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HOLOCENE DEGLACIAL AND SEA LEVEL HISTORY OF DOBBIN BAY, EASTERN ELLESMERE ISLAND, ARCTIC CANADA

by

SCOTT M. ROBERTSON ©

A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfillment of the requirements for the degree of Master of Science

Department of Earth and Atmospheric Sciences

Edmonton, Alberta

Spring 1999

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled "Holocene Deglacial and Sea Level History of Dobbin Bay, eastern Ellesmere Island, Arctic Canada" submitted by Scott M. Robertson in partial fulfillment of the requirements for the degree of Master's of Science.

Dr. J. England (Supervisor)

Dr. N. Rutter

Dr. M. Hickman

April 16, 1999

ABSTRACT

This thesis presents the Holocene deglacial and sea level history at the head of Dobbin Bay, east Ellesmere Island. An ice-contact delta at the fiord head contains two tills separated by deltaic foreset beds dated 23.3 ka BP. This provides a maximum age for the onset of late Wisconsinan glaciation. Marine limit at the fiord head is 71 m asl, and likely dates ≤7.3 ka BP.

Since the mid Holocene, two lateral moraines (upto 10 km in length) were deposited along the south margin of Eugenie Glacier, draining from Agassiz Ice Cap. The outer moraine was formed below sea level via 'ice-push' of glaciomarine sediments whereas the inner moraine is currently forming via 'ice-thrust' of frozen sediment in a subareal setting. The Eugenie Glacier was in contact with the outer moraine on the 1959 airphoto, however; formation predates the arrival of driftwood dated 4.6 ka BP. Although, the inner moraine contains shells dated 3.4 ka BP its formation postdates the 1959 airphoto. Three advances of Eugenie Glacier are documented between mid to late Holocene.

Well preserved marine shorelines on the distal slope of the outer moraine extend from 33 to 4 m asl. The highest shoreline was surveyed and has a tilt of 0.52 m/km indicating greater uplift towards Eureka Sound (western Ellesmere Island). This direct measurement of differential tilting is in agreement with regional postglacial isobases.

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For me, this thesis has been a long journey, which at times seemed unending. Throughout this time, my supervisor Dr. John England has been there, providing much needed support and encouragement. I can't thank you enough for your *thorough* editing and countless discussions which always kept me moving forward. The experience I gained wandering around Dobbin Bay will never truly be equalled. Thank you for this opportunity. As well, thank you to my committee (Dr. Nat Rutter and Dr. Michael Hickman) for their time and concise edit of this thesis.

I would like to thank Bonnie Robinson for her dedicated field assistance and conversations during our stay at Dobbin Bay. I would also like to thank Dr. Dave Evans for his contribution (both through conversation and comedic relief) to my understanding the dynamics of my field area.

I can't imagine a stranger bunch of misfits than the ones who shared the early years of my Master's. In particular, thanks to Mark Barling, John Woodward, Mark Skidmore. A special thanks to Craig Coburn for all of his help and encouragement both at U of A and SFU. As well, thanks to Vince Miller, you definitely made those early years something to remember. More recently, I would like to thank Stefan and Chelsea, Kirby and Chelsea, and Peter (truly the best man) for their support and encouragement.

Of course, this thesis would not have been possible if not for the support of my family. Thanks to my parents (Brian and Jeanette Robertson) for their love, support, encouragement, and unending faith in my ability to complete this endeavour. Thanks to my brother John, for his understanding and friendship.

Most of all, I would like to thank my partner, Lora. You have been a pillar of strength for me throughout this journey. Without your constant support, encouragement and love, I cannot imagine ever making it through. I am grateful for all of the support through both the highs and lows I've experienced while completing my thesis. It is truly because of your day to day encouragement and faith that I am able to finally thank you for this thesis. I only hope that one day I can give back to you the unbridled love and support you've shown me over the past years. Thank You, Lora.

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1.0 Study Area

This study focuses on the glacial geomorphology and late Quaternary /Holocene history surrounding Eugenie Glacier, at the head of Dobbin Bay, eastern Ellesmere Island (79° 45' N, 75° 00' W). The fiord head at Dobbin Bay is characterized by a ~2 km wide lowland which is encircled by mountainous terrain with a relief of ~1300 m. Dobbin Bay is ~5-10 km wide and extends approximately 50 km inland from Kane Basin (Figs. 1.1 and 1.2). It has served as an eastward conduit for past advances of outlet glaciers from the southern Agassiz Ice Cap. The Dobbin Bay syncline runs along the axis of the innermost part of the bay and is composed of Devonian siltstone and sandstone which are weak and susceptible to erosion.

Presently, there are ten separate glaciers entering the fiord head of Dobbin Bay. Of these, Eugenie Glacier is by far the largest, descending for ~50 km from the southern Agassiz Ice Cap, terminating at the head of Dobbin Bay. At its terminus, Eugenie Glacier is characterized by a 5 km. wide floating tongue, with a freeboard of ~10-15 m. The other glaciers, extend through confined valleys and form piedmont lobes on the lowlands of the fiord head.

1.1 Regional Context

Numerous studies have presented field evidence concerning the extent of ice during the last glacial maximum (LGM) in the Canadian High Arctic (England 1978, 1983, 1992, 1997; Paterson 1977; England and Bradley 1978; Hodgson 1985; Dyke and Prest 1987; Bednarski 1986; Retelle 1986; Lemmen 1989; Evans 1990; Lemmen et al 1994; Bell 1996).

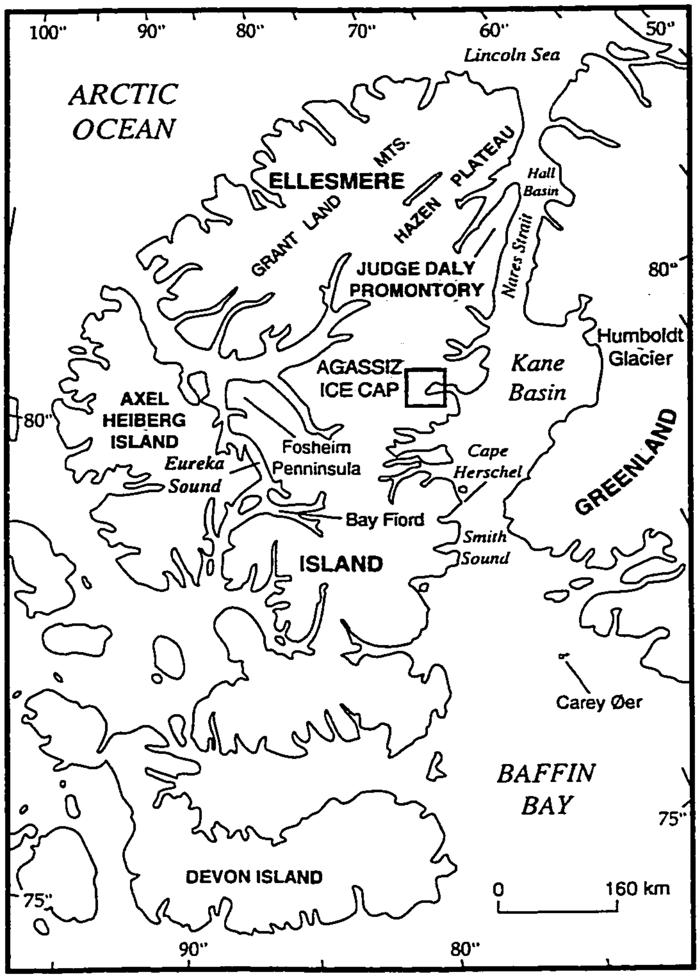

Figure 1.1 Location map showing the eastern Queen Elizabeth Islands. Study area is blocked out on the east coast of Ellesmere Island. Areas mentioned in the text are also labelled on the map. Shaded areas represent the modern limits of ice cover.

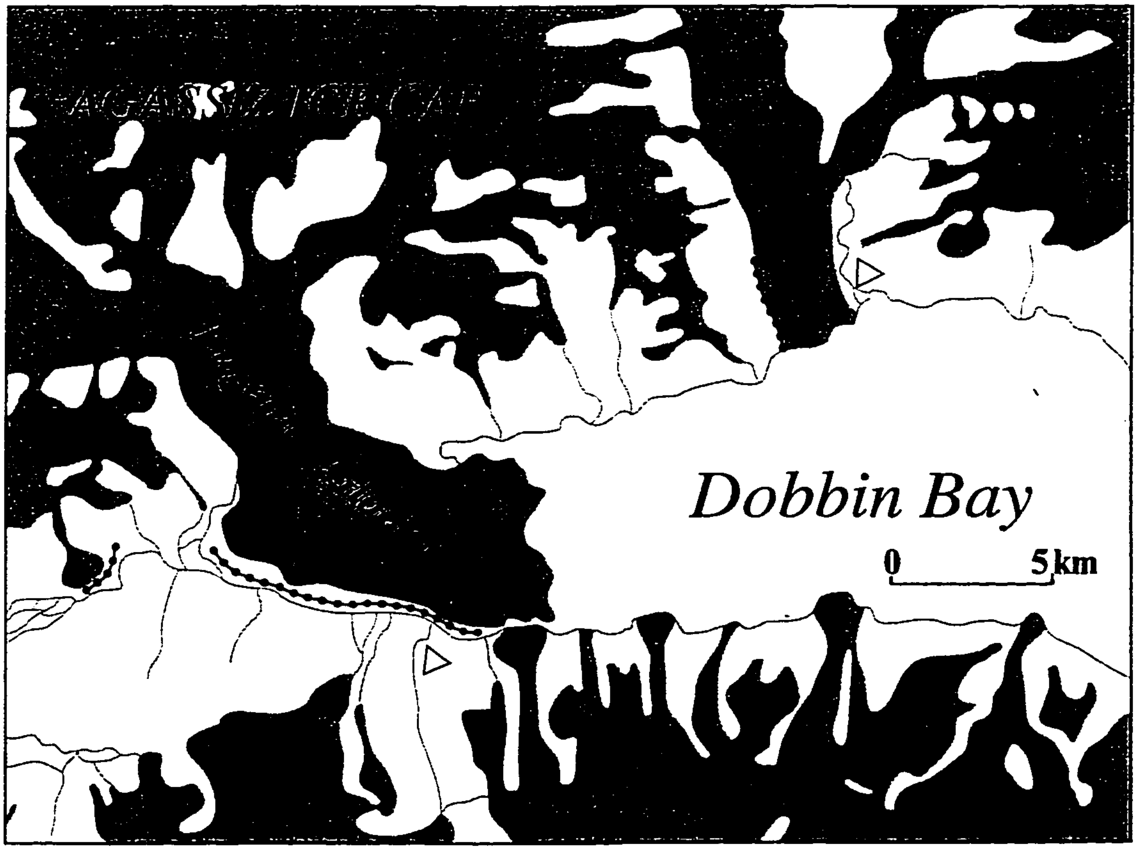

Figure 1.2 Dobbin Bay study area location map. Research focuses on the lowland located along the southern and western margin of Eugenie Glacier. The significant moraines in lowland (dotted line) and ice-contact delta (triangle) are marked on the map. Inner moraine not shown, however; it corresponds to the contemporary ice margin.

All of these studies emphasize the connection between former ice margins and sea level changes during the Holocene (Blake 1970, 1972, 1992a, 1992b; Walcott 1972; England 1976b, 1983, 1992; Evans 1990; Tushingham 1991; England et al. 1991). Until recently, two different models had been proposed for the late Quaternary history of the Queen Elizabeth Islands: one model favoured a pervasive ice cover (the Innuitian Ice Sheet, Blake 1970) whereas the other model favoured smaller local ice caps which left some land and adjacent marine channels ice-free (the Franklin Ice Complex, England 1976b).

During the completion of this thesis, England (1998, 1999) has undertaken a major revision of the eastern Ellesmere Island Quaternary record. This revision indicates that during the LGM, Greenland and Ellesmere Island ice were coalescent, infilling Nares Strait. This is based on more detailed mapping and new chronological evidence, including surface exposure dating (Cl-36) of bedrock and erratics (Zreda et al. 1999). AMS dating of shells within till, outwash and ice-contact marine deposits confirm a late Wisconsinan age whereas deglaciation of Nares Strait occurred between 10 and 7.5 ka BP (England 1997). Bednarski (1998) has also proposed that ice infilled Nansen Sound, between northern Ellesmere Island and Axel Heiberg island, during the LGM. This ice had to originate, in part, from Eureka Sound, hence it is now proposed that both major marine channels, west and east of Ellesmere island were occupied by ice during the LGM. Provisional isobases from England (1997) indicate that Dobbin Bay is located within a saddle between the Greenland uplift centre to the east and the Eureka Sound uplift centre to the west. Prior to deglaciation ice draining through Dobbin Bay extended into Kame Basin where it joined southward flowing trunk ice. The regional distribution of deglacial sites provided by England (1998), indicate that the sea penetrated Dobbin Bay during the between 8 and 7 ka BP. This new interpretation provides the context for the present thesis which emphasizes the late Quaternary and Holocene history at the head of Dobbin Bay.

Many of the Arctic glacial geomorphic studies have focused on mapping ice margins during the LGM based on moraines and ice-contact marine deposits. In addition, numerous sea level studies have been undertaken to determine the magnitude and pattern of postglacial rebound following the last glaciation (Andrews 1970; England 1976a, 1977). To achieve this goal, the highest shoreline, referred to as marine limit, is surveyed and dated along with lower shorelines recording the history of subsequent sea level regression. Marine limit for a particular area is a product of both the date of deglaciation and the distance from the former ice margin. Andrews (1968) noted that in Arctic Canada, deglaciation and the formation of marine limit is usually synchronous due to the late date of deglaciation which allowed postglacial emergence to exceed the rate of eustatic sea level rise (Andrews 1968). Because deglaciation in Dobbin Bay commenced at ~8 ka BP, the effect of eustatic sea level rise has been minimized. Based on the eustatic sea level curve of Fairbanks (1989), only ≤30 metres of sea level rise occurred after 8 ka BP.

1.2 Research Objectives

This thesis concerns the deglacial history at the head of Dobbin Bay and the history of two prominent lateral moraines flanking Eugenie Glacier, and along the outermost lateral moraine the profile of some exceptionally developed Holocene shorelines that extend for ~10 km long moraine. The research includes three objectives. The first objective involved the survey of glacial and marine landforms and sediments in the lowland. This included the history of relative sea level based on the elevation and age of shorelines at and below marine limit. Marine shells and driftwood from these raised shorelines were collected for radiocarbon dating. The first objective also included a detailed stratigraphy of an ice-contact delta located on the south side of Eugenie Glacier. The delta has been incised by a river, exposing a section ~40 m high and ~1.5 km long which contains several prominent depositional units. Previous research by J. England (1993), suggests that the delta is Holocene, as its elevation is similar to that of other deltas in the area which have been radiocarbon dated close to 7 ka BP (England 1996). In the middle of the delta section, there is a silt layer 1 - 2 m thick containing marine shells. The silt is laterally extensive, has a seaward dip, and underlies foreset beds. Initially, this unit was interpreted to represent proximal bottomset beds within a prograding delta. However, ^{14}C dating of a bulk sample of marine shells collected from the silt provided an age of 31,400 ±900 BP (GSC-5663, England 1996). This suggests that the silt bed either contained redeposited shells or that the unit itself is much older and has survived subsequent erosion. The purpose of this research was to re-examine the stratigraphy and sedimentology of the delta and resample the shells in order to test the validity of the initial radiocarbon date.

The second objective of this research was to study two large lateral moraines located along the southern margin of Eugenie Glacier. These are among the longest moraines (~10 km) observed in the Canadian High Arctic. Reconnaissance of these lateral moraines in 1993 indicated that they were composed of widespread marine sediments, likely displaced from below sea level (J. England, pers. comm., 1995). The stratigraphy of the moraines was reexamined in order to determine their mode of formation, and to clarify their history of deposition. Marine shells and driftwood were collected for radiocarbon dating.

Superimposed on the outermost moraine flanking Eugenie Glacier are a series of shorelines which mark former sea levels. A number of shorelines extend along the entire length of the moraine, thereby providing an excellent opportunity to survey the delevelling of synchronous water planes. Hence, the third objective of this study was to survey the shorelines along the moraine in order to determine the amount of differential titling that has taken place since deglaciation. Two shorelines (the uppermost and lowest) were surveyed in order to obtain the gradients of differential tilting which spanned the greatest time during deposition. The rate of emergence slows down as it approaches the present, and therefore, the lower and younger shorelines should have the lowest gradients. The shorelines were also investigated for embedded driftwood logs that could be radiocarbon dated, providing a maximum age for these shorelines.

Recent reconstructions of the regional isobases for eastern Ellesmere Island (8 ka BP, O'Cofaigh 1999) indicate that there was greater uplift to the west. The surveying of the shorelines along the Eugenie moraine will therefore be used to test this hypothesis directly. For example, previous regional isobases have widely utilized deglacial marine-limit deltas, which are widely separated (tens of kms.) and involve less precise chronological control (±100's of years). The ability to survey and radiocarbon date individual mid to late Holocene shorelines along the Eugenie moraine provides an unambiguous measurement of shoreline tilt upon which isobases can be drawn.

1.3 Research Methods

Fieldwork was preceded by the interpretation of landforms and sediments from airphotos. The airphotos used in this study were produced by the Department of Energy, Mines and Resources (Airphotos A-16612-48/49 and A-16613-82/83). The identified units were then transcribed onto a topographic map (Dobbin Bay, 39H & 29G, 1:250,000) and field sites were selected for closer inspection.

1.3.1 Field Methods

The collection of data in the field included standard stratigraphic and sedimentological techniques at exposed sections, as well as surveying key landforms by micro-altimetry, and level and stadia to determine their elevation of landforms. Elevations determined by microaltimetry (Wallace and Tiernan) were corrected for both temperature and pressure. When carefully corrected for these controls the accuracy of the altimeter is $\pm$ 2 m up to ~100 m asl and $\pm$0.5 m within 10 m asl (Evans 1990). All elevations were measured from high tide. Sea level was difficult to access because it was located seaward of a piedmont glacier whose calving margin is narrowly separated from Eugenie Glacier. For this reason, the elevation above sea level of the first base camp was established by helicopter. This benchmark was used for all subsequent surveys. Chronological control was provided by the collection of datable materials (predominantly shells) related to the formation of former sea levels and the displacement of Holocene marine sediments to form moraines.

For maximum precision, the well-preserved shorelines along the outer Eugenie moraine were surveyed by level and stadia. In many areas, the shorelines were represented by distinct notches on the moraine. Due to the large number of preserved shorelines along the moraine, it was important to survey those that showed little disruption by subsequent mass wasting and periglacial processes. Because all of the shorelines are occasionally intermittent, careful attention was given to continuing the survey along the correct, selected shoreline. This was accomplished by walking along the selected shoreline and carefully marking reference points for the surveying. The shorelines were then surveyed using standard levelling techniques, where backsight to foresight readings were taken, together with instrument height and the distance to the stadia on each backsight and foresight. This permitted the calculation of elevation and distance along the shoreline, from which the profile of that shoreline could be established.

1.3.2 Laboratory Methods

In the lab, shell and driftwood samples collected in the field were cleaned and prepared for radiocarbon dating. The shell samples were first washed in distilled water and placed in a sonic bath (Bransonic 220 Sonic Bath) for 5 minutes in order to remove any attached organics or sediment particles. The sample was then removed from the bath and any remaining particles were scraped off. The samples were then placed on tinfoil and allowed to air dry for 24 hours. Once dry, the samples were weighed and placed in new sample bags and sent for radiocarbon dating. Driftwood samples were cleaned by removing all surface wood coated with sediment and/or showing oxidation.

Three samples were dated at two different laboratories. A wood sample was dated by Beta Analytic (Miami) using conventional (beta counting) radiocarbon dating. The other two samples of marine shells (*Hiatella arctica)* were dated at IsoTrace Laboratory (University of Toronto) by Accelerator Mass Spectrometry (AMS). This technique of dating involves measuring the concentration of individual ^{14}C ions (Stuvier 1978a). The ions are accelerated in a cyclotron to extremely high velocities where they pass through a magnetic field which separates different ions (by mass) for analysis (Stuvier 1978a). This differs from the conventional method of ^{14}C dating which utilizes one of two different methods. The first method is known as the proportional gas counter technique. This involves converting the carbon to a gas (i.e. methane or carbon dioxide) which is placed in a 'proportional counter' capable of detecting β particles (Bradley 1985). The other method used in conventional dating is the liquid scintillation technique. This process involves converting the carbon to a benzene (or other organic liquid) and placing the sample in an instrument which detects scintillations (or flashes of light) produced by the β-particle emissions (Bradley 1985). Both of these methods require stringent controls on the influence of background atmospheric radioactivity (Bradley 1985). Beta Analytic used the liquid scintillation method to determine the age of the samples submitted.

1.4 Conclusions

The remainder of this thesis is subdivided into three chapters. Chapter two addresses the glacial and sea level history of the lowland south of Eugenie Glacier. Chapter three addresses the formation and importance of the moraines flanking the south margin of Eugenie Glacier. As well, Chapter three presents the survey of the shorelines superimposed on the outermost moraine. Chapter four concludes the thesis by integrating the preceding observations on the deglacial and postglacial geomorphology, stratigraphy, and chronology.

CHAPTER 2

2.0 Introduction

The glacial and sea level history at the head of Dobbin Bay is subdivided into two areas: 1) the lowland along the southern margin of Eugenie Glacier which is subsequently referred to as 'lower' Dobbin valley; and 2) the extension of the lowland farther west (inland) of Eugenie Glacier, which is referred to as 'upper' Dobbin valley. Because these two areas are contiguous, their separation here is simply for convenience.

2.1 'Lower' Dobbin Valley

The 'lower' valley is bounded by Eugenie Glacier to the north and by a large piedmont glacier (unnamed) to the east which separates the lowland from Dobbin Bay (Figure 2.1). The western (inland) boundary of the 'lower' valley is arbitrarily located where the lowland merges with the adjoining 'upper' valley. The southern boundary is comprised by local mountains supporting several glaciers. The margin of Eugenie Glacier is flanked by an exceptional lateral moraine which trimmed by well-preserved shorelines. The main river (unnamed) draining the lowland flanks the distal side of the outer Eugenie moraine and eventually breaches it at its eastern end near the unnamed piedmont glacier.

A large (~1 km wide) hanging valley extends through the local mountains and joins the 'lower' valley from the south (Site A, Fig. 2.1). Five kilometres south of Dobbin Bay, the interior of the hanging valley contains several piedmont lobes which previously coalesced. These glaciers formerly impounded proglacial lakes, one of which, displayed on the Topographic map and 1959 airphotos for Dobbin Bay (39H & 29G, 1959 (updated 1987)), has since drained. A prominent canyon, up to 70 m deep and ~10 m wide, has

Figure 2.1 Airphoto of 'lower' Dobbin valley (National Airphoto Library A-16612-48, 1959). Those features described in the texted are label with corresposnding letters on the photo. As well, the (former) ice dammed lake (IDL), outer piedmont glacer (OP), Eugenie Glacier (EG) and Dobbin Bay (DB) are marked. Note only the outer Eugenie Moraine is depicted on the 1959 airphoto. The inner moraine is currently forming at the contemporary ice margin.

incised the bedrock of the hanging valley and likely records both prolonged seasonal, as well as catastrophic drainage (jaukalaup) from the glacierized interior. Extremely coarse gravel bars below the canyon (boulders up to 1 m in diameter) may record the recent jaukalaup that resulted in the lake drainage (Site B, Fig. 2.1). The base of the canyon, at the junction with the lowland, contains a prominent raised marine delta whose incision also exposes a >40 m section of late Quaternary sediment (discussed later in this Chapter).

A prominent series of nested meltwater channels occupy the upper slopes of the hanging valley. The configuration of these channels record the successive retreat of a former trunk glacier southward from 'lower' Dobbin valley into the hanging valley (Site C, Fig. 2.1). Below these channels within Dobbin Valley are a series of high (~200-400 m asl.) lateral moraines located on the upper slopes immediately west of the hanging valley (Site D, Fig. 2.1). These moraines trend across the bedrock strike. Given the elevation and orientation of the moraines and adjacent meltwater channels, these moraines represent the coalescence of a large piedmont glacier with Eugenie Glacier. At this time, this coalescence would require an ice margin extending into Dobbin Bay well beyond the present ice margins.

2.1.1 Terrestrial Landforms

In the 'lower' valley, glacial landforms are sparse above marine limit whereas they are widespread adjacent to present day glaciers. The lack of well preserved glacial landforms above marine limit is likely due to the highly erodible bedrock which has produced abundant colluvium indicating an unstable platform for depositional landforms or adequate resistance for erosional ones. In localized settings, however; a sandy diamict containing 35% pebbles to boulders was found. Although the diamict is composed mostly of local clasts, 10-15% of the lithologies are limestone erratics. The provenance of these limestones is the Franklinian mobile belt located >20 km to the west, indicating the influence of glacial transport (Trettin 1989).

A large limestone erratic (1.5 x 2 m) occurs above marine limit (~73 m asl) at 127 m asl. It is found on Ordovician siltstone located to the southwest of Eugenie glacier. A mountain to the northwest of Eugenie Glacier (~4 km from the erratic) is the closest outcrop which requires a S to SE trajectory for the transporting ice. As well, within the 'lower' valley, at a much lower elevation (~36 m asl), two granite erratics occur between a local ice-contact delta and the eastern piedmont lobe (Site E, Fig. 2.1). These erratics (Fig. 2.2) are likely of Greenland provenance because the shield of SE Ellesmere Island does not underlie glacier catchments entering Dobbin Bay. Sea ice transport through Nares Strait is southward and likely these erratics were ice rafted either by sea ice or icebergs when rsl was >36 m asl in Dobbin Bay. Hence, without a firm connection to a relative sea level, these erratics could have been deposited anytime during or after glaciation (up until approximately the mid Holocene when relative sea level had fallen to 36 m asl).

