvon Member**

Unit 2 - Limestone. Medium grey weathering (N8), medium brownish-grey (5Y4/1), partially micritized skeletal grainstone (Figure 30); contains rusty
Figure 29 - Photomicrograph of Unit 1, calcareous quartz arenite. Two stages of cementation are seen. Early quartz overgrowths (O) are replaced by sparry calcite (C). Long edge of photograph measures 0.06 mm.
Figure 30 - Photomicrograph of partially micritized skeletal grainstone of Unit 2. Point contacts are common between micritized grains. Cement is sparry calcite. Long edge of photograph measures 0.06 mm.
weathering medium dark grey chert nodules up to 2 m in length; abundant large brachiopods and corals. This unit has a high resistance to weathering, and forms cliffs. Thickness is approximately 9 m.

Unit 3 - Dolostone. Light grey, tan and orange weathering (5Y6/1), medium-dark grey (5Y4/1), massive, finely crystalline dolostone (Figure 31); contains abundant white to dark-grey chert nodules. Unit 3 has a low resistance to weathering and forms a recessive trench between Units 3 and 4. Thickness is approximately 6 m.

Unit 4 - Limestone. Medium grey weathering (N7), medium grey-brown (5Y3/2), massive lime mudstone to wackestone (Figure 32); tan weathering, white to medium-grey weathering elongate chert nodules up to 10 cm. in length are common. Unit 4 has a great resistance to weathering and forms cliffs. Thickness is approximately 2 m.

Unit 5 - Dolostone. Light grey-tan and orange weathering (5Y6/1), dark grey (5Y5/2), massive finely crystalline dolostone (Figure 33). Unit 5 has a low resistance to weathering as did Unit 3. Thickness is approximately 4 m.

Unit 6 - Limestone. Light olive-grey weathering (N5), dark brown to dark grey (N2), micritized skeletal grainstone (Figure 34); grain size ranges from
Figure 31 - Photomicrograph of the finely crystalline dolostone of Unit 3. Dark grains are fine grained frambooidal pyrite. Long edge of photograph measures 0.06 mm.
Figure 32 - Photomicrograph of skeletal wackestone of Unit 4. Skeletal fragments are sparry and are suspended in a micritic matrix. Long edge of photograph measures 0.3 mm.
Figure 33 - Photomicrograph of the finely crystalline dolostone of Unit 5. Dark grains are fine grained framboidal pyrite. Long edge of photograph measures 0.06 mm.
Figure 34 - Photomicrograph of the micritized skeletal grainstone of Unit 6. Dark areas are micritized skeletal fragments. Light areas are sparry calcite cement. Long edge of photograph measures 0.3 mm.
less than 10 μm to 5 mm; modal grain size is approximately 0.5 mm. Unit 6 is highly resistant to weathering and is one of the main cliff forming units in the study area. Thickness is approximately 20 m.

Mount Head Formation - Marston Member

Unit 7 - Limestone. Medium grey-brown weathering (N5), dark grey (N2), coarse grained skeletal packstone to grainstone; contains abundant fossils including rugosan corals, crinoids, brachiopods and bryozoans; this unit is poorly sorted and has a modal grain size of approximately 2 mm; isolated dolomite crystals are found within the matrix of packstone samples; shaly beds occur near the stratigraphic base; this unit is moderately resistant to weathering and has a thickness of approximately 15 m. The percentage of matrix in this unit is seen to increase toward the northeast in the map area. The most southwesterly exposure of the unit is at the hinge of the anticline where it is a skeletal grainstone (Figure 35A). To the northeast, along the fold limbs, the unit is seen to be a packstone (Figure 35B).

Mount Head Formation - Loomis Member

Unit 8 - Limestone. Medium grey weathering (N5), medium grey (N5), massive, oolitic and skeletal
Figure 35 - Photomicrographs from Unit 7. A) At the hinge: Unit 7 is a skeletal grainstone with sparry calcite cement and point contacts between grains. B) In the limbs, northeast of the hinge, Unit 7 is a skeletal packstone with sutured contacts between grains. Long edge of photograph is 0.3 mm for (A) and 0.06 mm for (B).
grainstone (Figure 36); grains are typically partially micritized; the grain size mode is approximately 2 mm and grains are moderately well sorted; calcite cement is sparry and the point contacts between grains suggests that cementation occurred during early diagenesis, before significant compaction. This unit is highly resistant to weathering; the combined thickness of Units 9 and 10 is greater than 30 m.
Figure 36 - Photomicrograph of the oolitic and skeletal grainstone of Unit 8. Long edge of photograph measures 0.3 mm.
6. OBSERVATIONS ON THE MEGASCOPIC SCALE - HINGE STYLE

Lithologic units were chosen on the basis of mechanical properties as well as lithology. Each unit is mechanically unique, not only due to lithologic differences, but also due to physical properties such as bed thickness. While one unit was selected because it exhibited an unusual hinge style, another was chosen due to an unusual high or low density of certain mesoscopic structures. These features are controlled by the deformational processes which have acted on the rock at the granular scale, but are inextricably linked to the mechanical integrity of the entire package of folded rock. The unique mechanical properties of each of these lithotypes is displayed throughout the study area on all scales of observation.

On the megascopic scale, variations in mechanical properties are most readily viewed at the hinges of folds, where the most strain has been accommodated. Each unit is seen to have a unique hinge style (Section 5.2.1.). When the rocks are observed on the mesoscopic scale, fractures and stylolites are seen to be common. Some units show high densities of these structures whereas other, adjacent units, do not, indicating that the two units have accommodated strain in different ways. In other words, the strain has been partitioned differently between the two units. This strain partitioning is best observed on the microscopic scale, where it can be seen that each unit is distinct in the way in which strain has been partitioned between
intragranular and intergranular mechanisms. A comparison of the densities of the microscopic and mesoscopic structures aids in the determination of the mechanisms of deformation that have caused the folding of each unit.

6.1 **HINGE STYLE CONTROLS**

In the study area, hinge style appears to be governed by a few important properties of the individual units as well as properties of nearby units.

1. The mineralogy of the unit. Some units behave relatively ductilely and are able to bend around the hinge without a significant loss of cohesion. This category includes most of the limestone units (2, 6, 7, and 8). Other units lack ductility and are more likely to show shearing at the hinge. This category includes the dolostone and sandstone units (Unit 1, 3 and 5).

2. The thickness and massiveness of the unit. Thick massive units have no internal slip surfaces and often have broad, open hinges. These units appear to exert control over the shape of the fold, while other, thinner units conform to the shape prescribed by these thick, controlling units. This category includes some of the limestone units (Unit 6 and 8).

3. Proximity to nearby thick, massive units. This category includes limestones and dolostones of moderate thickness. Units overlying massive units will either follow the open hinge style (Unit 7) or will form a tighter hinge by slip
along the bedding surfaces or shaly beds (Unit 2, 3, 4 and 5).

6.2 HINGE STYLE - OBSERVATIONS

UNIT 1 - Quartz arenite.

The hinge of Unit 1 has a close chevron geometry with an interlimb angle of approximately 70° at the hinge (Figure 37). There is a great deal of shearing at this hinge, more so than at the hinges of the carbonate units. Limited exposure of this unit has not allowed thickness measurement.

Units 2 through 6 represent a continuous stratigraphic sequence:

UNIT 2 - Limestone.

Unit 2 has an open, circular hinge with an interlimb angle of approximately 90°. No loss of cohesion is apparent at this hinge (Figure 38). There has been no significant change in the thickness of this unit throughout the fold.

UNIT 3 - Dolostone.

Unit 3 is sandwiched between Units 2 and 4 and consequently has a fold geometry which is intermediate between the two. The interlimb angle is approximately 80° at the hinge. This unit shows much more fracturing than Unit 2 likely due to lithologic differences (Figure 38). There has been no significant change in thickness of this unit throughout the fold.
Figure 37 - Close chevron geometry of the hinge of Unit 1.
Figure 38 - Open, circular hinge of Units 2 and 3.
UNIT 4 - Limestone.

The hinge region of Unit 4 has a close chevron geometry with an interlimb angle of 65°. There is a great deal of shear fracturing along the hinge, likely due to the tight geometry. The tightness of this fold hinge is related to the decollement at the base of Unit 5 (Figure 39). There has been no significant change in the thickness of this unit throughout the fold.

