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THE RECORD OF EPISODIC HIGH-ENERGY SEDIMENTATION IN THE WESTERN OAK RIDGES MORAINE, SOUTHERN ONTARIO

by

Hazen Algar John Russell

A Thesis
Submitted to the School of Graduate Studies and Research
in Partial Fulfillment of the Requirements for the
Degree of Doctor of Philosophy in
Earth Sciences

Ottawa-Carleton Geoscience Centre
University of Ottawa
Ottawa, Canada

0-612-67220-4

Nothing in the world can take the place of PERSISTENCE,
Talent will not; nothing is more common than the unsuccessful man with talent.
Genius will not; unrewarded genius is almost a proverb.
Education alone will not; the world is full of educated derelicts.
Persistence and Determination alone are omnipotent...

Calvin Coolidge
1872-1933

Abstract

The 160 km long Oak Ridges Moraine in southern Ontario has been interpreted as an interlobate moraine, a subaerial braidplain-deltaic deposit, a subglacially formed ridge, or a polygenetic landform of subglacial and proglacial deposits. Because of these significant differences a detailed sedimentological study is needed in order to provide an improved depositional model. This study of the western 40 km of the moraine in the Humber River watershed challenges the concept that it was deposited from seasonal meltwater discharges, climatic modulated ice-marginal fluctuations or in an interlobate position. Instead it is interpreted to have formed in response to changes in the ice-sheet during deglaciation associated with subglacial ponding, episodic and catastrophic subglacial jökulhlaup discharge and seasonal meltwater discharge. The moraine probably formed as the ice-sheet profile and glacial hydraulic system re-equilibrated to the presence of a thinned ice-sheet and a subglacial lake in the Lake Ontario basin.

The Oak Ridges Moraine is interpreted to have been deposited in three discrete stages that record sedimentation from both high-energy and low-energy depositional processes. Stage I consists mostly of massive and diffusely-graded sand that is up to 50 m thick and forms the lowest tunnel channel infill. It was deposited rapidly from hyperconcentrated flows some distance downflow of a hydraulic jump that developed as jökulhlaup flow discharged from tunnel channels into a subglacial lake in the Lake Ontario basin. Low energy basinal sedimentation of Stage II is recorded by the upward increase in thin normally graded fine-sand, silt, and clay laminae forming varves. Stage III is characterized by rapid facies changes associated with subaqueous fan, esker and basinal sedimentation. Proximal fan sediment consists of heterogeneous gravel with sand intraclasts and planar-stratified gravel. Immediately downflow the supercritical to subcritical flow transition is recorded by steep-walled scours and diffusely-graded infill. Farther downflow the subcritical flow region of the fan is dominated by planar cross-stratified medium sand, climbing medium scale cross-stratified sand and small-scale cross-laminated fine sand.

Based on an incomplete varve chronology and general evidence of rapid and voluminous sedimentation it is estimated that the moraine was deposited over a short period of time, maybe as little as 100 years. This study provides an additional demonstration of the value of detailed sedimentological studies to advancing understanding of Laurentide meltwater and deglacial processes.

Résumé

L'origine de la moraine d'Oak Ridges, élément topographique de 160 km de long au sud de l'ontario, suscite la controverse. Elle a été interprétée comme une moraine interlobaire, un dépôt deltaïque à l'embouchure d'une rivière proglaciaire en tresse, une crête sous-glaciaire, ou un dépôt polygénétique sous-glaciaire et proglaciaire. La présente étude a été entrepris afin de développer un modèle de déposition et d'apporter des éléments nouveaux permettant d'en éclaircir son origine. Elle porte sur la partie ouest de la moraine Oak Ridges, soit le bassin hydrographique de la rivière Humber. Les résultats remettent en question les processus de formation de la moraine proposés ultérieurement; soit qu'elle a été mise en place par décharge saisonnière d'eaux de fonte, des fluctuations climatiques influençant les marges glaciaires, ou dans un contexte interlobaire. L'analyse sédimentologique détaillée révèle plutôt que la moraine d'Oak Ridges a été formée suite aux changements de la calotte glaciaire engendrés par la formation de lacs sous-glaciaires, par des déversements d'eaux de fonte épisodiques et catastrophiques (jökulhlaup), et des variations saisonnières du régime hydraulique. Plus précisément, la moraine a probablement été formée lors du rééquilibre des calottes glaciaires et système hydraulique suite à l'amincissement de l'extrémité de la calotte et à la présence d'un lac sous-glaciaire dans le bassin du Lac Ontario.

La moraine d'Oak Ridges a été déposée en trois phases témoins des processus de deposition de haute et de basse énergie. La première phase, avec un dépôt atteignant 50 m d'épaisseur, remplis la partie inférieure du chenal sous-glaciaire (ou intra-glaciaire ?) et est constituée presque entièrement de sables massifs et à granoclassement diffus. Les sables ont été mis en place rapidement par un écoulement saturé en sédiments en aval d'un ressaut hydraulique. Ce saut hydraulique résulte d'un écoulement jökulhlaup provenant de chenaux sous-glaciaires se déversant dans un lac sous-glaciaire du bassin du Lac Ontario. La deuxième phase, marquant une sédimentation de faible énergie, est caractérisée par

l'augmentation de la stratification fine varvée de sables fins, silts, et argiles. Des changements rapides de facies associés aux cônes alluviaux subaquatique, aux eskers, et à la sédimentation du bassin, caractérisent la troisième phase de déposition. Les sédiments des cônes proximaux sont constitués de graviers hétérogènes avec intraclastes de sable et de gravier à stratification planaire. Un peu plus en aval, des marques d'érosion à versants abrupts remplis par des dépôts à granoclassement diffus dénotent la transition d'un écoulement supercritique à un écoulement subcritique. Encore plus en aval, la section à écoulement subcritique du cône alluvionnaire est dominée par du sable moyen à stratification entrecroisée tabulaire et du sable fin à stratification entrecroisée ascendante de moyenne et petite taille.

D'après la chronologie sédimentaire des varves la moraine d'Oak Ridges a été mise en place en peu de temps, soit peut-être aussi peu que 100 ans, ce qui indique taux de sédimentation rapide et élevé. Cette étude démontre une fois de plus l'importante contribution des études sédimentologiques détaillées à l'avancement des connaissances sur les eaux de fonte de la calotte Laurentide et sur les processus de déglaciation.

Acknowledgements

... I am a part of all that I have met ...

From Ulysses

by Alfred Tennyson

This thesis would not have been possible without the encouragement and support of many people over a long period of time. I would like to use this opportunity to thanks both those who helped me during the course of this thesis and to some whose assistance and guidance preceded this exercise by many years. First and foremost I would like to thank my parents for their depth of support and encouragement since time immemorial. My interest in geology stems back to the initial curiosity fostered by high school teachers, particularly Chris McGill. Nevertheless, I would not have ventured into geology if it had not been for the accidental registration in a course titled *Geology in the Service of Man* that was taught with undying enthusiasm by Dr. Bob Stevens at MUN. It was following this course that I decided to study geology. It is thanks to a cadre of professors and graduate students at MUN that I continued in geology and finally developed an interest in glacial geology. Unsuspectingly Robert Hildebrand ensured through a field trip to the southwestern United States that I would never have an interest in olivine phenocrysts comparable to my interest in geomorphology and sedimentary processes. This was further fueled by two summers in the Torngat Mountain of northern Labrador with Dr. Bob Rogerson. I would like to thank Dr. Jaques Locat for his role in arranging the chance meeting on the shores of Lake Mellville, and a causal afternoon field trip along the Churchill Road, with Dr. David Sharpe, which established the foundation for involvement in this study. That brings us to the present and the host of friendships developed during the course of this thesis to whom I owe many thank, in particular, Shona, Mathew, Ruth, Bill, Elena, Sergy, Donna, Glen, Johannes, Simone, Don, and Jenny.

To my family I would like to express my gratitude for their patience during the course of this thesis and the many missed deadlines and moments *sur la lune*. More importantly, thanks to Heather and Nicholas for ensuring that the most important things in life were never forgotten. To Nicholas, now that this is done, I hope we will have more time to do things together.

During the course of this study Tracy Brennand, John Shaw and Bill Shilts, made many helpful contributions to my knowledge and perspectives on geology. At the Geological Survey of Canada I would like to thank, Charles Logan for database support, Rachelle Lacroix for tutelage and assistance with the images in this thesis, and Eric Boisvert for providing the software Gran. The quality of the French abstract is due to the diligent effort of Simone Dumas. To David Sharpe I owe many thanks for making my participation in the Oak Ridges Moraine project at the Geological Survey of Canada possible and for his support of this thesis. I also extend my gratitude to Dave for opening my eyes to the realm of glacial sedimentology through many entertaining field trips. To Bill Arnott there is no degree of thanks that I can extend for his diligent editing and continual support to finish this thesis. But most important was his continual good humour, professional attitude, and enthusiastic interest in sedimentology.

Funding for this project was provided by the Geological Survey of Canada as part of the Oak Ridges Moraine National Mapping Program (NATMAP) and hydrogeology project. This study was also supported by a Natural Science and Engineering Research Council research grant to R. W. C. Arnott.

Statement of Contribution

Assistance toward completion of the work presented in this thesis was obtained from the following people for the specified tasks.

Field Assistance: The following individuals provided field assistance through employment with the Geological Survey of Canada, Simone Dumas, Christopher Helmer, and Donald Cummings.

Core logging: Logging of core collected by the GSC and core donated by the Interim Waste Authority was completed at Golder and Associates temporary core-storage facility, Brampton, the old Ministry of Natural Resources compound on Dufferin Road, Maple, or the GSC Tunneys Pasture sample storage facility, Ottawa. Cores C34b-14, C34b-28 and C34b-5 were logged with the assistance of Tracy Brennand and David Sharpe. The remaining IWA core logged was completed with the assistance of Donald Cummings. Logging of GSC-Nob was completed with the assistance of David Sharpe, Donald Cummings, and Lucy Maurice.

Laboratory analysis: Grain size and total organic carbon analysis were completed by the sedimentology laboratory of the Geological Survey of Canada, under the Supervision of Patty Lindsay.

Database management: Archival data used in this project was obtained from the Oak Ridges Moraine NATMAP and hydrogeology project at the Geological Survey of Canada. Data management was provided at the GSC by Charles Logan.

Graphics: All images in this thesis were scanned by Rachelle Lacroix, Geological Survey of Canada. Figures 4.18 and 5.17 were drawn by John Glew.

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Appendix A-4_Photos. Photos for Gormley site

Appendix A-5_text. The complete thesis is located here in a PDF format

CHAPTER I

Introduction

1.1 Rational

Moraines form along ice-margins under a wide range of conditions and are morphologically and compositionally diverse. They extend up to 100s of kilometres in length and are commonly the largest depositional landforms in a glacial landscape. Because of their widespread occurrence and morphological and sedimentological variability there is general confusion concerning the origin and stratal characteristics of moraines. For example, are moraines composed of till (e.g. Karrow 1974; Mickelson et al. 1983), or stratified sediment (e.g. Hillaire-Marcel et al. 1981)? Much of this uncertainty relates to the general approach to mapping glacial deposits that is based largely on the dominant sediment characteristic at a shallow subsurface depth, commonly 1 m, and terrain analysis. In part, this can be attributed to the general poor exposure of unconsolidated sediment that makes up Pleistocene moraines and also the predominant interest of researchers in moraine geomorphology and their predisposed interest to studying glacigenic diamicton (till). This reflects a longstanding interest by researchers in the reconstruction of regional deglacial histories based on airphoto interpretation and till sheet stratigraphy (e.g. Karrow 1974). For example, in the over 100 year history of investigations of the Oak Ridges Moraine in southern Ontario, less than ~10 percent of all studies have investigated the stratigraphy of the moraine below several metres of the surface. As a consequence, the moraine has long been identified as an interlobate moraine based on geomorphology, surface geology and reconstructed regional ice-flow indicators (Chapman 1985; Chapman and Putnam 1984).

The inability of the proglacial environment to disperse debris from the glacier terminus faster than it is delivered results in a wide range of moraine types. Moraines are generally classified according to ice-front position (e.g. terminal), depositional environment (e.g. subaerial, subaqueous) and depositional processes (subaqueous fan, push moraines) (e.g. Benn and Evans 1998; Bennett and Glasser 1996). Where ice-margins terminate in a body of water and there is significant glacifluvial discharge, a variety of moraine types form, including subaqueous fans (Powell 1981; Rust and Romanelli 1975) and deltas (Price 1973). Where there are multiple efflux points and/or accommodation space is limited, polygenetic moraines form as deposition changes from subaqueous to deltaic (Barnett et al. 1998; Fyfe 1990). Subglacial or englacial conduit flow emerges from the glacial margin as a jet-efflux. The fluid dynamics of jet-effluxes have been investigated experimentally (Albertson et al. 1948) and applied to both deltas (Bates 1953; Wright 1977) and subaqueous fans (e.g.Gorrell and Shaw 1991; Powell 1990). Similarly, sediment facies and stratal architecture of delta and subaqueous deposits have been extensively studied in both marine (e.g. Powell 1990) and lacustrine environments (e.g. Gorrell and Shaw 1991). The character of these deposits, however, has been less extensively investigated within the context of large moraines (e.g. Fyfe 1990; Sharpe and Cowan 1990). Large moraines can be >200 m thick, > 10 km wide, and 100s km long (e.g. Barnett et al. 1998; Fyfe 1990; Sharpe and Cowan 1990), although both thickness and width may vary extensively along the moraine length. In fact, moraines are commonly discontinuous with individual moraine segments being 100s m to ten's km long, whereas the entire moraine length maybe 100s km long. Long moraine sections are commonly the result of the lateral coalescence of multiple discrete subaqueous fan or delta deposits (e.g. Fyfe 1990; Sharpe and Cowan 1990) or alternatively formed by broad sheet flood discharge deposits (Powell 1990; Sharpe and Cowan 1990). Additionally, moraines may extend several kilometres beyond their topographic expression beneath younger glacilacustrine basin strata and till sheets (e.g. White 1975).

Moraine formation has been interpreted to be controlled by variations in the rate of glacier retreat related to climate change (e.g. Hunter et al. 1996; Powell 1991), changes in ice sheet dynamics that are unrelated to climate (e.g. Hillaire-Marcel et al. 1981), and, or alternatively a combination of both factors (e.g. Fyfe 1990). Where moraines have consistent spacing and are small in size, annual climatic modulated deposition by seasonal melt is a plausible mechanism (e.g. Hunter et al. 1996). In the case of larger moraines, such as the Oak Ridges Moraine, due to a lack of detailed sedimentological information, the events controlling the depositional history are more poorly understood. Current understanding of large moraines suggests that they form in response to a number of events involving regional ice sheet dynamics (e.g. Barnett et al. 1998; Hillaire-Marcel et al. 1981; Sharpe and Cowan 1990). For example, Barnett et al. (1998) discuss four depositional stages for the Oak Ridges Moraine, each involving markedly different meltwater discharge fluxes ranging from large-scale jökulhlaup discharges, associated with tunnel channels to diurnal seasonal meltwater discharge and varve formation (see Gilbert, 1997).

Small to moderate meltwater discharges may be controlled by climatic-related seasonal ablation, anomalous precipitation events (Cowan et al. 1988; Warburton and Fenn 1994), or the draining of small supraglacial lakes (Kirkbride 1993; Russell 1993). By contrast, larger discharge events may be related to ice-sheet dynamics that involved the draining of large supraglacial reservoirs or subglacial reservoirs. Discharge may range from relatively small volumes of diurnal steady-state discharge to nonsteady-state large-scale reservoir meltwater releases that are regional in scale and significant in volume, for example 8.4×10^4 km^3 (Shaw 1996).

Over the past 10-25 years the sedimentological criteria to recognize such events have progressively emerged. For example, evidence of low energy suspension sedimentation in glacilacustrine environments has been widely documented (e.g. Ashley 1975). By contrast, depositional evidence of high energy episodic discharge events has been more poorly documented. Sedimentological studies of Recent (Holocene) and modern Icelandic outwash deposits, however, provide detailed lithofacies descriptions and facies associations correlated with seasonal discharge events or catastrophic jökulhlaups (Maizels 1989). Additionally, lithofacies deposited by hyperconcentrated dispersions (Powell 1990), supercritical flow (Brennand 1994), and hydraulic jumps (Gorrell and Shaw 1991) have been described from esker and subaqueous fan deposits.

Moraines also form an important component in the complex socio-economic framework of industrialized countries. During the past 20 years the fastest growing element of the Canadian mining industry has been aggregate extraction. Moraines in addition to eskers form two important sources of sorted sand and gravel used in the construction industry (e.g. Cowan 1984). Additionally, as upland areas, moraines serve an important role as the headwaters of streams and as regions of groundwater recharge (Gerber and Howard 1997; Hinton et al. 1998). Also, moraines commonly form the principal regional aquifers that provide potable water supplies for 10s to 100s thousands of people. As demand on the limited groundwater resource continues to increase, there is a need to understand better the size and distribution of the resource (Howard et al. 1995). Consequently, regional hydrogeological studies require the integration of geological data and depositional facies models to improve the development of regional hydrostratigraphic models, the determination of appropriate hydraulic parameters, and ultimately the construction and interpretation of reliable numerical flow models. The importance of such integration has been highlighted by Anderson (1989) for glacial deposits using sedimentological facies modelling techniques for conceptualizing large-scale hydrogeological trends. The lack of such integration has been observed (e.g. Fogg 1986; Fogg et al. 1998) to be a common problem with many site-specific and regional studies (LeGrand and Rosen 1998).

As noted earlier by Karrow (1973), no deposit as thick as large moraines will ever be fully understood from the study of 1-3 m deep road side-sections. To fully understand lithofacies variability, stratal geometry and the origin of large moraines, plus the implications for deglacial and hydrogeological models, details of moraine stratigraphy and depositional mechanisms must be improved. To date, however, only a small number of moraines have been investigated in sufficient detail, for example: Hartman Moraine (Sharpe and Cowan 1990; Sharpe et al. 1992); Salpausslka I Moraine (Fyfe 1990), and the Oak Ridges Moraine (Barnett et al. 1998).

1.2 Objectives

Because of the need for better stratigraphic and sedimentological control to constrain regional hydrogeological and stratigraphic models for the western Oak Ridges Moraine (ORM) this project was undertaken to complement the regional-scale Oak Ridges Moraine NATMAP and hydrogeology project conducted by the Geological Survey of Canada. Three principal objectives of the study were to:

i) *define the lithofacies of channel fill sediment beneath the moraine,*

thereby establishing the relationship between earlier tunnel channel events and formation of the moraine;

ii) *describe the lithofacies and stratal geometry of a subaqueous fan deposit in the moraine,*

thereby providing a framework to better understand facies variability and stratal geometry within the moraine and;

iii) *integrate objectives i and ii in order to understand the meltwater discharge flux responsible for moraine formation,*

thereby improving the understanding of the contribution of episodic jökulhlaups or seasonal meltwater discharges to moraine genesis.

1.3 The Study Area

The Oak Ridges Moraine is located in southern Ontario north of Lake Ontario (Fig. 1). The moraine extends ~160 km from the Niagara Escarpment eastward to Trenton. It forms the drainage divide between Lake Simcoe and Georgian Bay to the north, and Lake Ontario to the south. The study area is located in the western Oak Ridges Moraine, specifically that part of the moraine in the Humber River watershed (Fig. 1). The Humber River watershed is the largest watershed draining the southern flanks of the moraine, and extends 60 km from the Niagara Escarpment southward to Lake Ontario. The geological setting of the study area is reviewed in Chapter II.

1.4 Data

Data for this study consist of both, data collected by the author and archival data. These datasets consist of field mapping sites, core logs, geotechnical logs, and water well descriptions (Table 1.2). The focus of this study is on the sedimentology of a small number of sites, principally sections measured in two aggregate pits, three river sections, and eight cores (Fig. 1.2). The three river sections are located along the Humber River between Woodbridge and Bolton. Seven cores were obtained from the Interim Waste Authority (IWA) site investigations (e.g. Golder and Associates 1994) and relogged. An eighth core was collected by the Geological Survey of Canada (GSC) near Nobleton and logged. The archival datasets, field site descriptions, core descriptions, geotechnical data, and Ontario Ministry of Environment (MOE) water well data were used to provide a context for describing the spatial variability of moraine sediment in the study area and to permit mapping of the Oak Ridges Moraine stratal thickness (Fig. 1.3). Additional supplementary data consists of geological mapping by the Geological Survey of Canada (Russell and

White 1997; Sharpe and Barnett 1997) and Ontario Geological Survey (Barnett and Gwyn 1997; Cowan 1984; Karrow 1991) and a digital elevation model (Kenny et al. 1999). A small number of grain size analyses were completed and integrated with similar data from archival sources (Table 1.2).

1.5 Methods

A basin analysis approach (e.g. Miall 1984; Potter and Pettijohn 1963) was employed in this study of the Oak Ridges Moraine. The field program was conducted during the summers of 1994, and 1995, with minor data collection in the summer of 1996. Description of river and aggregate sections along with core from eight drillholes was carried out in bed-by-bed detail. These data are the basis of the depositional interpretations presented in this thesis. Moraine morphology was analyzed in a desktop GIS (MapInfo 1998; Vertical Mapper 1999) using a Digital Elevation Model and digital geological map coverage. Archival data were integrated with field data with the use of a simplified material code (Russell et al. 1998). Stratigraphic coding was completed manually in MapInfo and MSAccess (1997) using Borehole Mapper (Borehole Mapper 1998) as a viewing tool. This procedure permitted the production of the Oak Ridges Moraine stratal surface in MapInfo and Vertical Mapper. Grain size data were analyzed using standard sieve techniques and a Brinkmann particle size analyzer for the < 0.063 mm size particles (Lindsay et al. 1998). Archival grain size data from the IWA sites have been analyzed according to ASTM standards using standard sieve techniques and hydrometers (ASTM 1994). Although grain size results from different analytical techniques are known to differ no attempt has been made to quantify differences obtained by these two different techniques (e.g. Syvitski 1991). All grain size data were plotted and inclusive graphic parameters calculated in Granu (Boisvert, 1999). Total organic carbon analysis was completed using a Leco CR-12 Carbon Analyzer (Wang and Anderson 1998). Analysis for upward thinning or thickening trends within boreholes was completed using the runs up and run down (RUD) technique (e.g. Murray et al. 1996; Waldron 1987).

1.6 Thesis Organization

This thesis consists of six chapters, of which three are self contained chapters written in a paper format. Chapter two provides an overview of the background geology and glacial history of southern Ontario. Chapter three and four present lithofacies data of discrete depositional elements of the moraine. Chapter three discusses drillcore data from two tunnel channel fills that occur at the base of the Oak Ridges Moraine. Chapter four presents data from an aggregate pit and develops a model for the origin and sedimentation history of a high-energy subaqueous fan deposit. Chapter five incorporates data from chapters 3 and 4, plus data from other field and archival datasets, and landform analysis to discuss the formation of the western segment of the Oak Ridges Moraine. The final chapter summarizes the principal conclusions of the study.

Fig. 1.1 Location of the study area in southern Ontario with respect to other moraines in the area. Modified from Chapman and Putnam, (1984); Barnett, (1992). Eskers from Barnett et al. (1991).

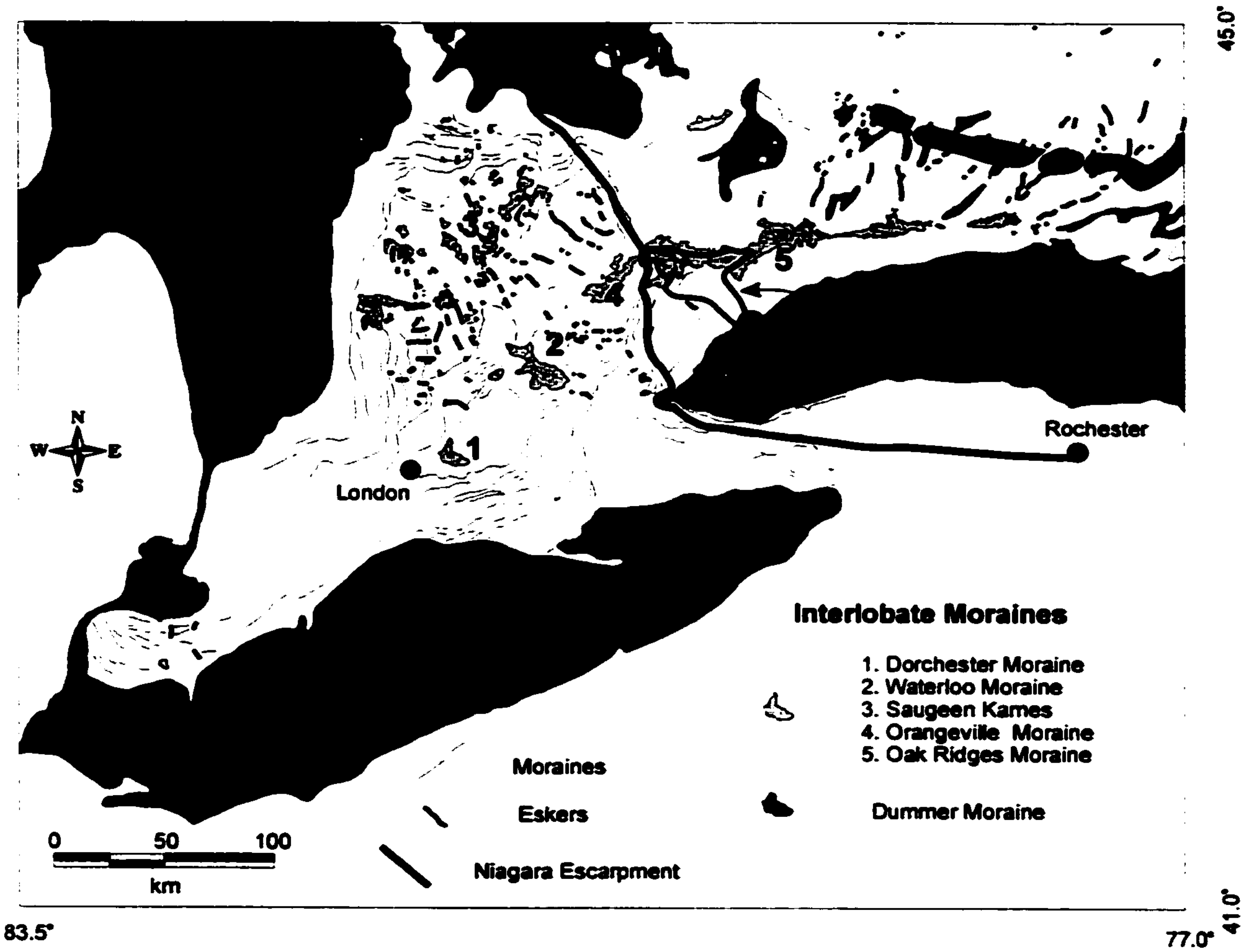

45.0°
5
2
1
N
W
E
S
London
Rochester
Interlobate Moraines
1. Dorchester Moraine
2. Waterloo Moraine
3. Saugeen Kames
4. Orangeville Moraine
5. Oak Ridges Moraine
Moraines
Eskers
Niagara Escarpment
Dummer Moraine
0 50 100
km
83.5°
77.0°
41.0°

Fig. 1.2: Location of principal study sites within the study area. Humber River watershed outlined and course of the Humber River shown.

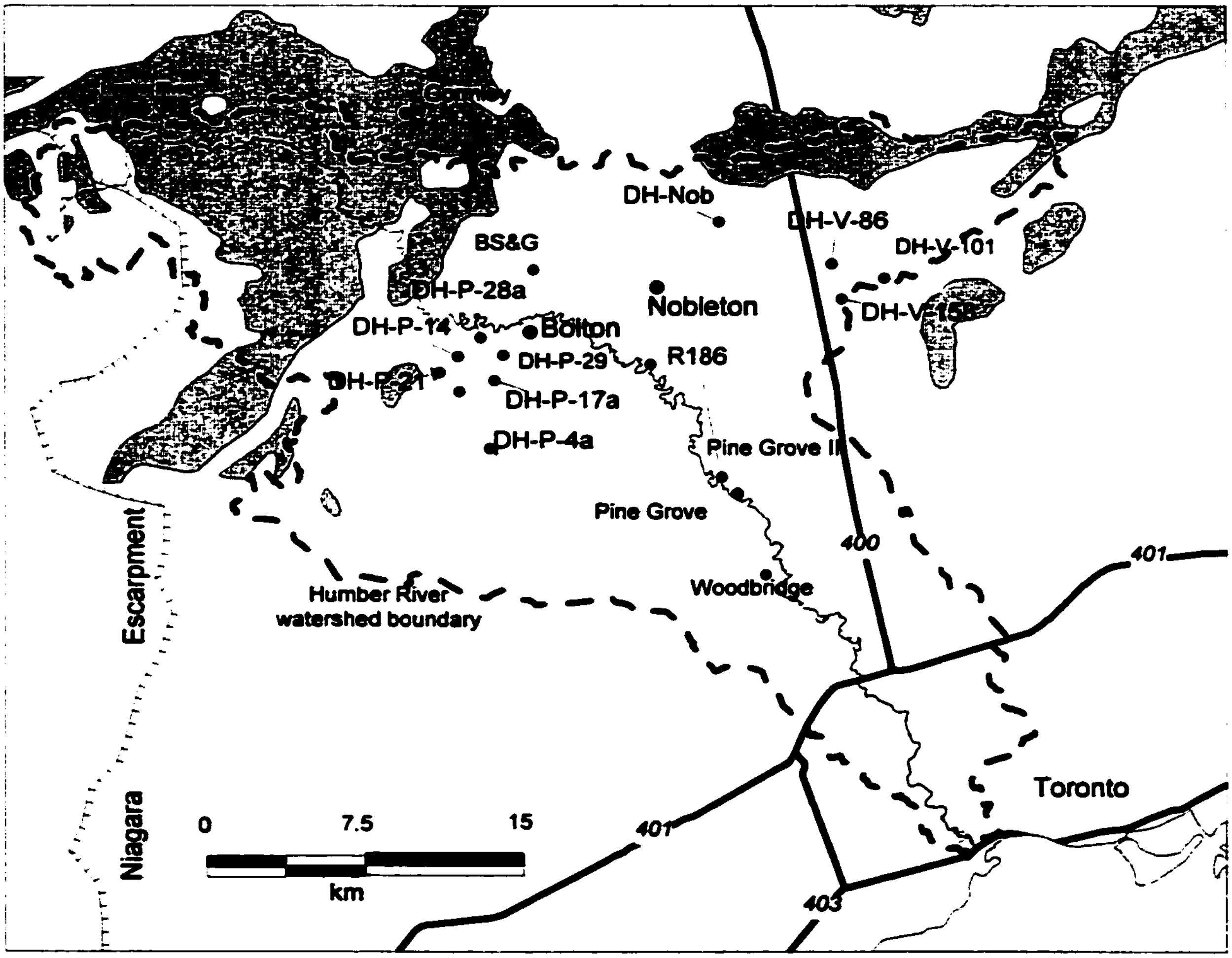

DH-Nob
DH-V-86
DH-V-101
BS&G
DH-P-28a
Nobleton
DH-P-14
Bolton
DH-P-29
R186
DH-P-21
DH-P-17a
DH-P-4a
Pine Grove II
Pine Grove
400
401
Humber River
watershed boundary
Woodbridge
Niagara
Escarpment
0
7.5
15
km
401
403
Toronto

Fig. 1.3. Location plots of secondary data sets used in this study to delineate the extent of the Oak Ridges Moraine.