Below marine limit, the most prominent glacial landforms are terminal and lateral moraines bordering present day ice margins (Site F & G, Fig. 2.1). Around the smaller glaciers in the 'lower' valley, the moraines are composed predominantly of sand and gravel and range from 5 to 10 m in height. The most conspicuous moraines, however, occur adjacent to Eugenie Glacier, where two parallel lateral moraines flank its southern margin. The innermost moraine marks the contemporary margin and is approximately 10 - 30 m high and reaches a surveyed elevation of 57 m asl. The inner moraine is discontinuous throughout

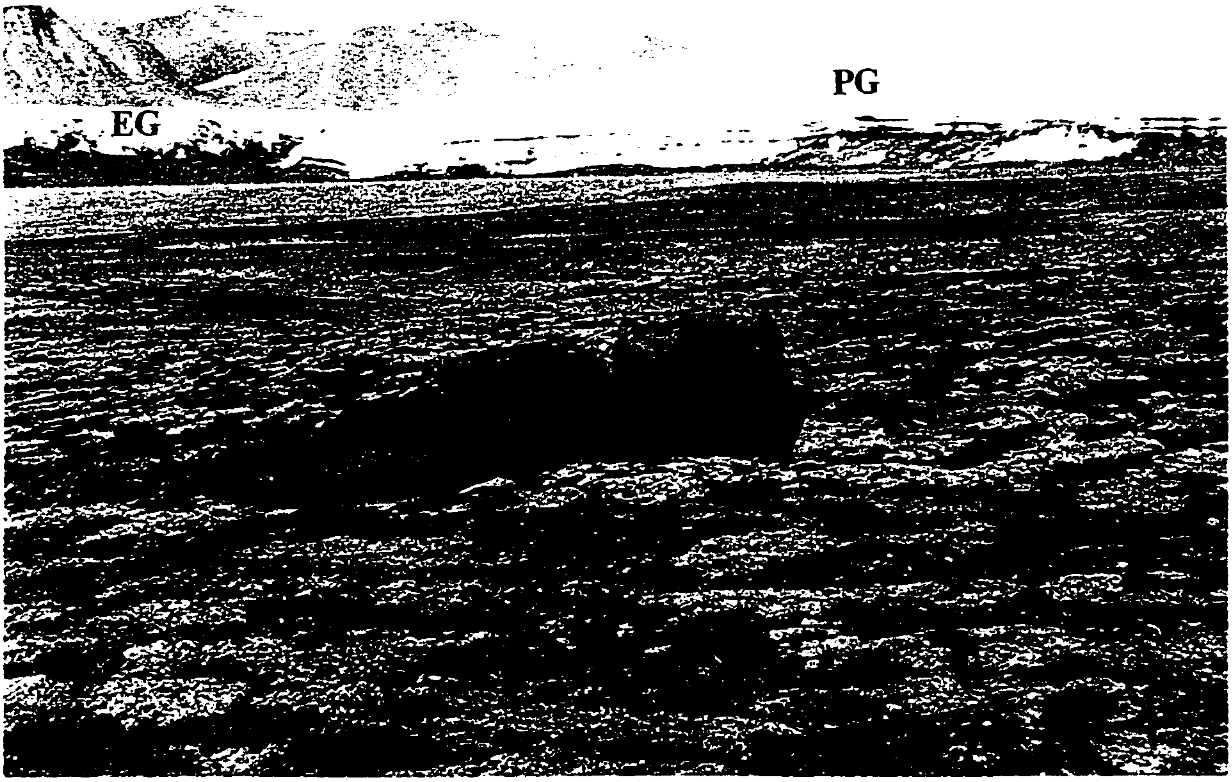

Figure 2.2 Photo of granite erratic in the lower Dobbin Valley. Erratic was found in the lowland south of Eugenie Glacier at ~36 m asl. Photo taken facing east towards Dobbin Bay. Eugenie Glacier (EG) and eastern Piedmont glacier (PG) in the background.

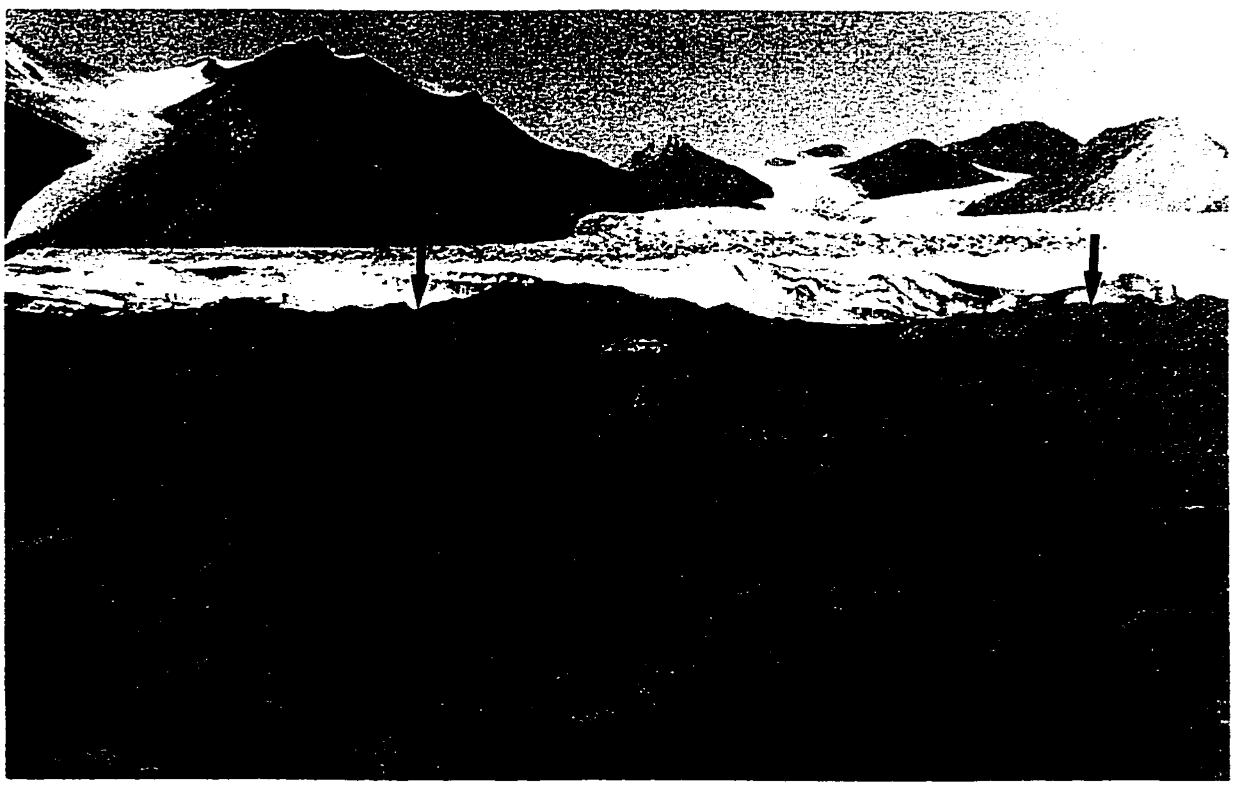

Figure 2.3 Photograph depicting both the Eugenie outer lateral moraine (open arrow) and the inner moraine (black arrow). Photograph taken towards the north, looking across Eugenie Glacier. Note the large erratic (circled) on distal slope of the outer moraine. The boulder is ~6x5x5 m. Prominent shorelines trim the base of the erratic and extend downslope.

its length but extends for ~6 km. This moraine is composed predominantly of silt and sand, with localized areas of angular boulders. Exposures in the moraine indicate that the silt is massive, containing both broken and single valves of marine pelecypods.

The outermost moraine is much larger than the inner moraine, both in length and relief. It ranges from 30 to 50 m high, reaching 54 m asl. The length of the outer moraine is also exceptional by Arctic standards, reaching 10 km whereas its width reaches >500 m (Site H, Figs. 2.1 and 2.3). The outer moraine is composed predominantly of marine silt and sand which contain abundant shell fragments, however; localized areas of gravel also occur. The shells found within the silt are commonly broken or occur as single valves. Exposures in this moraine indicate large scale folding although undisturbed sections of fine horizontally-bedded units also occur. The distance between the outer and inner moraines range between 0.2 - 1 km. In many locations between the moraines, buried glacial ice is exposed where marine silt has been removed. A more detailed discussion of these moraines is presented in the next chapter (Section 3.3).

2.1.2 Marine Landforms

The 'lower' valley contains widespread raised marine landforms and sediments extending to an ice-contact delta which marks marine limit at ~71 m asl. The morphology of this landform is consistent with other Holocene deltas within Dobbin Bay, however; its sedimentology and several radiocarbon dates on its marine shells suggest a more complex origin than simply Holocene progradation. Based on the morphology and foreset bedding within the deposit, this study will refer to this feature as a delta.

The ice-contact delta is located along the southern margin of the 'lower' valley where it joins the hanging valley (Site I, Fig. 2.1). Its surface is incised by an abandoned meltwater channel (spillway) which currently contains a kettle lake. The lip of the delta occurs at ~71 m asl, which is a minimum estimate for marine limit. This elevation is similar to other Holocene deglacial deltas deposited throughout Dobbin Bay (England 1996). However, The delta has been incised by the modern river, draining from the hanging valley to the south, exposing a section 1.5 km long and ~50 m high. Within this exposure, prominent foreset and topset beds were noted, however; it can be further subdivided into additional stratigraphic units which include outwash deposits, diamict, deformed sand and gravel, and marine silt.

In the outermost part of the delta, a thick bed (~2 m) of fossiliferous silt separates two prominent units of foreset bedding. The silt was originally assumed to be bottomset beds overridden by the prograding delta (England pers. comm. 1996). The silt contains abundant single valves of *Hiatella arctica* and *Mya truncata*. A bulk sample of these shells dated 31,400 ±900 BP (GSC-5663, England 1996). For this study, a single valve (*H. arctica*) collected from the silt was resubmitted for AMS dating, together with a second sample composed of a single valve (*H. arctica*) collected from the overlying foreset beds. These samples dated 36,290 ±440 BP (TO-5865) and 23,260 ±360 BP (TO-5864), respectively. The composition and genesis of the sediments that core the delta, and importance of these radiocarbon dates, will constitute the main focus of this chapter.

Elsewhere, ice-pushed beach ridges occur throughout the 'lower' valley. These ridges were produced when sea ice, driven by the wind, grounded against the shore (Taylor 1978). The highest ice-pushed ridge observed was located on the delta surface at ~74 m asl (Fig.

2.4) and is considered to be a maximum estimate for marine limit. Hence, marine limit is considered to lie somewhere between the elevation of the delta lip (71 m asl) and the uppermost ice-pushed ridge at ~74 m asl.

To the east of the ice-contact delta (Site I, Fig 2.1), an outlier of marine silt and clay separate the easternmost piedmont glacier from an adjacent glacier to the west (Site J, Fig 2.1). The surface of the deposit was surveyed at ~71 m asl. The surface of the deposit is wave-washed which coincides with the lip of the ice-contact delta ~2 km up valley. An exposure along the piedmont glacier indicates that the ridge is comprised of three distinct units. The lowest unit is a massive fossiliferous silt with ~15% rounded dropstones. Overlying this is a clay-rich diamict with a clast content of ~35 % gravel. The uppermost unit of the exposure is a massive non-fossiliferous silt.

The bottom unit in this section is interpreted to be a remnant of marine silt deposited in deep water during a period of widespread faunal activity. The overlying diamict is interpreted to record ice advance over the site, switching from a marine to subglacial environment. This unit is not considered to be an ice proximal debris flow due to the lack of diagnostic features (e.g., flow noses, grading, or gradational lower contacts) (O'Cofaigh 1999). The uppermost unit in the section, represents a return to marine conditions and the transgression of the sea over the deglaciated site. This unit is non-fossiliferous silt possibly recording shallow marine conditions or conditions which are too inhospitable for the entry of marine fauna due to high sedimentation rates or brackish waters. The exposure within the ice contact delta (described later) supports this interpretation with a similar sedimentology of marine silts overlying till which overlies fossiliferous marine silts.

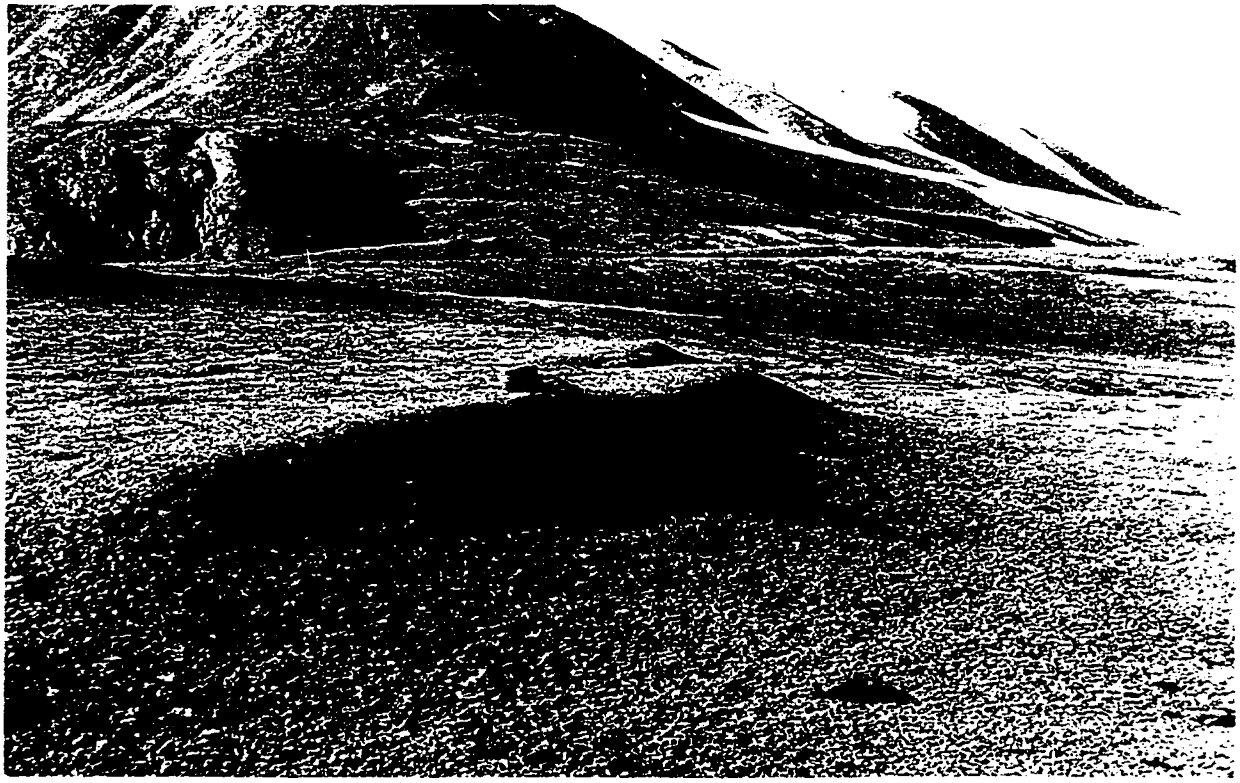

Figure 2.4 Photo of ice-pushed ridge developed in beach shingle superimposed on the outer surface of the ice-contact delta (74 masl). Ridge is ~50 cm high and 10 m long. Photo taken towards the west with hanging valley gorge in background.

Figure 2.5 Photo of regressive shorelines which trim the distal slope of the outer Eugenie moraine. The highest continuous shoreline (thin black arrow) and higher non-continuous shorelines

A delta at 64 m asl (Site K, Fig. 2.1) occurs 1 km west of the 71 m ice-contact delta (Site J, Fig. 2.1). This delta is comprised of fossiliferous silt that extends to its surface where it lacks foreset or topset bedding. Much of the delta appears to have been eroded by a former sandur, hence the 64 m surface is interpreted to lie below marine limit. Shell samples were collected from this silt, however they were not dated due to the lack of a clear relationship to a former sea level.

Within the 'lower' valley, prominent raised shorelines trim the distal-side of the outermost moraine adjacent to Eugenie Glacier (Fig. 2.5). These shorelines extend from ~50 to 4 m asl, and extend with exceptional continuity along the entire length of the moraine (~10 km). Shorelines below 4 m asl have been removed by the contemporary sandur which flanks the south side of the moraine. The significance of these shorelines will be discussed in the following chapter.

2.2 Description of Ice-Contact Delta (Hanging Valley)

The section through the delta at the mouth of the hanging valley is 1.5 km long and is described from its proximal to distal end. The exposure has been arbitrarily divided into three parts (S-1, proximal; S-2, intermediate; S-3, distal), which are described in terms of their stratigraphy and sedimentology. The delta incised by the modern river appears to be typical of former ice-contact deposits (including its spillway and kettle lake. Fig. 2.6). To aid in the interpretation of the delta, each unit within the three parts is assigned a number (U#) beginning at the base of the section.

Figure 2.6 Photo of the large exposure in the ice-contact delta in the lower Dobbin Valley. Sections S-2 and S-3 are marked by number whereas Section S-1 is located to the right, hidden behind foreground. Note the kettle lake (kl) and large sandur (sa). View is to the NE arcoss lower Eugenie Glacier.

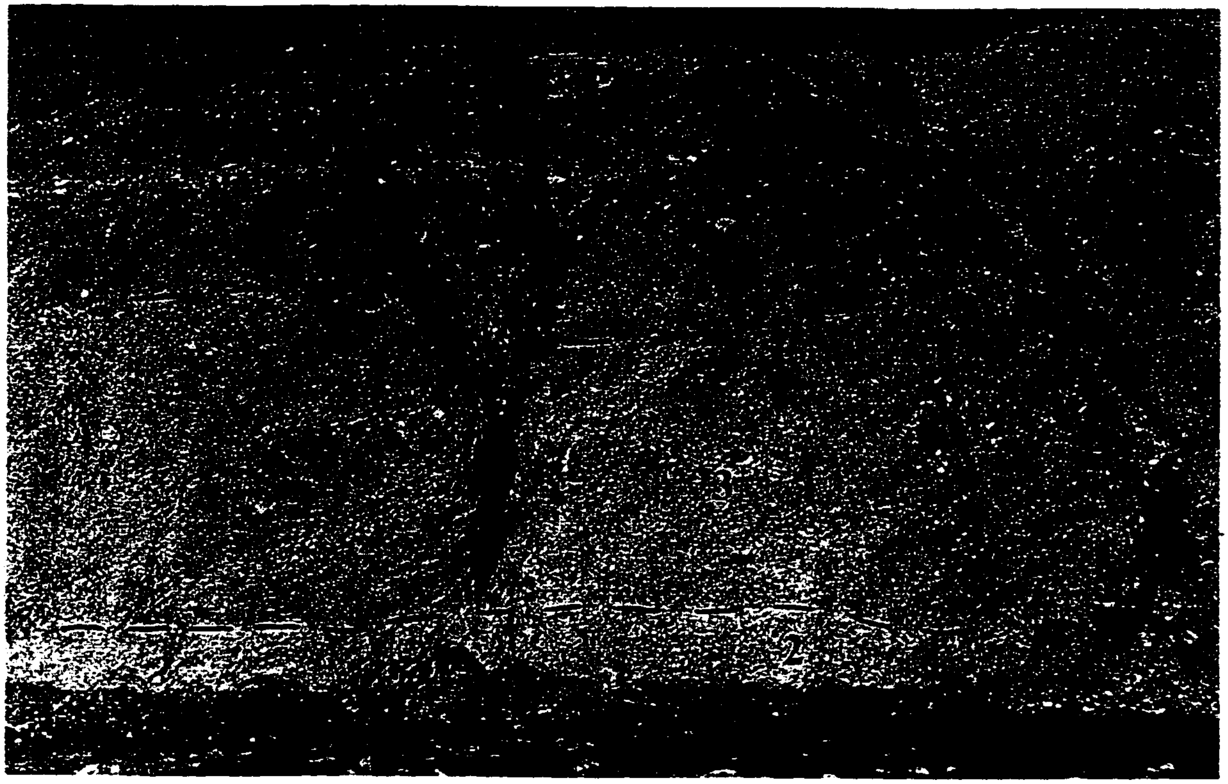

Figure 2.7 Photo of Section S-1. Note the separation of the different units by sharp erosional boundaries. The units described in the text are marked by unit number. The uppermost two units (U8-U9) are not shown.

2.2.1 Description of Site S-1

The proximal part of the exposure is located at the mouth of the canyon exiting the hanging valley (Fig. 2.7). This section (S-1) is subdivided into nine units, which have a total thickness of ~43 m (Fig. 2.8). The lowest unit (2.5 m thick) consists of clast-supported, pebble to cobble gravel which is massive throughout (U1). Overlying this is a indurated, massive silty clay diamict (~3 m thick, U2) containing clasts ranging from boulders (max. ~1 m A-axis) to gravel. Most of the clasts are angular to sub-angular and exhibit both striae and faceting. The next unit is bedded sand and gravel (~11 m thick) with small interbeds of silt and clay (U3). The beds within this unit range in thickness from 5 to 40 cm and have a seaward dip (~30°). There is also evidence of folding of these beds and the matrix is well cemented (likely by calcium carbonate). This unit also contains the occasional large boulder (~30 cm. A-axis). The next unit (U4, Fig. 2.8) is a thin, horizontally-bedded silt, ~1 m thick, containing numerous rounded to sub-rounded clasts (to ~50 cm). This is overlain by 6 m of massive silt containing angular to sub-angular clasts and boulders (U5). With this unit, there is evidence of coarse bedding which exhibits a steep seaward dip. The next overlying unit is gravel with a silt matrix (~1.5 m, U6). The gravel varies from pebbles to cobbles and shows weak horizontal bedding. The next unit (U7) is similar to U5 however; this unit contains a finer matrix. Unit U7 is ~2.5 m thick and dominated by angular to subangular clasts with a silty sand matrix. No visible structures or bedding are evident in this unit. Access to this unit was not possible due to the steepness of the exposure and the induration of the sediment. Hence, it was not possible to determine whether the clasts within this unit were faceted or striated. Overlying this is ~6 m of non-fossiliferous horizontally-bedded silt

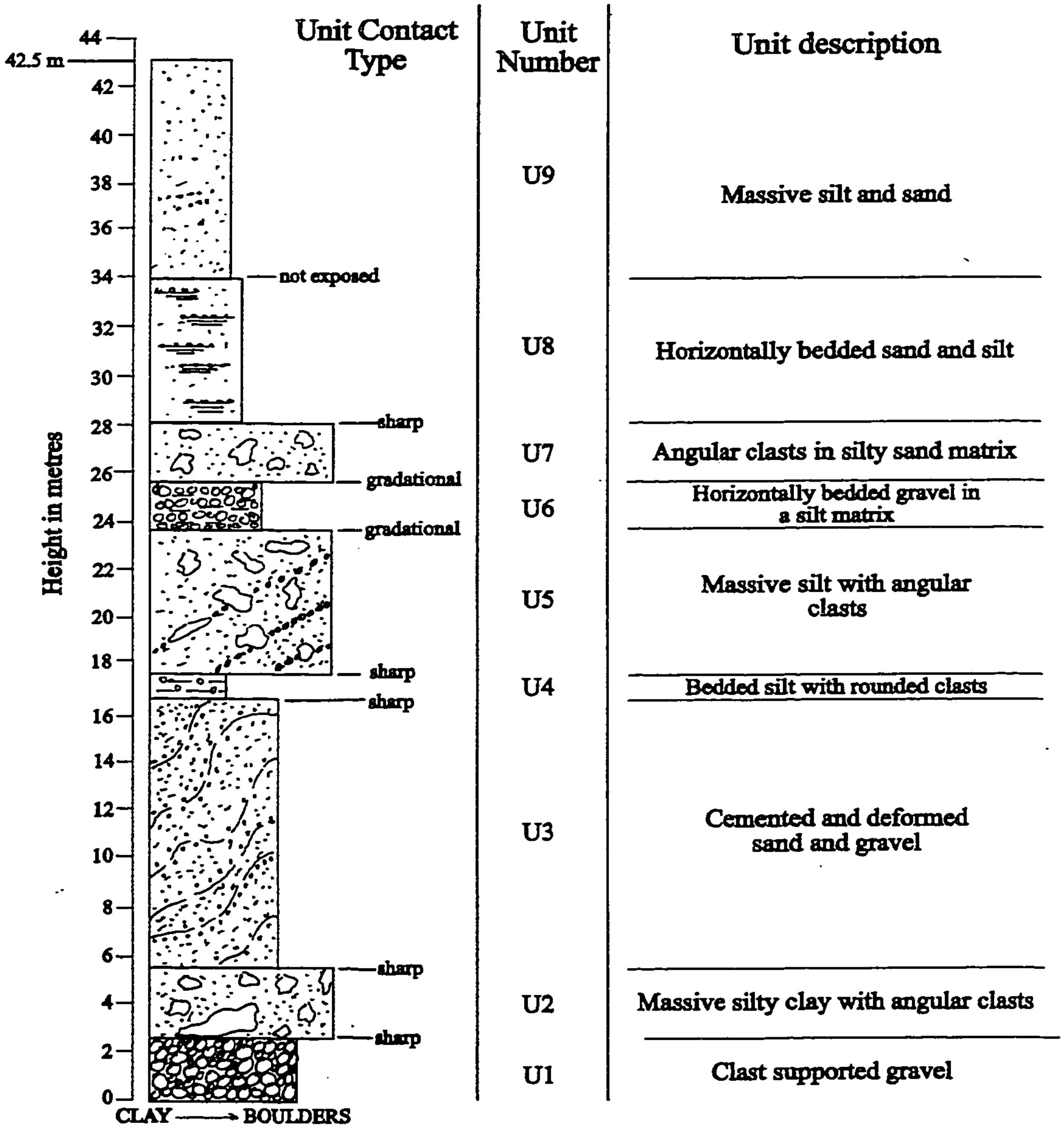

Figure 2.8 Composite log of Section S-1 from ice-contact delta exposure.

and finely bedded sand (U8). The uppermost unit of the exposure is ~9 m of massive silt and sand (U9) that appears gradational (conformable) with the upper boundary of U8. The elevation of the at the top of this section is ~75 m asl (Holocene ml).

2.2.2 Description of Site S-2

The central part of the section is located ~200 m north of S-1 and is ~300 m long and ~39 m high, reaching ~73 m asl. Three logs were constructed at this location (Figs. 2.9 and 2.10). The lowermost unit is composed of a rounded to subrounded gravel ranging from pebbles to cobbles (~2.5 m, U1). This unit is entirely clast-supported. The next unit (U2) is up to ~14 m thick and is predominantly composed of sand and gravel with thin inter-beds of silt and clay throughout. Generally, the dominant clasts are pebbles and cobbles, with occasional boulders. These beds have a seaward dip of ~20° with some beds folded and faulted dipping up to 30°. The next unit is a predominantly massive, rounded to subrounded gravel in a silt matrix that ranges in thickness from 2-2.5 m (U3). Within this unit occasional horizontal bedding of the gravel was noted.