UNIT 5 - Dolostone.

At the base of Unit 5 is a shale bed which has acted as a surface of decollement between Unit 5 and Unit 6. The hinge of Unit 5 has a close chevron geometry with an interlimb angle of 60°. The shale bed is discontinuous within the hinge suggesting that it has been locally thinned and thickened to adjust to the room problem caused by folding (Figure 39). There is a thickening of this unit at the hinge, mainly due to shear fracturing.

Unit 6 - Limestone.

Unit 6 is a thick, massive unit with an open circular hinge. The interlimb angle is approximately 90° at the hinge. As mentioned, the top of this unit is a surface of decollement, and the units above have significantly tighter hinges. There has been no apparent change in the thickness of this unit throughout the fold.
Figure 39 - Close chevron geometry of Units 4 and 5.
Unit 7 - Limestone.

Unit 7 is part of the well-bedded Marston Member sediments. The interbedded shales of this Member and the well bedded nature of the sediments has added to the ease of folding. In general the units of this Member conform to the dictates of the thick controlling units (Unit 8 and Unit 6). The hinge at Unit 7 is broad and has an open circular geometry with an interlimb angle of approximately $105^\circ$ (Figure 40). There has been no significant change in the thickness of this unit throughout the fold.

Unit 8 - Limestone.

Unit 8 belongs to the very thick and massive Loomis Member. This unit has exerted a large amount of control over the fold style of the anticline. The hinge is broad and has an open, circular geometry with an interlimb angle of approximately $110^\circ$ (Figure 43). There has been no significant change in the thickness of this unit throughout the fold.
Figure 40 - Open, circular geometry of hinge of Unit 7.
Figure 41 - Open, circular geometry of hinge of Unit 8.
7. **STRUCTURAL ANALYSIS, STRAIN PARTITIONING AND PERMEABILITY HISTORY**

7.1. **UNIT 1 - ROCKY MOUNTAIN FORMATION: QUARTZ ARENITE**

Unit 1 is unique in two respects. First, it is the only lithologic unit chosen which crops out at the hinge of the syncline within the study area. Second, it is the only siliciclastic lithology chosen for this study. The deformational mechanisms which have accommodated folding in this sandstone unit are, for the most part, the same as those utilized in the folding of the carbonates. It is the way in which the strain has been partitioned between mechanisms that differs, and provides a basis for comparison.

Hydraulic and shear fractures are the prevalent mesoscopic structures seen in Unit 1. Slickensides are commonly found on shear fracture surfaces. Only one stylolite seam was observed, displaying the relative unimportance of pressure solution as a deformational mechanism in this unit. The most common structures observed on the microscopic scale are microfractures, deformation lamellae within quartz grains, and flattened to concavo-convex grain contacts resulting from pressure solution.

7.1.1. **EXTENSION FRACTURES**

Mesoscopic extension fractures within Unit 1 are commonly unfilled. When fracture filling is observed it is typically composed of calcite or quartz, usually not both.
These filled fractures are rarely seen to cross-cut each other, but when they do, quartz filled fractures are seen to pre-date calcite filled fractures. This suggests that there was a change in the fluid composition over the deformation interval from a silica rich fluid, reflecting the composition of the surrounding rock, to a carbonate rich fluid which may have originated from the underlying carbonate rocks. The quartz filled fractures tend to have a syntaxial form, whereas the calcite filled fractures have an antitaxial form.

Change in fluid composition over the deformation interval is supported by the apparent cementation history of the sandstone. An early quartz cement forms overgrowths on quartz grains. Both the grains and the overgrowths have deformation lamellae, indicating that this cement was pre-deformational. The quartz cement has been partially replaced by a sparry calcite cement which is virtually unstrained, suggesting that its emplacement was post-deformational (Figure 42).

The shortage of cross-cutting relationships in Unit 1 has made the determination of the timing of the different episodes of extension fracturing impossible. However, some assertions can be made from fracture orientation and density. First, there is one fracture group which is found in all parts of the syncline as well as the anticline (Figure 43). This group has an orientation which is approximately normal to the megascopic fold axis. Due to
Figure 42 - Photomicrograph of the two stages of cementation in Unit 1. Post deformational sparry calcite cement (C) locally replaces early quartz overgrowths (O). Long edge of photograph measures 0.06 mm.
A) Poles to Mesoscopic Extension Fractures

Contour interval = 3 sigma (Kamb Method)

Figure 43 - Stereonet plots of mesoscopic and microscopic extension fractures within Unit 1. Solid great circle represents bedding; dashed great circle represents the ac-plane of the syncline.

B) Poles to Microfractures

C) Sample Locations and Cross-section

Figure 43 - Stereonet plots of mesoscopic and microscopic extension fractures within Unit 1. Solid great circle represents bedding; dashed great circle represents the ac-plane of the syncline.
the fortuitous orientation, this fracture has not been rotated from its original orientation, regardless of the timing of its development. Direct correlations can thus be made between the fold limbs. Cross-cutting relationships in Units 2 through 8 show that this fracture set formed in two pulses. The first pulse was early in the kinematic history of the rock, predating most fracture sets throughout the study area. These fractures are quartz filled in Unit 1 and may be pre-tectonic in origin or may have formed at a very early stage of folding. A second pulse occurred late in the fold history, these fractures are among the youngest in the study area. Fractures with this orientation are common in buckle folds and are believed to represent fold axis parallel extension resulting from the room problem created during folding.

Mesofracture sets which are perpendicular to bedding are found at most locations in the fold. It is possible that some of these fractures were formed during sedimentary loading, prior to deformation, although the obvious decrease in these fractures toward the fold hinge would tend to suggest that most were syn-deformational.

Another mesoscopic fracture set of interest is sub-parallel to bedding (Figure 43). If these fractures are concordant, then they were formed prior to, or at an early stage of the folding.

There is little variation in the densities of mesoscopic fractures between the fold limbs. There is
however a distinct decrease in the densities of these structures in all orientations at the fold hinge. There are several possible explanations for this phenomenon.

1. The permeability of the sandstone was low. Under this condition, fluids could not enter the rock, and pore pressure would not build up in cracks (strain hardening). As a result, shear failure instead of tensile failure would occur.

2. The fluid content of the sandstone was low. There would be no pore pressure effect on the stressed sandstone and thus very little tensile deformation.

3. The differential stress was too high at the hinge. This would result in the development of shear fractures at pore pressures lower than those required for tensile failure.

b) Microscopic Scale

Microfractures in Unit 1 are very short, typically only cutting through one or two quartz grains. Where the fractures have not been healed by quartz they are unfilled. Microfractures within the quartz grains typically have nucleated at grain boundaries and cut directly through grains, although commonly they have propagated along the contacts between grains and quartz overgrowths.

In general there is a very good correlation between the microfractures and the mesofractures. The variation in the orientations of microfractures in the different parts of the fold is illustrated in the stereonets in Figure 43B. Three trends are readily visible in the fold.

1. All parts of the fold show the fracture set which is sub-normal to the megascopic fold axis. The density of this fracture set decreases slightly toward the fold hinge.
2. There is a group of microfractures sub-parallel to bedding and a group of microfractures perpendicular to bedding in the fold limbs. These fractures are not found in the hinge.

3. There is a high density of microfractures which are perpendicular to bedding and parallel to the fold axis at the hinge, possibly due to extensional strain induced by the bending of the beds. This fracture is also found on the mesoscopic scale, only at the hinge.

The variation in the orientations of microfractures in the syncline suggests that the extensional direction at the time of folding varied throughout the fold. This observation is in accordance with computer simulations of folding by Dieterich (1970).

The total density of microfractures within Unit 1 is fairly constant throughout the fold. There is however, a variation in the density of individual fracture sets (Figure 43B). This suggests that microfracture development took place at a constant rate throughout the fold, and that microfracture orientations were governed by the local stress configuration.

In general, the correlation between the microfractures and mesofractures found in the sandstone unit is better than that found in the carbonate units. Microfracture orientation data from this unit also show less scatter than that from the carbonate units. Both of these observations are likely due to the lack of cleavage and twin planes in the quartz grains, structures which are known to act as loci for fracturing in the carbonates (Atkinson and Meredith, 1987). Microfractures found in quartz may thus
give a better representation of the true stress configuration at the time of their formation.