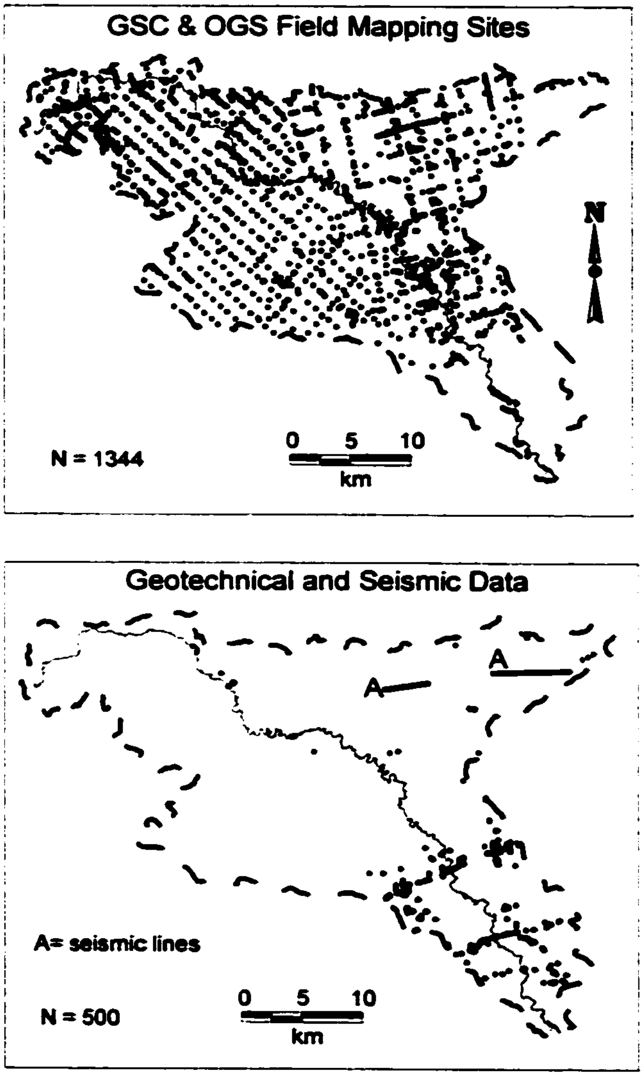

GSC & OGS Field Mapping Sites
N
N = 1344
0 5 10
km
Hydrogeology Drillhole Sites
A
B
A = Peel IWA sites 34b & 48
B= Vaughn IWA sites V3 & V4
N = 92
0 5 10
km
Geotechnical and Seismic Data
A
A
A= seismic lines
N = 500
0 5 10
km
MOE Water Well Data
N = 2028
0 5 10
km

Table 1.2: Listing of datasets and type of data used in this study

Category	Type	Source	Number sites	Texture	Structure	Grain-size	Depth top - bottom
this study	aggregate pit	Russell	2	yes	yes	yes	yes
this study	river section	Russell	3	yes	yes		yes
this study	field mapping	Russell	394	yes	yes		yes
this study	core logs	Russell	8	yes	yes	yes	yes
archival	field mapping	White (OGS)	1344	yes			yes
archival	core logs	Kelly (OGS)	4	yes			yes
archival	core logs	IWA	72	yes		yes	yes
archival	geotechnical	MTO	500	yes		yes	yes
archival	water wells	MOE	2028	yes			yes
published	Seismic	GSC	2				yes

Chapter II

Geologic Setting

2.1 Regional Geology

Southern Ontario is underlain by Paleozoic bedrock of Cambrian to Carboniferous age (Fig. 2.1) that forms part of the western St Lawrence Platform (Johnson et al. 1992). The region north of Lake Ontario occurs along the southern flank of the Algonquin Arch, a regional northeast-southwest trending structural high separating the Michigan Basin from the Allegheny Basin. To the west of Georgian Bay and Lake Ontario resistant Silurian dolostone of the Amabel Formation overlies more recessive strata of the Ordovician Queenston, and Georgian Bay formations (Johnson et al. 1992). Differential erosion of resistant dolostones and less competent underlying Ordovician shale, siltstone and interbedded limestone strata has formed the Niagara Escarpment (Fig. 2.1). This is a regional feature that can be traced from Rochester, New York northwestward to Manitoulin Island (Johnson et al. 1992). It forms the divide between Lake Ontario and Lake Erie wherein lake levels differ by ~100 m. The escarpment is cut by numerous reentrant valleys (e.g. Dundas, Credit River) that were formed, or enhanced by glacial and glacifluvial erosion (Kor and Cowell 1998; Straw 1968).

South of the Canadian Shield - Paleozoic margin (Fig. 2.1) glacigenic sediment thickness increases dramatically, reflecting the significant decrease in the erosional resistance of the Paleozoic strata. Consequently, southern Ontario is generally covered by up to 30 m (Karrow and Occhietti 1989) of Quaternary sediment (Fig. 2.2) that locally is up to 200 m thick (Barnett et al. 1991; Russell et al. 1998). The Quaternary succession consists of Illinoian and younger deposits. Pre Upper Wisconsinian strata, are generally poorly exposed except in deep, lake and river escarpments (Barnett 1992). From water well logs, water quality surveys and river bank outcrops, these strata are known to occur extensively across the Greater Toronto Area (Aravena and Wassenaar 1993; Karrow and Occhietti 1989). The oldest Quaternary strata in southern Ontario is the Illinoian York Till (Barnett 1992); however, this till is only recognized where it is overlain by Sangamonian deposits. Sangamonian Don Beds have been described from only two outcrops, the Don Valley brickyard and Woodbridge (Karrow and Occhietti 1989). Lower and Middle Wisconsinian strata are documented from some dozen sites with the most extensively described section cropping out in the Scarborough Bluffs along the shore of Lake Ontario. The Late Wisconsin maximum occurred at ~20 ka BP and deposited the Catfish Creek Till (Barnett 1992) and east of the Niagara Escarpment the Newmarket Till (Boyce et al. 1995). Overlying these diamictons and exposed extensively across southern Ontario is a predominantly glacilacustrine silt and silt-clay till (Fig. 2.2). Locally, however, areally restricted glacifluvial sand and gravel are associated with some moraines (Fig. 1.1, 2.2). Of particular note is a series of moraines that extend from London toward the northeast as far as Trenton (Fig. 2.2, Karrow and Paloschi 1996). The largest of these is the Oak Ridges Moraine, which extends from the Niagara Escarpment eastward for ~160 km to Trenton (Chapman and Putnam 1984).

2.2 Late Wisconsin Glacial History

The deglacial history of southern Ontario first outlined by Chapman and Putnam (1984) has been continually refined and most recently reviewed and updated by Karrow and Occhietti (1989) and Barnett (1992). This deglacial model provides a general history of ice sheet retreat with an emphasis on ice-marginal lakes. Consequently, the focus of the model is the documentation of ice-marginal positions and proglacial glacilacustrine sedimentation rather than Late-Wisconsin ice-sheet dynamics and glacial or glacifluvial processes. Following on work of Shaw and Gilbert (1990), Barnett (1992) integrated regional subglacial meltwater erosion of tunnel channels into the deglacial framework. Additional interpretations were presented by Shaw and Gilbert (1990) Who recognized large-scale meltwater events. These interpretations have not been integrated into models of regional deglaciation. Accordingly, the traditional deglacial model of Chapman and Putnam (1984) is reviewed first , followed by the meltwater model.

2.2.1 Deglacial model

The retreat of Laurentide ice from southern Ontario was not one of continuous northward migration but instead was characterized by the development of thicker ice in individual Great Lakes basins, local ice margin fluctuations and locally enhanced in situ wasting. Initial deglaciation was centered on high ground of the Niagara Cuesta in the interlobate position between Lake Erie, Lake Huron, Georgian Bay and Lake Ontario. Thin ice atop this bedrock high first melted near Orangeville (Chapman and Putnam 1984). From ~20-15 ka BP (Table 2.1) the Orangeville, Waterloo and Dorchester moraines were deposited in an interlobate glacilacustrine basin (Fig. 2.3). Ice margin fluctuations of Lake Erie, Lake Huron and Georgian Bay lobes prior to 14 ka BP are indicated by a succession of interbedded till and glacilacustrine and glacifluvial sediment (Table 2.1, Fig. 2.2). Where ice advanced over glacilacustrine silt the till is fine-grained. As ice margins fluctuated, each ice lobe constructed several moraines that form a highly complex stratigraphy (Fig. 1.1). These moraines have variable composition ranging from predominantly till to predominantly sand and gravel (Karrow and Occhietti 1989). Ponding of meltwater in the Lake Erie and Lake Huron basins formed a series of glacilacustrine lakes, Maumee I to Maumee IV, and subsequently Lake Whittlesey and Arkona. (Fig. 2.3). With progressive ice retreat at ~14 ka BP these lakes enlarged and drained into the Mississippi River system or southward across Michigan State (Barnett 1992). These and subsequent glacilacustrine lakes are significant because they permit shoreline trends and positions to be mapped, and lacustrine fauna dated. Furthermore, shoreline deposits permit the determination of the extent of ice-bounded and isostatic basins, which are key data used to calculate rates of post-glacial crustal rebound.

By ~14-13.5 ka BP the Ontario lobe had retreated to the Paris Moraine, a thick deposit of coarse gravel fronted by outwash channels (Cowan 1984; Fraser 1982) and overlain by a veneer of Wentworth Till (Karrow 1974). The Oak Ridges Moraine has been interpreted to have been deposited in an interlobate suture confined to the south and north by Ontario and Simcoe ice lobes and to the west by the Niagara Escarpment (Chapman and Putnam 1984). High glacilacustrine base levels were maintained by ice abutting against the Niagara Escarpment and outlet discharge that breached the escarpment in key channels (Chapman 1985). Diminished meltwater discharge resulted in a change to predominantly glacilacustrine and ice-marginal sedimentation and deposition of the Halton and Kettleby tills. Minor glacifluvial discharge was sustained from the Ontario Lobe as recorded by the Brampton Esker and isolated interbeds of sand and gravel within the Halton Till (Saunderson 1975). From about 14-13 ka, until the collapse of the Ontario lobe around 12 ka, the central part of the lobe may have been floating on a subglacial lake of ponded meltwater following earlier catastrophic discharges (Shoemaker 1991). Retreat of the Simcoe lobe and in situ ablation of the Ontario Lobe are marked by a series of ice-supported lakes following ~13 ka. South of the Oak Ridges Moraine, the Peel Ponds (White 1975) and subsequently Lake Iroquois formed, and north of the moraine Lake Schomberg (White 1975) and Lake Algonquin.

2.2.2 Meltwater events

Subglacial meltwater discharge from episodic reservoir drainage north of the Great Lakes Basin drained southward across the region. In the Georgian Bay area extensive bedrock erosion occurred (Kor and Cowell 1998; Kor et al. 1991). These regional meltwater events have been reviewed by Shaw and Gilbert (1990) and integrated into a regional outbreak event stratigraphy involving two major events, the Algonquin and Ontarian (Fig. 2.4). The Algonquin event is a south trending erosional discharge that most probably sculpted regional drumlin fields and incised tunnel channels. The later Ontario event is interpreted to have had a more westerly flow direction through the Lake Ontario and Lake Erie basins. Downflow of Georgian Bay, these flows deposited extensive coarse sediment in Michigan (Fisher and

Taylor 1999). In the western Lake Ontario basin erosion of Quaternary strata formed extensive erosional drumlin fields along the top of the Newmarket Till (Shaw and Sharpe 1987) and broad, steep-walled, southwest-trending tunnel channels (Barnett 1990; Sharpe et al. 1997; Sharpe et al. 1996). In the eastern Lake Ontario basin, on the other hand, the same erosional surface consists of sculptured bedrock and bedrock tunnel channels (Brennand and Shaw 1994; Pair 1997; Shaw and Sharpe 1987). Within central Lake Ontario and Lake Erie, a very thin stratal succession is present, in addition to drumlins and scoured bedrock (Lewis et al. 1999; Lewis et al. 1996; Lewis et al. 1997). These features can also be traced across northern New York State (Pair 1997) and into the Finger Lakes region (Mullins et al. 1996) and have been interpreted to have formed during a single meltwater event (Shaw and Gilbert 1990). Dating of these events is poorly constrained, but tunnel channels in the Finger Lakes of New York State, located up flow from the Valley Head Moraine, have been dated at ~14.5 to 14.0 ka (Mullins et al. 1996). These dates compare favourably with the estimated time of formation of the Oak Ridges Moraine around 13.5 ka BP (Barnett 1992). Using these time estimates and sedimentological evidence Barnett et al. (1998) suggested that the tunnel channel fill and moraine formation were related. The sequence of regional meltwater events that formed drumlin fields and tunnel channels and deposited various moraines illustrates the general lack of understanding of the deglacial event-sequence of southern Ontario and problems of integrating emerging meltwater evidence within the more established deglacial models.

2.3 Geology of the Greater Toronto Area

To facilitate regional subsurface mapping in the Oak Ridges Moraine and Greater Toronto (GTA) areas a regional stratigraphic framework was developed by Sharpe et al. (Sharpe et al. 1996, 1999) based on regional surficial geological mapping (Fig. 2.5), drillhole, and seismic studies (Pullan et al. 1994). This framework amalgamates all deposits between bedrock and overlying Newmarket Till into a single stratal unit informally termed the lower deposits (Table 2.2). Overlying the lower deposits, the model consists of four elements (Fig. 2.6): Newmarket Till, regional unconformity, Oak Ridges Moraine, and Halton Till. Each of the elements is reviewed briefly below followed by a review of the Oak Ridges Moraine depositional model developed recently by Barnett et al. (1998). Recent deposits, such as alluvium, are not explicitly treated in the framework because they are mostly restricted to river valleys and are generally < 5 m thick.

2.3.1 Bedrock surface

The bedrock surface is a regional unconformity that is not exposed in southern Ontario but is exposed to the north where it is concident with exposed surface of the Precambrian Shield (Fig. 2.1). Physiographically the surface consists of a number of buried and exposed escarpments and valleys (Karrow and Occhietti 1989). The most prominent topographic feature is the Niagara Escarpment, that crops out along the western border of the area. Immediately to the east is the Laurentian Channel, a buried bedrock valley that has been suggested to have connected Georgian Bay and Lake Ontario prior to the Quaternary (Eyles et al. 1985; Spencer 1890). Along the north shore of Lake Ontario a number of smaller valleys rise to the north from Lake Ontario (Brennand et al. 1997).

2.3.2 Lower deposits

This package of sediment consists of ~7 stratigraphic units described primarily from exposures in the Scarborough Bluffs and Don Valley Brickyards. These units are regionally extensive and are composed of a wide variety of lithofacies. The lowest unit is the York Till which has been identified in outcrop (Karrow 1967; White 1975) and tentatively correlated with strata in drillhole core (Russell and Pullan 1998; Sado et al. 1983).This silty sand till is overlain by silt and sand of the Don Formation. The Don Formation has been mapped along the Don Valley, in outcrop near Woodbridge (White 1975) and in core from the Nobleton area (Russell and Pullan 1998). On the basis of macrofossils, peat, and pollen in sand

and silt these sediment are assigned to the Sangamonian interglacial, (e.g. Eyles and Williams 1992; Karrow 1967). The Don Formation is overlain by the Scarborough Formation which consists of a lower silt-clay unit overlain by a sand unit (Karrow 1967). This formation has been correlated regionally on the basis of drillhole geophysics (e.g. Eyles et al. 1985), and the presence of abundant organic debris (e.g. Sharpe and Barnett 1997a). High dissolved organic contents in water wells (Aravena and Wassenaar 1993) can be used to infer either Don or Scarborough formations at depth. The Scarborough is interpreted to be a braided outwash and deltaic deposit (Eyles and Eyles 1983; Kelly and Martini 1986). The Sunnybrook Till is a regionally extensive < 30 m thick silt diamicton unit that crops out along the Scarborough Bluffs (Eyles et al. 1985; Karrow 1967). The depositional environment of this unit has been much debated and may be of glacial (Karrow 1967) or lacustrine origin (Eyles et al. 1985). The Scarborough Formation is in turn overlain by the Thorncliffe Formation. It has been most extensively mapped along the West Duffins Creek and consists mostly of sand (Sharpe and Barnett 1997a). In contrast, to the northwest beneath the Oak Ridges Moraine, it consists mostly of silt-clay rhythmites that are bioturbated along bedding planes (Russell and Pullan 1998). The Thorncliffe Formation has been variously interpreted as a deltaic deposit(e.g. Eyles and Williams 1992) or as a subaqueous fan deposit (Sharpe and Barnett 1997a).

2.3.3 Newmarket Till

Newmarket Till is exposed north of the Oak Ridges Moraine (Gwyn and DiLabio, 1973), and to the south (Sharpe et al. 1997). Beneath the Oak Ridges Moraine and Halton Till it has been intersected by boreholes, river sections, and interpreted from seismic data (Pugin et al. 1999). The base of the unit typically occurs between 200 and 220 m a.s.l. (above sea level) within the Uxbridge wedge (Sharpe et al. 1996). It is a dense, stony, silty sand till, that is 5-30 m thick in outcrop. Discontinuous sand horizons < 5 m thick and cobble-boulder horizons are common. In addition Newmarket Till is characterized by high seismic velocities, generally > 2000 m s^{-1} (Boyce et al. 1995; Pugin et al. 1999).

2.3.4 Regional unconformity

This surface is best exposed north of the Oak Ridges Moraine (Sharpe et al. 1996), where it overlies Newmarket Till. Erosional features consist of drumlins and tunnel channels (Sharpe et al. 1996). Based on downhole geophysics and seismic reflection surveys, the unconformity has been shown to extend southward beneath the Oak Ridges Moraine (Gwyn and Cowan 1978; Pugin et al. 1999). Where the Newmarket Till has been eroded the surface overlies lower deposits. In this case it is recognizable on the basis of distinct changes in organic carbon content, macrofossil, trace fossil assemblages, lithofacies, and geophysical properties (Eyles et al. 1985).

2.3.5 Oak Ridges Moraine

The Oak Ridges Moraine is a large polygenetic landform extending for 160 km eastward from the Niagara Escarpment (Barnett et al. 1998). It ranges from 0.1 to 10 km wide and where it overlies tunnel channels incised along the regional unconformity it may be up to 150-180 m thick. The moraine consists predominantly of gravel, sand, and silt (Barnett et al. 1998; Duckworth 1979; Gilbert 1997; Paterson and Cheel 1997). Based on surface morphology the Oak Ridges has long been interpreted as an interlobate moraine (Chapman and Putnam 1984). The majority of investigations of the Oak Ridges Moraine have been morphological with only a small number of sedimentological studies (Barnett et al. 1998; Duckworth 1979; Gilbert 1997; Paterson and Cheel 1997) (Table 2.3). The emphasis on morphology is largely because of the recessive erosional nature of sand-rich, near-surface strata of the moraine. Only in the past 30 years, principally because of extensive aggregate extraction and drilling, have sedimentological studies been possible. Duckworth (1979) was the first to apply modern sedimentological analysis to the western end of the moraine and concluded that the moraine was of braided fluvial origin and not an interlobate feature. More recent work, however, has suggested that the moraine is a polygenetic landform (Barnett et al. 1998) composed of subaqueous fan deposits (e.g. Paterson and Cheel 1997), basin sediment (Gilbert 1997) and that only the youngest strata were deposited by subaerial fluvial processes

(Barnett et al. 1998). The origin and depositional history of the Oak Ridges Moraine is discussed in section 2.4.

2.3.6 Halton Till

This unit includes the Halton and Wildfield tills south of the Oak Ridges Moraine and the Kettleby Till to the north of the moraine. South of the moraine, the Halton Till extends from Hamilton (Karrow 1963) northward across the GTA along the southern flanks of the Oak Ridges Moraine (Sharpe et al. 1997; White 1975). Where it onlaps the Oak Ridges Moraine the terrain typically exhibits a hummocky topography (White 1975). On the north side it is generally restricted to the Alliston and Newmarket map areas. Within the southern Humber River watershed it is < 50 m thick, but eastward thins to generally < 10 m in thickness (Russell and Arnott 1997). The Halton Till consists of silt to clay-silt till interbedded with lacustrine sediment and glacifluvial sand. The diamicton has been interpreted as subglacial debris flow and rain-out deposits (e.g. Russell and Arnott 1997; Sharpe et al. 1997).

2.4 Depositional Model of the Oak Ridges Moraine

Recent work by Barnett et al. (1998) suggests that the moraine is a polygenetic landform that formed in four discrete depositional stages: subglacial, subaqueous fan, deltaic and ice-marginal sedimentation (Fig. 2.8).

2.4.1 Stage I: subglacial sedimentation

Coarse gravel and thick deposits of sand were deposited by waning meltwater flows along the tunnel channel network (Fig. 2.8). Deposition may have started during the waning stages of the erosional event that formed the tunnel channels or from flows that reoccupied pre-existing channels. Channel deposits consist of large mesoforms up to 10 m high and wavelengths of 200 m (Shaw and Gorrell 1990); macroforms with similar dimensions have been documented from the Spokane outbreak floods of Washington (e.g. Baker 1978). Similarly, clinoforms imaged on seismic reflection data from channel fills beneath the Moraine also have comparable dimensions (Pugin et al. 1999). Several metre-high mesoforms indicate relatively deep flows of 30-40 m or more and paleohydraulic calculations from boulders in these deposits suggest flow speeds of 2-10 m s^{-1} (Shaw and Gorrell 1990).

2.4.2 Stage II: subaqueous fan sedimentation

Gravel and sand deposits, particularly massive sand and gravel deposits in addition to climbing bedforms are key facies described from sections in aggregate pits in the central part of the Oak Ridges Moraine (e.g. Paterson and Cheel 1997). In conjunction with the absence of topset and foreset beds, and the extensive evidence of rapid flow expansion (e.g. climbing small-scale cross-stratification) these strata have been interpreted as subaqueous fan deposits formed where a subglacial conduit entered a glacilacustrine environment (Fig. 2.8). Proximal sand and gravel rich subaqueous fan deposits grade rapidly into a rhythmic silt-dominated basin succession that locally consists of ~100 varves (Gilbert 1997).

2.4.3 Stage III: subaqueous fan to delta sedimentation

Broad, low-relief landforms consisting of sand and gravel strata, and finer silt-clay rhythmites have been interpreted as proximal and distal delta deposits respectively. Deltas formed due to a local fall of baselevel or because aggrading subaqueous fans locally filled the basin (Fig. 2.8). Elsewhere local subaqueous fan sedimentation continued.

2.4.4 Stage IV: ice-marginal sedimentation

Thick successions of interbedded massive diamicton, laminated silt and sand, and gravel have been interpreted as glacilacustrine ice-marginal deposits (Fig. 2.8). Meltwater discharge was dominated by diurnal meltwater fluxe and low rates of sedimentation. Intermittent grounding of glacier ice restricted subglacial glacifluvial discharge to conduits and deposition of isolated glacifluvial deposits (e.g.Brampton esker Saunderson 1976) that form interbeds in the thick glacilacustrine succession termed the Halton Till (Karrow 1963; Russell and Arnott 1997; Sharpe 1987; White 1975).

2.5 Geological Setting of the Humber River Watershed

Most of the western Oak Ridges Moraine occurs in the Humber River watershed, which is the focus of this study. The watershed covers an area of 900 km^2 (Fig. 2.9) and is the largest watershed in the GTA. It drains from headwaters along the Niagara Escarpment and flows southward into Lake Ontario. Detailed geological investigations in the area have been limited. The most extensive publically available work are regional surficial geological maps (1:50,000 scale Barnett and Gwyn 1997; Cowan 1976; Russell and White 1997; Sharpe and Barnett 1997b; White 1973; White 1975). In addition, local subsurface site investigations were conducted during the early 1990s for the Interim Waste Authority by R. Blair (Golder and Associates 1994) and by M. Gomer (Fenco MacLaren 1994). These studies focused primarily on the vertical changes in sediment grain size with little attention to other sedimentological characteristics. The following reviews watershed physiography, bedrock topography and the thickness and lithofacies of the key stratal units in the lower deposits, Newmarket Till and Halton complex. The Oak Ridges Moraine, glacifluvial, glacilacustrine, and alluvial units, are summarized in (Table 2.2).

2.5.1 Physiography and landforms

The headwater of the main branch of the Humber River rises along the Niagara Escarpment in the western margin of the watershed at 390 to 440 m a.s.l (Fig. 2.9). Along the crest of the escarpment are a number of small moraines, including the Singhampton and Gibralter moraines. East of the escarpment and north of the East Humber River are two areas of hummocky topography associated with the Oak Ridges Moraine (Unit 5, Fig. 2.5, 2.9). These areas are also extensively dissected by modern fluvial erosion. On the basis of the digital elevation model it is difficult to determine whether previously mapped hummocky terrain (White 1973) is a depositional landform element or the result of post-depositional erosion. South of the Oak Ridges Moraine an area of low-relief Halton Till has been incised by post-glacial drainage. This is due to changes in base level associated with Lake Iroquois and subsequent Lake Ontario water levels.

2.5.2 Bedrock geology, topography and drift thickness

The Ordovician Queenston and Georgian Bay formations (Liberty 1969), subcrop at the base of the Niagara Escarpment and crop out along the Humber River (White 1975). Silurian Amabel Formation and the Clinton and Cataract groups form the Niagara Escarpment (Sanford and Baer 1981). The Paleozoic bedrock has a highly variable surface topography (Fig. 2.10b). Some tributary valleys can be traced from the escarpment eastward to a buried bedrock valley extending from Georgian Bay to Lake Ontario, previously named the Laurentian Channel by Spencer (1890; see also Brennand et al. 1997; Eyles et al. 1985). Relief between the base of the Laurentian Channel and the crest of the Niagara Escarpment is ~300 m. Quaternary sediment is up to 200 m thick in the northeast corner of the watershed (Fig. 2.10b). The sediment is thickest in the northeast where the Oak Ridges Moraine overlies the Laurentian Channel; it thins both toward the Niagara Escarpment and southward toward the Humber River. Southwest of the Humber River a broad bedrock high commonly has a sediment cover of < 20 m.

2.5.3 Lower deposits

These strata have long been recognized to outcrop in a railroad embankment near Woodbridge (White 1975). Based on their stratigraphic position, organic content, and fossil content, York Till, Don Beds and Scarborough Formation sand have been described from this section (White 1975) or locally from drilling

(Kelly 1994).These strata, in addition to the Thorncliffe Formation, are also present in core DH-Nob (Russell and Pullan 1998). Thorncliffe Formation rhythmites have also been recognized during field mapping in river bank outcrop from east of Kleinburg, and possibly along the bottom of the Humber River. The continuous core from DH-Nob provides the most complete record of the lower deposits in the watershed and the following description is of that core (Fig. 2.1; Russell and Pullan 1998).

The York till overlies bedrock and is an~5 m thick unit of silt-sand till with ~10 % pebble size clasts. Near the base of the unit clasts are predominantly shale but igneous and metamorphic lithologies become more common upward. The till is mostly massive with minor sand interbeds and based on texture, and massive structure is interpreted to be a lodgement till (Russell and Pullan 1998). As it underlies Sangamonian strata it has been interpreted to be deposited during the Illinoian Stadial (White 1975).

The Don Formation abruptly overlies the York till and is ~23 m thick and consists of ~60 units that are up to 30 cm thick. It consists of pebble gravel, cross-bedded sand and silty fine sand. Organic material consists of shells, wood fragments, and mm-long detrital organic fibres dispersed in fine to coarse sand (Fig. 2.12). The Don Formation is interpreted to represent deposition in a shallow, high-energy, littoral environment (Eyles and Clark 1988).

The Scarborough Formation is ~39 m thick and consists of ~1260 units of fine sand to clay. Fine sand and silt are horizontally micro-laminated or small-scale cross-laminated and detrital organic debris is commonly dispersed throughout (Fig. 2.12). The succession has an overall upward-fining trend with secondary upward-coarsening cycles. On geophysical logs, particularly the gamma and conductivity logs, a well-defined asymmetric signature is interpreted to indicate an upward-coarsening trend (Fig. 2.11). Strata of the Scarborough Formation are interpreted to have been deposited in a proximal deltaic environment.

The Thorncliffe Formation is 65 m thick and consists of ~2300 units of fine sand, silt or clay (Fig. 2.13). The fine sand-silt part of couplets average ~5 cm thick and are capped by clay strata generally < 1 cm thick. Sedimentary structures are dominated by horizontal micro-laminations and small-scale cross laminated fine sand. Horizontal trace fossils are common along silt and clay bedding planes. The unit comprises a number of stacked upward coarsening cycles that become sandier and consist of thicker varves upward. This trend is well defined on borehole geophysical logs (Fig. 2.11). Based on lithofacies and lithofacies-association these strata are interpreted to have been deposited in a distal proglacial delta setting.

2.5.4 Newmarket Till

Newmarket Till crops out in the lower southeastern part of the watershed, locally along the Humber River (e.g. Kleinburg), and atop Mount Wolfe. Where observed in section the lower contact was slump covered and therefore the thickness is unknown, but it is at least 3 m thick. On the basis of seismic data Newmarket Till has been interpreted to be present across the northeastern corner of the watershed and to be up to 20 m thick (Pugin et al. 1999). The Newmarket Till is a dense silt-sand diamicton with 10 to 15% pebbles. Gravel clasts are mostly Paleozoic carbonate and shale with minor crystalline bedrock lithologies (Fig. 2.14). A possible correlative facies of the Newmarket Till occurs north of Mono Mills where the till is characterized by a red matrix with abundant dispersed Queenston Shale clasts. The Newmarket Till has commonly been interpreted to be lodgement till [Sharpe, 1997 #439].

2.5.5 Halton complex

The Halton Till is considered by Barnett et al. (1998) to be the final pulse of Oak Ridges Moraine sedimentation (Fig. 2.6, 2.8). Consequently, it is discussed here in some detail to highlight the important changes in sedimentation history in Late Wisconsinian ice-margin dynamics during the final stage of Oak Ridges Moraine deposition. Although it is commonly referred to as the Halton Till, it in fact consists of an interstratified succession of diamicton, sand and gravel, and silt-clay (Russell and Arnott, 1997). Within the Humber River watershed the Halton complex (defined herein) consists of three formal stratigraphic units, the Wentworth, Halton and Wildfield tills and associated interstratified sediment.

Subsurface work in the West Humber River watershed has shown the Halton complex to be up to 52 m thick (Golder Associates 1994a). The sandy-silt Wentworth Till is the lowest till unit and has been mapped from excavations by White (1975), and interpreted to be a coarse layer at the base of the Halton Till (Sharpe 1987). The silt-rich Halton Till was originally defined in the Hamilton area (Karrow 1963) and subsequently correlated with strata in the Bolton (White 1975) and Port Hope areas (Martini et al. 1984). In the Humber River watershed, the younger, fine-grained silt-clay Wildfield Till (White 1975) has been mapped southwest of Bolton. In the Hamilton area the Wildfield till was not differentiated from Halton Till (cf. Karrow 1963). The Halton complex forms a low-relief plain south of the Oak Ridges Moraine (Fig. 2.6, Chapman and Putnam 1984).

Diamicton lithofacies. This is the most common lithofacies and crops out extensively in the area (Figs. 2.9, 2.10). The lower contact is locally sharp, interbedded or gradational with underlying silt and sand. It is a massive to faintly laminated clay-silt to sand-silt diamicton with less than 20 % gravel clasts (cf. Golder Associates 1994), of which only 2% are coarser than granule size (Fig. 2.15). Pebbles are mostly subrounded to subangular and consist of carbonate, siltstone, shale, and granite gneiss. Intraclasts, although uncommon, are composed of 2-5 mm long silt and clay rip-ups and less commonly 5-10 cm sand clasts. Sharp-based, erosionally truncated, 40 cm long by 30 cm thick, cross-stratified sand lens crop out in river sections (Fig. 2.15). Stratigraphically upward the facies becomes more massive and finer-grained with a decrease in gravel content.

Sand and gravel lithofacies. The sand and gravel lithofacies occurs as discontinuous layers consisting of ripple-drift-laminated fine sand to medium-scale cross-bedded and planar-laminated pebbly coarse sand. In addition, there is a coarser subfacies composed of matrix- to clast-supported cross-bedded pebble and cobble that form beds up to 30 cm thick. The lower contact, is typically marked by an abrupt transition from diamicton. Intraclasts consist of 2-4 mm long clay clasts and less commonly 1-2 cm long diamicton clasts. Deformation is common and includes normal and reverse faults with 1-2 cm offsets and minor folds. Paleoflow was commonly toward the northwest to northeast. Minor organic detritus occurs in some beds.

Silt - clay lithofacies. This facies consists of fining-upward silt to clay rhythmites. Clay laminae are massive and commonly ~5 mm thick, rarely exceeding 1 cm. Silt deposits vary from 1 mm to 3 cm thick. Silt laminae are sharp-based and commonly have minor sand just above the basal contact. Rare gravel clasts are up to 2-3 cm in diameter.

The Halton complex was deposited during the latest stages of deglaciation of Lake Ontario(Karrow 1974). Diamicton is interpreted to have been deposited by a number of mechanisms, including debris flow, ice-proximal rain-out, and subglacial processes. The occurrence of isolated thin beds of diamicton is interpreted to mark the change from glacifluvial-deltaic (stage III) to ice-marginal (stage IV) sedimentation (Fig. 2.8). This is supported by paleoflow measurements in the sand and gravel strata along the Humber River, where paleoflow direction changes from dominantly westerly in stage III to northerly within stage IV (Halton complex). Sand and gravel is interpreted to have been deposited within subglacial conduits or as underflows in an ice-proximal setting. Silt-clay rhythmites highlight the variable dynamics of the glacial lake basin and indicate periods of low-energy sedimentation temporarily interrupted by higher energy events. The thickness of the Halton complex was probably influenced by the pre-existing lake bottom physiography, whereby thicker stratal successions accumulated in topographic lows. In the upper Humber River watershed, deposition of Halton complex indicates decreased meltwater discharge from sources to the north and south, and a change to predominantly ice marginal depositional processes within a glacilacustrine grounding-line zone.

Fig. 2.1. Simplified bedrock geology of southern Ontario classified by stratigraphic period. Digital data replotted from OGS (1993).

