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The effect of rock creep on the morphology of steep-sloped sections of the Niagara Escarpment (Ontario)

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THE EFFECT OF ROCK CREEP ON THE MORPHOLOGY OF STEEP-SLOPED SECTIONS OF THE NIAGARA ESCARPMENT

By

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THESIS

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Abstract

The thesis addresses the post-glacial development of the Niagara Cuesta between Hamilton and Collingwood. Conventional thinking on the escarpment during the Holocene suggests that the present morphology developed rapidly following deglaciation due to periglacial processes (Straw, 1966/Tovell 1992). The disruption of the preglacial drainage pattern by drift has meant the escarpment has not been subject to extensive fluvial action along its base, resulting in an extremely slow rate of retreat (Bird, 1980). The concept that the escarpment is a relict feature therefore pervades the modern literature.

The apparent motion of the blocks at the face of the escarpment (Hintz, 1997) suggests that in the absence of exogenetic processes, a slow development due to endogenetic processes has dominated the modern development of the escarpment. It has bee suggested by Hewitt (1997), that the present morphology found at cliffed sections of the escarpment may be due to deformation within the shale layers. In order to test this hypothesis, the strength properties of the rocks that form the escarpment were tested and compared to the gravitational stresses that would be experienced within the rock mass. The results indicate that both the Cabot Head Shale and the Queenston Shale posses compressive strengths that are below the principal gravitational stress expected within the escarpment. It is therefore concluded that in the absence of high confining stresses, as would be expected near the cliff face, deformation within these formations is occurring.

Acknowledgments

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Table of Contents

List of Figures

List of Tables

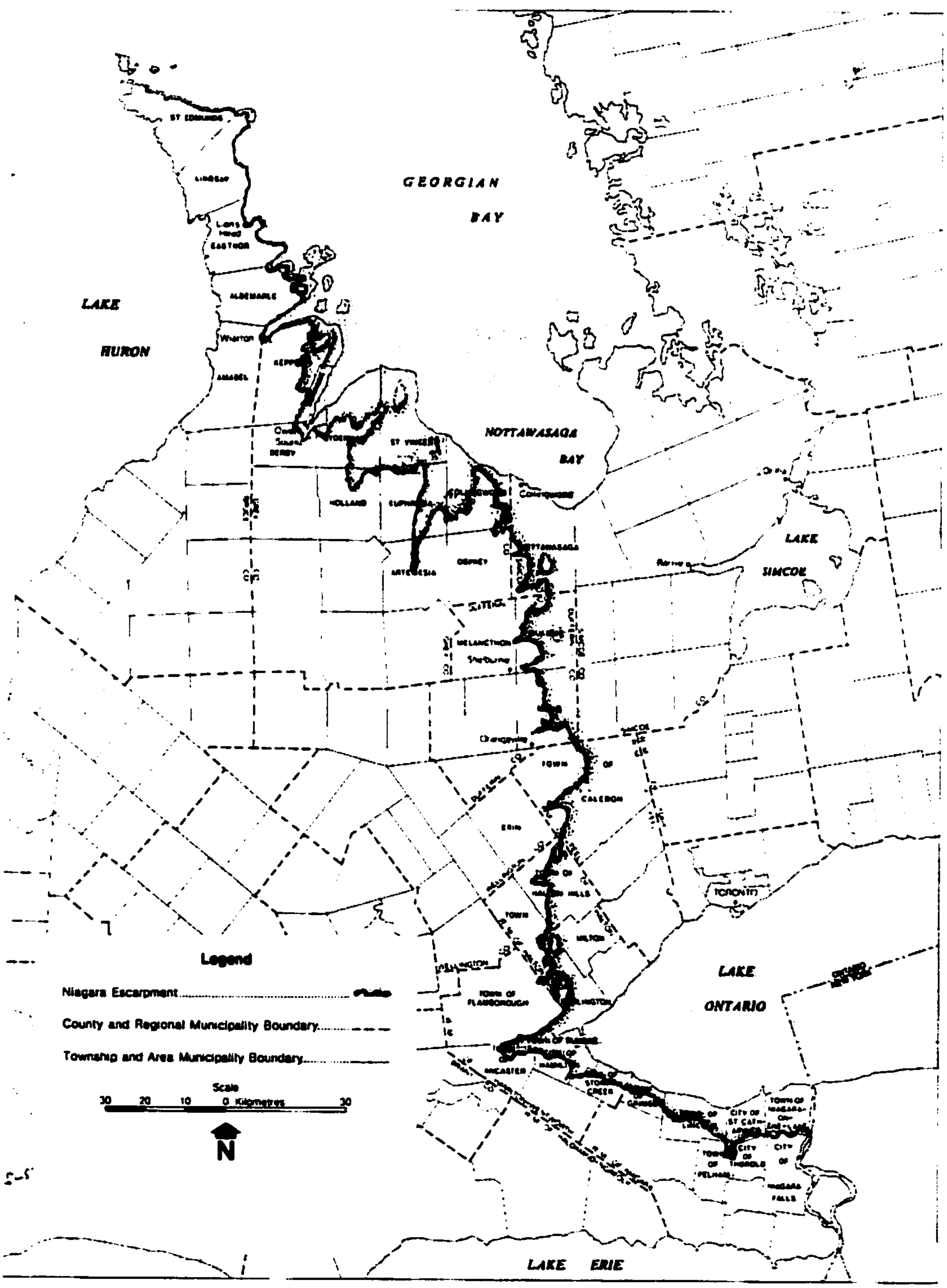
Chapter One: Introduction

1.1 Introduction:

The Niagara Escarpment of Southern Ontario is a cuesta landform of great importance to the geomorphology of the region. It is oriented along a roughly north-south axis, from the mouth of the Niagara Gorge to where it disappears beneath the waters of Lake Huron at the tip of the Bruce Peninsula (Figure 1.1). The escarpment is one of the few bedrock features in a landscape otherwise dominated by glacial and lacustrine landforms. It is also one of the few major topographic features within Southern Ontario. Drainage patterns are greatly affected, streams to the west of the escarpment generally drain parallel to the scarp face while those to the east tend to flow perpendicular to the escarpment (Bird, 1980). Along its length, the geometry of the escarpment is highly variable in both form and plan. In many areas, the escarpment is present as a vertical dolomite cliff with a steep talus slope below. However, long stretches are also identifiable as a gradual rise with the bedrock concealed beneath glacial deposits. In plan, the escarpment is incised by reentrants along its entire length. These have been interpreted as the result of glacial action on weak areas (Straw, 1968) or the remolding of preglacial valleys (Grabau, 1920), or a combination of both (Tovell, 1992).

The Niagara Escarpment is thought to be one of a series of erosional features associated with the lithological structure of the Great Lakes area (Bird,

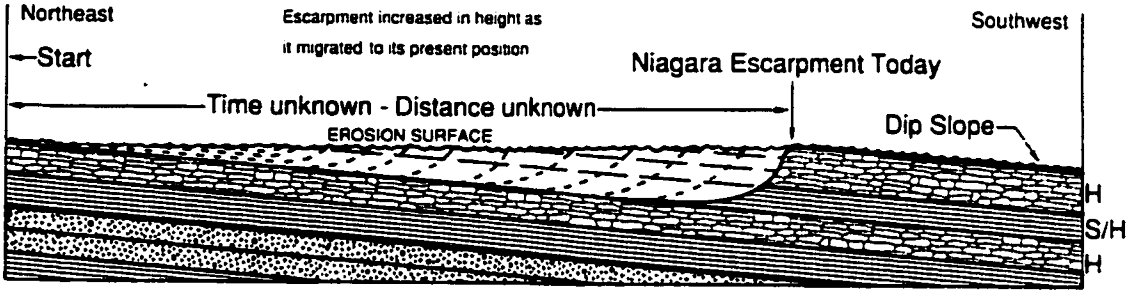
Figure 1.1: Geography of the Niagara Escarpment in Southern Ontario. Source: Tovell, 1992.



1980). The dolomite that caps the escarpment is underlain by a series of weaker shale strata interbedded with sandstone, limestone and dolomite. One interpretation of the landform suggests that the relative ease with which the shale strata are eroded compared to the dolomite results in sapping (Tovell, 1992). This sapping may take the form of either direct undercutting of the cap rock due to fluvial action or due to the chemical weathering of the shale layers via acidic groundwater. This implies that the escarpment has migrated to the west over time while increasing in height as depicted in figure 1.2. The presence of crevice caves and detached blocks on the crest of the cliffed sections of the escarpment has been attributed to post glacial pressure release and periglacial processes (Straw, 1966). These ideas will be more fully developed in sections 1.2 and 2.1.

There is however a significant amount of evidence which points to an alternate hypothesis for scarp development. The presence of cambering in the near scarp zone, combined with the apparent jostling of the blocks of the cliff face indicate that deep seated rock creep within one or more of the shale formations may be the controlling factor for mass movement along the Niagara Escarpment (Hewitt, 1997). The process may be aided or triggered by the hydrogeology of the area. Groundwater flowing through the escarpment may have a significant impact on the strength properties of the various rock formations and may induce weathering due to the removal of calcite and dolomite in solution. In addition to this, the hydraulic pressure within the groundwater system can have a significant effect on the stress environment found within the rocks of the escarpment. Such relationships between strong cap rocks squeezing out weak, clay rich, sub-strata have been noted in the United States (Radbruch-Hall, 1979), Great Britain (de Frietas & Watters, 1973), and elsewhere.

Figure 1.2: The Development of the Niagara Escarpment.
Source: Tovell, 1992

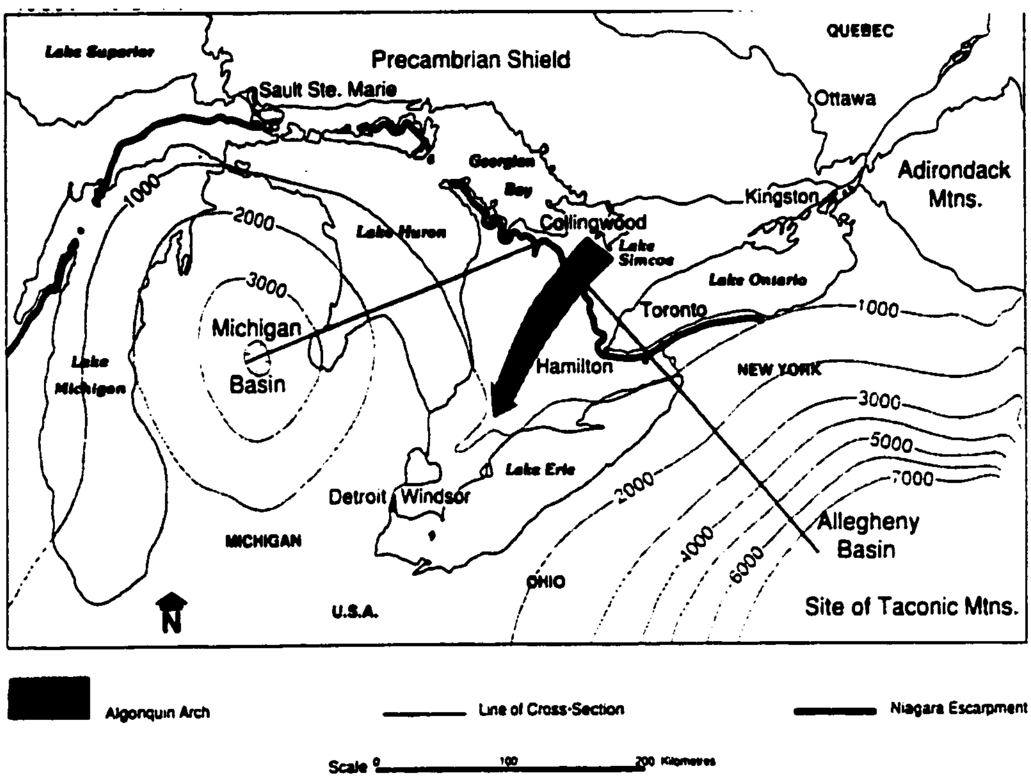


H: Hard, resistant rock; S: Soft, easily weathered rock

1.2 The Geological Setting:

Three main geological features dominate the bedrock geology of the Great Lakes Region: the Algonquin Arch, the Michigan Basin, and the Allegheny Basin. The Algonquin Arch is a large anticline, or upwarping of the earth's crust, which has its axis running from Chatham in the southwest to the area in and around Algonquin Park in the northeast (Chapman & Putnam, 1984). The anticline plunges to the southwest forming the spine of southwestern Ontario around which are arranged lakes Huron, Erie, and Ontario (Chapman & Putnam, 1984). Proximal to the Algonquin Arch are the two basins. To the west, located between lakes Huron and Michigan, lies the Michigan Basin, in which the underlying basement rock of the Precambrian shield is buried beneath roughly 4000 m of sedimentary strata. To the southeast lies the Allegheny Basin which contains around 7000 m of sedimentary rock (Tovell, 1992). The spatial arrangement of these features is illustrated in figure 1.3. As can be seen from this diagram, the formations north of Hamilton dip to the southwest into the Michigan Basin while those located along the Niagara Peninsula and in New York State dip to the south into the Allegheny Basin.

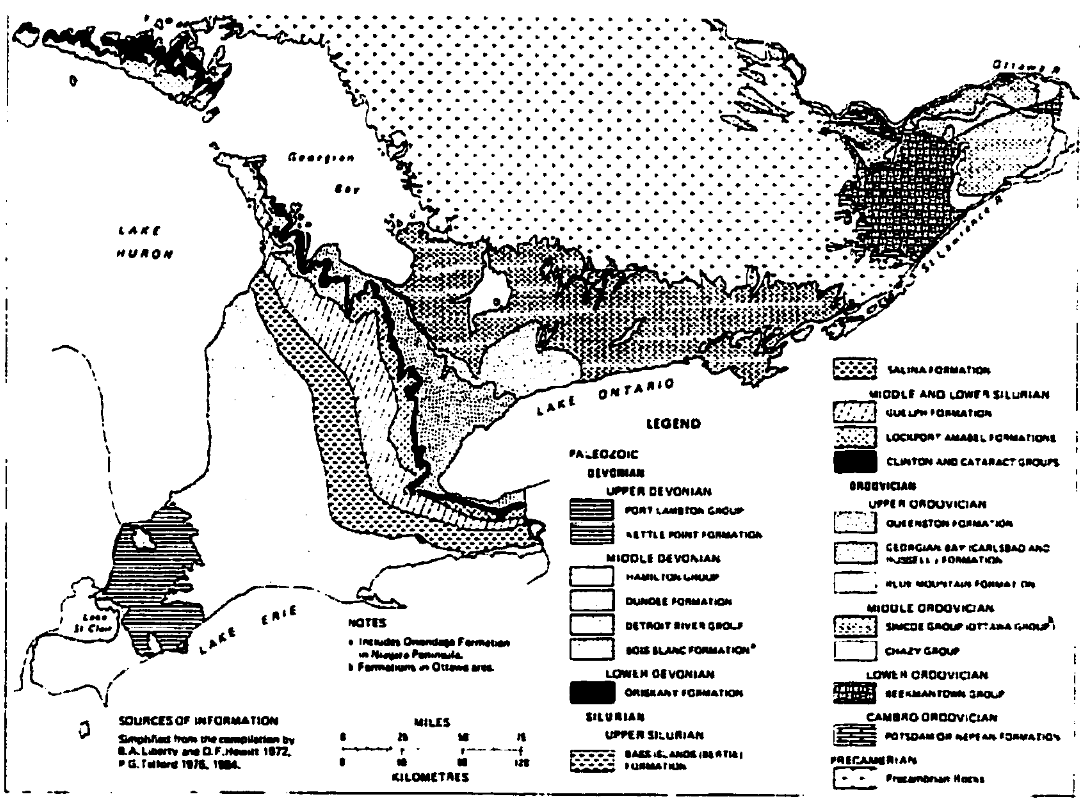
Figure 1.3: Major Geological Features of the Great Lakes Region.
Source: Tovell, 1992.



Also evident from figure 1.3 is the strong influence that these structures have on the position of the Niagara Escarpment. In New York State, the escarpment skirts the edge of the Allegheny Basin while in Ontario and Michigan State, the escarpment circles the edge of the Michigan Basin. The elevation of the escarpment is also greatly affected by these features. Indeed, the escarpment reaches its highest elevation (546 masl) where it intersects the Algonquin Arch 2.4 km north of Singhampton, overlooking Edwards Lake (Chapman & Putnam, 1984). From here it slopes down the flanks of the arch in both directions. To the north, at Tobermory, it disappears beneath the surface of Georgian Bay and to the south, at Hamilton, it has an elevation of just 150 masl.

The rocks of Southern Ontario are of the Precambrian, Ordovician, Silurian, and Devonian. The Precambrian rocks are those of the Canadian Shield and are exposed to the north of Georgian Bay and underlie the entire area. At the surface, the sedimentary rocks of the Ordovician, Silurian, and Devonian are arranged in concentric bands around the two basins as illustrated in figure 1.4 These formations are composed of sediment transported from the Taconic Mountains, which were being uplifted approximately 500 million years ago

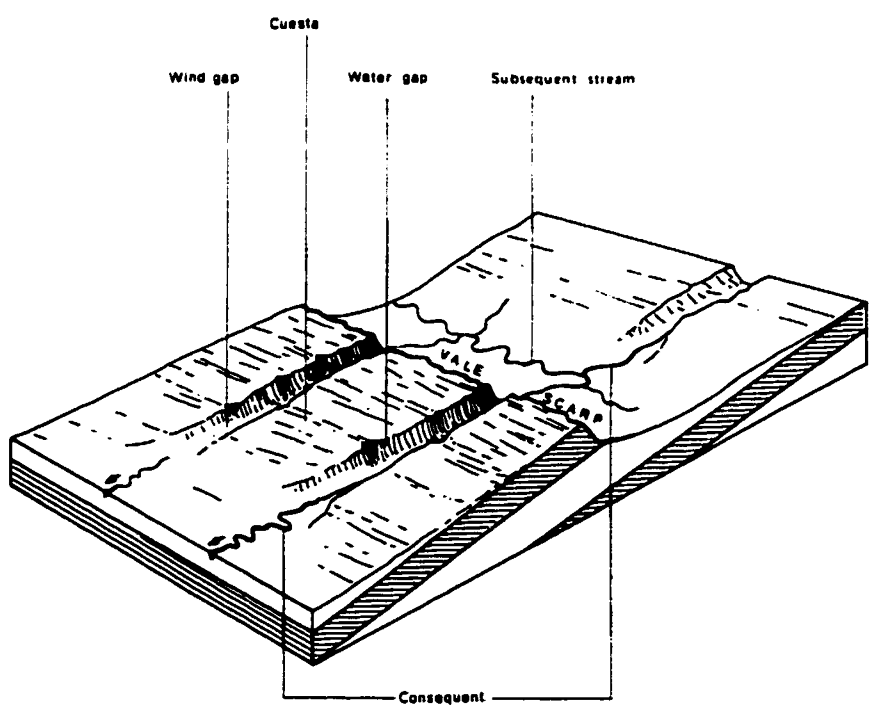
Figure 1.4: Bedrock Geology of Southern Ontario.
Source: Chapman and Putnam, 1984.



during the Taconic Orogeny. These mountains were the progenitors to the present day Appalachian Mtns. and it is estimated that roughly 600000 km^3 of sediment were removed during the orogeny and deposited in marginal seas (Wilcander & Monroe, 1993). The level of these marginal seas was subject to much fluctuation. This resulted in the periodic submergence of the sediments being brought down from the mountains. The formations that comprise the Niagara Escarpment along the entire section between Mt Nemo and Osler Bluff are the Ordovician shale of the Queenston Formation and the Silurian rocks of the Clinton and Cataract groups. At Osler Bluff and to the north at Blue Mountain the Queenston Formation is exposed from top to bottom and the underlying Georgian Bay Formation occupies the position at the base of the escarpment. The depositional environment envisioned for the various formations is a coastal delta in a shallow inland sea that experienced several fluctuations in sea level (Dennison, 1976). The thickness of the clastic formations increases to the southeast in the direction of what was once the Taconic Mountains (Tovell, 1992). To the north, the sediment becomes more calcareous reflecting the increasing importance of biogenic processes moving seawards in coastal waters. Periodic submergence of the sediments resulted in the deposition of thick limestone beds, which contain a large abundance of bioherms and other marine fossils. These have since undergone a high degree of dolomitization. A more detailed description of each of the formations involved will follow in section 2.4.

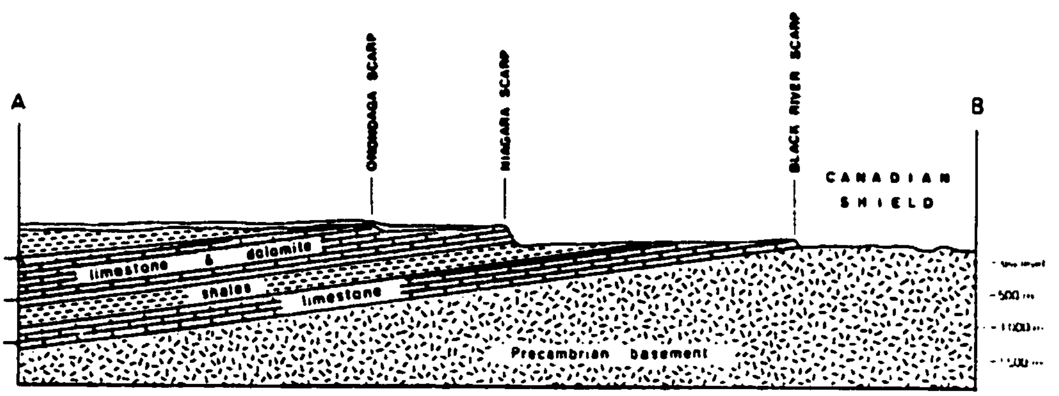
Based on other deposits within Ontario such as gypsum, salts, and carbonates it is known that the seas in which the Silurian rocks of the escarpment were deposited persisted beyond the end of the Paleozoic Era (Tovell, 1992). After these seas receded, the rocks of Southern Ontario were subjected to approximately 250 million years of erosion before the onset of the Pleistocene Glaciation (Chapman & Putnam, 1984). The evolution of the region was therefore primarily the result of fluvial action acting on the various rock formations as they followed a homoclinal recession into the Michigan Basin (Tovell, 1992). A prevailing theory of the geomorphic development of the region states that the rate of this recession would be proportional to the resistance exhibited by the various rock strata to the processes of weathering and erosion

Figure 1.5: Elements of the Scarp and Vale Topography.
Source: Bird, 1980



such that the weaker shale formations would rapidly recede and therefore undercut the more resistant carbonate deposits. The result is a scarp and vale topography, the major elements of which are shown in figure 1.5. As can be seen, the vales are interpreted as being due to the relatively quick rate at which the softer shale formations are removed by a series of subsequent streams developing parallel to the strike of the resistant formations. The consequent drainage continues to cut downward creating deep valleys through the cuesta (Bird, 1980). This mechanism seems to fit the bedrock geology of Southern Ontario, which, in essence, is a series of carbonate scarps separated by softer shales as depicted in figure 1.6. Three major scarps are illustrated in this figure: the Black River Scarp that is located

Figure 1.6: Simplified Geological Cross Section of Southwestern Ontario. Source: Bird, 1980.



between the Penetang Peninsula and the eastern end of Lake Ontario, the Niagara Scarp, and the Onondaga Scarp that runs through Southern Ontario to the west of the Niagara Escarpment.

The effect that the Pleistocene glaciation had on the geomorphology of the region was to bury much of the scarplands of the Great Lakes Region beneath a thick layer of drift. This has resulted in the masking of the Niagara Escarpment along much of its length between Milton and Collingwood, the most notable exception being the cliffs and outliers located at Mono. The Black River Scarp remains a somewhat prominent feature, reaching heights of between 15 - 25 m. The Onondaga scarp is also widely buried but outcrops along portions of the Niagara Peninsula, New York State, through Goderich on the shore of Lake Huron, and submerged beneath the waters of Lake Huron along Nine Fathom Ledge (Grabau, 1920). Perhaps the most important effect that the Pleistocene glaciation is thought to have had on the development of the Niagara Escarpment was the infilling of the vale to the east of the scarp face with drift, which is thought to have disrupted the preglacial drainage. The present day drainage to the east of the escarpment is therefore largely orthogonal to the scarp face (Bird, 1980). The obvious effect of this is to disrupt the preglacial process of scarp recession which has led some to the conclusion that the present day escarpment is a relict feature (Tovell, 1992).

1.3 Problem Statement:

The purpose of this investigation is to test and more fully define the hypothesis of deep-seated deformation (Hewitt, 1997) and its relationship to the geomorphology of the Niagara Escarpment. In particular, it is necessary to determine if plastic deformation is occurring within the lithological sequence of the escarpment, particularly the shale formations. If this type of deformation is taking place, it must be monitored in a quantitative way in order for a useful analysis of its effect on the geometry of the escarpment to be made. The effects that hydraulic pressure and chemical weathering may have on the system must also be accounted for as they enhance, or may act as a trigger for, ductile flow.

There is empirical evidence that indicates that the escarpment is neither a relict feature nor presently subject to extensive fluvial action. The presence of the crevice caves and the apparent motion of the dolomite blocks may reflect an ongoing subsidence along the escarpment face brought on by the removal of material at depth. Furthermore, the manner in which the Queenston Formation, which is quickly weathered when exposed, protrudes from the base of the escarpment suggests that it may be in motion laterally due to the heavy overburden and the lack of confining stress (Hewitt, 1997). Such a process has been termed a yielding flow landslide (Pei & Tianchi), and results from the reduction in confining stress near the faces of slopes. The present day escarpment may, therefore be developing due to the eastward expansion and subsidence of the cap rock resulting in recurrent toppling failure of blocks at the scarp face. Evidence for this mechanism can be found on the talus slope below which is typically littered with many large blocks indicative of large scale mass movement events.

1.4 Objectives:

In order to fully answer the questions posed above, the rheological behaviour of the various rock formations involved will be assessed. These will then be compared with the stress environment that applies for each formation within the escarpment. This is a function of the geometry of the slope, the unit weight of the various rock formations, the slope hydrology, and the tectonic stresses experienced within the study area. For the purposes of this project, only the unit weight of the rocks will be used to calculate overburden. While this will not provide any quantitative data as to the orientation of the principal stress, it will indicate its potential magnitude. The field sites used for this investigation are located along the section of the escarpment between Hamilton and Collingwood. The relative simplicity of the lithology in this area compared to that of the Niagara Peninsula is an important factor and it is away from the influence of wave action that occurs along much of the Bruce Peninsula. The study sites also give a good representation of north-south variations in the rock formations and of the mineralogy within formations. With these data and precise vertical profiles of the selected study sites, a conceptual model of slope processes along the escarpment can be created. The model, in turn, will be used to investigate the likely mechanism of slope failure along cliffed sections of the Niagara Escarpment.

The discussion will be developed in four sections:

1) A review of the relevant literature will first be given in order to put the present study into context.
2) A description of the field and laboratory methods used in this study.
3) A summary of the results obtained by these methods.
4) The conclusions that can be drawn from these results.

It is hoped that this work will stand as a first step towards a revised interpretation of the development Niagara Escarpment during the Holocene.

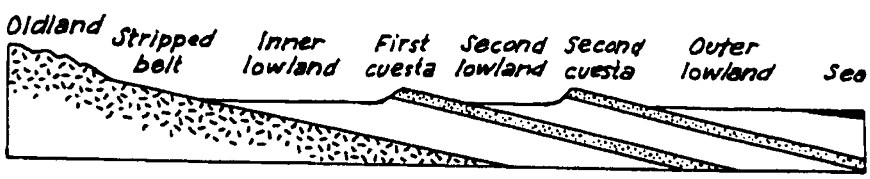
Chapter Two:
A Review of the Literature

2.1 Previous Work on the Geomorphology of the Niagara Escarpment:

Some of the earliest work done on the development of the Niagara Escarpment was that of William Morris Davis. In his book Physical Geography of 1898, Davis asserts that the Great Lakes region is, in actuality, an ancient coastal plain derived from sediment eroded from the Canadian Shield (Davis, 1898). The sedimentary rocks of the plains were exposed either by a period of uplifting, or by a lowering of the sea levels. The strata of this coastal plain dip to the south away from the "Old Land" of the shield due to a number of factors: the original seaward dip of the underlying Old Land, the up-warping of the "Old Land" during periods of uplift, and down-warping of the coastal areas due to the overburden of the deposited sediments (King & Schumm, 1980). Upon emergence, the plain would be capped by the uppermost strata of rock and drained by a series of consequent streams flowing down slope across this cap rock (King & Schumm, 1980). Over time, the coastal plain would begin to thin due to coastal erosion on the seaward side and fluvial erosion on the landward side. As the processes of erosion continued, the upper layer of sedimentary rock would have eventually been breached exposing the substrata. If some of the substrata possessed a greater resistance to weathering and erosion than the others, they would have been carved into cuestas (Davis, 1898). If no further periods of uplift were experienced, it was theorized that this zone would gradually be lowered to a peneplain, in which the "edges of the sloping underlying rocks are expressed less by relief than by belts of contrasting vegetation" (King & Schumm, 1980).

If however, the area is subjected to a further period of uplift, the result will be a rejuvenated coastal plain on the landward side and a new strip of coastal plain on the seaward side (King & Schumm, 1980). Here, the belts of contrasting vegetation would be joined by lines of cuestas moving down dip towards the sea as illustrated in figure 2.1. These cuestas would be the result of the drainage system in which subsequent streams would develop in the lowlands due to their higher susceptibility to weathering and erosion (Davis, 1898).

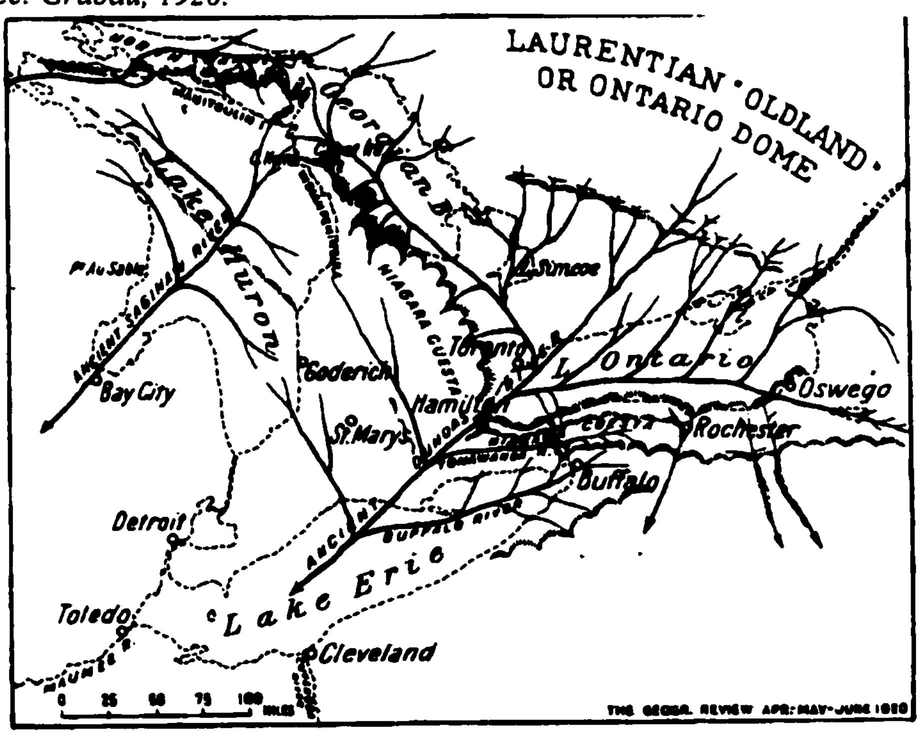
Figure 2.1: Coastal Plain Topography According to Davis.
Source: King & Schumm, 1980.