7.1.2. **SHEAR FRACTURES**

Shear fracturing appears to have been the most important mechanism for the accommodation of strain within Unit 1. The density of shear fractures in this unit is considerably higher than that of the underlying carbonate units, likely due to the greater plasticity of the carbonates. The importance of this deformational mechanism is illustrated by the increase in the density of coaxial shear fractures with proximity to the synclinal hinge.

Analysis of slickensided shear surfaces reveals two orientations for the axis of slip ($b$-axis). The predominant calculated $b$-axis is approximately parallel to the megascopic fold axis, as expected. In the overturned limb, a second $b$-axis has been calculated which is approximately $90^\circ$ from the megascopic fold axis. Slickensides on the latter shear surfaces approximately parallel the megascopic fold axis and are non-coaxial with folding (Figure 44). For reasons that are not clear, there is an increase in the abundance of non-coaxial shear surfaces with distance from the hinge (Figure 44). The presence of a non-coaxial phase of deformation is also demonstrated by the pressure solution contacts between quartz grains in this unit, as well as by the stylolite seams found in the underlying Etherington Formation (Section 7.1.4) and Mount Head Formation (Section 7.2 - 7.4).
a) Poles to shear fractures.

b) Kinematic axes for shear fractures with slickensides.

Figure 44 - Mesoscopic shear fractures within Unit 1. Dashed great circle represents the megascopic ac plane, solid great circle represents bedding.
7.1.3. **SEMI-BRITTLE SHEAR ZONES**

Shear zones are not abundant in Unit 1. Where they are observed (locations D and G from Figure 43), they follow the same shear trends observed in the nearby shear fractures (Figure 45).

1. Location D: two shear zones are found which define an axis of rotation (b-axis) parallel to the megascopic fold axis. This is the same shear geometry defined by the shear fractures at this location.

2. Location G: three shear zones define an axis of rotation (b-axis) which is non-coaxial with folding. This axis parallels the axis of rotation determined by the shear fractures at this location. No coaxial shear zones or shear fractures were found here.

7.1.4. **PRESSURE SOLUTION**

Pressure solution has been active throughout this quartz arenite unit, albeit of relatively little importance in the accommodation of strain during folding. Little evidence of pressure solution is visible on the mesoscopic scale. On the microscopic scale however, grain contacts are seen to have been flattened by pressure solution. This has affected both grains and quartz overgrowths, showing that dissolution occurred after quartz cementation. Calcite cement, on the other hand, post-dates pressure solution. Subhedral calcite crystals are often found along flattened grain contacts.

A crude fabric has developed by the flattening of the grains. In the fold limbs the fabric is very poorly
Pole to the shear zone (c-axis)

Slip linear of the shear zone, arrow gives the sense of shear on the upper surface of the zone.

Axis of rotation of the shear zone (b-axis), represents the line of intersection of the shear zone and the enclosed extension fractures.

Direction of shear on zone (a-axis), represents the line normal to the b-axis in the plane of the shear zone.

Megascopic fold axis

Figure 45 - Equal area projection of the mesoscopic shear zones within Unit 1. The great circle represents the megascopic ac plane.
Figure 46 - Poles to flattened contacts between quartz grains. Equal area projection, contour interval equal to 3 sigma (Kamb method). Dashed great circle represents the ac plane of the megascopic fold. Solid great circle represents bedding. For locations see Figure 7-2.
developed and parallels bedding. This suggests that its origin may be pre-tectonic compaction (Figure 46; Location A and G). At the hinge, however, the fabric is oblique to bedding and much better developed. Here it defines an event of non-coaxial compression which parallels the megascopic fold axis (Figure 46; Location D). The presence of this fabric at the hinge suggests that it formed after fold development, and that the tectonically enhanced permeability at the hinge caused pressure solution to be localized there.

Carbonate rocks belonging to the underlying Etherington Formation are exposed along the limbs of the syncline within the study area. Although these rocks are not among the lithologic units chosen for this study, structural data was collected which gives some insight into the deformational history of this fold. The carbonates within this Formation here, developed stylolite seams far more readily than the sandstone of Unit 1. Analysis of these seams, and the corresponding stylolite ridges, supports the deduction that there has been a post-folding event of compression sub-parallel to the megascopic fold axis.

There are two well developed sets of stylolites found in both limbs of the syncline (Figure 47). The first, and most abundant, is parallel to bedding. A second group lies in a plane which is sub-normal to the megascopic fold axis. Ridges along the first set of stylolites defines a $b$-axis which parallels the megascopic fold axis (Figure 47). Parallelism to bedding suggests that the stylolite
Northeast Limb

Southwest Limb

Figure 47 - Equal area projection of mesoscopic stylolite seams within the Etherington Formation. The dashed great circle represents the megascopic ac plane. The solid great circle represents bedding. Contour interval = 3 sigma (Kamb method).
originally nucleated during sediment loading and continued to grow over the deformation interval. The latter set delineates a non-coaxial b-axis, approximately 90° from the fold axis, sub-parallel to the non-coaxial b-axis determined by the shear fracture analysis. The fact that the non-coaxial stylolite set parallels the non-coaxial plane of flattening seen at the hinge of Unit 1, suggests that these features are contemporaneous. If so, this stylolite set was formed after fold development and implies that there was a relatively large post-deformational influx of fluid during an interval of fold-axis-parallel compression.

7.1.5. **DEFORMATION LAMELLAE IN QUARTZ**

There is an increase in the number of quartz grains containing deformation lamellae at the hinge of the syncline (Table 1). This indicates that the rocks are more highly strained in the hinge than in the limbs of the fold. Higher strain at the hinge is expected in folds with a chevron geometry such as this.

7.1.6. **TWINSING IN CALCITE CEMENT**

The calcite crystals that form the second stage cement in this unit are virtually untwinned (Figure 42). This indicates that this cement was crystallized after the deformational episode.

7.1.7. **SUMMARY – UNIT 1**

The obvious increase in the density of coaxial shear fractures at the synclinal hinge shows that this is the most
important deformational mechanism in the folding of sandstone near Overfold Mountain. Extension fractures have a marked density decrease at the hinge region on the mesoscopic scale, and no variation on the microscopic scale. Coaxial pressure solution was of minor importance and has little or no variation across the fold hinge (Table 1). All of these factors suggest a low porosity or fluid deficiency within the sandstone at the time of folding.

<table>
<thead>
<tr>
<th>TABLE 1 - DENSITIES OF MESOSCOPIC AND MICROSCOPIC STRUCTURES</th>
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<tbody>
<tr>
<td>Extension fracture</td>
</tr>
<tr>
<td>Mesoscopic(^1) (per m)</td>
</tr>
<tr>
<td>Microscopic(^1) (per cm)</td>
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<tr>
<td>Microscopic(^1) (per cm)</td>
</tr>
<tr>
<td>Shear Fractures</td>
</tr>
<tr>
<td>Coaxial (per m(^2))</td>
</tr>
<tr>
<td>Non-coaxial (per m(^2))</td>
</tr>
<tr>
<td>Stylolite Seams</td>
</tr>
<tr>
<td>0</td>
</tr>
<tr>
<td>Flattened Grain Contacts</td>
</tr>
<tr>
<td>Coaxial</td>
</tr>
<tr>
<td>Non-coaxial</td>
</tr>
<tr>
<td>Deformation Lamellae(^2) (per cm)</td>
</tr>
<tr>
<td>Calcite Twin Lamellae (in cement)</td>
</tr>
</tbody>
</table>

\(^1\) Combined linear densities of all fracture sets.
\(^2\) Linear density of quartz grains with lamellae.

It is apparent from the structural data presented that there was a major shift in the stress configuration after the development of the syncline at Overfold Mountain. This conclusion has been drawn by the stress analysis of
slickensided shear fractures, shear zones and flattened contacts between quartz grains, as well as stylolites within the underlying Etherington Formation.

1. Analysis of stylolite data and flattened grain contacts suggests that late in the deformational history, the maximum compressive stress was sub-parallel to the megascopic fold axis. The corresponding intermediate axis of compression (b-axis) is determined to be toward the northeast.