The next overlying unit is a massive diamict composed of sand (~7 m thick) containing a large number of angular to sub-angular boulders (to ~75 cm) and smaller clasts (U4). It was impossible to access this site due to the steepness of the slope and the induration of the sediment. No evidence of faceting or striations were noted. Within the proximal part of Section S-2, this units shows evidence of steep seaward dipping units of stratified sediment. These stratified beds are replaced with the structureless diamict described above, in the outer part of Section S-2. Overlying this, is ~4 m of massive diamict composed of sand containing thin gravel lenses (U5). This unit differed from U4 by the lack

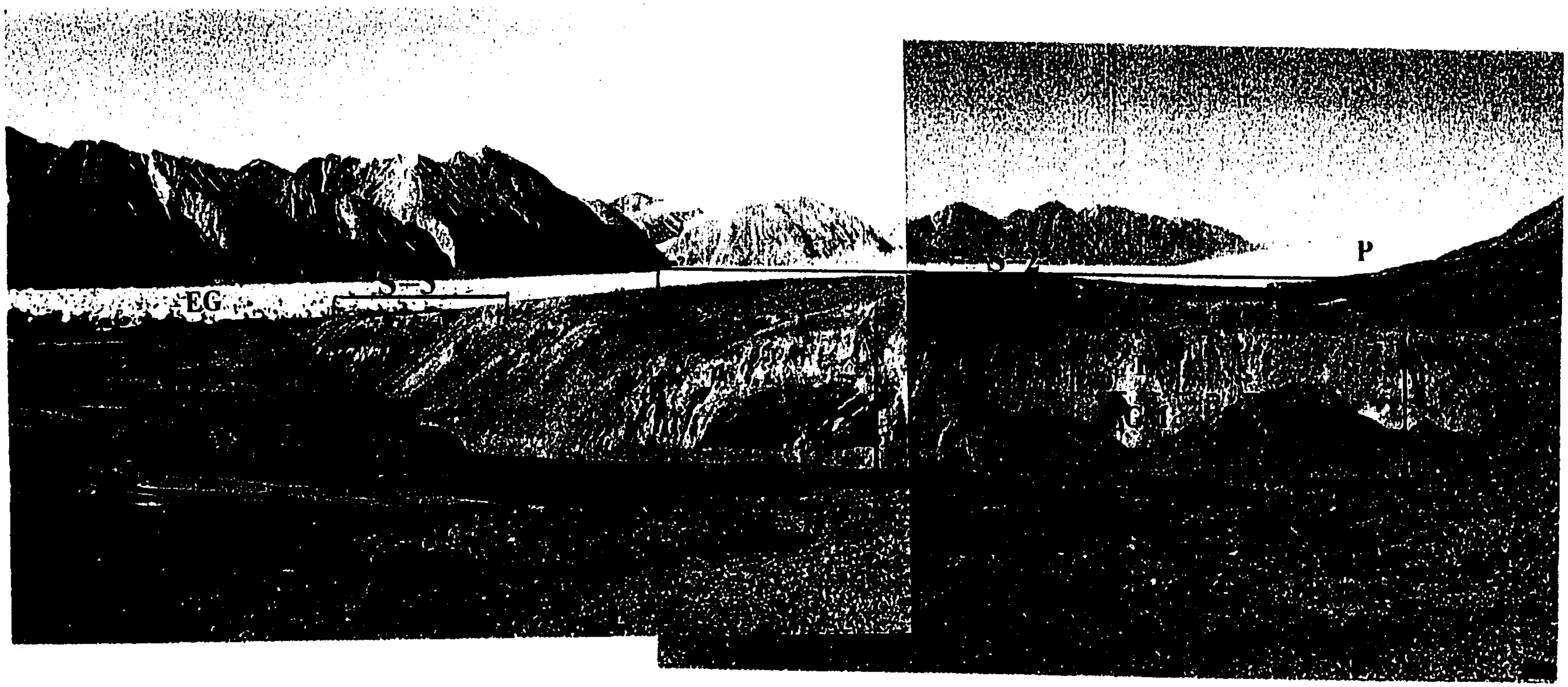

Figure 2.9 Photo of ice-contact delta in the 'lower' valley looking northeast out fiord Sections S-2 (right center) and S-3 (left center) are designated. Section S-1 occurs to the right of the photo. Note the sharp contacts between the different depositional environments: glaciomarine (GM); till (T); and marine (M). In Section S-3 note the light coloured silt bed where lower shells (31.4 ka BP, GSC-5663) were found. Outermost piedmont glacier (P) and Eugenie Glacier (EG) in distant background. Note the coarse sandur that drains from the hanging valley to the south occurs in the center foreground and is undercutting the section.

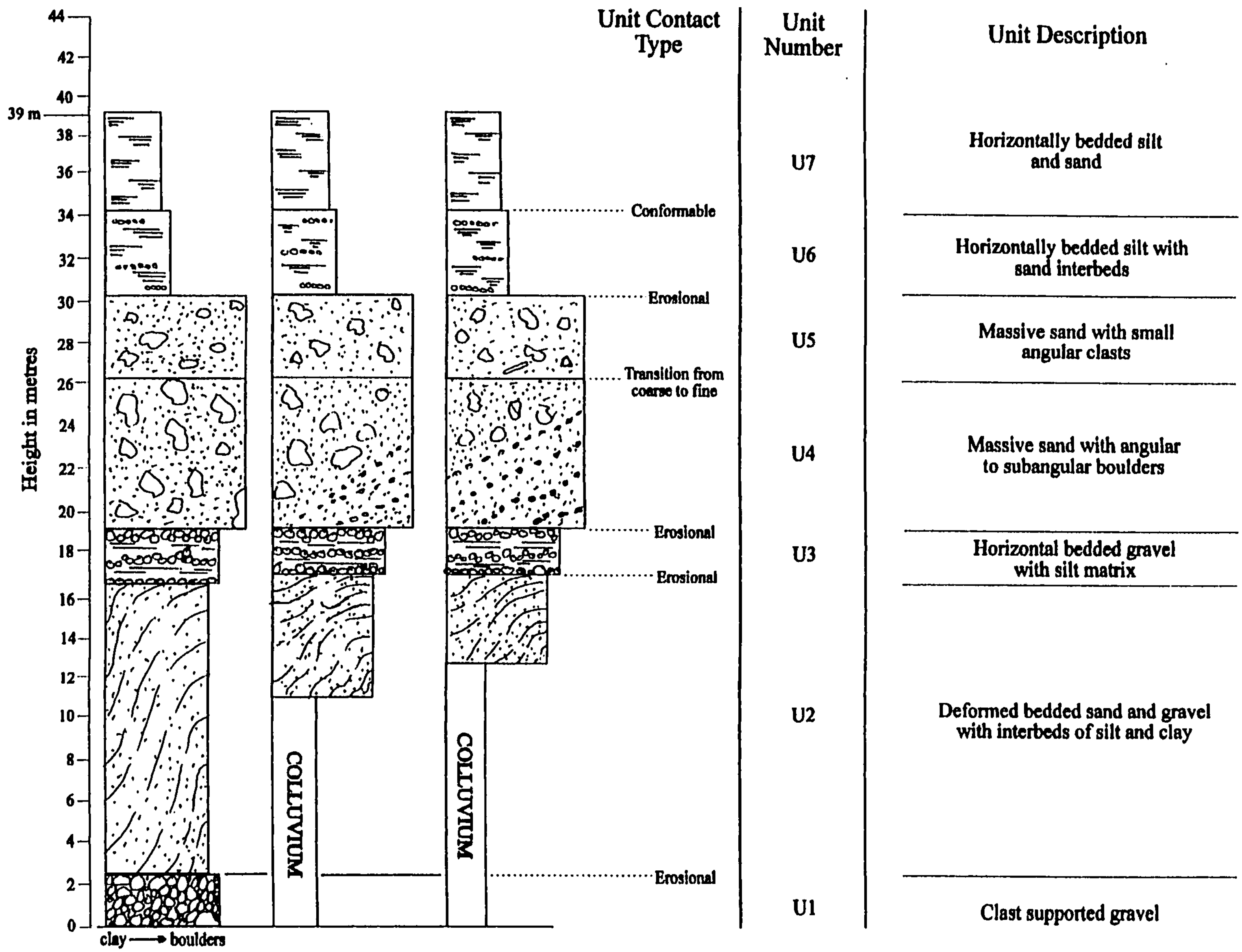

Figure 2.10 Composite log of Section S-2 from ice-contact delta exposure.

large clasts and distinct fining upwards in the sequence. This unit also contains steep bedding in the southern part of the section which is again replaced by a structureless diamict.

The next unit is a horizontally-bedded silt ~4 m thick (U6) containing small interbeds of sand which is succeeded, conformably by a 5 m thick unit of horizontally-bedded silt (U7). This unit lacks the clasts and the thin sand interbeds found below in U6.

2.2.3 Description of Site S-3

The final section is the most distal and is located ~300 m north of section 2. It extends for ~120 m, is ~23 m thick and reaches ~70 m asl (Figure 2.9 and 2.11). The lowest part of the section is obscured by colluvium that extends to unit 1 (U1) at ~42 m asl. This unit is >6.5 m thick and is composed of bedded sand and gravel dipping northward at ~20-30° (U1). This sequence contains normally graded sand and gravel with a maximum grain size of ~3 cm. (A-axis). The overlying unit is a massive fossiliferous silt, ~2 m thick (U2). This is one of the few locations, within the exposure, where abundant marine shells are found. A sample of *H. arctica* was collected from this unit and which AMS dated 36,290 ±440 (TO-5865, Fig. 2.12). The next unit is bedded gravel and silt dipping at ~20° (~5 m thick, U3). A series of distinct fining upwards beds are preserved within the unit, which has a maximum grain size of 12 cm (A-axis). A bivalve of *H. arctica* was collected from U3 and was AMS dated at 23,260 ±360 BP (TO-5864, Fig. 2.13).

The next unit is a massive gravel in a sand matrix, ~2 m thick (U4). This unit also contains numerous large clasts (~30 cm A-axis). The overlying unit is a non-fossiliferous diamict (~3.5 m thick) composed of striated and faceted angular to subangular gravel, containing a wide variety of clast sizes in a silt matrix (U5: Fig. 2.14). The largest of these

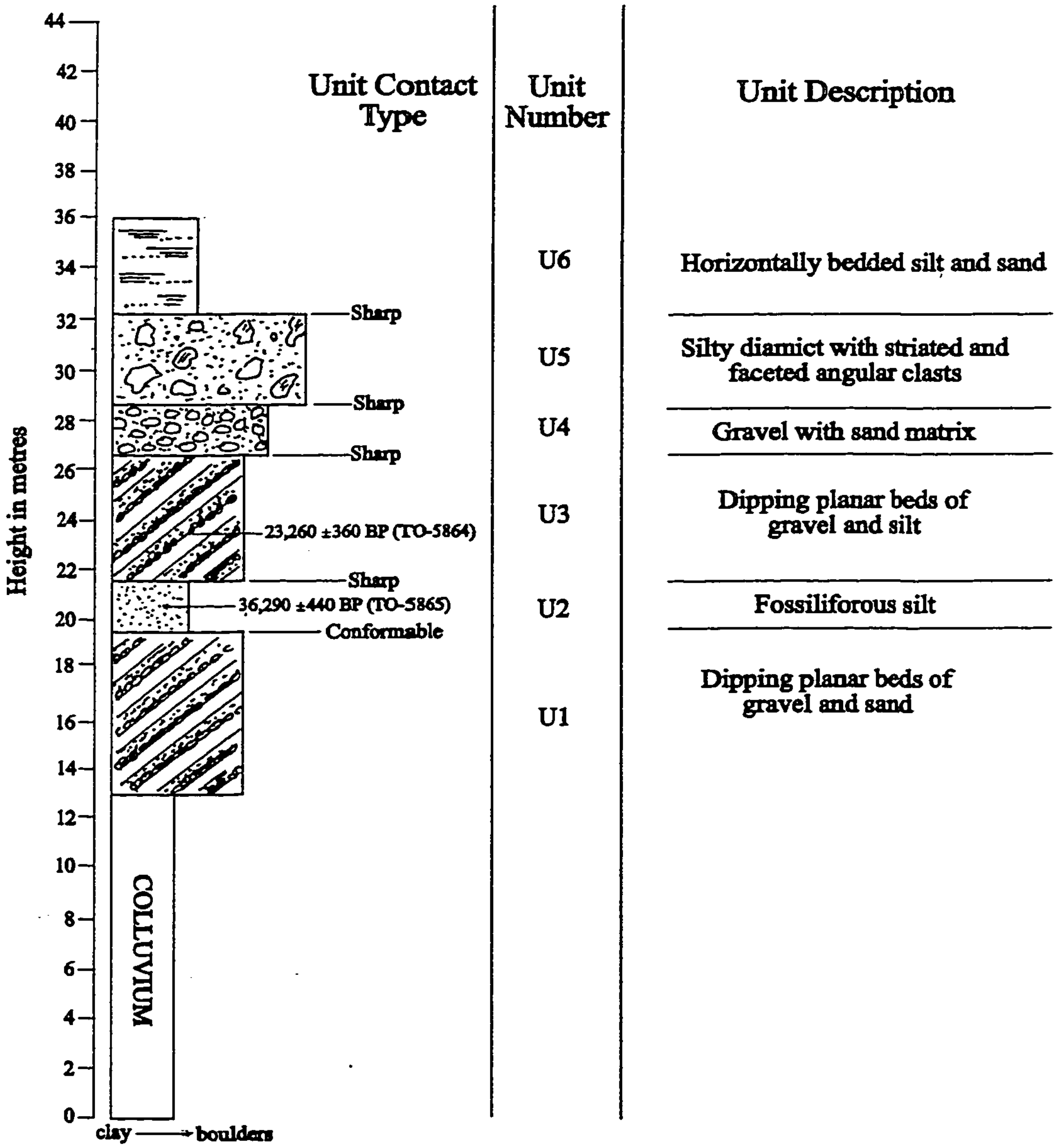

Figure 2.11 Composite log of Section S-3 from ice-contact delta exposure.

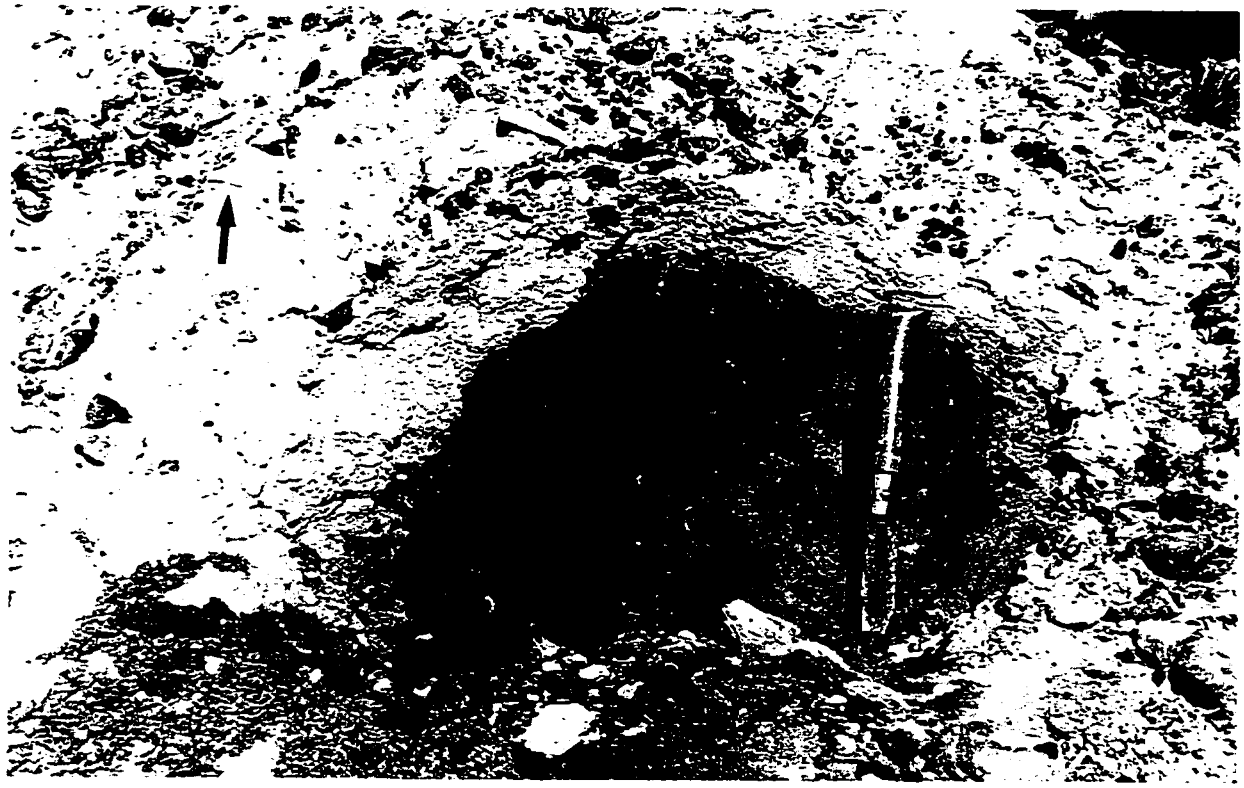

Figure 2.12 Photo of fossiliforuous silt bed within S-3 (U2) of the exposure. Note the presence of shell fragements on the surface and within the sediment (arrows). AMS dated 36,290 ±440 (TO-5865).

Figure 2.13 Photo of *Hiatellica arctica* bivalve found within the upper foreset beds of S-3 (U3). Note that the hinge is still attached within the fine silt matrix. An AMS date of 23,360 ±360 BP (TO-5864) was obtained on these shells..

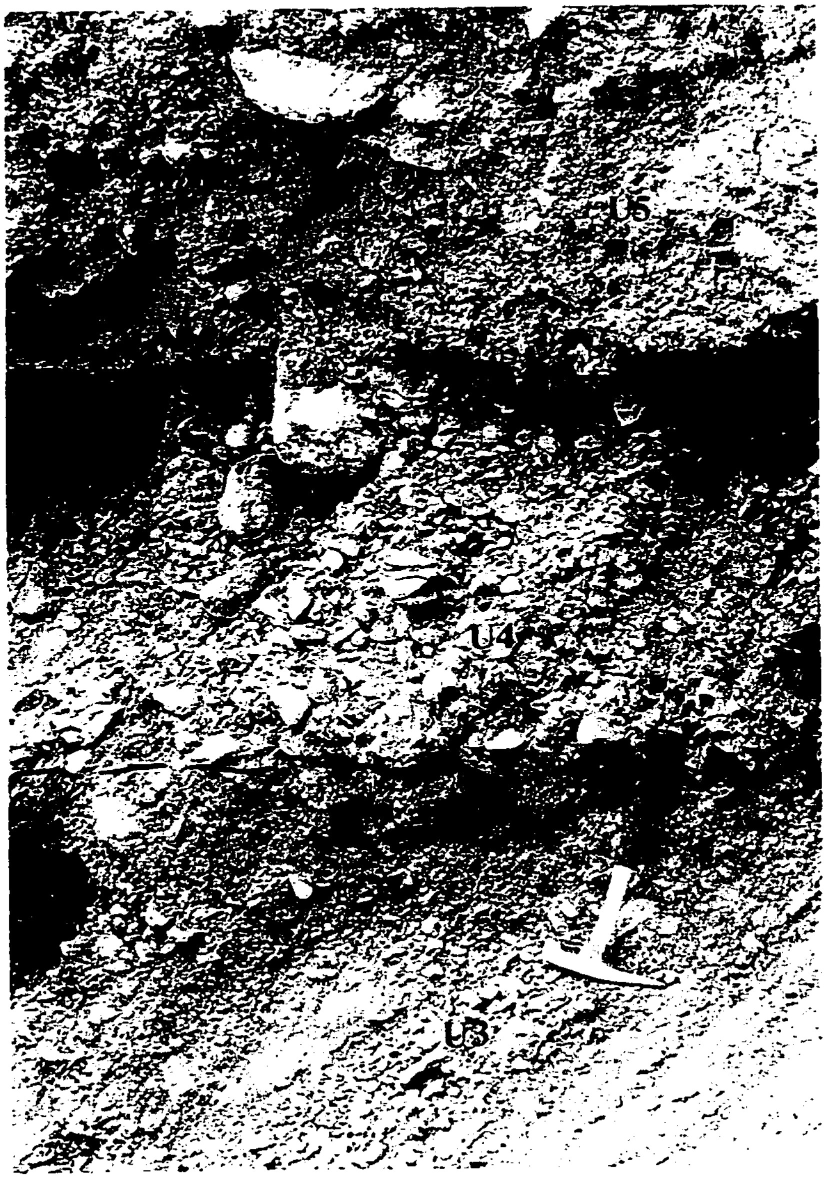

Figure 2.14 Photo of upper units U3-U5 of Section S-3. This photo shows the erosional boundaries between the units. The lower forset unit (U3) is overlain by a matrix supported gravel (U4). These units are inturn capped by till (U5).

clasts was ~20 cm. The uppermost unit is a non-fossiliferous, horizontally-bedded silt and sand, ~4 m thick (U6) with widespread, small rounded clasts.

2.3 Interpretation of Sedimentology

This section proposes depositional processes and environments inferred from the observed stratigraphy and sedimentology. The interpretation of these sections follows the same order of their original presentation (above).

2.3.1 Interpretation of Site S-1

The lowest gravel unit (U1) is interpreted to be a clast-supported glaciofluvial outwash similar to modern sandur in this environment. Unit 2 indicates that the sandur is overlain by a diamict interpreted as till due to its content of striated and faceted clasts. This represents the oldest ice advance recognized within the exposure.

The stratified and seaward-dipping sand and gravel overlying the till, is interpreted to be ice-proximal glaciomarine sediment (U3), likely constituting subaquatic outwash or the foreset beds of a delta. This bedded sediment exhibited a high degree of glaciotectonism and folding. These beds represent either a transgression following deposition of U2 (underlying till) or a transgression that occurred prior to the arrival of ice responsible for U5. The lower boundary of the overlying gravel unit (U4), truncates the underlying deformed sediments (U3). This gravel (U4) is interpreted as outwash from either a subareal or, more likely, a proglacial, subaquatic (marine) setting in front of an advancing ice front (Stewart 1990). This interpretation is based on the fining upwards sequence of sediment within the unit.

The diamict described in units U5-U7 fines upward and is partitioned by a thin gravel. There is no apparent need to separate U5 from U7 other than by a local event (U6) within or along a fluctuating ice margin. Hence, U5-U7 are treated as one glaciation, where U5 constitutes a basal till and U7 represents a supraglacial meltout till. The intervening gravel in U6 is likely subaquatic outwash along the margin of the ice front. This till records the arrival of the ice which was responsible for the deformation of the underlying glaciomarine beds. The two distinct till within Section S-1 are units U2 and combined units U5-U7. The uppermost units (U8 and U9) are interpreted as marine silt deposited during the Holocene transgression which coincides with deglaciation.

2.3.2 Interpretation of Site S-2

The lowest gravel unit (U1) is interpreted as a proglacial outwash plain or sandur. This unit is separated from the overlying unit by a distinct erosional contact. The overlying unit (U2) is a combination of highly tectonized (folded and faulted) ice-proximal marine sediments. The presence of interbedded sand, gravel, silt, and clay, suggests a depositional environment which adjusted to rapidly shifting subglacial effluxes from a nearby ice margin (Figure 2.15). The steep seaward dip of these beds suggest foreset beds of a prograding delta when rsl was >48 m asl. Based on the high degree of tectonization, it is assumed that this deposit was overridden by the ice responsible for this transgression and these overridden deposits may record a temporary pinning point for the ice margin as it proceeded toward Dobbin Bay. The next unit (U3) is interpreted as proximal proglacial outwash from a subglacial or proglacial marine setting. The overlying diamict (U4) is interpreted as a till deposited by the overriding ice. The southern part of the exposure contains steep seaward dipping beds which are interpreted as proximal subglacial effluxes from an advancing grounding line. Farther out the section, this unit is replaced by a dense basal till which

Figure 2.15 Photo of interbedded sand and gravel (sg) with a intermixed bedding of silt and clay (sc) matrix which comprise S-3, U2. This deposition suggests rapidly shifting subglacial effluxes within a proximal proglacial setting.

records ice advance over the site. The next unit (U5) is also a diamict, however; it is distinguished from the underlying till (U4) by its marked reduction in the amount of large clasts and weak bedding. This upper till (U5) is interpreted as a supraglacial meltout till from englacial and supraglacial sources (Figure 2.16). The final two units (U6 and U7) are interpreted as marine silt recording the Holocene marine transgression accompanying deglaciation.

2.3.3 Interpretation of Site S-3

Approximately half of the northern part of the exposure is masked by colluvium. The seaward-dipping sand and gravel (U1) above the colluvium is interpreted to be foreset beds of a prograding delta. Overlying the foreset beds, is fossiliferous marine silt (U2) which dated 36.3 ka BP. This unit is unusual, as it is inconsistent with the normal deltaic sequence deposited into a regressing sea level which tends to coarsen upwards. Rather, this fining upwards sequence is interpreted to record an ongoing transgression or simply a change in the trajectory of the clastic sediment source responsible for the underlying foresets (U1). The latter would simply require a change in the meltwater routing whereas the former would require increasing water depth to produce the finer sediment of U2. The overlying unit (U3) is interpreted as renewed deposition of foreset beds into a relative sea level at least 50 m asl (the upper limit of foreset deposition). A shift in meltwater routing would most easily accommodate this sequence (U1 through U3). The foresets of U3 were dated by a single valve at 23.3 ka BP. The overlying stratified gravel (U4) is interpreted as an ice-proximal proglacial outwash from either a subglacial or subareal setting. Given that the underlying unit (U3) represents deposition into the sea, and that the overlying unit (U5) is a till

Figure 2.16 Detailed photo showing bottom units of Section S-2. Note the highly deformed interbedded sediment (U1) which corresponds to Figure 2.15. This unit is overlain by a proglacial outwash (U2) which forms an erosional boundary upper. The lower units are overlain by a structureless till (U3).

containing striated angular clasts in a compacted matrix, it is assumed that the gravel of U4 represents subaquatic outwash along the advancing ice margin (Stewart 1998). The uppermost silt (U6) records the postglacial marine transgression when the sea invaded to marine limit (~73 m asl).

2.4 Discussion of Sedimentology

The formation of the ice contact delta in the 'lower' Dobbin valley based on the stratigraphic and sedimentological interpretations outlined above (Fig. 2.17). Much of the sediment which cores the delta was deposited along the margin of an northward advancing ice front out of the southern hanging valley. The lowermost units at both S-1 and S-2 record deposition of a glaciofluvial outwash. These units either record the earliest evidence of ice advance into the area or they record an interglacial or interstadial sandur. The extent of this unit north of S-2 is unknown due to thick colluvium mantling the lower half of S-3. At S-1, the outwash (U1) is truncated by till, recording the oldest record of glaciation in the exposure. The arrival of this ice may have been responsible for the transgression recorded by the overlying foreset beds which were deposited during and early stage of ice retreat.