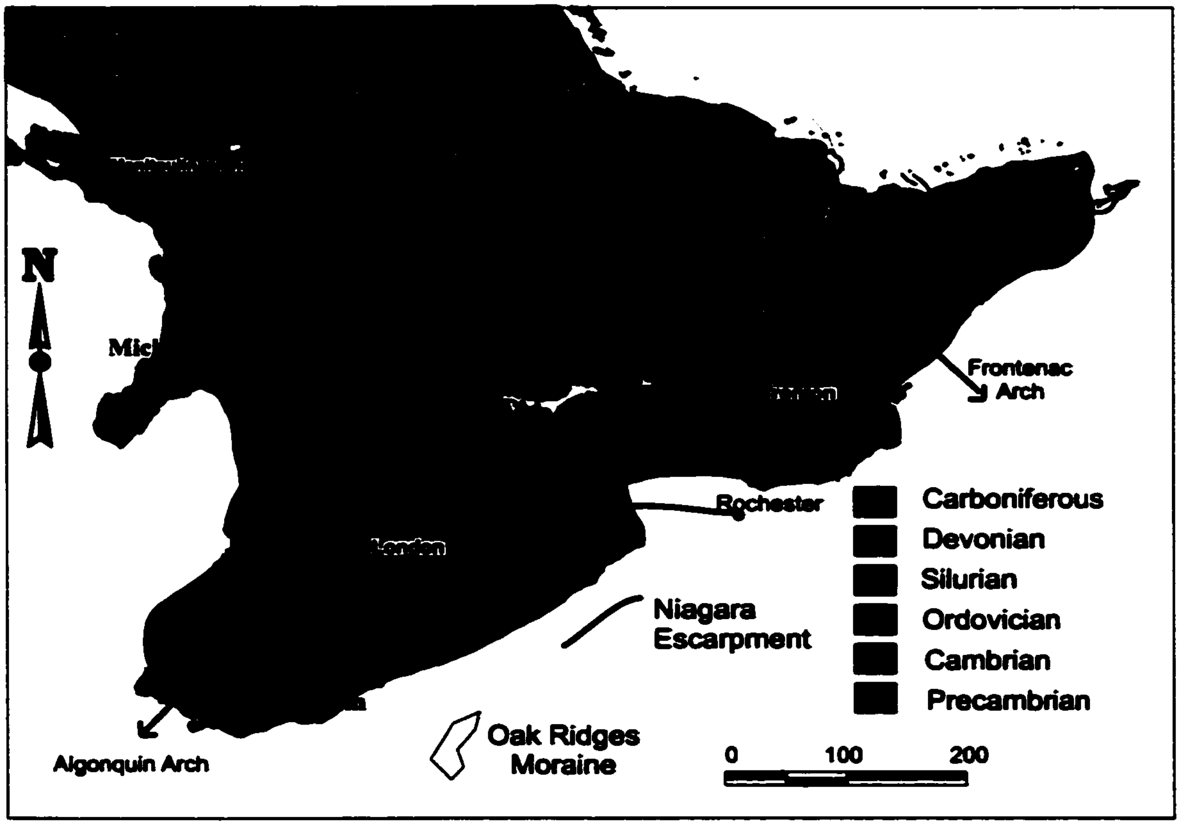

N
Frontenac
Arch
Rochester
Carboniferous
Devonian
Silurian
Ordovician
Cambrian
Precambrian
Niagara
Escarpment
Oak Ridges
Moraine
Algonquin Arch
0
100
200

Fig. 2.2. Surficial geology of southern Ontario, replotted from OGS digital dataset 17 (OGS, 1997)

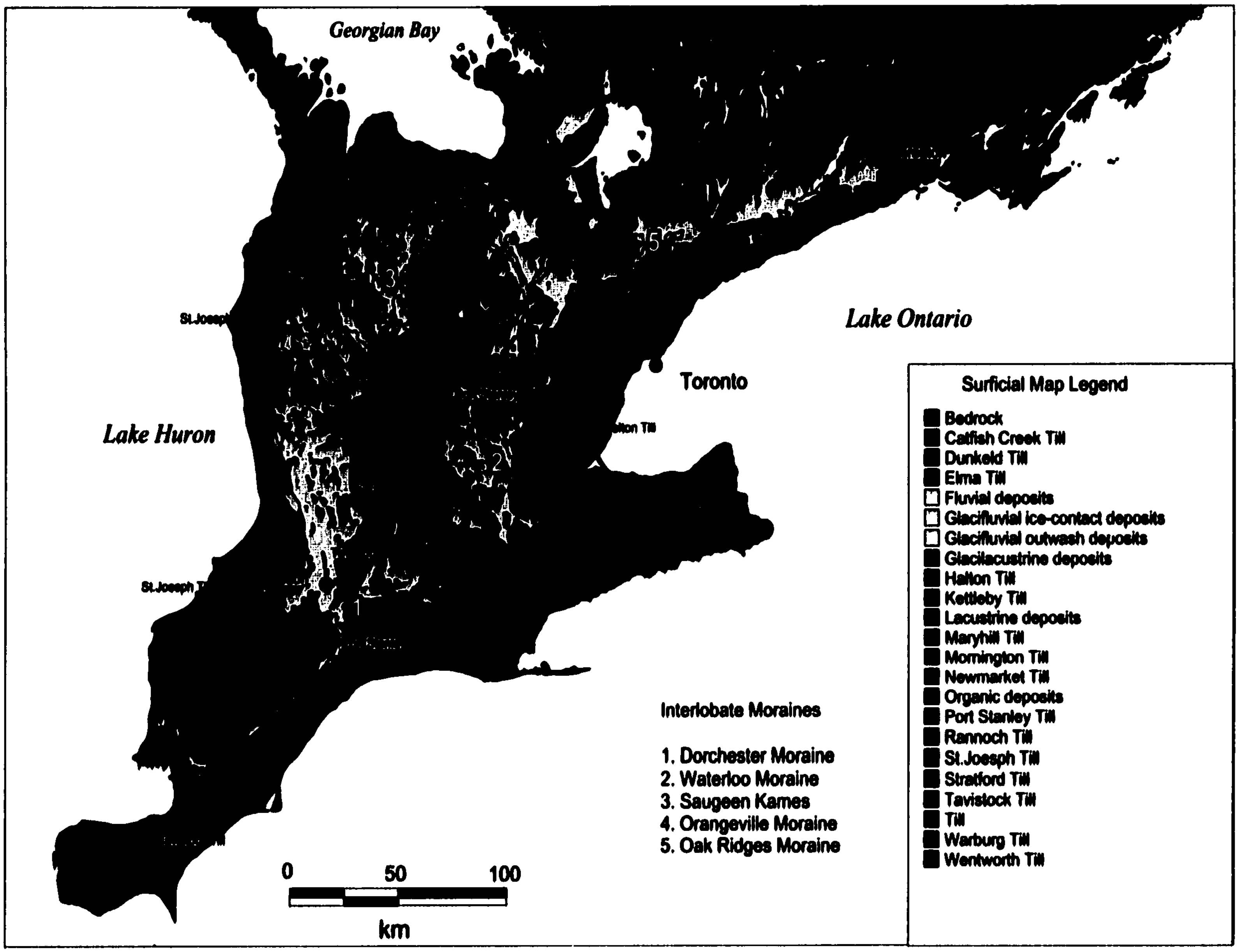

Georgian Bay
Lake Huron
Lake Ontario
Toronto
St.Joseph
St.Joseph Till
Surficial Map Legend
Bedrock
Catfish Creek Till
Dunkeld Till
Elma Till
Fluvial deposits
Glacifluvial ice-contact deposits
Glacifluvial outwash deposits
Glacilacustrine deposits
Halton Till
Kettleby Till
Lacustrine deposits
Maryhill Till
Mornington Till
Newmarket Till
Organic deposits
Port Stanley Till
Rannoch Till
St.Joesph Till
Stratford Till
Tavistock Till
Till
Warburg Till
Wentworth Till
Interlobate Moraines
1. Dorchester Moraine
2. Waterloo Moraine
3. Saugeen Kames
4. Orangeville Moraine
5. Oak Ridges Moraine
0
50
100
km

Fig. 2.3. Selected stages of deglaciation of southern Ontario, from Chapman and Putnam (1955). (a) Formation of the Orangeville and Waterloo moraines. (b) Initiation of Ontario Island. (c) Lake Maumee stage and formation of Singhampton Moraine. (d) Formation of late stages of Oak Ridges Moraine.

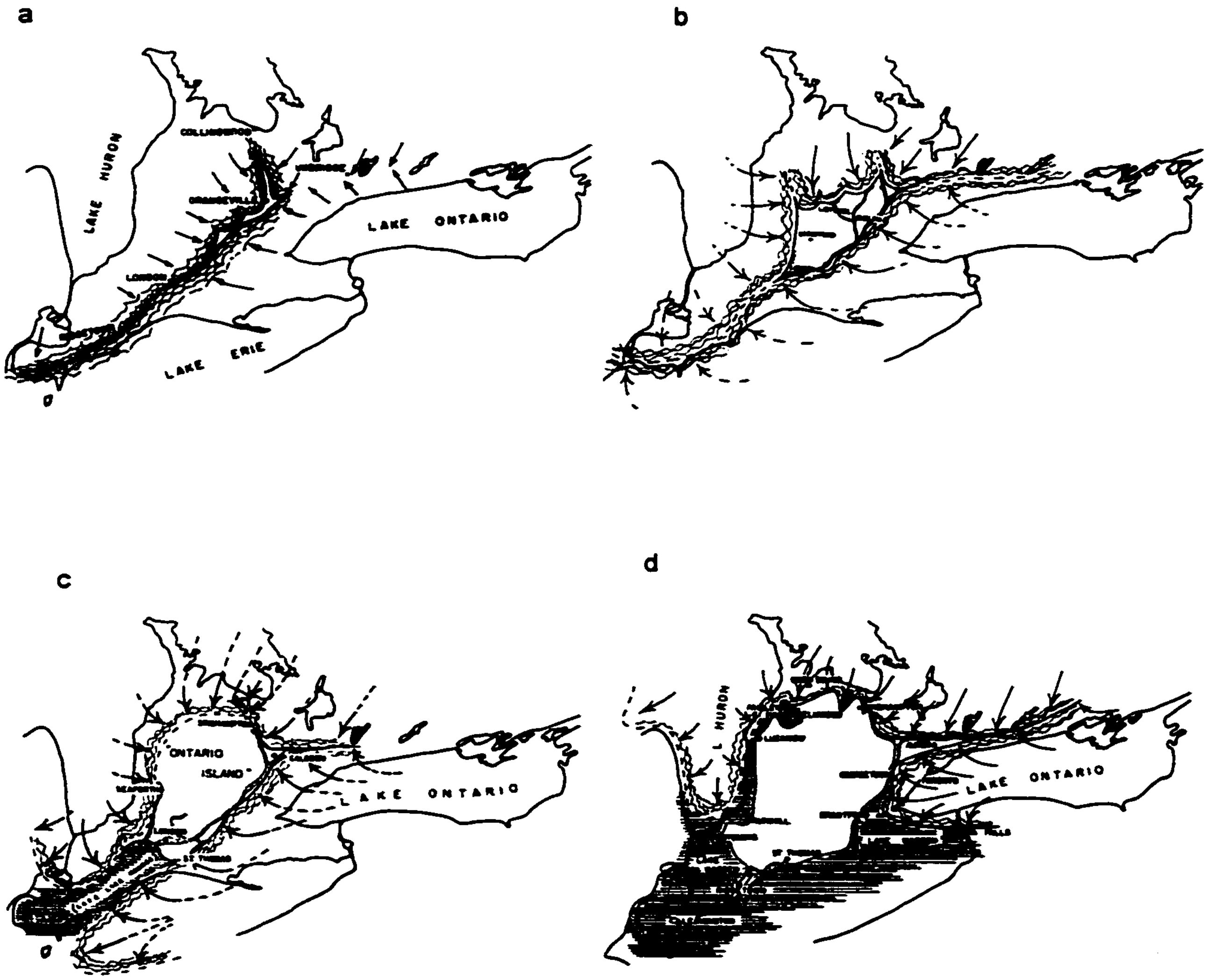
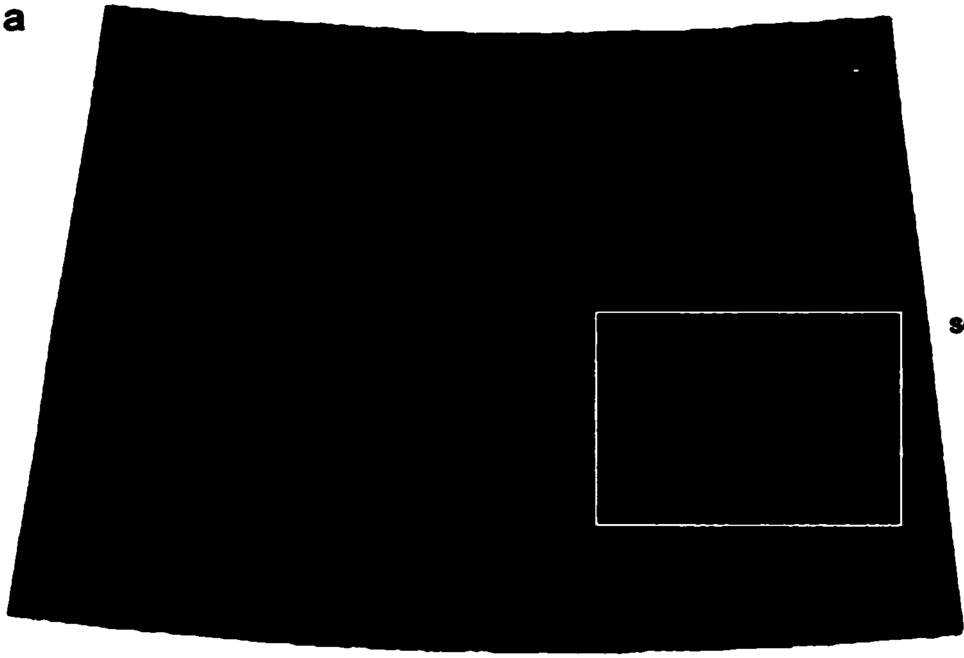

a
b
c
d
LAKE HURON
LAKE ONTARIO
LAKE ERIE
ONTARIO ISLAND
LAKE ONTARIO
L HURON
LAKE ONTARIO

Fig. 2.4. (a) Digital Elevation Model of Great Lakes region, red is high and blue is low elevations (from Lewis et al. 1999). (b) Proposed meltwater flow paths for the Late Wisconsin erosional outbreak meltwater events across southern Ontario, (from Shaw and Gilbert 1990).

a



see b

b

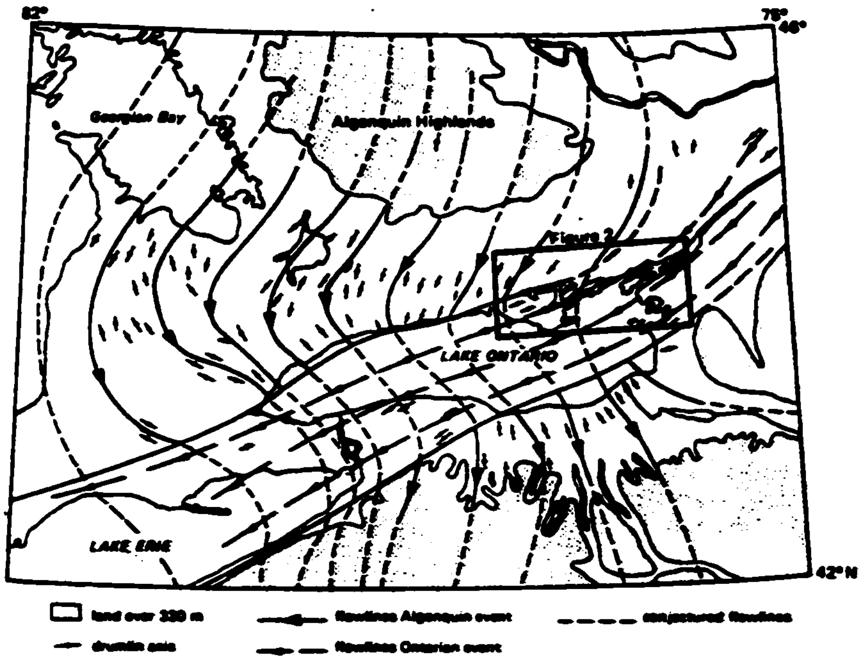

82°
78°
46°
Georgian Bay
Algonquin Highlands
Figure 2
LAKE ONTARIO
LAKE ERIE
42° N
land over 330 m
flowlines Algonquin event
conjectured flowlines
drumlin axis
flowlines Ontarian event

Fig. 2.5. (a) Legend for the surficial geology map of the Greater Toronto Area. (b) Surficial geology map of the Greater Toronto Area, from Sharpe et al. (1997).

LEGEND

QUATERNARY PERIOD (~last 2 million years)

11 — **Recent Deposits:** sand, gravel and diamicton: 1 - 3 m thick; includes wind-derived, landslide, slope, groundwater sapping, lakeshore deposits and fill

10 — **River Deposits:** sand and gravel
- a. gravel, sand, silt, clay, muck; 1-2 m thick; occurs on modern floodplains
- b. gravel, sand, silt, clay; 1-8 m thick; forms river deltas and terraces of early post-glacial age

Organic Deposits: peat, muck and marl, 1-7 m thick; occurs in wetlands

8 — **Glacial Lake Deposits**[1]**:** sand and gravel (minor diamicton)
- a. sand and silty sand, 1 to >50 m thick; occurs in basin lows and nearshore flats
- b. gravely sand and gravel, 1-5 m thick; raised shorelines or bars

Glacial Lake Deposits[1]**:** silt and clay, massive to laminated
- a. silt and clay interbedded with diamicton and some lone stones, 1 -10 m thick; occurs in basins
- b. silt and clay; 1-5 m thick; laminations deformed in basin fills

6 — **Glacial River Deposits**[2]**:** sand and gravel (minor diamicton)
- a. sand; 1-15 m thick; occurs as eskers, valley fills and terraces
- b. gravel; 1-15 m thick; occurs as eskers, valley fills and terraces

Moraine Deposits: fine sand to gravel
- a. fine sand, some gravel, minor silt, clay and diamicton; 1-50 m thick; rhythmic beds common
- b. medium to coarse sand and gravel and diamicton; 1-20 m thick; channels common

(a and b occur in disorganized hills, depressions and eskers)

Glacial Deposits (till): clayey silt to silt, 1-2% stone content; 1 -15 m thick; occurs in till or lake plains often with interbedded fine sand, silt and clay
- a. Wildfield / Kettleby[3]
- b. Halton
- c. Tavistock

Glacial Deposits (till): sandy silt to sand, > 3% stone content; stratified interbeds; 1-50 m thick; forms uplands
- d. Wentworth
- e. Port Stanley
- f. Newmarket/northern /Bowmanville

Lower (drift) Deposits: till, fine-medium sand, and laminated silt and clay, 1-50 m thick; exposed in bluffs
- g. Upper Thorncliffe Formation / Clarke beds; h. Seminary / Meadowcliffe / Bondhead tills;
- i. Lower Thorncliffe Formation / Clarke beds; j. Sunnybrook / Port Hope till;k. Scarborough Formation;
- l. Don Formation; m. York Till; n. Stratified sediment, dominantly sand; o. Stratified sediment, dominantly silt and clay

Unconformity ____________________________ (interval with no deposits and/or major erosion)

PALEOZOIC (rocks >400 million years in this area)

Bedrock: limy mudrock and clastic sedimentary rock
- a. bedrock-drift complex
- b. clastic (sandstone or shale)
- c. carbonate

1. Peel/Schomberg ponds above, Lake Iroquois/Algonquin below raised shorelines
2. Subglacial and/or proglacial outwash
3. Wildfield south of ORM and Kettleby north of ORM; sub units in units 2-4 are listed in stratigraphic order: a-o

Note: The legend is not strictly stratigraphic.

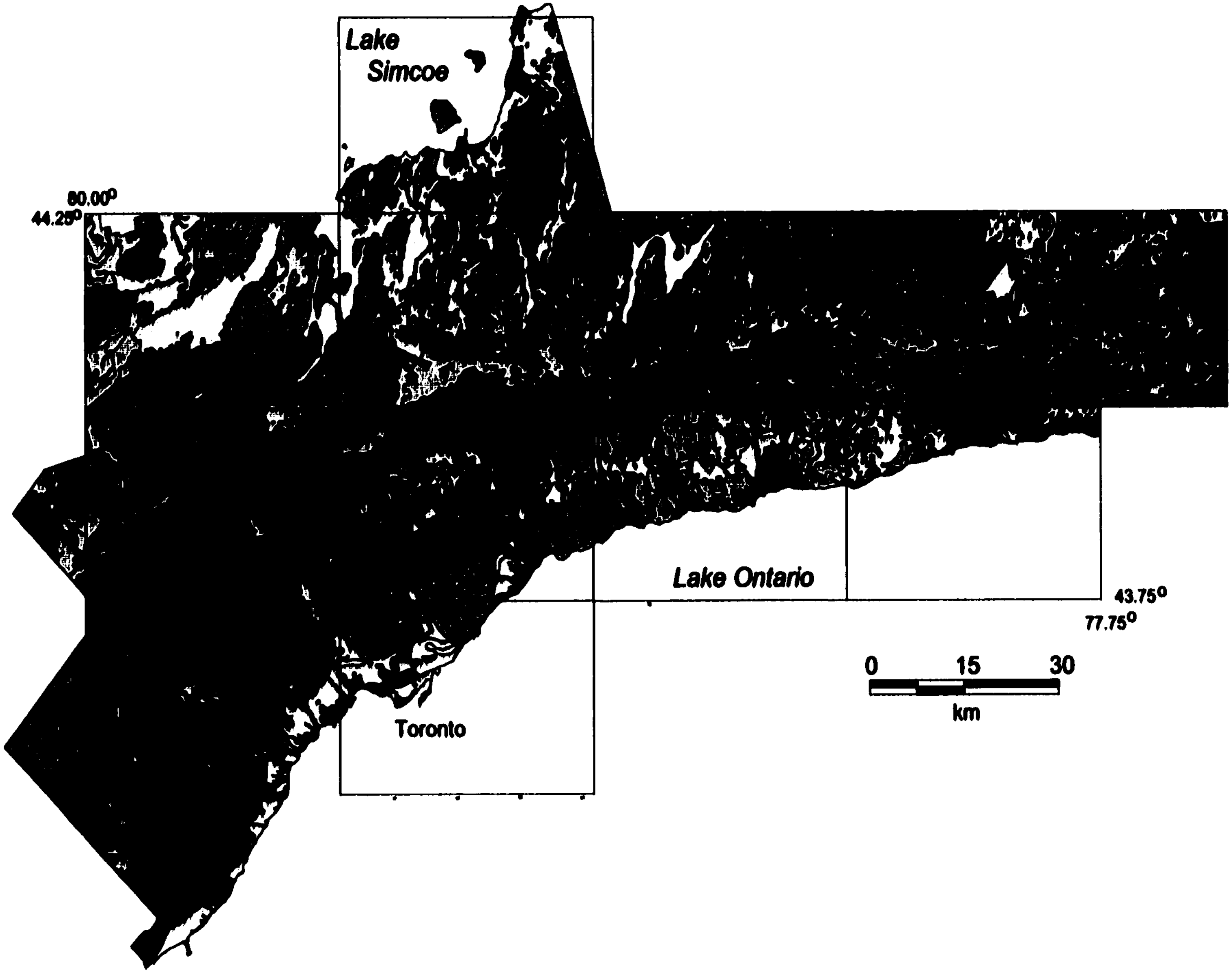

Lake
Simcoe
80.00°
44.25°
Lake Ontario
43.75°
77.75°
0
15
30
km
Toronto

Fig. 2.6. Schematic block model of the regional Quaternary stratigraphic framework of the Greater Toronto Area, modified from Sharpe et al. (1999). Note regional Late Wisconsin unconformity along upper surface of drumlinized and channelized Newmarket Till. Oak Ridges Moraine sediment overlies unconformity. For clarity channel fill layer has been omitted. Drawn by J. Glew.

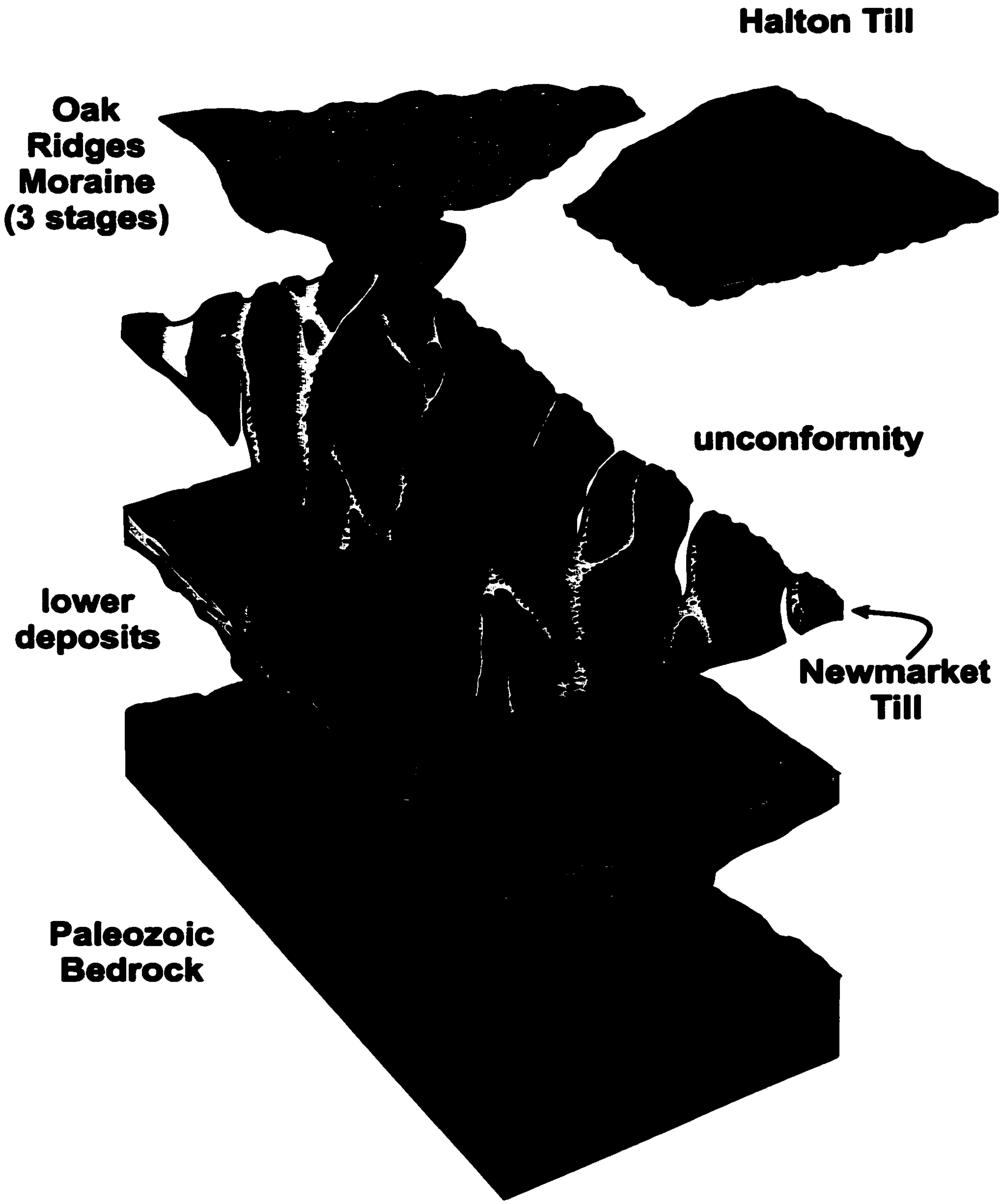

Halton Till
Oak Ridges Moraine (3 stages)
unconformity
lower deposits
Newmarket Till
Paleozoic Bedrock

Fig. 2.7. Distribution of mapped tunnel channels north of the Oak Ridges Moraine, modified from Sharpe et al. (1996). Note that surface expression of channels is truncated by Oak Ridges Moraine. For subsurface channel geometry, see Fig. 2.11.

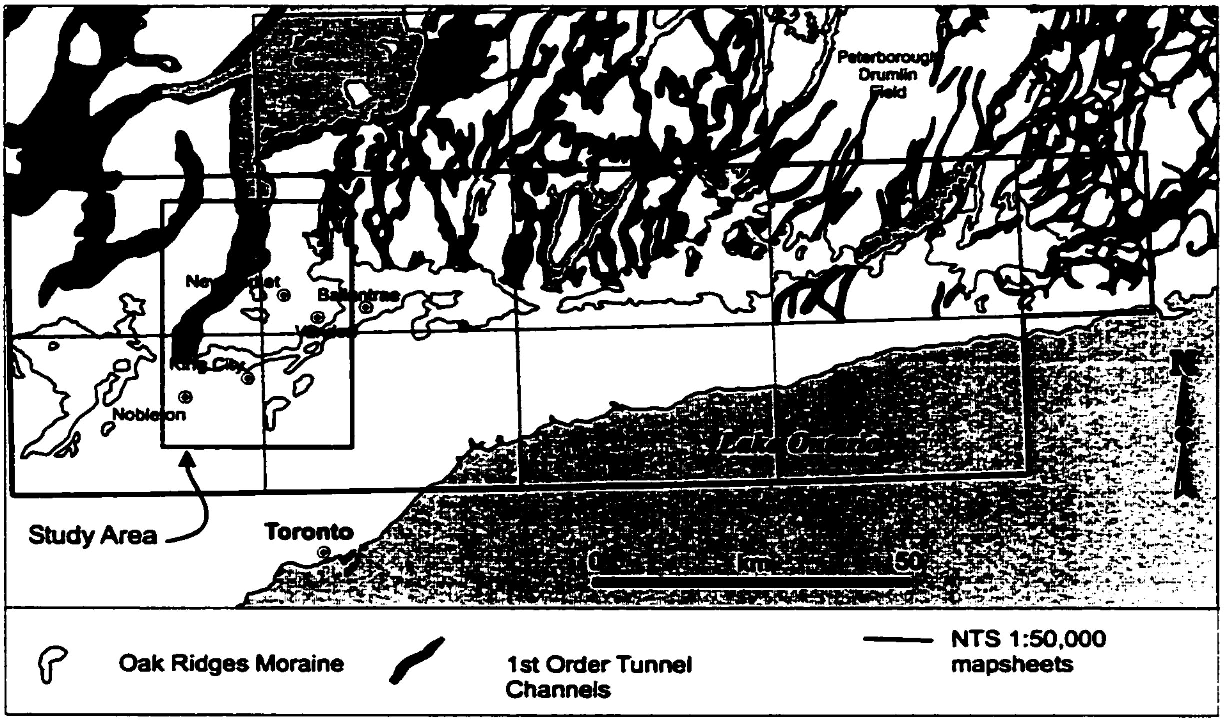

Peterborough
Drumlin
Field
Study Area
Toronto
Oak Ridges Moraine
1st Order Tunnel
Channels
NTS 1:50,000
mapsheets

Fig. 2.8. Conceptual depositional model of the four formative stages of the Oak Ridges Moraine, from Barnett et al., (1998). Depositional order is indicated by roman numbers in top right corner of inset boxes. Drawn by J. Glew, 1997.

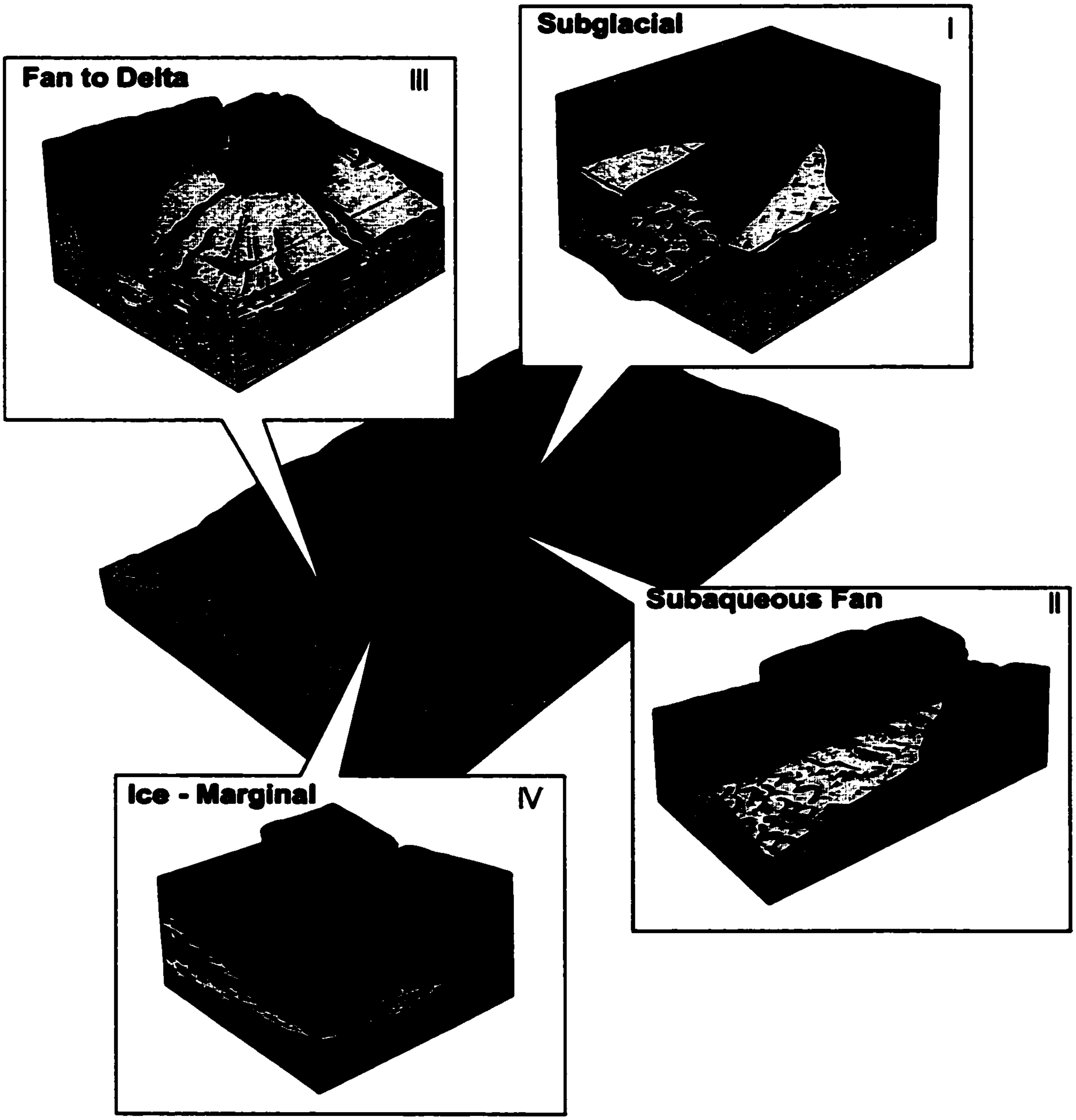

Subglacial
I
Fan to Delta
III
Subaqueous Fan
II
Ice - Marginal
IV

Fig. 2.9. Surficial geology (from Sharpe et al., 1997) draped on a digital elevation model of the Humber River watershed study area (from Kenny et al. 1999). Note irregular topography associated with western Oak Ridges Moraine and areas of Halton Till.