2. Analysis of shear fracture data delineates two orientations for an axis of rotation (approximately equal to the intermediate compressive stress). The most common orientation is parallel to the megascopic fold axis, as expected. However, at some time(s) in the deformational history this axis appears to have switched to a position approximately 90° from this fold axis, toward the northeast.

3. Analysis of extension fractures indicates that at several times in the deformational history, the minimum compressive stress was sub-parallel to the megascopic fold axis. This phenomena is coaxial with folding and is commonly observed in flexural-slip folds.

There appears to have been an influx of fluids into the system at the time of the latter, non-coaxial, episode. This is evidenced by the increased incidence of pressure solution, both in Unit 1 and in the underlying Etherington Formation carbonates. Pressure solution in Unit 1 is localized at the synclinal hinge, possibly due to the enhanced permeability there. This enhanced permeability may also explain the absence of non-coaxial shear fracturing at the hinge.
7.2. **UNITS 2 THROUGH 6 - UPPER CARNARVON MEMBER**

Units 2 through 6 represent a continuous stratigraphic sequence of interbedded limestone and dolostone within the Lower Carnarvon Member of the Mount Head Formation. Each of these units has a unique aspect that has caused it to behave differently from the others during folding. Evidence for this is seen throughout the fold as a variation in the densities of fractures, stylolites and shear zones in adjacent units.

There are two factors which appear to have controlled the way in which strain has been accommodated within this sequence. First is lithology - limestones simply accommodate strain differently than dolostones. Second, and more important in the study area, is the thickness of the unit. Thick, massive units such as Unit 6 have controlled the style of folding of the thinner beds above and below.

Dolostone units are seen to have behaved differently than the adjacent limestone units. This is principally due to the contrast in the behavior of the mineral constituents, dolomite and calcite. Calcite is more soluble, and therefore will undergo pressure solution more easily than dolomite. Also, calcite forms deformational twin lamellae much more readily than does dolomite. As a result, limestone is inherently more ductile than dolostone, a factor which is evident on all scales of observation.
Vein densities may be used qualitatively as a guide to the permeability of the rock at the time of their formation (Section 7.2.1). Dolostones are seen to have a very high density of thin calcite filled or unfilled microfractures. Limestones, on the other hand, tend to have a low density of thick calcite filled fractures. This dissimilarity is a result of the different permeabilities of the rock at the time of fracture formation. A high fracture density marks deformation under conditions of low permeability. Under these conditions fluid escape is impossible, and pore pressure will build up at many discrete points within the rock. Microfractures formed at these points may remain unfilled, as fluids cannot migrate into them. These unfilled microfractures impart a secondary porosity to what would otherwise be a relatively impermeable rock and may be largely responsible for the fracture porosity found in many hydrocarbon reservoirs.

Structural trends found in the syncline are repeated in the anticline. Two distinct phases of deformation are found in the anticline, a coaxial phase and a non-coaxial phase. During the non-coaxial phase, the maximum compressive stress appears to have been sub-parallel to the megascopic fold axis. Cross-cutting relationships show that stylolites formed during the non-coaxial phase formed later than the coaxial stylolites. The higher density of non-coaxial stylolites at the hinge further enforces the belief that these structures are post-folding.
Cross-cutting relationships between vein and stylolite sets of different ages suggest that there may have been several episodes of fluid influx into the deforming system. A period of fluid influx is marked by the presence of a young, well developed structure which cuts an older, poorly developed, low density structure. Fractures formed during these events are coaxial and do not have a consistent orientation. Thus, they cannot be related to a single episode of fluid influx. Instead, they appear to have formed continuously over the interval of deformation.

7.2.1. EXTENSION FRACTURES

Extension fractures within Units 2 through 6 are typically calcite filled. Calcite crystals within these fractures have two forms: fibrous and blocky. Fibrous calcite, the growth form, is a valuable kinematic indicator (Chapter 4). Unfortunately, most of the calcite within the fractures has been recrystallized to a blocky form which is useless as a kinematic indicator. Where fibrous calcite is observed in veins it is perpendicular to the vein wall and does not show any indication of progressive strain (Durney and Ramsay, 1977). Under these conditions, the pole to the vein surface represents the minimum compressive stress at the time of its formation, and the maximum and intermediate principal stresses lie within the plane of the vein.

Stylolites delineate two phases of deformation near Overfold Mountain (Section 7.2.4.). Extension fracturing associated with both of these phases is expected.
Unfortunately, it is impossible to separate the veins associated with the later phase, which was non-coaxial with folding, from those formed in the late stages of the coaxial phase. During the non-coaxial phase, the maximum compressive stress was sub-parallel to the megascopic fold axis and the intermediate stress appears to have been nearly horizontal and to the northeast. If veins were formed during this phase they would dip shallowly to the southeast, as do many of the veins associated with the coaxial phase. Cross-cutting relationships show no consistent late stage vein with this orientation.

Cross-cutting relationships between coaxial veins of different orientations describe a complex history of extension fracturing which locally differs from the ideal model. Some consistent fracturing is found however. The earliest fracture found in the study area is perpendicular to bedding and parallel the fold axis (Figure 48). This vein is believed to have formed prior to deformation, due to sediment loading. The earliest stylolites found are perpendicular to these veins (parallel to bedding), and likely formed at the same time. The second vein found with some consistency throughout the study area is parallel to bedding (Figure 48). It is believed that this vein formed during coaxial shortening just prior to the onset of folding. Stylolites associated with this vein are perpendicular to bedding and parallel to the fold axis. A third vein set is perpendicular to the fold axis. This set
Stage 1 - Sediment Loading

Stage 2 - The Onset of Tectonic Stresses

Figure 48 - Orientations of fractures and stylolites formed during compaction (Stage 1), and early stages of folding (Stage 2).
is very well developed throughout the fold. Cross-cutting relationships show that this set formed continuously over the interval of folding.

Total fracture density and bulk dilation for microfractures and mesofractures have been plotted for the different units in the different parts of the fold (Figure 49 and 50). There is a decrease in fracture density at the hinge and a corresponding increase in bulk dilation. This infers that the permeability of the rocks at the hinge was higher than that in the limbs during fold development. In rocks of high permeability, fluid pressure will not build up and few hydraulic fractures are nucleated. As a result, those which are nucleated will continue to propagate due to stress corrosion at the fracture tips (Section 3.2.1.). Also as a result of high permeability, fluids can readily migrate into the fractures as they open and the width of the fracture can grow in proportion to its length. These thick mesofractures are responsible for the accommodation of a great deal of extensional strain. In comparison, unfilled microfractures, such as those commonly found in the study area, accommodate very little permanent strain. The dilation associated with these fractures was elastic, and lasted only as long as the high stresses were in place.

Thin sections impregnated with fluorescent dye revealed no unfilled microfractures in the thick massive limestone units (Units 2 and 6). The dolostone units (Units 3 and 5) are seen to have an extremely high fracture permeability,
Figure 49 - Mesoscopic extension fracture density, microfracture density and bulk dilation plotted against location within the anticline, for two dolostone units within the Lower Carnarvon Member (Units 3 and 5). Bulk dilation has been estimated from the linear densities and thickness of all sets of extension fractures.
Figure 50 - Mesoscopic extension fracture density, microfracture density and bulk dilation plotted against location within the anticline, for three limestone units within the Lower Carnarvon Member (Units 2, 4, and 6). Bulk dilation has been estimated from the linear densities and thickness all sets of extension fractures. Fold locations refer to Figure 49.
Figure 51 - Thin section of Unit 3 impregnated with fluorescent dye. Fluorescence reveals fracture porosity along many grain boundaries not visible under plane light. A) Plane light; B) Blue light. Long edge of photograph measures approximately 0.06 mm.
Figure 52 - Thin section of Unit 5 impregnated with fluorescent dye. Fluorescence reveals abundant fracture porosity surrounding dolomite crystals and only limited fracture porosity within a remnant calcite clast. A) Plane light; B) Blue light. Long edge of photograph measures 0.06 mm.
Figure 53 - Thin section of Unit 4 impregnated with fluorescent dye. Fluorescence reveals fracture porosity along grain boundaries, especially around disseminated dolomite rhombs. A) Plane light; B) Blue light. Long edge of photograph measures 0.06 mm.
fluorescence surrounding nearly every grain (Figure 51 and 52). Unit 4, also limestone but thinner than Units 2 and 6, was found to have some unfilled microfractures, though not nearly to the extent of the dolostone units (Figure 53).