The evolution described by Davis survived with little modification until 1920 when Grabau pointed to a "now vanished continent" which stood where the present day Atlantic Coast is located as the source area for the Paleozoic sediments of the Great Lakes Region (Grabau, 1920). This "vanished continent" is of course the Taconic Mountains. Citing the fact that remnants of the Ordovician and Silurian Beds can be found upon the surface of the present day shield rock, and to the north of it around James Bay; Grabau stated his belief that the formations of the Niagara Escarpment once extended from their present location across the whole of Ontario to the James Bay Region and beyond

(Grabau, 1920). The Algonquin Arch or "Ontario Dome" was thought to have been elevated at the end of the Paleozoic and subsequently peneplained during the Mesozoic such that the shield rock was once again exposed while the sedimentary rocks of the Paleozoic Era remained at the periphery (Grabau, 1920). These were then carved into the scarp and vale topography that is present today. The streams that sculpted the lines of cuestas followed a different drainage to that of today. Grabau postulated that the drainage would be to the south as predicted in the original model by Davis (Grabau, 1920). Therefore, the consequent streams drained to the south through a series of gaps in the Niagara Escarpment as shown in figure 2.2. Draining along the faces of the cuestas in the weaker shale

Figure 2.2: Preglacial Drainage Pattern of the Great Lakes Area.
Source: Grabau, 1920.



strata, acting as tributaries to the three major rivers were the subsequent streams. As illustrated, the major rivers of the time were the Ancient Dundas which passed through the present day Dundas Valley, the Ancient Saginaw which passed through the gap between what is now the tip of the Bruce Peninsula and Manatoulin Island, and the Ancient Genesee which drained across what is now the eastern portion of Lake Ontario and into the Genesee Valley of New York State (Grabau, 1920).

The present drainage emerged as a post-glacial system following the Wisconsinan Glaciation (Tovell, 1992). In Southern Ontario, the glacial period began some 1.65 million years ago. It consisted of four stages, the last of which was the Wisconsinan. It is thought that this extended period of glaciation is responsible for transforming the region from a structurally controlled landscape into one that has been strongly modified by glacial deposition (Bird, 1980; Tovell, 1992). The weight of the ice served to lower the area significantly, and indeed isostatic rebound continues to this day (Chapman & Putnam, 1984). Furthermore, some preglacial valleys were over-deepened and filled with morainic deposits (Grabau, 1920). This succeeded in blocking some of the major drainage systems which, combined with the lowering of the area, could have resulted in a reversal in the drainage direction within the old valleys (Grabau, 1920). This remains a topic of much debate in terms of both the nature of the preglacial drainage patterns and indeed the very origins of the valleys or reentrants that dissect the Niagara Escarpment. Straw (1968) points out that within the Dundas Valley, the bedrock descends towards Lake Ontario and the present elevation of the bedrock around Brantford Ontario implies that if the Ancient Dundas River did exist, it would have had to cut the escarpment at the level of the present day crest. This indicates that either the Dundas system was draining into the St. Lawrence system, or as Straw suggests, the reentrants along the Niagara Escarpment are primarily of glacial origin (Straw, 1968). As evidence, Straw compares the general orientation of the reentrants with axis of ice advance and notes a close correlation between the two (Straw, 1968). The morphology of the reentrants is also of interest as they are both straight and deep with much evidence of ice molding (Straw, 1968).

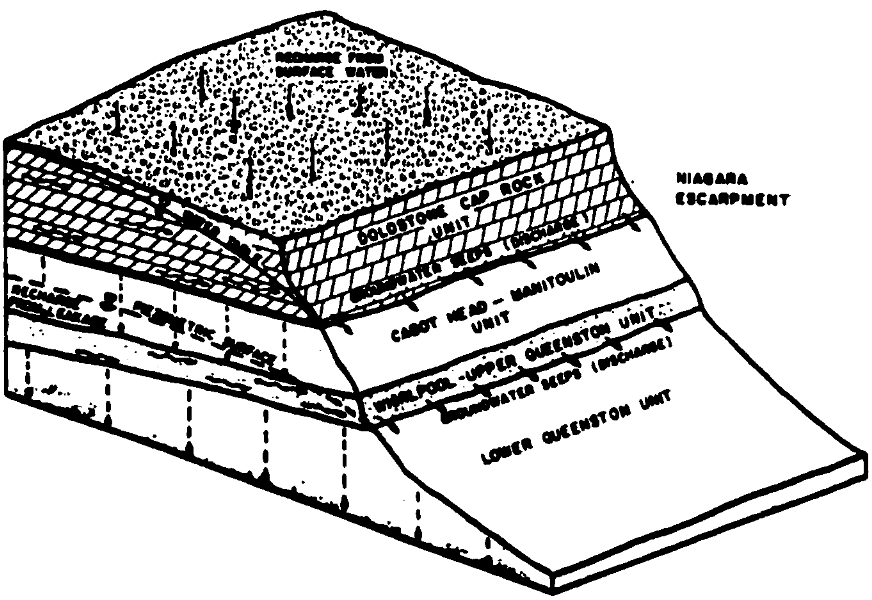
Irrespective of the preglacial drainage, the present day Niagara Escarpment appeared from beneath the retreating glaciers roughly 13000 - 12500 years ago (Tovell, 1992). The reentrants now support a series of obsequent streams draining eastwards from headwaters located atop and behind the cuesta. The preglacial vale, theorized to exist to the east of the escarpment would have been infilled with glacial drift or submerged beneath Georgian Bay and Lake Ontario effectively neutralizing the primary method of scarp retreat (Bird, 1980). The assumption that the Niagara Escarpment is a relict feature therefore pervades the modern literature (Bird, 1980, Tovell, 1992). The majority of recent studies have focused on small scale processes affecting various portions of the escarpment.

2.1.1: The Hydrology of the Niagara Escarpment

The hydrogeology of the Niagara Escarpment is dictated by its lithology. This is due to the presence of highly permeable dolomite and sandstone interbedded with less permeable shales. A review of the literature in this area reveals that few studies have been done. However, the work of Cowell & Ford (1975), Nadon & Gale (1984), and Tovell (1992) provide some insight into the subject. The northern temperate continental climate of the area results in an annual temperature variation in excess of 50° C each year. Precipitation in the area ranges from 759 mm/yr along the Niagara Peninsula to 701 mm/yr at the Bruce Peninsula (Environment Canada, 1995). Peak periods of precipitation occur in the spring and fall; while intermittent melting during the winter and spring, and convective storms during the summer also contribute to run-off.

Nadon & Gale (1984) produced a model for the flow pathways that water takes within the escarpment at Milton. This illustrates the presence of two major aquifers separated by shale of low permeability as depicted in figure 2.3. The relatively high permeability of the Amabel and Whirlpool formations is attributed

Figure 2.3: Groudwater Flowpathways Through the Rocks of the Niagara Escarpment.
Source: Nadon & Gale, 1984.



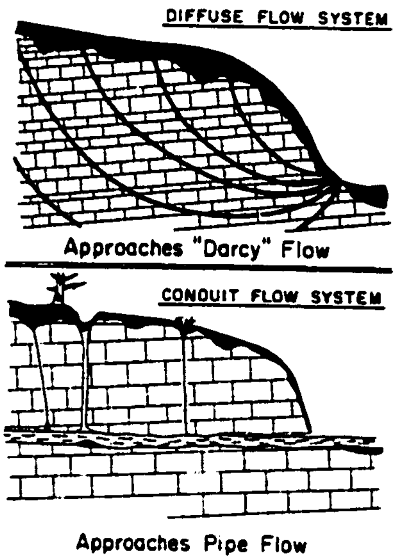
to a network of fractures. Within the Amabel Formation, these had an orientation that favoured a N60°E and N120°E pattern. This is consistent with neo-tectonic and pressure release jointing that has been observed in the area (Gross &

Engelder, 1991). The fractures of the Whirlpool formation trend towards N85°E. These fracture patterns cause the permeability of the formations to be anisotropic in the horizontal plane at a scale of a few meters. However, the ubiquitous nature of the fractures within the rock mass makes it behave in an isotropic manner at a scale of 10+ meters (Nadon & Gale, 1984). The low permeability of the Cabot Head and lower Queenston Formations causes them to act as an aquitard. The upper portion of the Queenston Formation is highly fractured (Russell & Harman, 1985), causing it to become more permeable as well. The upper portion of the Queenston Formation is therefore part of the Whirlpool aquifer. The source of water for this aquifer is seepage through the overlying Cabot Head Formation. This arrangement results in a series of springs along the base of the escarpment. Those issuing from the base of the Amabel formation are typically located near the top of the talus slope whereas those issuing from the upper Queenston Formation can be found below the talus slope along the lower slopes of the escarpment.

In some areas along the escarpment the fracture patterns within the cap rock have been connected and opened due to karst processes. One of the largest karst systems is that of the Wodehouse Creek Karst, located on the western side of the Beaver Valley near Kimberley, Grey County and has been discussed by Cowell & Ford (1975). The system is postglacial and therefore less than 10,220 yrs old. It has been formed due to the solubility of the dolomitic cap rock and the very steep groundwater hydraulic gradient caused by the abrupt scarp face. Karst systems are quite rare along the Niagara Escarpment, except along the Bruce

Peninsula where many large solution caverns and underground drainage systems operate (Tovell, 1992). However, the presence of karst systems does illustrate the potential importance that preferential flow pathways can have on the hydrogeology of the area. The escarpment is therefore subject to both of the end-member flow systems common to carbonate aquifers (Schuster & White, 1971). The model produced by Nadon & Gale for the cliffed sections of the escarpment near Milton is an example of a diffuse flow system whereas the karst systems are examples of a conduit flow system. Both the diffuse flow system and the conduit flow system are illustrated in figure 2.4. As can be seen, the diffuse flow system is described by Darcy's law which states the rate at which water will flow through a porous medium is a function of the hydraulic conductivity of the medium and hydraulic gradient (Brassington, 1990). Conversely, the conduit flow system illustrates how groundwater can make use of preferential flow pathways as it travels through a rock mass. The importance of this to the hydrogeology of an area is evident. Aquifers that experience a diffuse flow system will have a longer residence time and fairly even dispersal for their groundwaters. This means that more of the rock mass will be subject to the effects of chemical weathering. In the few areas that

Figure 2.4: The Two End-Member Flow Systems of Carbonate Aquifers Source: Schuster & White



show evidence of a conduit flow system, the groundwater will have a lower residence time and the weathering will be focused upon enlarging the flow pathway.

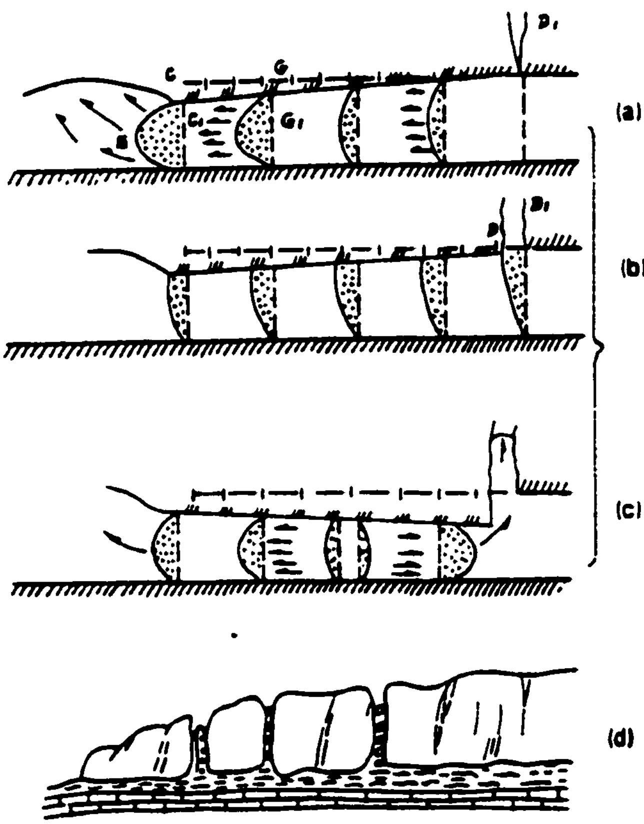
2.1.2: Slope Processes Studies on the Niagara Escarpment

Studies done by Moss & Nickling (1980) and Fahey & LeFebure (1988) have focused on some aspects of slope stability at site-specific locals. Moss & Nickling investigated several sections of the escarpment in the Hope and Barrow Bay area of the Bruce Peninsula. They concluded that previous catastrophic failure of the cap rock had occurred due to glacial oversteepening, periglacial processes immediately following the retreat of the Wisconsinan glaciers, and wave action upon the weaker fossil hill dolomite which underlies the cap rock at this local. However, the present slope morphology within their study area is attributed to the interaction between the slope materials and various plant species, most notably White Cedar (Moss & Nickling, 1980). Fahey & LeFebure also investigated the escarpment on the Bruce Peninsula. They instrumented a site near Wiarton to determine the effect that freeze/thaw and frost wedging has upon the rocks of the escarpment. They concluded that frost wedging within preexisting fractures is the primary process by which material is weathered from the scarp face (Fahey & LeFebure, 1988). This process however, only seemed to be effective to a depth of 5 cm and resulted in small fragments of only a few centimeters per axis. By collecting the debris falling from a known area of the scarp face they calculated that the scarp had retreated approximately 0.1 mm during the winter of 1983-84 (Fahey & LeFebure, 1988). This obviously does not account for the huge blocks, which can be found within the talus slopes of the cliffed portions of the escarpment. Such a block, 1 m in width, would be equivalent to 10000 years-worth of this type of weathering. Clearly, a process of higher magnitude and lower frequency is responsible for the production of such blocks.

One case of large-scale mass movement along the Niagara Escarpment is suggested by Straw (1966) for the prominent outcropping between Meaford and Kimberley called the Griersville Rock. Inspection of the base of the dolomitic blocks that cap the area was made possible due to a road cut across the top of the site. These blocks appear to have separated from one another and moved down slope under the influence of gravity (Straw, 1966). The mechanism suggested for this by Straw were periglacial conditions, in which the groundwater was exposed to extreme cold. Freeze/thaw cycles then acted to break up the upper layers of the shale and assist in the movement of the dolomite blocks (Straw, 1966). Subsequent warmer and wetter conditions resulted in the plastic deformation of the underlying Cabot Head Shale such that the blocks continued to move downslope. Evidence for this deformation is given by intrusion of the shale upward into the spaces between the blocks and the rotation of the blocks on top of the shale. The thickness of the dolomitic cap rock in this area is roughly 7 m (Straw, 1966), suggesting that even a minimal amount of overburden will result in the plastic deformation of the Cabot Head Shale near the scarp face.

Hewitt (1997) has further developed the idea of plastic deformation within the shale strata. The similar morphologies of the cliffed sections of the escarpment between Hamilton and Collingwood seem to indicate a commonality of process in the post-glacial development of these areas. The absence of the subsequent streams dismissed the possibility of undercutting, so an endogenetic process seemed to provide the best explanation for the landforms present (Hewitt, 1997). These landforms include the deep crevice caves which run parallel to the strike of the scarp face, the divergence of the dolomite blocks at the face of the scarp from the regional dip due to rotation, the presence of many large dolomite blocks on the talus slope at the base of the cliffs, and the projection of the weaker shale strata of the Queenston Formation from the base of the escarpment. It has been argued that the opening of these crevice caves must be due to the rotation of the blocks away from the scarp face which necessitates some kind of deformation at depth (Hewitt, 1997). Due to the brittle nature of the dolomite which makes up the cap rock, it is probable that this deformation is occurring within the weaker shale formations that underlie the cap rock (Hewitt, 1997). Repeat surveys conducted on some of these blocks indicates that they are still in motion (Hintz, 1997). This process is identical to the "yielding flow" landslide described by Pei and Tianchi and is illustrated in figure 2.5. The letters along the right hand side of the figure denote the sequence of events. Part (a) denotes the initial form, with a large block of material resting on weak sub-strata. The weaker material is squeezed out towards the free face where it bulges up and out. Parts (b) and (c) show the two possibilities for block motion that exist. In the first instance, the weaker material is squeezed out laterally towards the cliff face inducing the block to tilt out away from the rest of the rock mass. Part (c) shows material being squeezed out laterally but also vertically into the major jointing plane. This results in subsidence and rotation of the block outwards at the base so as to form true crevice caves. Finally, (d) illustrates an example of one type of landform that may result from these processes.

Figure 2.5: Occurrence of a Yielding Flow Landslide
Source: Pei & Tianchi



Other examples of this process can be found at Nant Gareg-Iwyd, near Blaenrhondda, North Glamorganshire in Great Britain (de Freitas & Watters, 1973), and at Ben Lomond, Tasmania (Caine, 1982). At Nant Gareg-Iwyd, the Rhondda Sandstone has toppled due to subsidence of the underlying shale creating a large crevice between the toppled mass and the rest of the formation. A similar occurrence has occurred at Ben Lomond, where the Dolerite cap of the plateau is underlain by softer shales. Once again, deformation within the shales has resulted in toppling. Hewitt (1997) points out that both type (b) and (c) in fig 2.5 are evident along the cliffed sections of the escarpment. This suggests that the escarpment is not simply a relict feature but has continued to develop throughout the Holocene according to the rheological properties of its shale layers. The type of behaviour exhibited by the shales being a function of stress and hydrological conditions prevalent within the escarpment forms (Hewitt, 1997).

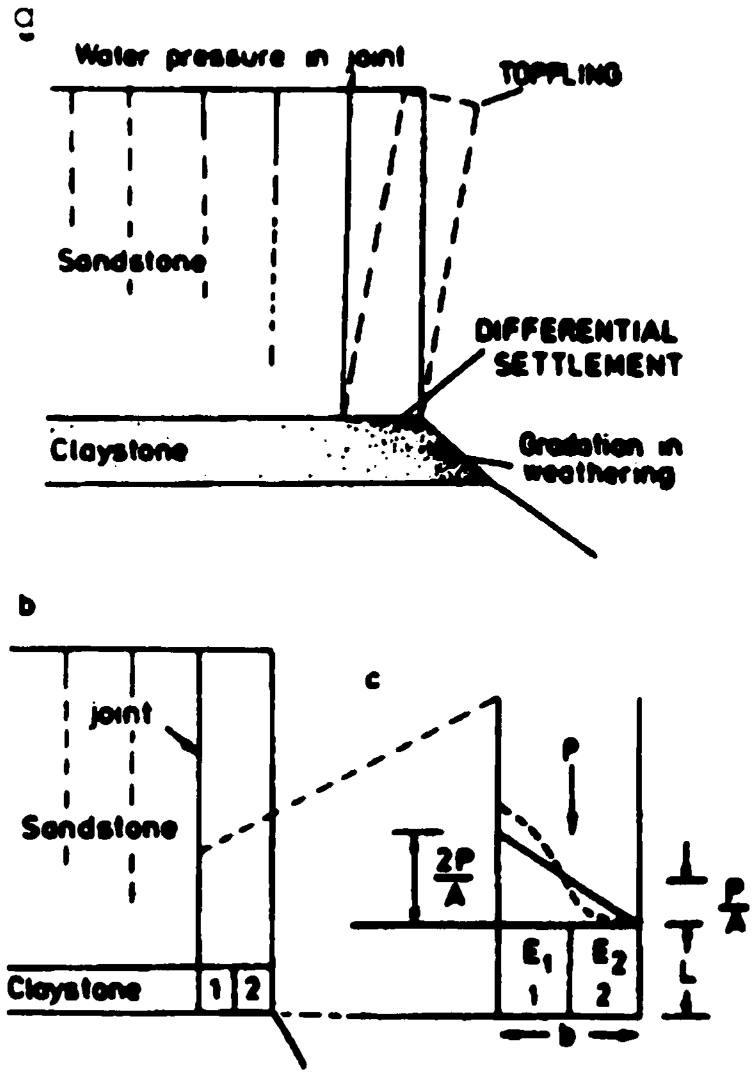
2.2 Studies on Similar Escarpments:

Studies done on other escarpments of similar structure to that of the Niagara Escarpment have illustrated the importance that this landform type can have on the geomorphology of regions throughout the world. The most intensively studied area containing a cuesta-dominated landscape is the Colorado Plateau of the Southwestern United States. The basic geological framework in which these cuestas have evolved is similar to that of the Great Lakes Region. The rocks of the Colorado Plateau are a series of sandstones, limestones, and Basalts interbedded with weaker shales and siltstones. Tectonic activity has created a series of anticlines in the region around which the cuestas have developed (Schmidt, 1989). The cuestas therefore follow the same type of homoclinal recession as that theorized for the scarplands of Southern Ontario. The primary mechanism of scarp recession is considered to be sapping of the cap rock via the weakening or removal of the shale formations by groundwater. Unlike the Niagara Escarpment, there is ample evidence of this type of sapping and indeed many of the canyons that incise the escarpments of the Colorado Plateau have been created via this process (Howard *et al*, 1988). The fact that no such features are evident along the Niagara Escarpment suggests that such processes are not significant in this area. The probable reason for this is the differences in lithology that exist between the two areas. While the Niagara Escarpment is capped by massive carbonates that act to neutralize acidic groundwater, those in Colorado are capped with sandstone which is highly permeable and has minimal buffering capacity.

An excellent body of work exists as to the slope processes of the Illawarra Escarpment in Southern Australia. Here, the stratigraphy is very similar to that of the Niagara Escarpment, being a series of brittle sandstones interbedded with claystones. The slope morphology is also similar with a series of large crevice caves running parallel to the scarp face. A rheological investigation into the properties of the various formations combined with a finite element analysis of the stress environment indicated the possibility of plastic deformation within the claystone layers at the base of the escarpment (Valliappan & Evans, 1982). This outward squeezing of the claystone set up zones of tension within the cap rock of the Illawarra Escarpment which in turn resulted in the crevice caves (Valliappan & Evans, 1982). The morphology of the escarpment is therefore probably an expression of endogenetic processes.

Another possibility theorized for the geomorphology of the Illawarra Escarpment was that of differential settlement (Evans, 1981). This assumes a gradation of weathering within the claystone formations inward from the surface. This gradation is accompanied by a corresponding gradation in the strength of the formation such that the rock at the surface of the scarp face is significantly weaker than that at depth. The result is the isolation and rotation of blocks of sandstone at the scarp face, assisted by hydraulic pressure within the joint planes (Evans, 1981). This mechanism is illustrated in figure 2.6. In part 'a' the outer-most block of sandstone is separated from the rest of the rock mass by a joint and rests

Figure 2.6: Differential Settlement Model
Source: Evans, 1981



upon the underlying claystone. The stress due to overburden applied to the top of the claystone is P/A (Pressure/Area) and is evenly distributed. As the weathering front moves through the rock, the strength properties of the claystone will change such that column 2 in part 'b' of the figure will begin to yield. Part 'c' shows the change in loading across the two columns. The inside edge of column one has a normal stress of 2P/A while the outside of column one has no normal stress. This results in a 'yielding flow' similar to that described by Pei & Tianchi. The process continues until the block passes its centre of gravity and topples to the ground (Evans, 1981).

2.3 The Nature of Rock Slopes:

The stability of rock slopes is governed by both endogenetic and exogenetic processes. A major factor that determines which of these is of greater importance, is the rate at which weathering and mass wasting products are removed from the base of the slope. In cases where the rate of deposition exceeds the rate of removal, exogenetic processes generally dictate the geometry of the slope. Over time, this leads to an accumulation of deposits and the burial of the slope. As a result, most rock slopes are found in environments that are weathering limited. For this reason it is typically endogenetic processes which dictate slope stability on rock slopes. Weathering and erosion do however play an important role in modifying the geomechanical properties of rock slopes and therefore can act as a trigger to change the endogenetic processes involved.

2.3.1: Exogenetic Processes

The study of exogenetic processes is intrinsic to all of geomorphology and will therefore be covered in brief. Those that have the greatest effect on the stability of rock slopes are weathering and erosion. Weathering of rock masses has been discussed in great detail by Curtis (1976), Selby (1982), Faure (1991), and others. It results from the interaction between intact rock and the earth's atmosphere, hydrosphere, and biosphere. Weathering can be physical, chemical or biological in nature. The relative importance of each of these is dependent upon the chemistry of the rock and the climate in which it is located.

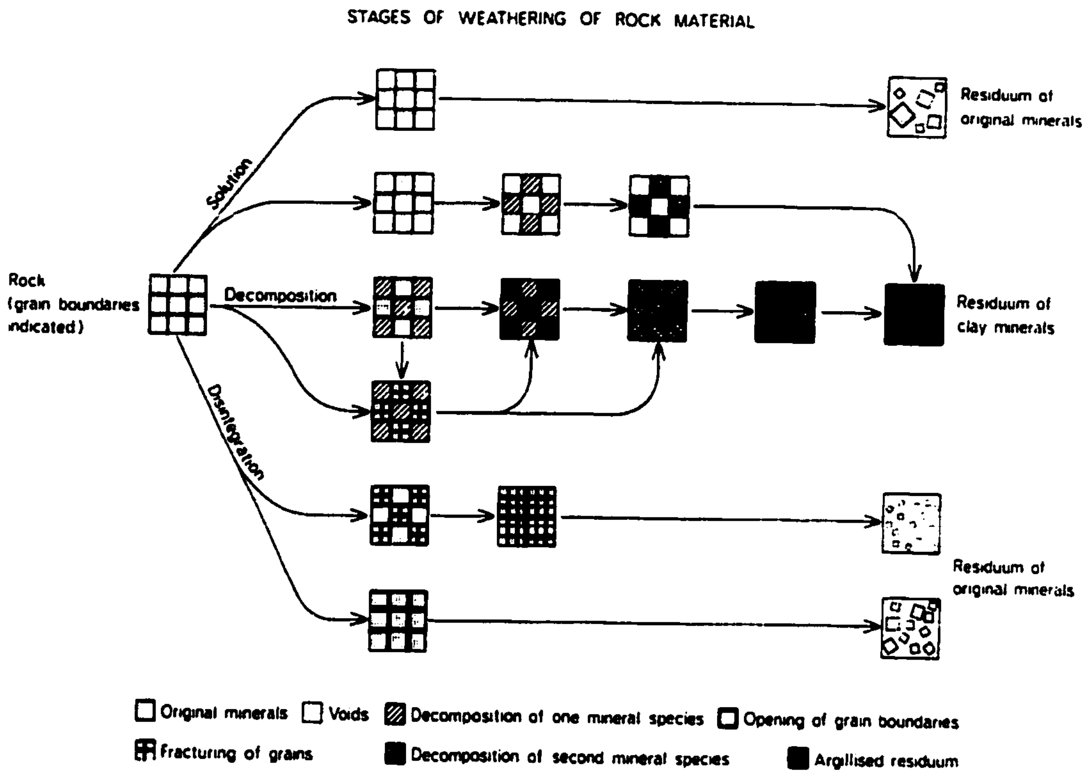
Physical weathering involves the disaggregation of a rock mass into smaller units, which are of the same chemistry as the original. Important types of physical weathering on rock slopes are hydrofracturing, frost action, salt weathering, cyclic wetting and drying, wave action, and insolation. Hydrofracturing results from ice formation within rock fractures. In very narrow fractures, ice formation may force films of liquid water into the tip of the fracture. The hydraulic pressure created by this process can cause the opening and extension of the fracture. Frost action results from the oscillation of the ambient temperature across the freezing point. The more regular this oscillation, the more prone the rock becomes to this kind of weathering. Rocks that are only partially wet are most susceptible. Rapid freezing may result in the collapse, or cavitation, of air pockets within the water filling rock pores. The pressures generated by this process have been calculated to range from 100 MPa - 10 GPa (Selby, 1982). These pressures are easily large enough to fracture rock. Salt weathering involves the formation of salt crystals within rock pores. The salts are typically transported into the pore by solution. Here they accumulate until crystallization occurs. Rocks then tend to fail due to the tensile stress induced by crystal growth, crystal hydration, or the thermal expansion of crystals. The action of cyclic wetting and drying has a highly destructive effect on fine-grained rocks (Mugridge & Young, 1983). The true nature of this process is poorly understood, but the ability of water molecules to separate the sheet silicates, changes in pore pressure, and the swelling properties of some clay minerals are all thought to play a part. Wave action is of great importance to the stability of coastal cliffs. The hydraulic pressure caused by wave impact and the abrasion of particles held in suspension within the wave cause the disaggregation of the rock at the area of impact. The wave also acts to erode the lose materials resulting in the undercutting of the cliff. Finally, insolation weathering involves the expansion and contraction of rocks in the near surface in response to changes in temperature. In many cases, this results in microfracturing, splitting, and spalling of surficial rocks. The process is of greater effectiveness if the change in temperature is rapid such as the cooling of heated rock through precipitation.

Chemical weathering usually takes the form of solution, however, the processes of hydrolysis, and oxidation/reduction can also be of importance. Solution involves the solubility characteristics of the rock mass. Water entering a rock mass is almost always acidic due to its reaction with atmospheric CO_2 to form carbonic acid. The pH can be lowered further due to the production of organic acids within decomposing organic material. The elements that are highly soluble in acidic water include Ca, Mg, Na, and K. Minerals that possess these elements in abundance such as calcite, dolomite, and feldspar are therefore more susceptible to chemical weathering than are others. Hydrolysis is the process by which water acts as a reactant with a mineral instead of as a solvent. This includes the exchange of cations located within the crystal structure of minerals with H^+ and the removal of OH^- molecules from crystals due to bonding with H^+. Oxidation and reduction are the processes by which electrons are either added to an atom (reduction) or removed from an atom (oxidation). This is of importance to the weathering of rock slopes, as the reduction/oxidation of cations can result in the destabilization of the charge balance within the crystal structure of some minerals, most notably the sheet silicates.

Biological weathering is a combination of both physical and chemical weathering produced due to the presence of life. The growth of roots and the burrowing activity of animals can cause the fracturing of rock. Chemicals produced by organisms by way of their life cycles can also have an effect on the chemistry of both the groundwater and the rock. As such, biological weathering can be of great significance to the stability of rock slopes, particularly in temperate and tropical areas.

The various processes discussed eventually lead to the destruction of the rock mass on which they are acting. Once weathered, the material is available to be transported away by the erosive properties of wind and water. The processes of weathering and their effect on rock are shown diagramatically in figure 2.7. As can be seen, the process is typified by the weakening of the rock due the opening of grain boundaries or the removal of a cementing agent, which leaves an insoluble residuum. These processes lead to a change in the rheological properties of the material and can therefore lead to an endogenetic response.

Figure 2.7: The Progression of Weathering Due to the Effect of Various Processes.
Source: Selby, 1982.