Throughout the delta, the exposure of foreset beds record a period of ice-proximal glaciomarine sedimentation when relative sea level was at least 42 m asl. These sands andgravels were later deformed by the advance of ice over the site which deposited the upper till. The overriding of the foreset beds is recorded in all three sections (S-1 to S-3). These foreset beds are tectonized at sites S-1 and S-2, however; they remain undeformed at site S-3. The boundary between the deformed glaciomarine sediments and the outer foreset beds is located somewhere between S-2 and S-3. Assuming that the foreset at all three sites are

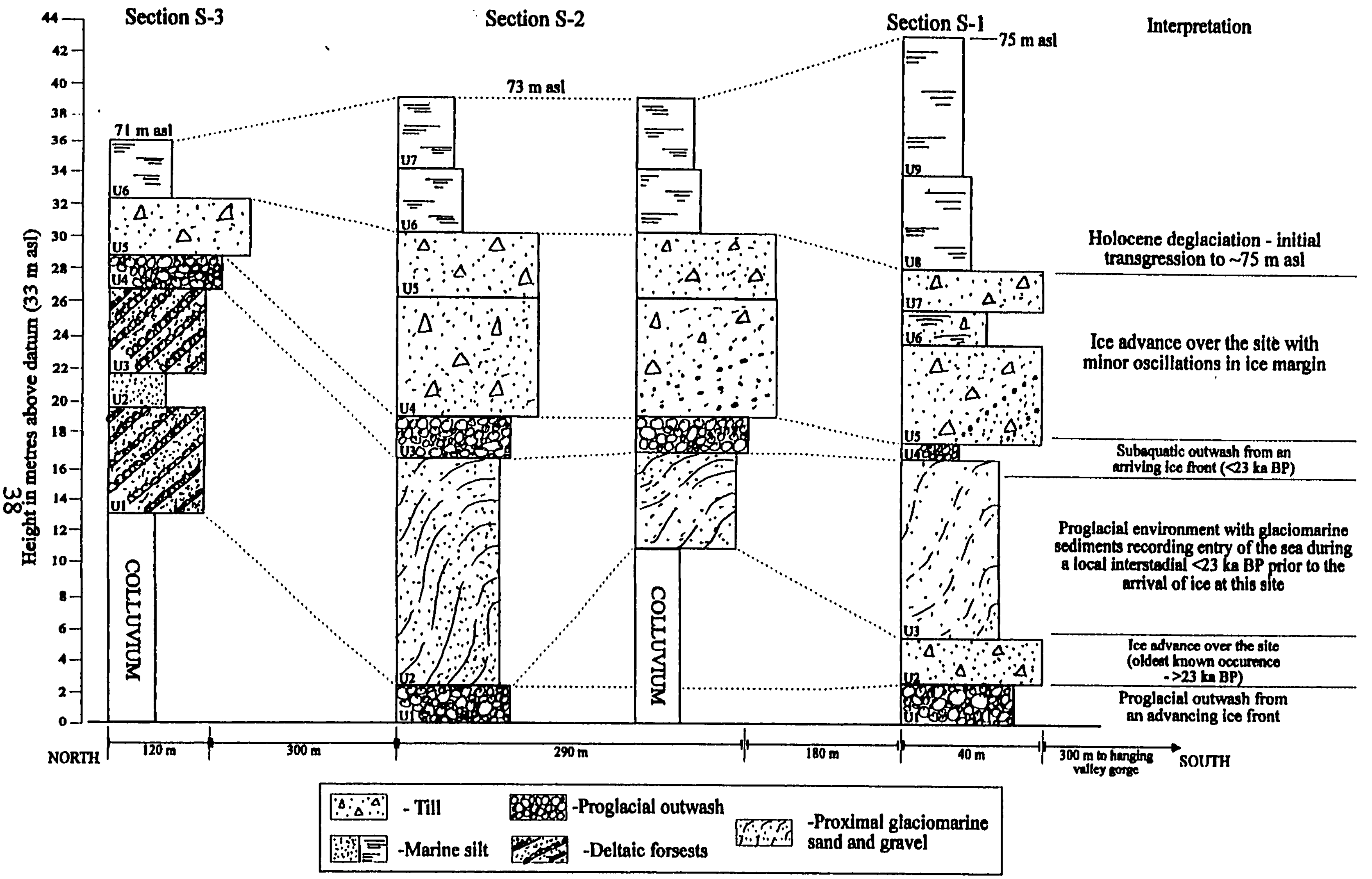

Figure 2.17 Lithofacies reconstruction of ice-contact delta in 'lower' valley.

coeval, then they were deposited sometime after 23.3 ka BP (the age of the youngest valve, TO-5864). This provides an important constraint in the age of the last advance and supports other evidence on Ellesmere island for late ice buildup (Blake 1992a; England 1998).

Separating the two foreset beds at S-3 is a unit of marine silt. This sequence has been interpreted to represent the deposition of a transgressive delta. Elsewhere in the Canadian High Arctic silt beds over gravel have been interpreted to record a transgression (Bednarski 1995). Such evidence is rare, as most of the deltas within the Canadian High Arctic were deposited as a result of sea level regression during deglaciation rather than sea level transgression during ice advance or buildup. Although the fossiliferous silt (U2, site S-3) between the two foreset beds (U1 and U3, site S-3) may also record such a transgression, it is possible that they record a single depositional event within a delta with changing sediment supply trajectories. Within other proximal glaciomarine sediments similar shifts in sediment supply trajectories have been noted (Mackiewicz et al. 1984;Powell 1990;O'Cofaigh 1999).

If units U1 to U3 are coeval, the range in AMS dates (36.3 ka BP in U2 and 23.3 ka BP in U3) are misleading. The original bulk sample collected from unit U2 by J. England in 1993, provided an age of 31,400 ±900 BP (GSC-5663). This date was interesting, as the elevation and position of the delta suggested that it was entirely Holocene. During the current study, I collected a single valve (*H. Arctica*) from this same silt which dated 36.3 ka BP. However, the best estimate on the deposition of units U1-U3 (if coeval) would be the younger date (23.3 ka BP). The preservation of the 23.3 ka BP foresets indicate limited erosion by trunk ice that advanced out the fiord during the LGM. Similar records of preserved older sediments are recorded elsewhere on eastern Ellesmere Island (Blake 1992b;

England 1996). However, this does not explain why the shell date from the silt is so much older (36.3 ka BP) unless some shells were redeposited from a nearby ice margin. The redeposition of shells within this unit is supported by the distinct lack of paired valves. This is also suggested by the 5000 year range in age of the two samples (36.3 ka BP and 31.4 ka BP) within the silt unit (U2). Because this remains uncertain, the alternative interpretation is that units U1 through to U3 represent an interval of glaciomarine deposition spanning 36.3 to 23.3 ka BP. This would imply a prolonged mid Wisconsin ice cover adjacent to this site (producing the required glacioisotatic depression to >45 m asl). This ice margin would then have advanced out Dobbin Bay during the late Wisconsinan (England 1999).

Throughout the entire exposure and truncating all of the previously described foreset beds is proglacial outwash recording an advancing ice front or topset beds associated with the deposition of the underlying foreset beds. Based on the radiocarbon dates discussed above, this advance must be late Wisconsinan. The ensuing arrival of ice is recorded by the deposition of proximal subglacial outwash and till. Although, this till varies both in composition and in thickness but is laterally continuous throughout the exposure (S-1 to S-3). The proximal part of the overlying unit is characterized by steep seaward dipping beds which may record the advance of a grounding line over the site. In the central part of Section S-2, these steep beds are replaced with a structureless till of both basal and supraglacial origins.

Capping all of the lower sediments within the exposure is a thick deposit of marine silt. This unit represents the last transgression of the sea to its Holocene limit following deglaciation. The age of this sediment is undetermined, however; deglacial deltas at a similar elevation in inner Dobbin Bay dated ~7 to 7.5 ka BP (England 1996).

The radiocarbon dates from the foreset and silt units at S-3 constrain the last ice advance at the head of Dobbin Bay to <23.3 ka BP. Dates on deglacial deltas in outer Dobbin Bay indicate that most of the fiord was ice free by ~7 ka BP (England 1996). These dates suggest that the late Wisconsin ice advance lasted for a duration of less than 16 ka within Dobbin Bay proper. This late initiation of ice advance in Dobbin Bay supports other studies which have found similar evidence for late ice buildup (<20 ka BP) along eastern Ellesmere Island (Blake 1992a; England 1998, 1999).

On Spitsbergen, late ice buildup was also recorded by Mangerud et al (1992). Mangerud et al. found sub-till marine silt which dated 36,100 ±800 (T-5211), providing a maximum age for the last glaciation. Based on amino acid ratios, Mangerud et al. (1992) concluded that the last buildup of ice on Spitsbergen occurred rapidly sometime after 36 ka BP, and possibly after 25 ka BP, lasting <10,000 years.

2.5 'Upper' Dobbin Valley

In terms of geomorphology, the 'upper' valley varies dramatically from the 'lower' valley. The open terrain of the 'lower' valley is replaced by steep, confined terrain encircling a valley ~11 km long and 2 km wide. Here, cliffs of the Dobbin Bay syncline are ~1000 m high (GSC Map 1358A 1972). The western limit of the 'upper' valley is blocked by the snout of a valley glacier, approximately 7 km west of Eugenie Glacier. The eastern boundary of the 'upper' valley is arbitrary, and extends from the western boundary of the 'lower' valley into the mountainous terrain mentioned above (Fig 2.18). Along the north wall of the 'upper'

Figure 2.18 Airphoto of 'upper' Dobbin valley (National Airphoto Library A-16613-83, 1959). Those landforms described in the text are labelled with corresponding letters on the photo. Also of note is Eugenie Glacier (EG), the modern sandur (MS), Piedmont glacier #1(PG1), and valley head piedmont glaciers (VHG). Again this photo depicts only the outer Eugenie moraine.


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valley, several piedmont glaciers descend from the local mountains onto the valley floor. The easternmost piedmont lobe (#1, Fig. 2.18) is located ~3 km west of Eugenie Glacier whereas the others occur 3 km farther west (Fig. 2.18).

2.5.1 Terrestrial Landforms

The most prominent glacial landforms in the upper valley are moraines found adjacent to the margins of piedmont glacier #1 and Eugenie Glacier. Within the upper valley, the inner moraine abutting Eugenie Glacier is intermittent, whereas the outer moraine, located ~600 m beyond the ice margin, is exceptionally developed (Site A, Fig. 2.18). The terminal moraine along the margin of piedmont glacier #1 (Site B, Figs. 2.18 and 2.19) is similar in both morphology and sedimentology to the outer lateral moraine flanking Eugenie Glacier. The crest of this moraine ranges from ~44 to 52 m asl., and reaches a maximum of 500 m in width. Exposures through the moraine indicate that it is composed primarily of highly deformed bedded sand and silt of marine origin. These sediments exhibit large and small scale folding and faulting. This deformation will be described in greater detail in the following chapter. Like the outer moraine of Eugenie Glacier, the moraine flanking piedmont lobe #1 contains prominent and well-preserved raised beaches on its distal side.

A series of rock glaciers extend several kilometres along the base of the northern wall (Site C, Figs. 2.18 and 2.20). Elevations recorded on the outer lips of the rock glaciers varied from 92 m (immediately east of the piedmont lobe #1) to 80 m asl. The westernmost rock glacier, located on the west side of piedmont lobe #1, has been undercut by a stream exposing foliated glacier ice overlain by angular boulders (Figure 2.21). This mantle of boulders is the result of supraglacial meltout (forming ablation till) as well as rockfall from

Figure 2.19 Photo of the terminal moraine which flanks the piedmont lobe #1 ('upper' valley). This moraine extends for ~1.5 km around the margin of the glacier and is similar in terms of composition and sedimentology to that of the outer Eugenie moraine. 47 m asl delta located on western corner (arrow).

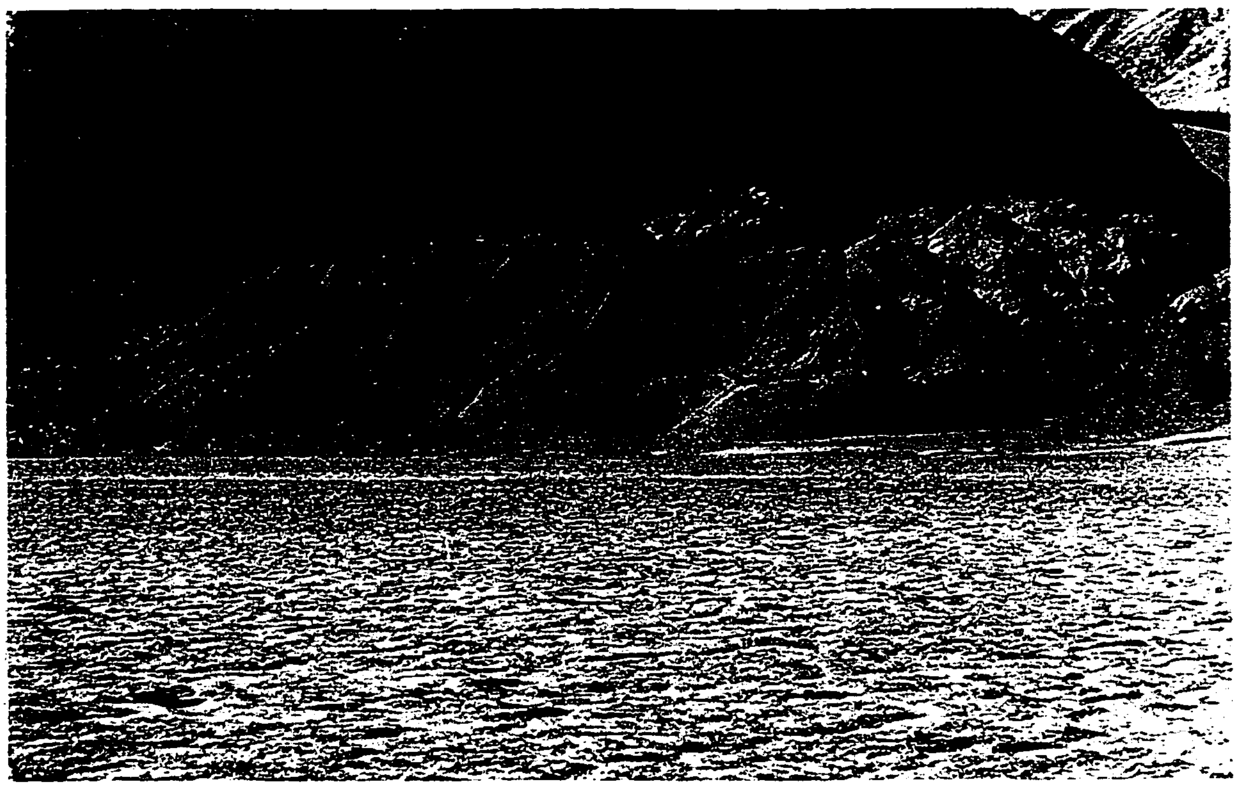


Figure 2.20 Photo taken facing north of the easternmost three rock glaciers which occupy the northern wall of the 'upper' valley (~1 km west of Eugenie Glacier. The lip of the rock glaciers ranged from 92 m (single arrow) to 80 m asl (double arrow). The surface was composed of argular boulders which rested at the angle of repose.

the adjacent slope onto the surface of the glacier when it was ~ 0.5 km beyond its present margin. It is assumed that the remaining rock glaciers are similarly cored with glacier ice which has been preserved and remobilized by a carapace of talus.

On the floor of the 'upper' valley, a rockfall occurs below a conspicuous scar of unoxidized bedrock approximately 750 m asl on the north bedrock wall. The rockfall deposited dozens of large boulders (up to ~350 m^3) distributed in a lobate pattern which extends ~1.5 km from the base of the north wall and covers a width of ~1 km (Site D, Figs. 2.18 and 2.22). The timing of this event is unknown, however it likely occurred prior to sea level regression (from the adjacent valley) due to the strongly confined lobate pattern of the rockfall which suggests entrapment by a shallow marine embayment (below 28 m asl). The age of this relative sea level is ~4.6 ka BP (see later section on raised shorelines).

Along the southern wall of the 'upper' valley, a large meltwater channel cut into shale descends to ~75 m asl (Site E, Figs. 2.18 and 2.23). Below this elevation no evidence of meltwater incision was found, likely indicating the elevation of marine limit which constituted the local base level during meltwater incision. At present, a misfit stream flows through the meltwater channel, and onto a fan deposited on the adjacent lowland (Site F, Figs. 2.18 and 2.23).

2.5.2 Marine Landforms

Marine limit in the upper valley is marked by a bench composed of silt that accumulated to ~75 m asl (Site G, Fig. 2.18, Fig. 2.23). This is accordant with the lower limit of the meltwater channel as well as with wave limits found throughout the upper valley. Marine silt containing *H. arctica*, was found along the southern wall of the upper valley at

Figure 2.21 The internal conposition and structure of the westernmost rock glacier (~3 km west of Eugenie Glacier). Note the foliated glacier ice capped by both fine silt and coarse angular fragments. It is assumed that the remaining rock glaciers along the northern wall are also ice-cored.

Figure 2.22 Photograph of large rockfall which covers a portion of the 'upper' valley floor. Note the conspicuous scar located ~750 m asl. The lobate pattern of the rockfall covers ~1.5 km from the northern wall. An outermost block (1) is ~350 m^3 (arrow).

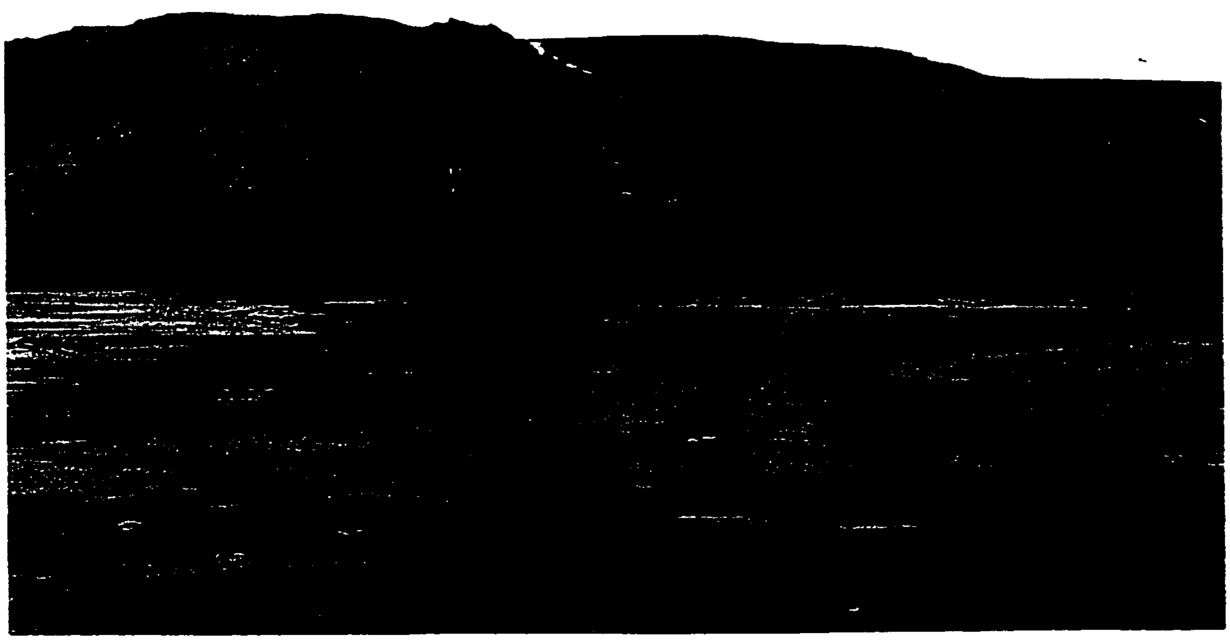


Figure 2.23 Photo of the southern wall of the 'upper' valley (View taken from the north). This photo shows the large meltwater channel (black arrow) which drains through the bedrock ridges. As well, the silt bench (open arrow) which marks marine limit in the 'upper valley' and a large fan (F) are also recognizable.

46 m asl (Site H, Fig 2.18). This site is located downslope from the 75 m bench. Because the shells represent the highest fossiliferous silt observed in the upper valley they would likely provide the best minimum age estimate for marine limit. Although many of the shells were fragmented, a few bivalves were found in growth position. The sample remains undated. The similarity of marine limit (~71 m asl) in both the upper and lower valleys indicates rapid deglaciation of the entire field area inland of Dobbin Bay.

Along the northwestern part of the moraine flanking piedmont glacier #1, an ice-contact delta occurs at ~47 m asl (Site I, Figs. 2.18 and 2.24). Incision by the adjacent river has exposed well-defined foreset and topset beds, however no datable material was found. The lip of the delta is accordant with the highest shoreline on the piedmont moraine. This ice margin was attained when relative sea level (47 m asl.) was well below marine limit (~71 m asl.), hence it records either a mid Holocene readvance or a stillstand at this position, close to the modern ice margin. On the moraine, well preserved shoreline extend to 31 m asl.

The most prominent marine landforms in the upper valley are the preserved shorelines which range from ~8 to 45 m asl. Along the lower slopes of the northern wall, the shorelines are intermittently preserved. For example, along the outer slope of the large rock glaciers, shorelines are preserved up to ~30 m asl. In other areas the shorelines are difficult to identify. A more detailed discussion of the shorelines is included in the next chapter.

2.6 Conclusions

The interpretation given above suggests that the marine limit for the Dobbin Bay lowland was ~71 m asl. This elevation also closely corresponds to marine limit reported from outer Dobbin Bay (79 - 83 m asl; England 1996). The highest marine limit (83 m asl)

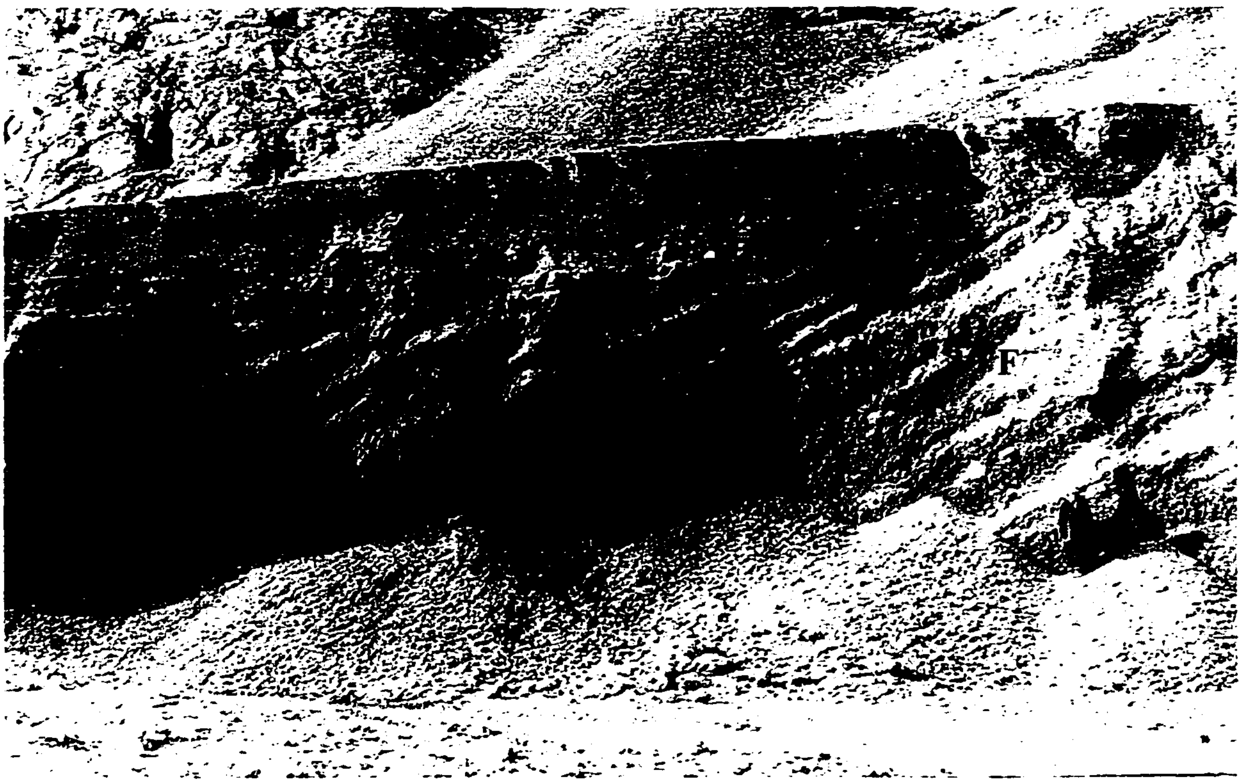


Figure 2.24 Photo of the internal structure of an ice-contact delta on the northwest corner of the piedmont glacier #1 moraine. Photo taken facing north. Note the distinct foreset (F) and topset (T) bedding typical of a prograding delta. Delta was examined, however; no material was found for radiocarbon dating. The delta surface was ~ 47 m asl. Person for scale is 1.5 m.

at the eastern end of Dobbin Bay was dated 6770 ±110 BP (GSC-5654), however; this is likely a minimum age for this relative sea level because accordant dates of 7.3 ka BP were obtained on a similar marine limit 20 km to the north in Scoresby Bay (England 1996) and on Darling Peninsula (Gualtieri and England 1998).

The ice-contact delta described in the 'lower' valley is important for understanding the overall history of late Wisconsinan ice buildup along eastern Ellesmere Island. The preservation of glaciomarine sediments beneath late Wisconsinan till within the delta indicates that ice advance postdates 23.3 ka BP. Dates on deglacial Holocene deltas throughout Dobbin Bay indicate that the bay was ice free by ≤7.3 ka BP and a similar age likely applies to marine limit at the fiord head where marine limit was ≥71- ≤74 m asl. North of Dobbin Bay, at the head of John Richardson Bay, a similar Holocene marine limit was found. This marine limit (67 m asl) dated 6.7 ka BP (England 1996). These three fiord heads were ice free by ~7.3 to 6.7 ka BP which confines the late Wisconsin glaciation to an interval of approximately 16 ka (<23 to ~7 ka BP).

Chapter 3

3.0 Introduction

Along the margin of Eugenie Glacier, two lateral moraines of exceptional length extend inland from the sea and terminate along the north wall of the 'upper' valley. Within the 'upper' valley, a smaller terminal moraine abuts piedmont glacier #1. The composition and stratigraphy of the moraines indicate their mode of formation whereas prominent shorelines along the outer Eugenie Moraine record the gradient of differential postglacial emergence. The shoreline profiles are compared with regional reconstructions of differential postglacial emergence across Ellesmere Island (England and O'Cofaigh 1998; O'Cofaigh 1999).

3.1 Dobbin Bay Moraines

Within the Canadian High Arctic, moraines are generally not well developed or widespread, and rarely extend for more than a few hundred metres. Most of the moraines are formed as a result of glacial transport or rockfall to a stable margin where deposition occurs, however; two other moraines also occur in the Dobbin Bay field area: 'ice-thrust' and 'ice pushed' moraines.

'Ice-thrust' moraines form due to the bulldozing of pre-existing frozen sediment (Kalin 1971; Evans and England 1991; Bennett et al. 1998). This process involves the movement of large blocks of consolidated sediment by thrust-faulting in a proglacial setting. In the High Arctic, this commonly involves glaciofluvial or glaciomarine sediments occupied by permafrost (Boulton 1986). In western Canada, thrusting commonly involves sedimentary bedrock (Fenton 1987). On Axel Heiberg and Ellesmere islands, approximately twenty 'ice-thrust' moraines have been mapped (Evans and England 1991), however; their occurrence is widespread.