7.2.2. SHEAR FRACTURES AND FAULTS

Evidence for bedding parallel shear is found throughout the anticline in the form of slickensided bedding surfaces in the fold limbs and bedding parallel decollement at the hinge (at the base of Unit 5). Shear fractures which are oblique to bedding, however, are rare, both in the dolostones and the limestones in the Lower Carnarvon Member. This may result from the tendency for carbonates to accommodate strain by ductile processes (twinning and stylolites) such that the critical stress for shearing is not reached. Faults which are oblique to bedding, on the other hand, are relatively abundant. This would suggest that shearing at an angle to bedding is restricted to zones along which relatively large amounts of movement have taken place. Two phases of faulting are seen in these units. A phase which is coaxial with folding is common in the overturned limb of the fold (Figure 54). Slickensides on these fault surfaces reveal a $b$-axis which is sub-parallel to the megascopic fold axis. Normal faults, which represent a later, post-folding episode of extension, are evident in the vicinity of the hinge (Figure 55). These structures may have formed contemporaneously with late stage extensional
Figure 54 - Three common shear geometries for faults and shear fractures formed during folding near Overfold Mountain. Modified after Hancock (1984).
Figure 55 - Orientation of post-folding normal faulting near Overfold Mountain.
structures such as the Flathead Fault, located to the east. No faults or shear fractures were found in these units that correlate with the episode of fold axis parallel compression delineated by stylolites.

7.2.3. **SHEAR ZONES**

Semi-brittle shear zones are far more common in the limestones than in the dolostones near Overfold Mountain. This may be a function of the relatively low permeability and low ductility of the dolostone units. Both factors would cause the dolostones to accommodate shear strain by brittle mechanisms (unfilled microfractures and shear fractures) rather than semi-brittle mechanisms (shear zones).

Shear zone orientations (Figure 56) are consistent with the orientations of shear fractures and faults within the study area (Figure 54). All of the shear zones found are coaxial with fold development and appear to have formed at some time between the initiation of folding and the closure of the fold. These zones are locally abundant in the limbs of the anticline and absent at the hinge. Extension fractures within shear zones are either perpendicular to the fold axis (rare), or have a fracture to bedding angle between 45° and 60° (common). The stress configuration described by the latter zones suggests that they are associated with bedding parallel shear such that the maximum compressive stress was at an angle of approximately 45° to bedding at the time of their formation. This would have
Figure 56 - Semi-brittle shear zone orientations found within the limbs of the anticline at Overfold Mountain.
occurred at some time after the onset of folding, when the beds had been rotated to a position $45^\circ$ from the maximum compressive stress. In this orientation, the bedding planes would have been in the field of maximum shear stress, and a high percentage of strain would have been accommodated by bedding parallel shear. Since bedding planes at the hinge were never rotated into a position of maximum shear stress, shear zones did not form there.

7.2.4. PRESSURE SOLUTION

Stylolites within the Mount Head Formation have similar trends to those found in the overlying Etherington Formation. As with the overlying rocks, two phases of stylolite development are seen, a coaxial phase and a non-coaxial phase. In Units 2 through 6, the phase of pressure solution which is coaxial with folding is well developed throughout the fold whereas the non-coaxial phase is mainly found in the overturned limb and the hinge.

Several factors are responsible for the density variation between stylolites of different ages. Most important are lithology, permeability, and the amount of compressional strain the rock has been required to accommodate. An attempt has been made to determine the importance of this mechanism in the accommodation of compressional strain in the different lithologies, in different parts of the fold, and the implications for permeability and fluid flow over the deformation interval.
The amount of material removed along a stylolite can be estimated from the amplitude of the stylolite sutures. In some cases, the amplitude of the sutures can give a better estimate of the strain accommodated by a group of stylolites than the stylolite density. This is particularly true in lithologies of low permeability and low clay content such as Units 2 through 6. In rocks such as these, stylolite nucleation is inhibited by the lack of a diffusion pathway for the dissolved material. Once a stylolite has nucleated and a small amount of clay is concentrated within its selvage, the rate of diffusion along the seam is increased, and so the rate of dissolution is increased (Engelder and Marshak, 1985).

Evidence for clay catalysis of dissolution in stylolite formation is found throughout the study area as a variation in the densities and orientations of the stylolites of different ages. As the beds are rotated during folding, previously formed stylolites are moved to a position that is no longer normal to the maximum compressive stress. One might expect to find younger stylolites which have nucleated in response to the new stress field. In fact, this is rarely the case. Where younger stylolites are found, they are very poorly developed. There are two possible explanations for this phenomenon. First, it is possible that the rock was dry after the onset of folding and there were no fluids to initiate pressure solution. Second, and more likely, is that compressive strain was taken up by the
pre-existing stylolites. This would support the idea that
the activation energy required for the dissolution of
carbonates in the absence of clay is significantly higher
than that for the reaction which has been catalyzed by even
a small amount of clay (Engelder and Marshak, 1985; Geiser
and Sansone, 1981). As a result, rocks with a low
permeability and low clay content will tend to have a low
stylolite density, but the stylolites formed will have a
large suture amplitude. Since the stylolite sutures found
in the study area have a ridge structure rather than a
columnar structure, and since the ridges parallel the
ac plane of the fold; the sutures are perpendicular to the
seam and thus there is no kinematic marker to show that the
stylolites grew in an orientation which was oblique to the
maximum compressive stress (Figure 57).

Suture amplitudes from the stylolites found in the
limbs and hinge of the anticline have been contoured in
Figure 58. In the overturned limb, three concentrations of
stylolites with large sutures are evident. In order of
importance they are:

1. Parallel to bedding.

2. Perpendicular to bedding and parallel to the fold
axis.

3. Sub-normal to the fold axis.

Group 1 and 2 stylolites are coaxial with folding.
Group 3 is non-coaxial. The most common stylolite observed
is parallel to bedding (Figure 58 and 48). These
Figure 57 - Rotation of a bedding parallel stylolite relative to the principal stresses from nucleation through fold closure. Note that the minimum strain direction along the stylolite seam does not vary throughout rotation (B and C). This direction parallels the stylolite ridges as well as the proposed direction of fluid movement. A) Nucleation: a clay selvage developed along a bedding parallel stylolite during compaction. B) During folding: fluid movement along pre-existing clay selvage has begun to carve "ridges" along the stylolite seam. The ridge crests parallel the direction of fluid movement. C) Fold closure: ridge crests still parallel the minimum strain direction.
Figure 58 - Poles to stylolites within Units 2 through 6 in the overturned anticline near Overfold Mountain. Contour intervals represent 1 mm of amplitude on stylolite sutures. 1 = Bedding parallel stylolites; 2 = Stylolites parallel with fold axis and perpendicular to bedding; 3 = Stylolites perpendicular to the fold axis.
stylolites are believed to have nucleated during early compaction of the sediments. Stylolites perpendicular to bedding and parallel to the fold axis developed during a period of layer parallel shortening during the initial stages of folding (Figure 58 and 48). As the beds were rotated during folding, both of these sets of coaxial stylolites were active and accommodated the compressive strain required in the development of the fold; few new stylolites were nucleated because the critical stress required for nucleation was not reached. This would explain the absence of a stylolite set which is axial planer to the fold. Analysis of suture orientations of these coaxial stylolites reveals a kinematic b-axis parallel to the megascopic fold axis (Figure 59).

Group 3 stylolites are non-coaxial with respect to the folding of the beds. These stylolites are sub-normal to the fold axis and represent a period of compression which was sub-parallel to the fold axis. Cross-cutting relationships between coaxial and non-coaxial stylolites show the latter to be a younger structure thus the non-coaxial phase was post-folding.