2.3.2: Endogenetic Processes:

Endogenetic processes are a function of the response of the slope material and structure to the stress environment. The type of response is governed by the rheological properties of the rock, the orientation and distribution of fractures in relation to the slope face, and the geometry of the slope. The stresses incident on any point within a rock mass are a combination of overburden, confining pressure, tectonic forces, hydrostatic pressure, and residual stresses inherent to the rock.

The science of rheology studies the behaviour of materials under stress. Stress is a function of the force per unit area acting on an object. It is measured in newtons per square metre (N/m^2) or pascals (Pa) such that $1N/m^2 = 1Pa$. When under stress, materials experience strain. Strain is the ratio of change in a rock from its initial condition to its stressed condition. It is therefore dimensionless, but gives an indication of the magnitude of the deformation that a material will experience at a given stress (Means, 1976). The types of stresses that are of importance to slope processes are shear, tensile and compressive. Perhaps the most important of these is shear stress. This is a force which acts to deform a body by having one part slide over another. This usually results in a change in the shape of a rock mass without a significant change in its volume. Shear stress is applied either by two forces acting not in line or by compression (Selby, 1982). Tensile stress is composed of two divergent stress vectors. These act together to pull apart a rock mass. In rocks, tensile stress in usually accompanied by brittle fracture. Lastly, the weight of overlying materials on a volume of rock causes a compressive stress. Failure occurs due to the internal collapse of the rock's structure. This includes the fracture of mineral grains and movement along grain and crystal boundaries. The compressive strength of rocks is usually greater than the shear strength which, in turn, can be up to ten times greater than the tensile strength (Jaeger & Cook, 1979).

The rheological responses that a rock has to the different types of stress depend on the geotechnical properties of the rock, the magnitude of the stress, and the relation of the principal stress to the rest of the stress environment experienced by the rock. As illustrated in figure 2.8, there are generally three types of behaviour exhibited by rock under stress. Elastic deformation is typically the first type of behaviour exhibited. This is deformation which is recoverable if the stress is removed. If more stress is applied to a material, it may begin to deform plastically. The point at which this occurs on a stress versus strain curve is known as the yield point. Plastic deformation results in the shape of the rock mass being permanently altered. In many rock slopes, plastic deformation can have a significant impact on slope processes, especially for slopes that possess rigid cap rocks underlain by more deformable mudstones (Radbruch-Hall, 1979). Still more stress will typically result in the brittle failure of the rock. Since rocks are generally brittle in nature, the transition from elastic deformation to brittle deformation is usually quite rapid. The strength of rock for the purpose of slope stability analysis is usually taken to be the stress at which brittle failure occurs. The behaviour for rock is illustrated in figure 2.8 in the form of a stress (σ) versus strain (ε) curve. Modifications to the experimental curve of marble due to changes in confining pressure and to that of granite due to changes in temperature are shown in figure 2.9. From this it can be seen that presence of high confining pressures on rock alters its strength characteristics. The numbers shown above each line represent the confining pressure used in that experiment. The lowest two confining pressures used result in a curve that is similar to figure 2.8, however, as the confining pressure increases the

Figure 2.8: Stress Strain Curve for Any Hypothetical Material.
Source: Judson & Kauffman, 1990.

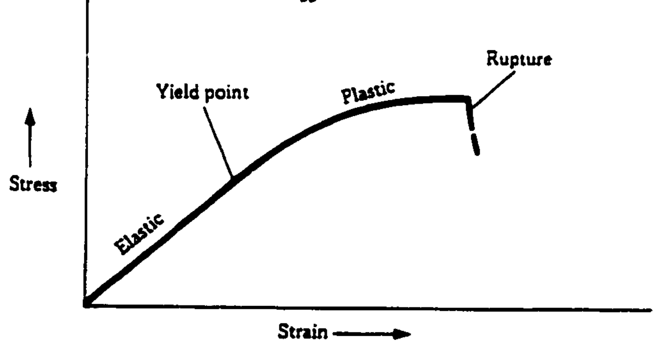
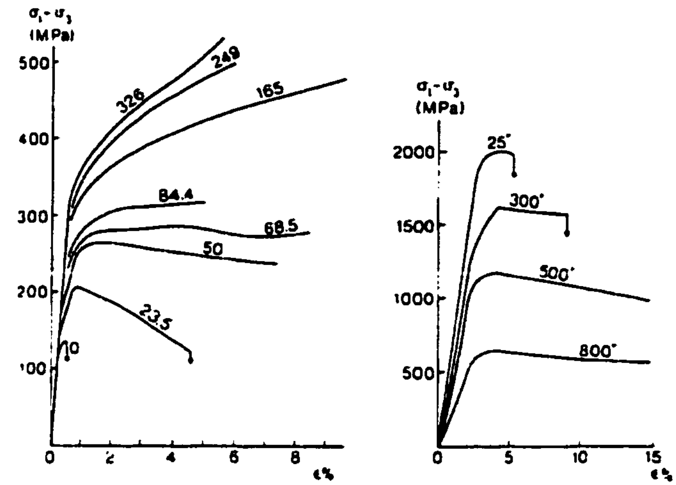


Figure 2.9: Modifications to Stress-Strain Curve Due to Changes in Confining Stress (left) and Temperature (right).
Source: Ranalli, 1995.



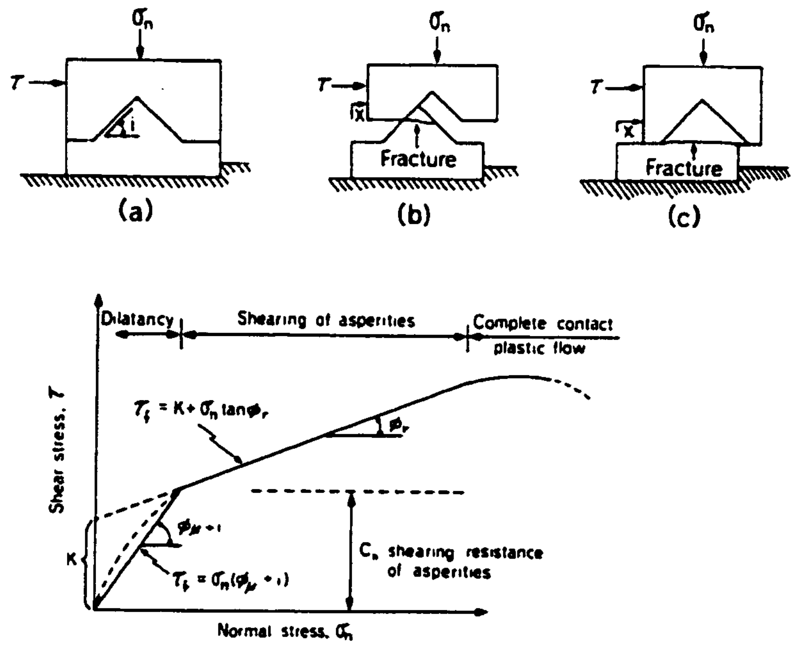
curves indicate an increase in rock strength accompanied by a greater tendency toward plastic deformation. Figure 2.9 also shows the effect of increasing temperature. As can be seen, this tends to result in a decrease in strength and a greater propensity toward plastic deformation instead of brittle failure. On figure 2.8, this would equate to a lowering of the yield point and the extension of the zone of plastic deformation. This type of behaviour under high temperatures and/or pressures shows that rocks may act as a viscous material. This is, however, of little significance to rock slopes except for those that are subject to volcanic action. Another significant variable that alters the strength properties of rock is the pore water pressure. The usual effect of increasing pore pressure is to increase the rock's propensity toward plastic deformation instead of brittle failure (Jaeger & Cook, 1979).

While the strength and behaviour of rock does exert some control on slope processes, in most cases it is the nature and orientation of partings and planes of weakness within the rock mass that exert the primary control on slope stability.

Many different types of partings within rock can be cited; they typically result from either the lithification process, planes of weakness innate to the particular rock type, or post-lithification stresses. The mode of their formation dictates the roughness of the interior planes of rock fractures and therefore has an influence on the shear strength of the fracture plane. Fractures caused by tensile stress are typically rough whereas those resulting from shear stress are smooth. The compressive strength of a joint under a stress that is perpendicular to the plane of the fracture is virtually the same as that of the intact material. The tensile strength of fractures is usually considered to be zero. Evaluating the shear strength of fractures has been of critical importance to many slope stability investigations and has been discussed in great detail by researchers such as Terzaghi (1962), Jeager (1971), Young (1972), Barton (1976), Barton & Choubey (1977), Jaeger & Cook (1979), and Selby (1982, 1987). The shear properties of a fracture are thought to be dependent upon its roughness, the degree to which it has opened, the amount of water present, the amount and type of infill, and the amount of weathering that has taken place within it.

As previously stated, the initial roughness of a fracture is dependent upon the way in which it occurred. Roughness has a large effect on the shear strength of fractures due to the interlocking of asperities located on the fracture surface. The greater the roughness, the greater the number of these asperities, and therefore the higher the shear strength. The relationship between these asperities

Figure 2.10: The Relationship Between Asperities and Fracture Shear Strength.
Source: Selby, 1987.



and the shear strength of a fracture is shown in figure 2.10. From this it can be seen that an increase in shear stress along the plane of the fracture results in a three-step process. The initial phase begins at low stress and results in the dilation of the asperities as they ride up and over one another. As stress increases, the asperities are sheared off. Following this, the rock masses will slide over one another with little increase in the amount of shear stress according to the angle of friction of the rock. At low stresses, when the asperities are still intact, the shear strength of the rock is represented by:

$$\tau_f = \sigma_n \tan(\phi_u + i) \qquad (1)(\text{Selby}, 1987)$$

Where: τ_f is the shear stress at the onset of displacement over an asperity.
σ_n is the normal stress
ϕ_u is the angle of frictional sliding resistance along a plane surface
i is the inclination of the asperity to a common tangent to the base of the asperities.

This relationship is shown graphically in figure 2.3.2.3. As the asperities begin to shear through this relationship changes to:

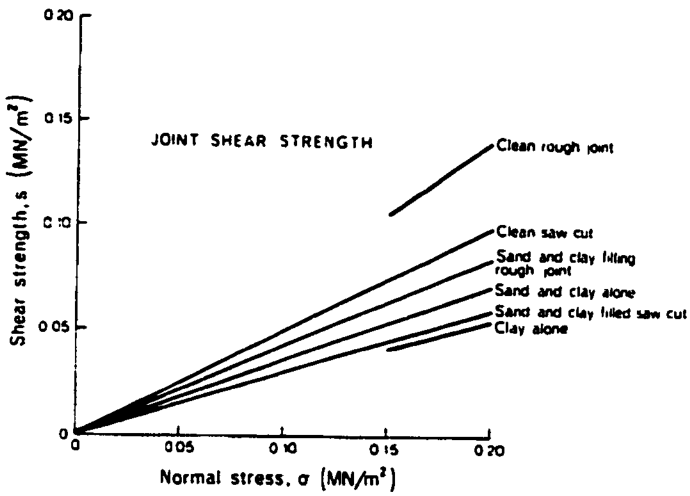
$$\tau_f = K + \sigma_n \tan\phi_r \qquad (2)(\text{Selby,1987})$$

Where: ϕ_r is the angle of residual shearing resistance
K is a constant, describing the apparent cohesion between the rock masses.

The shear strength of a fracture can therefore be drastically reduced once the asperities have been destroyed. The opening up and weathering of the fracture, its infilling with weathering products, and the movement of water within it induce further reductions in shear strength. The opening up of a fracture has obvious implications as less of the rock on the insides of the fracture is in contact.

Once open, a fracture becomes susceptible to weathering processes and infilling. Weathering has the effect of softening the rock along the surfaces of fractures and thereby reducing the shear strength of the fracture as a whole. It also contributes to the accumulation of sediment within a fracture. This process of opening and infilling eventually leads to the shear strength of the fracture approximating that of the disaggregated material filling it. Experimental results proving this relationship are shown in figure 2.11. From this figure it is clear that clean fractures are much stronger than those that have fill. The exact nature of this relationship must be determined experimentally.

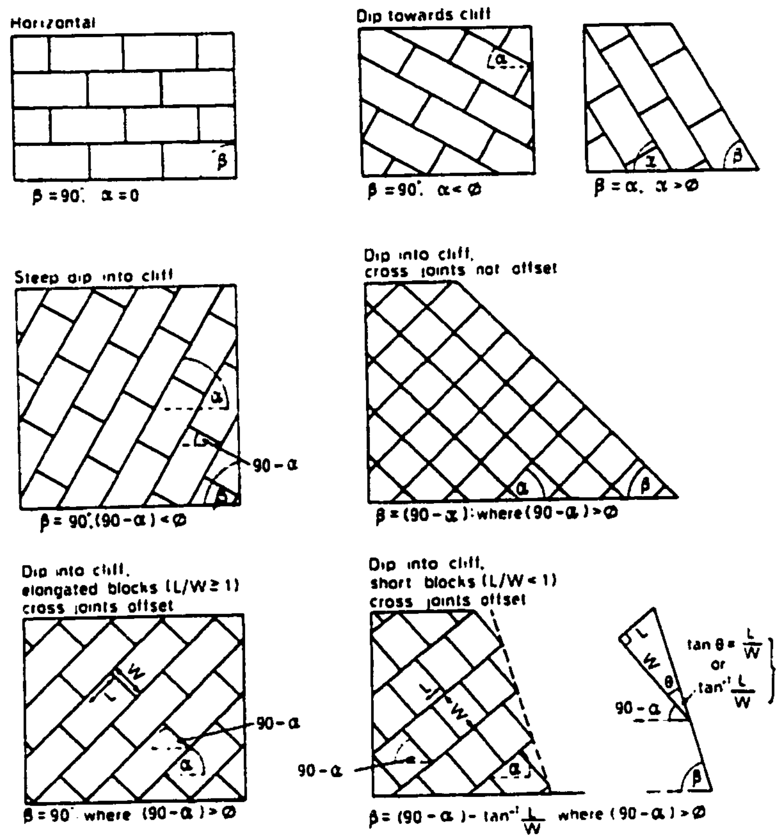
Figure 2.11: Experimental Results Illustrating the Relationship Between Fracture Shear Strength & Fill. Source: Selby, 1982.



The movement of water within a fracture system can have a drastic effect on the shear strength of the fractures involved. This is because water acts as both a lubricant to aid in the sliding of rock fractures and the hydraulic pressure of water at depth alters the stress environment in which the fracture is located. The hydrostatic pressure created by water within the rock mass has a pronounced effect on slope stability. The amount of hydrostatic pressure within the fractures of a rock slope can vary considerably. This is due to the fact that rocks have low permeabilities and therefore tend to channel water into the fracture system. Since the total volume of water which can be held in this system is usually quite low, the large input of water following a precipitation event can cause a dramatic increase in the height of the water table with a corresponding increase in hydrostatic pressure. This can lead, for example, to blocks being pushed towards a cliff face and/or in an upward direction depending on the orientation of the fracture planes.

The shear strength of fractures is of great importance to many assessments of slope stability. The orientation of the fracture planes in relation to the surface of the slope determines the type of stress to which a fracture will be subjected. This relationship between slope stability and the orientation of the fracture planes is shown in figure 2.12. In this figure the fractures are bedding planes and joints. The angle at which the bedding planes dip into or out of the cliff is defined as angle α, the slope angle is β, the angle of friction for the material involved is ϕ, the spacing of the bedding joints is the distance L, and that of the cross joints is W. As can be seen from figure 2.12, a number of situations are possible. The first is the most stable and has bedding planes horizontal to the plane of the earth and perpendicular to the slope face ($\alpha = 0$, $\beta = 90$). The strength of this structure is such that vertical cliffs are often possible. In the next scenario, the bedding planes dip out of the slope at an angle that is less than the angle of friction. This is also a strong arrangement and $\beta = 90°$. In the third case, the bedding planes dip into the slope but the angle of this dip is greater then the angle of friction. This results in failure along the bedding planes such that the angle of the slope becomes the same as the angle of the bedding planes. The next diagram shows the response of a cliff to bedding planes, which dip steeply into its face and in which the cross-joints are offset. Here the angle of the slope will be controlled by the angle of the cross joints ($90-\alpha$) and the angle of friction such that $90-\alpha < \phi$. The fifth example shows bedding planes dipping steeply into the cliff with the cross-joints not offset. The slope angle in this situation is determined by the angle of the cross-joints in a similar fashion to that of the third example. The last two examples show the significance of the spacing between the cross-joints for slope stability. In example six, the ratio of length to width is greater than one. In this

Figure 2.12: Relationship Between the Orientation of fracture Planes to that of the Surface of the Slope and Slope Stability. Source: Terzaghi, 1962.



case, the stability of the slope is dependent upon the angle of the joints in the same way as in example four. In example seven however, the cross-joints are closer together such that the ratio of length to width is less than one. In this case, the angle of the slope is a function of the dip of the cross-joints and the angle created by setting a right triangle with L and W as the short sides. The cases shown in figure 2.12 are all an oversimplification of the natural system, and do not take into account the effects of weathering along joints, variations in block shape, or hydrostatic pressures. However, they are a good illustration of the controlling endogenetic factors that govern slope stability.

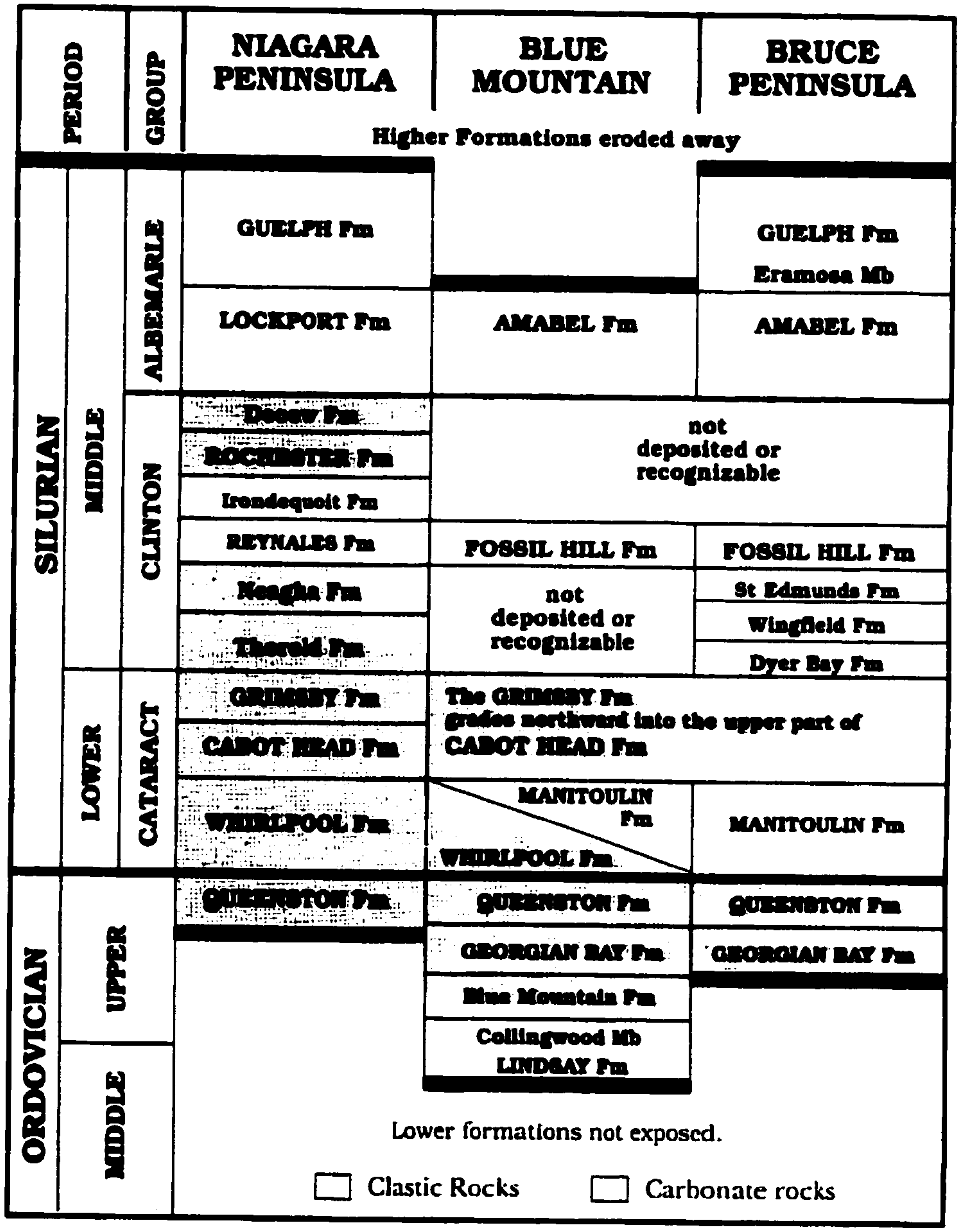
2.4 The Rocks of the Niagara Escarpment:

As previously stated, the stratigraphy of the Niagara Escarpment is extremely complex. The following is an introduction to the various formations in order to put them in context. A detailed description of each of the major formations will follow. The various strata vary in age from upper Ordovician to the middle Silurian and are therefore between 425 and 400 million years old. A simplified schematic of the stratigraphy is shown in figure 2.13. The Ordovician formation that is positioned at the base of the escarpment from the Niagara Gorge to the tip of the Bruce Peninsula, with the exception of Blue Mountain, is the Queenston Shale. The origin of the deltaic sediments that form the Queenston Shale is thought to be the Appalachian Mountains that were undergoing upheaval during the Taconic Orogeny. It has been estimated that roughly 600000 km^3 of sediment were removed from the Appalachians during this period and deposited in marginal seas (Wilcander & Monroe, 1993). This deposition also corresponded with a recession in sea level due to tectonic uplift, which explains the Queenston Formation's large aerial extent relative to its thickness. Decreasing sea level would result in the growth of extensive mudflats over a shallow basin instead of the infilling of a basin due to delta growth (Dennison, 1976). Following the deposition of the Queenston Formation, sea levels rose to submerge it.

Above the Queenston Formation are the rocks of the lower to middle Silurian. These can be broken down into three groups: the Cataract Group, the Clinton Group, and the Albermarle Group. The Cataract Group consists of the Whirlpool, Manitoulin, Cabot Head, and Grimsby Formations. The Whirlpool

Formation is a white to light gray sandstone, which follows a sharp contact over

Figure 2.13: Stratigraphy of the Niagara Escarpment
Source: Tovell, 1992



the Queenston Shale. It is also considered to be deltaic in origin although not as widely spread as the Queenston Formation. To the north, the Whirlpool Formation grades into the Manitoulin Formation that is composed of fossiliferous dolomite. It overlies the Queenston Formation from the tip of the Bruce

Peninsula to Osler Bluff near Collingwood. South of Osler Bluff, the Manitoulin Formation begins to thin and is underlain by the Whirlpool Formation, which generally thickens as one moves south, such that at the Dundas Valley the Manitoulin dolomite has disappeared (Bolton, 1957). Such a facies change is interesting when one considers that the two formations were lain down contemporaneously. The Whirlpool formation was deposited as a delta on the coastal area of a shallow sea while the Manitoulin Formation was being biogenically produced further off shore. A slow reduction in the supply of clastic sediments supplied from the southeast is thought to be responsible for the southward expansion of the Manitoulin Formation over the Whirlpool Formation (Tovell, 1992). Over the Whirlpool/Manitoulin Formations lies the Cabot Head Formation. This is a 15 m layer of greenish-gray and red shale interbedded with limestone. It too is deltaic in origin, although the facies change indicates periodic decreases in sediment supply such that the depositional environment switched to the biogenic. The upper part of the Cabot Head Formation grades laterally into the Grimsby Formation, which is a red and green sandstone, interbedded with shale. The Grimsby Formation is 14 m thick at the Niagara Gorge but thins to the north so that it is only 9 m thick at Hamilton and 1 m thick at the Nottawasaga River. This formation is therefore considered to be deltaic as well, receiving nourishment from the south-east.

Occupying the stratigraphic position above the Cataract Group is the Clinton Group, which is highly complex both vertically and laterally. It is composed of many shales interbedded with limestone and dolomite. In all, 16 formations make up the Clinton Group some of which are statigraphically equivalent although not all of them are present in Ontario (Bolton, 1957). A general trend within the Clinton Group is the increasing proportion of carbonate deposits as one moves north. In the Niagara Peninsula, the rocks of the Clinton Group are 80-85 % clastic and 15-20 % carbonate; whereas in the Bruce Peninsula they are 35-40 % clastic and 60-65 % carbonate (Tovell, 1992). This trend is consistent with that of the Cataract Group, which infers a deltaic accumulation of clastic sediments in the south and a biogenic accumulation in the north within a sea that was marginal to the Appalachians of the Taconic Orogeny.

The Albermarle Group represents the cap rock of the present day Niagara Escarpment. It consists of the massive dolostones of the Lockport-Amabel Formation and the Guelph Formation. The Lockport-Amabel Formation makes up the cliffs of the Niagara Escarpment along its entire length. The formation is composed of the Lockport Formation to the south of Hamilton and the Amabel Formation to the north. They are both highly fossiliferous and contain many bioherms along the length of their outcropping. The reefs present consist of patch reefs, barrier reefs, and pinnacle reefs indicating the presence of a number of marine habitats at the time of deposition (Tovell, 1992). The Guelph Formation overlies the Lockport-Amabel Formation and is exposed to the west of the escarpment face. It too is a gray dolostone of the middle Silurian containing multiple bioherms (Bolton, 1957).

2.4.1 The Queenston Formation

The Queenston Formation is a shale layer that outcrops all along the Niagara Escarpment. In the region from the Dundas Valley to the Forks of the Credit, the Queenston Formation is present as a rather broad belt of red shale. Where exposed on the basal slopes of the escarpment, the shale is usually highly eroded (Chapman & Putnam, 1984). At the Forks of the Credit, the shale begins to rise in elevation as it climbs the eastern flank of the Algonquin Arch such that at Blue Mountain, it is found 300 m above Georgian Bay (Tovell, 1992). The sediments from which the shale is derived originated in the Ordovician and were lain down as an expansive coastal delta during a period of receding sea level (Dennison, 1976). Due to this mode of deposition, the Queenston Formation thickens to the south such that it is 243 m underneath the north shore of Lake Erie but only 61 m thick at the base of the Bruce Peninsula (Brogly *et al*, 1998). Throughout the region, the Queenston Shale overlies the shale of the Georgian Bay Formation, which occupies the position at the very base of the escarpment at Blue Mountain near Collingwood.

The Queenston Formation is mainly composed of clay minerals; being 60 % clay minerals and 40 % rock forming minerals. The most abundant of the clay minerals is illite, followed by chlorite with only traces of vermiculite. The most significant of the rock forming minerals are quartz, calcite, and dolomite with traces of feldspar. The facies typically encountered at the base of the escarpment is the C1 facies identified by Brogly *et al* (1998). This is a medium bedded shale with thin beds of calcareous siltstone. At Milton, and elsewhere, facies C3 can also be found. This is a thickly bedded red shale with thin beds of bioclastic siltstone (Brogly *et al*, 1998). The most distinctive feature of these outcrops is the presence of horizontal and vertical layers of blue shale. These are the result of circulating ground waters, containing concentrations of organic acids, that reduce the iron within the clay minerals from Fe^{3+} to Fe^{2+} (Guillet, 1967).

2.4.2 The Whirlpool Formation

Overlying the Queenston Formation with a sharp contact is the Whirlpool Formation. This is a Silurian Sandstone that ranges in colour from white to light pinkish-brown. It is composed of a thickly bedded, fine grained quartz cemented by calcite (Bolton, 1957). The formation is 6 m thick at the Niagara Gorge and thins to the north such that it disappears north of Osler Bluff (Tovell, 1992). It outcrops at many locations along the base of the Niagara Escarpment most notably at Milton, where it can be found in a series of old quarries, and around Collingwood, where it is often present with the Manitoulin formation forming a secondary scarp below the main scarp.

2.4.3 The Manitoulin Formation

The Manitoulin Formation is a section of blue-grey dolomitized limestone of variable bedding (Bolton, 1957). The formation is 20 m thick over the Bruce Peninsula (Tovell, 1992), and thins to the south such that it disappears at Stoney Creek (Bolton, 1957). North of Osler bluff, the Manitoulin formation overlies the Queenston formation but to the south of this location it is underlain by the Whirlpool Formation. To the north of the Bruce Peninsula, the Manitoulin formation extends across Manitoulin Island and into Northern Michigan State

Bolton, 1957). The exposures evident along the base of the escarpment from Hamilton to Collingwood often contain some grey shale partings between bedding plains (Bolton, 1957). Combined with the Whirlpool formation, the Manitoulin formation often caps a secondary scarp at the base of the main escarpment, particularly in the Collingwood area.

2.4.4 The Cabot Head Formation

The Cabot Head Formation is a blue-green shale interbedded with layers of limestone (Tovell, 1992). The base of the formation grades into the Manitoulin Formation with which it shares roughly the same area of extent (Bolton, 1957). At the base of the Bruce Peninsula, the Cabot Head Formation is roughly 36.5 m thick and thins to the south such that it is only 15 m thick at Stoney Creek (Bolton, 1957). The upper contact is sharply defined with the Reynales Formation and Fossil Hill Formation but to the south it is gradational with the sandstones of the Grimsby Formation (Bolton, 1957). Exposures of the Cabot Head Formation are few in number and usually quite poor in the Milton-Hamilton area. The best exposure is that reported by Kor (1991) for the outcrop located along the streambed in Spencer Gorge. Very little has been written regarding the Cabot Head Formation, which is perhaps a reflection of its lack of exposure. There is however a general consensus that the Cabot Head shale becomes plastic when wet (Tovell, 1992, Bolton 1957).

2.4.5 The Reynales Formation

This dense, thickly bedded dolomite underlies the Amabel Formation from Rochester, New York to Georgetown. The formation retains a fairly uniform thickness of 2.5 - 3 m and contains few fossils (Bolton, 1957). Outcrops of the Reynales Formation can be found on the floor of the abandoned Quarry located within the Kelso Heights Conservation Area at Milton (Kor, 1991), as well as in several other dolostone quarries located throughout the area (Hewitt, 1960). There are very few natural exposures along the Niagara Escarpment due to burial beneath the talus slope.

2.4.6 The Fossil Hill Formation

The Fossil Hill Formation represents the northern equivalent to the Reynales Formation in the stratigraphic sequence. The transition takes place to the south of Caledon where the Fossil Hill Formation follows a gradational change into the Reynales Formation (Bolton, 1957). It is similar in appearance to the overlying Amabel Formation, being an unevenly bedded tan-brown dolomite (Tovell, 1992). The contact with the Amabel Formation is sharp and clearly defined as is that with the underlying Cabot Head Shale (Bolton, 1957). The Fossil Hill Formation is generally thinly bedded and fine grained with a large number of fossils including corals and brachiopods (Bolton, 1957).

2.4.7 The Amabel Formation

The Amabel Formation is the northern equivalent of the Lockport Formation and the two grade into each other from Hamilton to Georgetown. They are typically grouped together to make the Lockport-Amabel Formation (Tovell, 1992). It forms the cap rock of the escarpment from here up to the northern portions of the Bruce Peninsula where it is overlain by the Guelph Formation. For the purposes of this study, it is therefore the uppermost formation in the lithological sequence. It is a thin to massively bedded dolomite with an abundance of bioherms and other fossils (Tovell, 1992). Along the Niagara Peninsula and the Bruce Peninsula the Formation is composed of a series of distinct members but the section in between is generally considered a single unit (Kor, 1991b). There is however a great deal of variation in the appearance of the rock. In many areas, is a massive crystalline dolostone that is grey in colour with many fossils. Elsewhere, it is a brownish-grey dolostone of fine-grained texture with no visible fossil structures.