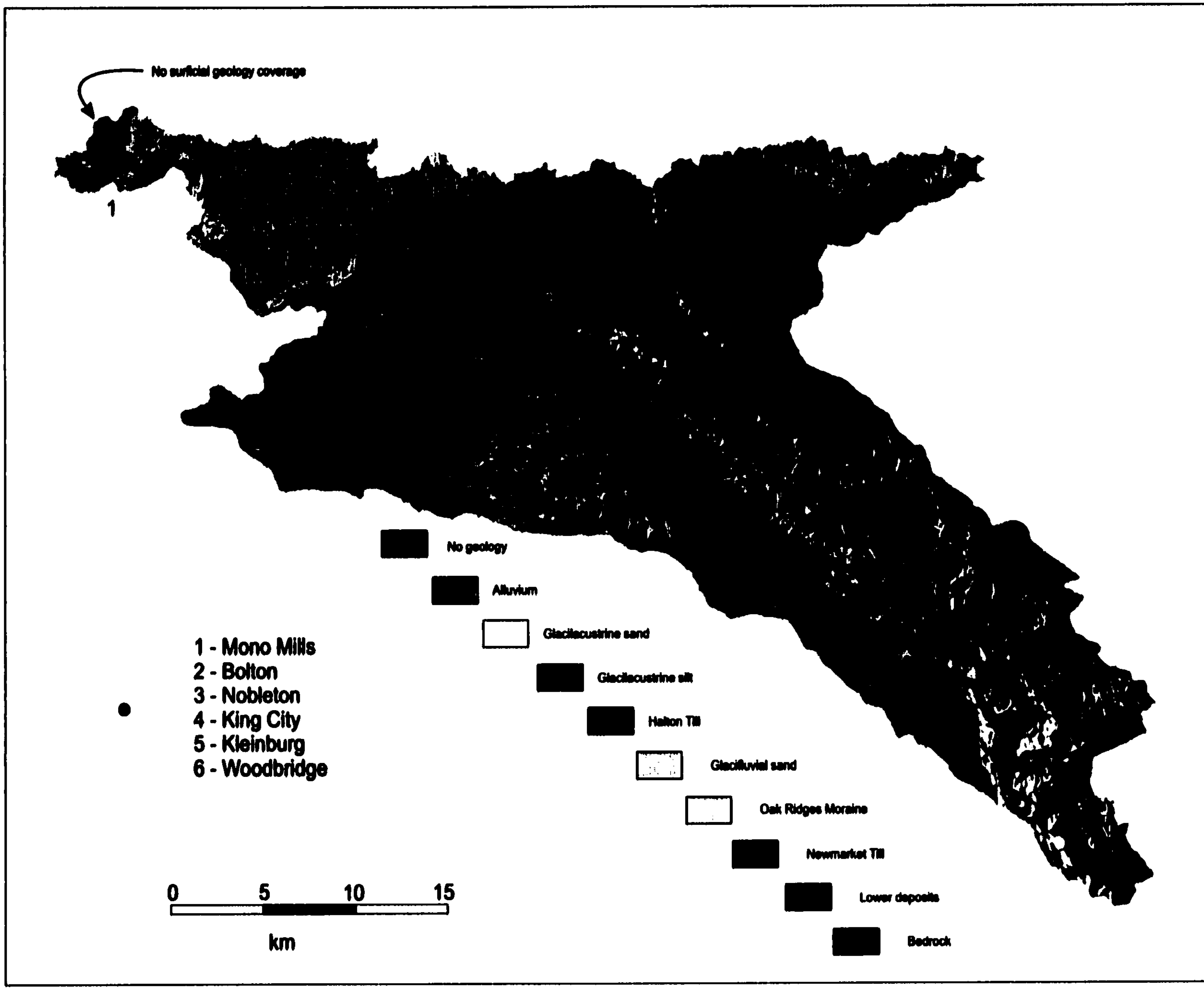

No surficial geology coverage
1
No geology
Alluvium
Glaciolacustrine sand
Glaciolacustrine silt
Halton Till
Glaciofluvial sand
Oak Ridges Moraine
Newmarket Till
Lower deposits
Bedrock
1 - Mono Mills
2 - Bolton
3 - Nobleton
4 - King City
5 - Kleinburg
6 - Woodbridge
0
5
10
15
km

Fig. 2.10. (a) Bedrock geology of the Humber River watershed. Across most of the watershed the geology subcrops beneath up to 200 m of Quaternary sediment. Geology from OGS digital data set 6 (b) Bedrock topography of the Humber River watershed, modified from Brennand et al. (1997). The western margin of the Laurentian Channel is coincident with the lowest areas along the eastern edge of the watershed. (c) Quaternary geology of the study area, geology from Sharpe et al. (1997) (d) Sediment thickness of the Humber River watershed, modified from Russell et al. (1998). Note general trend to thinner Quaternary sediment toward the south. Scale and orientation of all figures are the same as (a).

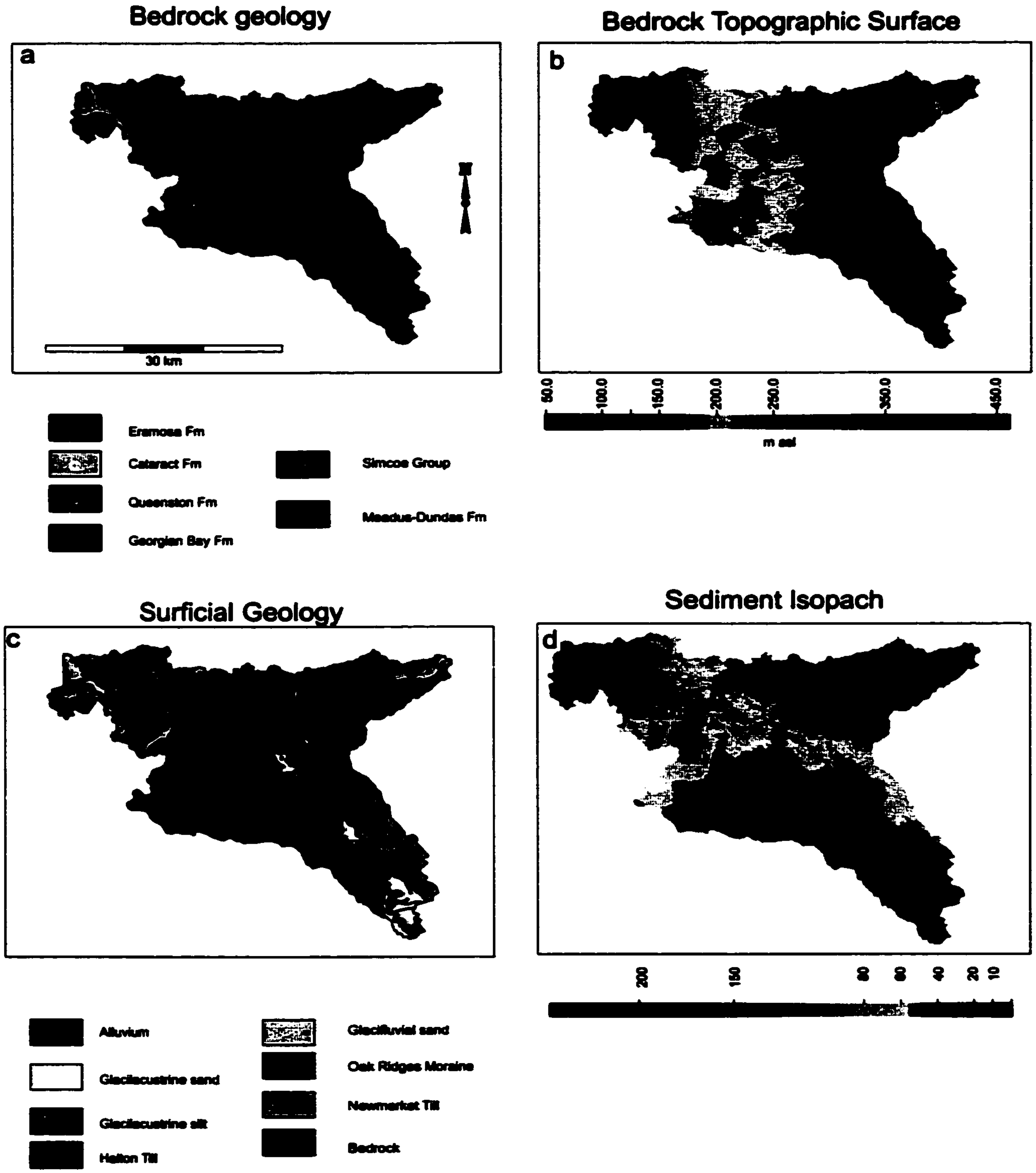

Bedrock geology
a
30 km
Eramosa Fm
Cataract Fm
Queenston Fm
Georgian Bay Fm
Simcoe Group
Meadus-Dundas Fm
Bedrock Topographic Surface
b
50.0
100.0
150.0
200.0
250.0
m asl
Surficial Geology
c
Alluvium
Glaciolacustrine sand
Glaciolacustrine silt
Halton Till
Glaciofluvial sand
Oak Ridges Moraine
Newmarket Till
Bedrock
Sediment Isopach
d
200
150
40
20
10

Fig. 2.11. (a) Sediment log and downhole geophysics of the Nobleton drillhole, from Russell and Pullan, (1998). Note strong geophysical trends for Scarborough and Thorncliffe formations compared to Oak Ridges Moraine sediment. (b) Reflection seismic profile adjacent to the Nobleton drillhole, seismic data from Pugin et al. (1999). Note strong horizontal reflectors of lower deposits truncated west of the drillhole. Truncation surface is interpreted as base of buried Holland Marsh tunnel channel and part of the Upper Wisconsin regional unconformity.

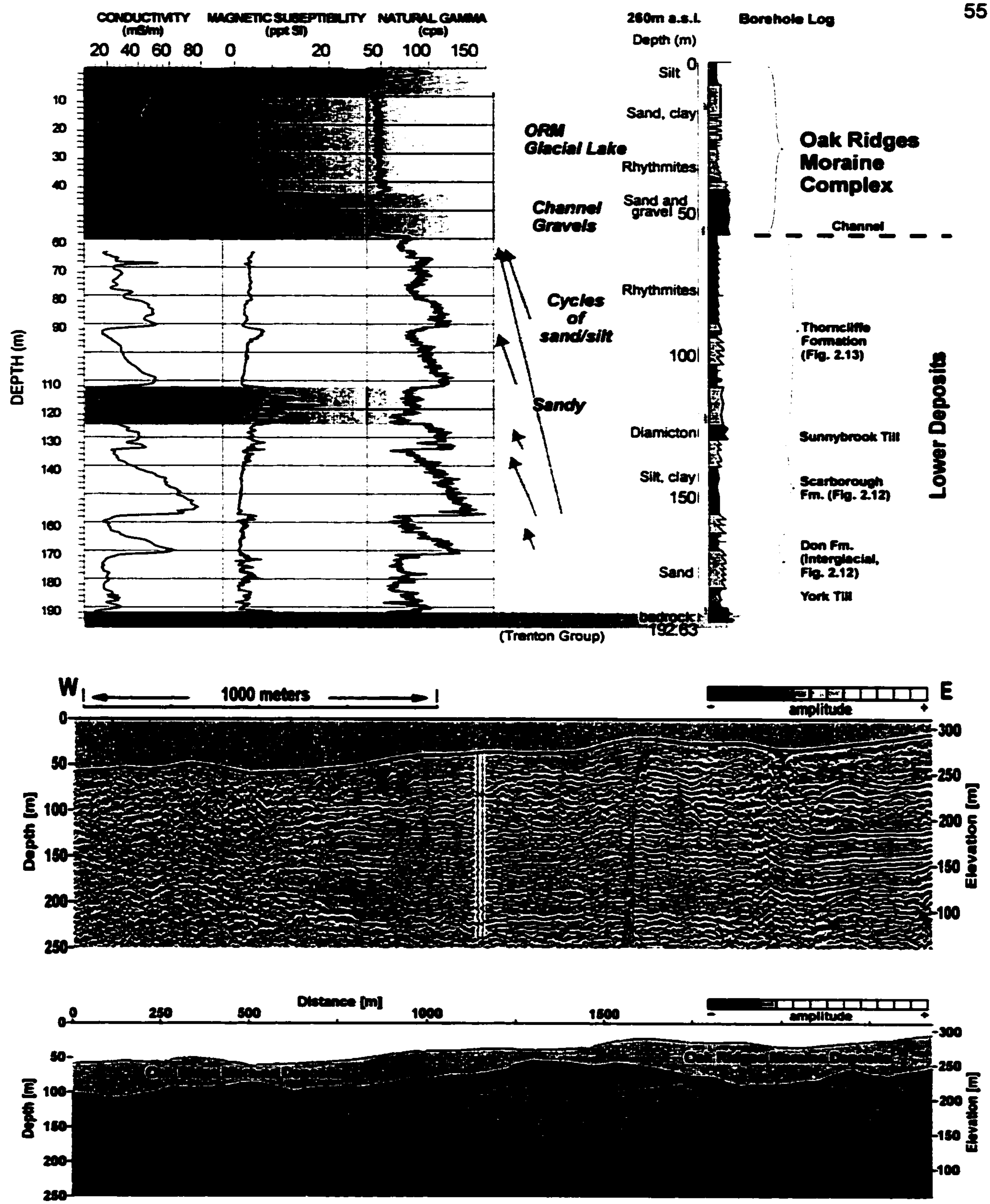

CONDUCTIVITY (mS/m)
MAGNETIC SUSEPTIBILITY (ppt SI)
NATURAL GAMMA (cps)
20 40 60 80 0 20 50 100 150
DEPTH (m)
10 20 30 40 60 70 80 90 110 120 130 140 160 170 180 190
ORM Glacial Lake
Channel Gravels
Cycles of sand/silt
Sandy
(Trenton Group)
260m a.s.l.
Depth (m)
Borehole Log
Silt
Sand, clay
Rhythmites
Sand and gravel
0
50
100
150
Rhythmites
Diamicton
Silt, clay
Sand
bedrock
192.63
Oak Ridges Moraine Complex
Channel
Thorncliffe Formation (Fig. 2.13)
Sunnybrook Till
Scarborough Fm. (Fig. 2.12)
Don Fm. (Interglacial, Fig. 2.12)
York Till
Lower Deposits
W
1000 meters
E
amplitude
Depth [m]
Elevation [m]
0 50 100 150 200 250
300 250 200 150 100
Distance [m]
0 250 500 1000 1500
amplitude
Depth [m]
Elevation [m]
0 50 100 150 200 250
300 250 200 150 100

Fig. 2.12. (a) Don Bed Formation sand with shell fragments of the lower deposits at the base of the Nobleton core. (b) Scarborough Formation sand and organic debris of the lower deposits from the Nobleton core.(Russell unpublished).

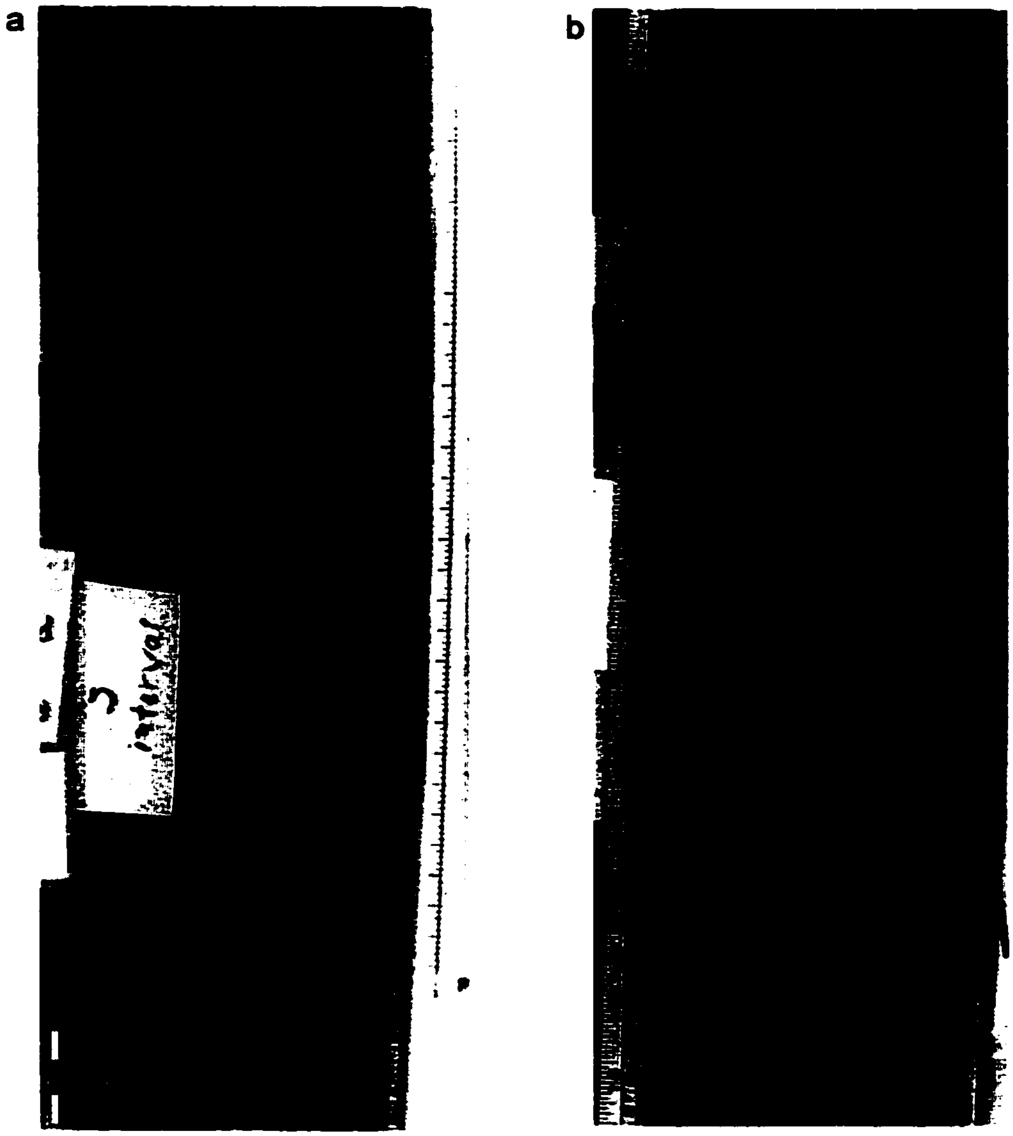
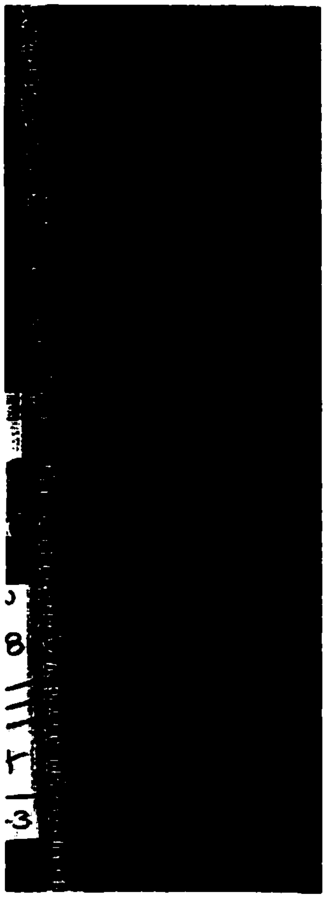
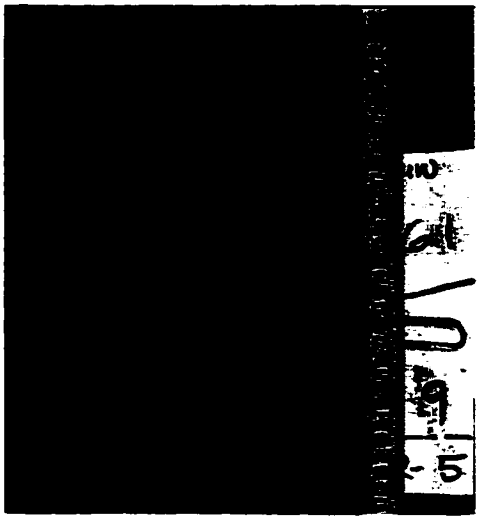


Fig. 2.13. Thorncliffe Formation silt-clay rhythmites of the lower deposits. (a) Rhythmites in the Nobleton core. (b) Close-up of rhythmites in the Nobleton core showing fine planar laminae, not from (a). (c) Outcropping rhythmites near Kleinburg.

a



b



c

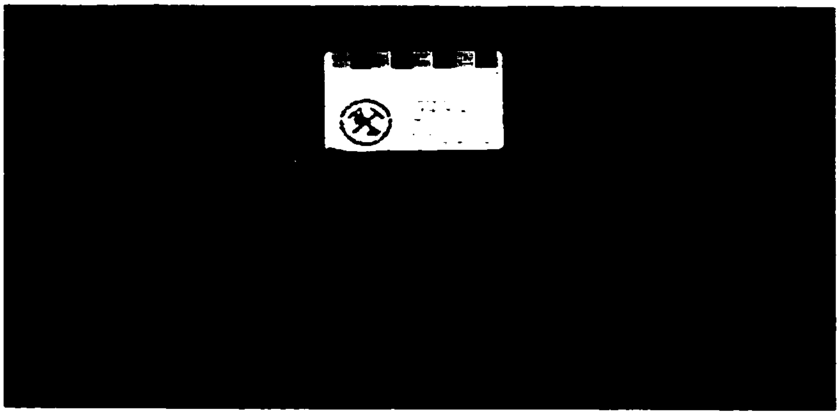
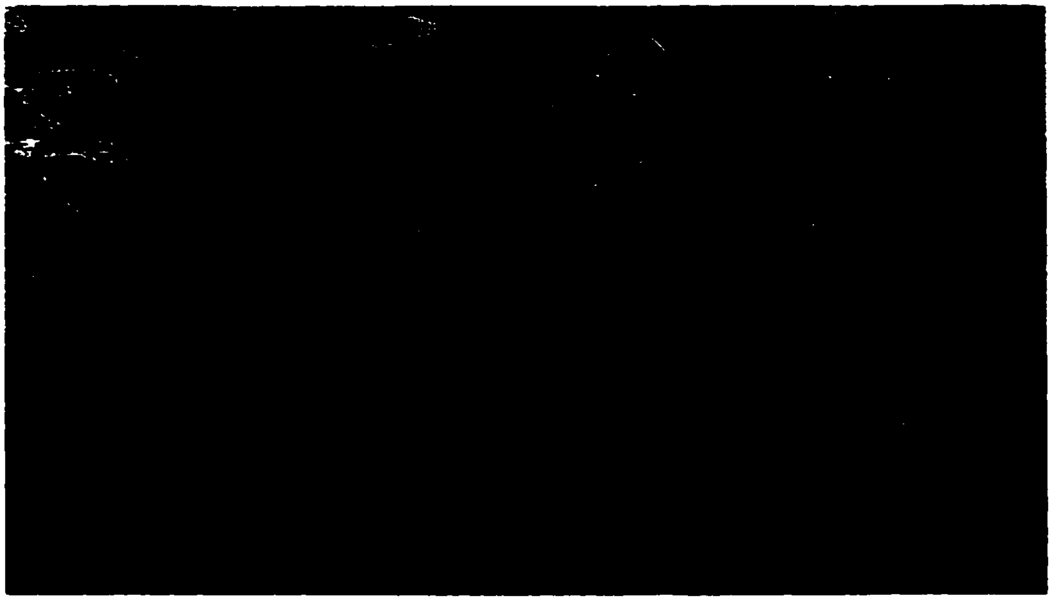


Fig. 2.14. Newmarket Till outcrop near Kleinburg. Note abundance of oversized clasts and sandy matrix. Wooden handle is ~ 1 m long.

Fig. 2.15. Halton Till, (a) massive silt diamicton with granules, (b) lens of cross-stratified sand in a silt-sand diamicton. Intervals on scale is 10 cm.

a



b

Table 2.1. Regional correlation for the Stratford-Toronto region (modified from Karrow, 1974, ages from Barnett, 1992).

Age (ka)	Time Classification*	Stratford (Huron lobe?)	Listowel (Georgian Bay lobe?)	Galt-Hamilton (Erie-Ontario lobe)	Toronto (Ontario lobe)
10	Holocene				
	Wisconsinan *Late*				
13	Port Huron Stade			Halton Till Wentworth Till	Halton Till
13.4	Mackinaw Interstade				Oak Ridges Moraine
14.8	Port Bruce Stade	Till M Stratford Till[†]	Elma Till[†] (N)		
		Tavistock Till[†] Till H	Morington Till[†] (D) Tavistock Till[†] (C) Stirton Till[†] (B)	Port Stanley Till (W) Maryhill Till[†] (P)	
15.5	Erie Interstade				
20	Nissouri Stade	Catfish Creek Till	Catfish Creek Till Older Till	Catfish Creek Till Older Till	NewmarketTill
22	*Middle* Plum Point Cherry Tree		Older tills?		Upper Thorncliffe Fm
		Fossiliferous sed	Fossiliferous sed		Meadowcliffe Till Middle Thorncliffe Fm
40	Port Talbot				Seminary Till Lower Thorncliffe Fm
60	*Early* Guildwood St. Pierre Nicolet	Older tills?	Older tills?		Sunnybrook Till Pottery Road Fm Scarborough Fm
115	Sangamonian				Don Fm
>135	Illinoian				York Till

Notes: Upper case letters in parentheses following full names are the informal designations of Karrow (1971)
Tills designated only by a letter retain informal nomenclature of Karrow (1971).
**From Dreimanis and Karrow, 1972.*
[†]New name.

Table 2.2 Stratigraphic sequence of Quaternary deposits in the Oak Ridges Moraine area. Modified from Sharpe et al., (1997).

Stratigraphic unit	Explanation/description	Age years ka	Regional Stratagraphic Framework
Quaternary period	present to 1.8 million years ago - glacial time		
Recent (Holocene) epoch	interglacial since 10,000 BP	< 10 ka	
Alluvium	sediment deposited by modern rivers		
Older alluvium	sediment deposited by rivers in terraces and deltas		N/A
Organic deposits	wetlands		
Pleistocene epoch	10,000 to 1.8 million years ago		
Late Wisconsinan		~12-25	
Glacial lake deposits	Iroquois/Algonquin/Peel/Schomberg (sand, silt, clay)		
Glaciofluvial deposits	deposits of high-energy streams (sand and gravel)		Halton Till
Wilfield / Kettleby tills[2]	interbedded glaciolacustrine diamicton		Halton Till
Halton Till	interbedded with clay, silt and fine sand		
Wentworth / Upper Leaside tills	may be part of Halton		
Oak Ridges Moraine sediments	thick glacial lake deposits overlie regional unconformity		Oak Ridges Moraine
Port Stanley / Tavistock tills	Lake Ontario / Lake Huron derived till		N/A
unconformity			
Newmarket Till	Late Wisconsin Maximum, regional till		Newmarket Till
Middle to Early Wisconsinan		~25-100	
Upper Thorncliffe Fm / *Clarke Beds*[3]	deltaic / subaqueous fan sands / lacustrine varves		lower deposits
Seminary/Meadowcliffe/*Bondhead tills* diamicton separating Thorncliffe sands	thin diamictons		
Lower Thorncliffe Fm / *Clarke Beds*	sand and clay rhythmites		
Sunnybrook / *Port Hope tills*	may form regional till sheets	~30-50	
Scarborough Formation	glacilacustrine,detrital organics (~122 m asl at bluffs)	~90	
Sangamonian		>115	lower deposits
Don Formation	warm-climate (+3°C) sand, silt and clay		
Illinoian			lower deposits
York Till	till on bedrock		
unconformity	no deposits and/or major erosion		unconformity
Paleozoic Era	543-230 million years ago		
(Ordovician- Devonian)	carbonate; sandstone or shale		N/A

[2]Wilfield occurs south of ORM; Kettleby occurs north of ORM.
[3]Units in *italics* refer to strata found east of Oshawa; see Brookfield *et al.* 1982.

Table 2.3 Development of ideas on extent and origin of Oak Ridges Moraine (ORM). Modified from Barnett et al. (1998).

Author	Contribution
Bigsby (1829)	Noted "Oak Ridge," the height-of-land, north of Toronto
Logan (1863)	Suggested moraine extended from Niagara Escarpment to east of Trent River
Taylor (1913)	Formally defined "Oak Ridges Moraine"; extended from King City and Maple to the Trent River; inferred interlobate and overlapping nature (Lake Ontario Lobe, younger, and Lake Simcoe Lobe, older)
Chapman and Putnam (1943)	Modified Taylor's definition of ORM; included smaller moraines named by Taylor extending the moraine to the Niagara Escarpment and to approximately Trent River (e.g., Logan 1863); agreed to interlobate origin (most accepted definition of ORM)
Gravenor (1957)	ORM interlobate and may be old and, in part, palimpsest, having been overridden by a readvance from the north
Mirynech (1962, 1978)	Eastern ORM moved west of Trent River (Castleton); ORM was not overridden
White (1975)	Redefined ORM in west based on sediment character; identified a separate Palgrave Moraine (Lake Ontario ice only); ORM to include only interlobate high-level sands in King Township
Gwyn (1972)	Sandy till underlies ORM north and south (Newmarket; Bowmanville Tills)
Gwyn and DiLabio (1973)	Palgrave Moraine extended to all areas on south flank of ORM covered by Halton Till
Duckworth (1975, 1979)	Palgrave Moraine extended into Newmarket area; ORM not interlobate; a braided stream deposit fed from the south
Gwyn and Cowan (1978)	ORM not an interlobate moraine; it is a young feature (~13 ka)
Chapman (1985)	Identified strong regional control on water level and sedimentation within ORM
Barnett (1992, 1993, 1994)	ORM is interlobate and composite (e.g., Taylor 1913); subaqueous fans, deltas, end moraines of both northern and southern ice lobes; one till, Late Wisconsinan, underlies ORM; linked tunnel valleys and ORM genesis
Gorrell and McCrae (1993)	ORM has subglacial origin in eastern area; ORM linked to eskers, tunnel channels
Brennand and Shaw (1994)	Early part of ORM was deposited in a subglacial depositional environment (esker)
Gilbert (1997)	ORM sediments at Vandorf formed in 100 m deep lake in about 100 years
Paterson (1997)	Subaqueous fan model for part of the Uxbridge wedge
Barnett et al., (1998)	Has several depositional environments: subglacial, ice-marginal, and proglacial lacustrine; ORM comprises smaller landforms, including Palgrave Moraine; ORM is linked to channel filling; sedimentation was rapid, a few hundred years at most

Chapter III

Sedimentology of a Tunnel Channel Fill

3.1 Introduction

Erosional landscapes formed by large, catastrophic meltwater discharge events from subglacial drainage networks were once considered anomalous (e.g. Bretz 1969). Episodic drainage processes are now recognized to be common events during the waning of the Laurentide Ice Sheet (Shaw 1996), and other continental (e.g. Piotrowski 1994; Sawagaki and Hirakawa 1997), and outlet glaciers (e.g. Fleisher et al. 1997). One element of this meltwater landscape is tunnel channel (valley) networks. These networks have been mapped from extensive regions once covered by Pleistocene ice sheets, particularly in North America and western Europe (e.g. Cofaigh 1996). The Great Lakes basin of central North America (Fig. 3.1),the Scotia Shelf, North Sea and northern Germany are areas in which extensive channel networks have been reported. The plan and cross-sectional geometry of tunnel channels have been well defined by landform analysis (e.g. Brennand and Shaw 1994; Fisher and Taylor 1999), seafloor mapping (e.g. Loncarevic et al. 1992) and seismic reflection surveys (e.g. Pugin et al. 1999).