Stylolite density and suture amplitudes for coaxial and non-coaxial stylolites have been plotted in Figure 60. Of interest is the inverse relationship between coaxial and non-coaxial stylolites at most stations. At the hinge, Unit 5 (dolostone) is seen to have a high density of coaxial stylolites and no non-coaxial stylolites. On the other
Northeast Limb

Hinge

Southwest Limb

Figure 59 - Kinematic axes of stylolite seams in Units 2 through 6 (within the Lower Carnarvon Member of the Mount Head Formation), in the overturned anticline near Overfold Mountain.
Figure 60 - Stylolite densities and suture amplitudes for Units 2 through 6, at the hinge of the anticline. A) Stylolites coaxial with folding. B) Stylolites formed post-folding (non-coaxial with folding).
hand, Units 3 (dolostone) and 4 (limestone) have a very low density of coaxial stylolites and a high density of non-coaxial stylolites. Units 2 and 6 (both limestones), the controlling units, have intermediate values of both. This infers that the permeability of the beds were altered by the deformational mechanisms which caused the folding, prior to the second phase of deformation. In order to ascertain the reasons for the variations in permeability, one must consider all of the structures which may have affected the permeability: veins, stylolites and unfilled microfractures. In the dolostone units (Units 3 and 5) there is a good correlation between microfracture density and non-coaxial stylolite density (Figure 61). Unit 3 has a large density of unfilled coaxial microfractures, structures which would form under low permeability conditions. Under these conditions, one would expect to find a low stylolite density, as is the case for the coaxial phase. The fracture porosity imparted into the rock prior to the non-coaxial phase of deformation is reflected in the high density of non-coaxial stylolites found at the hinge. Unit 5, on the other hand, has a lower density of microfractures, most of which are filled with calcite. This unit accommodated a great deal of extensional strain, mainly parallel to the fold axis. The relatively low density of unfilled microfractures in Unit 5 suggests that it had a high permeability at the hinge during the coaxial phase of deformation, but a low permeability during the non-coaxial
Figure 61 - Densities of microfractures (per cm) and two phases of stylolites (per m) at several locations within the anticline for two dolostone units within the Lower Carnarvon Member (Units 3 and 5).
Figure 62 - Densities of microfractures (per cm) and two phases of stylolites (per m) at several location within the anticline for three limestone units within the Lower Carnarvon Member (Units 2, 4, and 6). Fold locations refer to Figure 61.
phase of deformation. Therefore no non-coaxial stylolites were formed.

Units 2 and 6 are very similar in character. Both are thick, massive limestones that appear to have had similar permeability histories over the deformation interval. In both of these units, the shape of the density curves for the coaxial and non-coaxial stylolites are nearly parallel (Figure 62). This suggests that there was little strain induced change in permeability during folding. There is a fairly good correlation between microfracture density and non-coaxial stylolite density. This suggests that fracture permeability, imparted during folding, may have caused the development of the stylolites which were formed after folding.

Unit 4 is a relatively thin limestone bed sandwiched between the somewhat stiffer units (Units 2, 3, 5, and 6). As a result, it has been forced to accommodate these units, mostly by extension parallel to the fold axis. The high density of filled coaxial microfractures at the hinge suggests that this unit was relatively impermeable over the interval of folding. The very small amount of compressional strain accommodated by this unit at the hinge, suggests that it was mainly in an extensional regime and thus may have acquired a fracture permeability not visible by conventional petrographic microscopy. This was confirmed by fluorescence microscopy which shows abundant microfractures along grain boundaries and twin planes (Figure 53).
It can thus be observed through two stages of deformation that stylolite development was closely linked to permeability in the study area. Also, it becomes even more apparent that deformational mechanisms affect permeability, by lowering the permeability of permeable beds, and increasing the permeability of impermeable beds.

7.2.5. **TWINNING IN CALCITE AND DOLOMITE**

Twinned Calcite is found in the limestone units throughout the fold. The average linear density of calcite twins has been determined for each station. The density of these twins decreases with distance from the fold hinge (Figure 63). Some scatter of data is caused by the localized micritization of grains; in particular, Unit 2 has been diagenetically micritized at the hinge. The small grain size of the micrite has made twin measurement impossible thus the density of calcite twin lamellae appears to be low. In fact, the density of twin lamellae in the micrite is probably quite high due to the high density of dislocations at grain surfaces and the high surface to volume ratio.

Dolomite twinning is rare in the study area; none is seen at all at most stations. The dolostone units (Units 3 and 5) do show some dolomite twinning at the anticline hinge, where the rocks are the most highly strained. Figure 64 shows the density of the twins in dolomite as a function of distance from the anticline hinge. It should be noted that calcite crystals within veins in the dolostones
Figure 63 - Variations in the average density of calcite twin lamellae (per cm) as a function of location within the anticline for three limestone units (2, 4, and 6) of the Lower Carnarvon Member.
Sample Locations

Figure 64 - Variations in the average density of dolomite twin lamellae (per cm) as a function of location within the anticline for two dolostone units (Units 3 and 5) of the Lower Carnarvon Member of the Mount Head Formation.
do show twinning, the density of which increases with the age of fracture but decreases with distance from the hinge.
Unit 7 is a massive, very thick packstone within the Marston Member of the Mount Head Formation. Mechanically, the Marston Member as a whole is unique in comparison with the other members of this Formation. It is composed of interbedded limestone and dolostone, as is the Carnarvon Member, but the beds are typically thinner and there is a higher percentage of dolostone and shale. As a result of the large viscosity contrast between the beds, most of the strain associated with folding has been accommodated by the less competent dolostones and shales.

Rocks of the Marston Member are sandwiched between the very thick and massive limestone of the Loomis Member (Units 9 and 10) and the thick limestone unit at the base of the Carnarvon Member (Unit 6). These massive units have controlled the style of folding in the more thinly bedded rocks of the Marston Member.

The lithology of Unit 7 varies between outcrops at the hinge and those of the limbs. In the hinge region, this unit is a well cemented skeletal grainstone. Point contacts between grains suggest that cementation was an early diagenetic feature, long preceding deformation. In the limbs, however, the rock is a skeletal packstone. There is a general lack of cement, and pressure solution has affected all grain contacts. In the limb areas, Unit 7 is relatively rich in clays in comparison to the other units chosen for this study. Clay catalysis of pressure solution has been
operative, causing dissolution along grain boundaries and stylolite formation. The effect of clay catalysis of pressure solution has been discussed in Sections 3.2.2. and 7.2.4., and will not be detailed further here. Pressure solution is observed to be the main deformational mechanism in this unit. Clay catalysis has, in fact, been so effective as to all but negate the need for other deformational mechanisms. In particular, there is no evidence of shear fractures, faults or shear zones within this unit. Veins are found throughout the unit but occur at a considerably lower frequency than in the other carbonate units studied, suggesting that clay catalysed pressure solution has taken place under a lower effective stress than that required for all other mechanisms.

7.3.1. PRESSURE SOLUTION

Although pressure solution has been determined to be the most important deformational mechanism in the folding of Unit 7, there are some exceptions that need to be considered. In the limbs of the fold, almost every grain contact is sutured (Figure 35B). This is not the case at the hinge, where the unit is well cemented and grains often are seen to have point contacts (Figure 35A). The lack of sutured grain contacts at the hinge suggests that the calcite cement was emplaced early in diagenesis, before sufficient depth of burial was reached for the onset of pressure solution. In the fold limbs, however, the ubiquitous pressure solution parallel to bedding suggests
that the rock had a high permeability during compaction and thus stylolites were nucleated at many grain contacts.

All mesoscopic stylolites observed are sub-parallel to bedding, again emphasizing the importance of a pre-existing clay selvage in the catalysis of pressure solution. Stylolites oblique to bedding are found on the microscopic scale only. In the overturned limb of the fold, a set of micro-stylolites was seen which is perpendicular to bedding and parallel to the fold axis. This set is believed to represent the episode of shortening which took place during the initial stage of folding (Figure 48). A second set of micro-stylolites is seen at the hinge only. This set is sub-normal to the fold axis, and parallels the non-coaxial stylolites found in other parts of the fold.

Relationships between the three stylolite sets and the permeability of the rock at the time of their formation are similar to those observed in Units 2 through 6. Early cementation resulted in a low permeability in the rock which is now at the hinge of the fold. Few stylolites were developed either before or during folding. The low permeability during fold development caused microfracturing which increased the finite permeability of the rock. Fluids moving through the rock along fractures allowed pressure solution to take place during post-folding, fold-axis-parallel compression resulting in the development of micro-stylolites normal to this axis. In the fold limbs, on the other hand, this unit had a high permeability during
compaction and the onset of folding. This permeability was decreased to almost nil during folding, as pressure solution progressively closed existing intergranular pores. As a result, no late stage non-coaxial stylolites are found.