Chapter Three: Methods

3.1: Introduction

The progression followed through this project was one of field reconnaissance, field surveying, sample taking, sample preparation and testing, and data manipulation/integration. Field reconnaissance was conducted during the spring and summer of 1998. Site surveying took place during the fall of 1998, while sampling occurred during both of these periods. Sample testing and data manipulation were accomplished during the spring and summer of 1999.

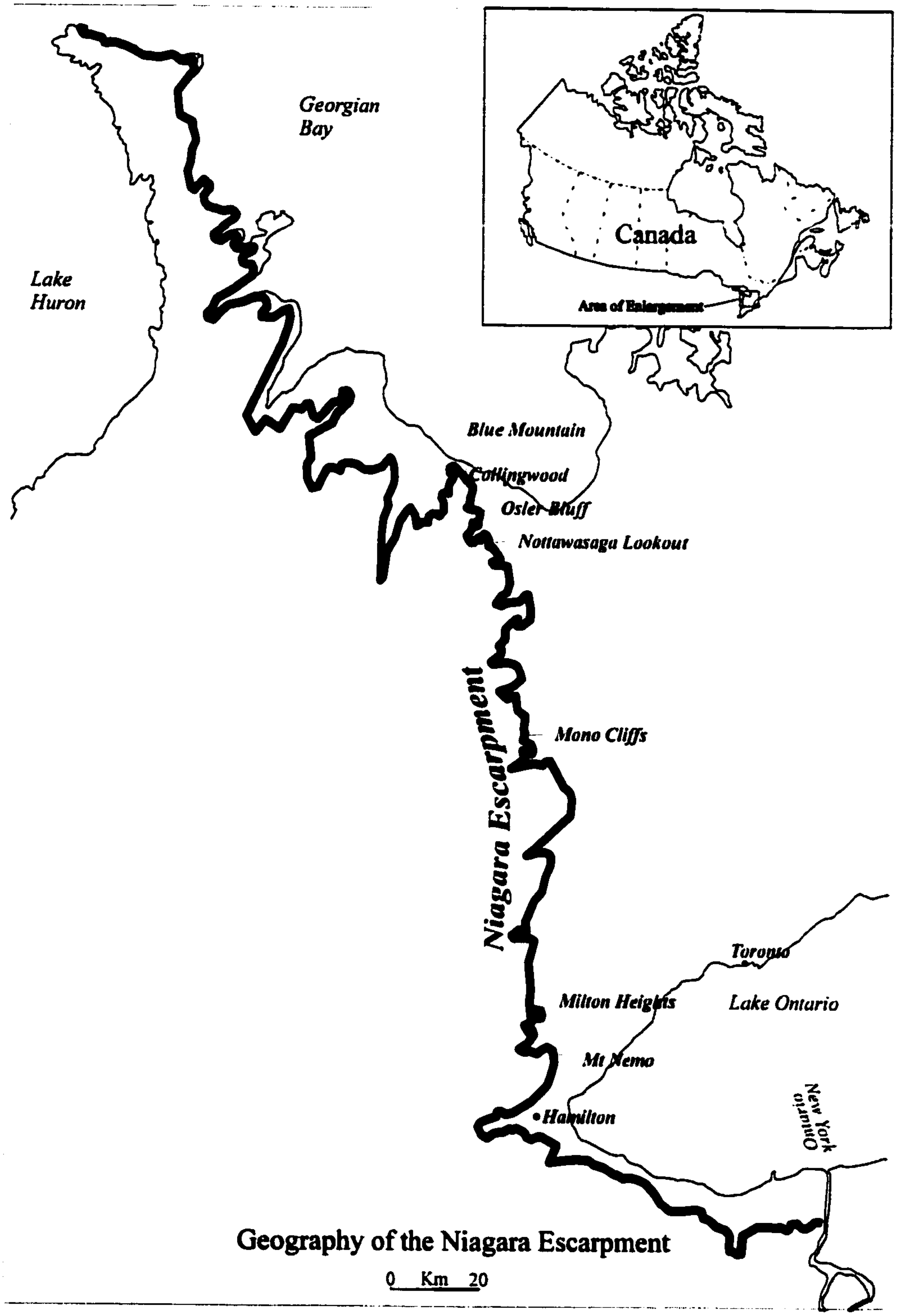
3.2 Field Site Selection

Field reconnaissance was conducted along all of the cliffed sections of the Niagara Escarpment between Mt. Nemo to the south and Blue Mountain to the north (Fig. 3.1). In particular, five areas were examined in detail: Mt. Nemo, The Milton Outlier and surroundings, the Mono Cliffs, the Nottawasaga Lookout, and Osler Bluff/Blue Mountain. A brief description of each will follow. The sites selected for further research were those at Mt. Nemo, Milton, and Osler Bluff.

3.2.1 Mt. Nemo

The site at Mt. Nemo is the eastern portion of an "attached outlier" separated from the main escarpment by a narrow channel that has subsequently been filled with glaciofluvial deposits. This site represents the southernmost portion of the escarpment that retains a relatively simple stratigraphy. It possesses a number of well-developed crevices, high cliffs, a blocky talus slope

Figure 3.1: Site Locations Along the Niagara Escarpment



and an abandoned quarry at the brow of the escarpment. The area was inspected and documented by Kor (1991b). Outcrops beneath the Amabel Formation are buried beneath the talus slope but the surrounding area has several sites at which the various formations can be sampled. The most notable of these is Spencer Gorge, located to the southwest, which shows the entire Silurian stratigraphy from the Lockport-Amabel Formation to the Whirlpool Formation and the underlying Queenston Formation (Kor, 1991a). Located to the northwest of Mt. Nemo, near the village of Kilbride, there is a good exposure of the Silurian - Ordovician contact. This occurs at a small waterfall at the end of #8 sideroad. Unweathered outcroppings of the Queenston Shale and the Whirlpool Sandstone may be found in this area.

3.2.2: Milton Heights

Milton Heights is located on the northern portion of the Milton Outlier and separated from the main body of the escarpment by the Nassagaweya Channel (Tovell, 1992). The site is part of the Kelso Conservation Area, and contains tall cliffs in the Amabel formation with accompanying crevice caves and blocky talus slopes. Exposures of all of the formations within the escarpment are present (Kor, 1991c). A number of abandoned quarries exist on the brow and at the base of the escarpment within the conservation area. Along the base of the talus slope, a series of abandoned quarries are cut into the Whirlpool Formation and the Manitoulin Formation. The Milton Limestone quarry is located just outside the conservation area and occupies a large proportion of the outlier between Kelso and Rattlesnake Point to the south. The quarry cuts through the Amabel

Formation and is floored in the Reynales Formation (Hewitt, 1960). Also outside the conservation area is the Canada Brick quarry which is cut into the Queenston Formation. The exposed shale within is typically weathered and large samples of intact rock can not be found. Along the talus slopes are areas marked by blue clay, which denote the presence of the underlying Cabot Head Shale. Unfortunately, these exposures are too weathered to sample

3.2.3: Mono Cliffs

Mono Cliffs represents the only major cliffed section of the escarpment between Milton and the Collingwood Area. It is located near Mono Township to the north of Orangeville. A detailed inventory of the exposures in the area is given by Kor (1991c). The area also has well developed crevice caves and talus slopes that are ubiquitous to all of the cliffed sections of the escarpment between Hamilton and Collingwood. The area is made up of a long series of low cliffs with two outliers located to the east. The presence of groundwater can be detected throughout, and the base of the slopes is typically wet with many ponds and marshy areas. In some places, particularly the southern outlier, the talus slopes show a marked degree of soil formation indicating a transport-limited environment. While this area does share the same basic morphology as the rest of the steep sloped sections of the escarpment, it is not as well developed as the other sites and was therefore discarded.

3.2.4 Nottawasaga Lookout

This long section of cliffs is located on the southern side of the Pretty River Valley, to the south of Osler Bluff. There are exposures of the Amabel, Whirlpool, and Manitoulin Formations along the top and base of the slope (Kor, 1991d). The latter two of these are typically encountered as a secondary scarp at the base of the main scarp. The crevice caves found in this area are the largest and deepest of any of the sites. However, the blocks do not show as much evidence of cambering as those encountered at Osler Bluff, and seem to be simply expanding to the east. This site was therefore overlooked in favour of the site at Osler Bluff.

3.2.5 Osler Bluff

Located between Blue Mountain to the north and the Pretty River to the South, Osler Bluff is capped by a series of Cliffs in the Amabel formation. This marks the northernmost extension of the Whirlpool Formation and possesses outcroppings of the Amabel, and Manitoulin Formations. The Manitoulin Formation is typically encountered in the form of a secondary scarp at the base of the talus slope. The blocks slumping from the face of the escarpment show a high degree of cambering with bedding planes clearly dipping to the east. This cambering has been observed to continue for up to 150 m into the scarp face (Hewitt, 1997). The base of the escarpment here is in the Georgian Bay Formation with the Queenston shale overlying it. Good exposures of the Queenston shale and the Fossil Hill Formation can be found to the north at the Blue Mountain lookout.

3.3 Sampling

Representative samples were taken of each formation for each study site. The exception to this being the Queenston and Cabot Head Formations for Mt.

Nemo and Milton. Here, due to the proximity of these sites to one another, the same sample location was used for both. Obtaining the samples required the use of a geological hammer to dislodge a useable piece from the rock mass. In each case, care was taken to use as little force as possible and to mark the orientation of the sample *in situ*. The sample size was generally determined by the need to maintain a 2:1, height : width ratio for the cores. Ideally, samples with a height of between 10 and 15 cm were preferred. However, in cases where the parent material was too thinly bedded to allow this, samples as thin as 8 cm were taken. This of course necessitated the cutting of smaller cores.

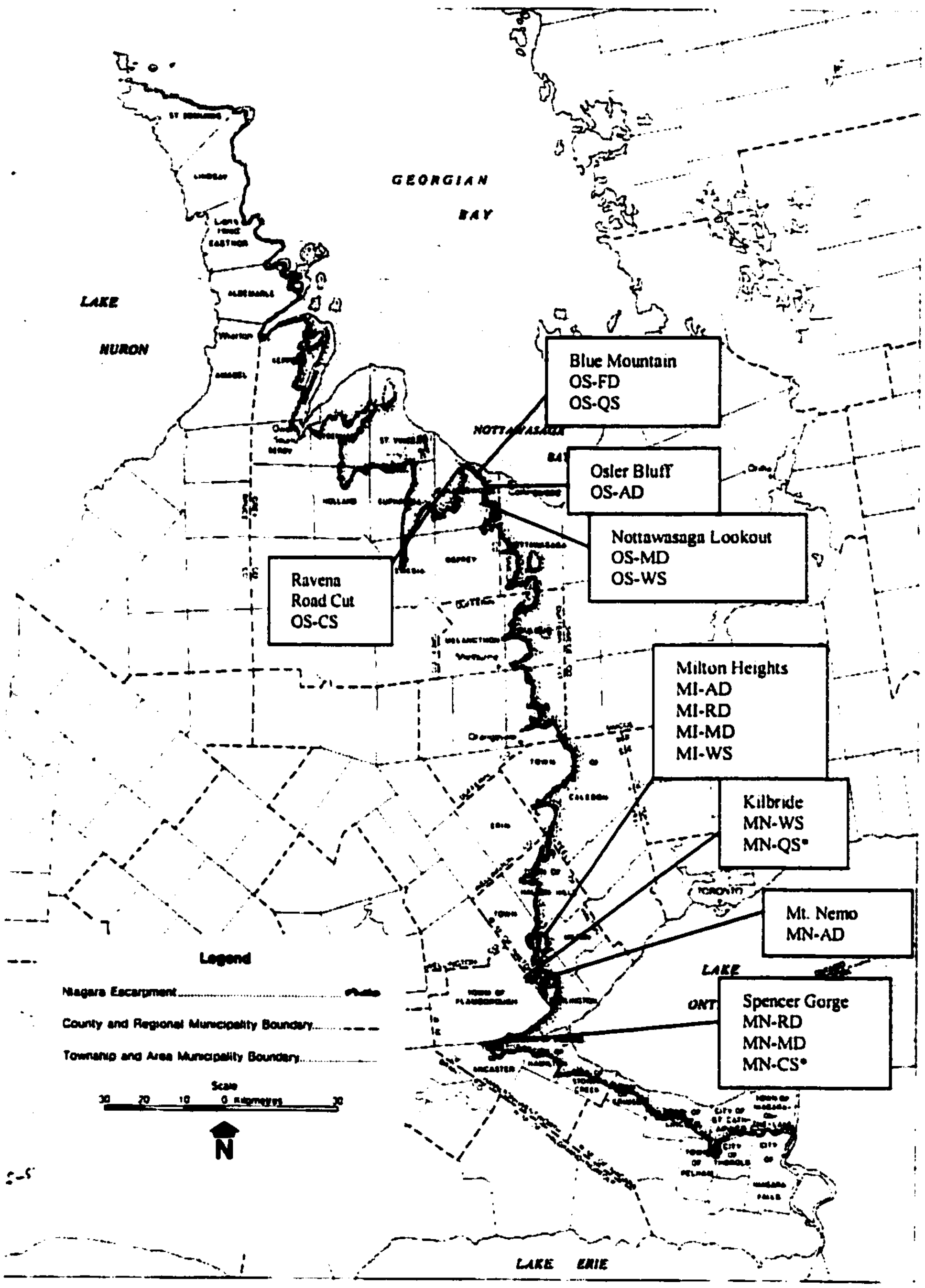
The location of the sampling was done at the closest known outcrop to the study site. These Sampling Locations are shown in Figure 3.1. Once removed, samples were wrapped in plastic in order to retain their moisture content and transported to the rock lab for coring. The sample identification system consisted of a six-character code along the following lines:

1st Two Letters Designate Location
- MN – Mt. Nemo
- MI – Milton
- OS – Osler Bluff

2nd Two Letters Designate Formation
- AD – Amabel Dolomite
- RD – Reynales Dolomite
- FD – Fossil Hill Dolomite
- CS – Cabot Head Shale
- MD – Manitoulin Dolomite
- WS – Whirlpool Sandstone
- QS – Queenston Shale

3rd Two Numbers designate Sample Number

Figure 3.2: Sample Site Locations
Source: Tovell, 1992 (Basemap)



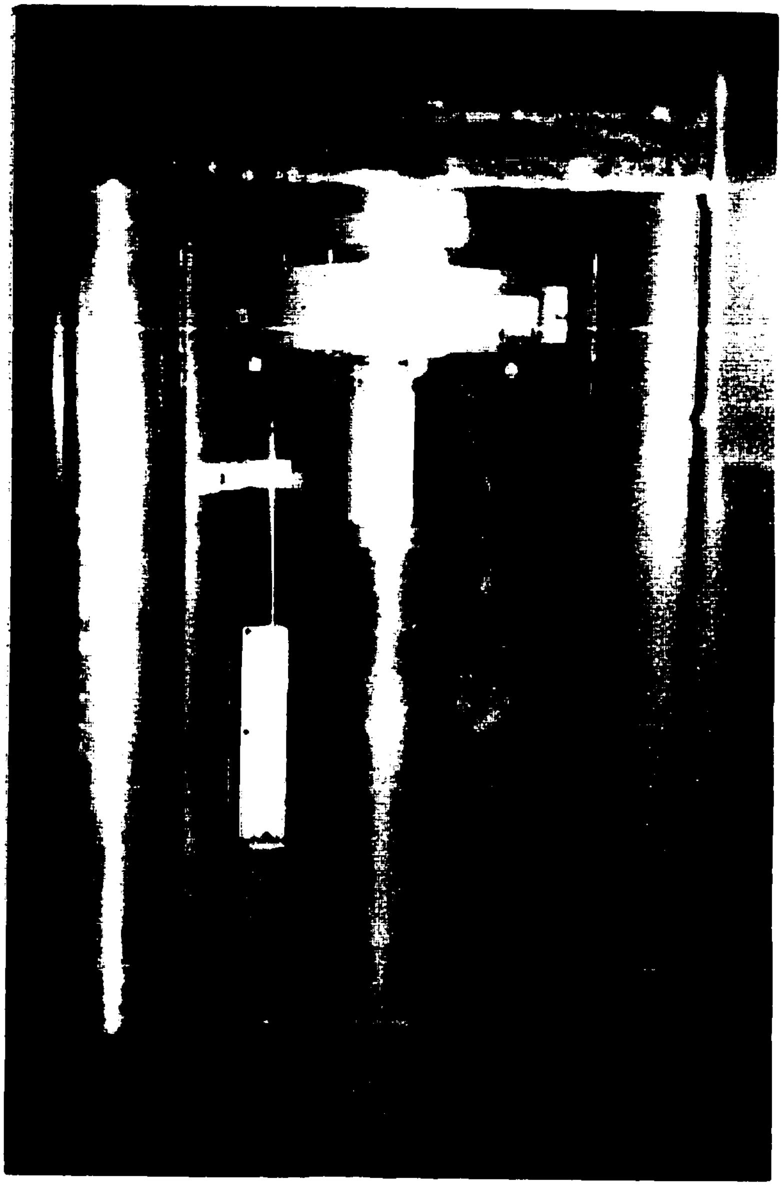
Note: * Used for Milton/Mt. Nemo

3.4 Sample Preparation & Testing

Samples were cored such that the long axis of the core was parallel to the principal stress that would have been experienced by the sample *in situ*. Three different core diameters were used: 4.4 mm, 3.8 mm, and 3.2 mm. The diameter used on any given core depended on the thickness of the sample, as a 2:1 length to width ratio had to be maintained (Dhir & Sangha, 1973). Although the samples taken were as large as possible, in many instances the rock was too thinly bedded to allow for the larger core size. Cores were then capped using a surface grinder.

Once prepared, cores of the dolomite and sandstone formations were dried in an oven at 80 °C for 3 hrs and weighed. They were then saturated with deionized water in order to simulate their position beneath the water table. Once saturated they were re-weighed and compressed in the uniaxial compression machine (Figure 3.3). This involved inserting the core between two plates and increasing the stress at a fixed rate until failure occurred. The amount of strain for a given stress could then be plotted (Appendix 1-3). See Hawkes and Mellor (1970), for an extremely detailed discussion of uniaxial compression testing. The weights were used in conjunction with the calculated volume of the core to determine each sample's dry and wet weight in g/cm^3. The two shale formations could not be dried due to their high susceptibility to slaking. They were therefore weighed while saturated and compressed in the uniaxial compression machine. The bulk density of the dry shale formations was done through displacement techniques.

Figure 3.3: Testing a Core in the Uniaxial Compression Machine.



3.5 Geological Cross Section

In order to gain a better understanding of the stratigraphy at each of the study sites, well water records were obtained through the Ontario Ministry of the Environment. However, due to their poor quality, the well records provided no usable information on the stratigraphy of the area. Therefore, this had to be approximated from geological maps and data found within the literature. When combined with detailed cross sectional surveys of the three study areas, a fairly accurate representation of each was created. The surveys were done using a theodolite and in some instances, where the vegetation was too thick for line of sight, a tape and inclinometer.

The cross sections were then used to produce a plot of overburden stresses within the escarpment. This was done by calculating the weight of a column of rock, based on the dry weight, on points within the cross-section. When analysed using Microsoft Surfer's linear interpolation software, these plots created detailed cross-sectional maps of the stress due to overburden within the entire cross section. Recent work has shown that the actual stress environment acting on a portion of rock within a slope is determined by several factors. These include overburden, slope geometry, tectonic forces and the rheological properties of the rock (Silvestri & Tabib, 1983). Overburden alone therefore does not give an indication of the true stress environment. However, it does give an excellent idea as to the magnitude of the principal stress acting on each of the formations. In order to determine the orientation if the stresses, methods such as finite element analysis would be required at a substantial cost.

Based on the rheology of the rock formations, the amount of overburden, and the empirical evidence described by Hewitt (1997) a conceptual model was developed to explain the endogenetic processes within the escarpment. For this model, the following assumptions are inherent:

1. The rocks were considered to be continuous and homogeneous.
2. Hydraulic Stresses were considered to be nil.
3. Tectonic Stresses were considered to be nil.
4. All rocks were considered saturated with water due to the estimated position of the water table.

Of these, numbers 1-3 are conservative in that they serve to increase the stability of the slope. Number 4 is based on the model put forward by Nadon & Gale (1984), and from field observations. Due to the high hydraulic gradient near to the cliff face, a portion of the cap rock may be unsaturated in this area. However, the majority of this formation remains below the water table and the inherent strength of the Amabel formation, even when saturated, justifies this assumption. The stresses used to develop the model are therefore the minimum that would be expected within the escarpment.

Chapter Four: Results

4.1 Introduction

In this chapter, the field and laboratory work produced for each of the three field sites will be conveyed and discussed. Although the types of analysis done on each of the field sites were identical, they are submitted separately with a final discussion at the end. Each section will contain a map of the study site, a geological cross section, the geotechnical and rheological data for the various rock formations, and the analysis of overburden within each cross section.

4.2 Mt. Nemo Study Site

A map of the Mt. Nemo study site is shown in figure 4.1. This is a detail of the northern portion of the Mt. Nemo "attached" outlier. The location of the cross section is also shown. It runs in a northerly direction, from near the northwest corner of the abandoned quarry, towards Britannia Rd. The geological cross section of Mt. Nemo is illustrated in figure 4.2. In this, the southern-most cross section, the morphology is fairly simple. The cap rock is broken by a series of crevices running parallel or sub-parallel to the escarpment face (figures 4.2 - A. & B.). Below the cliff face a steep talus slope composed of Amabel Dolomite drops to the basal slope below (figure 4.2C.). The size of the blocks ranges from loose gravel to blocks that are several meters long on each axis. The convex shape of the talus slope in profile is highly unusual (Selby, 1993). This may be

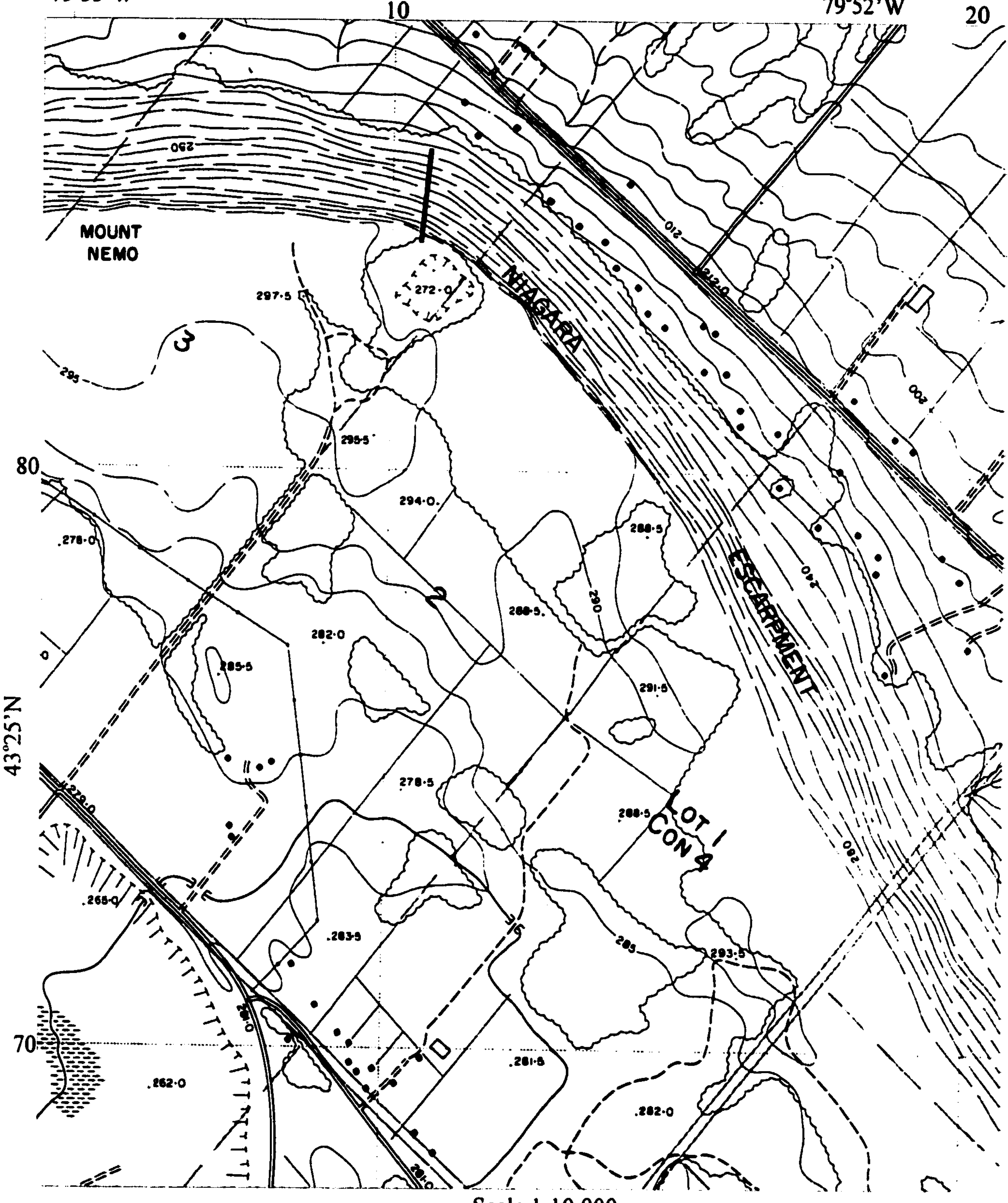


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Figure 4.1: Mount Nemo Study Area

Legend

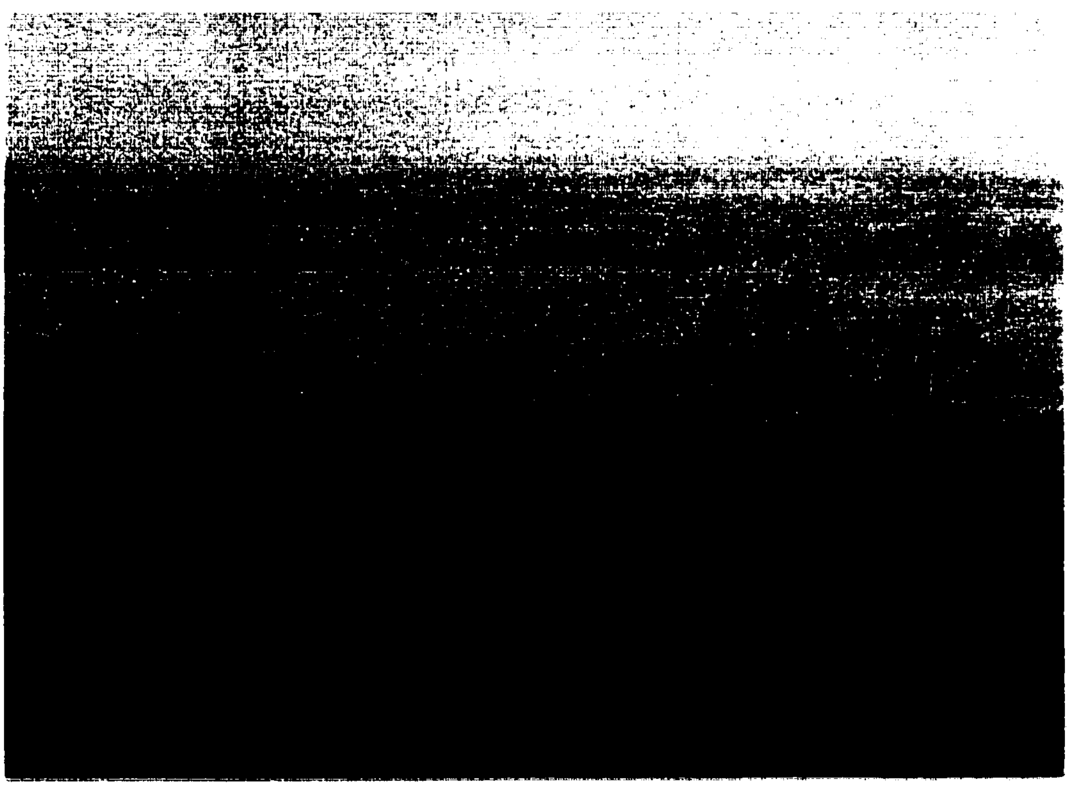
Location of Cross Section

Detail from O.B.M. 10 17 5900 48050

Notes:
North American Datum 1927
Universal Transverse Mercator (6°) Projection
Zone 17. Central Meridian 81°W
Grid Interval 1000 m

Figure 4.2C: Talus Slope at Mt. Nemo.

Figure 4.2D: Basal Slope, Mt. Nemo



explained however by the lateral expansion of the Cabot Head formation, which is stratigraphically equivalent to the talus slope. The talus slope is also heavily vegetated with both white cedar and a mixed deciduous forest. The basal slope is more gently inclined and is in the Queenston Formation (figure 4.2D).