The sedimentary fill of tunnel channels is poorly understood. In many cases the stratigraphic detail of the thick fill of buried channels has been delineated by seismic facies analysis (e.g. Boyd et al. 1988; Mullins et al. 1996), and low-quality water well boring descriptions (e.g. Ehlers and Linke 1989). Outcrop and continuous drill core studies and acquisition of high quality sedimentological data, on the other hand, are comparatively rare (e.g. Eyles and McCabe 1989; Shaw and Gorrell 1990; Wellner et al. 1996). Seismic data resolve large-scale stratal geometries, but generally lack the resolution to identify small-scale sedimentary features necessary for depositional process interpretations. In contrast to architectural information provided by seismic data, tunnel channels mapped by water wells rely on textural changes from wash-boring records (Ehlers and Linke 1989; Patterson 1994; Piotrowski 1994). Sediment descriptions obtained from wash-boring holes come from monitoring of drilling fluid and sediment flushed to the surface during the drilling process. Accordingly, in the absence of downhole geophysics, few data are available on either the stratal succession or its three dimensional geometry. As a consequence, although detailed paleohydraulic models have been developed for some channels based primarily on water well data (Piotrowski 1997), detailed depositional geological models for channel fills remain undeveloped. Where thick, deeply buried channel fill sediment has been cored, the full succession has typically been partially intercepted (e.g. Boyd et al. 1988; Wellner et al. 1996). Similarly, outcrop studies have been restricted to near-surface studies of generally late stage channel fills. The resultant scarcity of detailed lithofacies descriptions from the lower part of the thick channel fill succession continues to constrain models of channel erosion and infill sedimentation.

Tunnel channels are interpreted to form by one of three mechanisms: i) subglacial sediment deformation (Boulton and Hindmarsh 1987), ii) time-transgressive formation close to the ice margin (e.g. Mooers 1989), and iii) catastrophic subglacial meltwater floods (Shaw and Kvill 1984; Wright 1973). These models have been developed primarily on the basis of erosional landforms, mapping of adjacent sediments, and glacial hydraulic modelling. Integration of sedimentological information on tunnel channel fills has been limited by a paucity of detailed data and an inability to relate erosional and depositional flow events. Consequently, each of these models has a number of deficiencies regarding processes that formed and filled them that further study is still needed (Cofaigh 1996).

This paper presents sedimentological data from two continuously cored drillholes supplemented with seismic data from Pugin et al. (1999). The cores are located in a tunnel channel network previously defined by regional mapping (Sharpe et al. 1997), seismic profiling (Pugin et al. 1999), and borehole analysis (Russell unpublished) of the Oak Ridges Moraine area, southern Ontario. Preliminary analysis of the two cores (Russell and Pullan 1998; Russell et al. 1998) has defined the base of the channel fill succession and described the relationship between channel fills and the Oak Ridges Moraine (Barnett et al. 1998). In one core the lowest channel fill is gravel and in the other it is predominantly massive sand. A detailed lithofacies description of this lower channel fill succession is presented along with seismic data that permit the lateral geometry to be defined. Depositional processes are interpreted for individual lithofacies and a depositional model is presented for the complete succession. These data are then compared with previously reported seismic facies and from deeply buried tunnel channel fills to develop the stratal architecture and depositional mechanism of tunnel channel fills.

3.1.1 Geological setting

North of Lake Ontario the Paleozoic bedrock of south-central Ontario is overlain by a thick sequence (< 200 m) of Quaternary sediment deposited after the Illinoian glaciation (Karrow 1974). In the Toronto area, a simplified regional stratigraphy has been established that consists of four principal stratal units. These include from oldest to youngest: i) pre Upper Wisconsinan deposits; and Upper Wisconsin, ii) Newmarket Till, iii) Oak Ridges Moraine, and iv) Halton Till (Sharpe et al. 1996). In this stratigraphic framework a regional unconformity is recognized as drumlins and channels on the upper surface of the Newmarket Till (Sharpe et al. 1997). In a regional event-stratigraphic model of the area, Shaw and Gilbert (1990) recognized two late Wisconsinan meltwater erosion events termed the Algonquin and Ontarian. The older Algonquin event is represented by the unconformity on the exposed upper Newmarket Till surface north of the Oak Ridges Moraine (Shaw and Sharpe 1987) and that extends beneath the Oak Ridges Moraine. Part of the erosional topography is an anabranched pattern of tunnel channels that has been mapped north and east of the Oak Ridges Moraine (Fig. 3.1). Channels are 10s km long and are up to 4 km wide (Barnett 1990; Brennand and Shaw 1994). Surface relief of channels is generally < 50 m but they are up to 160 m deep for a combined depth of ~200 m (Barnett 1990). Based on seismic reflection surveys (Pugin et al. 1996; Pugin et al. 1999) and drillholes (Barnett et al. 1998) the channels have been mapped beneath the Oak Ridges Moraine. For example, the southward downflow extension of the Holland Marsh tunnel channel (Sharpe et al. 1997) has been mapped by seismic surveys (Pugin et al. 1999) and intercepted by one of the drillcores described in this paper (Russell and Pullan 1998). Regionally the channel infill consist of coarse-grained gravel and sand (e.g. Barnett et al. 1998; Shaw and Gorrell 1990), overlain by late stage proglacial lacustrine silt and fine sand (e.g. Sharpe et al. 1997). These strata are characterized by a number of seismic facies types that include: large, steeply dipping sigmoidal reflectors, irregular hummocky reflectors, and incoherent, chaotic reflectors (Pugin et al. 1999). These seismic facies are interpreted to be bedform foresets, high energy scour and fill, slump or mass-flow deposits and fine-grained suspension deposits, respectively (Pugin et al. 1999). The overlying Oak Ridges Moraine is interpreted by Barnett et al. (1998) to have formed in four principal depositional stages: i) subglacial, ii) subaqueous fan, iii) fan to delta, and iv) ice-marginal sedimentation (Fig. 3.2b).

3.1.2 Study site and methodology

The dataset consists of two continuous, 9 cm diameter drillcores from the western Oak Ridges Moraine, approximately 35 km north-northwest of metropolitan Toronto (Fig. 3.1, 3.2a). Both cores penetrate the complete Quaternary succession and terminate in bedrock. Drillcore recovery is generally > 90 %, but is significantly lower in gravel intervals. Drillhole DH-Nob is collared at 260 m asl (above sea level) and bedrock was intercepted 190 m beneath the surface (70 m asl). Drillhole DH-Nob was positioned by the Geological Survey of Canada to provide core control for a seismic line that trends normal to the projected southward continuation of the Holland Marsh (Fig. 3.2a). The second drillhole (DH-V-158) is located ~7 km to the southeast. It is collared at 276 m asl and intercepts the bedrock at a depth of ~155 m (~121 m asl). Both cores were logged in bed-by-bed detail, photographed, and samples collected for grain size and total organic carbon content. Drillcore DH-V-158 was originally collected and logged by M. Gomer (Fenco MacLaren 1994) to ASTM protocol for a hydrogeology study. The stratal log presented here is a synthesis of subsequent logging completed when the core was moist and relogging when it was dry. Grain size data have been analyzed using the graphic technique of Folk (1974). In this study the ASTM clay size sediment classification (ASTM 1994) of < 2 micron (9 phi) is used rather than the < 3.9 micron (8 phi) defined by the Wentworth scale. The silt, sand, and gravel intervals are the same in both grain size classifications. Total organic carbon (TOC) content was determined with a Leco CR-12 Carbon Analyzer (Wang and Anderson 1998).

3.2 Sedimentology

From the base up the general succession of DH-Nob (Fig. 3.2b) is fossiliferous sand, organic-rich, fine-medium sand, rhythmic silt-clay with horizontal trace fossils, gravel, silt-sand, and interbedded silt and silt diamicton. On the basis of texture, organic content and bioturbation these strata have been correlated with regional stratigraphic units, that from oldest to youngest are: Don Beds, Scarborough Formation, Thorncliffe Formation, Oak Ridges Moraine, and Halton Till (Fig. 3.2b). The Don Beds, Scarborough Formation and Thorncliffe Formation termed the lower deposits (Sharpe et al. 1996) are pre Upper Wisconsinan deposits and occur below the regional unconformity described above. The contact between fine sand to clay varves of Thorncliffe Formation and overlying gravel is interpreted to be the erosional base of a tunnel channel that coincides with the unconformity. Regional seismic reflection data (Pugin et al. 1999) and abrupt changes in downhole geophysical signatures (Eyles et al. 1985; Russell and Pullan 1998) support this interpretation.

The core at DH-V-158 consists predominantly of fine to medium sand, silt and clay (Fig. 3.2b). A 5 m thick massive silt-sand diamicton overlies bedrock. Below 171 m asl the succession is mostly massive fine and medium sand. From 171 to 207 m the succession consists of ~62 rhythmic sand-silt clay couplets that fine and thin upward. Clay strata become less common above 200 m and the succession is dominated by small-scale cross-laminated sand and microlaminated silt. The upper 20 m consists of interbedded silt diamictons and fine sand.

The Oak Ridges moraine strata in DH-V-158 can be correlated with the four stage regional depositional model of Barnett et al. (1998). In DH-Nob the gravel unit and in DH-V-158 the sand strata below 171 m are correlated with Stage 1(Fig 3.2b). Stage 2 strata in DH-V-158 consist of sand-silt varves deposited in a distal subaqueous fan environment. The fan to delta stage 3 sand and silt form an overlying 50 m thick succession that can be correlated between DH-V-158 and DH-Nob. The final stage corresponds to the fine grained ice-marginal glacilacustrine Halton Till that represents the surface unit in drillcore.

The planform of the tunnel channel (Fig. 3.2a) and the stratigraphic cross-sections (Fig. 3.2b,c) constructed between the drillholes has been developed from a variety of datasets. For the Holland Marsh and DH-Nob the cross-section is a synthesis of seismic reflection data (Pugin et al. 1999), regional mapping (Sharpe et al. 1997), drillcore logs, and analysis of water well logs. The Holland Marsh tunnel channel is 5-6 km wide and ~170 m deep (Fig. 3.2c). A second subparallel channel, defined by drillcore, water well logs and surface mapping is ~4-5 km wide and 100-140 m deep (Fig. 3.2b).

Detailed lithofacies descriptions and interpretations are presented below for the gravel strata in DH-Nob and the diamicton and sand strata below 171 m (asl) in DH-V-158 that correspond with stage 1(Fig. 3.2).

3.2.1 Diamicton

The up to 5 m thick diamicton facies subcrops at the base of DH-V-158 where it overlies Paleozoic bedrock. It has 10-15 % oversized clasts (> 4 mm) dispersed in a dense, massive, clay-silt to silt-sand matrix. Both the matrix grain size and oversized clast size increase upwards. Clasts range in size from pebbles to boulders. Boulder size was estimated from the maximum dimension of drilled clasts (25 cm). Oversized clasts are subrounded to angular. Clasts are shale, limestone, dolostone, granite and granite gneiss. Angular shale clasts are more abundant in the lower part of the facies, whereas subrounded granite and granite gneiss increase in abundance upward. Total organic carbon (TOC) values are < 0.23 %.

Interpretation: On the basis of its dense and massive character, clast lithology and shape, and stratigraphic position, the diamicton is interpreted to be a till of probable subglacial lodgement origin. The shale-rich basal diamicton indicates local subglacial erosion and consequently is correlated with the York Till, which crops out ~10 km to the southwest, along the Humber River (White 1973). Upward change in matrix texture and clast composition most likely indicate a change in sediment provenance from local shale-dominated bedrock to more distal sources. Alternatively, but less likely, the change in clast composition represents incorporation of local Newmarket Till from slump along tunnel channel walls. Poor core recovery precludes a definitive determination of the depositional mechanism for the upper part of the facies.

3.2.2 Gravel

The gravel facies is present only in DH-Nob and occurs at a depth of 33.3-50.0 m (~204-221 m asl) and overlies silt and clay of the Thorncliffe Formation (Fig. 3.2b). Core recovery from the 16.7 m interval of gravel was poor, ranging from ~3 to ~40%, and consequently preservation of gravel texture and structure is poor (Fig. 3.3). For the most part, little or no fine-grained matrix was recovered. Variable core recovery and changes in drilling speed, drilling fluid loss, and logging of recirculated drilling fluid suggests that the gravel sequence is stratified and consists of a number of upward-fining cycles (G. Gorrell, pers. com. 1998). Sediment ranges from coarse sand-granule to polymodal and possibly bimodal pebble - cobble gravel (Fig. 3.3). The presence of a primary sand matrix is inferred from its presence in the recirculated drilling mud. Clasts consist predominantly of limestone, granite gneiss and gneiss (< 20 %), plus minor dolostone. Clasts are mostly subrounded with a minor subangular fraction. A small percentage of clasts are larger than the 9 cm core diameter and therefore represent only a minimum size estimate.

Seismic reflection data indicate that the gravel interval is a west-dipping tabular unit of highly reflective, horizontal to subhorizontal and hummocky, continuous to discontinuous reflectors (Fig. 3.3d). Laterally within the unit, the reflection pattern changes from continuous subhorizontal reflectors to irregular discontinuous reflectors. In addition, reflectors less continuous in the eastern part of the unit. To the east of DH-Nob, the reflectors onlap a weaker, less coherent seismic facies (Fig. 3.3).

Interpretation: Interpretations of depositional process are inferred from drilling characteristics, grain-size and seismic signatures because primary sedimentary structures were not observed in core. Based on the abrupt change from silt in the underlying Thorncliffe Formation to gravel, and truncation of seismic reflectors, the base of the gravel is interpreted to be the floor of a channel that forms part of a regional unconformity (Pugin et al. 1999). The less coherent reflection patterns observed in the eastern area are interpreted to represent several stacked channel-base mesoforms (i.e. subaqueous dunes) that possibly formed part of bank-attached macroforms. Based on the vertical seismic resolution of ~ 4 m (Table 3.1) individual mesoforms are interpreted to be greater than 4 m thick. The absence of sigmoidal reflectors associated with an angle of repose slip-face is, most probably, the result of the transverse orientation of the seismic profile with respect to the paleoflow direction. The western, more coherent, horizontal reflectors are interpreted to be low relief gravel bedload sheet deposits. These strata occur closer to the channel axis, and were probably deposited under higher energy flow conditions than the mesoforms deposits that formed along the channel margins.

3.2.3 Graded-massive sand

The graded - massive sand facies is the most common facies below 171 m asl in drillhole DH-V-158. It consists predominantly of silty, medium sand with minor coarse sand (Fig. 3.5). The facies shows a general upward fining from fine sand to coarse silt (Fig. 3.6). Strata are reverse graded, normal graded, or massive. Graded beds occur predominantly in strata that consist of upper-fine sand. Beds are generally < 3 cm thick and form cosets < 20 cm thick. Bed contacts range from sharp, typically with overlying heavy mineral lamina, to diffuse where no heavy mineral concentration is present. In contrast, massive fine sand beds are up to several metres thick and have dispersed heavy minerals and < 2 % coarse sand and granules. Grain size analysis shows the sand to be moderately well-sorted to poorly sorted, and fine-skewed mesokurtic or leptokurtic. Silty layers are generally more poorly sorted. Clay content is < 0.5 %, even where silt dominates (Fig. 3.6). Massive, medium to coarse sand has a coarse tail of dispersed granules and pebbles (Fig. 3.6). Isolated silt intraclasts are generally < 0.5 cm in diameter. At ~133 m asl, there is a layer of abundant, 0.5-3.0 cm thick, silt clasts within massive medium sand. The largest silt intraclasts have no surface armour and are concentrated in a horizon 12 cm thick (Fig. 3.5d). Rare (< 0.25 % of the facies) silt lamina, < 0.5 cm thick, and with gradational basal contacts occur in otherwise massive sandy silt. The laminae are either continuous across the core, loaded, or discontinuous. When the core dries, massive sand beds commonly have either a minor horizontal desiccation fracture pattern, a vertical fracture network, or a more three dimensional blocky fracture network (Fig. 3.6). Zones of vertical desiccation structures have thicknesses > 2m and fracture spacing ranges from ~0.5 to 1 cm. In one instance graded fine sand beds curve downward 15-20 cm along one side of the core (Fig. 3.7c).

Interpretation: The diffusely graded-massive lithofacies is interpreted to have been deposited some distance downflow of a hydraulic jump by a succession of depositional processes responding to changes in flow speed, sediment concentration and rate of deposition. The transition from supercritical to subcritical flow in a closed conduit system is generally smooth and a hydraulic jump does not develop. If the region of flow expansion has an adequately large volume relative to the discharge volume and the higher density discharge can expand unhindered by confining pressure then a hydraulic jump will occur. In this case development of the jump requires an adequate density contrast between the sediment laden flow and ambient basin waters. A subglacial lake centred in the Lake Ontario basin may have provided a location for flow expansion. The depositional mechanisms of massive and reverse-graded strata have received extensive treatment in the sedimentological literature during the past 30 years and are presently the focus of vigorous debate (e.g. Kneller and Branney 1995; Shanmugam 1996; Shanmugam 1997; Sohn 1997). These strata have been variously interpreted to be deposits from, rapid suspension sedimentation (Hiscott 1994a), sweep-fallout processes (Hiscott 1994b), sandy debris flows (Shanmugam

1996), grain-flows (Lowe 1982), traction carpets (Sohn 1997; Todd 1989), hyperconcentrated flows (Kneller and Branney 1995), and surge deposits (Vrolijk and Southard 1997). Because of the general lack of internal scours, the absence of "spaced stratification", or other evidence for turbulent flow, the sweep-fall out model inadequately explains the origin of these strata. Also, it is improbable that sediment was deposited directly from debris flows because the clay-size fraction is generally < 5 % (Fig. 3.6) and therefore is insufficient to provide adequate cohesive strength to the flow (Hampton 1975). Common to the last five mechanisms is a highly concentrated, generally non-turbulent region of the flow where fluid turbulence provides negligible sediment support. It is crucial to recognize that no one mechanism was likely active along the full depositional length of the flow and that temporal and spatial variations in flow energy and sedimentation rate would have resulted in changes in the active depositional processes. In this study, deposition in the most proximal region, where sediment concentrations were highest, was likely from rapid suspension sedimentation due to loss of transport capacity (e.g. Hiscott 1994a) or from a laminar sheared layer (e.g. Vrolijk and Southard 1997). The predominant depositional mechanism, however, is interpreted to have been a traction carpet (e.g. Sohn 1997)(Fig. 3.8). With progressive decrease in sediment concentrations and a change in the basal flow region from hyperconcentrated to a fully turbulent and fluidal flow, deposition was by tractional processes.

Traction carpets are a transient part of stratified flows with a lower hyperconcentrated dispersion and an upper fluidal zone. Depending upon the sediment load and flow velocity, the traction carpet changes in thickness and rheology (Sohn 1997). The principal support mechanisms in the traction carpet are hindered settling and buoyancy. Where fine sand is the dominant grain size, dispersive pressure is not a significant support mechanism due to the small mass of individual grains and insufficient grain momentum (Lowe 1976). Additionally, because of high sediment concentration, secondary turbulence is suppressed but large-scale turbulence may be present even at sediment concentrations > 15 % by volume (Pierson and Scott 1985). Traction-carpets ideally consist of two superjacent layers: a basal frictional zone and an overlying collisional zone (Fig. 3.8) (Sohn 1997). Boundaries between the zones are defined by sediment concentration of ~50 % and 9 % by volume in the lower frictional zone and upper collisional zone, respectively (Fig. 3.8). Grain mobility is limited in the lower friction zone where grains are in near continuous contact. In contrast, the upper collisional zone is a region of highly mobile particles and active sediment sorting because of intense shear exerted by the overlying turbulent flow and large gradients in particle concentration. The relative thickness of these two regions is dynamic and determined by three principal variables: applied shear stress, flux of settling sediment, and grain size (Sohn 1997). Thus, the thickness of the frictional zone increases with decreasing shear stress and increasing sediment flux. Additionally, flows dominated by finer sediment (sand) develop thicker frictional regions.

Deposition of inverse to normally graded beds requires that the traction-carpet have a thin frictional layer and a relatively thick collisional layer (Sohn 1997). This flow structure provides the greatest amount of time for sediment sorting in the upper collisional zone. By contrast, massive beds were most likely deposited from flows with a thick frictional region. These relationships correspond with differences in vertical sediment fluxes and high rates of sedimentation that produced more massive strata. Deposition from the traction carpet occurs because of frictional freezing and takes place from the base up (Kneller and Branney 1995; Sohn 1997). This mechanism differs from earlier traction-carpet models where deposition was by en masse plug deposition (Lowe 1982). Because of this important difference, the relationship between flow thickness and deposit thickness, as previously reported by Lowe (1982), can not be applied to these strata.

The traction-carpet model (Sohn 1997) thus provides a single depositional mechanism for both inversely-normally graded strata and massive strata. Massive beds, however, can also be formed by a variety of other syn- or post-depositional processes related to rapid sedimentation or post-depositional dewatering. For example, some massive beds have been interpreted to be the result of rapid suspension sedimentation and inadequate time for sediment sorting at the bed (Arnott and Hand 1989). Under fully turbulent flow conditions, Arnott and Hand (1989) observed a transition from laminated to diffusely laminated and eventually massive beds with increasing rates of suspension deposition. This transition was reported to be the result of reduced lateral segregation of sediment grains because of rapid burial and steep angles of climb of any persistent low-amplitude bed forms, such that otherwise distinct laminae degraded into diffuse textural bands. Massive conditions were achieved at aggradation rates of ~4 cm min^{-1} and at relatively low sediment concentrations, probably less than <10 %, for an upper fine sand (Mz = 0.23 mm;) (Arnott and Hand 1989), this is a grain-size similar to much of the sediment in this study. One important constraint for considering this mechanism is that flow conditions in the Arnott and Hand experiments were critical; however, evidence of upper plane-bed stratification in this study was observed at only one location (Fig. 3.4, 129 m). At this location planar-laminated sand graded upward into massive sand. More commonly interbedded dune-scale cross-stratification is observed, suggesting subcritical flow conditions.

Post depositional dewatering is also a possible mechanism for the formation of massive sediment, either soon after deposition (e.g. Lowe 1982; Lowe and LoPiccolo 1974), or alternatively during sampling. Rapidly deposited sand has a loosely packed framework that as a consequence is susceptible to post-depositional disturbance (Kneller and Branney 1995; Lowe 1982). Evidence for syndepositional dewatering, including dish or pillar structures, is generally absent. The absence of dish structures is possibly related to the low clay content and absence of micaeous minerals in the sand. Where dish or pillar structures are observed in glacifluvial sediment, the sand is poorly sorted and has thin clay lamina (e.g. Cheel and Rust 1986).

3.2.4 Cross-stratified sand

The cross-stratified sand lithofacies consists of fine to medium sand overlying abruptly and gradationally graded to massive sand. Cross-strata above 171 m are predominantly small-scale cross-laminated, whereas below 171 m medium-scale cross-strata occur exclusively and form rare interbeds in graded-massive sand strata (Fig. 3.5, 3.9). Small-scale cross-laminated sets are < 2-3 cm thick with coset thickness > 15 cm. Measurement of the maximum coset thickness was obscured by drilling induced dewatering structures and the dry crumbly nature of the sand. Small-scale cross-lamination occurs in both non-climbing and climbing sets. Climbing sets are stoss-erosional and have low angles < 15° of climb. Detrital organic particles are < 1 mm in size and commonly occur along foresets. Medium-scale cross-beds are < 3 cm thick and form isolated sets up to 15 cm thick. Cross-strata are highlighted by heavy mineral concentration along foresets and are normally graded. Three samples from this facies are moderately well sorted upper fine - medium sand with 3-8 % silt, and exhibit finely skewed leptokurtic grain-size distributions (Fig. 3.7). Total organic content of this facies is < 0.001% (Fig. 3.5). Deformation in this facies is restricted to minor faults with < 1-2 cm offset and folded laminae. Pipes form vertical structures up to 70 cm long that grade from a coarse-grained core to finer grained limbs (Fig. 3.7a). With distance above the base of larger pipes heavy mineral concentration typically decreases and the sediment has a more uniform texture.

Interpretation: This facies was deposited under turbulent flow conditions and traction sediment transport (e.g Harms 1982; McDonald 1972). Above 171 m asl in DH-V-158 there is an upward transition from medium-scale cross-bedding to small-scale climbing cross-laminae coincident with an increase in silt content and frequency of clay interlaminae. Within the lower 50 m of channel fill interbedding of the cross-stratified and graded-massive facies indicates deposition from unsteady non-uniform flow. The general lack of thick, medium-scale cross-laminated sets likely indicates that flows were unsteady and passed rapidly through the dune stability field or alternatively, were too shallow, or dune development was suppressed by high sediment concentration. The cross-stratified facies was thus deposited episodically from a flow that fluctuated between deposition by tractional bedload transport and traction carpet mechanisms (e.g. Kneller 1995). Stratigraphically higher cosets of small-scale climbing cross-lamination record the seasonal meltwater discharge following the end of the jökulhlaup discharge along the tunnel channels. The low-angle of climb and stoss-erosional form of small-scale cross-lamination indicates that sedimentation was dominated by traction and not suspension sedimentation (e.g. Jopling 1968).

Deformation in this facies was penecontemporaneous or alternatively was disturbed during sample collection. Deformation of small-scale cross-lamina is interpreted to have been induced by drilling based on the similar grain size of sand coating the core perimeter and sand found within adjacent fractured parts of the core. Vertical pipe structures on the other hand, are less clearly drilling induced. Pipes are similar to pillar structures described by Lowe (1974) and indicate fluidization of the sediment (Fig. 3.7). A definitive assignment, however, to in situ versus sampling related disruption, remains equivocal.

3.4 Depositional Model

The lateral and vertical arrangement of the lithofacies described and interpreted above record different styles of deposition within a dynamic and evolving subglacial tunnel channel environment. Hereafter the cross-laminated sand and graded-massive sand facies are collectively termed the sand facies association. The following discussion first reviews the evidence that suggests the sediment forms part of a tunnel channel fill, and then develops a five-stage depositional model for channel erosion and fill.

3.4.1 Tunnel channel setting

North and south of the Oak Ridges Moraine drumlins and tunnel channels along the top of Newmarket Till form part of a regional Late Wisconsin unconformity. The tunnel channels form a network of steep walled, 0.5 to 6 km wide, southwest-trending valleys (Barnett 1990; Brennand and Shaw 1994; Sharpe et al. 1997). Their southward downflow continuation beneath the Oak Ridges Moraine has been confirmed by seismic reflection surveys (Pugin et al. 1999). Drillhole DH-Nob occurs near the eastern margin of the buried Holland Marsh valley tunnel channel and DH-V-158 is located ~10 km to the southeast in a subparallel channel (Fig. 3.2). Regionally the lower deposits of pre-Upper Wisconsinan age have a characteristic total organic carbon content (TOC) that for the Scarborough Formation ranges from 0.01 -1.35 % (Eyles and Williams 1992). Locally, Scarborough Formation silt and sand in DH-Nob have TOC values of 0.38-0.65%. In contrast, maximum TOC values reported from DH-V-158 are 0.17 % with an average of 0.034%. Interestingly, the highest TOC value in DH-V-158 is from a basal diamicton overlying bedrock. The values reported here are thus interpreted to represent regional background values for sediment derived, in part, from the Scarborough Formation. In DH-Nob the Thorncliffe Formation does not contain significant organic material, although the fine-grained muddy sediment is commonly intensely bioturbated. Additionally, at DH-Nob the Thorncliffe Formation consists of a thick succession of approximately 1000 silt-clay couplets interpreted as varves. No similar bioturbated sediment or varve succession is present in DH-V-158 (Fig. 3.2b).

The regional unconformity truncates Newmarket Till, or where the till has been eroded, Thorncliffe or Scarborough formation silt and sand. At DH-Nob, lacustrine silt and clay of the Thorncliffe Formation have been truncated and are overlain by gravel deposits. The erosional surface can be traced westward on seismic records where it descends to bedrock and truncates horizontal reflectors in the Thorncliffe and Scarborough formations (c.f. Pugin et al. 1999). East of DH-Nob the erosional contact rises and incises the Newmarket Till (Pugin et al. 1999). This unconformity is identified on downhole geophysics (Eyles et al. 1985; Russell and Pullan 1998) by abrupt changes in downhole gamma and conductivity profiles. Downhole gamma logs have well developed inverted, bell-shaped curves beneath the unconformity, whereas above the unconformity the data have a random jagged pattern (Russell and Pullan 1998).

3.4.2 Setting - subglacial or proglacial

A depositional model for these deposits must address a number of questions concerning the location of channel formation, for example, whether it was proglacial or subglacial. Additionally, if subglacial, did the channels form near the ice margin further up-ice. In this study the channels are interpreted to have been cut and subsequently infilled in a subglacial setting for the following five reasons: i) inferred position of the regional paleo-ice margin, ii) elevation of the gravel, iii) depth and form of the channels, iv) presence of overlying Halton Till, and v) absence of proglacial fluvial or lacustrine sediment successions.

Regional models of deglaciation constrain ice margin position at this time and do not indicate proglacial subaerial environments in the study area prior to formation of the Oak Ridges Moraine (e.g. Barnett 1992). Additionally, these models suggest that the ice margin was south of Lake Ontario and thus channel formation developed at least 60-100 km up-ice north of the ice-margin. If the area was deglaciated prior to formation of the Oak Ridges Moraine, regional isostatic constraints and an ice-marginal position abutting the Niagara Escarpment would probably have maintained glacilacustrine conditions. As demonstrated by Chapman (1985), lake levels would have been above the 200 m elevation at which gravel presently occurs in DH-Nob. Furthermore, on the basis of channel form (e.g. undulating thalweg) Brennand and Shaw (1994) suggested that erosion of channels to the east of the study area was by pressurized subglacial meltwater flows. In this study, both channel fill successions subcrop beneath Halton Till, which represents a fine-grained glacilacustrine diamicton complex deposited in an ice-marginal or subglacial position (Russell and Arnott 1997). The Halton Till is conformable with the earlier depositional stages of Oak Ridges Moraine (Barnett et al. 1998). The presence of Halton Till required ice to have been in the area to sustain elevated lacustrine environments.

Alternative sedimentary environments for deposition of channel fill successions include: braided fluvial, subaqueous fan, and glacilacustrine. Each of these depositional environments has unique lithofacies and lithofacies associations that can differentiate one from the other. Sandy proglacial environments are typically characterized by braided fluvial systems that consist of channels, traction bedforms, and overbank deposits (e.g. Church 1972). Sand dominated braided rivers have abundant erosional surfaces, trough and planar cross-bedding, and cross-lamination; massive sand is rare (Miall 1977). In sandur deposits dominated by jökulhlaup flows, which represent short-lived catastrophic fluidal and hyperconcentrated events, sandur deposits have distinct lithofacies and lithofacies associations of which a significant element is massive sand (Maizels 1993). Aside from thickness, massive sandur sand deposits and deposits of this study would be difficult to differentiate in core. Massive sand is also a common element of subaqueous fan deposits where it is commonly associated with abundant small and medium-scale cross-stratification, and particularly climbing cross-stratification (e.g. Gorrell and Shaw 1991). No such association is observed in the sand facies association presented. In contrast to the proximal jökulhlaup-dominated environments, deep-water lacustrine deposits are generally characterized by cyclic successions 10's - 100's cm thick that consist of rhythmically stratified sand and silt with interstratified clay horizons (e.g. Gilbert 1997). Deposits similar to these occur above 171 m asl in DH-V-158 and are interpreted to represent a later low-energy stage of the fill succession (Fig. 3.2). The gravel and sand facies associations are thus interpreted to be unrelated to proglacial sandur outwash deposits or glacilacustrine basin fill sediment on the combined criteria of regional ice margin positions and sedimentology. Consequently, the stratal succession described here is interpreted to have been deposited by rapid subglacial sedimentation.

3.4.3 Depositional dynamics

The erosion-depositional model must thus account for a subglacial channel origin and subsequent subglacial sedimentation. Lithological similarities with outwash jökulhlaup deposits support the interpretation that the depositions was from episodic outbreak floods or jökulhlaups. A five stage model is developed (Fig. 3.10). Each stage is characterized by a single major physical process that together forms a continuum.

Stage I - Initial Conduit: Models of tunnel channels produced by subglacial meltwater suggest a systematic progression from sheet floods and drumlin formation, to channelization of the unsteady waning sheet floods and associated channel deepening (Shaw 1996). Collapse of the erosional meltwater sheet flow that formed the drumlins was routed through a system of channels, cavities and connecting constrictions along the drumlinized Newmarket Till surface (e.g. Brennand 1994). The composite cross-section (Fig. 3.2c) illustrates the relationship between drumlinized Newmarket Till (D), with shallow channels (TC^2), and broader over-deepened channels (TC^1). The relationship between drumlins and shallow channels is most clearly visualized on a digital elevation model of the area (c.f. Kenny 1998).

During development of tunnel channels variations in the geometry of subglacial meltwater pathways formed areas of high- and low-speed flow. The gravel lithofacies is interpreted to have been deposited during the early formative stage of the channel system (Fig. 3.10a). Gravel was transported primarily by downstream migrating dunes and bedload sheets. Channel switching, channel erosion, channel margin collapse or collapse of the conduit roof may have resulted in rapid-velocity changes, even locally, resulting in channel abandonment and gravel deposition (Fig. 3.2, 3.10a). Maximum flow depths estimated from the height of the channel during this stage were generally < 40 m (Fig. 3.2). Depths of this magnitude are consistent with flow depth estimates commonly interpreted for subglacial tunnel channel flows based on scaling relationships of dune-scale cross-stratified gravel (e.g. Shaw and Gorrell 1990). Flow depths may have been less but there are no data to further constrain these estimates. Gravel deposition at DH-Nob occurred prior to incision of the deeper channel to the west of the drillhole (Fig. 3.2c).