'Ice-pushed' moraines involve the deformation (bulldozing) of unconsolidated sediment. This also occurs in a proglacial setting, where unfrozen sediment is overrun and deforms under the pressure of the advancing ice front (Fig. 3.1). The principle differences between ice-thrust and ice-pushed moraines lies in the degree of deformation of the displaced sediment and whether its redeposition includes incorporation of glacier ice or segregated ground ice which can subsequently melt. In ice-thrust moraines, the structure of the pre-existing frozen sediment is commonly recognizable and commonly with the incorporation of ice blocks. In ice pushed moraines, the original sediments are amalgamated or deformed due to their lower internal strength. If ice-push occurs in a subaquatic marine setting, the salinity of the water would precluded the incorporation of glacier ice as its buoyancy would cause the ice to float.

3.1.1 Inner Eugenie Moraine

The inner moraine along Eugenie Glacier marks its contemporary margin (Fig. 3.2). The moraine is characterized by a single crest ~10-30 m high, which extends for ~6 km. Moraines of this length are rare in the High Arctic. The moraine is composed of both diamicton and blocks of recognizably bedded silt and sand. Throughout much of its length, the moraine is composed of silt and sand, however; coarse angular boulders dominate at some sites. At many locations where the moraine is composed of imbricated blocks of silt and sand, it is clear that large blocks of sediment (1-2 m^2) were thrust into the ridge. These blocks were distinguishable from the surrounding sandy matrix, however; they contained no

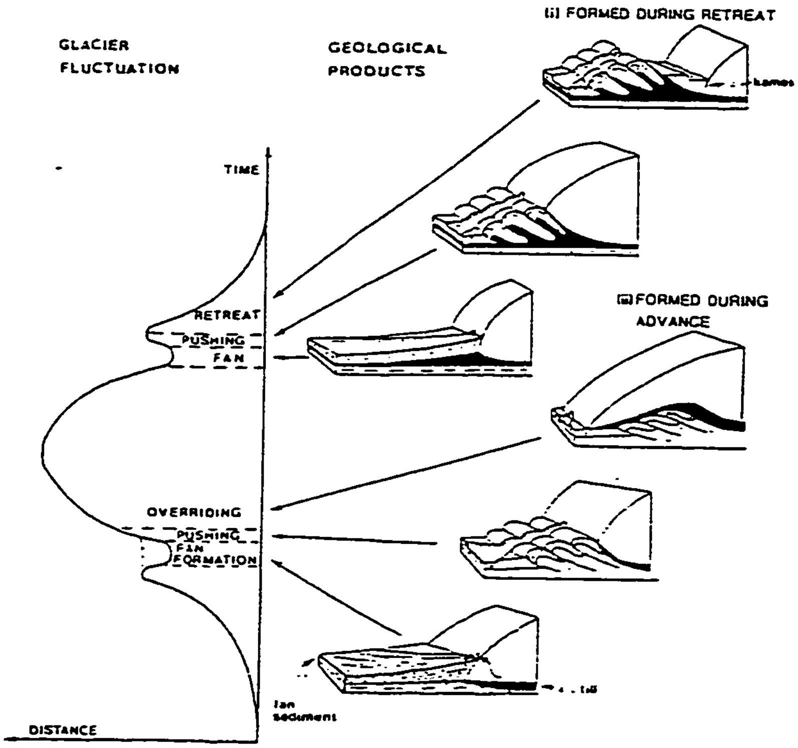


Figure 3.1 Formation of 'ice-push' moraines under varying glaciological conditions. Note the folding of sediment away from the advancing ice front. Also note the extension of folded sediment under the ice. (Modified from Boulton 1986).

Figure 3.2 Photograph showing the inner lateral moraine (arrow) of Eugenie Glacier. This moraine is actively forming within 30 m of the margin. Foreground of the photo is the melt-out topography described within the text. View to the east, parallel to moraine axis.

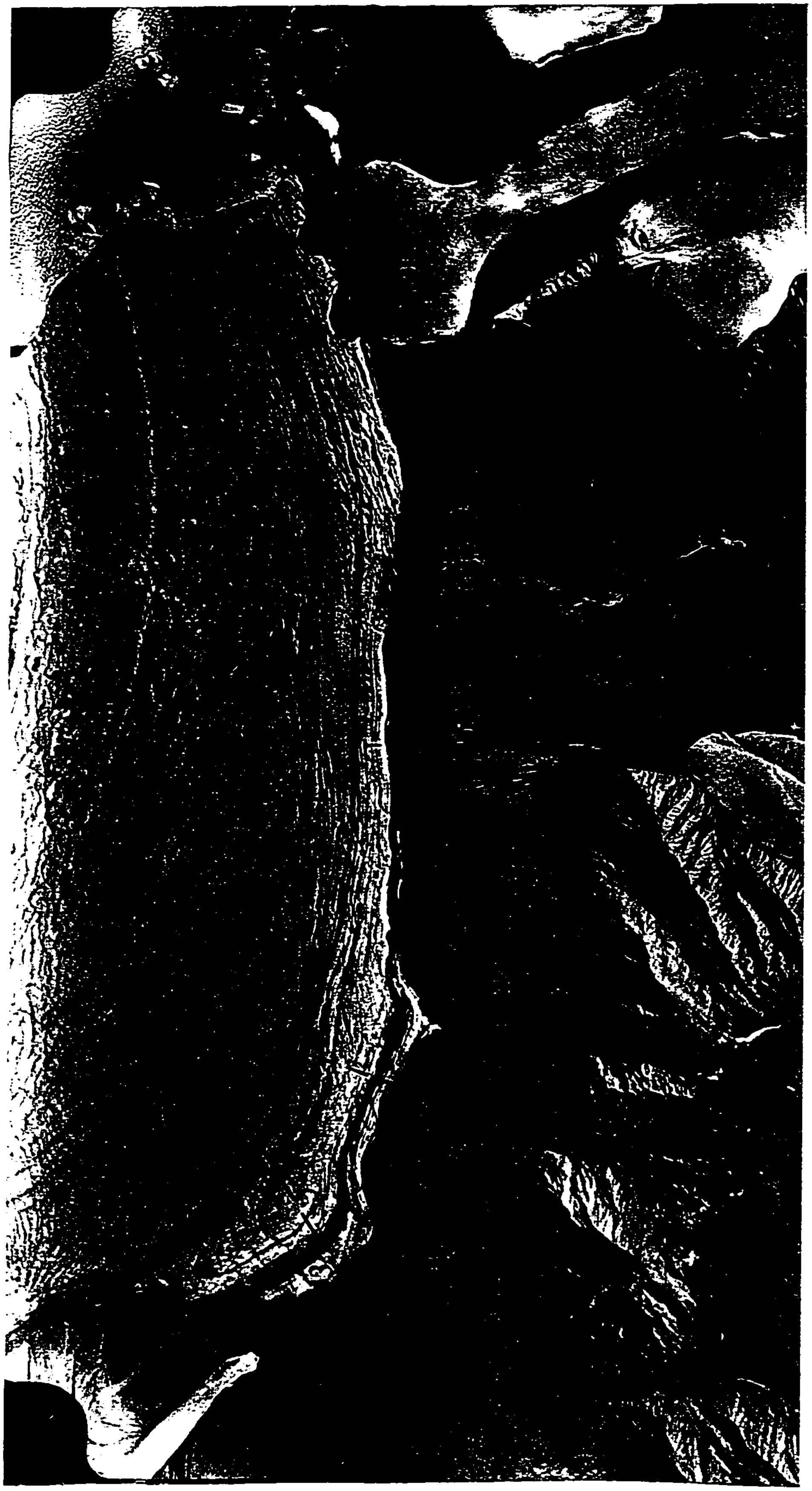
internal structure themselves. Small exposures at the base of the moraine indicate that buried glacier ice is common within the moraine as well. The diamicton within this moraine is structureless and is characterized by a sand matrix with angular boulders showing no internal structure. Where the moraine is dominated by silt, complete valves (Hiatellica arctica) and shell fragments are abundant. Only one shell sample was radiocarbon dated which provides a useful measure of the maximum age for moraine formation. This sample was taken from within ice-thrust silt at ~38 m asl and dated 3460 ±70 BP (TO-4189; England 1996). The implications of this date will be address later within this chapter.

3.1.2 Outer Eugenie Moraine

The outer moraine is separated from the inner moraine by >100 m, and across this distance meltout is producing a hummocky terrain with widespread ponds. The outer moraine is approximately 10 km long and up to 0.5 km wide (Fig. 3.3). Its height ranges from 30-50 m asl (its summit). The outer moraine is dominated by a single crest, however; in some locations it is characterized by wide benches enclosing smaller, secondary ridges. The moraine is narrowest at its eastern (seaward) end, with maximum width attained in its center from which it tapers again to its western end.

The eastern part of the moraine has been breached by the modern river, however; an aerial photograph taken in 1959 shows the moraine to be continuous from the sea to its inland limit. Subsequently, the main river to Dobbin Bay has breached the moraine from its distal side cutting through approximately 250-400 m of sediment, ~30 m high. The timing of this incision may coincide with the drainage of a formerly ice-dammed lake within the hanging valley due south of the breached site. Based on the topographic map of Dobbin Bay,

Figure 3.3 Airphoto of Eugenie Glacier and large lateral moraine which flanks its southern margin. Inner moraine postdates the airphoto. Landforms indicated on photo include: Ice-contact delta (ICD); Outer piedmont Glacier (OP) and; Piedmont glacier #1 (PG1). Also depicted on figure are the positions of three transects which provide cross sections of the moraine complex geomorphology (----- #).



this lake drained sometime after 1959, its last noted occurrence. The outlet channel for this lake drains through a deep gorge and onto a large gravel and boulder sandur just above and aligned with the location of the moraine breach. The dramatic increase in river drainage from the northward flowing channel would have approached perpendicular to the moraine, causing effective erosion. Alternatively, the progradation of the sandur due to lake drainage into the main river may have resulted in increased erosion by the now deflected main stream and the subsequent moraine breach. Either way, the hydrologic history of the hanging valley seems to be a contributing factor to the recent migration of the main river.

The distal side of the moraine immediately east of the breach shows little sign of post-depositional erosion. The proximal side has undergone significant erosion due to breaching by the river which now exposes buried glacier ice between the moraine and present ice margin. This ice does not core the moraine, rather it is the remnant sole during a period of glacial retreat. The moraine is composed of a massive silty clay matrix with >40 % coarse fragments. The clasts were dominated by local lithologies, however limestone and a few localized, small granite erratics were also noted.

The central part of the outer moraine extends from the river breach westward until the moraine begins to bend sharply to the north. This part of the moraine is ~500 m wide at its base and reaches ~50 m asl. The apex of the moraine has multiple crests and is ~100 m wide. Although steep, the distal slope exhibits minimal disturbance by mass wastage and periglacial processes. The proximal slope is presently steep due to widespread modification by meltout occasioned by recent ice retreat to the north. Exposures in the central part of the moraine indicate that the sediment is predominantly sand and silt. The exposures also show that some of the sediment is horizontally-bedded whereas other parts exhibit highly deformed bedding, including both large and small scale folds (Figs. 3.4 and 3.5). Despite these exposures, very few marine shells were found within these beds. Clast content was generally <5% within the exposures sampled, however; on the surface of the moraine a lag gravel is widespread likely due to prolonged deflation.

The western end of the moraine has a single crest ~400 m wide and 25 m high, reaching ~45 m asl. Again the distal slope is steep but remains undisturbed as indicated by well defined shorelines. In contrast, the proximal slope is steep and has experienced widespread post depositional disturbance (slumping). Exposures within this part of the moraine indicate that the sediment is dominated by silt with less than 10% coarse fragments. Exposures also show intense deformation of bedded silt that was likely subhorizontal during its initial deposition. This silt is of marine origin and contains localized occurrence of marine shells.

3.2 Description of Moraine Complex Geomorphology

This section will describe the margin of Eugenie Glacier where three cross-sections were constructed to aid in the description of the moraines (Fig.3.3). The eastern (#1) and central (#2) transects run south to north starting from the main river and extending to Eugenie Glacier. The western transect (#3) was measured from the lowland bordering the outer moraine and extends across the surface exposed by the greatest post 1959 ice retreat. The cross sections were constructed by measuring the gradient and ground distance in short intervals over the moraine complex. As well, sedimentological data was collected to complement this work.

Figure 3.4 Photograph of large scale folding (arrow) of marine sediment within the outer Eugenie moraine. The person in the photo is 1.5 m tall. View to the east.

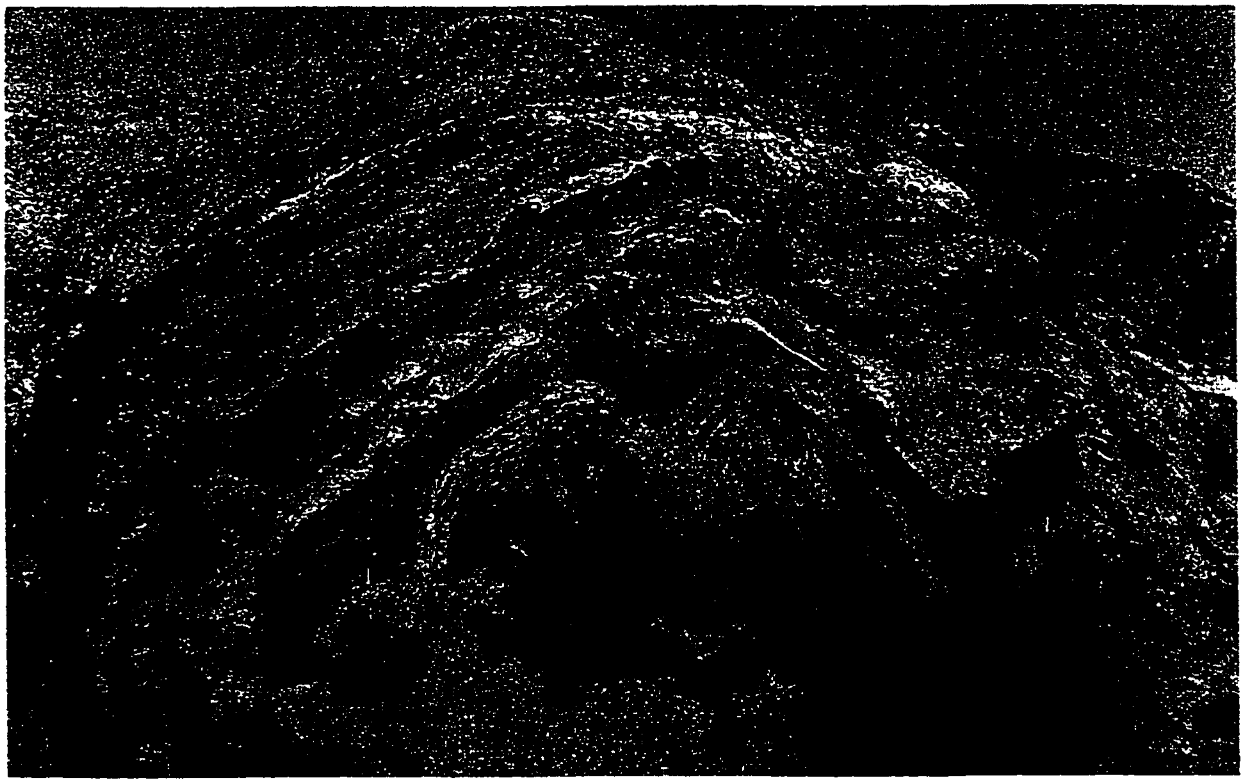


Figure 3.5 Photograph of folded marine sediment. Scale not given however this exposure is located adjacent to Fig. 3.4. Photo covers an area of ~15 m in height.

Transect #1 is located ~1.5 km west of the river breech and is marked by a large erratic (~10 m high) located on the distal slope of the outer moraine (Figs. 3.6 and 3.7). The distal slope has remained unchanged since areal photography. The proximal side of the moraine is marked by an ~8 m scarp. Between the outer and inner moraines, dead-ice topography extends for 110 m (Fig. 3.13). This area is occupied by numerous small hummocks surrounded by small ponds. The sediment is dominated by silt with isolated marine shells on the surface. It extends to the inner moraine which is ~40 m wide, and 10 m high. The remaining 25 m of the transect, between the inner moraine and Eugenie Glacier, is occupied by a small proglacial pond.

The second transect is located near the widest part of the outer moraine. This area is marked by a change in the moraine's axis from E-W to NW-SE (Figs. 3.8 and 3.9). Again, the distal slope (~350 m long) is predominantly unmodified as indicated by the continuity and preservation of the raised shorelines. This transect indicates that the crest of the outer moraine is ~100 m wide. The proximal slope of the outer moraine is 18 m high and ~100 m long. The next 150 m of the transect is again occupied by a plain of hummocks and ponds. Numerous small exposures indicate that the area is underlain by buried glacial ice. The final 75 m of the transect crosses the inner moraine. At this location, the inner moraine has a relief of ~7 m and is characterized by a single ridge (~15 m wide) comprised of blocks of marine silt.

The third transect covers a distance of 800 m (Figs. 3.10 and 3.11). From the base to the apex of the outer moraine, the distal slope is 27 m high and ~180 m long. This slope is uniform except for a small bench (~10 m wide) that occupies the lower slope. This bench

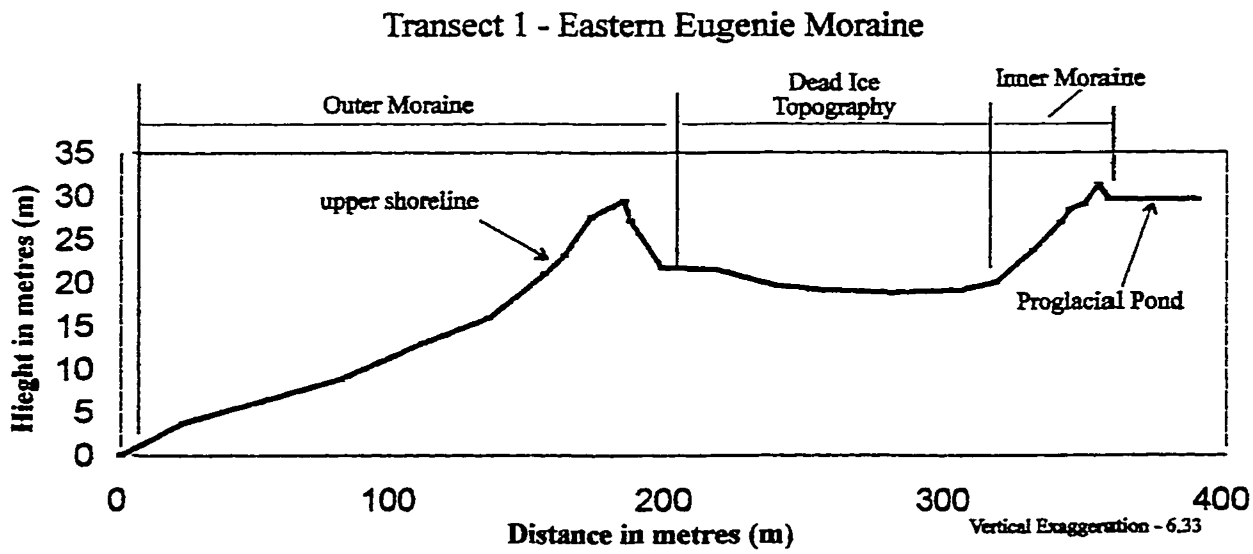


Figure 3.6 Cross section of moraine complex along Transect #1.

Figure 3.7 Photograph showing the position of Transect #1 (---). The inner (thick arrow) and outer (open arrow) moraines are clearly defined in this photo. The uppermost continuous shoreline on distal slope (thin arrow). View to north across Eugenie Glacier.

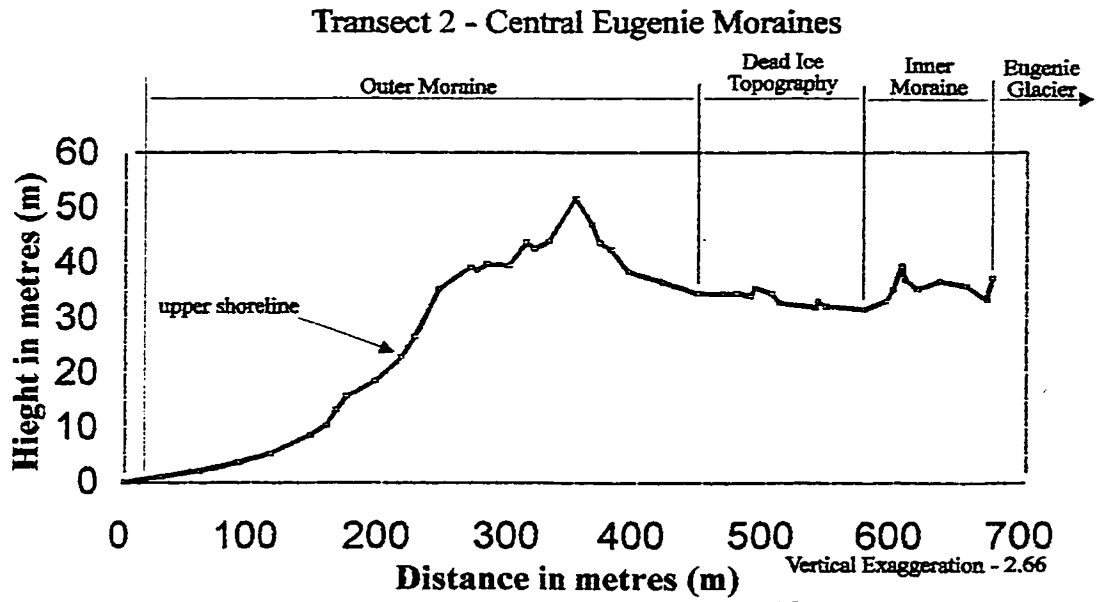


Figure 3.8 Cross section of moraine complex along Transect #2.

Figure 3.9 Photograph of the position of Transect #2 (—). The inner (thick arrow) and outer (open arrow) moraines are clearly defined. This transect has a significantly larger dead ice zone (thick arrow) than the last and is evident on the photo. View to north across Eugenie Glacier.

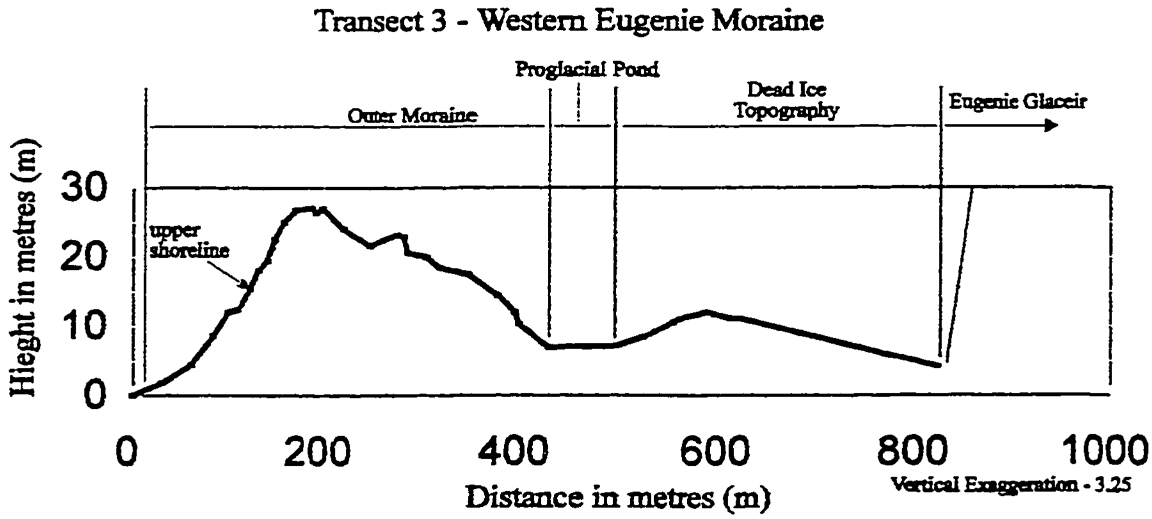


Figure 3.10 Cross section of moraine complex along Transect #3.

Figure 3.11 Photograph showing Transect #3 (---). The outer moraine is large in this transect. The inner moraine is either very small or discontinuous. Note the regressive shorelines which trim the distal slope of the outer moraine (arrow).

extends for ~100 m and is likely part of the shoreline system above and below it. The proximal slope of the outer moraine is 20 m high and 220 m long. At the base of the slope is a small proglacial pond (~75 m wide). The remaining 300 m of the transect is characterized by gently sloping marine silt underlain by buried glacier ice. Close to the Eugenie Glacier (~200 m distant) this plain displays several smaller ridges (~50 cm high) which parallel the ice front. It is believed that these ridges are forming as a result of pressure exerted by Eugenie Glacier which is again advancing (D. Evans, personal communication 1995). This terrain is frozen (to an unknown depth) which would facilitate transfer of this pressure. The inner moraine is absent from this transect, but is found in other location along the western side of the glacier. On average, the inner moraine north of Transect #3 has a maximum relief of ~4 m.

3.3 Mode of Deposition and Age of Eugenie Lateral Moraines

The composition and structure of the inner moraine indicates that it was formed by ice-thrusting. The presence of large blocks of undeformed sediment suggest ice advance into a frozen setting. Compressive stress from the advancing ice resulted in the thrusting and superposition of sediment blocks which partly form the inner moraine. Blocks of glacier ice are also incorporated into the inner Eugenie moraine. The inner moraine is <40 years old (based on its absence on the 1959 airphoto) which limits the amount of permafrost which is able to aggrade into the sediment. The composition of the moraine suggests that the thrust blocks of silt are ~2 m thick, likely recording the thickness of permafrost during formation.

The outer lateral moraine along the margin of Eugenie Glacier contains highly deformed fine-grained marine sediment. The scale and widespread nature of this deformation indicates that the sediment was unfrozen during the advance of the ice. Pressure from the advancing ice forced the sediment to fold and deform during emplacement of the outer moraine. The fact that the sediment was unfrozen is indicated by the subsequent formation of shorelines along its entire length (indicating that it originated from below sea level) and by the exceptional preservation of these shorelines subsequently which precludes melt-out of buried glacier ice (which would not be expected in a subaquatic marine setting).