Samples taken from the nearest outcroppings of the formations involved revealed the following characteristics:

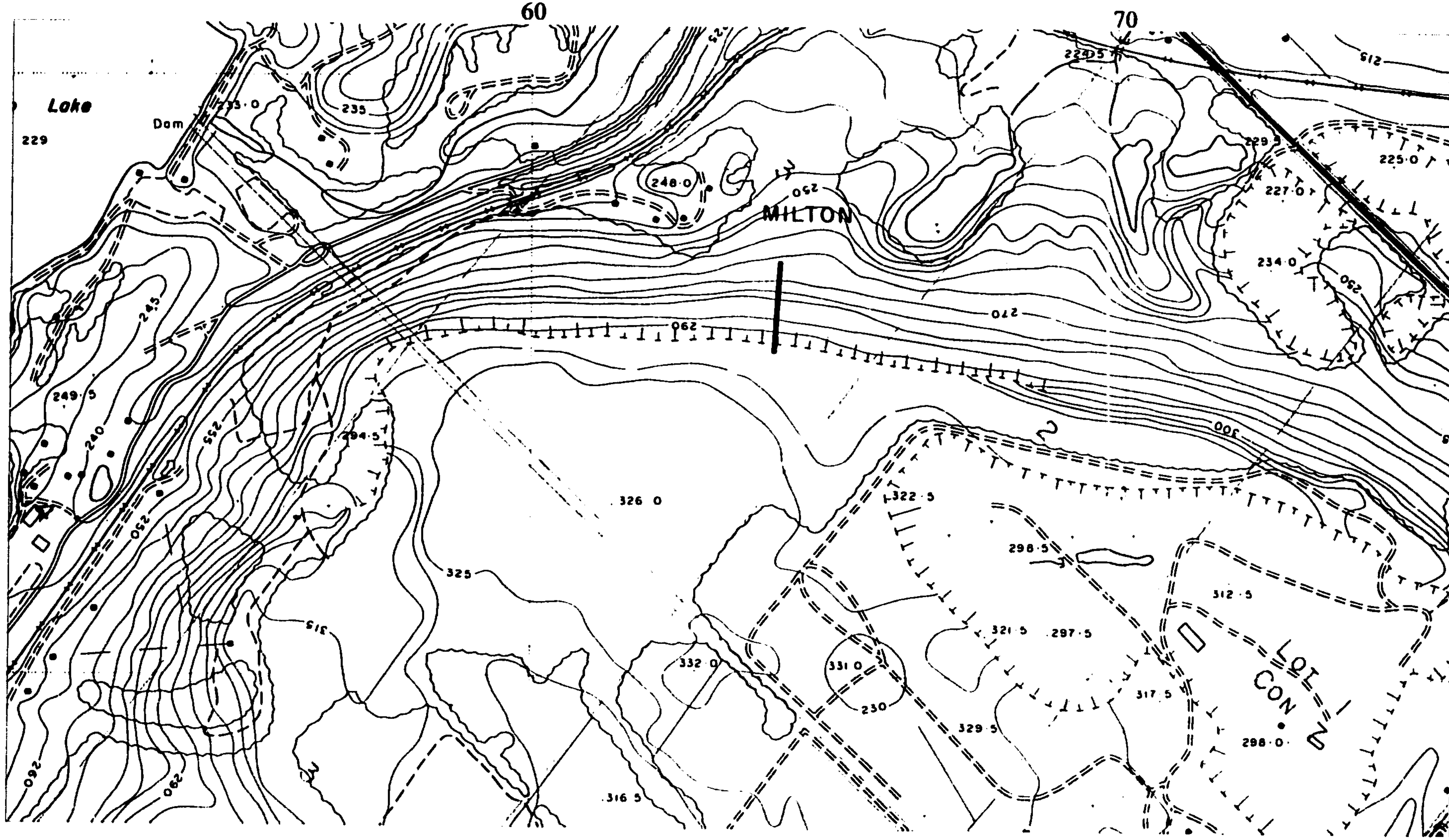
Table 4.1: Rheological data for Mt. Nemo Formations.
(MN=Mount Nemo, AD=Amabel Dolomite, RD=Reynales Dolomite, CS=Cabot Head Shale, MD=Manitoulin Dolomite, WS=Whirlpool Sandstone, QS=Queenston Shale, #'s=Sample Number)

Sample #	Yield Pt. (MPa)	Failure (MPa)	Dry Wt. (g/cm^3)	Saturated Wt. (g/cm^3)	% Strain @ Failure
MN-AD-01	-	76.871	2.700	2.7211	0.96
MN-AD-02	-	101.062	2.699	2.7192	1.11
MN-AD-03	-	151.072	2.642	2.6604	0.89
MN-AD-04	-	54.516	2.646	2.6675	1.11
Form. Ave.	-	**95.880** **+55.192** / **-41.364**	**2.672**	**2.6921**	**1.02**
MN-RD-02	-	163.568	2.764	2.7975	1.02
MN-RD-04	-	257.622	2.787	2.8101	1.51
MN-RD-05	-	312.856	2.787	2.8117	1.53
Form. Ave	-	**244.682** **+68.174** / **-81.114**	**2.779**	**2.8064**	**1.35**
MN-CS-01	0.212	0.3789	2.270	2.2737	5.77
MN-CS-02	0.164	0.2786	2.339	2.3422	6.53
MN-CS-03	0.163	0.2812	2.341	2.3452	6.87
Form. Ave.	**0.18**	**0.3129** **+0.066** / **-0.0343**	**2.317**	**2.3204**	**6.39**
MN-MD-01	-	97.284	2.688	2.7153	0.88
MN-MD-02	-	157.615	2.757	2.7899	0.96
MN-MD-04	-	140.775	2.676	2.7059	0.93
MN-MD-O5	-	124.438	2.785	2.8060	0.94
MN-MD-06	-	91.764	2.669	2.7012	0.98
Form. Ave.	-	**122.375** **+35.240** / **-30.611**	**2.715**	**2.7437**	**0.94**
MN-WS-01	-	96.195	2.343	2.4106	1.10
MN-WS-03	-	96.134	2.234	2.3933	1.08
MN-WS-04	-	83.401	2.243	2.3295	1.03
MN-WS-05	-	88.764	2.331	2.4055	0.79
MN-WS-06	-	80.419	2.344	2.4097	1.19
Form. Ave.	-	**88.983** **+7.212**	**2.299**	**2.3897**	**1.04**

			-8.564			
MN-QS-01	1.028	1.108		2.495	2.4984	0.62
MN-QS-02	0.674	0.734		2.401	2.4076	0.54
MN-QS-03	1.352	1.726		2.438	2.4439	1.42
Form. Ave.	**1.018**	**1.189**	**+0.537**	**2.445**	**2.4500**	**0.86**
			-0.455			

Figure 4.3 shows the calculated stress due to overburden within the Niagara Escarpment. While this does not include the effects of tectonics or hydrology on the stress environment, both of these would result in an increase in applied stress. The omission is therefore conservative and the effects of hydraulic stresses and tectonic stress would only serve to enhance any deformation resulting from overburden. The overburden plot also takes no account of topography such that the principal stresses indicated are all aligned with the Y-axis. In reality, the principal stresses would tend to align themselves with the surface of the slope. The magnitude of the principal stresses beneath the cap rock would however be the same as those indicated by overburden.

When compared to the stress due to overburden shown in figure 4.2.3, it can be seen that only the two shale formations have uniaxial compressive strength below that which is experienced within the Niagara Escarpment. The unconfined compression strength of the Cabot Head Shale was found to be between 0.2812 MPa and 0.3789 MPa, with an average yield point of 0.18 MPa. Beneath the cap rock, the Cabot Head formation experiences roughly 0.62 MPa at the top of the formation and 0.91 at its base. The unconfined compressive strength of the Queenston Shale was determined to be between 1.726 MPa and 0.734 MPa, with an average yield point of 1.018 MPa. From figure 4.2.3, the Queenston Formation has an overburden that is roughly 1.10 MPa at the top of the formation



Scale 1:10 000
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Legend

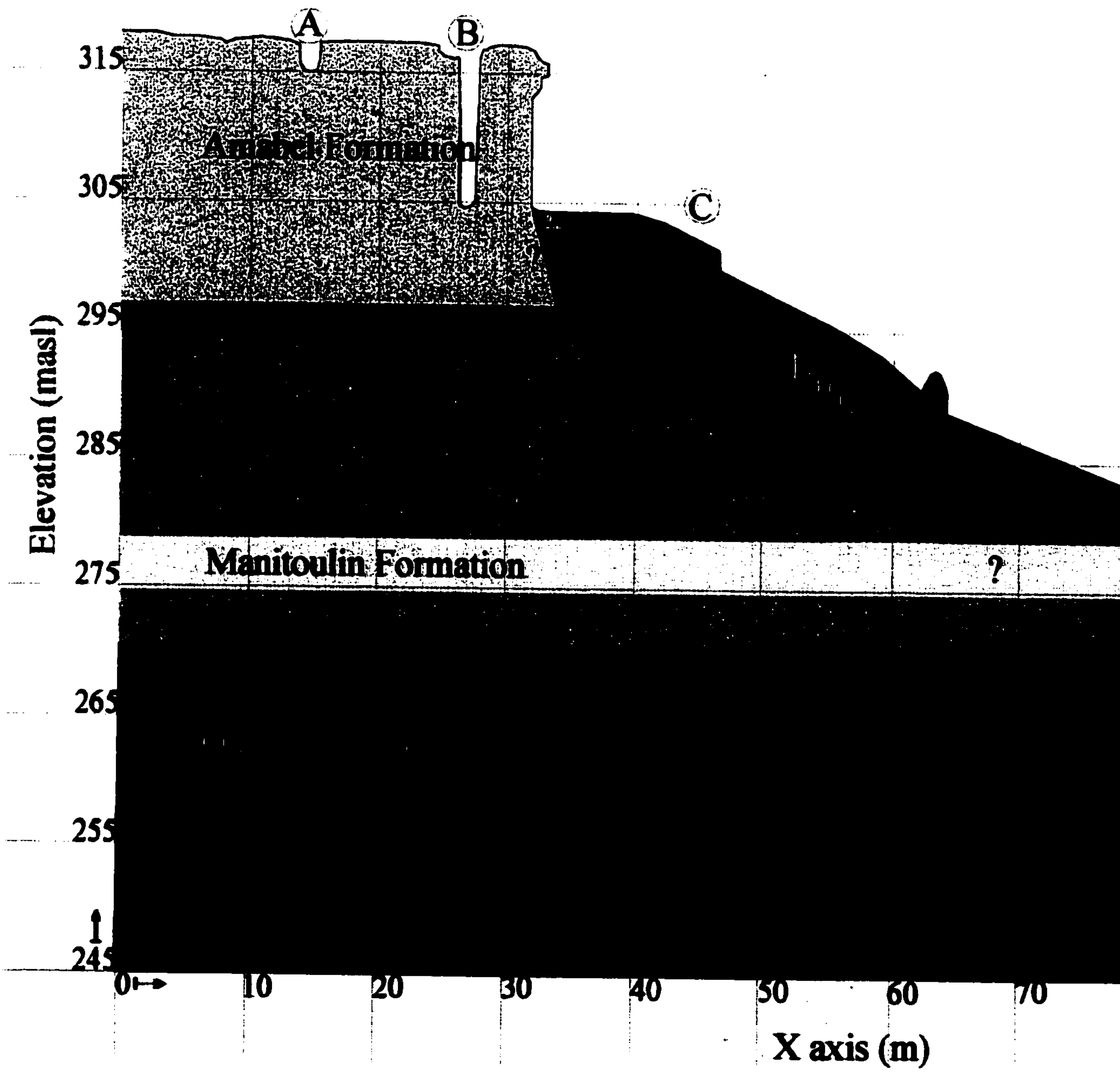
Location of Cross Section

Figure 4.4: Milton Heights Study Area

Detail from O.G.S. Map 10 17 5850 48150

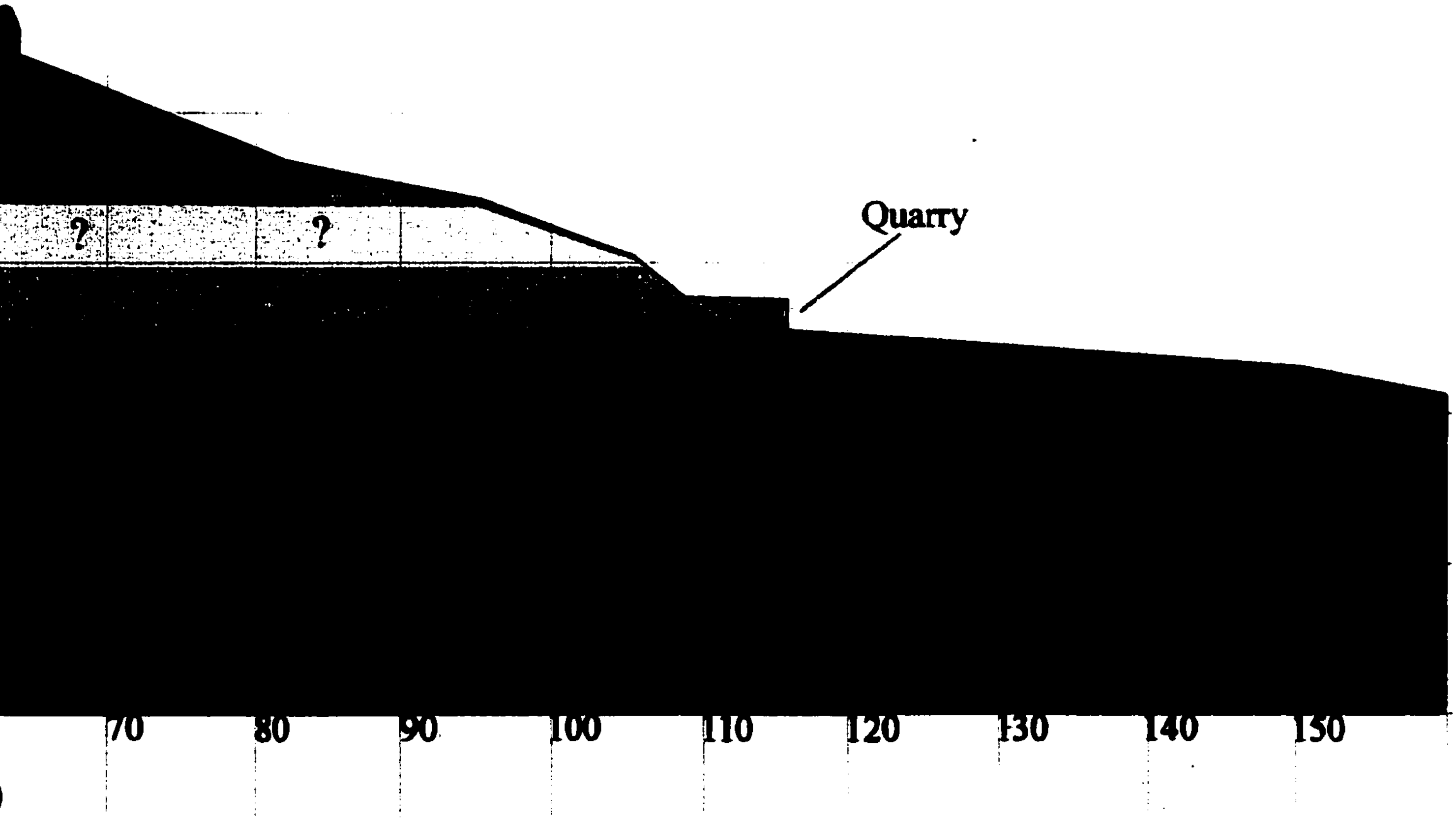
Notes:
North American Datum 1927
Universal Transverse Mercator (6o) projection
Zone 17. Central Meridian 81oW
Grid Interval 1000 m

Figure 4.5: Geological Cross section of M



section of Milton Heights

Scale: 1:500



study area, with deep fractures isolating blocks from the main rock mass (Figures 4.5 - A. & B.). The talus slope, while similar in appearance to Mt. Nemo (figure 4.5C.), is longer in the X-axis, and extends out from the cliff face to the top of a secondary scarp composed of the Manitoulin and Whirlpool Formations. The edge of this secondary scarp has been modified in many areas by now abandoned quarries.

Rock samples were taken from the nearest outcropping and subjected to uniaxial compression tests. The shale formations were sampled at the same location as for Mt. Nemo and the results obtained were used for both. Tests on the brittle rocks revealed the following characteristics:

Table 4.2: Rheological Data for Milton Heights Formations.
(Mi=Milton Heights, AD=Amabel Dolomite, RD=Reynales Dolomite, MD=Manitoulin Dolomite, WS=Whirlpool Sandstone, #'s=Sample Number)

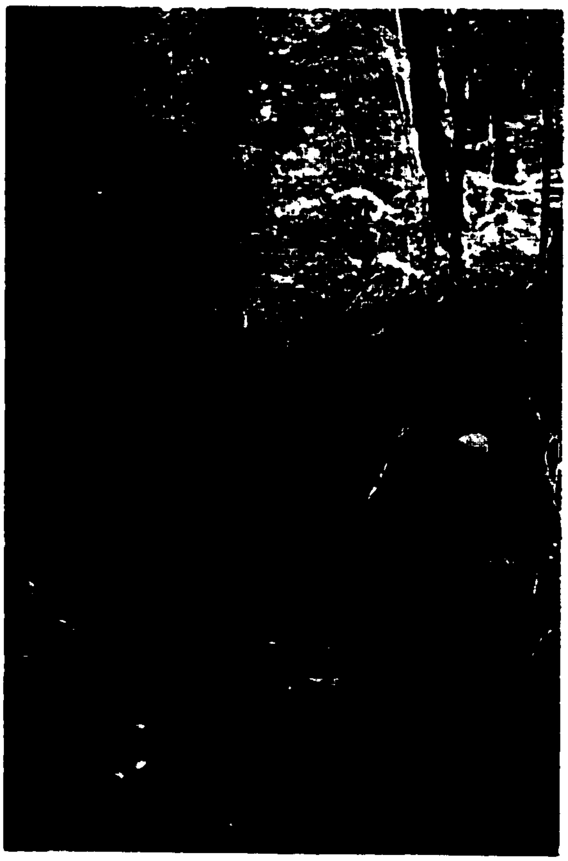
Sample #	Yield Pt. (Mpa)	Failure (Mpa)		Dry Wt. (g/cm^3)	Saturated Wt. (g/cm^3)	% Strain @ Failure
MI-AD-02	-	133.056		2.6642	2.6920	0.84
MI-AD-03	-	97.137		2.6236	2.6580	0.69
MI-AD-04	-	64.001		2.5898	2.6286	0.60
MI-AD-05	-	124.802		2.6042	2.6399	0.66
MI-AD-07	-	69.766		2.6126	2.6521	0.83
Form. Ave.	-	**99.75**	**+35.304** **-33.751**	**2.6189**	**2.6541**	**0.72**
MI-RD-01	-	147.175		2.7523	2.7785	0.94
MI-RD-06	-	119.760		2.7341	2.7605	0.96
MI-RD-07	-	103.40		2.7918	2.8137	0.78
MI-RD-08	-	82.559		2.7829	2.8060	0.87
MI-RD-09	-	142.667		2.7680	2.7902	0.94
MI-RD-10	-	91.585		2.7859	2.8084	0.83
Form. Ave.	-	**114.52**	**+32.651** **-31.965**	**2.7692**	**2.7929**	**0.89**
MI-MD-04	-	112.532		2.3947	2.4561	1.01
MI-MD-05	-	89.356		2.3719	2.4342	0.91
MI-MD-06	-	80.150		2.3931	2.4502	0.82
MI-MD-07	-	94.807		2.3929	2.4499	1.06
MI-MD-08	-	87.139		2.3846	2.4460	0.95

Figure 4.5C: Talus Slope, Milton Heights

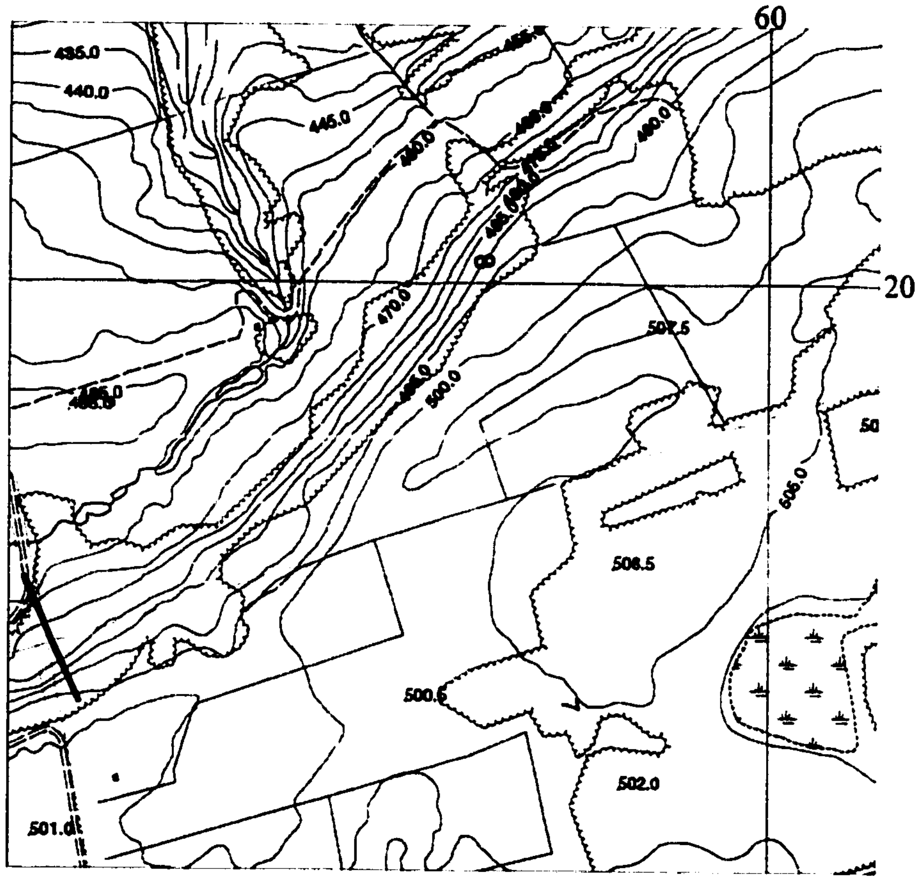
4.4 Osler Bluff Study Site

The Osler Bluff Site is located on the northern flank of Osler Bluff, near Collingwood. The cross section runs from the top of the cap rock down and across the access road running between Gibraltar Sideroad and County Road 19 (Fig 4.7). The morphology of Osler Bluff differs significantly from the previous two sites. While the cap rock is incised by the same type of crevices as are found at the southern sites (Fig 4.8 – A & B), they are much more developed. As can be seen in figure 4.8, the crevices isolate two major blocks from the rest of the rock mass. These have both been "jostled" to some extent and the bedding planes show a marked divergence from the regional dip. Repeat surveys of some of the blocks at this site have illustrated that they are in motion relative to one another at a rate of up to 1 cm/yr (Hintz, 1997). This trend becomes more pronounced the closer it gets to the cliff face and is indicative of the cambering phenomenon (Hewitt, 1997).

Figure 4.8A: Major Crevice Behind Cliff Face, Osler Bluff.



The Talus slope is much longer in the X-axis than that of the two southern sections and has a much gentler gradient. It is made up of many large blocks, some of which seem be remnants of a previous cliff face and have lost their uppermost beds due to toppling failure (figure 4.8C). As with the Milton Height's site, the talus slope overlies a secondary scarp of Manitoulin Dolomite



Scale: 1:10000
Contour Interval 5 metres

Figure 4.7: Osler Bluff Study Area

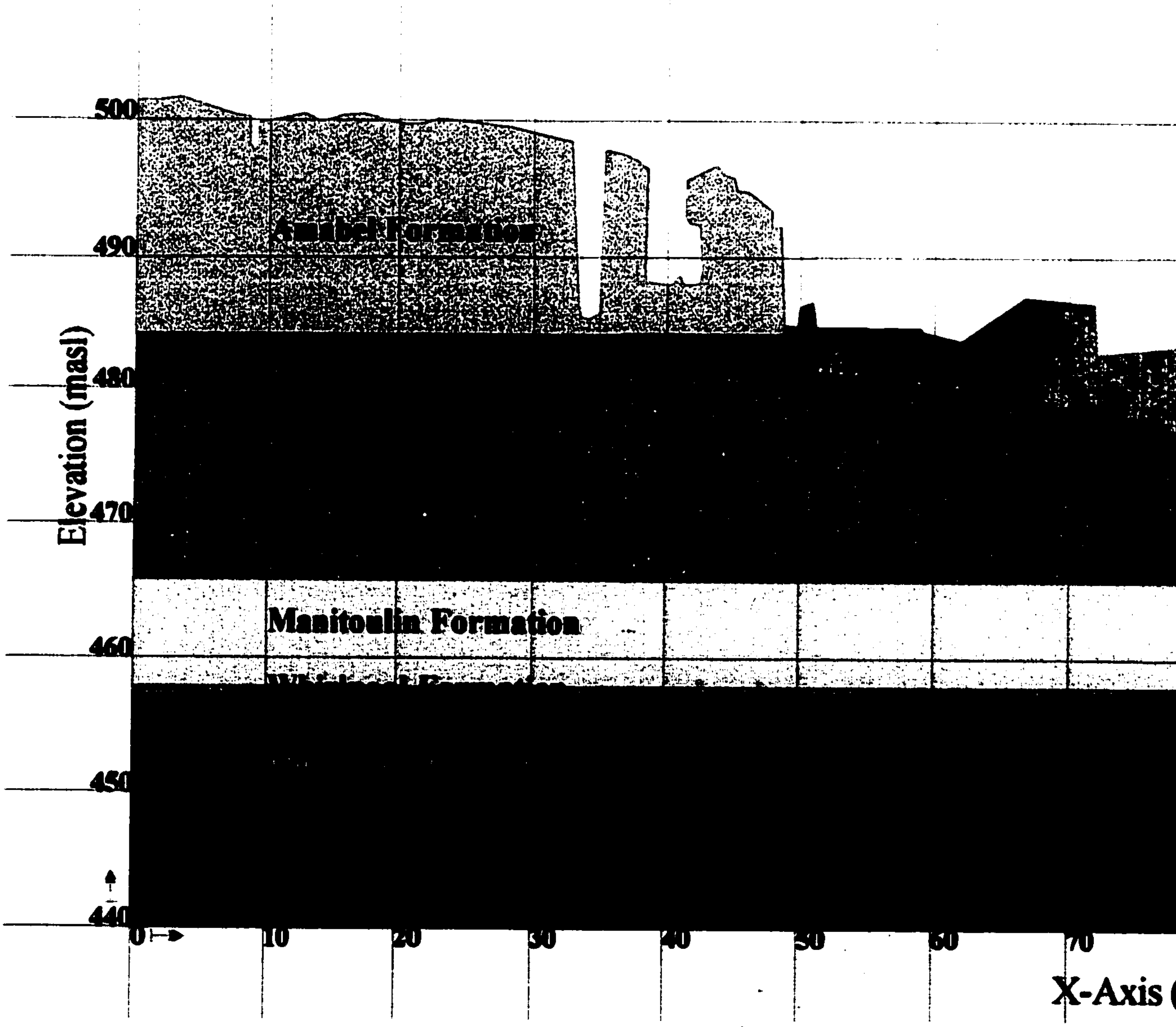
Legend

Location of Cross Section

Detail from O.G.S. Map 10 17 5550 49200

Notes:
North American Datum 1927
Universal Transverse Mercator (6°) Projection
Zone 17. Central Meridian 81°W
Grid Interval 1000 m

Figure 4.8: Geological Cross Section



ss Section of Osler Bluff

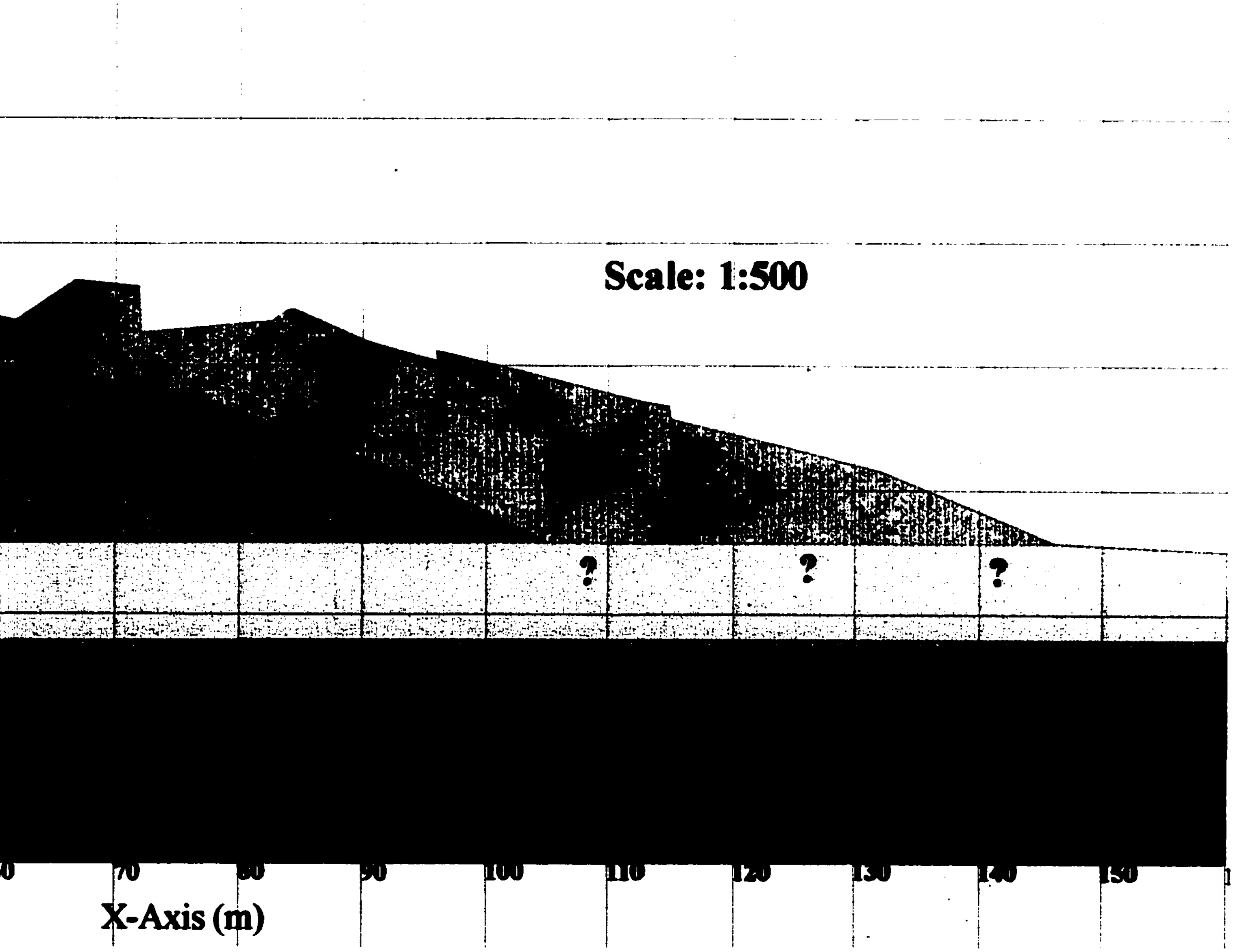


Figure 4.9: Calculated Stress due to Overburden for Osler Bluff (Mpa)

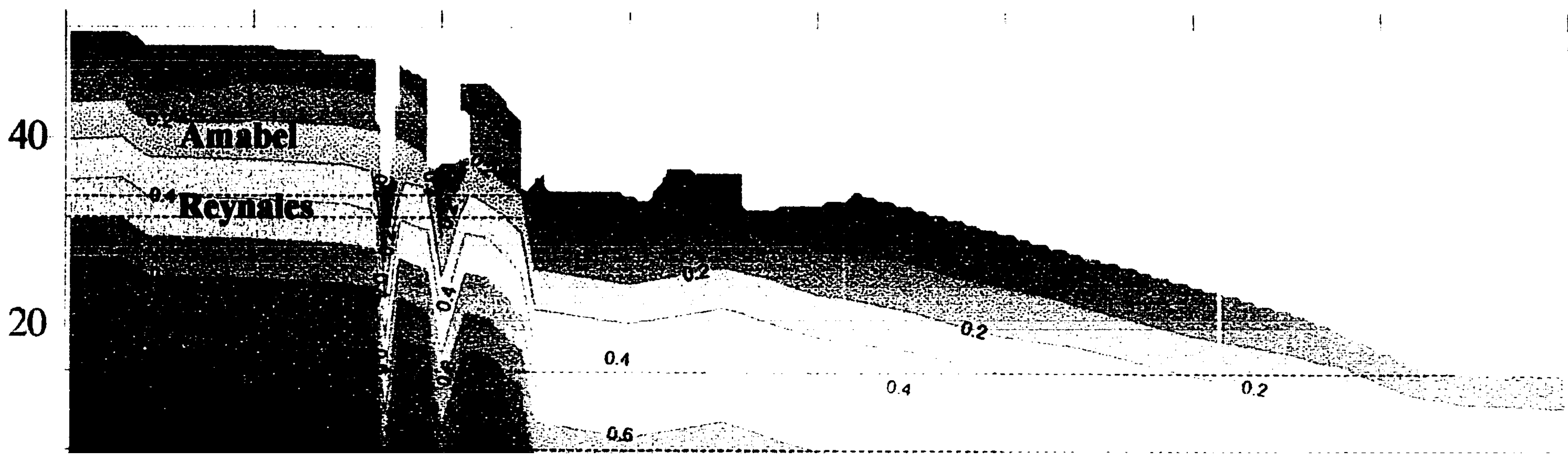


Figure 4.8B: Crevices Isolating Blocks near the Cliff Face, Osler Bluff.