Stage II- Erosion of Newmarket Till: The regionally extensive Newmarket Till is the substrata of the initial tunnel channel network. Regional studies have shown the thickness of Newmarket Till to be highly variable (Sharpe et al. 1999). Stone lines, plus sand and silt interbeds suggest earlier, probably smaller-scale, meltwater events occurred during deposition of the Newmarket Till (Sharpe and Barnett 1997). This dense, relatively consolidated, heterogeneous till was generally resistant to meltwater erosion. Local scouring occurred where large vortices were generated in zones of intense shear, such as at channel confluences. Erosion at confluence scours commonly forms depressions of the order of 5 times the depth of the channel system (e.g. Best 1987). Confluence scours are the deepest point of erosion along a channel system (Bristow et al. 1993). These scour holes would have locally eroded the Newmarket Till and unroofed older, less consolidated sediment of the Thorncliffe and subsequently Scarborough formations. Channel enlargement would have accelerated significantly once these less competent sediment were exposed.

Stage III - Channel deepening: Shallow channels overlie Newmarket Till, whereas deeper channels are eroded into the strata of the Thorncliffe or Scarborough formations, and in places eroded to bedrock (Fig. 3.2). At DH-Nob, the eastern channel flank is underlain by a thick succession of rhythmic silt-clay (Russell and Pullan 1998). This indicates that lateral facies changes within the substrata from cohesive silt clay to less cohesive sand likely influenced the location of channel deepening and produced the asymmetrical profile of the channel cross-sections (Fig. 3.10c). Depending on the erosional mechanism of the scour, an equilibrium depth would be achieved. Further erosion was then only achieved by a combination of sediment piping and/or slumping. Further scour required either deepening of the channel upstream or more probably headward erosion of the scour (c.f. Nordin 1963).

Scour dimensions increased as the scour eroded into permeable sand sediment underlain by less permeable sediment and piping within the sand aquifers occurred. An important constraint on piping is the rate and quantity of subglacial water infiltration. Under a glacihydraulic head, infiltration was probably equivalent to or greater than modern subaerial infiltration rates (e.g. Piotrowski 1997). Also, numerical modelling by Piotrowski (1997) suggests a relationship between the magnitude of groundwater flow and tunnel channel formation. Consequently, pore fluid pressures within confined aquifers of the lower deposits possibly became over-pressured once overlying cohesive, low-permeability confining sediment was eroded (Fig. 3.10). Large tunnel channels may thus coincide with areas that had the highest potential for piping and related sediment excavation. Even today substrates exposed along riverbanks and lakeshores undergo extensive slope failure as a result of sediment piping (e.g. Barnett and Kenny 1997). As overlying cohesive sediment was undercut, slumping along the scour margins occurred. Turbulence generated in the scour caused more sediment to be eroded and mixed into the overlying flow and transported downflow (Fig. 3.11). The dense clouds of sediment added into the flow produced flows with high suspended loads. These highly concentrated dispersions rapidly became stratified with a basal hyperconcentrated layer. Consequently, this stage provided a mechanism to rapidly scour and entrain sand from Thorncliffe and Scarborough formations. The resultant unsteady flow would have been highly sediment charged, and at the limit of its transport capacity.

Stage IV - Sand facies deposition: Deposition is inferred to have occurred at a zone of flow expansion, possibly along the grounding line of a subglacial lake centred in Lake Ontario. At this zone, particularly if flow was supercritical and a hydraulic jump developed, intense turbulence, rapid dissipation of flow velocity, and diminished transport capacity resulted in high rates of suspension sedimentation (e.g. Hiscott 1994a). Some distance downflow deposition was from a laminar sheared layer (e.g. Vrolijk and Southard 1997) and then with increased flow stratification a traction carpet (e.g. Sohn 1997). As the stratified flow developed a basal hyperconcentrated region and an overlying dilute fully turbulent region deposition was from either the laminar shear or traction carpet. Inverse-normal graded and massive sand layers were deposited by these two mechanisms (Fig. 3.9). So long as the laminar surge or traction-carpet was sustained, inverse-graded or massive sediment was deposited. During periods when large volumes of fine sand were being supplied to the traction-carpet, the traction carpet thickened with diminished sorting efficiency within the collisional layer and thickening of the basal frictional layer. This in turn resulted in the deposition of massive sand (Sohn 1997). If the flux of fine sediment was reduced significantly, deposition was replaced by a thin traction or bedload layer (Sohn 1997). A subsequent increase in the rate of suspension sedimentation regenerated the hyperconcentrated flow, and consequently a return to inverse-normal or massive sand deposition. The hyperconcentrated region is considered to have existed in the lower part of the flow, and did not necessarily extend the full flow depth.

The ~ 50 m thickness of the sand facies association and general absence of evidence for discontinuous sedimentation or erosion, suggest it was deposited rapidly and probably quasi-continuously. Previous work by Arnott and Hand (1989) suggested that massive sand deposition occurred under conditions of rapid suspension fallout and related bed aggradation of ~ 4 cm min^{-1}. Intermittent rates of bed aggradation may have been equal to or higher than those observed by Arnott and Hand. Accumulation rates for the 50 m succession, which includes rare layers of cross-stratified sand, and silt laminae were probably lower. The evidence suggests that the basal 50 m succession was deposited during a single meltwater season and likely from a single quasi-continuous unsteady flow with temporal fluctuations ranging from seconds to days (e.g. Russell and Knudsen 1999).

Stage V: *Low-Energy Stage*: From 170 m to ~205 m in DH-V-158 (Fig. 3.3) a fining upward succession dominated by climbing small-scale cross-laminae (ripple-drift cross-stratification) and micro laminated silt - clay rhythmites occur. Upward, rhythmites become progressively thinner and finer grained. The sand facies association is not present above 179 m (Fig. 3.4). The rhythmite succession marked the end of rapid channel infill and the change to a lower energy seasonal meltwater discharge regime (Fig. 3.10d) that was controlled by diurnal and climatic influences (e.g. Gilbert 1997). In DH-Nob the absence of basinal varves (Russell et al. 1998) above the gravel suggests that ice became reattached to the substratum at this site earlier than along the deeper parts of the channel axis.

3.5 Discussion

3.5.1 Interregional lithofacies and channel fill correlation

Investigations of deeply buried channel fill strata have been limited to seismic surveys, drillhole geophysics, and water-well wash-boring records. Although voluminous, these datasets provide inadequate resolution to constrain the origin of tunnel channel fills or provide a detailed understanding of their depositional infill history. Thus it is important to link these studies to detailed sedimentological data such as presented in this paper.

Seismic attributes of tunnel channel fills have been well documented from the Scotian shelf (Boyd et al. 1988), Finger Lakes (Mullins and Hinchey 1989; Mullins et al. 1996), Oak Ridges Moraine (Pugin et al. 1999), and, the North Sea (Wingfield 1990). In all of these studies, however, cores have failed to penetrate the deeper parts of the channel fill and consequently interpretations based on seismic facies lack critical ground-truthing. Accounting for variations in seismic resolution, each of these studies describes relatively similar seismic facies forming the channel fills (Table 3.1). On the basis of reflector characteristics, channel-fill sediment was interpreted to be silt, sand or till (Table 3.1). Seismic data from the lower 120 m (below 200 m asl) of the Holland Marsh tunnel channel in the Oak Ridges Moraine area have been interpreted to document deposits of rapid suspension sedimentation from a hyperconcentrated dispersion (Pugin et al. 1999). Strata with similar seismic facies have been interpreted to be deposits of rapid suspension sedimentation in a subglacial environment (Mullins and Hinchey 1989), subglacial till tongues (Mullins et al. 1996), proglacial sedimentation (Mullins et al. 1996), and plunge pool sedimentation (Wingfield 1990). The contrasting interpretations of Mullins and Hinchey (1989) and Mullins et al. (1996) for the same strata succession illustrate the problems associated with seismic facies interpretation from deep buried channel-fills lacking drillcore control.

The sand facies association in DH-V-158 fills the lower 50 m of a ~100 m deep tunnel channel (Fig. 3.2) This tunnel channel trends subparallel to, and forms part of the same channel network as the Holland Marsh channel intercepted by DH-Nob. Infill of the channel network, at least locally, is interpreted to have been synchronous. Based on detailed sedimentology the sand facies association is interpreted to be deposited from a hyperconcentrated dispersion. Because of the thick succession of massive sand with subordinate traction deposits, and general absence of major erosional surfaces, (Fig. 3.4) the sand facies is likely lithologically similar to strata in the deeper part of the Holland Marsh tunnel channel. The later are seismically typified by low-amplitude reflectors or an absence of reflectors and on this basis have been previously interpreted as deposits of hyperconcentrated dispersions (Pugin et al. 1999). On the other hand, the sand facies association provides no direct sedimentological evidence for the origin of the chaotic seismic facies commonly reported from tunnel channel fills (Pugin et al. 1999; Wingfield 1990). Seismic data indicate that the chaotic facies varies in lateral extent and vertical distribution and becomes less common with increasing distance from the margins of the tunnel channel (c.f. Pugin et al. 1999).

Although the depositional model presented here suggests that slumping was an important channel widening process, only limited evidence for slumping was observed in core. One possible explanation for this apparent inconsistency is that extensively dewatered strata potentially cause extensive back scattering of seismic energy, which was misinterpreted to be a chaotic reflection signature related to slump blocks. Alternatively, DH-V-158 may have penetrated a stratal succession beyond the channel margin apron where slump deposits are common.

Tunnel channels eroded into both bedrock and unconsolidated sediment are infilled with similar seismic facies. The sediment - seismic facies correlation and the depositional model presented suggest rapid and near instantaneous erosion and sedimentation of the channel. The model is most applicable to tunnel channels in unconsolidated sediments. Where tunnel channels are eroded in bedrock, additional mechanisms are necessary to produced enhanced erosion (e.g. structural control). Additionally it is improbable that infill sediment can be derived directly from the erosional event. Consequently it is necessary to identify additional sediment sources up flow and subglacially. Widespread distribution of the seismic facies suggests that rapid sedimentation in tunnel channels is a common event related to channel formation and probable jökulhlaup flow that remains to be fully elucidated. In channels incised into unconsolidated sediment, rapid deposition from sediment charged flows is possibly partially controlled by rapid scouring and over-deepening and widening of channels. Both the Holland Marsh and the buried channel intercepted by DH-V-158 are two of the widest and deepest tunnel channels within the Oak Ridges Moraine channel network (Fig. 3.1). Deep channel fills elsewhere in the Oak Ridges Moraine area are, by contrast, predominantly traction-flow deposits (e.g. Pugin et al. 2000).

3.5.2 Channel fills and channel origin

The fill of tunnel channels has been suggested by Cofaigh (1996) to provide little information from which to deduce the process of valley incision. This conforms with the generally held belief that sediment overlying an unconformity commonly has no genetic relationship with the erosional event that formed the unconformity. In the Oak Ridges Moraine area, however, the proposed channel fill model suggests an intimate link between erosion and fill of the tunnel channels. Regional Laurentide glacial models indicate that the channels were eroded into Upper Wisconsinan Newmarket Till. The Newmarket Till was deposited during or subsequent to the Late Wisconsinan maximum at ~20 ka (Barnett 1992). Sediment overlying the Newmarket Till, principally the Oak Ridges Moraine is interpreted to have been deposited around 13.5 ka (Barnett 1992). Consequently, the erosion and fill of the tunnel channel network is constrained within an ~6-7 ky period. To the south, erosion and fill of the New York Finger Lakes tunnel channels has been interpreted to have occurred between ~14.5 to 14.0 ka BP (Mullins et al. 1996). Assuming a similar or younger age for channel formation in Oak Ridges Moraine area suggests a close link between erosion and infill events during a period of <1 ky. Additionally, the depositional model proposes a continuum of events that link channel erosion and subsequent channel fill. It is interpreted that high energy scouring eroded sand from underlying deposits and redeposited it as channel fill. The sediment facies and depositional model indicate that the flows were related to episodic, catastrophic meltwater outbreak floods or jökulhlaups. This suggests that the tunnel channels were formed by non-steady state processes and not steady state processes, and thus supports the conclusion by Shoemaker (1991) that the Laurentide ice sheet did not generally drain by steady state processes.

3.6 Conclusion

The basal fill succession of two buried tunnel channels beneath the western Oak Ridges Moraine consists of a gravel and a sand dominated facies associations. The gravel facies is ~ 17 m thick and occurs along the stepped eastern margin of a 4-5 km wide channel with a ~170 m deep thalweg to the west. The gravel facies is interpreted to have been deposited from turbulent tractional flows prior to an episode of significant channel deepening. Laterally from the channel margin the gravel facies records increased flow velocities as inferred from the change from stacked mesoforms to low relief bedload sheets. Initial scour of the Newmarket Till and channel deepening occurred by differential erosion by large vortices. These vortices may have been associated with channel confluences or large hairpin vortices along drumlin margins. Unroofing of less competent strata beneath the till established conditions for more rapid channel deepening and widening. In addition, high pore fluid pressures in underlying sand beds developed horizontal groundwater movement toward low pressure zones that coincided with the scours. This groundwater flow initiated sediment piping and transport of sand into the channel. Sand mobilized from these underlying beds became part of the sediment load entrained by the turbulent flow related to the scouring. Deposition of the sand facies association occurred downflow of a grounding line within a subglacial lake. Flow expansion of a supercritical flow at the grounding line terminus of the conduit formed a hydraulic jump. As flow rapidly lost transport capacity and became laden with sediment the flow became stratified. The basal layer of this stratified hyperconcentrated flow formed a traction carpet. Rapid sedimentation occurred from this basal layer of the flow where turbulence was suppressed. During low rates of discharge and sediment concentration, cross-stratified sand was deposited from bedload traction transport. During higher rates of discharge and sediment concentration, inverse-normal graded or massive sand, were deposited by a basal traction carpet. The sand facies association of cross-stratified and graded or massive sand forms a 50 m thick succession that infills the lower 50 m of a 100 m deep channel (Stage 1, Fig. 3.2b). The remaining channel fill of sand-silt-clay rhythmites was deposited under lower energy conditions of associated traction and suspension sedimentation (Stage 2, Fig. 3.2). This marks the end of jökulhlaup discharge and a return to diurnal seasonal meltwater discharge.

The sand facies association of cross-stratified, inverse-normal graded or massive sand is correlated with low-amplitude reflectors and chaotic seismic facies (Pugin et al. 1999) from the thick, deeply buried channel fill west of DH-Nob. Additionally, the sand facies association is interpreted to be analogous to low amplitude seismic reflector facies reported from other deep channel fills in eastern North America (Boyd et al. 1988; Mullins et al. 1996) and the North Sea (Wingfield 1990), and provides for the first time direct sedimentological evidence for the origin of this common but enigmatic, seismic facies.

Fig. 3.1. Location of study area in Great Lakes region and distribution of first order tunnel channels north of the Oak Ridges Moraine (dark grey). Upper figure modified after Sharpe et al. (1996, fig. 5). Numbers on inset map refer to selected tunnel channel studies in the Great Lakes region, [1]Wright, 1974; [2]Muller and Hinchley, 1989; [3]Mooers, 1989; [4]Barnett, 1990; [5]Shaw and Gorrell, 1990; [6]Brennand and Shaw, 1994; [7]Patterson, 1994; [8]Pugin et al. 1996, 1999; [9]Pair, 1997; [10]Muller et al. 1997; [11]Fisher and Taylor, 1999.

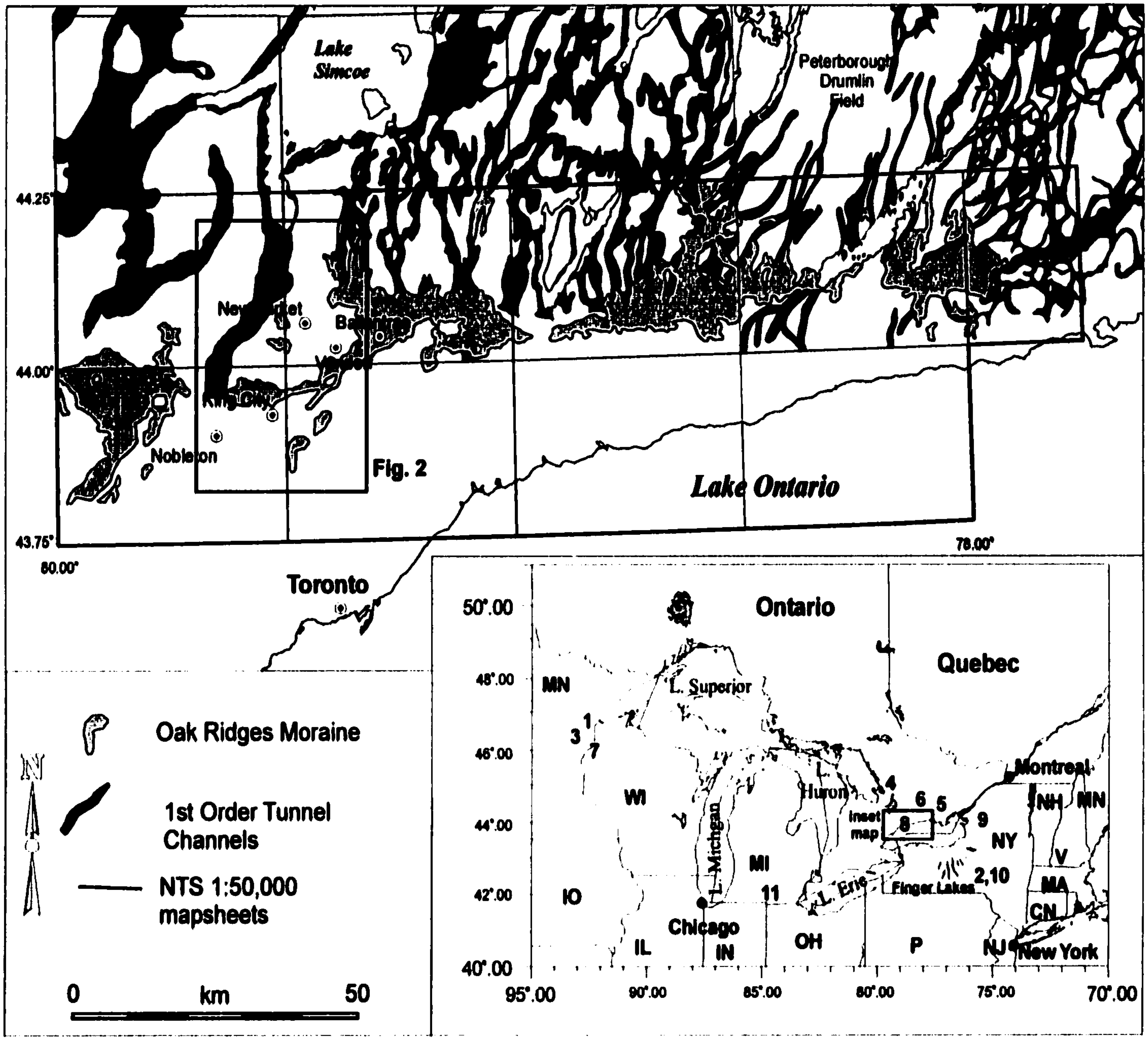

Lake Simcoe
Peterborough Drumlin Field
44.25°
44.00°
43.75°
80.00°
78.00°
Newmarket
Nobleton
King City
Fig. 2
Lake Ontario
Toronto
Oak Ridges Moraine
1st Order Tunnel Channels
NTS 1:50,000 mapsheets
0
km
50
N
Ontario
Quebec
L. Superior
Huron
L. Michigan
L. Erie
Finger Lakes
Montreal
Chicago
New York
inset map
MN
WI
IO
IL
IN
MI
OH
P
NY
NJ
NH
MN
V
MA
CN
1
2,10
3
4
5
6
7
8
9
11
50°.00
48°.00
46°.00
44°.00
42°.00
40°.00
95°.00
90°.00
85°.00
80°.00
75°.00
70°.00

Fig. 3.2. Geometry of tunnel channels and location of drillhole data with respect to channels. Data synthesized from Pugin et al. 1999; Pullan unpublished; Kenny et al. 1999; Sharpe et al. 1997. a) map of tunnel channels in the vicinity of the study area, b) stratigraphic cross-section across channel interflueve, c) cross-section across the buried Holland Marsh tunnel channel showing deeply incised first order (TC^1)and shallow second order (TC^2) channels of Newmarket Uplands. Note that where the Newmarket Till is eroded the tunnel channels are deepest.

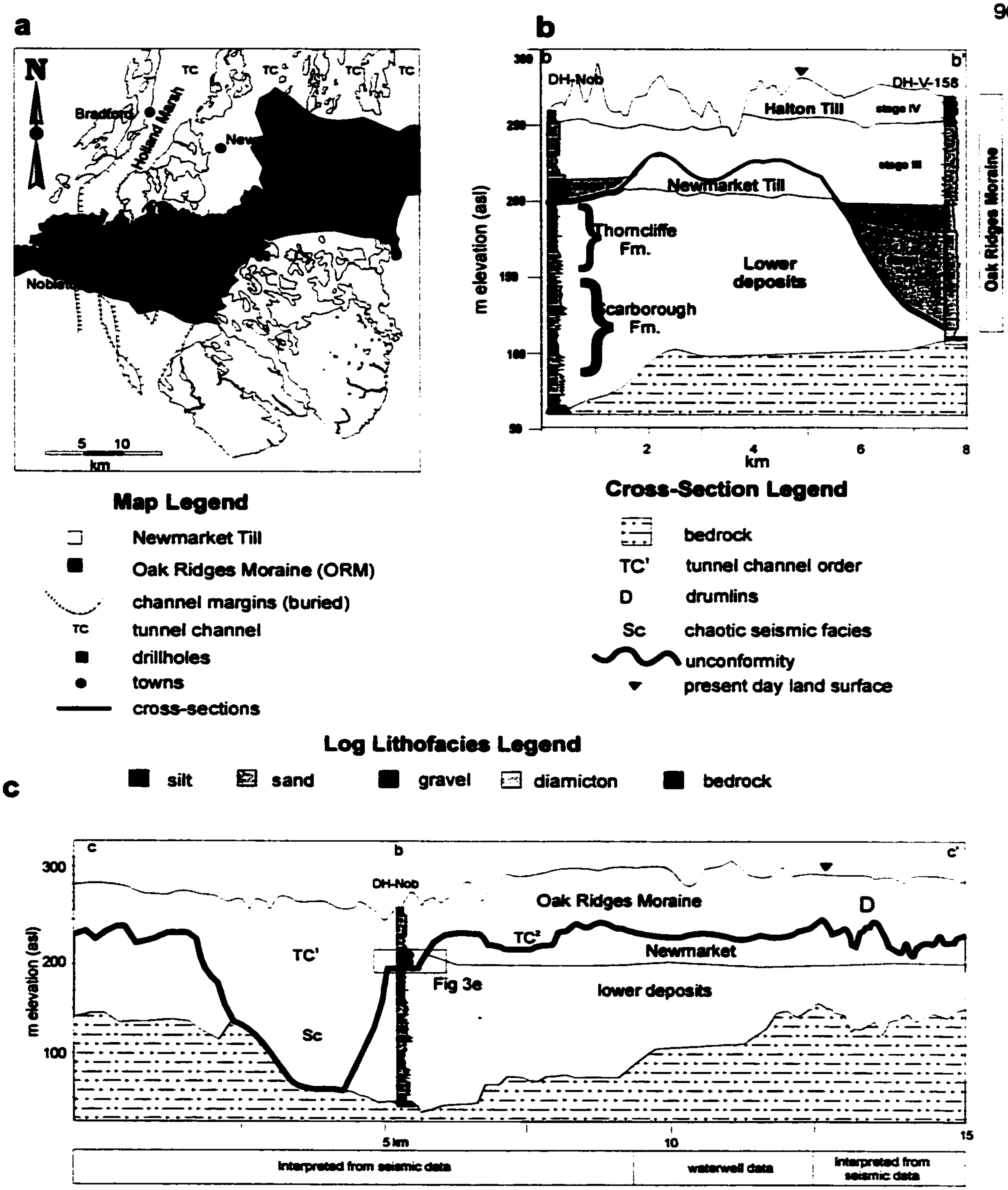

a
N
TC
Bradford
Holland Marsh
Nobleton
5 10
km
b
b
b'
DH-Nob
DH-V-158
Halton Till
stage IV
stage III
Newmarket Till
Thorncliffe Fm.
Lower deposits
Scarborough Fm.
Oak Ridges Moraine
m elevation (asl)
2
4
6
8
km
Map Legend
Newmarket Till
Oak Ridges Moraine (ORM)
channel margins (buried)
tunnel channel
drillholes
towns
cross-sections
Cross-Section Legend
bedrock
TC' tunnel channel order
D drumlins
Sc chaotic seismic facies
unconformity
present day land surface
Log Lithofacies Legend
silt
sand
gravel
diamicton
bedrock
c
c
b
c'
300
200
100
m elevation (asl)
DH-Nob
Oak Ridges Moraine
D
TC'
TC²
Newmarket
Fig 3e
lower deposits
Sc
5 km
10
15
Interpreted from seismic data
waterwell data
Interpreted from seismic data

Fig. 3.3. Gravel lithofacies in DH-Nob; a) graphic log, b) granule gravel, c) pebble gravel, d) seismic facies at DH-Nob of gravel on eastern margin of tunnel channel. Note high-amplitude signal (H) in downhole seismic-velocity trace indicates ~ top of gravel. Seismic data are from Pugin et al. 1999.

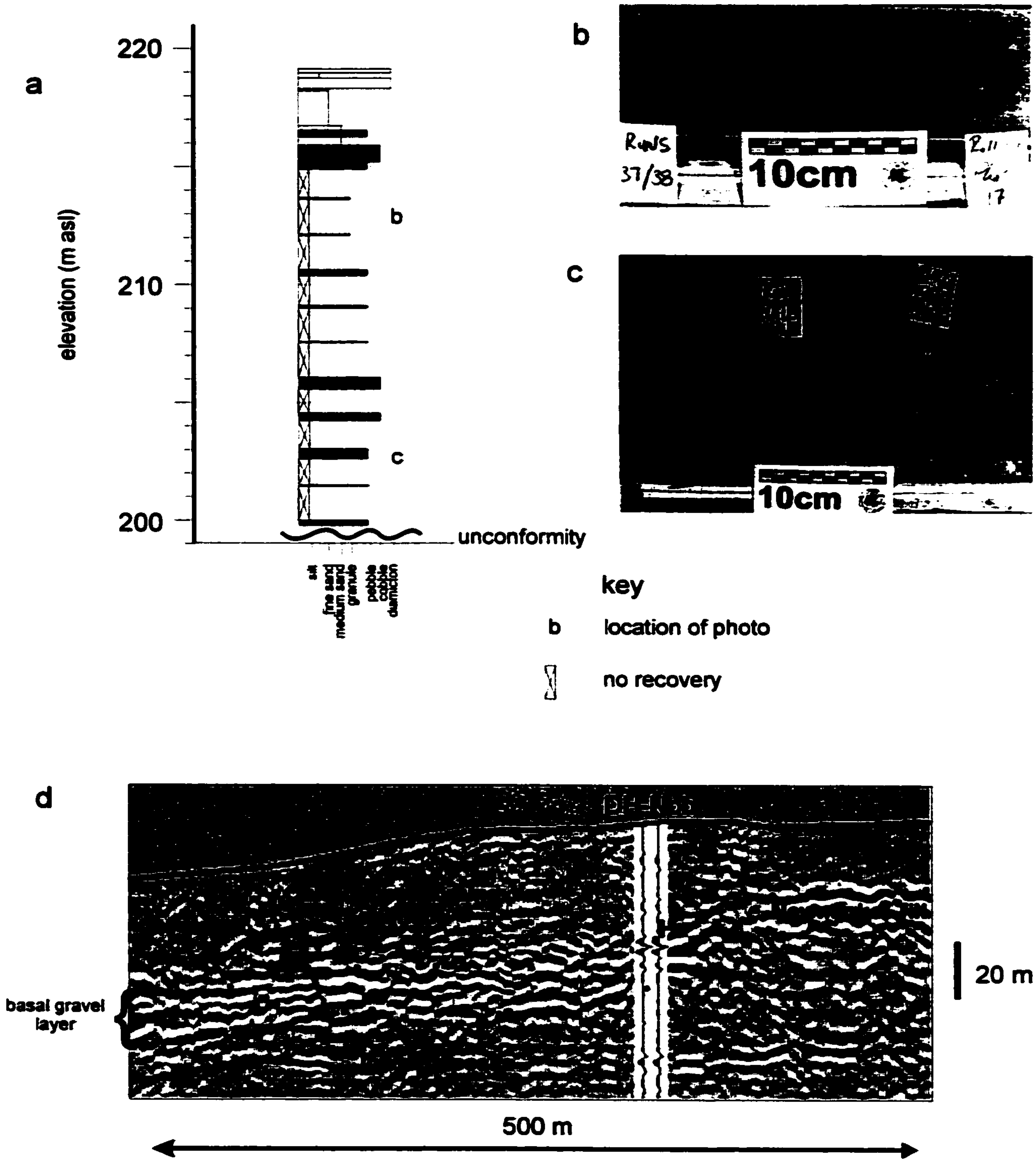

a
220
210
200
elevation (m asl)
b
c
unconformity
silt
fine sand
medium sand
granule
pebble
cobble
diamicton
b
10cm
c
10cm
key
b location of photo
no recovery
d
basal gravel layer
20 m
500 m

Fig. 3.4. Graphic sedimentological log of lower 60 m of core from drillhole DH-V4-158 plotted with total organic carbon and grain-size. Detailed stratigraphic log of drillcore provided for reference. Depth from surface is in parenthesis, elevation is masl.

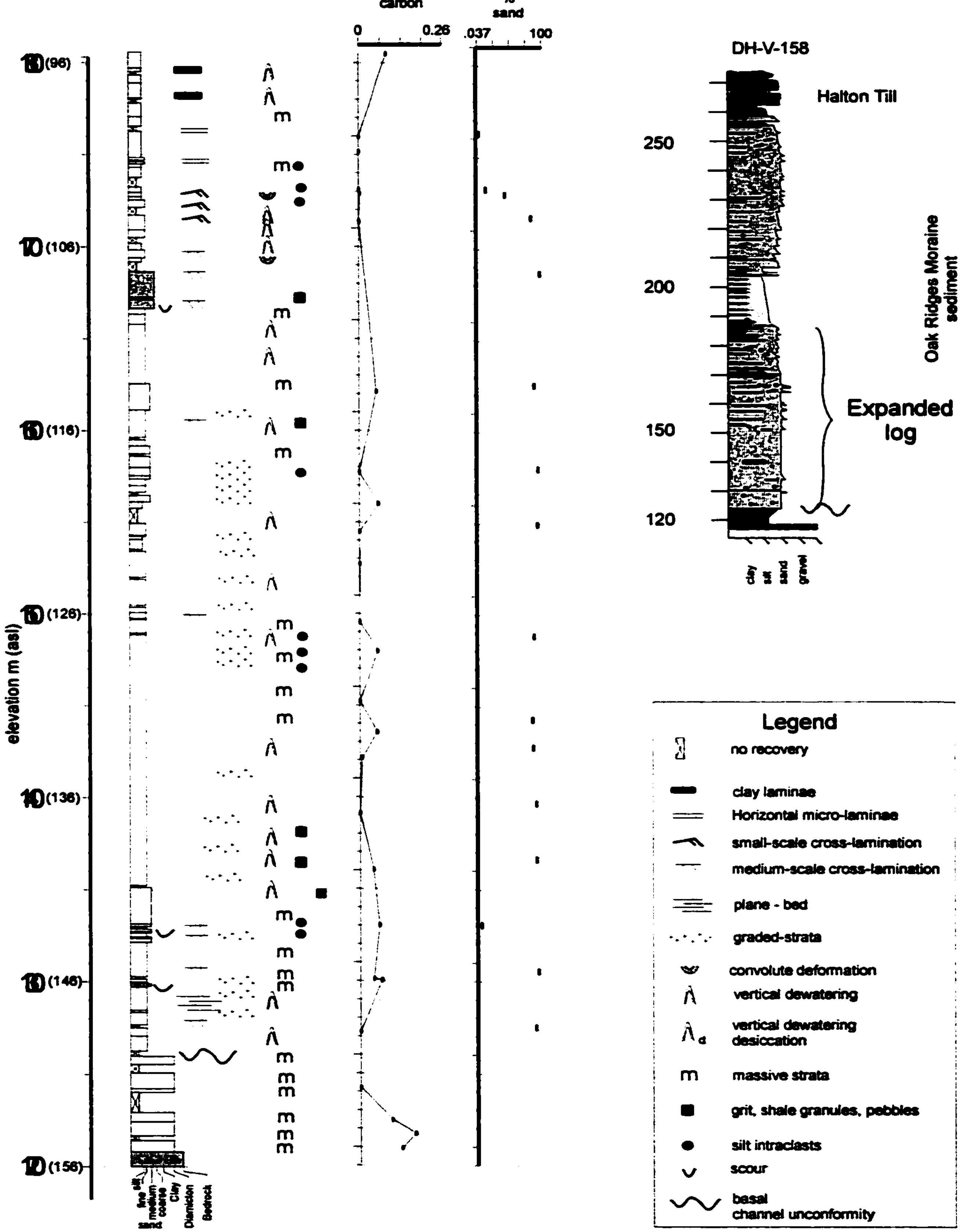

total organic carbon
0
0.26
Grain-size % sand
.037
100
elevation m (asl)
(96)
(106)
(116)
(126)
(136)
(146)
(156)
silt
fine
medium
coarse
sand
Clay
Diamicton
Bedrock
DH-V-158
Halton Till
250
200
150
120
Oak Ridges Moraine sediment
Expanded log
clay
silt
sand
gravel
Legend
no recovery
clay laminae
Horizontal micro-laminae
small-scale cross-lamination
medium-scale cross-lamination
plane - bed
graded-strata
convolute deformation
vertical dewatering
vertical dewatering desiccation
massive strata
grit, shale granules, pebbles
silt intraclasts
scour
basal channel unconformity

Fig. 3.5. Graded-massive sand facies, (a) arrow (1) indicates basal scour overlain by medium sand. Multiple reverse -normal graded fine sand beds with heavy mineral concentrations, arrows (2). (b) Massive fine sand with horizontal desiccation fractures. (c) Massive medium-coarse sand with isolated pebbles and traces of medium scale cross-laminae. (d) Silt intraclasts in massive medium sand. Scale is in cm and core top is to top of page, elevation is masl, depth from surface is in parenthesis.