When Eugenie Glacier advanced to deposit the outer Eugenie moraine, the head of Dobbin Bay was occupied by the sea where fossiliferous marine silt blanketed the submerged lowland. During this advance of Eugenie Glacier across the lowland, these unfrozen sediments were bulldozed subaquatically into a ridge close to or slightly above the sea level at that time (≥50 m asl). This is required by the fact that regressive shorelines trim the moraine from ~50 m asl. Furthermore, the surface of the moraine, to its highest point, has apparently been wave-washed and contains scattered shell fragments on its surface. The shorelines are assumed to be of marine versus lacustrine origin because they extend continuously, and are closely-spaced, from the moraine crest to within a few metres of present sea level. The continuity of these shorelines is more in keeping with progressive postglacial emergence than it is with the intermittent and catastrophic lowering of an ice-dammed proglacial lake. Furthermore the measured age of the uppermost continuous shoreline (mid-Holocene) is consistent with the corresponding mid-Holocene sea level in this area. For example, on the distal side of the outer moraine, a driftwood sample was located on a well-developed shoreline tread at ~33 m asl (Fig. 3.12). This sample dated 4600 ±70 BP (Beta-

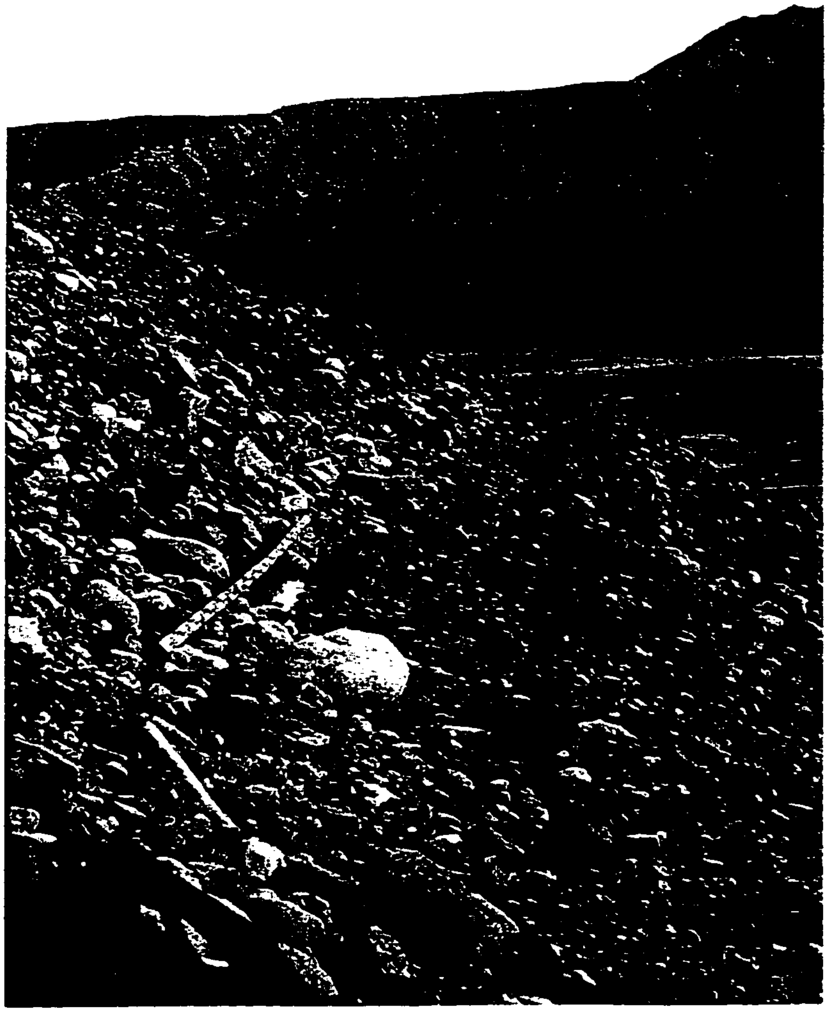


Figure 3.12 Photograph of driftwood sample (arrow) found on the highest continuous shoreline. The sample was found resting on the surface of the shoreline in an undisturbed state. Sample is 65 cm long. Also note the boulder lag which marks the shoreline. View to the south across Dobbin lowland.

91862). This sample provides a minimum date for the deposition of the moraine which clearly predates even higher relative sea levels (to ~50 m asl.).

Aerial photographs of this area indicate that Eugenie Glacier was still in contact with the outer lateral moraine in 1959 (Fig. 3.3). Subsequently, Eugenie Glacier has retreated up to 1 km to the north. The greatest amount of retreat has occurred along the central and western parts of the moraine. In these locations, the margin has retreated ~900-1000 m from its position on the airphoto. At the eastern (seaward) end of the moraine, the margin has retreated only ~50 -100 m from its airphoto location. Exposures in the area between the inner and outer moraines show localized plains of buried, foliated glacier ice capped by a thin (<1 m) layer of silt and fine sand (Fig 3.13). Numerous large proglacial ponds also occupy this area, the largest of which is ~500 m in diameter.

The rapid retreat of Eugenie Glacier can be attributed to the status of ice contacting the outer Eugenie moraine evident on the 1959 airphoto (Fig 3.14). The airphoto indicates that the outer southwest margin Eugenie Glacier has been severed from its former ice source above the bedrock constriction immediately upstream from this site (Fig. 3.14). Given the severance of this ice mass from its original upvalley source, it was clearly poised to undergo rapid retreat since this stagnant part of the glacier lies entirely below the Equilibrium Line Altitude (ELA, locally ~1000 m asl, Miller et al. 1975). The presence of stagnant ice at the outer moraine on the 1959 airphoto may record a preceding surge which caused thinning of the glacier through the bedrock constriction.

It is noteworthy that the contemporary margin of Eugenie Glacier which is readvancing, coincides with the westernmost flowline which extends from the lowland

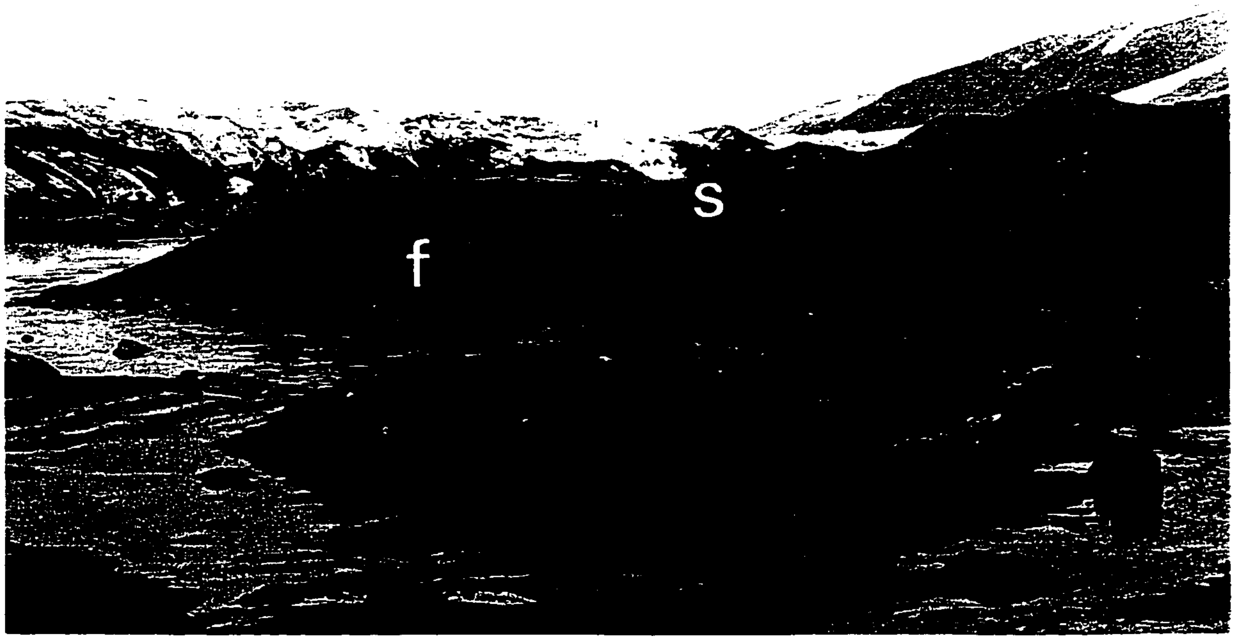
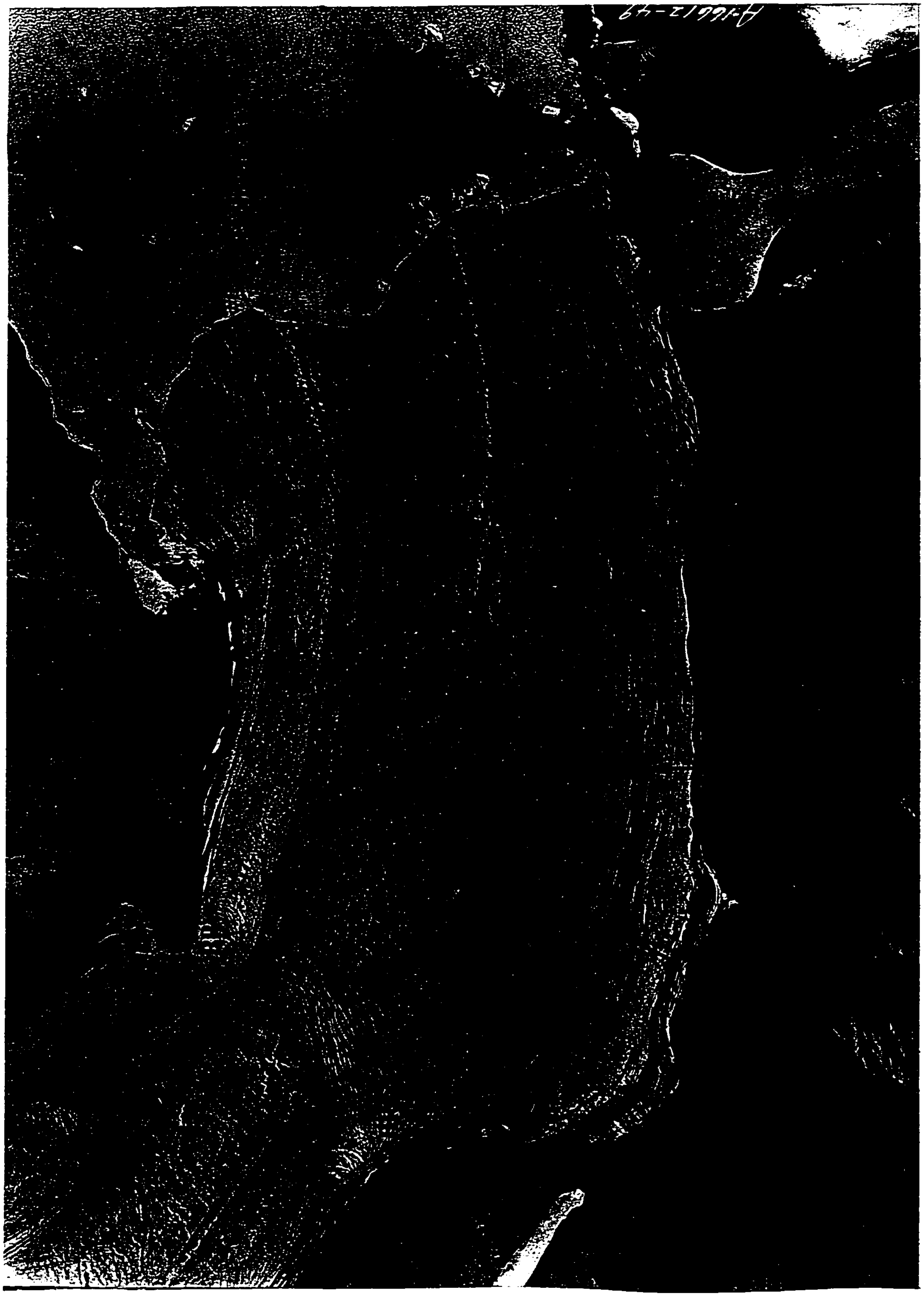


Figure 3.13 Photograph showing the buried glacial ice within the inner and outer lateral moraines. A thin layer of silt and sand (S) cap the foliated ice (F) which is presently being exposed and melting. The depth of this ice is unknown. View to east along the axis of the moraine.

Figure 3.14 Airphoto of recent Eugenie Glacier ice margin fluctuation. Present day margin indicated by dashed line (----). Area between dashed line and outer moraine is presently ice free. The severence of ice from its source as the glacier thinned through a bedrock constriction (BR) caused the stagnation of a large part of the glacier, which subsequently retreated.



through the bedrock constriction to an active supply of ice upglacier (on the 1959 airphoto). Evidence of this readvance is recorded by the entrainment of brecciated blocks of glacier ice at the base of Eugenie Glacier where it is re-encroaching on terrain exposed by the recent (post 1959) retreat (Benn and Evans 1998).

The inner moraine which marking the contemporary ice margin is comprised of large blocks of sediment which exhibit primary bedding that have been thrust away from the ice. This 'ice-thrust' moraine is formed by the faulting of frozen sediment due to the compressive force exerted by advancing ice. Given their different environments of formation, the inner moraine differs from the outer moraine in its sedimentology and geomorphological stability. The outer moraine was formed in a marine embayment from unfrozen, unconsolidated sediment and was never ice-cored. The inner moraine is forming in a terrestrial setting occupied by permafrost and includes considerable buried glacial ice within the thrust blocks.

3.4 Mid-Holocene to Modern Ice Margins

Since the mid-Holocene, Eugenie Glacier has experienced more ice marginal fluctuations than was previously recognized. As discussed earlier in this chapter, the outer moraine is >4.6 ka BP and likely ≥5.5 ka BP (based on the associated relative sea level of ≥50 m). Furthermore, the 1959 airphoto indicates that Eugenie Glacier was abutting this moraine. Since 1959 the ice margin has retreated up to ~1 km and is presently readvancing at some sites.

The initial interpretation of the mid-Holocene glacial history of Dobbin Bay envisioned that Eugenie Glacier advanced to the position of the outer moraine and *remained there* until sometime after 1959, when it began to retreat. The inner moraine represents the contemporary ice margin with a corresponding age of <40 years (i.e. sometime after aerial photography). This interpretation, however; requires revision because an AMS date on a marine shell collected from the ice-thrust silt within the inner moraine dated 3460 ±70 BP (TO-4189, England 1996). This sample postdates the formation of the outer moraine (>4.6 ka BP), indicating that the sea must have inundated the lowland some distance north (inside) of the inner moraine sometime after deposition of the outer moraine (mid-Holocene). The distance the ice margin retreated at this time (3.5 ka BP) is unknown. The date of 3.5 ka BP is a minimum estimate on ice retreat following the deposition of the outer Eugenie moraine. Additional AMS dates on shells from the inner Eugenie moraine would test this chronology (i.e. what are the oldest and youngest shells found within the moraine). The duration of this retreat is also unknown and additional AMS dates would refine the timing of marine occupation. Furthermore, the shell dated 3.5 ka BP must have been overridden by Eugenie Glacier when it readvanced to the outer moraine as shown on the 1959 airphoto. The 3.5 ka BP shell date from the inner moraine is clearly a maximum age for its formation because it is a recent landform that does not appear on the 1959 airphoto.

This sequence of ice advance and retreat along the lower south side of Eugenie Glacier is reconstructed on a time distance diagram of this margin (Fig. 3.15). This reconstruction represents the marginal fluctuations required by the present database, additional dating could reveal other adjustments not yet identified. Evans and England (1992) also report mid to late Holocene readvacnes on NW Ellesmere Island, followed by recent ice retreat. Blake (1981) has provided evidence for a readvance at ~1 ka BP in Makinson Inlet. The age of the readvance to the position shown on the 1959 airphoto is also

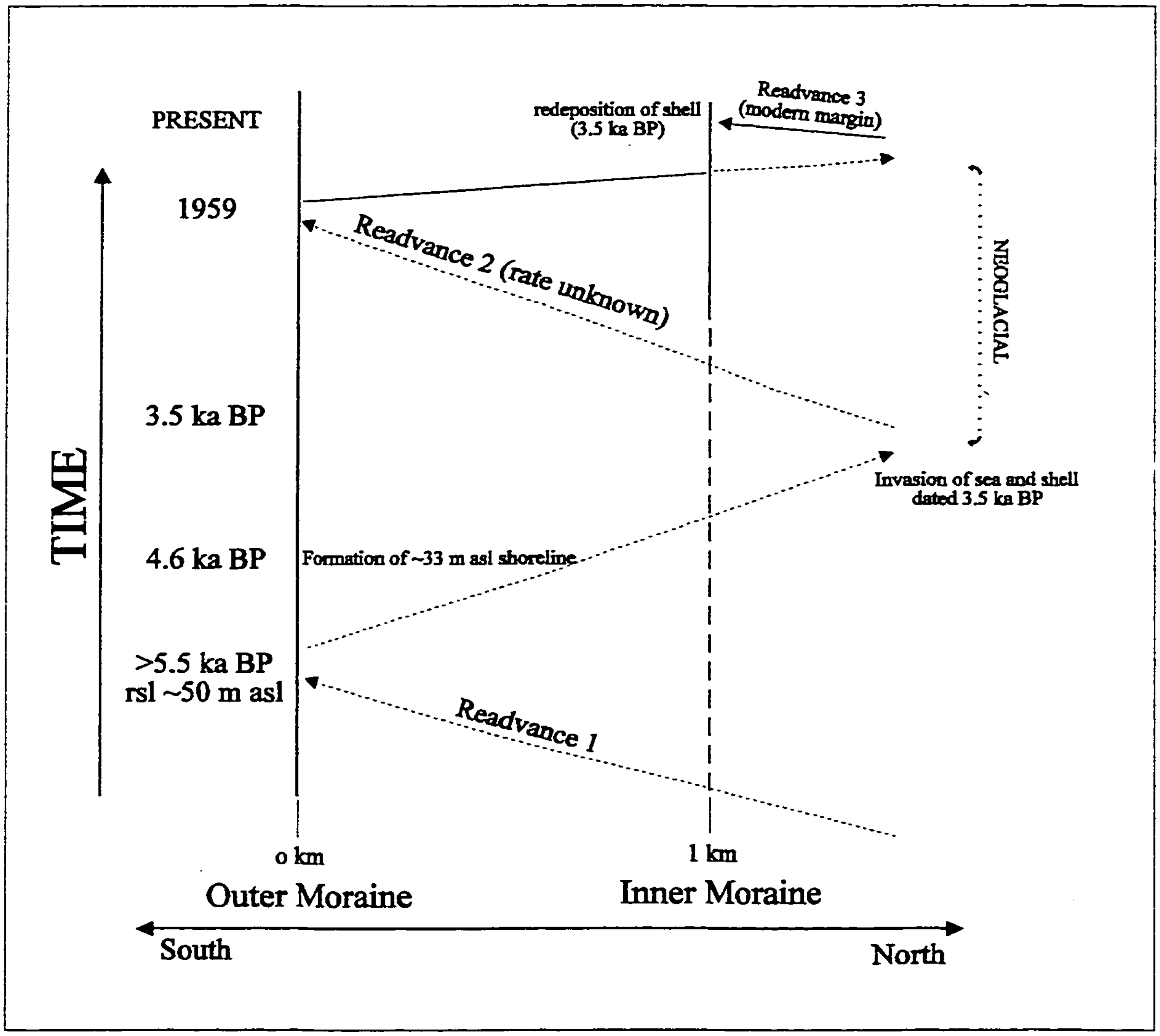


Figure 3.15 Ice margin fluctuations of Eugenie Glacier throughout the last 6 ka BP. Ice margin fluctuations based on two radiocarbon dates and recent airphotos of Dobbin Bay.

unknown, however; additional AMS dates could test whether a more widespread 1 ka BP ice advance occurred in this region (Blake 1981).

It is considered fortuitous that the modern ice margin readvanced sometime prior to 1959 to reattain the 4.6 ka BP moraine and not cause any modification of it. The outer moraine may represent an approximate equilibrium position for Eugenie Glacier due to geomorphic (i.e. the bedrock constriction upvalley) and bathymmetric (i.e. effectiveness of calving) controls within this broad valley at the fiord head.

This study emphasizes the need for more dating of single valves from moraines bordering modern ice margins. These provide evidence of past incursions of the sea >5 km inside modern glaciers. Blake (1987a) has provided ages on shells along and inside modern ice margins on SE Ellesmere Island. In one case, shells as young as 4 ka BP occur 20 km *inland* of the modern margin of Leffert Glacier. Additional radiocarbon dates on shells and organics (Blake 1981) incorporated in moraines could be more widely investigated to reconstruct late Holocene glacier fluctuations and to determine their regional variability with respect to climatic forcing. This would be relevant to late Holocene paleoclimatic changes especially along the margins of glaciers from which isotopic records have been derived (Agassiz Ice Cap, Devon Island, and NW Greenland). To date, little has been done to connect the isotopic signals from these ice cores with marginal fluctuations of the same glaciers. Hence, elaboration of late Holocene glacier fluctuations warrants further research.

3.5 'Upper Valley' Piedmont Terminal Moraine

Within the 'upper' valley, a terminal moraine flanks a small piedmont lobe (#1, Figs. 2.18 and 3.16). Exposures indicate similar depositional structures to that of the outer

Eugenie moraine. The piedmont moraine is cored with fine inter-bedded silt and sand which have been deformed into recumbent folds (Fig. 3.17). A similar history of ice-push, which was invoked for the formation of the outer lateral moraine, is also applied here. As the piedmont glacier readvanced into the shallow marine embayment, unfrozen sediment was deformed into a ridge. As in the case of the outer Eugenie moraine, moraine formation in the sea is recorded by subsequent wave-washed surfaces and regressive beaches extending from the upper part of the moraines distal slope. The timing of this event is unknown, however, it likely corresponds to the formation of the outer lateral moraine (mid Holocene) when rsl was at least 45 m asl. This is also indicated by the ice-contact delta abutting the piedmont lobe (47 m asl).

3.6 Dobbin Bay Lowland - Regressive Shorelines

The elevation of the 4.6 ka BP shoreline will be compared to other shorelines of similar age (mid-Holocene) across Ellesmere Island. This profile can then be compared to regional isobases previously published for the eastern Queen Elizabeth Islands and NW Greenland (cf. England 1997; England and O'Cofaigh 1998; O'Cofaigh 1999). It should be noted that the regional isobases are based on the surveying of widely separated marine limit deltas (of varying age) and necessarily lack the precision of a continuous shoreline for documenting differential postglacial emergence. The importance of the Dobbin Bay shorelines, from a regional perspective, is that they provide a synchronous water plane that and be directly measured over a distance of >5 km.

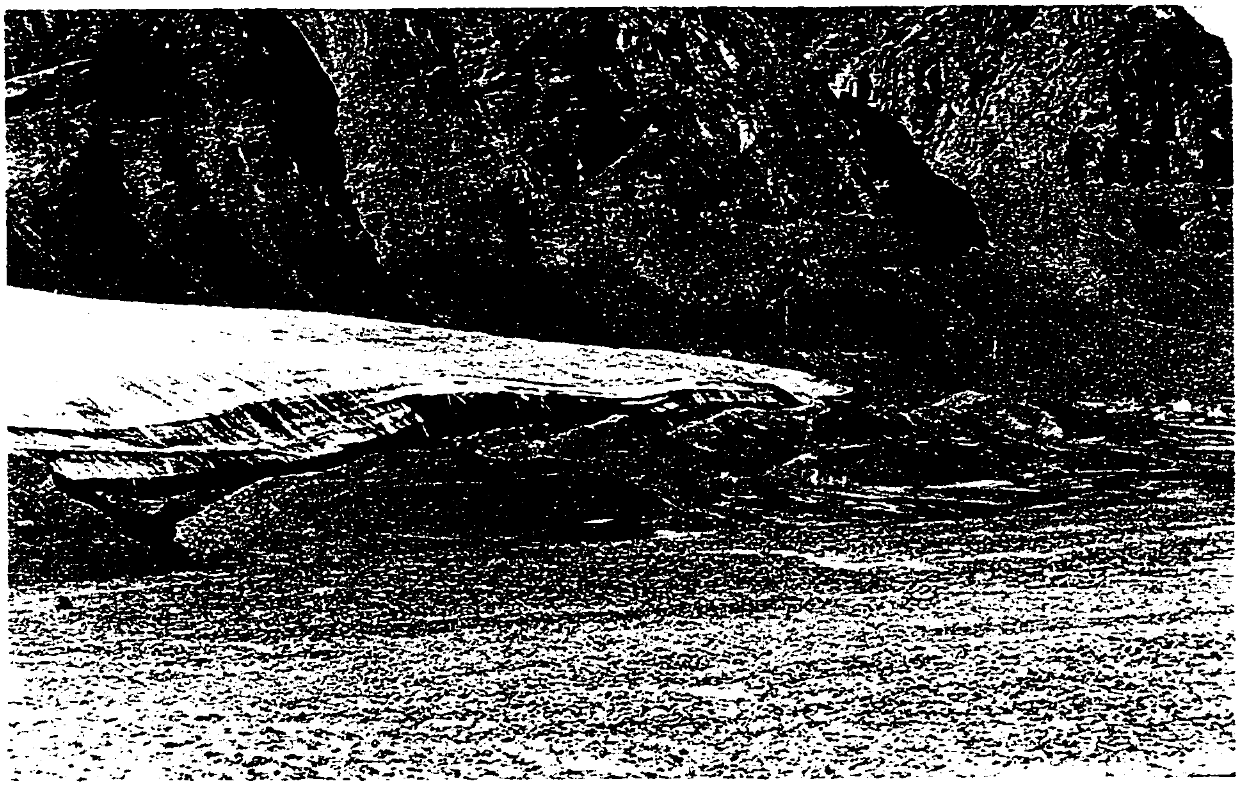


Figure 3.16 Photograph of the terminal moraine flanking the 'upper' valley piedmont lobe. This moraine has been highly eroded by fluvial processes. View to east towards Eugenie Glacier.