Figure 4.8C: Isolated Blocks of Amabel Dolostone at the top of the Talus Slope, Osler Bluff.

that Projects out several Hundred Metres beyond the main cliff face. A basal slope of Queenston Shale slopes down from the secondary scarp to the valley floor below and represents the largest portion of the total relief from the cliff top to the vale. This bench below the main scarp has poorly drained areas and is incised by several small streams.

Uniaxial compression tests performed on the various formations of the escarpment resulted in the following:

Table 4.3: Rheological Data for Osler Bluff.
(OS=Osler Bluff, AD=Amabel Dolomite, FD=Fossil Hill Dolomite, CS=Cabot Head Shale, MD=Manitoulin Dolomite, WS=Whirlpool Sandstone, QS=Queenston Shale, #'s=Sample Number)

Sample #	Yield Pt. (MPa)	Failure (MPa)		Dry Wt. (g/cm^3)	Saturated Wt. (g/cm^3)	% Strain @ Failure
OS-AD-02	-	66.264		2.5018	2.5521	0.58
OS-AD-03	-	71.682		2.4544	2.5165	0.56
OS-AD-04	-	93.083		2.5346	2.5806	0.69
OS-AD-05	-	54.898		2.5059	2.5608	0.66
OS-AD-06	-	69.936		2.4781	2.5349	0.65
Form. Ave.	-	**71.173**	**+21.910 -16.215**	**2.4950**	**2.5490**	**0.63**
OS-FD-01	-	142.010		2.6558	2.6797	0.80
OS-FD-02	-	107.762		2.6652	2.6875	0.66
OS-FD-03	-	160.837		2.6012	2.6286	0.93
Form. Ave.	-	**136.870**	**+23.267 -29.108**	**2.6407**	**2.6653**	**0.80**
OS-CS-01	0.182	0.2869		2.340	2.3438	6.40
OS-CS-03	0.179	0.2743		2.339	2.3431	6.72
OS-CS-04	0.175	0.2850		2.342	2.3442	6.21
Form. Ave.	**0.179**	**0.2821**	**+.0048 +.0078**	**2.340**	**2.3437**	**6.44**
OS-MD-02	-	62.087		2.6879	2.7110	0.71
OS-MD-03	-	102.582		2.7531	2.7688	0.83
OS-MD-04	-	173.925		2.7648	2.7789	0.79
Form. Ave.	-	**112.865**	**+61.06 -40.495**	**2.7353**	**2.7529**	**0.78**
OS-WS-02	-	55.559		2.1377	2.2843	0.76
OS-WS-03	-	52.962		2.1179	2.2680	0.72
OS-WS-04	-	81.271		2.1481	2.2896	0.83
OS-WS-05	-	66.057		2.1372	2.2804	0.77
Form. Ave.	-	**63.962**	**+17.309**	**2.1352**	**2.2806**	**0.77**

			-11.00			
OS-QS-01	0.582	0.738		2.3598	2.3632	1.80
OS-QS-02	1.016	1.102		2.3602	2.2641	1.27
OS-QS-03	1.112	1.136		2.3617	2.3644	0.77
Form. Ave.	**0.903**	**0.992**	**+0.144** **-0.254**	**2.3606**	**2.3306**	**1.28**

Once again, when compared to the calculated overburden in figure 4.9, only the shale formations show any potential for deformation. Beneath the cap rock, the Cabot Head Formation experiences between 0.42 MPa and 0.90 MPa due to overburden. This is several times higher than the formational yield point, which is 0.179. The Queenston Formation has a minimum overburden of 1.04 MPa beneath the cap rock. The yield point for the formation was found to be 0.903 MPa.

4.4 Discussion

It is evident that the lithology is similar at each of the study sites and they share many of the same landforms (Table 4). The major differences between the

Table 4: Summary of Morphological Features for the Three Sites:

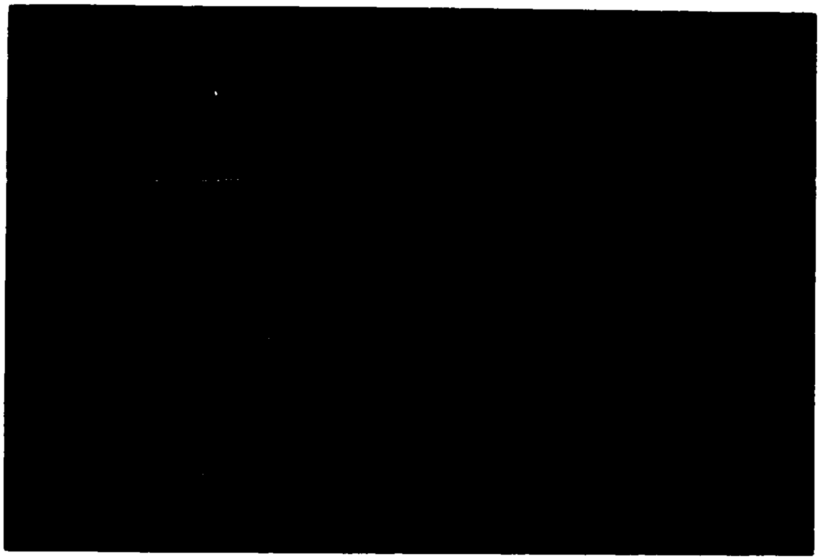
Site	Mt. Nemo	Milton Heights	Osler Bluff
Crevices	Yes	Yes	Yes
Convex Talus Slope	Yes	Yes	Yes
"Bench" Stucture	No	No	Yes
Secondary Scarp	No	Yes	Yes
Steep Basal Slope	No	No	Yes

field sites occur within the basal slope zones. The morphology of the cliff face is basically identical for all of the sites, although the crevices are more developed at Osler Bluff. A secondary scarp that extends from the base of the talus slope is present at both Milton Heights and Osler but not at Mt. Nemo. Furthermore, at the Osler Bluff site, the basal slope is much steeper than at the two southern sites. Although this is not shown on the 1:500 scale cross section, the basal slope descends at an angle of roughly 30 degrees. The upper portion of this slope is in the Queenston formation, while the lower portion is composed of the Georgian Bay Shale. The reasons for these similarities and differences are discussed below.

4.5.1 Rock Rheology

The stress Vs strain curves for all of the samples are given in appendix 1 - 3. They show that the rheology of the rock formations is highly variable. This is particularly true for the dolomite formations, which exhibited large variations. They range up to 149.288 MPa in the case of the Reynales Formation at Mt. Nemo. This is probably due to the influence of bioherms and other fossils, found throughout the Silurian carbonates of the Clinton and Cataract Groups (Bolten, 1957). These have the effect of altering both rock fabric and pore space, which play a key role in the unconfined compressive strength of rocks (Selby, 1993). The Whirlpool Formation shows much less variation in compressive strength than do the carbonate rocks, probably due to the absence of these structures. However, the Whirlpool Formation sampled near the Osler Bluff site did show some variation in its unconfined compressive strength. The exact reasons for this cannot be determined without microscopic and mineralogical investigation but it is probably due to fluctuations in the amount of calcite in the samples.

Figure 4.10: Brittle Fracture Pattern, Amabel Formation.

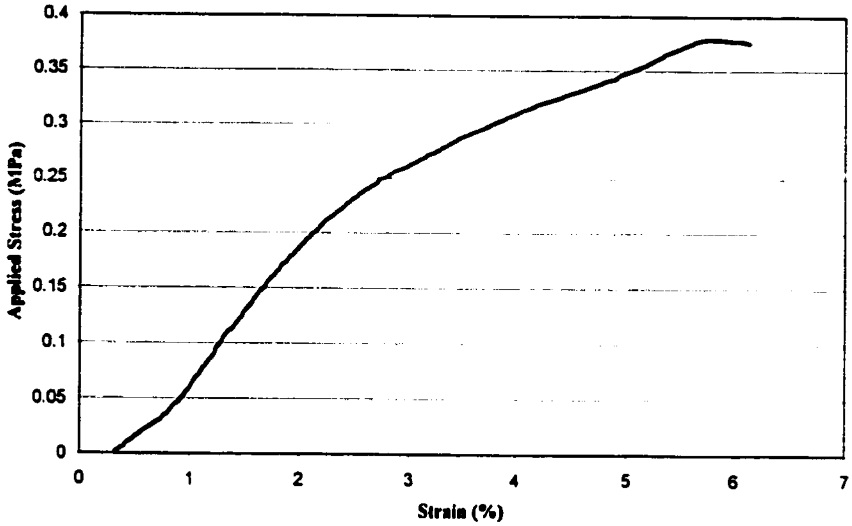


None of the dolomites, nor the Whirlpool formation, displayed an unconfined compressive strength below that which can be attributed to overburden pressures within the escarpment. Furthermore, all displayed a brittle fracture pattern upon failure (figure 4.10). Meanwhile, if deformation is occurring within the shale layers, these rocks will be in motion relative to one another and may experience shear and/or tensile stresses that do exceed their failure thresholds. This seems to be the most likely explanation for the propagation and expansion of the crevices near the cliff face.

The shale formations of the escarpment exhibit a very different type of behaviour. A stress Vs Strain curve for the Cabot Head Shale is shown in figure 4.11. The Cabot Head Shale possesses extremely low strength characteristics and a distinct yield point between 0.163 MPa and 0.212 MPa. This is several times lower that the stress due to overburden experienced at each of the study sites.

Furthermore, when under stress, the Cabot Head Shale showed a high degree of ductility with an average of 6.39 % strain at failure at the southern sites and 6.44

Figure 4.11: Uniaxial Compression Test: Cabot Head Shale

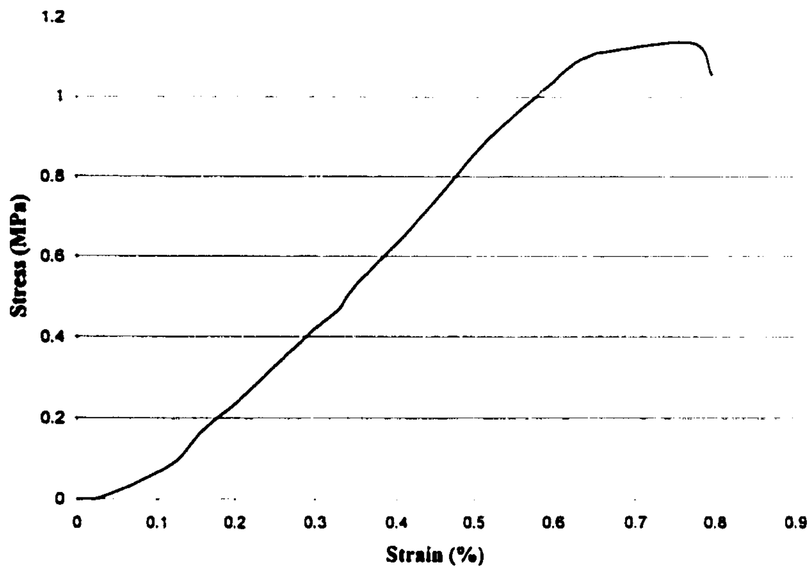


% strain at failure for the Osler Bluff site. A core of Cabot Head Shale after compression is illustrated in figure 4.12. As is evident from this picture, the Cabot Head Formation has not shattered but expanded laterally in response to the applied stress.

Figure 4.12: Cabot Head Shale after Unconfined Compression Test.

The Queenston Formation is somewhat stronger than the Cabot Head Formation (Fig 4.13) having an average yield point of 1.018 MPa for the southern

Figure 4.13: Uniaxial Compression Test: Queenston Shale



sites and 0.903 for the Osler Bluff site. This is just below the minimum overburden stresses experienced within the escarpment, which are roughly 1.10 MPa for the two southern sites and 1.04 for Osler Bluff. However, the Queenston formation behaves in a brittle manner when unconfined, although some plastic deformation does occur between the yield point and failure. There is also some variation in the strength properties resulting from the individual tests (Appendix 1,3). These can most likely be attributed to variations in the calcite content within the formation. In general, the calcite and dolomite content of the Queenston formation increases from the bottom of the formation to the top. This suggests leaching of calcite and dolomite from the overlying carbonates (Guillet, 1967).

4.5.2: Morphology

From the geological cross sections it can be seen that the Niagara Escarpment most closely resembles the horizontal bedding in figure 2.12. The major discontinuities are therefore either parallel or perpendicular to the principal stress resulting in a very stable landform. Mass movements within the escarpment can therefore not be attributed to a shear plane as is common to many rock slope failures (Selby, 1993). The creation of the crevice caves and the present day motion of the blocks along the escarpment face is therefore probably due to deformation within the Cabot Head Formation. The potential stresses within the escarpment are more than adequate to initiate failure within this formation. Furthermore, the geometry of the slope allows for less confining stress within the Cabot Head Formation. This is basically due to the gradient of the upper slope zone which results in the portion of the Cabot Head Shale directly under the cap rock being closer to a free face. The stiffening effect that confining stress has on rock would require a two to three fold increase in compressive strength before the Cabot Head Shale would become stable. The Queenston Formation however is constrained to a much higher degree laterally and possesses a much higher intrinsic strength. Deformation within this formation is therefore probably minimal.

The Niagara Escarpment can therefore be envisioned as a series of two escarpments, each dependent on the rheological properties of the underlying shale formations. The main escarpment is undergoing a recurrent toppling failure due to deformation within the Cabot Head Shale. This toppling has two primary mechanisms. This first involves a block gradually leaning out due to subsidence in the Cabot head shale along the scarp face, in a manner similar to the yielding flow process described by Pei & Tianchi. The second involves blocks that rotate out at the base due to intrusion of shale into the parting between the block and the main rock mass. This results in the "true" crevice caves that are closed at the top.

As it retreats over time it reveals the underlying secondary scarp made up of the Manitoulin and Whirlpool Formations which is supported by the more stable Queenston Formation. In areas where the rate of retreat has been more rapid, a "bench" has been created by the Manitoulin/Whirlpool Formations, which are only partially obscured by the talus slope. In areas of slower retreat, this "bench" is absent due to burial beneath the talus slope, however, the secondary scarp may still be present. This "bench" is clearly responsible for the projection of the Queenston Shale out from the main scarp at Blue Mountain and Osler Bluff as it protects the shale from weathering. In both instances, this projection is capped by the secondary scarp, and the Queenston Formation slopes steeply down from the base of this scarp. This is best evidenced at the Blue Mountain Lookout, where the Manitoulin Formation caps the Queenston Shale, which slopes steeply down to the vale below.

The relative rate of retreat is probably dependent upon the stratigraphy of the Cabot Head Shale. To the south, the upper part of this formation overlaps with the Grimsby Formation and therefore possesses a higher degree of sandstone and limestone interbeds (Bolton, 1957). These would have the effect of reducing the total amount of deformable clay shale within the formation as a whole. To the north, the Cabot Head Formation does not possess such a high proportion of nondeformable beds and therefore would deform more rapidly when under stress. Another control on the rate at which the blocks topple is the buttressing effect of the talus slope. As is evident from the cross sections, the base of the cap rock is buried beneath several meters of talus. This would act to inhibit the amount of lateral motion that the blocks will undergo and has been noted as a controlling factor in other areas (Cain, 1982).

Chapter Five: Conclusions

5.1 Conclusions:

1. From a consideration of the compressive strength data on the rock formations of the Niagara Escarpment, it is probable that the Cabot Head shale is undergoing plastic deformation due to the weight of the overlying rocks. This deformation is essential in explaining the morphological features that are ubiquitous to the steep-sloped landforms of the Niagara Escarpment between Mt. Nemo and Blue Mountain.
2. On these grounds, too, the hypothesis of Hewitt (1997) is essentially correct, except that these data identify the deformation associated with crevice cave formation and cambering as occurring within the Cabot Head formation rather than the Queenston Formation. The crevice caves and cambering described by Hewitt (1997) are an expression of the rheological response of the Cabot Head shale to its stress environment. The reason why the crevices are more developed in the northern areas is probably because the total thickness of clay shale within the Cabot Head Formation increases in a northerly direction.
3. The morphology of the secondary scarp is variable but generally more prominent in northern areas. This acts to armour the underlying Queenston Formation, which is highly susceptible to weathering and erosion (Barlow, 1995), and creates a "bench" at the base of the main scarp. This bench is clearly visible at Osler Bluff and Blue Mountain above the ski slopes. The

lower aquifer described by Nadon & Gale (1984) flows through this area creating a saturated zone that is common to major breaks in slope (Selby, 1993). Furthermore, this water has been buffered by the carbonates in the cap rock and is typically basic (Barlow, 1995) and therefore has limited weathering capacities.

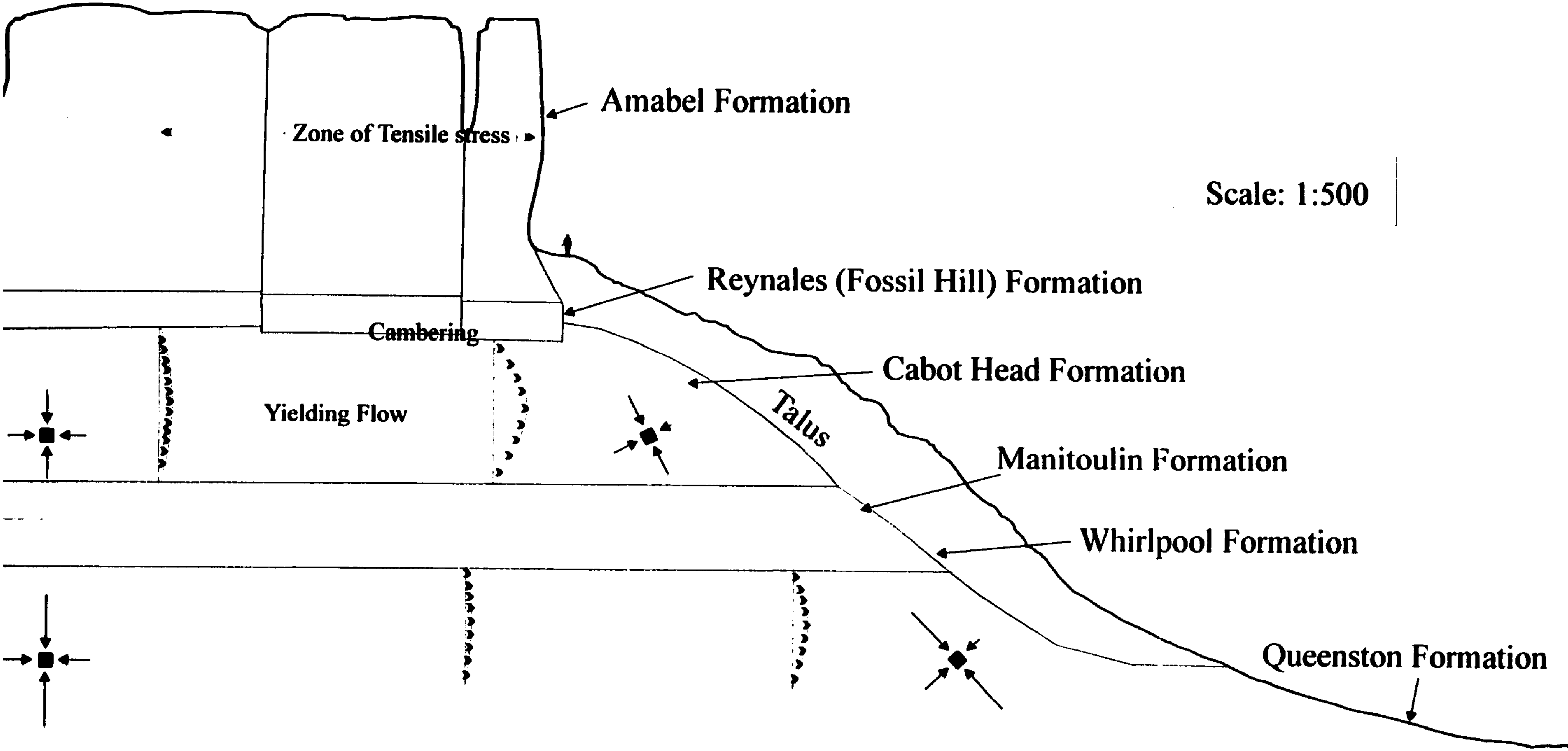
4. The main scarp and the secondary scarp are two separate landforms retreating at differing rates due to the rheological properties of their shale layers. The rate of retreat is probably higher for the main scarp due to the more deformable Cabot Head Shale. The actual rate of retreat would most likely slow over time due to the buttressing effect of the talus slope.
5. Based on the first three conclusions, a conceptual model of scarp development can be created. This is shown in figure 5.1.1. This is basically the yielding flow landslide envisioned by Pei & Tianchi applied to a more complex lithology.

5.2 Future Research:

The preceding study points to the following areas for future research in order to validate and expand the conclusions listed above:

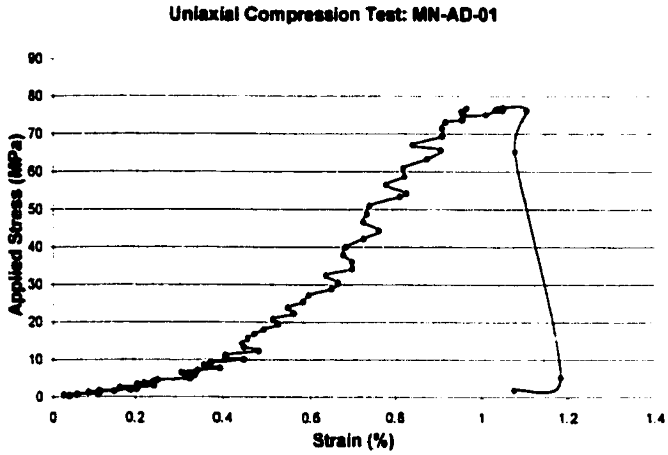
1. The development of more accurate geological cross sections through the drilling of boreholes: This will allow verification of the total thickness of clay shale within the Cabot Head Formation, and will establish the true dip of the formations due to cambering. It would also facilitate hydrological testing and measurement of tectonic forces.

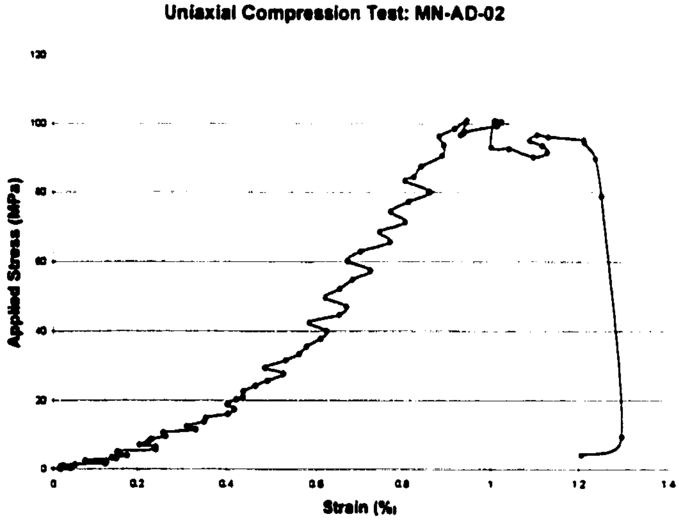
***Figure 5.1:** Conceptual Model: Endogenetic Processes Within the Niagara Escarpment*

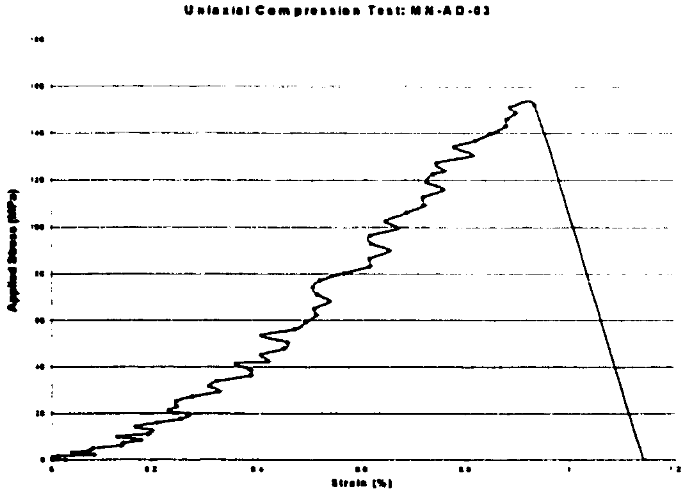


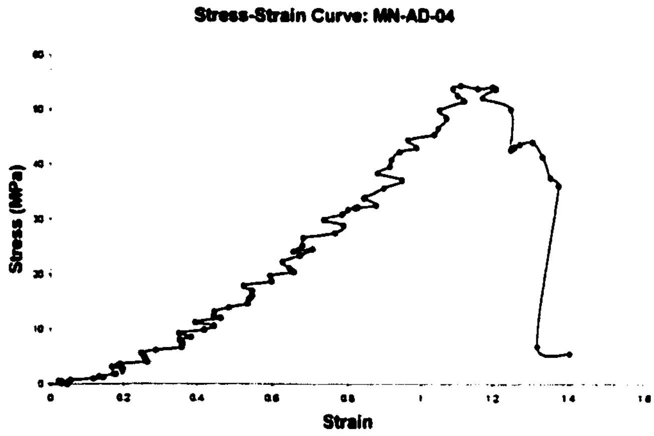
2. More extensive rheological studies of the formations involved: Rheological properties such as Young's modulus and Poisson's ratio would allow exact determination of lateral stress within the escarpment. These tests require lateral strain gauges to be affixed to the sample during compression. In addition, triaxial compression tests should be conducted to ascertain the affect that increased confining stress has on the strength properties of the shale formations. Samples taken from the boreholes (above) instead of at the surface would result in superior geotechnical data.
3. Mathematical modelling of the stress field within the escarpment: A finite element analysis of the stress field within the escarpment would yield the true orientation of the principal and secondary stresses and could delimit the actual zones in which deformation is to occur.
4. Hydrological Studies: The role of groundwater should be quantified through a detailed analysis of the flow pathways that the water takes through the escarpment. This can affect both the stress environment and the rheological properties of the rocks. This could then be integrated into the finite element analysis to get an accurate representation of the stress field. The geochemical effects of groundwater on the rocks at depth could also be ascertained through the use of boreholes.