7.3.2. EXTENSION FRACTURES

Throughout Unit 7, extension fractures have acted as loci for the deposition of material which has been removed along stylolite seams. As has been the case elsewhere in the study area, the importance of hydraulic fracturing in the deformation of the folding beds is dependent upon the permeability at the time of deformation. The permeability of Unit 7 varied laterally prior to deformation such that the hinge area had a low permeability and the limbs had a high permeability. As a result, the importance of hydraulic fracturing has varied throughout the fold, not only due to variations in the amount of strain accommodated, but also due to lithologic variations.

Trends in the densities of microfractures and mesofractures (veins) observed in this unit (Figure 65) reflect the permeability of the rock. At the hinge, a high microfracture density, accompanied by a low mesofracture density expresses the low permeability there. The inverse is found at the limbs, where the permeability was probably high during folding.

When considering the strain implications of extension fractures, one must look not only at the fracture densities, but also at the amount of extension accommodated by each
Figure 65 - Average linear densities of microfractures and mesofractures (veins) within a limestone unit of the Marston Member of the Mount Head Formation (Unit 7). Locations are in the limbs and the hinge of the anticline near Overfold Mountain.
fracture. Although the density of unfilled microfractures is high in beds of low permeability, most of the strain accommodated by them at the time of their formation is elastic in nature, and the amount of permanent dilation is small. Mesofractures (veins), on the other hand, typically occur at a low density, relative to microfractures, but can accommodate large amounts of permanent dilation. Bulk dilation can be estimated for each observed fracture set if the average fracture width and the linear density of each set is known (Figure 66). By adding the bulk dilation of all of the fracture sets observed, one can quantitatively estimate a total bulk dilation accommodated by extension fracturing. This estimate is useful in two ways. First, it can be used to compare the importance of extension fracturing between rocks with varying degrees of strain. It can also be used as an estimate of the volume increase associated with extension fracturing. If the deforming system was closed, then this volume increase should be equal and opposite to the volume decrease caused by pressure solution (bulk negative dilation).

Bulk dilation has been plotted for the mesoscopic and microscopic fractures and stylolites in the limb and hinge regions of the anticline (Figure 67). Negative dilation for the stylolites is taken from the length of the stylolite "teeth" and represents the minimum amount of material removed by pressure solution. If the system was closed, the bulk dilation estimated for the stylolites should be equal
Total extension from VEIN SET 1 = $W_1 D_1$

Original length perpendicular to VEIN SET 1 = $1 - W_1 D_1$

Bulk dilation from VEIN SET 1 = $\frac{W_1 D_1}{1 - W_1 D_1}$

Total bulk dilation from all vein sets = $\frac{W_1 D_1}{1 - W_1 D_1} + \frac{W_2 D_2}{1 - W_2 D_2}$

Figure 66 - Method of calculation of bulk dilation.
Figure 67 - Bulk dilation from veins and stylolites within a limestone unit of the Marston Member of the Mount Head Formation (Unit 7). Locations are in the limbs and hinge of the anticline near Overfold Mountain.
to or less than that for the fractures. Indeed, this is seen at the hinge and in the southwest limb of the anticline. In the northeast limb, however, there appears to have been a significant volume decrease.

7.3.3. **TWINNING IN CALCITE**

In general, calcite twins are less abundant in Unit 7 than the other limestone units in this study. As was observed in other units, the density of calcite twins decreases with distance from the hinge. The density of calcite twins appears to be proportional to microfracture density in this unit (Figure 68). This would be expected inasmuch as the density of both structures is related to the amount of strain the rock has been subjected to and the presence of dislocations in the calcite crystal lattice. An inverse relationship is observed between stylolite density and calcite twinning. This suggests that clay enhanced dissolution takes place at a lower stress than the critical resolved shear stress for calcite twinning.

There is a high density of calcite twins in the northeast limb of the syncline which corresponds to the high strain estimated for the stylolites seams. This would suggest that, at some point in the deformation, the strain rate may have increased such that both calcite twinning and pressure solution were utilized. It is unclear why this has occurred only in the northeast limb.
Figure 68 - Average linear densities of calcite twin lamellae and microfractures (per cm) within a limestone unit of the Marston Member of the Mount Head Formation (Unit 7). Locations are in the limbs and hinge of the anticline near Overfold Mountain.
7.4. **UNIT 8 - LOOMIS MEMBER**

The Loomis Member consists of thick, massive, cliff-forming, grey limestone. The stiffness of this Member has caused it to have a broad, open hinge geometry, a geometry which is mimicked by the thinly bedded rocks of the overlying Marston Member (including Unit 7). Unit 8 consists of the uppermost beds of the Loomis Member, at the contact with the Marston Member.

Unit 8 is a well cemented, oolitic grainstone (Figure 6 and 36). Point contacts between grains suggest that cementation occurred very early in diagenesis, before significant compaction. As a result of early cementation, this rock had a very low primary porosity throughout the deformational history. The size and densities of veins, unfilled microfractures, and stylolites reflect the low porosity of this unit. Unfilled microfractures in particular, have increased permeability during deformation, and have enhanced the finite permeability.

7.4.1. **EXTENSION FRACTURES**

Extension fracturing has been of less importance in the accommodation of strain in Unit 8 than in the other units in this study. This is probably due to the extremely low porosity of this unit, a factor which may have prevented fluids from entering the rock, and thus prevented the development of the high fluid pressures required for hydraulic fracturing.
Vein sets follow similar trends to those found elsewhere in the fold. The individual veins are thinner, however, resulting in a lower bulk dilation. Two common vein sets are found in this unit. The first is parallel to bedding, and corresponds to those veins formed at the onset of folding (Figure 48). This vein set is most common at the hinge and in the northwest limb. The second common vein set is perpendicular to the fold axis and is correlative with similar veins found in all units studied.

Low permeability during deformation has resulted in an abundance of unfilled microfractures in Unit 8. Fluorescence microscopy reveals that the fractures commonly follow grain boundaries and calcite twin lamellae (Figure 69). The interconnecting network formed by these microfractures would greatly enhance the permeability of this unit. This enhancement is low in comparison to the dolostone units (Units 3 and 5), however, where fracture porosity is found at nearly every grain boundary (Figures 51 and 52). The discrepancy in the abundance of microfractures within these two lithologies is partly due to the mineralogy of the grains - calcite may deform by intra-granular glide mechanisms rather than fracturing. Of more importance however, is the shape of the grains. The sharp points and edges of the dolomite rhombs are very effective stress concentrators. Fractures are more easily initiated at these points of high stress. No sharp surfaces exist in the oolitic grainstone.
Figure 69 - Thin section of Unit 8 impregnated with fluorescent dye. Fluorescence reveals fracture porosity along grain boundaries as well as twin planes within grains. A) Plane light; B) Blue light. Long edge of photograph measures 0.06 mm.
Bulk dilation and bulk negative dilation have been estimated from the veins and stylolites in the limbs and hinge of the fold (Figure 70). Since the curve for the bulk negative dilation consistently lies well below that for the positive dilation, it can be inferred that this unit has remained at constant volume throughout folding. The large amount of positive and negative dilation in the northeast limb of the anticline indicates that these rocks are more highly strained than rocks in other parts of the fold. This phenomenon was also found in the rocks of the Marston Member and the Lower Carnarvon Member. Reasons for this are not clear.

7.4.2. SHEAR FRACTURES

Shear fracturing is a dominant deformational mechanism in Unit 8, much more so than the other limestone units studied. In fact, in many ways this unit has behaved more like the dolostone units (Units 3 and 5) than the other limestone units (Units 2, 4, 6, and 7). This is a result of the low porosity of this unit, a factor which has prevented the utilization of the deformational mechanisms which require fluids: pressure solution and hydraulic fracturing. These mechanisms were very active in the deformation of the limestones which had a higher porosity, thus strain was accommodated without shear failure.