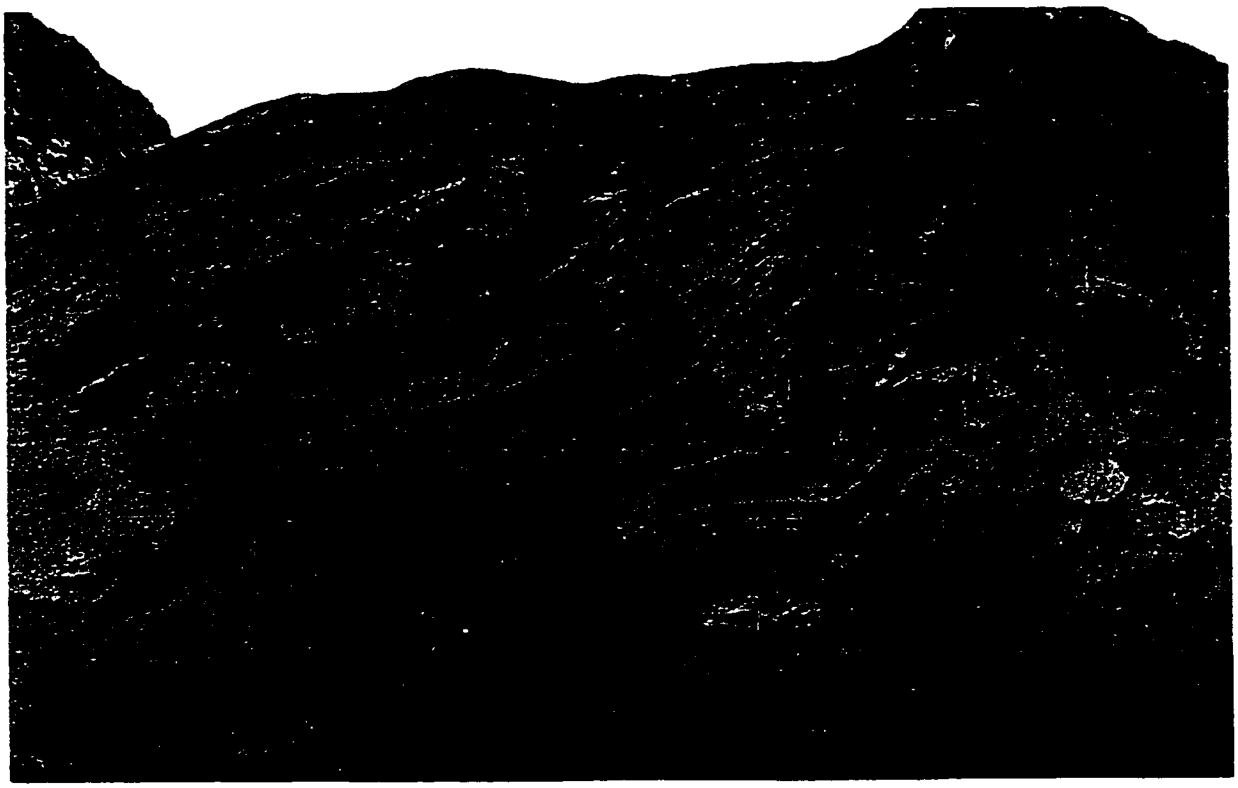


Figure 3.17 Photograph showing the internal structures within the piedmont moraine. Note the intense folding and deformation of the beds. Person for scale is ~1.5 m tall. The folding is again dipping away from the ice front.

3.6.1 Description of the shorelines

Along the distal slope of the outer Eugenie moraine, a series of ~20-30 regressive beaches extend from >45 m down to ~4 m asl. Those shorelines below 4 m asl have been eroded by the modern river which flanks the outer moraine. Those shorelines above ~31 m asl are discontinuous and cannot be accurately traced due to their poor preservation. Between 31 and 4 m asl, the shorelines are continuous along the length of the moraine with only small areas of disturbance. In general the continuous shorelines are separated vertically by 1-2 metres. On the 'upper' valley moraine in front of piedmont glacier #1, other shorelines were noted to 33 m asl (Fig. 3.18). Shorelines on both moraines are represented by a flat bench (tread) notched into the moraine and covered by a small boulder lag (Fig. 3.19). At some locations, the treads are so well developed that they form a prominent notch extending to a ~30 cm scarp cut into the moraine.

3.6.2 Formation of the regressive shorelines and source of 4.6 ka BP driftwood

The shorelines preserved on the outer Eugenie moraine are considered to be the result of glacioisostatic uplift that caused continuous sea level regression during the Holocene. Alternatively, these shorelines could have formed during the progressive lowering of a proglacial (ice-dammed) lake, however; most field evidence argues against this. The coalescence of the eastern piedmont glacier (lower valley) with lower Eugenie Glacier, supports the damming of a large inland lake. At present the separation of these two glaciers is ~30 m. However, the deposition of the outer Eugenie moraine likely occurred in a submarine setting as indicated by the widespread occurrence of marine shells and the inundation of the moraine to it crest. The exceptional preservation of the shorelines also

Figure 3.18 Photograph of the 'upper' valley piedmont moraine and the shorelines on its distal slope. The arrow marks the position of the highest continuous shoreline. View to the north across the 'upper' valley with the main river in foreground.

Figure 3.19 Photograph showing the distinct shoreline notches on the moraine. The lower shoreline (black arrow) is a combination of gravel lag and small scarps. The uppermost shoreline (open arrow) is characterized by a bench and distinct gravel lag. View to the east along the eastern end of the outer moraine.

indicates that the moraine is free of buried ice which is also consistent with its formation below sea level. Furthermore, the continuity of the shorelines from ~50 m down to ~4 m asl would require the lake to be (a) very recent, and (b) consequently, to have no discernible shoreline gradient. The fact that the shorelines are tilted isostatically (see later section) precludes their recent origin and the fact that they are evenly spaced is unlike the depositional history of most proglacial lakes that tend to drain catastrophically and intermittently.

The likelihood that the shorelines are marine is further indicated by one radiometric date which is fully consistent with the regional history of Holocene emergence. Driftwood found on the highest continuous shoreline dated 4600 ±70 BP (Beta-91862) . This sample was found resting on a prominent tread and was aligned with the tread scarp at its backside (Fig. 3.12). The driftwood provides a maximum age for this shoreline as it was rafted into the valley sometime after the tree died and, whether it originated in North America or Russia (the predominant source) it would have been transported across the Arctic Ocean by the Transpolar Drift, which requires at least three years (Häggblom 1982; Dyke et al. 1997). Furthermore, it could have been redeposited from a higher elevation, making it that much older than the 32 m asl shoreline. That wood of this age (4.6 ka BP) occurs at ~32 m asl, however; is consistent with the reported height of other 5 ka BP shorelines in the Queen Elizabeth Islands, which are commonly >25 m asl (Blake 1970).

3.7 Shoreline Survey

The shorelines on the outer Eugenie moraine were surveyed in order to calculate the amount of delevelling that has occurred since their formation. Two shorelines were surveyed. The uppermost continuous shoreline (~30 m asl) was measured using level and stadia, as well as by Wallace and Tiernan micro-altimeter at four benchmarks along it. This shoreline had an orientation of 290°. A lower shoreline (~5 m asl), with the same orientation, was also surveyed over a shorter distance (1.2 km) by level and stadia only (Fig. 3.19). The tilt of these shorelines is compared as an indication of changing rates of uplift during the mid to late Holocene. The more rapid the rate of uplift, the greater is the tilt of an individual shoreline (Andrews 1970). Because emergence tends to be most rapid during deglaciation, and slows progressively to the present, older shorelines have been found to have a steeper tilt than younger ones.

The 30 m asl shoreline on the outer moraine represents the highest continuous shoreline noted within Dobbin valley and therefore was chosen for surveying. This constitutes one of the longest continuous surveys (~6 km) of a shoreline in the High Arctic. This shoreline also extends upvalley to the moraine of piedmont glacier #1, where its orientation changes to 240°. The uppermost continuous shoreline on the outer Eugenie moraine occurs at 30.5 m asl at its eastern end and rises to 33.1 m asl at its western end. Collectively, the measured tilt of the uppermost shoreline averages 0.52 m/km based on the level and stadia method. This tilt was calculated over a distance of 5.6 km (Appendix A; and Figs. 3.20 and 3.21). The micro-altimetry indicated a shoreline tilt of 0.63 m/km measured across a distance of 4.2 km. Across the same distance, the two methods produces only minimal differences. The level and stadia method indicated that after 4.2 km, the elevation changed from 30.5 to 32.6 m asl, whereas the micro-altimetry method indicated a change from 30.5 to 33.1 m asl (Figs. 3.20 and 3.21). The difference is due to the lower order of precision for microaltimetry. As stated in Chapter 1, micro-altimetry is only accurate to within ±2 metres up to 100 m asl. The altimeter is subject to the inherent limitations of any mechanical device trying to measure the differences in atmospheric pressure between points separated vertically by only a few metres.

The uppermost continuous shoreline on the piedmont moraine rises from 33 m asl in the east to 33.4 m asl in the west. This provides a measured tilt of 0.40 m/km across a surveyed distance of 1.2 km (Appendix B; Fig. 3.21). This delevelling is less than the upper shoreline on the outer Eugenie moraine (0.52 m/km), however; it is not possible to compare these two shorelines accurately because they have different orientations, which results in differing amounts of divergence from the true isobase gradient. Nonetheless, the tilt of the shoreline bordering piedmont glacier #1 accords with the westward rise recorded by the upper shoreline on the outer Eugenie moraine.

The last shoreline which was surveyed occurs at ~5 m asl at the eastern end of the outer Eugenie moraine. It recorded a measured tilt of 0.31 m/km over a distance of 1.3 km (Appendix C; Fig. 3.20). This shoreline is close to modern sea level and although its age is unknown, it is presumed to be somewhat older than 1 ka BP. Throughout this area, paleoeskimo sites (Thule, 1 ka BP) are commonly ~3 m asl and hence must be somewhat younger than the 5 m relative sea level (J. England, pers. comm. 1998).

3.8 Comparison of Dobbin Bay Shoreline Tilting

The shoreline tilts recorded for Dobbin Bay can be compared to other measured and calculated shoreline tilts on Ellesmere Island. These previously published records are based either on direct measurements or isolated, discontinuous strandlines assumed to be of similar

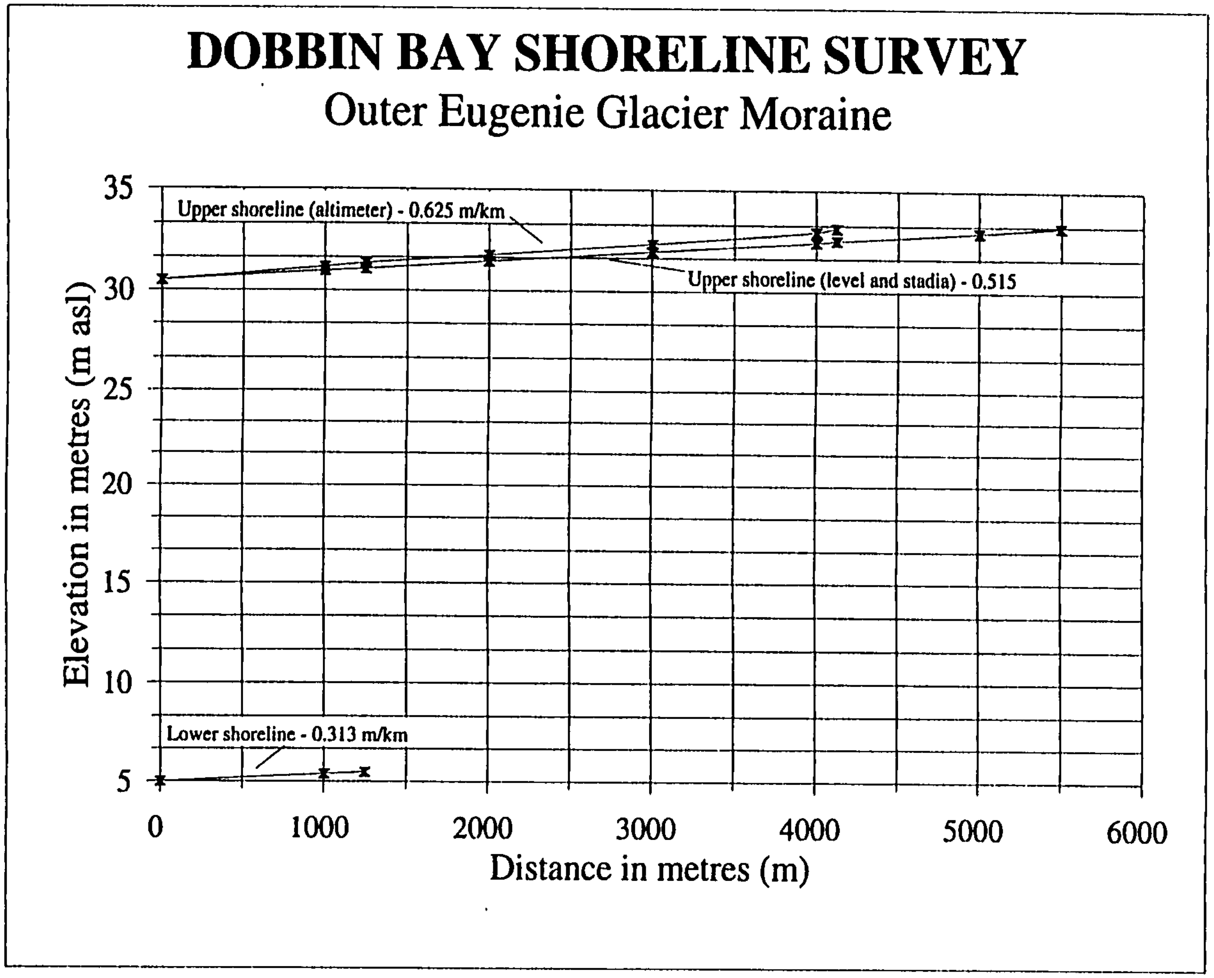


Figure 3.20 Graph depicting the surveyed shoreline gradients from outer Eugenie moraine. These gradients include the upper shoreline (level and stadia), upper shoreline (altimetry), and lower shoreline (level and stadia).

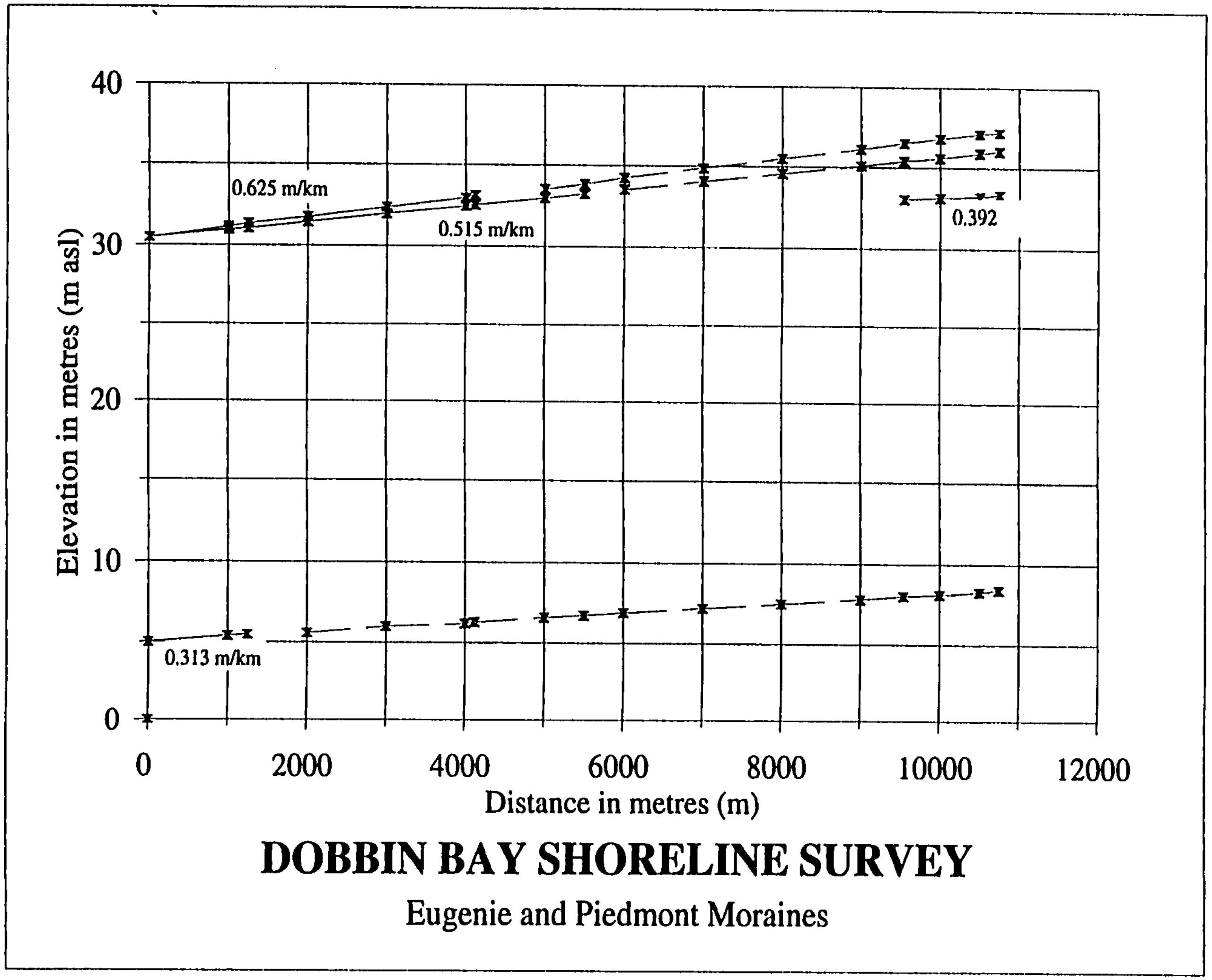


Figure 3.21 Graph depicting the surveyed and projected shorelines for Dobbin Bay lowland. These gradients include Figure 3.21, upper piedmont moraine shoreline, and the projected gradient of the Eugenie shorelines extending to the end of the piedmont moraine.

age. On southern Ellesmere Island, Blake (1970) recorded a pumice line which dated ~5 ka and had a recognizable westward tilt of 0.06 m/km. This is a significantly lower gradient than the tilt of the 4.6 ka BP shoreline surveyed in Dobbin Bay (0.52 m/km), however; Blake'smeasurement is based on isolated pumice elevations. Dyke (1998) indicates that the distribution of these sites is not at right angles to the 5 ka BP isobases (which would lower their gradient).

Previous studies of shoreline tilt on the northern coast of Ellesmere Island were based on widely separated deglacial deltas whose chronology was established by radiocarbon dates (Evans 1990). Evans reported that the profile of the 8.5 ka BP shoreline (close to the time of deglaciation) was curvilinear and ranged from 0.75 (outer coast) to 0.85 m/km (fiord head). In Clements Markham Inlet, Bednarski (1986) found that the 8.6 ka BP shoreline tilts at 0.75 m/km whereas the 10.5 ka BP tilt is even greater (0.88 m/km). These shorelines on northern Ellesmere Island clearly predate those found in Dobbin Bay, hence the younger shoreline tilt reported here (0.52 m/km, 4.6 ka BP) reinforces the evidence that tilt decreases with age (Andrews 1970). An even greater decrease in shoreline tilt in the late Holocene is recorded by the lower shoreline in Dobbin Bay (0.3 m/km at >1 ka BP).

3.9 Comparison to Other Mid-Holocene Shorelines: Eastern Ellesmere Island

This section compares the upper Dobbin Bay shoreline (~33 m asl) to previous observations reported on shorelines of similar age from southern and eastern Ellesmere Island. Based on the driftwood date of 4.6 ka BP it is concluded that relative sea level at this time was ~33 m asl and that this shoreline has been delevelled at a gradient of ~0.5 m/km (orientated WNW - ESE). England (1996) also collected a surface shell sample (5160 ±100

BP, GSC-5712) within the 'upper' Dobbin Bay valley in what he interpreted to be bottomset beds to a delta at ~31 m asl. The driftwood sample on the 33 m asl shoreline (4.6 ka BP) nearby would suggest that the 5.2 ka BP shells (GSC-5712) relate to a former sea level >33 m asl. Elsewhere on southern Ellesmere Island, Blake (1975) indicated that the 4.6 ka BP shoreline at Cape Storm, southern Ellesmere Island, is ~20 m asl based on a well dated emergence curve there. To the north of this site, at Makinson Inlet, SE Ellesmere Island, Blake (1993) indicated that the 4.6 ka BP shoreline lies between 19 and 21 m asl and this is very close to its height (~20 m asl) both at Cape Storm and Cape Herschel (~200 km south of Dobbin Bay, Blake 1992). Because, these sites all occur at the same elevation (20 ±1 m asl), they define an isobase oriented SW - NE. Based on the elevation and date of the driftwood, the 4.6 ka BP shoreline for Dobbin Bay is at least 33 m asl which places this site up-isobase from the other 4.6 ka BP shorelines discussed above.

Slightly older sea levels have also been reported from Bietstad Fiord and Sverdrup Pass. At Sverdrup Pass a shoreline dated 4990 ±70 BP (GSC-3929) occurs at an elevation of 35 m asl, and according to Blake (1987a) this date could relate to a delta surface as high as 50 m asl (Blake 1987a). Recent work by England (pers. comm. 1999), however; demonstrates that the 50 m delta nearby is close to 6.0 ka BP. At Bietstad Fiord, shells underlying a 35 m asl delta dated 5520 ±90 BP (GSC-3360; Blake 1987a) suggesting that this site is down-isobase (i.e. lower) than the head of Dobbin Bay. The highest shorelines in Dobbin Bay are ~ 50 m asl and are estimated to be ≥5.5 ka BP (assuming a rate of emergence of 1 m/century between 4.6 ka and 6.0 ka BP) which indicates greater emergence within Dobbin Bay compared to the southern fiord heads. This supports Blake's (1970,

1975) proposal that a SE-NW rise in the regional isobases records maximum glacial unloading toward Eureka Sound (along western Ellesmere Island).

3.10 Comparison to Other Mid-Holocene Shorelines: Western Ellesmere Island

Studies on western Ellesmere Island have shown that uplift was greater than on the east coast. Blake (1970) emphasized that emergence since 5 ka BP reached >25 m asl in Eureka Sound. Based on this elevation, on both coasts of Ellesmere Island, he indicated a zone of maximum uplift centered over Eureka Sound (Blake 1970). England (1976) showed that the 6 ka BP isobases over Eureka Sound reached 34-54 m asl (after removing his eustatic correction of ~6 m). Subsequently several authors have shown that the 5 ka BP shoreline, exceeds 30 m asl and possibly reaches 50 m asl (O'Cofaigh 1999). These elevations are based on local reconstructions within Blind Fiord and Irene Bay (O'Cofaigh 1999) where the 5 ka BP shoreline is >31 m and ~50 m asl, respectively (O'Cofaigh 1999). Sea level curves reported by England (1992) for Greely Fiord (NW Ellesmere Island) indicate that the 5 ka BP shoreline reaches ~45 m asl which corresponds to these recent findings by O'Cofaigh.

Farther west, on Axel Hieberg Island, the 5 ka BP shoreline was reported to be >35 m asl (Lemmen et al. 1994). One problem with the 4.6 ka BP shoreline gradient from Dobbin Bay, is that its extension ~200 km to the west (at a rate of 0.5 m/km) gives an unacceptably high elevation of ~130 m asl. This suggests that either the surveyed shoreline is too steep, or, if correct, that its gradient lowers to the west, in the direction of Eureka Sound. Regionally it is concluded that the western coast of Ellesmere Island had greater uplift than on the east coast indicating that the ice centre over Eureka Sound played the dominant role in the differential emergence of Ellesmere Island.

3.11 *Comparison to Regional Isobases*

Delevelled shorelines have been used routinely to define former centres of maximum ice thickness whereby the highest shorelines coincide with the greatest glacioisostatic unloading (Andrews 1970; Blake 1970; Walcott 1972; England 1976; Tushingham 1991; England and O'Cofaigh 1998). In the high Arctic, the reconstruction of postglacial isobases has been based almost exclusively on isolated marine limit deltas which provide only generalized measurements of differential emergence (Andrews 1970). In Arctic Canada, isobases drawn on these shorelines have been widely published (Andrews 1970; Walcott 1972; Blake 1975; England 1976, 1997). Furthermore, geophysical models of postglacial shorelines have also been used to estimate former ice thickness (Tushingham 1991; Tushingham and Peltier 1991; England et al. 1991). Nonetheless, the lack of continuous and well dated strandlines has limited our ability to test the accuracy of isobases drawn on deglacial deltas or younger shorelines from isolated locations. Direct measurements of the shoreline in Dobbin Bay test these existing isobase maps based on deglacial deltas that commonly show a range of ages (±200 years).

Based on existing isobase maps, two major ice-centres are recognized to have affected Ellesmere Island during the LGM (England 1997). A cell of maximum emergence has long been recognized over Eureka Sound based on the elevation of the 5 ka BP (above) shoreline as well as marine limit elevations (Fig. 3.22: Blake 1970; Walcott 1972). This zone of maximum uplift was termed the Innuitian uplift in recognition of the Innuitian Ice Sheet which was responsible for the crustal unloading (Walcott 1972). This zone of uplift extended from NW Greenland southwestward to Eureka Sound (Walcott 1972; England

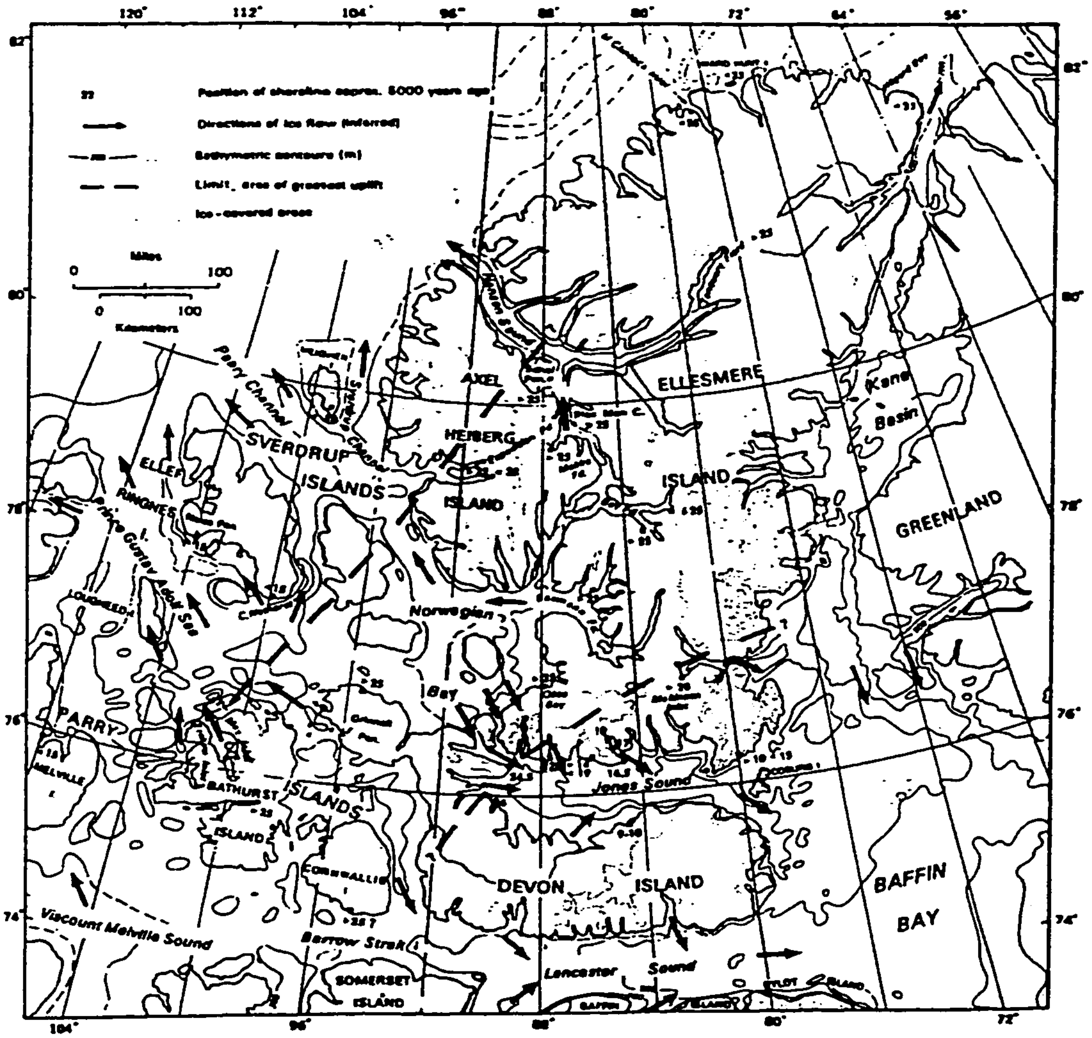


Figure 3.22 Reconstruction of 5 ka BP isobases for east-central Queen Elizabeth Islands. Figure shows the position of the 5 ka BP shoreline (~25 m asl) and the predicted outflow pattern from the Innuitian Ice Sheet (Modified from Blake 1970).