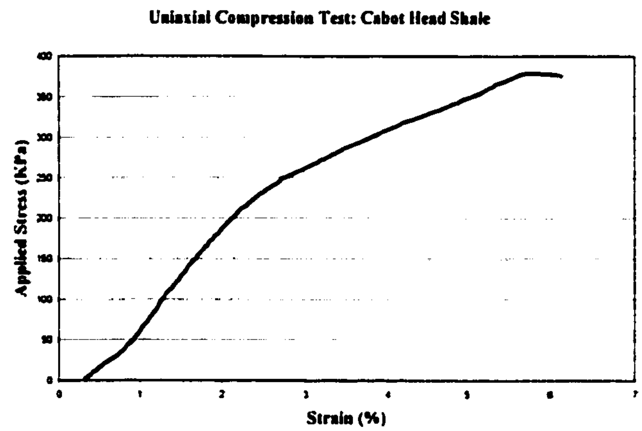
Appendix 1: Stress vs Strain Curves for Mt. Nemo

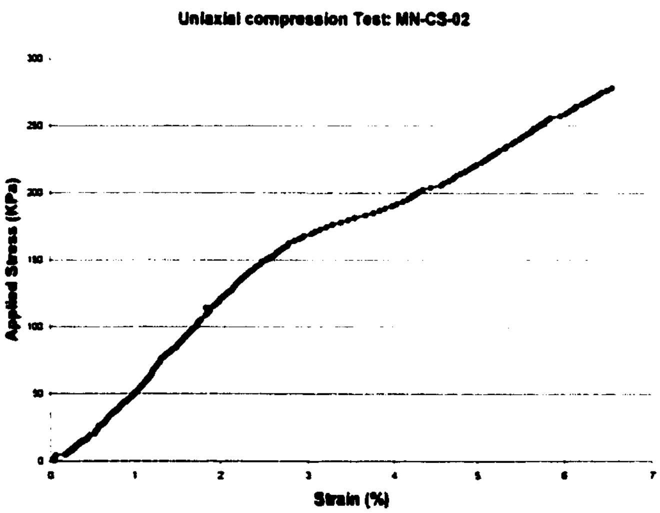




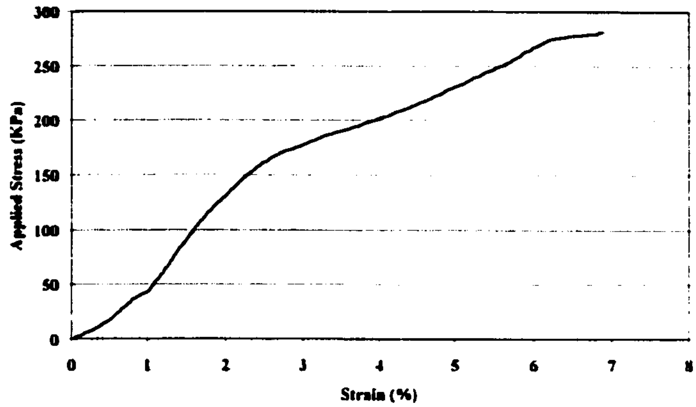



Stress-Strain Curve: MN-AD-04
Stress (MPa)
Strain

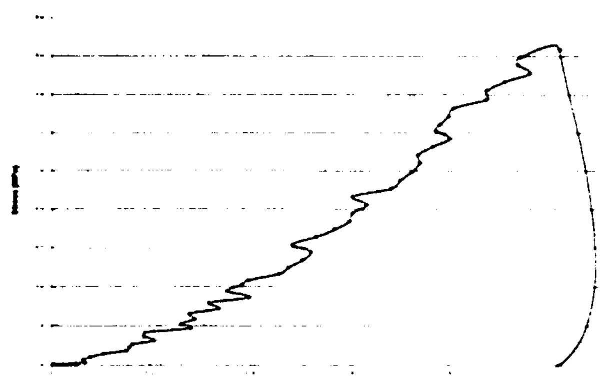

Uniaxial Compression Test: Cabot Head Shale
Applied Stress (KPa)
Strain (%)


Uniaxial compression Test: MN-CS-02
Applied Stress (KPa)
Strain (%)

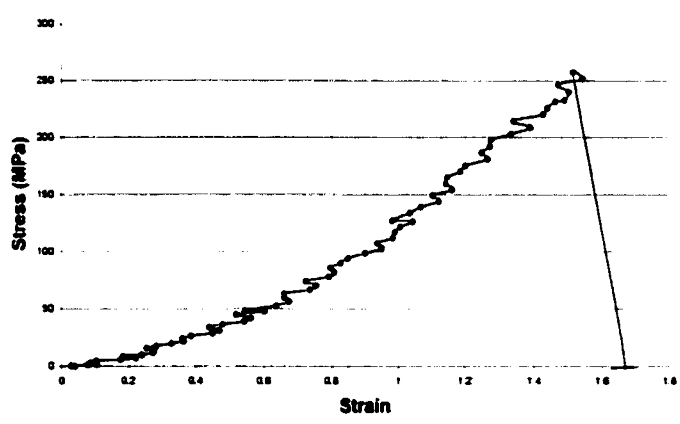
Uniaxial compression Test: MN-CS-03

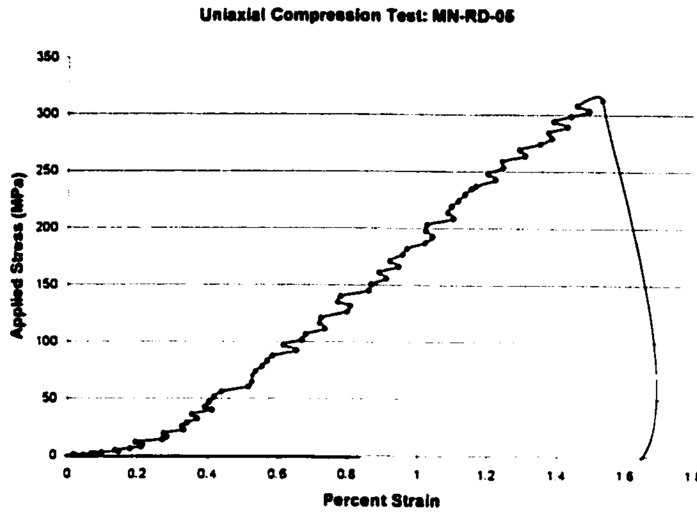


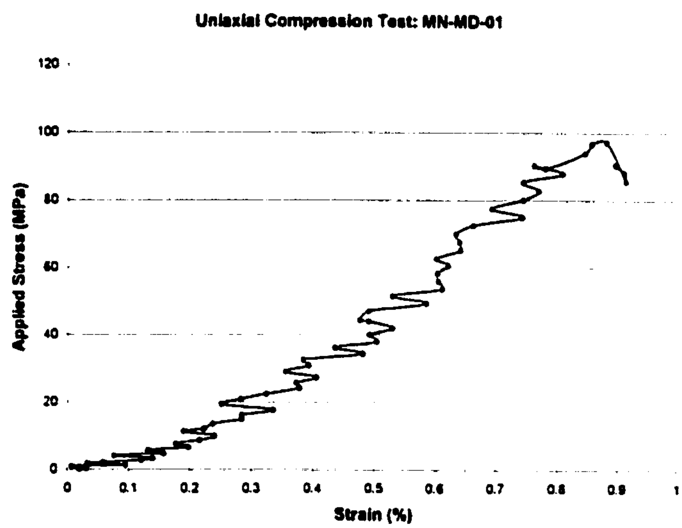
Stress-Strain Curve: MN-RD-02

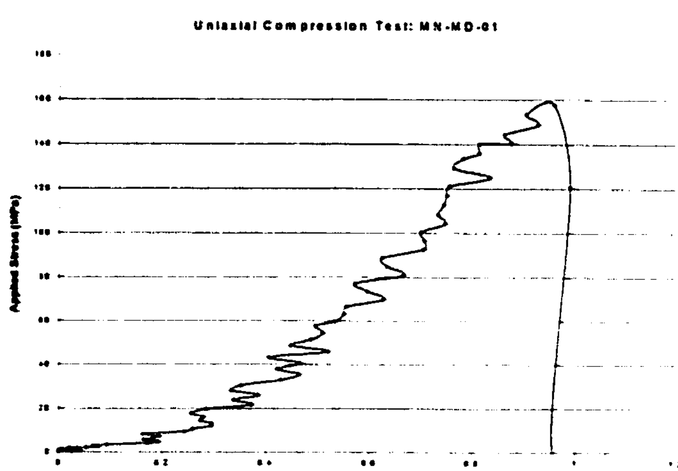


Stress-Strain Curve: MN-RD-04

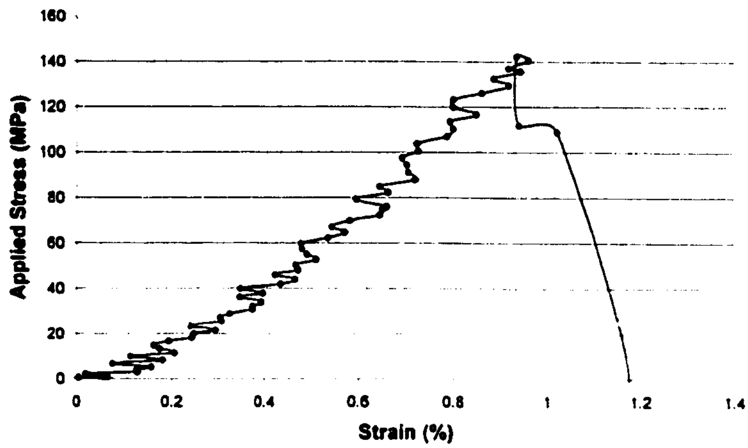



Uniaxial Compression Test: MN-RD-06
Applied Stress (MPa)
350
300
250
200
150
100
50
0
0
0.2
0.4
0.6
0.8
1
1.2
1.4
1.6
1.8
Percent Strain

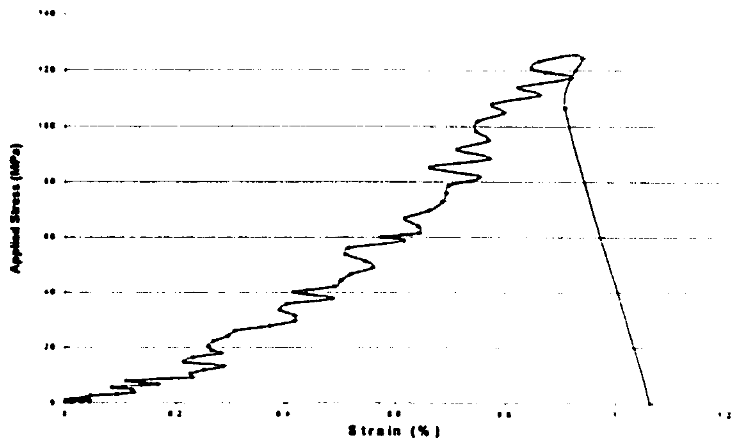

Uniaxial Compression Test: MN-MD-01
Applied Stress (MPa)
120
100
80
60
40
20
0
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
1
Strain (%)


Uniaxial Compression Test: MN-MD-01

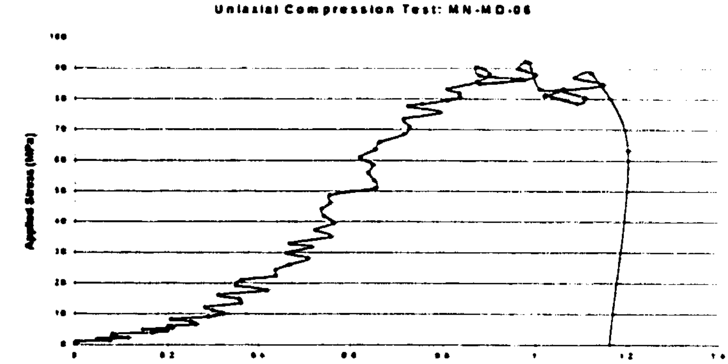
Uniaxial Compression Test: MN-MD-04



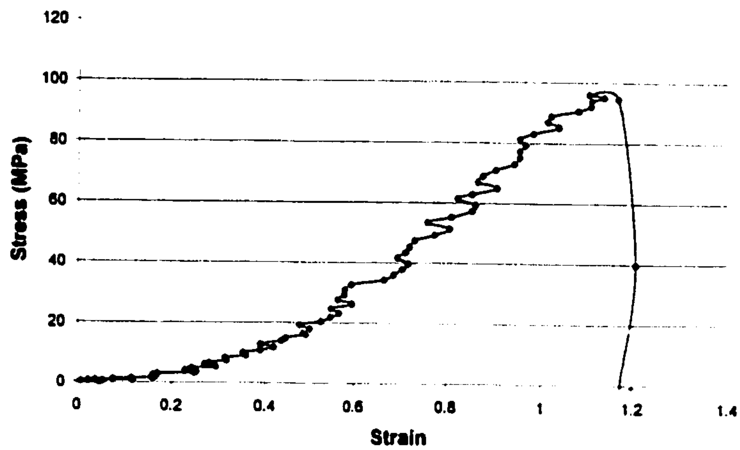
Uniaxial Compression Test: MN-MD-05



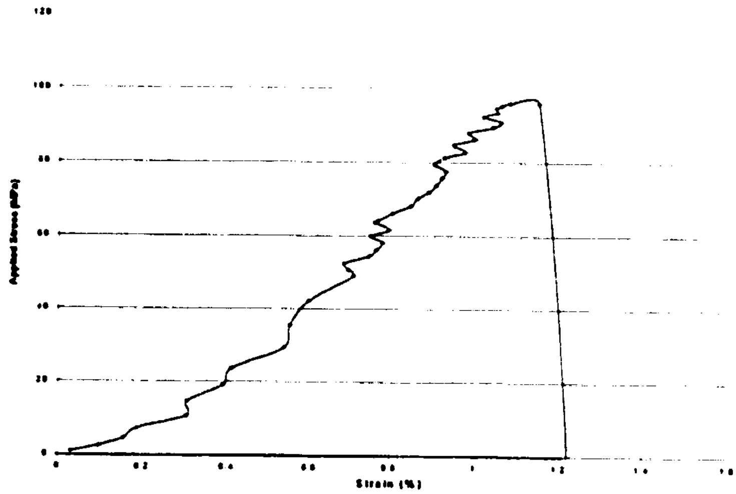
Uniaxial Compression Test: MN-MD-06



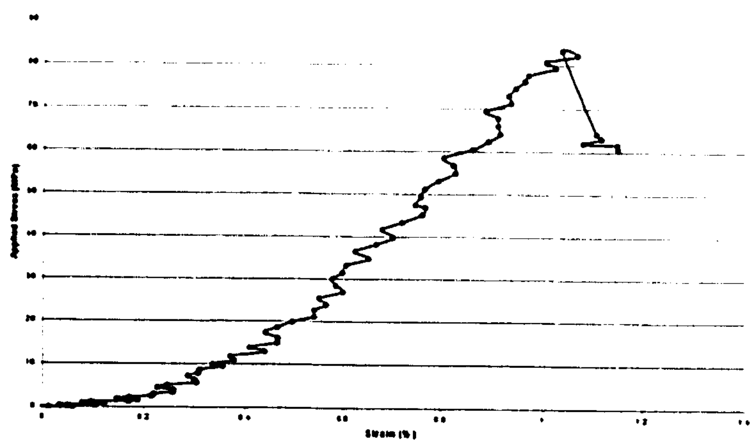
Stress-Strain Curve: MN-WS-01



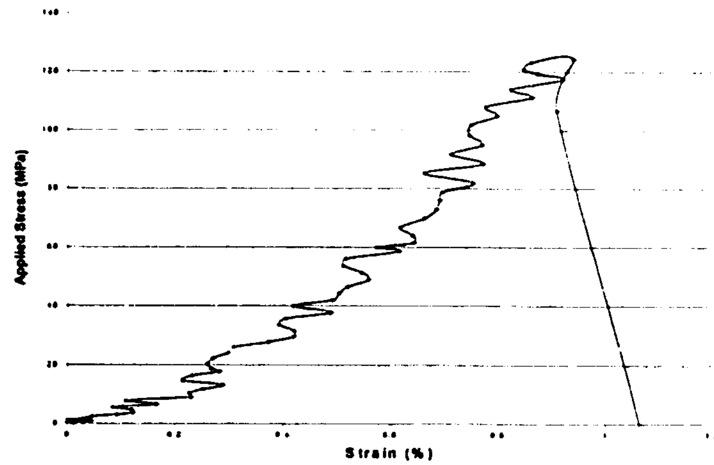
Uniaxial Compression Test: MN-WS-03



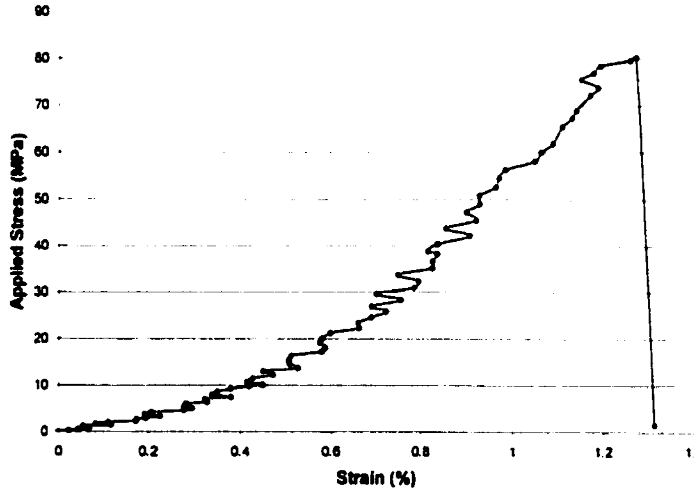
Uniaxial Compression Test: MN-WS-04



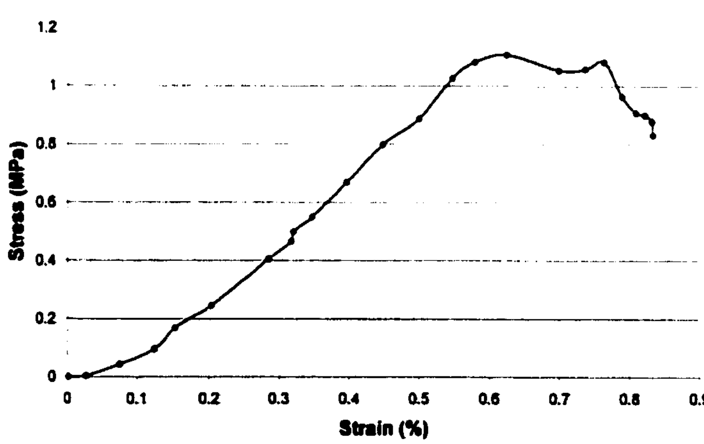
Uniaxial Compression Test: MN-MD-05

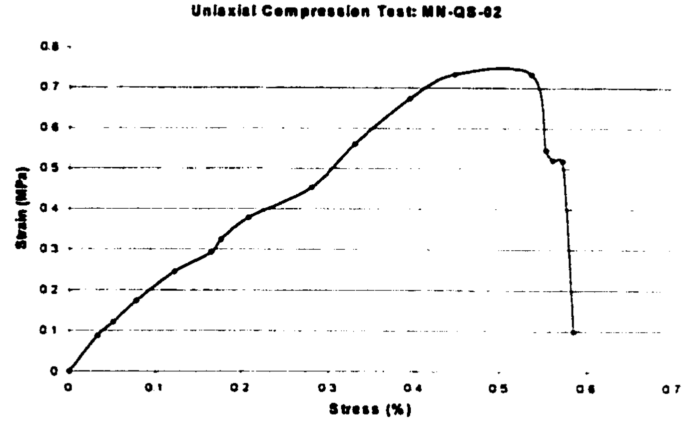


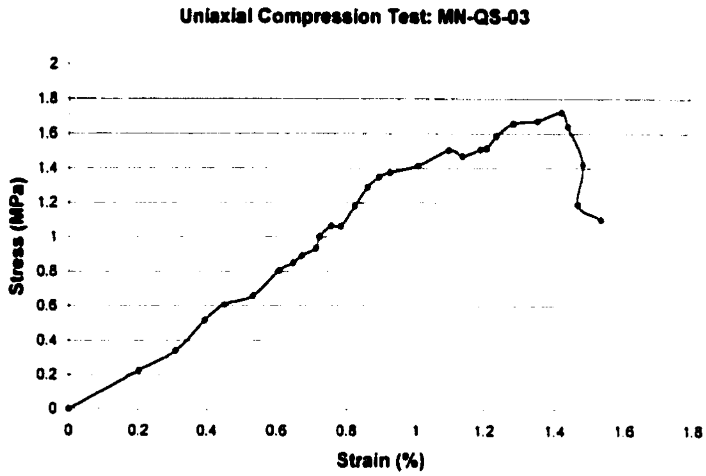
Uniaxial Compression Test: MN-WS-06



Uniaxial Compression Test: MN-QS-01

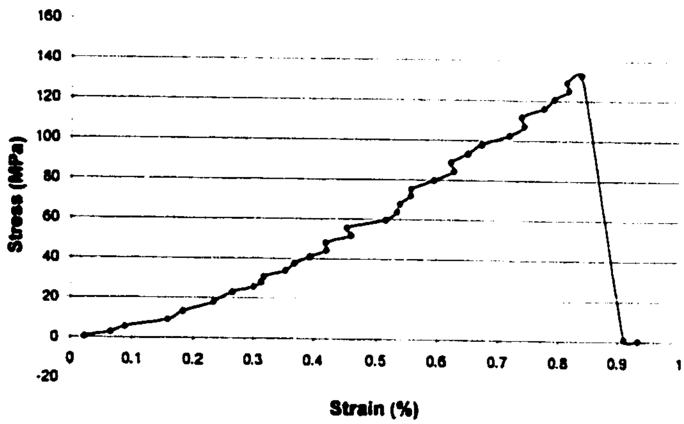



Uniaxial Compression Test: MN-QS-02
Strain (MPa)
Stress (%)
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8

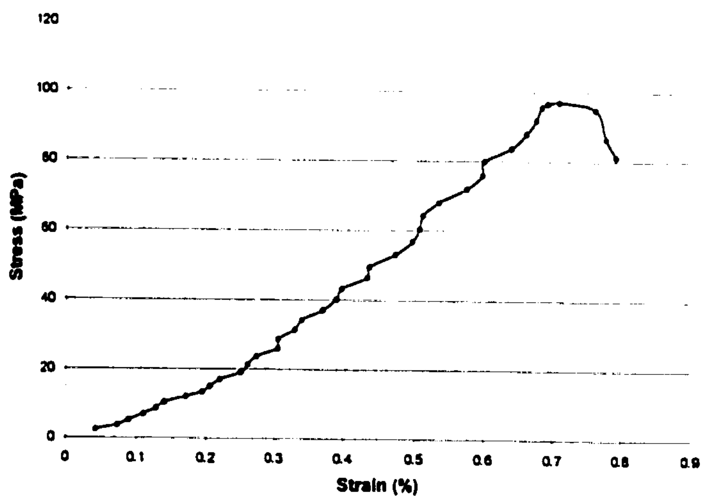

Uniaxial Compression Test: MN-QS-03
Stress (MPa)
Strain (%)
0
0.2
0.4
0.6
0.8
1
1.2
1.4
1.6
1.8
2

Appendix 2: Stress vs Strain Curves for Milton Heights

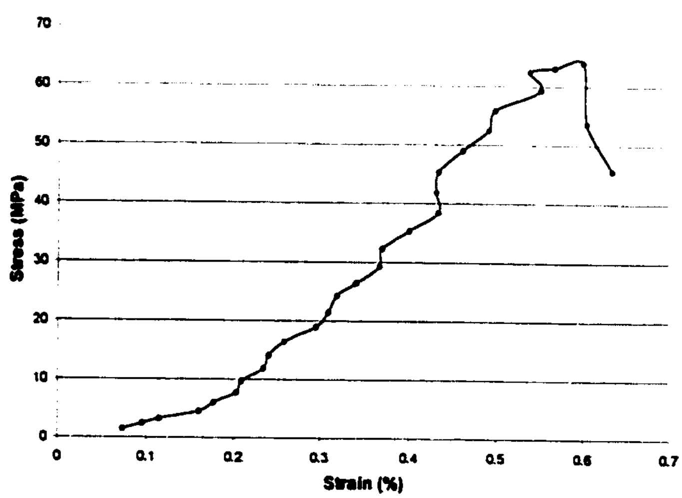
Uniaxial Compression Test: MI-AD-02

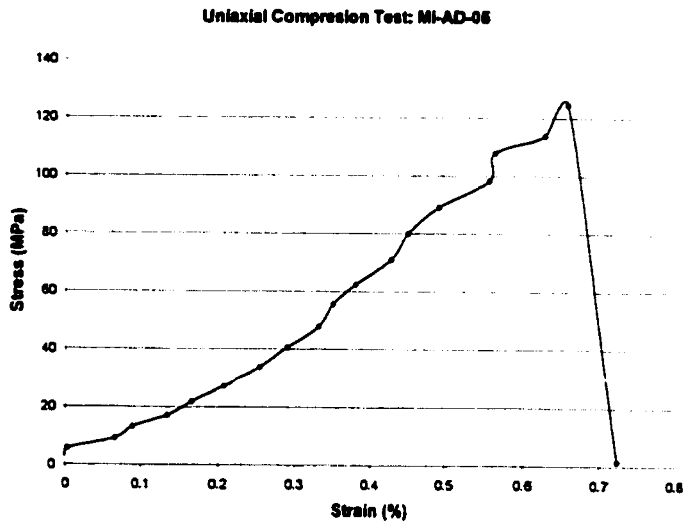


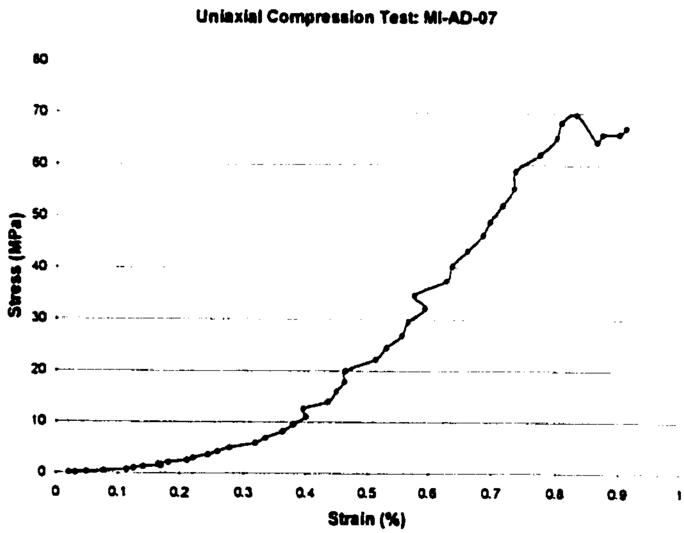
Uniaxial Compression Test: MI-AD-03

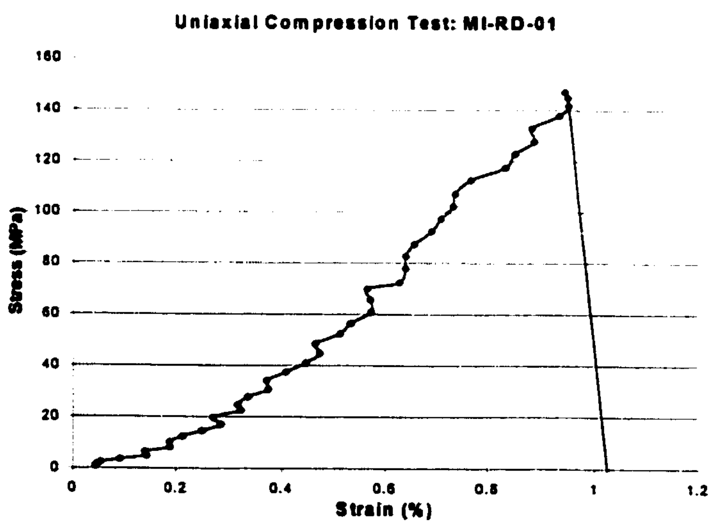


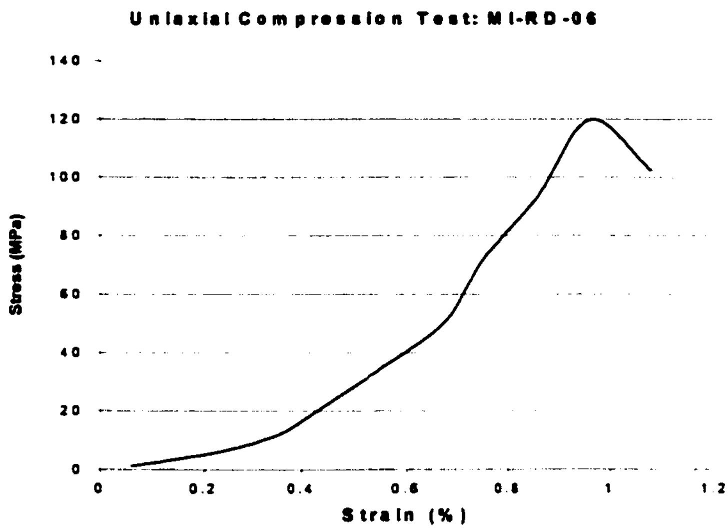
Uniaxial Compression Test: MI-AD-04

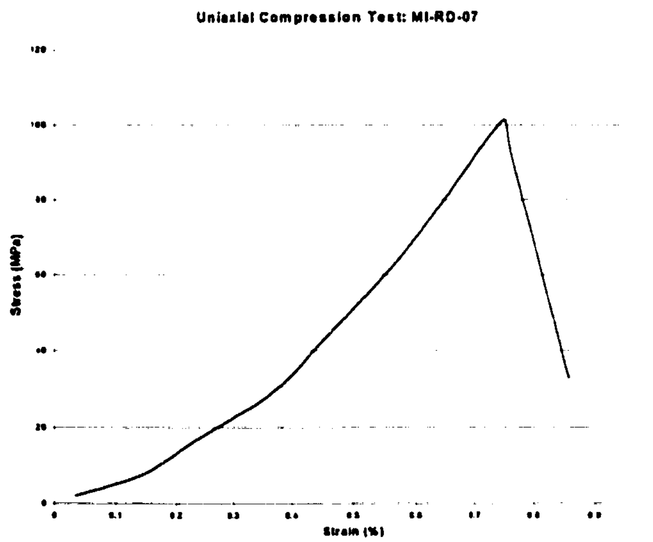


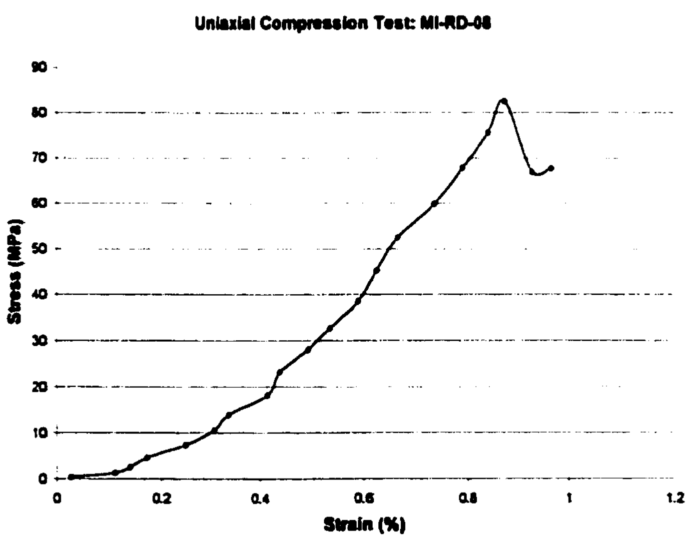

Uniaxial Compression Test: MI-AD-06
Stress (MPa)
140
120
100
80
60
40
20
0
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
Strain (%)

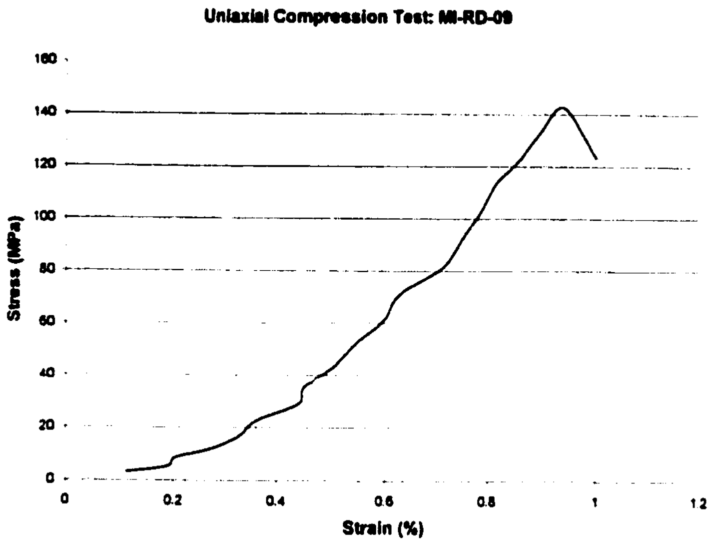

Uniaxial Compression Test: MI-AD-07
Stress (MPa)
80
70
60
50
40
30
20
10
0
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
1
Strain (%)

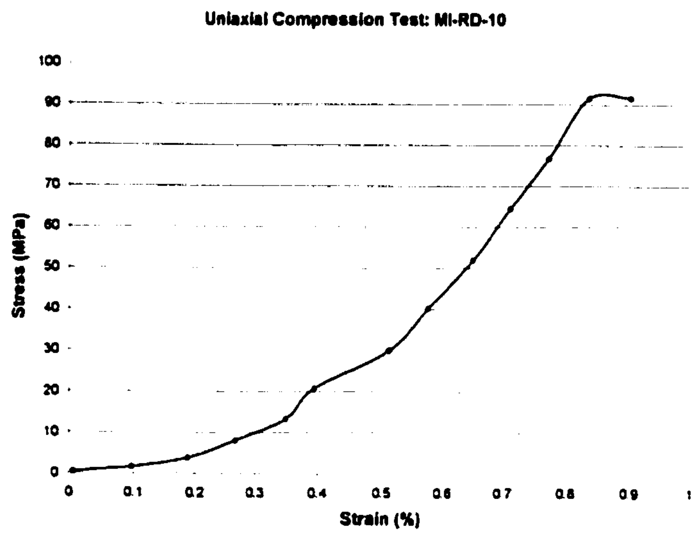

Uniaxial Compression Test: MI-RD-01
Stress (MPa)
160
140
120
100
80
60
40
20
0
0
0.2
0.4
0.6
0.8
1
1.2
Strain (%)

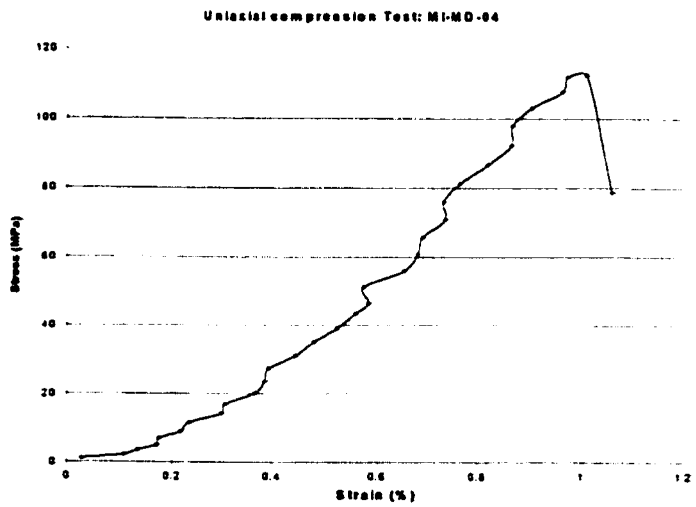

Uniaxial Compression Test: MI-RD-06
Stress (MPa)
Strain (%)
140
120
100
80
60
40
20
0
0
0.2
0.4
0.6
0.8
1
1.2