Two phases of deformation are delineated by the shear fractures in Unit 8 (Figure 71). This is in agreement with shear fractures found throughout the study area. The
Figure 70 - Bulk dilation from veins and stylolites within a limestone unit of the Loomis Member of the Mount Head Formation (Unit 8). Locations are in the limbs and hinge of the anticline near Overfold Mountain.
Northeast Limb  Southwest Limb

Figure 71 - Equal area projection of poles to shear fractures in a limestone unit within the Loomis Member of the Mount Head Formation (Unit 8). Information is given for the limbs of the overturned anticline near Overfold Mountain. The dashed great circle represents the megascopic ac plane. The solid great circle represents bedding. Contour interval = 3 sigma (Kamb method).
earliest phase is coaxial with the folding of the beds. Kinematic analysis of slickensides on these fractures reveals an axis of rotation (b-axis) which parallels the megascopic fold axis. A later, post-folding, phase of shear fracturing is found in the overturned limb of the anticline. Kinematic analysis of slickensides on these fractures reveals an axis of rotation toward the northeast.

7.4.3. **SEMI-BRITTLE SHEAR ZONES**

No semi-brittle shear zones were found in this unit. This again is an effect of low permeability and porosity. Lack of fluid in the rock has prevented the development of hydraulic extension fractures, a crucial element in the development of shear zones. Differential stress is thus allowed to increase to a point of shear failure.

7.4.4. **PRESSURE SOLUTION**

As was found in the overlying units, stylolites within Unit 8 delineate three phases of stylolite nucleation. The earliest stylolites are parallel to bedding and were nucleated by compaction during sediment loading. The second group of stylolites parallel the fold axis and are perpendicular to bedding. These stylolites developed during bedding parallel shortening at the onset of folding. Analysis of both of these stylolite sets reveal a kinematic b-axis which parallels the megascopic fold axis (Figure 72), confirming that these stylolites are coaxial with fold development. The youngest stylolites were formed after the
Northeast Limb  Hinge  Southwest Limb

- Pole of the stylolite seam (c-axis)
- Segment of the plane defined by the pole to the stylolite seam and the ridge crests along the seam (ac plane).
- Pole to the ac plane defined by the stylolite seam (b-axis)
- Megascopic fold axis

Figure 72 - Poles to stylolite seams in a limestone unit within the Loomis Member of the Mount Head Formation (Unit 8). Data is for the overturned anticline near Overfold Mountain. Contour interval equal to 3 sigma (Kamb method). Dashed great circle represents the ac plane of the megascopic fold. Solid great circle represents bedding.
folding of the beds and dip steeply to the northwest. Analysis of these stylolites reveals a kinematic b-axis which parallels the non-coaxial stylolites found throughout the study area.

As mentioned previously, the number of the stylolites nucleated at any one time is a function of the permeability of the rock. A low permeability allows the nucleation of only a few stylolites. In Unit 8, a history of low primary porosity, and a progressive increase in secondary (fracture) porosity, is reflected in the densities of stylolites developed before, during, and after folding.

Figure 73 shows the densities of the three stylolite sets at various stations in the limbs and hinge region of the fold. Of interest is the inverse relationship often found between the density of one stylolite set and the density of the set which preceded it in time. For instance, where there is a low density of stylolites which are coaxial with folding (stylolites 1 and 2), such as in the northeast fold limb, there is a high density of the later, non-coaxial, stylolites (stylolite 3). The high density of these later stylolites indicates that they nucleated at many sites, a condition which would have required an enhanced permeability. Microfractures formed during the folding of the beds are an obvious source of this secondary permeability. The inverse relationship between the densities of succeeding stylolite sets is more obvious, and more confusing, in the hinge region, where the permeability
LOCATIONS

Stylolite 1 - parallels bedding
Stylolite 2 - perpendicular to bedding and parallel to the fold axis
Stylolite 3 - perpendicular to the fold axis.

Figure 73 - Densities (per m) of three sets of stylolites in a limestone unit from the Loomis Member of the Mount Head Formation (Unit 8). Locations are in the limbs and hinge of the anticline near Overfold Mountain.
appears to have varied spatially, over a distance of several meters, as well as temporally.

7.4.5. **CALCITE TWIN LAMELLAE**

The abundance of microfractures and calcite twin lamellae decrease proportionally with distance from the hinge (Figure 74). This trend is also found in the other carbonate units studied. Both of these structures indicate that the hinge region has accommodated more strain on the granular scale than the fold limbs. One reason for this is that strain in the fold limbs has been accommodated dominantly by bedding parallel shear, a mechanism which has not been important at the hinge due to the orientation of the bedding planes with respect to the maximum and minimum compressive stresses.
Figure 74 - Average linear densities of calcite twin lamellae and microfractures (per cm) within a limestone unit of the Loomis Member of the Mount Head Formation (Unit 8). Locations are in the limbs and hinge of the anticline near Overfold Mountain.
8. CONCLUSIONS

The deformational mechanisms that have been operative at Overfold Mountain include solution processes (hydraulic fracturing and pressure solution), shear fracturing, and intragranular mechanisms (mechanical twinning, dislocation glide and microfracturing). Fluids within the rock have regulated which of these mechanisms has been utilized in the accommodation of strain. When fluids were present, pressure solution and hydraulic fracturing ensued. These processes appear to have taken place at stresses lower than those required for shear fracturing and calcite twinning, as there is an inverse relationship between the abundance of structures developed by solution processes and that of those from the other processes. When rocks were fluid deficient, they commonly failed by shear and by microfracturing.

The mechanical behavior of each of the lithologies studied has been governed not only by permeability, but also by the mineralogy of the grains within the unit. Mineralogy has controlled whether an individual grain behaved in a brittle or ductile fashion. Ductile mechanisms utilized in the rocks of the study area are pressure solution and intracrystalline glide. Microfractures found along grain boundaries and within grains are evidence of brittle deformation. In the study area, the limestones and dolostones have deformed in a semi-brittle manner, utilizing both brittle and ductile mechanisms. The quartz
arenite unit, on the other hand, has mainly deformed by brittle mechanisms; in this case shear fracturing. Calcite grains within the limestone units commonly have deformed by twinning and twin glide. Dolomite crystals have also formed twins, though not to the extent of calcite. Quartz has not deformed significantly by intracrystalline glide mechanisms due to the strong covalent bonds between $\text{Si}^{4+}$ and $\text{O}^{2-}$. The ease with which each of these minerals is dissolved by pressure solution also varies. Again, the covalently bonded atoms in the quartz lattice do not dissociate easily, and strain accommodated by pressure solution is relatively minor. The ionic bonds in the carbonate lattices, on the other hand, are relatively easily broken and pressure solution occurs readily.

Many of the deformational mechanisms studied have affected the finite permeability of the carbonates and sandstones. As mentioned, the solution processes have been most active in rocks with a high initial permeability. Pressure solution in these rocks has served to close existing pore spaces and decrease the finite permeability. The abundant thick veins formed in these rocks, as well as the fine grained, clay rich selvage along stylolite seams, also serve as effective barriers to fluid movement. The opposite effect is found in rocks which had a low initial permeability. Abundant microfractures in these rocks have served to increase the finite permeability. This effect is
most impressive in the dolostone units, where microfracturing is found along most grain boundaries.

The role of fluids in the deformation of rocks near Overfold Mountain is displayed in the type and density of veins, stylolites, shear fractures, microfractures and calcite twin lamellae. Kinematic analysis of these structures has delineated three phases for their development: pre-deformational compaction, folding, and post-folding. Permeability and porosity has varied through each of these phases, causing a variation in the way in which strain has been partitioned between the aforementioned structures. Analysis of the structures developed in each phase has thus given insight into how and why the permeability has varied over the deformation interval of the rock.

Variation in the permeability of the carbonate units during the three phases of deformation is best documented by the density of stylolites developed during each phase (stylolite density is proportional to permeability). Microfractures formed in units of low permeability during one phase increase the permeability of the rock during the next phase. Enhanced permeability during this later phase is indicated by the development of a higher density of stylolites.

Permeability has therefore governed the way in which strain has been partitioned between the deformational mechanisms employed throughout the strain history of the
rock. Mesoscopic and microscopic structures formed before, during and after the folding of the beds near Overfold Mountain show that the permeability has varied over this interval. Permeable beds have had a permeability decrease by the closure of pore spaces by pressure solution. Impermeable beds have had a permeability increase by the development of an interconnecting network of microfractures. The extent to which permeability varies is dependent upon the mineralogy of the unit and the amount of strain the rock has accommodated.
REFERENCES