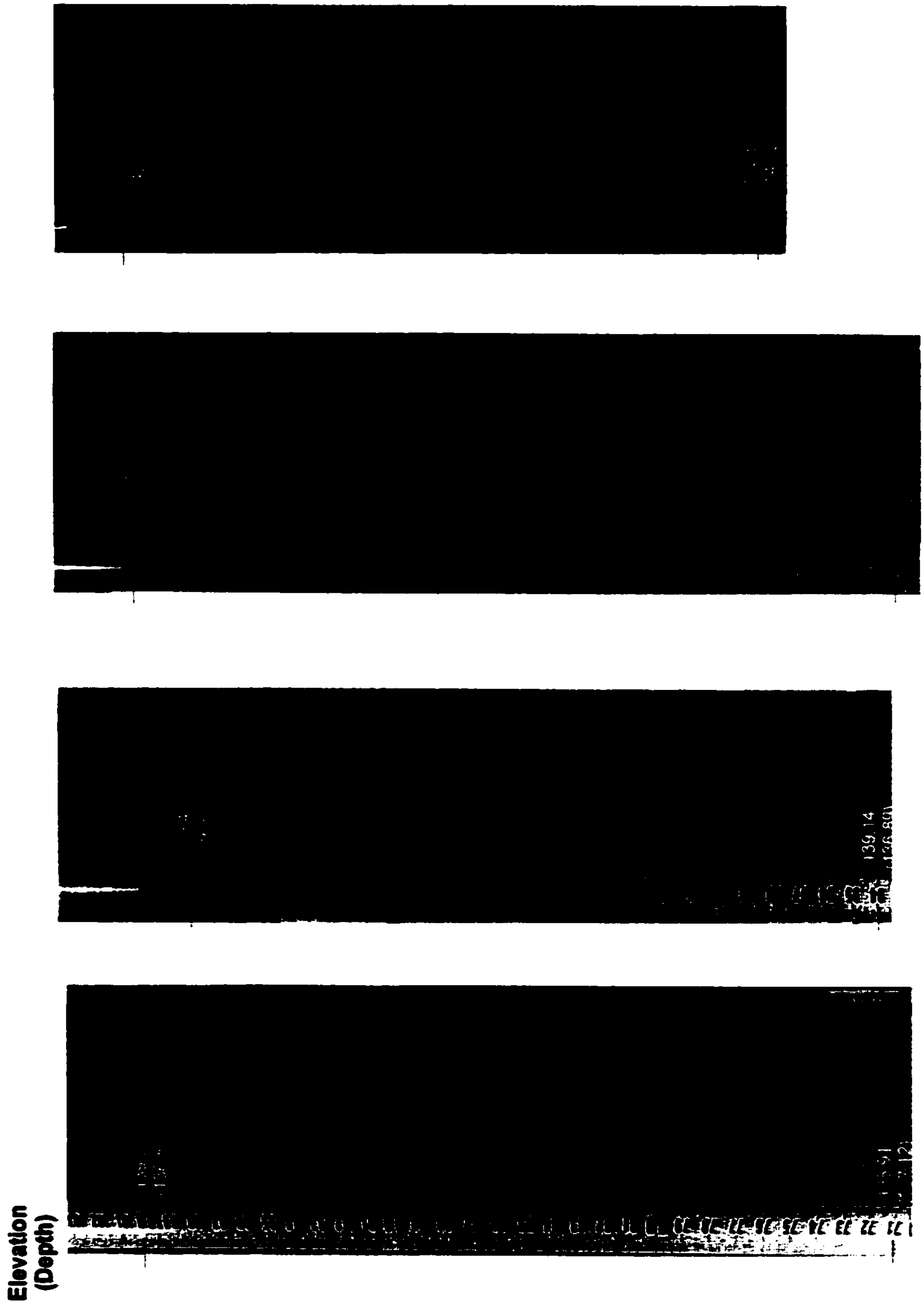

139.14
Elevation
(Depth)

Fig. 3.6. Grain size data. (a) Cross-laminated sand facies. (b) Graded massive sand facies. (c) Graded/massive sand facies with coarse tail. Abbreviations for grain size description are, P - pebble, G - granule, SVC - very coarse sand, SC - coarse sand, SM - medium sand, SF - fine sand, SVF - very fine sand, SiU - coarse silt, SiM - medium silt, SiF - fine silt, SiVF - very fine silt. Clay boundary is at 2 μm.

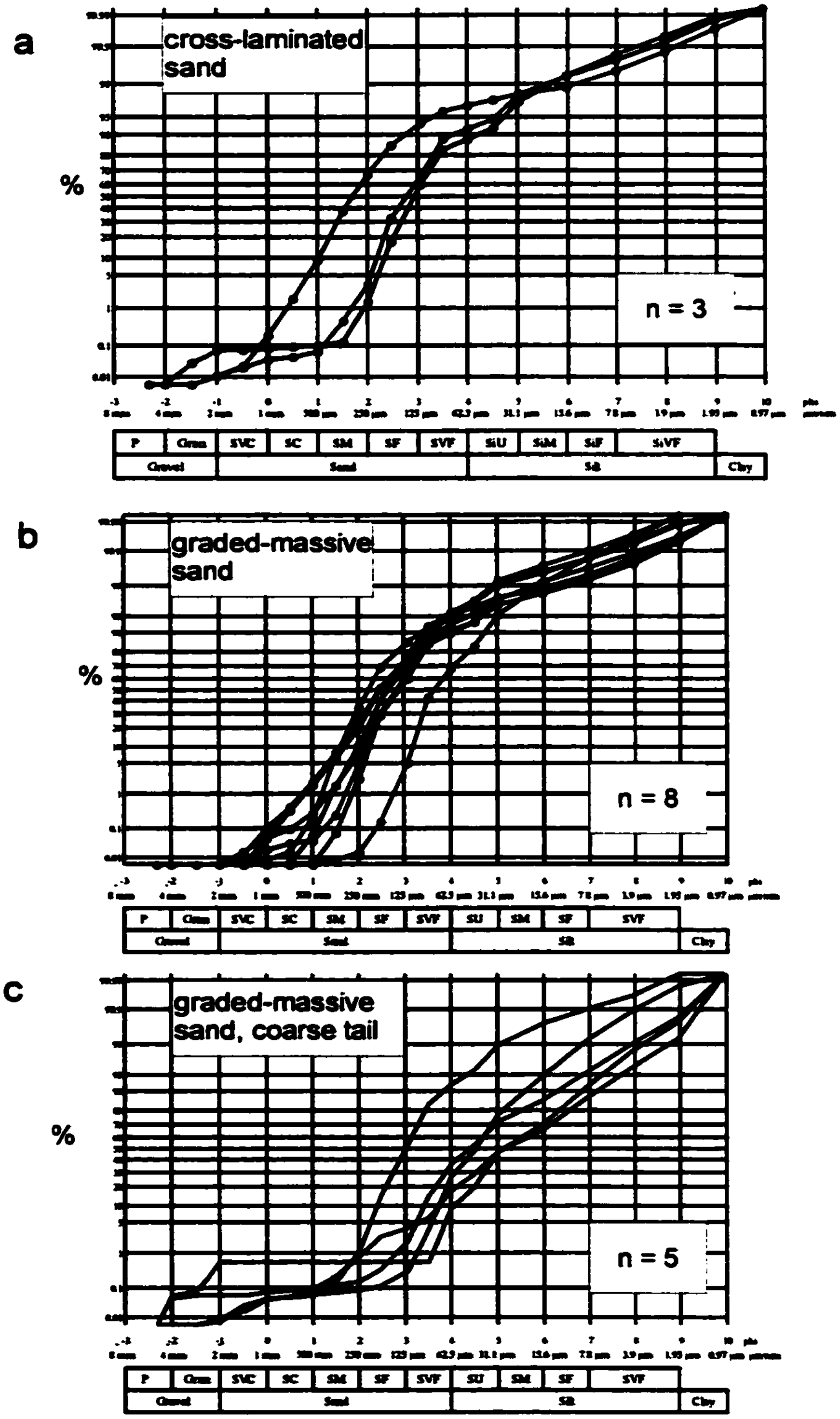

a
cross-laminated sand
%
n = 3
b
graded-massive sand
%
n = 8
c
graded-massive sand, coarse tail
%
n = 5

Fig. 3.7. Deformation features in sand facies. (a) Vertical pipe dewatering structures. (b) Vertical fracture pattern in massive sand. (c) Downward curved laminae in fine sand. Scale is in cm., elevation is masl., depth from surface is in parenthesis.

Elevation
(Depth)

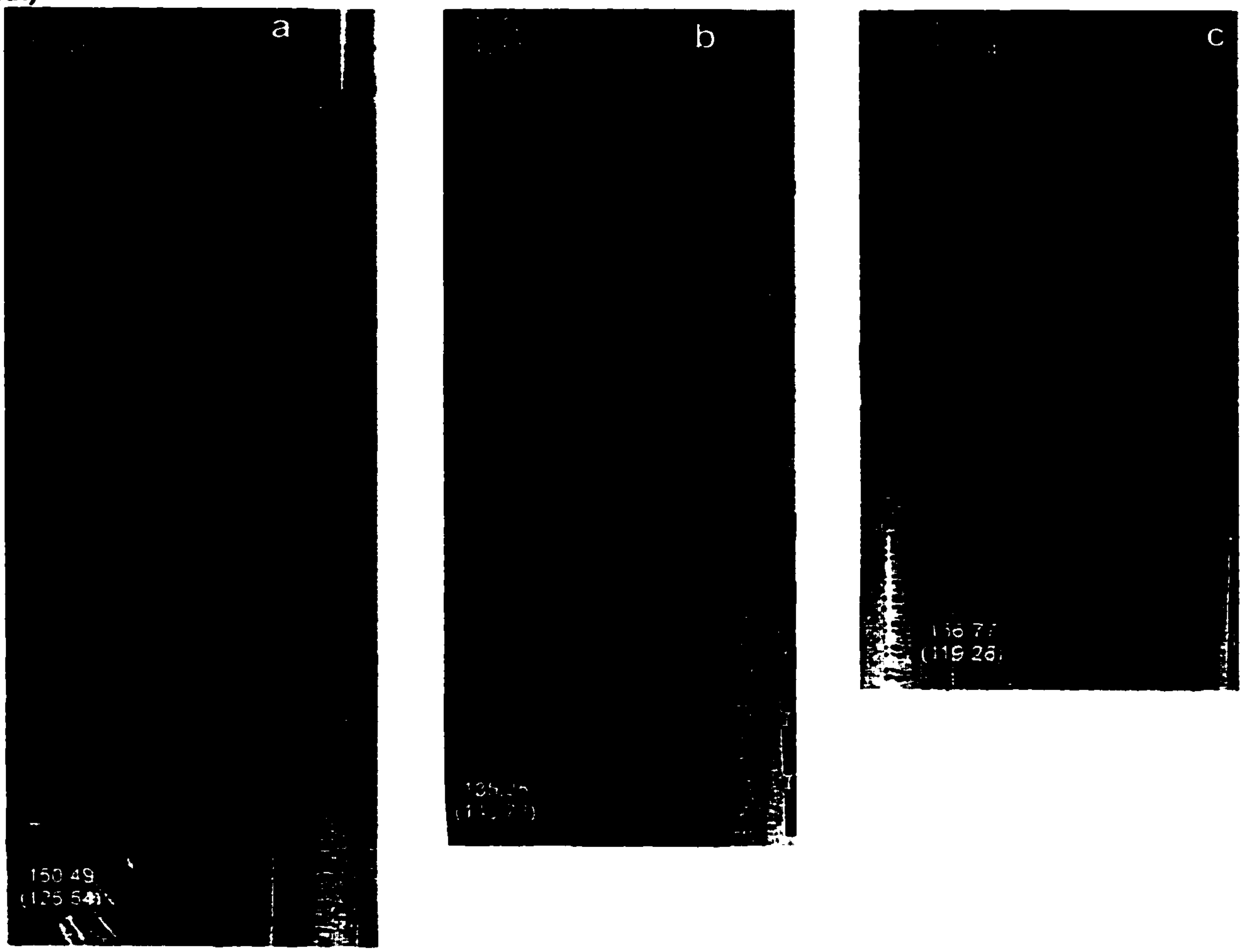


Fig. 3.8. Schematic illustration showing zonation within a traction-carpet (modified from Sohn 1997).

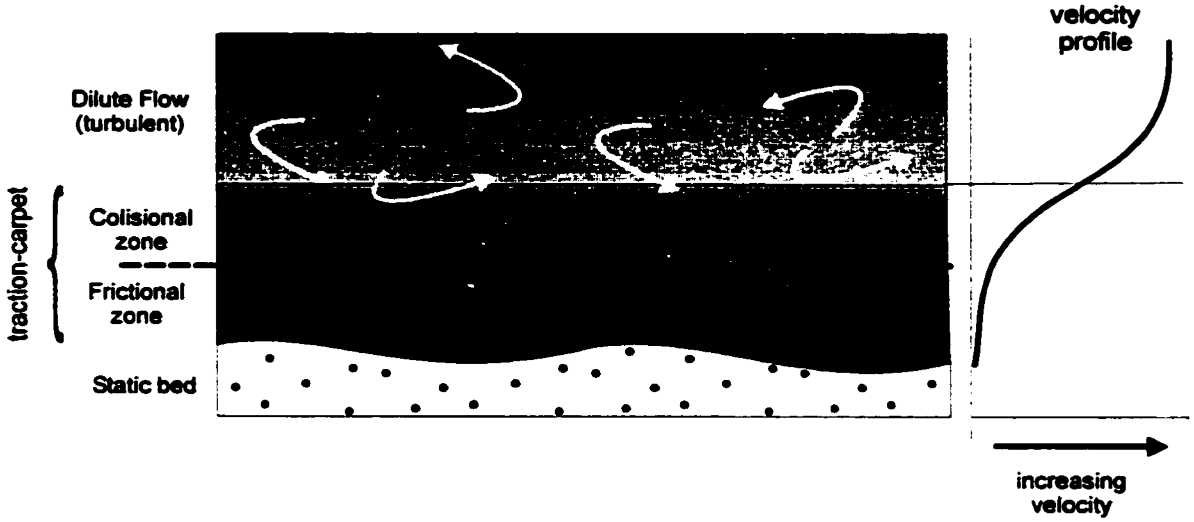
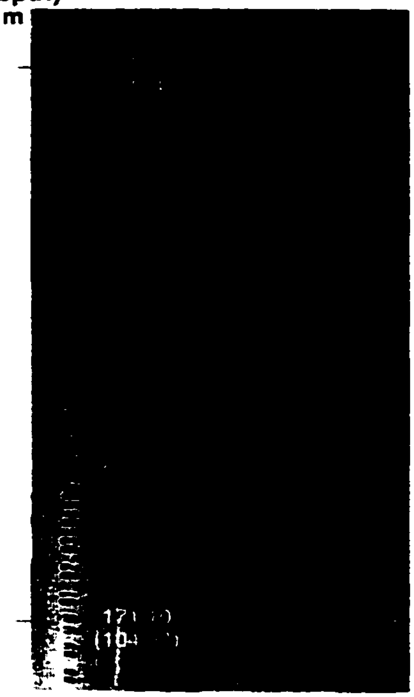

Dilute Flow
(turbulent)
traction-carpet
Colisional
zone
Frictional
zone
Static bed
velocity
profile
increasing
velocity

Fig. 3.9. Cross-laminated sand facies.(a) Small-scale cross-laminated fine sand coset with detrital organics on foresets and minor deformation. (b) Medium-scale cross-laminated set. Scale is in cm, elevation is masl, depth from surface is in parenthesis.

Elevation
(Depth)
m



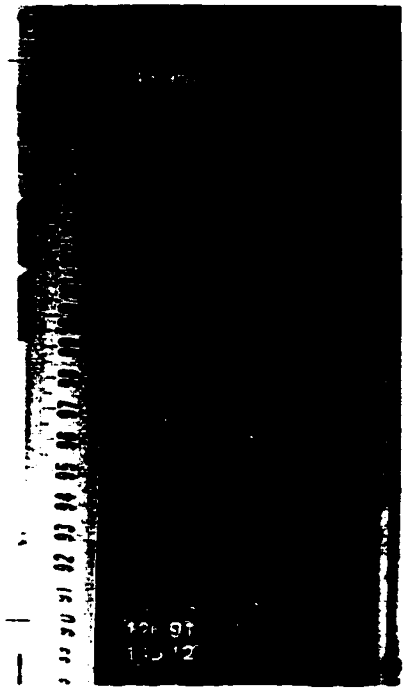


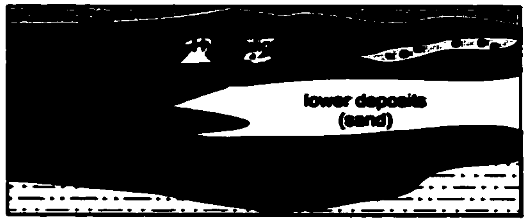
Fig. 3.10. Five stage model showing the sequential erosion and fill of a tunnel channel. Sections drawn perpendicular to flow direction.

a) **Stage I Subglacial channel development**

- high energy turbulent flow
- gravel deposition

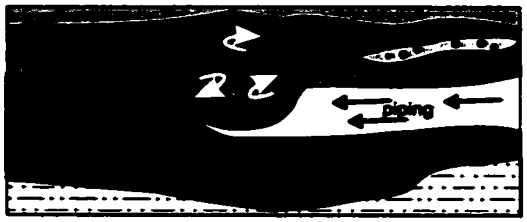
b) **Stage II Erosion of Newmarket Till**

- differential erosion beneath hydraulic jumps
- Unroofing of underlying less consolidated sediment



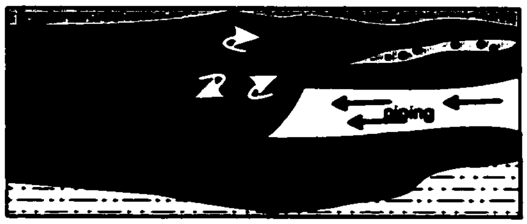
c) **Stage III Channel deepening**

- rapid channel deepening
- groundwater flow and sediment piping



d) **Stage IV Sand facies deposition**

- sediment entrainment
- development of hyperconcentrated flows
- rapid sedimentation from traction carpets



e) **Stage V Basin sedimentation**

- waning of meltwater flow
- transition to low energy sedimentation

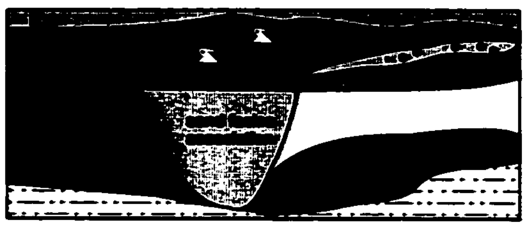


Fig. 3.11. Schematic illustration showing an oscillatory hydraulic jump within a subglacial conduit. Jump develops where flow rapidly expands due to change in conduit diameter (h to h'). Substratum is scoured and sediment entrained into flow by oscillating jet.

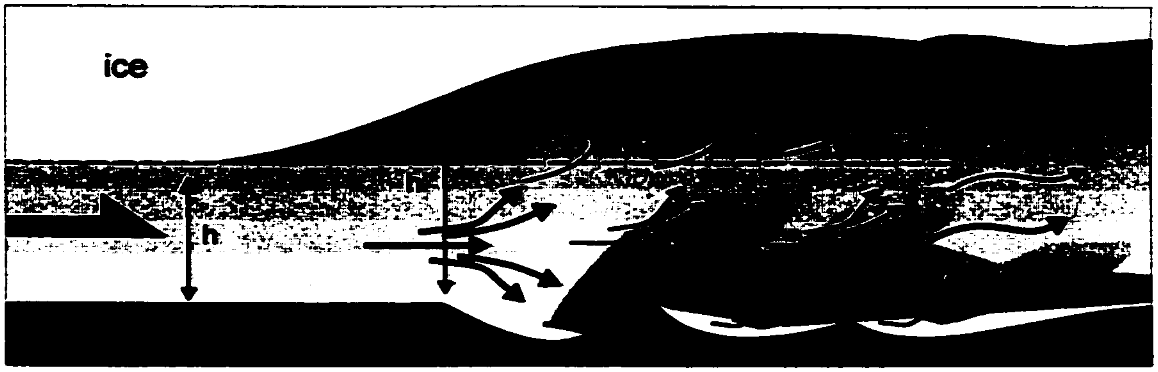

ice
h

Table 3.1 Comparison of seismic facies mapped from deep channel axis fills. Data from, [1]Mullins and Hinchey 1989; [2]Mullins et al. 1997; [3]Boyd 1988;[4]Pugin et al.1999; [5]Wingfield 1990.

Region	Seismic Technique	Frequency /resolution (m)	Seismic		Interpretation	Texture verified / (inferred)
			facies	Thickness m		
Finger Lakes[1,2]	high-resolution seismic (marine)	1000 Hz / 1m	reflection-free , locally diapiric	< 53	rapid suspension sedimentation[1,2] or till tongues[2]; subglacial[1,2] or proglacial[2]	(silt or till)
Scotian Shelf[3]	high-resolution seismic (marine)	1000 Hz/ 2 m	low - continuity reflectors; variable amplitude; chaotic configuration	260	subglacial	unknown
Oak Ridges Moraine[4]	exploration seismic (land)	150-200 Hz /4 m	incoherent, chaotic, angular	100	subglacial slump, suspension deposition from hyperconcentrated flow	(sand, silt)
North Sea[5]	high-resolution seismic (marine)	not stated	chaotic, discontinuous reflectors, hyberbolic reflectors	< 150-200	backfill as plunge-pool incision occurred	(sand and gravel)

Chapter IV

Lithofacies and stratal geometry of a subaqueous fan: Oak Ridges Moraine, southern Ontario

4.1 Introduction

Moraines mark a perturbation in the deglacial history of an area and form when sediment supply to the glacial terminus exceeds both the rate of retreat, and the ability of the proglacial system to redistribute the sediment. Where the ice terminus is bounded by a proglacial water body, sediment is deposited rapidly as the glacifluvial system quickly loses transport capacity. Meltwater influx to the basin occurs at the water surface of the basin, producing a delta and/or at depth forming a subaqueous fan (Rust and Romanelli 1975). Large continental moraines (e.g. Oak Ridges Moraine, Barnett et al. 1998; St-Narcise Moraine, Burbidge and Rust 1988; Salpausslka I, Fyfe 1990; Hartman Moraine, Sharpe and Cowan 1990) typically have a polygenetic origin that reflects both of these depositional processes. In the case of subaqueous fans, however, although they are a key element in many large moraines (e.g. Barnett et al. 1998; Sharpe and Cowan 1990; Warren and Ashley 1994), previously published depositional moraine models that include a subaqueous fan element are commonly highly schematic and include a great deal of implicit information concerning depositional processes and stratal architecture (e.g. Fyfe 1990; Sharpe and Cowan 1990). In part, this is because the sand-rich character of these deposits make them prone to recessive weathering which causes them to be poorly exposed in the field and therefore difficult to study. In contrast to more northerly or isolated moraines, the Oak Ridges Moraine in southern Ontario has extensive aggregate extraction operations that provide an opportunity to conduct detailed sedimentological studies of freshly excavated, unslumped sections. Furthermore, the moraine was deposited in a glacifluvial-glacilacustrine setting (Barnett et al. 1998) and consists, in part, of coalesced subaqueous fans (Barnett et al. 1998; Paterson and Cheel 1997). Consequently, it is composed of sediment deposited in a similar setting to many other large interior continental moraines.

The most completely developed subaqueous fan models are based on studies of Pleistocene glacimarine deposits (e.g. Hunter et al. 1996) or modern analogues from fjord tidewater glacier environments (e.g. Powell 1990). These models are considered poor analogues for glacilacustrine subaqueous fans because the density contrast between the jet efflux fluid and the non-saline ambient basin water is much reduced. The higher density contrast in glacimarine environments results in generally shorter distances of underflow and development of sub-vertical buoyant plumes (e.g. Powell 1990; Syvitski 1989). These differences in turn produce different facies assemblages and fan geometries compared to their glacilacustrine counterparts (e.g. Powell 1990). For example, large barchanoid bars form at the zone of jet detachment in marine fans as sediment rains from the rising plume. In light of the greater focus of research on glacimarine fans, plus the improved understanding of plane-wall jet (e.g. Gorrell and Shaw 1991) and hyperconcentrated-flow deposits (Brennand 1994; Plink-Bjorklund and Ronnert 1999; Powell 1990), it is considered pertinent to first review the current state of knowledge of the lacustrine subaqueous fan model.

4.1.1 A review of the glacilacustrine subaqueous fan model:

A glacilacustrine subaqueous fan model is synthesized here (Fig. 4.1) using the plane-jet efflux process of Bates (1953) and Powell (1990) and the plane-wall jet-efflux process (e.g. Gorrell and Shaw 1991; Rajaratnam 1981). This model which incorporates idealized stratigraphic columns constructed from lithofacies descriptions from the literature (Table 4.1) is intended to serve as a reference for the facies and architectural descriptions presented later. Consequently a number of simplifications have been made, principally: i) the efflux is from a subglacial conduit; ii) the efflux develops into an underflow; iii) diamicton facies are ignored; and iv) the discussion focuses only on the zone of jet efflux. For a continental glacier with a near vertical terminus in a glacilacustrine basin it is assumed that most sediment is supplied from subglacial conduits. Assuming a subglacial conduit position permits the jet efflux to be considered simply as an underflow. Interflow and overflow events are ignored (see, Powell 1990). The Oak Ridges Moraine (ORM) has very few diamicton interbeds, aside from the overlying Halton complex (Russell et al. 1998). As a result, diamicton facies of subaqueous fans are not considered here (see McCabe and O'Cofaigh 1994). Also, this review does not discuss conduit - eskerine processes that have been previously developed by Banerjee and McDonald (1975) and Brennand (1994), or distal, deeper water, suspension - underflow sedimentary processes (see review by Ashley et al. 1985). Additionally, this review does not discuss the morphological or spatial associations of eskers and subaqueous fans (see Warren and Ashley 1994).

The subaqueous fan model presented here is divided conduit, jet efflux, and basin components (Fig. 4.1a). Within each environment, facies and facies-architecture associations are controlled by unique hydrodynamic and ice sheet processes, including: ice front stability, conduit stability, conduit diameter, sediment flux, and flow velocity (Powell 1991).

Conduit flow

To deliver sediment beyond the ice-margin (grounding line) ice sheets terminating in a glacilacustrine basin must maintain pipe-full discharge in subglacial conduits. If the hydraulic head within the glacier falls below the glacilacustrine surface, basin water migrates into the conduit and sediment delivery to the basin is inhibited. Sediment deposited within the conduit consists of imbricated boulder gravel to small-scale cross-laminated sand (e.g. Banerjee and McDonald 1975; Brennand and Shaw 1994). These esker deposits differ from subaqueous fan deposits by being morphologically more linear and steeper sided; they contain more extensive ice let-down structures, have a stronger clustering of paleoflow directions, and have fewer indications of deposition related to rapid flow expansion.

Jet efflux

Sedimentation from the jet efflux into the basin is the principal constructional fan process (Fig. 4.1). If the conduit is subglacial and a subcritical jet emerges along the basin floor, then it is modelled best as a plane jet (Fig. 4.1a, Powell 1990). If the jet is supercritical then it is referred to as a plane-wall jet (Fig. 4.1b, e.g. Rajaratnam 1981). Three stages of decay occur as the jet progressively entrains ambient basin fluid and loses momentum. For a plane jet these are: i) zone of flow establishment (ZFE); ii) zone of flow transition (ZFT); and iii) zone of established flow (ZEF) (Bates 1953). The downflow length of each zone relates closely to conduit diameter, discharge velocity and fluid density. In general the plane jet is parabolic in plan, and its downflow extent is 2000 times the conduit diameter (2000D; for example 20 km for a 10 m wide conduit; Fig. 4.1c). Laterally the plane jet expands to a width approximately three times the square root of the distance downstream of the conduit mouth (Bates 1953). In addition, the velocity profile across the jet efflux has a Gaussian form (Albertson et al. 1948); and therefore velocity rapidly diminishes laterally outward from the axial core. These ideal dimensions are reduced when the effect of bottom friction, surface roughness and suspended sediment are considered (Powell 1990). Where friction along the base and/or upper surface of the flow is high the ideal form is dramatically modified and a broader and more triangular shaped form develops (Fig. 4.1d, Wright 1977). In the original treatment of jet efflux by Bates (1953) and in the review by Powell (1990) only the plane jet was considered. It is assumed here that for a plane-wall jet the rapid transition from supercritical to subcritical flow and a hydraulic jump development will correspond with the zone of flow transition (ZFT). For simplicity basinward flow distances in this review are all measured with respect to an inertia dominated plane jet.

Zone of flow establishment (ZFE): The ZFE is the region that extends downflow from the conduit mouth approximately 4 times the conduit diameter (4D, Powell 1990). The width of this zone is comparable to the conduit diameter at the mouth because minimal mixing occurs with basin water, and as a result the velocity of the jet is similar to that in the conduit and narrows to the jet axis in a streamwise direction. Consequently, the lithological character of the deposits is also similar. The ZFE contains the coarsest sediment in the fan and consists of open-framework imbricate boulders (c.f. Powell 1990; Rust and Romanelli 1975), heterogeneous massive to cross-bedded gravel, and plane-bed and cross-bedded sand (c.f. Gorrell and Shaw 1991). Stratal geometries range from tabular gravel sheets (Powell 1990) to channel fills composed of a range of meso- and macro-forms (dunes, low-relief gravel bars). If the emerging flow is supercritical, then in-phase waves may develop, resulting in the deposition of antidune cross-stratified sand or gravel (Brennand 1994). If the conduit is stationary, backset beds develop in response to fan aggradation (e.g. Johansson 1976).

Zone of flow transition (ZFT): Depending upon whether conduit flow is subcritical or supercritical (Froude number, $1.0 < Fr < 1.0$) the zone of transition (ZFT) is modelled, respectively, as a plane jet (Bates 1953) or plane-wall jet (Fig. 4.1a,b) (Gorrell and Shaw 1991; Rajaratnam and Subramanyan 1986). For a plane jet the ZFT extends downflow from the conduit mouth for a distance of 6-8D. This zone is characterized by flow deceleration as ambient basin fluid is entrained progressively inward toward the collapsing jet efflux boundary (Fig. 4.1c). The downflow end of this zone is marked by the development of a standard basinward rate of decay of the axial jet flow (Bates 1953). Deposits within this zone consist of coarse cobble-boulder gravel, and trough and planar cross-bedded sand (Fig. 4.1c). The reduced flow velocity and increased flow depth result in meso- and/or macro-form development. Dipping foresets have been interpreted to be large transverse macroforms (e.g. Banerjee and McDonald 1975; Warren and Ashley 1994) or mesoform accretion surfaces (Johansson 1976).

Development of a plane-wall jet requires supercritical flow through the ZFE (Fig. 4.1b). Similar flow conditions are not uncommon during flood conditions in fluvial systems (e.g. Pierson and Scott 1985) and are probably more common in subglacial conduits where flows are fully confined compared to open channels (e.g. Simons et al. 1965). Using the experimental results of Rajaratnam and Subramanyan (1986), Gorrell and Shaw (1991) suggested that a hydraulic jump is characterized by intense bed erosion, and flow thickening followed by rapid downflow sedimentation as the flow rapidly losses transport competence (Fig. 4.1b). This process has been discussed in detail by Komar (1971) for turbidites at the transition from submarine canyon to fan. As a consequence, a thick density-stratified dispersion is formed with a highly concentrated basal layer and a much thicker, more dilute fully-turbulent overflow. Rapid sedimentation occurs within the steep-sided scours or as the dispersions evolve downflow of the jump into significantly more dilute turbidity currents. Deposits consist of massive or diffuse-graded sand and gravel with large intraclasts eroded from the scour margins (Gorrell and Shaw 1991). Alternatively, other authors have suggested that similar steep-sided scours and sand fills were the result of slumping and evolution to hyperconcentrated grainflows (e.g. Cheel and Rust 1982; Postma et al. 1983).