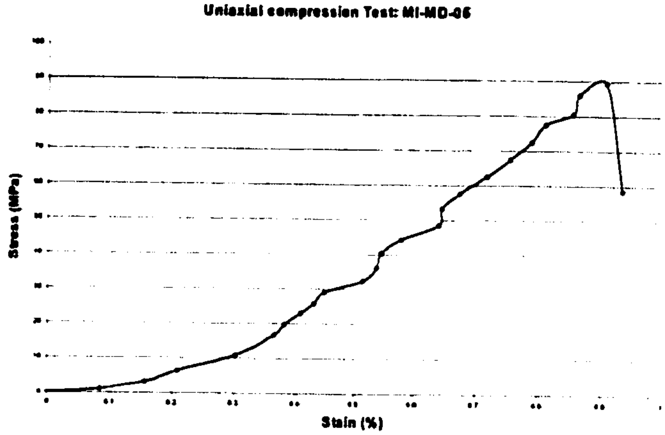

Uniaxial Compression Test: MI-RD-07

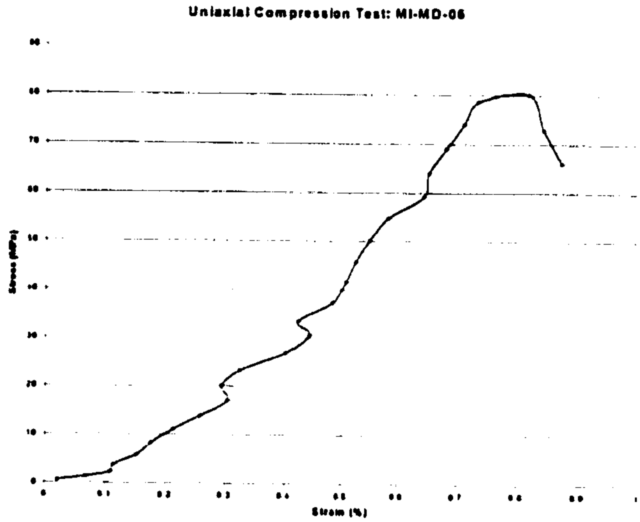

Uniaxial Compression Test: MI-RD-08
Stress (MPa)
Strain (%)
90
80
70
60
50
40
30
20
10
0
0
0.2
0.4
0.6
0.8
1
1.2

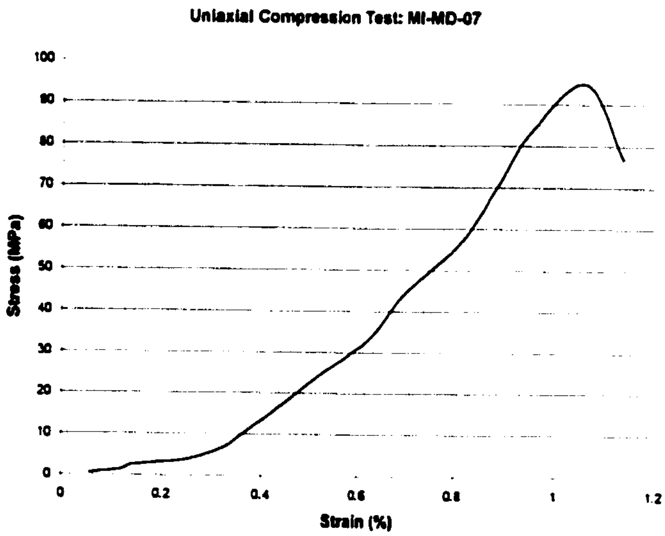

Uniaxial Compression Test: MI-RD-09
160
140
120
100
80
60
40
20
0
Stress (MPa)
0
0.2
0.4
0.6
0.8
1
1.2
Strain (%)

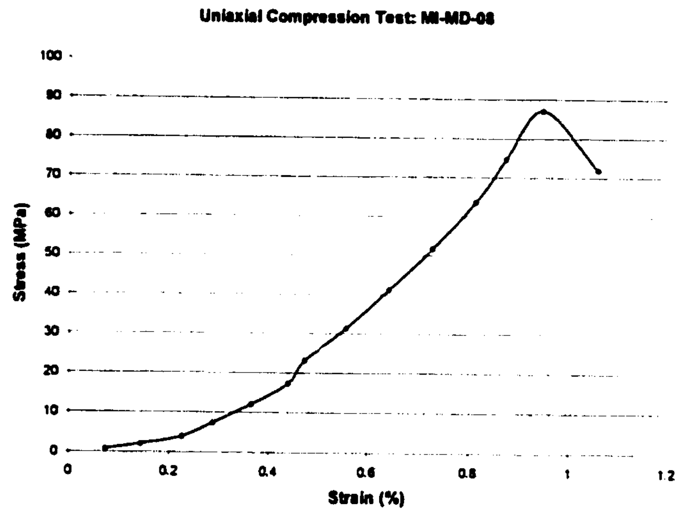

Uniaxial Compression Test: MI-RD-10
100
90
80
70
60
50
40
30
20
10
0
Stress (MPa)
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
1
Strain (%)

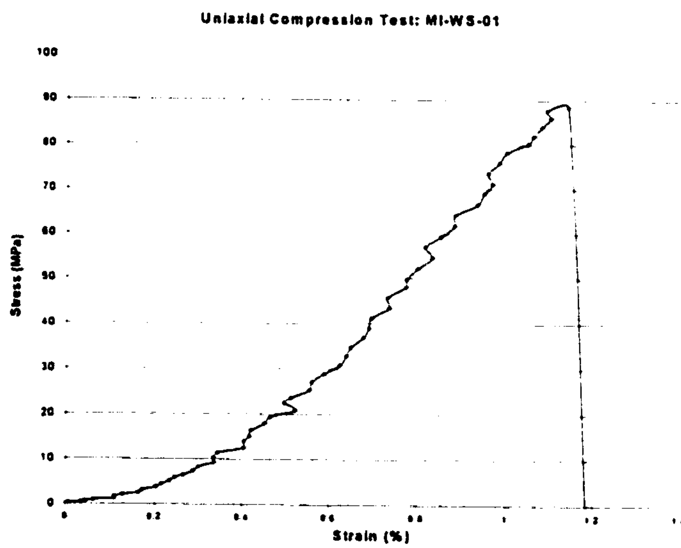

Uniaxial compression Test: MI-MD-04
120
100
80
60
40
20
0
Stress (MPa)
0
0.2
0.4
0.6
0.8
1
1.2
Strain (%)

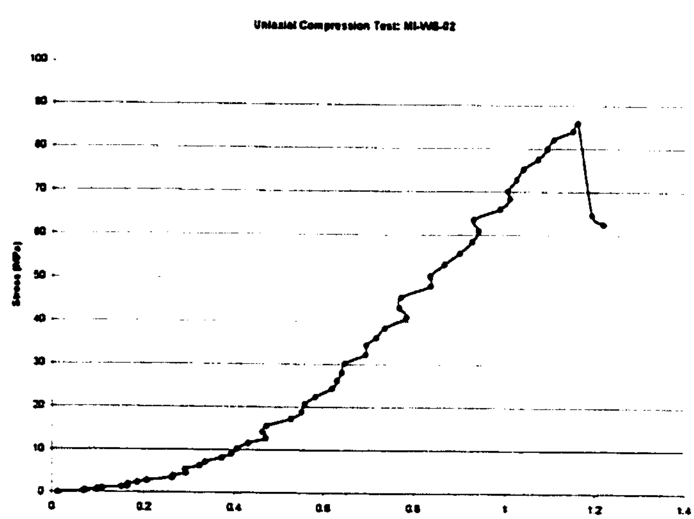

Uniaxial compression Test: MI-MD-05
Stress (MPa)
Stain (%)

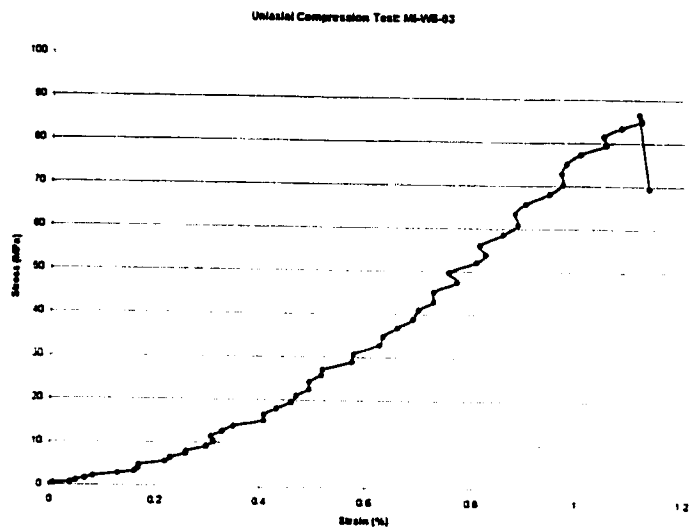

Uniaxial Compression Test: MI-MD-06

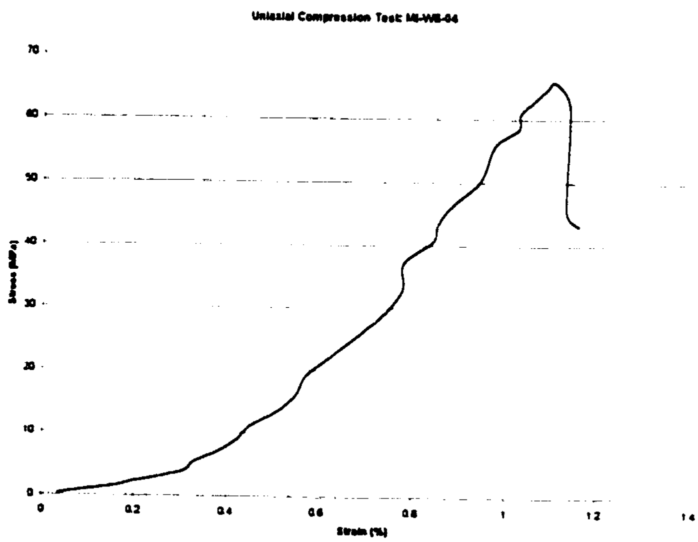

Uniaxial Compression Test: MI-MD-07
100
90
80
70
60
50
40
30
20
10
0
Stress (MPa)
0
0.2
0.4
0.6
0.8
1
1.2
Strain (%)


Uniaxial Compression Test: MI-MD-08
Stress (MPa)
Strain (%)
100
90
80
70
60
50
40
30
20
10
0
0
0.2
0.4
0.6
0.8
1
1.2

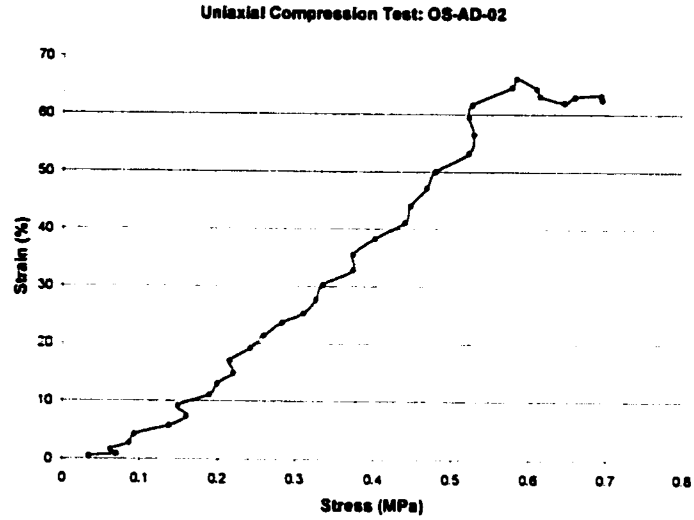

Uniaxial Compression Test: MI-WS-01
Stress (MPa)
Strain (%)
100
90
80
70
60
50
40
30
20
10
0

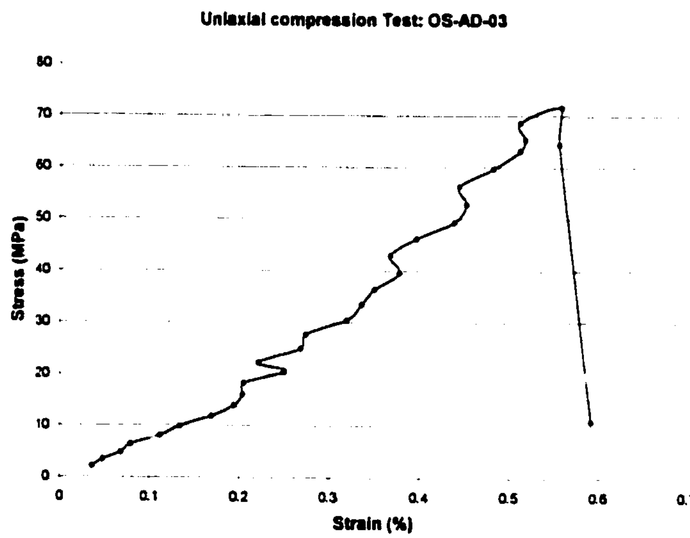

100
90
80
70
60
50
40
30
20
10
0
0
0.2
0.4
0.6
0.8
1
1.2
1.4

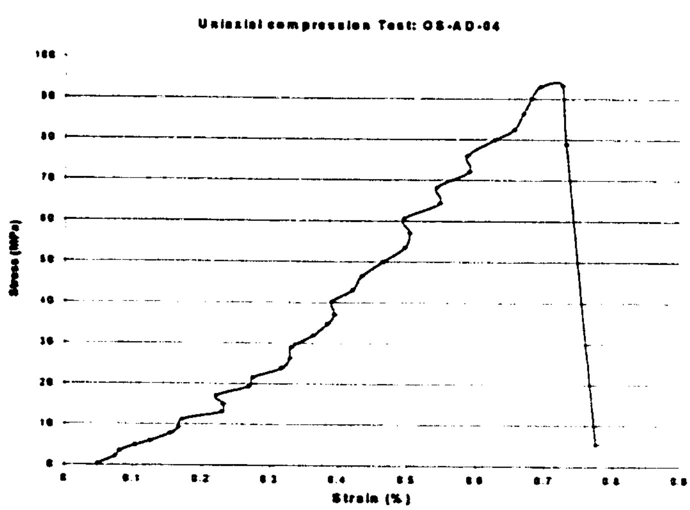

Uniaxial Compression Test: MI-WB-03
Stress (MPa)
Strain (%)
0
10
20
30
40
50
60
70
80
90
100
0
0.2
0.4
0.6
0.8
1
1.2


Uniaxial Compression Test: MI-WB-04
Stress (MPa)
Strain (%)
0
10
20
30
40
50
60
70
0
0.2
0.4
0.6
0.8
1
1.2
1.4

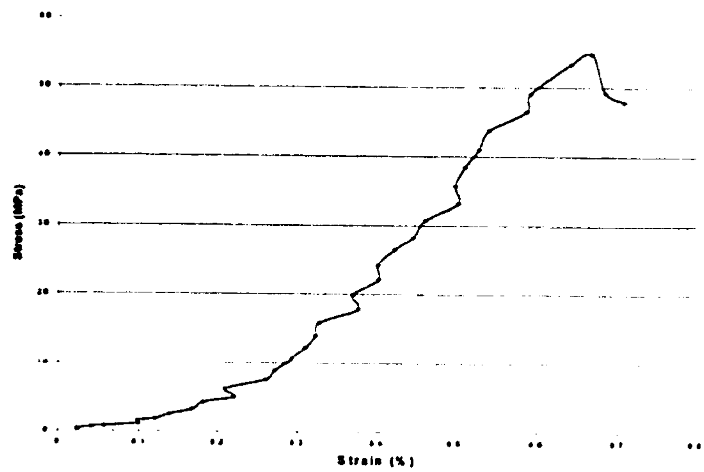
Appendix 3: Stress vs Strain Curves for Osler Bluff



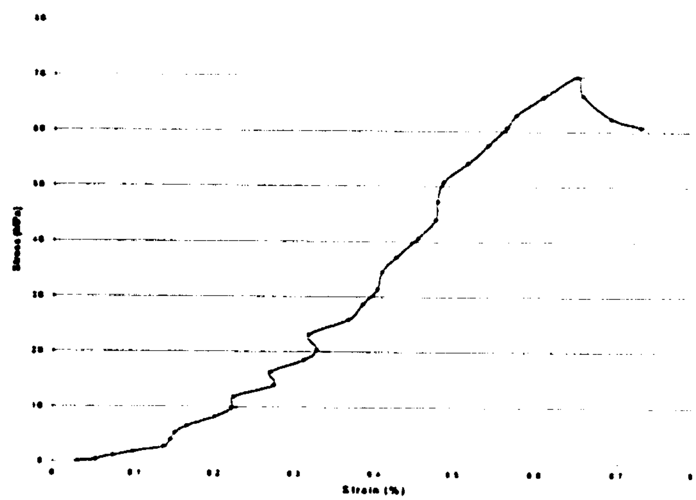




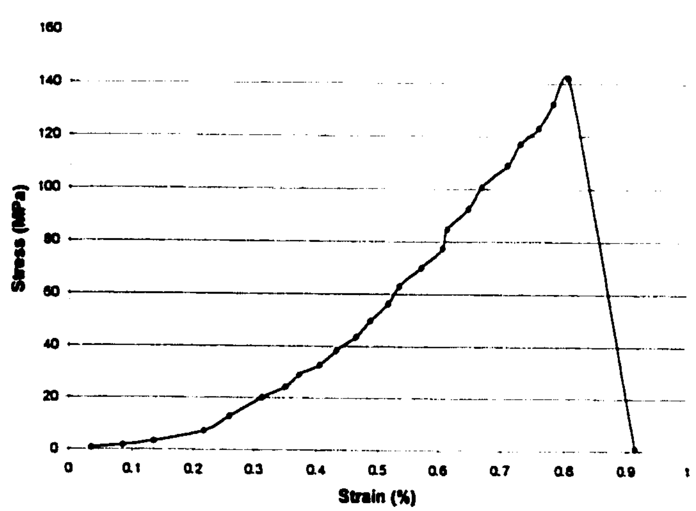
Uniaxial Compression Test: OS-AD-05

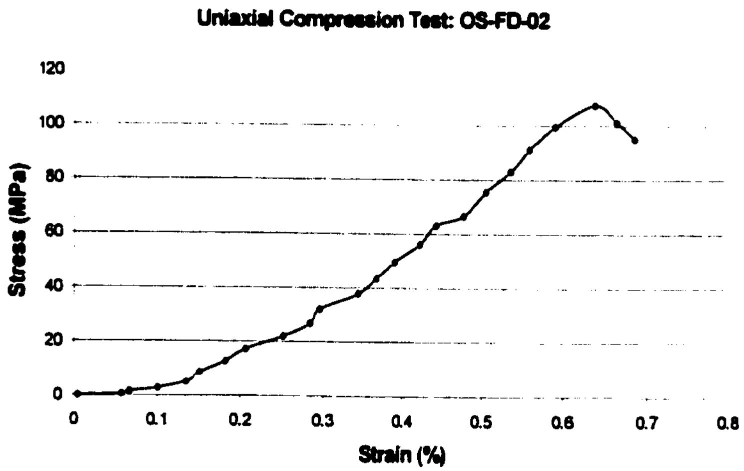


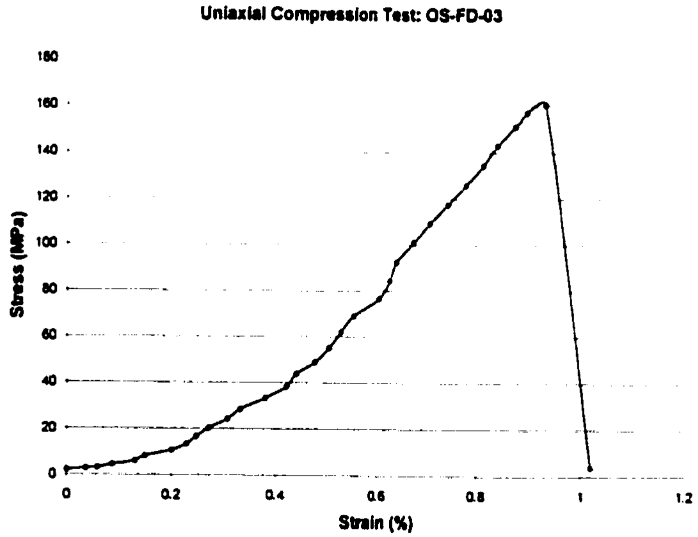
Uniaxial compression Test: OS-AD-06

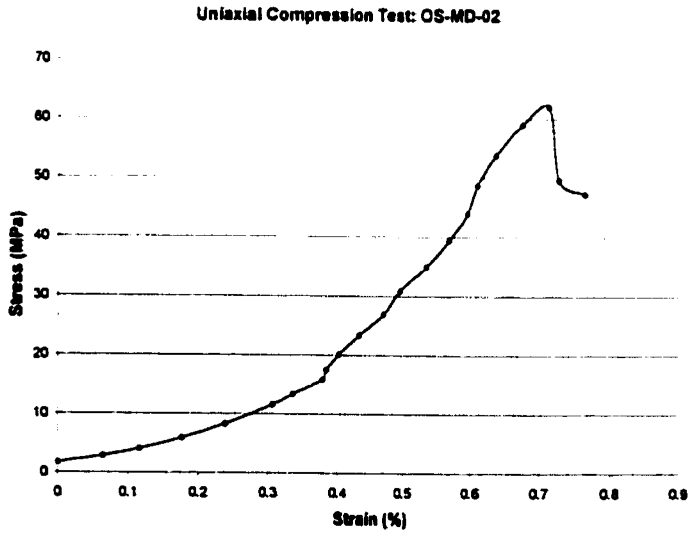


Uniaxial Compression Test: OS-FD-01

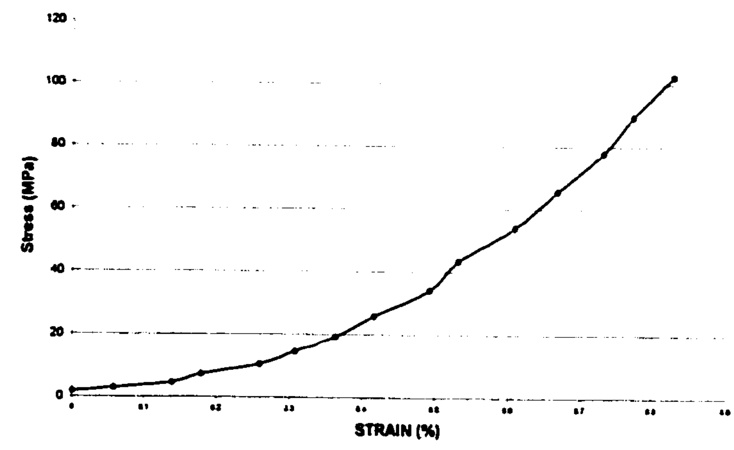



Uniaxial Compression Test: OS-FD-02
Stress (MPa)
Strain (%)
120
100
80
60
40
20
0
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8

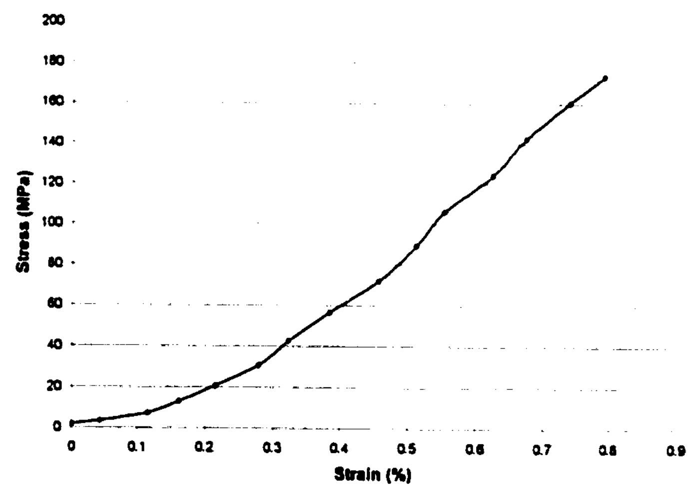

Uniaxial Compression Test: OS-FD-03
Stress (MPa)
Strain (%)
180
160
140
120
100
80
60
40
20
0
0
0.2
0.4
0.6
0.8
1
1.2


Uniaxial Compression Test: OS-MD-02
Stress (MPa)
Strain (%)
70
60
50
40
30
20
10
0
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9

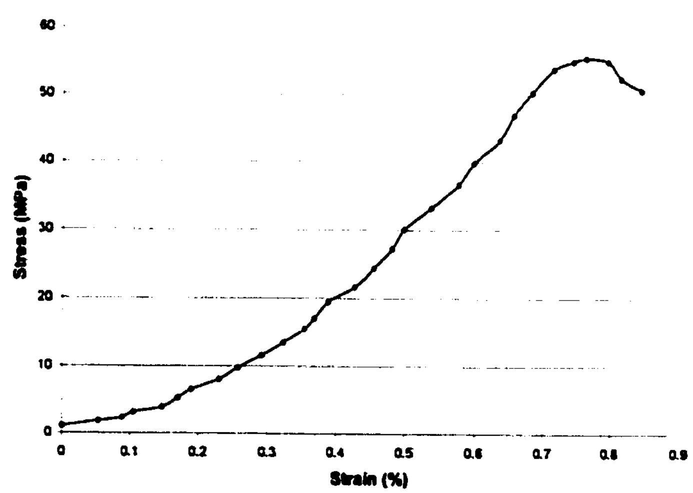
Uniaxial Compression Test: OS-MD-03



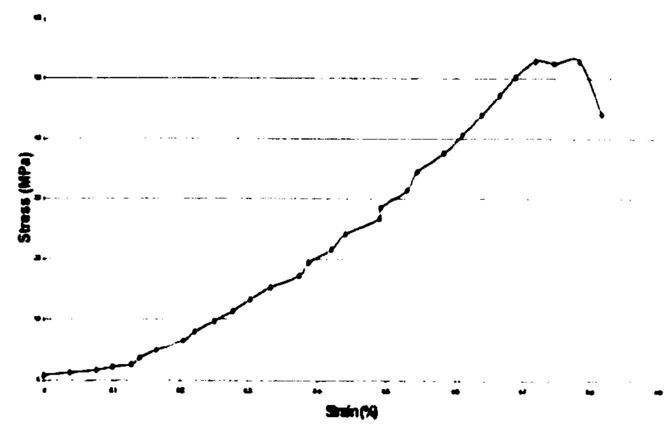
Uniaxial Compression Test: OSMD04



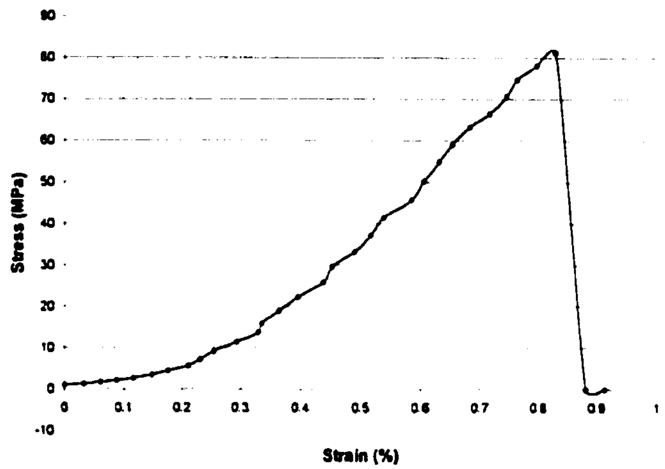
Uniaxial Compression Test: OS-WS-02



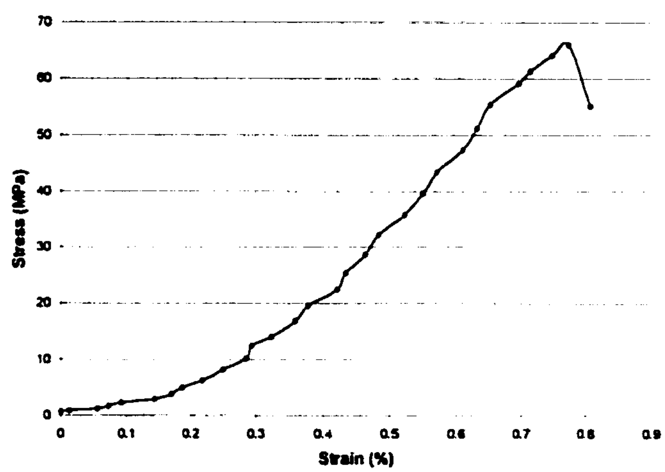
Uniaxial compression Test: OS-WS-03

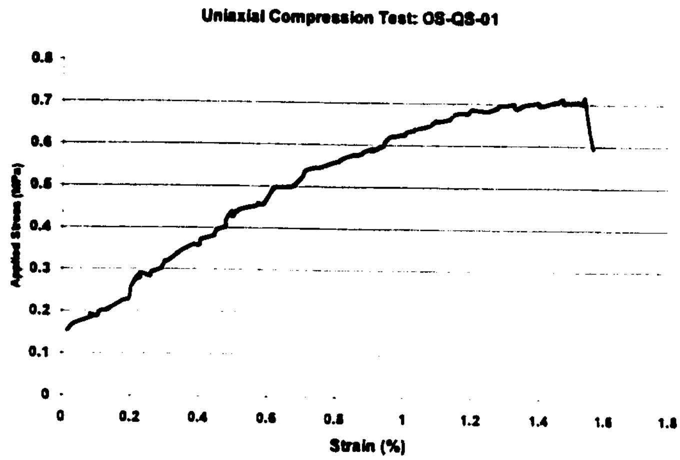


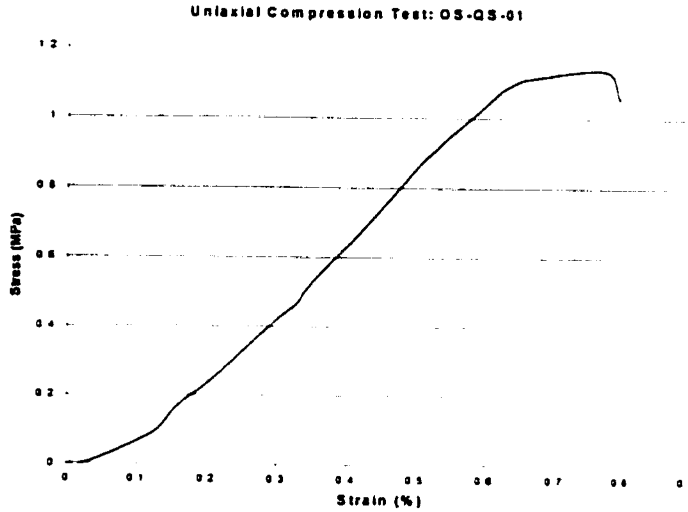
Uniaxial Compression Test: OS-WS-04



Uniaxial Compression Test: OS-WS-05




Uniaxial Compression Test: OS-QS-01
Applied Stress (MPa)
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
0
0.2
0.4
0.6
0.8
1
1.2
1.4
1.6
1.8
Strain (%)


Uniaxial Compression Test: OS-QS-01
Stress (MPa)
1.2
1
0.8
0.6
0.4
0.2
0
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
Strain (%)

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