1976). Earlier studies (Blake 1970; Walcott 1972) and more recent studies (England and O'Cofaigh 1998; O'Cofaigh 1999) confirm that the greatest amount of emergence and highest marine limits on Ellesmere Island occur along Eureka Sound (Fig. 3.23). The second uplift centre corresponds to NW Greenland which extends to a saddle between it and Eureka Sound.

The keel of the saddle between NW Greenland and Eureka Sound roughly coincides with the head of Dobbin Bay and hence shorelines there provide an opportunity to clarify the nature of the transition between these two uplift centres. Although a direct test of the older isobases (8 ka BP) cannot be conducted at the head of Dobbin Bay, the general trend for the ~5 ka BP shoreline there can be used for comparison. Based on observations of the 5 ka BP shoreline elsewhere on SE Ellesmere Island (Cape Storm, Makinson Inlet, Cape Herschel), as well as surveying of the 4.6 ka BP shoreline at the head of Dobbin Bay, it is concluded that shorelines of mid-Holocene age lower to the east of inner Dobbin Bay. Hence, the keel of the saddle likely runs N-S through the outer fiords of east-central Ellesmere Island, or even farther east in Nares Strait. However, it is possible that earlier isobases (~8 ka BP) do record the axis of the saddle over Dobbin Bay during deglaciation and that this has migrated eastward as the Greenland uplift centre decayed and also shifted eastward. This may also imply that the Eureka Sound uplift centre has also migrated eastward (towards the fiords of western Ellesmere Island) during the Holocene.

3.12 Conclusions

This chapter presents evidence on the formation of moraines within the lowland at the head of Dobbin Bay. The formation of the lateral moraines which flank Eugenie Glacier

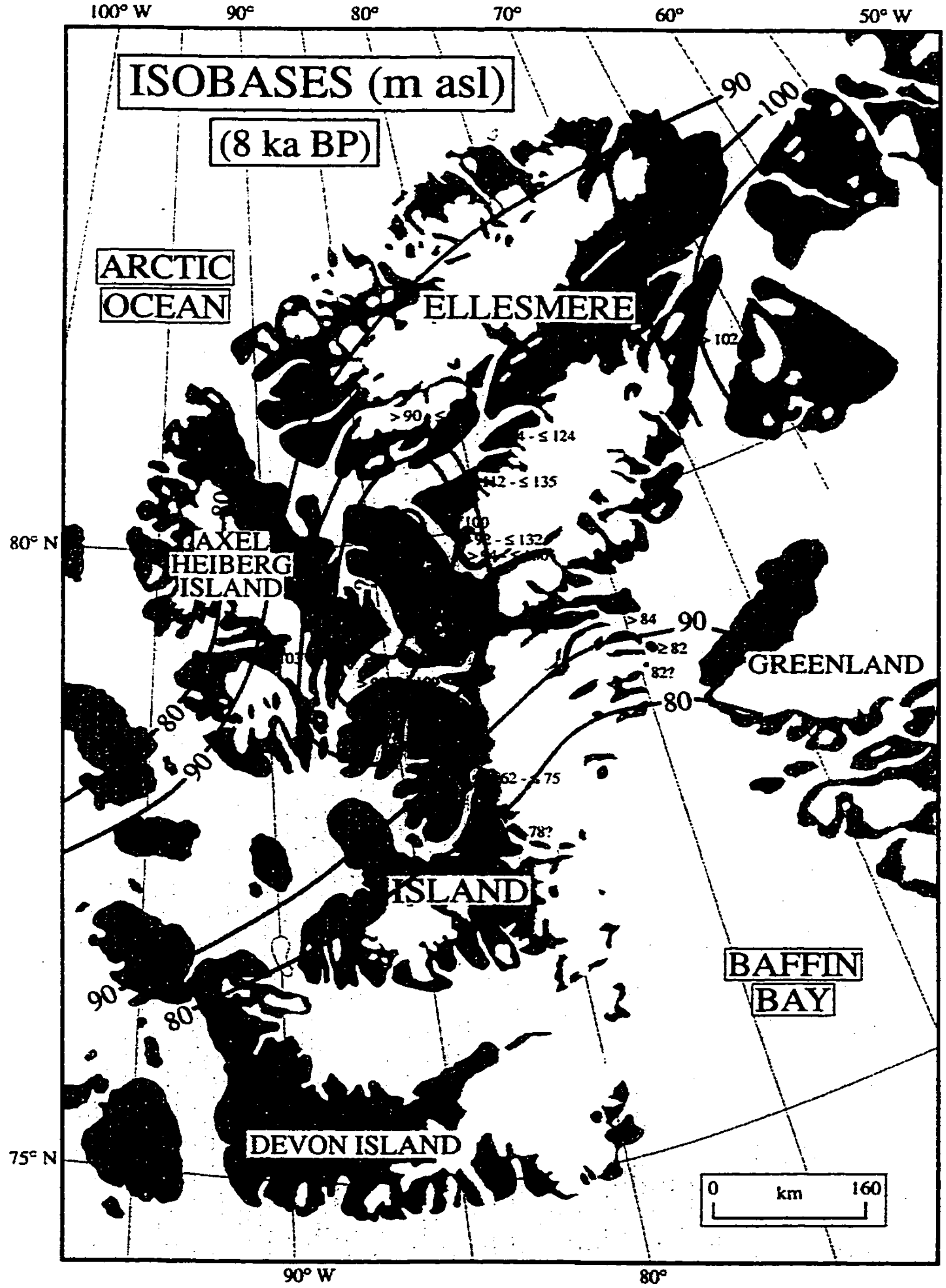


Figure 3.23 Unpublished 8 ka BP isobases from Ellesmere and Axel Hieberg Islands. Note the centre of maximum uplift over Eureka Sound and NW Greenland. Between these centres, a saddle (described in the text) is produced along eastern Ellesmere Island. The 5 ka BP isobase pattern is assumed to be approximately the same. Source: J. England 1999.

are among the longest (~10 km) observed anywhere in the High Arctic. Based on morphological and stratigraphic evidence it is concluded that the outer moraine was formed in a submarine environment under 'ice-push' conditions. Conversely, the inner lateral moraine was and is forming in a terrestrial environment containing widespread permafrost, hence involves ice-thrust. Radiocarbon dates on the outer and inner moraines require ice retreat from the outer moraine prior to 3.5 ka BP followed by a readvance (Neoglacial) sometime later, back to the position of the outer Eugenie moraine (Fig 3.15). Airphotos of Dobbin Bay indicate recent glacial retreat since 1959.

After the deposition of the outer moraine, a series of regressive shorelines trimmed its distal slope. These shorelines are important because they are among the longest recognized and surveyed in the Canadian High Arctic. The uppermost continuous shoreline (~30 m) has a tilt of ~0.52 m/km which is considered consistent with other mid-Holocene shorelines. Furthermore, this tilt precludes a recent, lacustrine origin for these shorelines. These shorelines also confirm that the maximum glacial unloading (since 5 ka BP) occurred to the NW of Dobbin Bay (toward Eureka Sound) and also indicates that the saddle between the Ellesmere Island and Greenland uplift centres lies somewhere east of Dobbin Bay.

Chapter 4

4.0 Conclusions

This thesis presented the Holocene deglacial and sea level history of Dobbin Bay, east central Ellesmere Island. In conjunction with other recent studies conducted along eastern Ellesmere Island (England 1996, 1997, 1999; Gualtieri and England 1998), this study adds to our understanding of late (<23 ka BP) ice buildup during the late Wisconsinan. Other observations from the Dobbin Bay lowland provide further evidence concerning: 1) the elevation of marine limit and hence the approximate date of deglaciation; 2) the mode of formation and ice marginal fluctuations of Eugenie Glacier since the mid-Holocene; and 3) the reconstruction of shoreline tilting measured directly on the outer Eugenie moraine.

A detailed survey of the ice-contact delta at the fiord head revealed a complex depositional history. Within a large exposure, two distinct tills were noted. These tills were separated by a thick deposit of glaciomarine sediment whose uppermost foresets radiocarbon dated ~23.3 ka BP. The overlying (uppermost) till was capped in turn by marine silts which extend to marine limit (71 m asl). Based on the stratigraphy and the age of the underlying foreset beds, it is interpreted that the upper till represents the late-Wisconsin ice advance. Locally, this confines late-Wisconsinan ice buildup at a site close to the modern ice margin to sometime *after* 23.3 Ka BP. This late ice buildup supports other studies along the eastern coast of Ellesmere Island that also indicate ice buildup after 20-25 ka BP (Blake 1992a, 1993; England 1996,1997, 1999).

The marine limit of inner Dobbin Bay is ~73 m asl based on the survey of the ice-contact delta within the 'lower' valley and the presence of wave-washing to ~74 m asl. The lip of the delta is 71 m asl which coincides with elevations from similar deglacial deltas out the fiord (England 1996). Furthermore, this elevation is accordant with the marine limit established in the 'upper' valley of the Dobbin Bay lowland (~ 75 m asl). Based on the age of similar marine limits in this area, it is concluded that the head of Dobbin Bay became ice-free ~7.3 ka BP.

The outer moraine flanking Eugenie Glacier was formed via ice-push processes while below the level of the mid Holocene sea (≤50 m asl). The intense deformation of the sediment, lack of subsurface ice, the widespread occurrence of shells, and the washing of sediment indicate that it was formed within a submarine setting. Conversely, the inner moraine, which marks the contemporary ice margin, is forming via ice-thrusting in a subareal setting. This is recorded by the incorporation of significant amounts of buried glacier ice as well as large blocks of sediment whose preserved primary bedding requires the presence of permafrost.

The 1959 airphoto indicates that Eugenie Glacier contacted the outer moraine, which has a minimum age based on a driftwood date of 4.6 ka BP and more likely dates ≥5.5 ka BP based on shorelines extending to ~50 m asl. Since 1959, the ice margin has retreated up to 1 km. This retreat involved a portion of the glacier that was severed from its original source by the bedrock constriction along the northern wall of the valley. This stagnant or inactive ice mass, which may have been delivered by a preceeding surge, has retreated rapidly since 1959. At present, the ice is readvancing from the westernmost flowline that was still connected to an upglacier source through the bedrock constriction (on the 1959 airphoto, Fig. 3.14). This readvance is forming the inner Eugenie moraine. A radiocarbon date (3.4 ka BP) in the inner moraine, however; indicates that the mid Holocene ice margin responsible for the deposition of the outer Eugenie moraine fluctuated more significantly than was originally thought. Based on this one date alone, three distinct glacial advances are recognized (Fig. 3.15). This emphasizes the importance of obtaining more AMS radiocarbon dates from shells in moraines near modern ice margins.

The distal slope of the outer moraine support a series of marine shorelines which extend from ~50 to 4 m asl. The highest continuous shoreline (30 m asl) was surveyed over ~5.5 km and tilted at 0.52 m/km. The fact that this shoreline is tilted precludes it from being of recent lacustrine origin. As well, the tilt of this shoreline is consistent with other tilts recorded for southern and northern Ellesmere Island (Blake 1970: Bednarski 1986; Evans 1990). Although each of these studies present shorelines of different ages, the general trend of decreasing gradients throughout the Holocene is supported here.

The age and elevation of the uppermost Dobbin Bay shoreline is also consistent with other shorelines reported along both the east and west coasts of Ellesmere Island. Previous studies, which extend from Jones Sound to Cape Herschel, eastern Ellesmere Island, indicate that the 4.6 ka BP shoreline is ~21 m asl. Within Dobbin Bay, 200 km to the northwest, the 4.6 ka BP shoreline is ≥33 m asl. This supports the presence of greater uplift to the NW (Eureka Sound) at ~5 ka BP (Blake 1970). This is further supported on western Ellesmere Island, where the 4.6 ka BP shoreline increases to >30 m asl and is possibly as high as 50 m asl (England 1992: O'Cofaigh 1999).

Based on the 8 ka BP isobases (Fig. 3.23), two uplift centres affected eastern Ellesmere Island. The Greenland uplift produced isobases which decrease westward from its NW coast, whereas the Eureka Sound uplift decreases to the east. These two uplift centres are separated by a saddle along the east coast of Ellesmere Island. Based on the survey of the 4.6 ka BP shoreline at the head of Dobbin Bay, the keel of this saddle was located somewhere to the east, possibly as far as Nares Strait by the mid-Holocene. This suggests that at ~5 ka, uplift in Eureka Sound dominated the emergence at the fiord heads of east-central Ellesmere Island. However, it is possible that at ~8 ka BP, the keel of the saddle was located farther inland (within inner Dobbin Bay) and that it migrated eastward as the Greenland uplift centre decayed. This would require a similar eastward migration of the Eureka Sound uplift centre.

4.1 Future Research

As a result of this study, three possible areas for future research in Dobbin Bay include: 1) greater sampling of shells in the moraines for AMS radiocarbon dating; 2) additional radiocarbon dating of shells found in growth position within silts associated with marine limit upvalley would provide a minimum age for deglaciation prior to the mid-Holocene readvance to the outer Eugenie Moraine; and 3) dating of a sample collected from silt below till in 'lower' Dobbin valley would help to further test the age of late-Wisconsin ice buildup.

Regionally, shoreline surveys should be conducted wherever possible. These profiles prove to be important for directly testing previously constructed regional isobases. Although preservation of such shorelines is rare, every effort should be made to measure shorelines ≥1 km, as they add to the regional database required to draw isobase maps.

The most important focus for future research from this study, which has regional significance, is the radiocarbon dating of recent (Neoglacial) moraines bordering modern ice margins. As shown in this study, these are commonly fossiliferous as these ice margins reoccupied the late Holocene sea, or redeposited its emergent sediments. In the case of inner Dobbin Bay, it has been shown that one date alone can greatly affect the reconstruction of Neoglacial ice margins. The radiocarbon dating of such moraines, combined with the isotopic record from adjacent ice cores (Koerner and Fisher 1990), would improve our understanding the paleoclimatic conditions favouring high Arctic glacier fluctuations. This research should be completed on both eastern and western Ellesmere Island, in order to investigate possible differences in regional paleoclimatic change.

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APPENDICES

Shoreline Gradient Database

Appendix A - Upper Shoreline, Outer Eugenie Moraine

Appendix B - Upper Shoreline, Piedmont Glacier #1 Moraine

Appendix C - Lower Shoreline, Outer Eugenie Moraine

E	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	P	Q	R	S
1	couplet #	back.up	back.mid	back.low	instrument	for.up	for.mid	for.low	bac-ins dis	for-ins dis	cos-angle	meas.dis	actual.dis	change E	change F-I	total change			
2	1	236	186	135	149	193	158	124	101	69	0.999	170	169.83	37	-9	28			
3	2	212.5	156.5	102	162	182	163	144	110.5	38	0.996	148.5	147.906	-5.5	-1	-6.5			
4	3	193.5	161	128.5	157	170	142	114	65	56	0.996	121	120.516	4	15	19			
5	*4	71		38		31.8		26.8	33	5	0.105	38	3.99	0	0	0			
6	*5	140		104.5		210		116	35.5	94	0.994	129.5	128.723	0	0	0			
7	6	214	150	87	158	203	148	93	127	110	0.438	237	103.806	-8	10	2			
8	7	205	163	121	157	187	141	96	84	91	0.999	175	174.825	6	16	22			
9	8	244	175	107	153	250	216	183	137	67	0.999	204	203.796	22	-63	-41			
10	9	177	159	141	160	107	69	32	36	75	0.974	111	108.114	-1	91	90			
11	10	111	64	17	150	166	123	81	94	85	0.985	179	176.315	-86	27	-59			
12	11	178	142	105	157	211	174	136	73	75	0.99	148	146.52	-15	-17	-32			
13	12	298	268.5	239	149	148	103	58	59	90	1	149	149	119.5	46	165.5			
14	13	183	148	114	149	175	143	111	69	64	0.996	133	132.468	-1	6	5			
15	*14	135		20.5		139		8	114.5	131	0.777	245.5	190.7535	0	0	0			
16	15	183	94	4	146	170	131	93	179	77	0.423	256	108.288	-52	15	-37			
17	16	210	136	63	159	182	139	95	147	87	0.993	234	232.362	-23	20	-3			
18	17	149	122.5	96	150	224	175	128	53	96	0.996	149	148.404	-27.5	-25	-52.5			
19	18	248	180	114	149	189	152	115	134	74	0.978	208	203.424	31	-3	28			
20	19	185	153	122	152	185	106	28	63	157	0.868	220	190.52	1	46	47			
21	20	189.5	133	76	163	279	192	106	113.5	173	0.985	286.5	282.2025	-30	-29	-59			
22	21	287	245	203	148	234.5	215	196	84	38.5	0.994	122.5	121.765	97	-67	30			
23	22	234	142	49	149	172	123.5	75	185	97	1	282	282	-7	25.5	18.5			
24	23	125	64	3	146	183	131	79	122	104	0.999	226	225.774	-82	15	-67			
25	24	170.5	124.5	78.5	142	192.5	162	131	92	61.5	0.995	153.5	152.7325	-17.5	-20	-37.5			
26	25	139	111.5	84	147	115	90	64	55	51	0.999	106	105.894	-35.5	57	21.5			
27	26	140	107	75	151	139	96	52.5	65	86.5	0.988	151.5	149.682	-44	55	11			
28	*27	173		85		177		111	88	66	0.946	154	145.684	0	0	0			
29	*28	132		37		169		101	95	68	0.819	163	133.497	0	0	0			
30	29	192	157	124	153	192	157	122	68	70	0.998	138	137.724	4	-4	0			
31	30	162	140	117	158	158	142	128	45	28	0.984	73	71.832	-18	16	-2			
32	31	61		28		313.5		290.5	33	23	0.984	56	55.104	0	0	0			
33	*32	109		49		177		130	60	47	0.961	107	102.827	0	0	0			
34	*33	220		94		85		44	126	41	0.94	167	156.98	0	0	0			
35	34	300	243	186	146	106.5	86	65	114	41.5	0.984	155.5	153.012	97	60	157			
36	35	231	192	153	156	175	151	128	78	47	0.995	125	124.375	36	5	41			
37	36	167.5	157.5	146	156	179	150	121	21.5	58	0.995	79.5	79.1025	1.5	6	7.5			
38	37	175	147	119	154	181	155	129	56	52	0.914	108	98.712	-7	-1	-8			
39	38	202	150	99	155	174	152	131	103	43	0.999	146	145.854	-5	3	-2			
40																			
41													5564.314			286.5			

APPENDIX A - UPPER EUGENIE MORAINE SHORELINE

E	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	P	Q	R	S
43																			
44				COLUMN	FULL TITLE														
45				A	Couplet Number				H	Forsight lower cross hair reading									
46				B	Backsight upper cross hair reading				I	Backsight to insturment distance									
47				C	Backsight middle cross hair reading				J	Forsight to instrument distance									
48				D	Backsight lower cross hair reading				K	Total Distance									
49				E	Instrument hieght				L	Change in hieght from instument to backsight									
50				F	Forsight upper cross hair reading				M	Change in hieght from instrument to forsight					Normal			Reversed	
51				G	Forsight middle cross hair reading				N	Total change in hieght					Percent Slope	Slope (cm/100 m)			
52													0.00		0.000000	0.0000		0	0
53													16983.00	28.00	0.164871	16.4871		14585.4	-2
54													31773.60	-6.50	-0.043947	-4.3947		24456.6	-8
55			* - represents level and stadia postitions where shoreline was approximated. These are no										43825.20	19.00	0.157655	15.7655		32366.65	7.5
56			used in the overall calculation of shoreline tilting, rather used only for distance measurements										44224.20	0.00	0.000000	0.0000		44804.35	41
57													57096.50	0.00	0.000000	0.0000		60105.55	157
58													67477.10	2.00	0.019267	1.9267		75803.55	0
59													84959.60	22.00	0.125840	12.5840		88086.25	0
60													105339.20	-41.00	-0.201182	-20.1182		91596.65	0
61													116150.60	90.00	0.832455	83.2455		98779.85	-2
62													133782.10	-59.00	-0.334628	-33.4628		112552.2	0
63													148434.10	-32.00	-0.218400	-21.8400		125901.9	0
64													163334.10	165.50	1.110738	111.0738		140470.4	0
65													176580.90	5.00	0.037745	3.7745		155438.6	11
66													195658.25	0.00	0.000000	0.0000		166028	21.5
67													206485.05	-37.00	-0.341681	-34.1681		181301.2	-37.5
68													229721.25	-3.00	-0.012911	-1.2911		203878.6	-67
69													244561.65	-52.50	-0.353764	-35.3764		232078.6	18.5
70													264904.05	28.00	0.137644	13.7644		244255.1	30
71													283956.05	47.00	0.246693	24.6693		272475.4	-59
72													312176.30	-59.00	-0.209070	-20.9070		291527.4	47
73													324352.80	30.00	0.246378	24.6378		311869.8	28
74													352552.80	18.50	0.065603	6.5603		326710.2	-52.5
75													375130.20	-67.00	-0.296757	-29.6757		349946.4	-3
76													390403.45	-37.50	-0.245527	-24.5527		360775.2	-37
77													400992.85	21.50	0.203033	20.3033		379850.5	0
78													415961.05	11.00	0.073489	7.3489		393097.3	5
79													430529.45	0.00	0.000000	0.0000		407997.3	165.5
80													443879.15	0.00	0.000000	0.0000		422649.3	-32
81													457651.55	0.00	0.000000	0.0000		440280.8	-59
82													464834.75	-2.00	-0.027843	-2.7843		451092.2	90
83													470345.15	0.00	0.000000	0.0000		471471.8	-41
84													480627.85	0.00	0.000000	0.0000		488954.3	22
85													496325.85	0.00	0.000000	0.0000		499334.9	2
86													511627.05	157.00	1.026063	102.6063		512207.2	0
87													524064.55	41.00	0.329648	32.9648		512606.2	0
88													531974.80	7.50	0.094814	9.4814		524657.8	19
89													541846.00	-8.00	-0.081044	-8.1044		539448.4	-6.5
90													556431.40	-2.00	-0.013712	-1.3712		556431.4	28

F	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	P	Q
1	couplet #	back.up	back.mid	back.low	instrument	for.up	for.mid	for.low	bac-ins dis	for-ins dis	cos-angle	meas.dis	actual.dis	change I-B	change I-F	total change	confirmed
2	1	184	161	139	150.5	159	131	103	45	56	1	101	101	10.5	19.5	30	
3	2	219	129	40	154	215	172	129	179	86	1	265	265	-25	-18	-43	
4	3*	211	138	64	158	218	190.5	163	147	55	1	202	202	-20	-32.5	-52.5	
5	4*	148.5	138.5	128.5	151	150	130	110	20	40	1	60	60	-12.5	21	8.5	
6	5	245	162	78	157	182	164	140	167	42	1	209	209	5	-7	-2	
7	6	197	144	91	154.5	185	158	130	106	55	1	161	161	-10.5	-3.5	-14	
8	7*	176	99	21	165	211	189	166	155	45	1	200	200	-66	-24	-90	
9	8	205	183	161	157	151	112	73	44	78	1	122	122	28	45	71	
10	9	148	139	131	166	169	134	98	17	71	1	88	88	-27	32	5	
11																	
12													1198			-163	47
13		COLUMN	FULL TITLE														
14		A	Couplet Number				H	Forsight lower cross hair reading						Differential tilt (m/km)			0.392321
15		B	Backsight upper cross hair reading				I	Backsight to insturment distance									
16		C	Backsight middle cross hair reading				J	Forsight to instrument distance									
17		D	Backsight lower cross hair reading				K	Total Distance									
18		E	Instrument hieght				L	Change in hieght from instument to backsight									
19		F	Forsight upper cross hair reading				M	Change in hieght from instrument to forsight									
20		G	Forsight middle cross hair reading				N	Total change in hieght									
21																	
22		* - represents level and stadia positions where shoreline was approximated. These site are not used in the															
23		overall calculation of shoreline tilting, rather used only for distance measurements.															

APPENDIX B - UPPER PIEDMONT MORAINE SHORELINE

A	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	P
1	couplet	back.u	back.mid	back.low	instrument	for.up	for.mid	for.low	bac-ins dis	for-ins dis	cos-angle	meas.dis	actual.dis	change I-B	change I-F	total change
2	1	175	121.5	68	150	199	108	17	107	182	1	289	289	-28.5	42	13.5
3	2	201	114	27	147	169.5	130.5	92.5	174	77	1	251	251	-33	16.5	-16.5
4	3	183	125.5	69	145	172.5	127	82	114	90.5	1	204.5	204.5	-19.5	18	-1.5
5	4	160	114	68	145	181.5	139	96.5	92	85	1	177	177	-31	6	-25
6	5	116.5	139	162	145	171	103	36	-45.5	135	1	89.5	89.5	-6	42	36
7	6	175	141	108	154	199	108	17	67	182	1	249	249	-13	46	33
8																
9													1260			39.5
10		COLUI	FULL TITLE													
11		A	Couplet Number				H	Forsight lower cross hair reading								
12		B	Backsight upper cross hair reading				I	Backsight to Insturment distance								
13		C	Backsight middle cross hair reading				J	Forsight to Instrument distance								
14		D	Backsight lower cross hair reading				K	Total Distance								
15		E	Instrument hleght				L	Change in hleght from Instument to backsight								
16		F	Forsight upper cross hair reading				M	Change in hleght from Instrument to forsight								
17		G	Forsight middle cross hair reading				N	Total change in hleght								
18															Percent Slope	Slope (cm/100 m)
19													0	0	0.00000000	0
20													28900.0000	13.5	0.04671280	4.671280
21													54000.0000	-16.5	-0.06573705	-6.573705
22													74450.0000	-1.5	-0.00733496	-0.733496
23													92150.0000	-25	-0.14124294	-14.124294
24													101100.0000	36	0.40223464	40.223464
25													126000.0000	33	0.13253012	13.253012
26																
27																
28													Reversed			
29													0.0000	0		
30													24900.0000	33		
31													33850.0000	36		
32													51550.0000	-25		
33													72000.0000	-1.5		
34													97100.0000	-16.5		
35													126000.0000	13.5		

APPENDIX C - LOWER EUGENIE MORAINE SHORELINE