Zone of established flow (ZEF): The ZEF extends downflow of the conduit mouth to approximately 2000D, the distance at which the flow velocity has decayed to a negligible value. In this zone turbulent mixing takes place throughout the entire jet (Bates 1953). Idèally, the jet expands downflow as a symmetrical parabola, although local fan topography may deflect it. Deposits close to the flow axis consist of planar and trough cross-bedded sand (Powell 1990) Produced by migrating 2-D and 3-D dunes. Laterally and downflow these cross-beds grade to climbing small-scale cross-laminated sand, commonly referred to as ripple-drift cross-stratification (c.f. Gorrell and Shaw 1991; Jopling and Walker 1968). Within this zone subaqueous braided systems have been interpreted to consist of channel levee deposits of small-scale, cross-laminated sand and channel fills of cross-bedded medium sand (e.g. Burbidge and Rust 1988).

Basin

At the basinal terminus of the inertial jet efflux, suspended sediment underflows may develop and flow farther basinward. These flows are not controlled by the inertial flow of the jet efflux but rather by local differences in fluid density and basin slope. Deposits are dominated by cosets of small-scale, cross-laminated fine sand (Burbidge and Rust 1988) that occur in association with graded silt strata and clay laminae forming Bouma T_{cde} divisions (e.g. Gilbert 1997; Weirich 1986). Along the jet efflux boundary reverse circulation eddies form as the ambient basin water is entrained into the jet and produces small-scale cross-laminae with paleoflow directions oriented opposite to the jet efflux flow direction (e.g. Gorrell and Shaw 1991).

In summary, this glacilacustrine fan model subdivides the jet efflux into three main units based on the basinward loss of velocity and sediment transport competence. The following discussion of a small glacilacustrine subaqueous fan that forms part of a large moraine complex describes the constituent facies and facies-associations and stratal geometry in the context of the jet efflux model (Fig. 4.1). In addition, evidence of deposition from high-energy flows, including supercritical flow, hydraulic jumps and hyperconcentrated dispersions, suggests that the fan was deposited within a single meltwater season from a *single* outburst flood or jökulhlaups. The possible source of the meltwater is also discussed.

4.1.2 Study area and geological setting

The study area is in an aggregate pit approximately 60 km northwest of Toronto, Ontario (Fig. 4.2). It is located on the north side of the western Oak Ridges Moraine (ORM), and is 15 km east of the Niagara Escarpment. The Oak Ridges Moraine is a major physiographic feature that extends for approximately 160 km from the Niagara Escarpment eastward to the vicinity of Trenton (Chapman and Putnam 1984) and is 2-20 km wide and up to 160 m thick (Barnett et al. 1998). In the western part of the ORM which includes the study area, sediment consists predominantly of gravel, sand and silt with less than 2 %

interbedded diamicton (Russell et al. 1998).

The ORM is interpreted to have formed in an ice-supported glacilacustrine basin (water surface ~420 m asl, depth up to 200 m) that was bounded to the west by the Niagara Escarpment (Chapman and Putnam 1984). This interlobate basin was <40 km wide at the escarpment and narrowed eastward. The moraine has been interpreted to consist of glacifluvial and glacilacustrine sediment deposited in four stages: i) subglacial, ii) subaqueous fan, iii) subaqueous fan/delta, and iv) ice-marginal (for details see Barnett et al. 1998). These four stages are interpreted to record extreme changes in the magnitude of meltwater discharge from regional subglacial reservoir outbreak floods (Barnett et al. 1998; Brennand and Shaw 1994) to seasonal, climate-modulated precipitation events (Gilbert 1997). Only the third stage, represented by the upper 20-60 m of the moraine, was deposited within a glacilacustrine setting dominated by subaqueous fan sedimentation (Barnett et al. 1998). This study discusses sediment deposited during that stage.

This study was conducted in an active aggregate pit from 1993-1998. Sections were 6-10 m high and extended laterally for approximately 250-300 m subparallel to the mean paleoflow direction (east-west). In addition, a number of shorter sections oblique to transverse to flow were measured (Fig. 4.2). Field observations of texture, bedding contacts, sediment structures, etc., have been grouped into facies (e.g. Miall 1996). Study of the deposit architecture is based on photo-mosaics (e.g. Miall 1996).

4.2 Sedimentology

The 6-10 m high and ~300 m of east-west oriented sections were predominantly sand and gravel overlain by 1-3 m of laminated-massive silt. The sand and gravel had a strong east-west sorting with coarser deposits exposed in the eastern sections, whereas more westerly sections were progressively dominated by finer sand. The sections were generally characterized by upward-fining trends. Very little diamicton was observed at the study site. From these exposed sections eight principal facies have been identified (Table 4.2).

4.2.1 Cross-stratified gravel (Gt, Gp)

The cross-stratified gravel facies forms 1-3 m thick tabular deposits (Fig. 4.3, 4.4, 4.5) that overlie a sharp, planar to scoured lower contact that truncates quasi-planar stratified or cross-stratified sand. The facies consists of normal-graded medium-scale cross-bedded, bimodal pebble to heterogeneous sandy pebble gravel (Fig. 4.6). Trough cross-bedded heterogeneous gravel is more poorly sorted and is generally thicker, forming sets 0.5-1 m thick (Fig. 4.6). Maximum pebble size is ~10 cm and averages ~1-3 cm (Fig. 4.6a). The largest clasts are rare intraclasts of silty sand that are up to 40 cm long and 5 cm thick and generally have a sub-horizontal fabric. Planar cross-bedded bimodal gravel form sets 4-10 cm thick and cosets 25-30 cm thick. Cross-beds dip at 20-25°, are well sorted, and are normally graded (Fig. 4.6b). Single sets are typically mantled by a fine silty sand.

Interpretation: This facies was deposited by migrating two- and three- dimensional subaqueous dunes (e.g. Harms et al. 1975) or bars (e.g. Boothroyd and Ashley 1975). Normally graded cross-stratification reflects gravity sorting during slipface avalanching.

4.2.2 Planar stratified gravel (Gh)

The planar stratified gravel facies occurs as laterally extensive 0.5-2 m thick tabular deposits overlying, and less commonly interbedded with, the cross-stratified gravel facies. It is most extensive in section M-2a where it was traced for 8-10 m parallel to flow (Fig. 4.3). Upflow it grades into the heterogeneous gravel facies (see below) and downflow is interbedded or grades into quasi-planar stratified or cross-bedded sand. Basal bed contacts are planar to undulatory with shallow scours < 10 cm deep. Scours are generally infilled by coarse bimodal gravel. The facies varies from open framework unimodal pebble beds to sandy pebble beds commonly forming upward fining couplets (Fig. 4.7). Open framework gravel beds are generally 1-3 cm thick (Fig. 4.7). Sandy pebble beds are 2-10 cm thick, more poorly sorted, and more laterally continuous. The maximum observed clast size is ~8 cm. Clast clusters occur locally and commonly have a cobble-size clast at the core with pebble and sand stoss and lee deposits. The stoss side deposits are characterized by imbricate pebbles with a medium sand matrix. In contrast, the lee side deposits tend to be finer grained and a more disorganized fabric. Clast fabric is predominantly a(t) b(i) to flow and indicates paleoflow toward the northwest.

Interpretation: The upward fining couplets, open framework beds, clast clusters, transverse clast fabric, and an absence of cross-bedding are all considered to be evidence for deposition from a turbulent high-energy flow (Collinson and Thompson 1989). These characteristics have been attributed to antidune flow conditions under both supercritical (Blair and McPherson 1994; Iseya and Ikeda 1987; McDonald and Day 1978) and critical flow conditions (Koster 1978). Gravel couplets with open-framework have been interpreted to be antidune stratification from alluvial fans formed under high-energy flow conditions of $\sim 1.5 < Fr < 1.7$ (Blair 1999). To account for the absence of cross-strata, Blair (1999) Invoked an autocyclic mechanism involving over steepening of the surface wave, rapid breaking of the curling wave crest followed by rapid downslope shooting (washout) of water. During washout, large volumes of sediment are momentarily carried into suspension and the antidune structure is partially or completely eroded. Subsequent rapid sedimentation forms the finer-grained upper portion of each gravel couplet. Clast clusters are possibly transverse ribs that form under upstream migrating hydraulic jumps (McDonald and Day 1978), or by antidunes (Koster 1978). The transverse ribs are marked by the largest clasts and consequently are interpreted to have formed beneath breaking standing waves when finer sediment was transported downflow. The succession of gravel couplets is interpreted to have been deposited by a single flow event as shown by studies of modern alluvial fan deposits that are 5 m thick, and consist of

15 gravel couplets that were deposited from a single flash flood that lasted for only 5 hours (Blair 1987).

4.2.3 Heterogeneous gravel with Intraclasts (Gd)

The heterogeneous gravel facies consists of matrix to clast supported heterogeneous gravel (Fig. 4.3, 4.4) and diffuse-graded sandy gravel. Bedding is generally poorly developed with gradational contacts and poor lateral continuity of units (Fig. 4.8). Common to these deposits are silty sand intraclasts up to 3 m long (Fig. 4.9a,b). The facies is up to 3 m thick, and sharply overlies the sand and cross-stratified gravel facies. The facies generally grades downflow to plane-bedded and cross-bedded sandy gravel and coarse sand (Fig. 4.4). The heterogeneous gravel is generally poorly bedded or massive, with minor interbeds of planar-stratified gravel facies, particularly toward the top (Fig. 4.8a,b). The dominant clast size in these strata is pebble with minor cobbles. Diffuse-graded sandy gravel was observed at one location downflow of section M-3. The lower contact was slump covered, as was the lateral extent of the strata. The bed fined-upward to coarse sand and the matrix was reverse and normal graded. Local grain-size clusters were always matrix supported (Fig. 4.9b). Boulder-sized intraclasts of fine and medium sand generally had a subhorizontal fabric and were stratigraphically low in the outcropping strata. The larger intraclasts are mostly enveloped in a sand rich matrix that coarsens away from the intraclast (Fig. 4.9b). These finer-grained envelopes are displaced toward the west along the upper margin of the core intraclast. Upward in the facies the size and abundance of intraclasts generally decreases. Clast fabric of these small intraclasts is random near larger intraclasts but with distance becomes more organized and planar.

Interpretation: On the basis of intraclasts, and massive or poorly defined bedding, the heterogeneous gravel and diffuse-graded sandy gravel are interpreted to be traction carpet deposits. Traction carpets are interpreted to form at the base of stratified flows consisting of a subordinate basal hyperconcentrated frictional layer and an overriding dilute fully turbulent collisional layer (Sohn 1997). The thickness of each zone relates to variations in applied shear stress, grain size and downward sediment flux (Sohn 1997). Carpets with a thick collisional zone form best in coarser-grained flows and deposit better sorted more distinctly graded strata. In contrast, a thicker frictional zone produces more massive deposits. Sediment is interpreted to be supported by a number of mechanisms including shear supplied by the overlying flow, dispersive pressure, buoyancy, hindered settling, and turbulence (Costa 1988; Smith and Lowe 1991). In heterogeneous gravel with abundant gravel clasts, dispersive pressure and hindered settling were the primary support mechanisms. In contrast, in the sand-rich gravel, dispersive pressure was less significant because of the smaller mass of individual grains (Lowe 1976) and hindered settling was likely a more important support mechanism.

Traction carpet deposits aggrade grain by grain (Sohn 1997) and, depending on flow dynamics, matrix supported or framework supported gravel were deposited. Massive or normally graded, poorly bedded, matrix supported gravel was previously interpreted to be deposits of hyperconcentrated flows (Lowe 1982; Smith 1986). Less commonly clast supported gravel was interpreted to have been deposited by hyperconcentrated flows (Dorsey and Falk 1998; Russell and Knudsen 1999). Variations in clast support is most likely related to differences in traction carpet dynamics related to changes in sediment flux, suspended sediment concentration, support mechanisms, etc. A number of characteristics differentiate the sandy gravel deposits from the heterogeneous gravel. In the sandy gravel deposits, intraclasts are enveloped by finer sediment, there is a broader range of intraclast sizes, matrix grading is more distinct and pebble intraclast fabric is more disorganized (Fig. 4.9b). Deposition from non turbulent, high concentration flows is indicated by the preservation of poorly consolidated intraclasts and grading in intraclast matrix haloes. Asymmetric relationship between the core intraclast and its of the fine matrix envelope suggests that the flows were not emplaced as a rigid plug. The displacement of the envelopes and vertical grading are interpreted to indicate deposition from a traction carpet. Increased frictional strength was possibly associated with disaggregation of silt intraclasts and a commensurate increase in local sediment concentration (e.g. Postma et al. 1988). Similar facies have been described from other subaqueous fan deposits as gravity flow deposits (Rust 1988) and to be deposits of hyperconcentrated flows (Brennand and Shaw 1996).

Erosion of large unconsolidated friable sand intraclasts was probably related to hydraulic jump scouring or scouring at the base of the flow. The rapid dissipation of flow energy and generation of high volumes of entrained sediment some distance downflow of a hydraulic jump would have developed stratified flows and caused rapid sedimentation. Sand intraclasts are commonly observed in esker (e.g. Brennand 1994), subaqueous fan (e.g. Gorrell and Shaw 1991) and jökulhlaup outwash deposits (Russell and Knudsen 1999) and have been attributed to scour within a hydraulic jump. Less commonly such intraclasts are occur fluvial settings where they may originate by slumping along channel margins (e.g. Martin and Turner 1998) or erosion from the channel bottom (Wan and Wang 1994).

4.2.4 Cross-stratified sand (St, Sp)

The cross-stratified sand facies crops out extensively in section M-3 and M-4 (Fig. 4.5, 4.10). It is sharp-based, predominantly medium-scale, planar-tabular and trough cross-stratified medium to pebbly coarse sand. Planar cross-sets are < 50 cm thick and form cosets 2-3 m thick (Fig. 4.10, 4.11a). The thickest sets occur in the lower part of cosets and thin upward. In section M-4 strata dip at a low angle (< 10°) to the west (Fig. 4.10). Trough cross-sets are generally coarser than planar cross-sets, are coarse-tail graded, and form cosets 30-50 cm thick. At one location, 10-20 cm thick climbing cross-sets form a coset 1-1.5 m thick (Fig. 4.11b). Deformation is rare, but where present consists predominantly of small-scale normal faults with 2-5 cm offset, and minor convolute bedding. Paleoflow measurements of predominantly trough cross-beds from section M-1 are to the southwest, whereas planar cross-beds in section M-4 indicate flow varying from the southwest to northwest (Fig. 4.2). The combined measurements from section M-1 and M-2 have a bimodal distribution with a primary southeast mode and a secondary northwest mode.

Interpretation: This facies was deposited by traction sediment transport and migrating dunes bedforms (Church and Gilbert 1975; McDonald and Vincent 1972). Tabular cross-stratification is the product of 2-D dunes that had limited erosion at the separation-eddy reattachment point (Saunderson and Lockett 1983). Trough cross-stratification was formed by 3-D dunes with intense downflow scouring (Harms et al. 1975). Climbing, medium-scale cross-stratification indicates high rates of suspended sedimentation under waning flow conditions (e.g. Ashley et al. 1982; Gorrell and Shaw 1991). It is useful to note that climbing medium-scale cross-stratification is not common in the geological record, most probably because of the high bed load transport rates in combination with downflow scouring associated with these dunes.

4.2.5 Quasi-planar stratified sand (Sh)

Cropping out extensively in section M-3 (Fig. 4.5) the quasi-planar-stratified (low-angle parallel to divergent strata) sand facies is up to 3.5 m thick (Fig. 4.12, 4.13) and with the exception of the silt-clay facies, overlies any facies. It is interbedded with diffuse-graded or cross-bedded sand, and is transitional upflow with heterogeneous gravel. It consists predominantly of well-sorted medium sand with minor coarse sand (Fig. 4.12) and minor moderately sorted pebbly sand. Strata are 0.1-1 cm thick, graded or massive, form cosets <40 cm thick, but locally up to 1.5 m thick, and overlie parallel or low-angle (< 15°) erosion surfaces (Fig. 4.12b,c, 4.13). Low-angle dipping strata onlap underlying erosional surfaces toward the east. Within a coset, laminae dip at a low-angle and are parallel or diverge slightly. Interbeds of massive sand 2-3 cm thick occur locally. Laterally strata are continuous for up to 8-10 m and pinch and swell along strike. Coarse sand occurs as discontinuous 30-40 cm long interlaminae within medium sand or fill shallow scours that are < 2 cm deep and 5 cm long. Larger scours are 20-30 cm deep, 1-2 m long and filled with massive, medium sand. Locally, low-angle downflow and upflow dipping laminae are preserved (Fig. 4.12b).

Interpretation: Based on stratal geometry, onlap and dip direction, normal grading and changes in bed contact conformity, two depositional processes are interpreted. Where strata are planar, normally graded, and conformable to the underlying bedding surface, deposition under upper-flow-regime, plane-bed conditions is interpreted (Cheel 1990). Conversely, where strata are bounded by low-angle erosion surfaces, and laminae dip at shallow angles into the inferred paleoflow direction, strata are interpreted to be deposited under antidune flow conditions. Both the plane-bed and low-angle strata were deposited from high-energy fluidal flows. Only the low-angle strata are discussed further.

Low-angle cross-stratification dipping <10° up-paleoflow and the paucity of high-angle cross-stratification in the section suggest that deposition occurred under supercritical flow conditions ($Fr \sim 1$) and migrating in-phase wave conditions (Cheel 1990; Middleton 1965). In-phase waves probably developed within an undular ($1 < Fr < 1.7$) or less likely a weak ($1.7 < Fr < 2.5$) hydraulic jump when a supercritical flow underwent rapid flow expansion (e.g. Chow 1959). Unsteady flow conditions may have fluctuated between the undular and weak hydraulic jump and as a result changed the local position of the hydraulic jump and, consequently, the location of erosion and deposition. Massive beds within the succession were possibly deposited under conditions of high sediment concentrations and rapid sedimentation. Breaking in-phase waves in fluvial environments have been reported to produce near-bed suspended sand concentration of up to 80% (Simons et al. 1965). Such elevated sediment concentrations may have momentarily suppressed local fluid turbulence and deposited massive beds.

4.2.6 Diffuse-graded/massive sand (Sd)

Diffuse-graded sand is exposed in section M-2a and M-3 (Fig. 4.3, 4.5) and abruptly overlies and is interbedded with quasi-planar-stratified sand. Locally it overlies small-scale cross-laminated sand. It fills irregular scours that are up to 3 m deep, have margins dipping up to 60°, and locally overhang. Strata occur also in sharp-based tabular beds 5-40 cm thick. Sediment ranges from medium sand to granule coarse sand (Fig. 4.14a). Granules and isolated clay and silt intraclasts occur only in scour fills (Fig. 4.14a). Strata are massive or diffuse-graded where sand is faintly planar stratified. Within the scour-fill beds are < 10 cm thick and average ~5 cm thick. Stratification is most clearly defined toward the scour-fill margins and upward in the fill (Fig. 4.14). Locally subhorizontal undulatory stratification is traceable for 10-100s cm until it becomes very faint or even massive. Within thicker fills multiple erosional surfaces are curvelinear and have apparent dips of up to 70°. The facies also occurs as sharp-based tabular strata < 40 cm thick that grade downflow into, or are overlain by, quasi-planar-stratified sand.

Interpretation: The infill of steep-walled scours, with local overhanging margins, and the diffusely graded or massive texture are interpreted to represent rapid deposition from sediment-charged flows. Although both scour-fill and tabular facies of diffuse-graded sand are interpreted to indicate rapid deposition, there are important differences. Scour-fill units were deposited immediately downflow of a hydraulic jump. Conversely, tabular units were deposited under high-energy, possibly upper-plane-bed conditions. Scours eroded into unconsolidated medium sand with margins dipping at up to 60° and local overhangs must have been eroded and filled rapidly. In the absence of extensive evidence suggesting liquefaction, slump scars or slump deposits, or faults in underlying strata, erosion is interpreted to have been beneath a hydraulic jump. Hydraulic jumps form where supercritical flows rapidly transform to subcritical flows, as a consequence of rapid flow expansion (Gorrell and Shaw 1991; Komar 1971). Intense erosion within the hydraulic jump, particularly in high Froude number oscillatory jumps, erodes deeply and entrains large quantities of sediment. The thicker, lower-energy flow downflow of the jump has greatly reduced transport capacity, which results in rapid sedimentation. Beds of diffuse-graded or massive sand were previously attributed to traction carpets (Sohn 1997), sandy debris flows (Postma et al. 1983; Shanmugam 1997), hyperconcentrated flows (Gorrell and Shaw 1991; Kneller and Branney 1995; Maizels 1989; Smith 1986), laminar sheared layers (Vrolijk and Southard 1997), or grain flow (Cheel 1982; Lowe 1982). Collectively, these processes involve highly concentrated sediment dispersions, and suppressed turbulence.

Sediment support mechanisms for each of the processes are variable and may include one or a combination of the following: turbulence, dispersive pressure, hindered settling, and buoyancy. Deposition of this facies in locally steep-sided scours is interpreted to have occurred almost simultaneously following scour erosion, and rapid loss of transport capacity by the flow. The downflow transformation of the flow probably involved a number of depositional processes. Where suspension sedimentation was highest, and the sediment flux was predominantly normal to the bed, deposition was possibly from a surge and a simple laminar sheared layer (e.g. Vrolijk and Southard 1997). With decreasing rates of suspension sedimentation, flow stratification and the transfer of shear from the overlying dilute fluidal flow, deposition was from sustained traction carpets (Sohn 1997) (see heterogeneous gravel with intraclasts).

Where tabular diffusely stratified sand grades laterally or is overlain by quasi-planar-stratified sand, the lack of stratification in the diffusely-stratified part was probably related to higher rates of suspension sedimentation. In a flume study, Arnott and Hand (1989) showed that under upper plane-bed conditions, stratification became progressively more diffuse with increasing rates of bed aggradation. The reason for this change was attributed to reduced lateral segregation of grains, and a steeper angle of climb of persistent low-amplitude bed forms. Consequently, the interbedded character of this facies with the quasi-planar stratified facies can likely be related to spatial and temporal variations in sediment concentration and local differences in rates of sedimentation and flow speed.

4.2.7 Small-scale cross-laminated fine sand: (facies Sr)

The small-scale, cross-laminated fine sand facies is the most extensive facies in section M-5 (Fig. 4.15) forming 4-5 m thick units; however it also crops-out in sections M-2a, M-3 and M-4. It abruptly overlies the cross-stratified sand facies, or less commonly, the gravel facies. Strata consist of small-scale cross-laminated fine sand composed predominantly of stoss-erosional or stoss-depositional small-scale cross-laminae; (Fig. 4.16, ripple-drift cross-lamination, Jopling and Walker, 1968). Less common are sinusoidal laminae with steep-angles of climb (>60°) and no heavy mineral laminae. Cross-laminated, silty fine sand to fine sand forms sharp-based sets < 4 cm thick, and cosets < 3 m thick. Where silty the coset thickness is generally thinner. Scours and intraformational clay clasts are rare.

Interpretation: Small-scale, cross-laminated silty fine sand was deposited by low-energy current ripples. Climbing cross-lamination is the result of combined traction and suspension sedimentation and rapid bed aggradation as the flow loses transport capacity (Ashley et al. 1982; Jopling and Walker 1968). Rapid bed aggradation and climbing cross-stratification commonly occur under expanding-flow conditions of density underflows, turbidity currents (Walker 1992) and overbank sedimentation (Harms et al. 1975). A jet efflux from a subglacial conduit would undergo rapid flow expansion as it entered a glacilacustrine environment and rapidly entrain ambient basin water (Fig. 4.1). Thick sequences (100s cm) of small-scale cross-laminated sand are common in glacial subaqueous fan deposits (e.g. Gorrell and Shaw 1991; Rust and Romanelli 1975). Such stratification is comparatively less common in deep submarine fans (e.g. Pickering et al. 1989; Walker 1992) where it is generally described as c-division of Bouma turbidites and is generally < 10s cm thick. This reflects the difference between quasi-continuous efflux flows in glacial settings compared with surge-type turbidity currents in deep-sea submarine fans.

4.2.8 Fine silt - clay: (facies F)

Fine silt-clay up to 2-3 m thick is stratigraphically highest and in most places gradationally overlies the small-scale cross-laminated facies. Fresh sections are mostly massive; however, upon drying 1-3 cm thick, sharp-based, fine sand or silt beds are generally discernible. Rare, < 1 cm thick small-scale cross-laminated fine sand beds occur at the base of individual couplets. Couplets fine upward and form rhythmic successions that are up to 20-30 cm thick. A series of graded sand-silt couplets are commonly overlain by 0.5-2 cm thick clay stratum that is faintly normal graded or massive.

Interpretation: This facies is interpreted to have been deposited predominantly by weak density underflows and suspension sedimentation beyond the zone of jet efflux (e.g. Gilbert 1997). Rhythmic, normal-graded sand-silt-clay strata were deposited by underflows and suspension sedimentation. The thicker clay strata were deposited from suspension. Suspension sedimentation of clay requires low turbulence in the water column and is interpreted to have occurred under winter conditions of low meltwater discharge and surface ice cover of the basin (Banerjee 1973). Consequently, thicker clay strata are interpreted to mark the end of an annual sedimentary cycle and the top of a varve. This facies overlies the coarser fan sediment and marks a shift from jet efflux dominated sedimentation to basin sedimentation following shutdown of the conduit or migration of the conduit efflux.

4.3 Facies Associations

The facies identified in this study (Table 4.2) have been organized into five facies associations (Table 4.3); each is discussed below and the associations are summarized in figure 4.17 and figure 4.18.

4.3.1 Fan-core association

The fan-core association consists of heterogeneous gravel and planar-stratified gravel with minor cross-stratified and quasi-planar stratified sand. It is tabular and erosionally overlies cross-bedded sand and gravel. In outcrop it is ~10-20 m wide, 7-8 m thick, and has a length parallel to flow of more than 40 m (Fig. 4.3, 4.4, 4.17). Upward the strata change from medium-scale cross-bedded gravel, to heterogeneous gravel with sand intraclasts and planar stratified gravel facies. Locally, there is well developed lateral gradational transitions from heterogeneous gravel with sand intraclasts to interbedded planar-stratified gravel, and further downflow to cross-bedded gravel and cross-bedded sand. This upward change in stratal succession is interpreted to reflect deposition from progressively higher energy flows and flows of increased sediment concentration, whereas the downflow transitions indicate rapid flow expansion and loss of transport capacity. Rapid downflow facies changes, lack of definable vertical succession and erosional contacts characterize this association.

4.3.2 Flanking-core association

The flanking-core association occurs lateral to the fan-core association and consists of cross-bedded sand and quasi-planar-laminated sand and minor diffusely graded/massive sand (Fig. 4.5, 4.17). It ranges from 2-8 m thick depending upon the depth of incision by the steep-walled scour association (see below). Beds either offlap the fan-core with a dip of 5-10° or grade laterally from the fan-core. Offlapping beds form a succession of quasi-planar stratified and diffuse-graded sand with multiple low-angle truncations and small scours (Fig. 4.13). In contrast, deposits that grade laterally are generally tabular, horizontal gravelly sand beds that fine laterally away from the fan-core (Fig. 4.5).

4.3.3 Steep-sided scour association

The steep-sided scour association consists of steep-sided, irregular scours eroded into the flanking fan association (Fig. 4.5,4.17) and infilled with diffusely graded sand facies. The largest scour is 10 m wide, 3 m deep and has margins dipping 10-60° (Fig. 4.14a). Locally the scour margins consist of overhangs 20-30 cm long (Fig. 4.14b). Elsewhere the substrate of the scour margins is offset suggesting minor slumping (Fig. 4.14b). The dip of the scour surface generally decreases upward and laterally, eventually becoming subhorizontal and overlain by cross-bedded sand. The coarser scour-fill sediment is poorly exposed but consists of sandy gravel with silty sand intraclasts that fine-upward to diffuse-graded sand The predominant infill sediment is diffuse-graded medium sand. Preservation of the steep-walled scour geometry, and especially the overhangs in a cohesionless sandy substrate indicates that erosion was followed immediately by deposition of the diffuse-graded/massive sand.

4.3.4 Large gently inclined bed set association

The gently inclined bedset association consists of a thick succession of gently dipping surfaces that are overlain by planar cross-stratified sand or small-scale cross-laminated fine sand facies (Fig. 4.10, 4.15, 4.17, Table 4.3). The association is 6-8 m thick and extends downflow subparallel to the paleoflow direction for >150 m. Individual bedsets are up to 60 cm thick and generally thin upward. Bed surfaces dip at 5-10° toward the west - northwest and the dip angle increases gradually basinward (Fig. 4.15). Small-scale cross-stratified sand becomes increasingly more abundant basinward. 1-2 m wide and 1 m deep scours occur locally and are infilled with cross-stratified medium sand.

4.3.5 Shallow channel association

The shallow channel association consists of stacked and nested channel fills of cross-stratified, small-scale cross-laminated sand and minor diffuse-graded/massive sand (Fig. 4.5, 4.17). It erosionally overlies the other stratal associations and is thickest (3 m) over sand-dominated stratal associations (Fig. 4.5, 4.10). Channels are at least 1-2 m deep and 5-20 m wide. Channels are deepest and widest near the base of the association and become progressively shallower upward and also decrease in size downflow. The bases of the channels have rare overdeepened troughs, and coarse lags are rare. This facies association has a very consistent vertical and downflow facies transition to progressively lower-energy, finer grained silt and clay strata.

4.4 Depositional Model

Based on the facies and facies associations discussed above, strata in the study area are interpreted to have been deposited in the proximal region of a glacilacustrine subaqueous fan. Subaqueous fan deposits occur at the downflow terminus of eskers (e.g. Diemer 1988; Warren and Ashley 1994), onlap or are lateral to eskers (e.g. Banerjee and McDonald 1975; Brennand and Shaw 1994; Gorrell and Shaw

1991), or form parts of large moraines (e.g. Fyfe 1990; Sharpe and Cowan 1990). Fan deposits have been interpreted to form in response to seasonal steady-state (e.g. Powell 1990), or episodic glacial hydraulic events (e.g. Gorrell and Shaw 1991). In some cases moraine deposits represent the coalescence of multiple adjacent fans (e.g. Rust and Romanelli 1975). Although no esker deposit was observed to merge into the fan, the study area is in the upland region of the Oak Ridges Moraine, which is a large polygenetic glacial landform (Barnett et al. 1998). Because the moraine is a positive topographic landform, deposition in a glacilacustrine basin required ice-support and a glacial meltwater system for sediment delivery. This ice-confined glacilacustrine basin was bounded by ice and to the west by the Niagara Escarpment. Basin geometry and depth was controlled by the location where ice impinged along the escarpment and by the elevation of outlet channels that breached the escarpment. On the basis of valley morphology, sedimentary deposits and elevation of moraine deposits, Chapman (1985) identified four key outlet channel elevations at 420, 380, 367 and 290 m asl. For the study area, with a surface elevation of 290 m, these outflow levels probably drained a lake 70-120 m deep. The approximate surface extent of the lake can be estimated from the location of the drainage channels, the fan location and surface extent of Halton Till. On this basis a rough estimate of the north-south width of the lake is estimated to have been 30-40 km. The east-west length of the lake basin between the northern Simcoe and southern Ontario ice lobes is more difficult to estimate but probably was < 100 km.

The meltwater-sediment supply to the subaqueous fan was most probably from unsteady episodic flood discharge, termed jökulhlaups, rather than steady seasonal flow. A jökulhlaup interpretation is supported by the lack of stratigraphic evidence for annual sedimentation and the high-energy nature of the facies assemblage that consists of heterogeneous gravel and diffuse-graded sand deposited from hyperconcentrated flows and steep-walled scours eroded beneath hydraulic jumps. In addition, the absence of clay interlaminae, either in situ or as intraformational clasts within the coarse-grained fan succession, suggests that the fan was constructed during a single meltwater season. Within the fill small-scale scours occur with dip angles < 70°.

4.4.1 Meltwater sources

Glacial outbreak flood events (jökulhlaups) span a broad range of discharge magnitudes that have been related primarily to the volume of stored water (Clague and Mathews 1973). Peak discharges range from 1-2 times up to 10 times normal flow rates (e.g. Church 1972) and contribute, in exceptional cases, up to 90% of the total annual sediment yield for a basin (e.g. Maizels 1997). These models indicate that deposition related to jökulhlaups dominate the sedimentary succession in affected basins. Analogues of Laurentide deglaciation meltwater processes, of which jökulhlaups are one component, are based on temperate alpine or piedmont glaciers, arctic icecaps, or sediment deposits described from these localities. These analogues, however, are not ideal because modern glacial environments are not of comparable scale and form under different climatic conditions. For these reasons the role, extent, and magnitude of meltwater in ancient continental glaciers remain highly equivocal. Nevertheless, using the Malaspina glacier as an analogue for the southeastern margin of the Laurentide ice-sheet, Gustavson and Boothroyd (1987) suggested that the glacial hydrologic system was the principal sediment pathway to the glacial terminus. This assertion is supported by studies of large continental moraines such as the St-Narcise Moraine, (Burbidge and Rust 1988); the Harricana Moraine (Brennand and Shaw 1996), the Hartman Moraine (Sharpe and Cowan 1990), or the Salpausslka I (Fyfe 1990) that are composed predominantly of glacifluvial sediment.

Possible jökulhlaup meltwater sources include subglacial, englacial and supraglacial reservoirs with large precipitation events (rain) potentially contributing a significant volume of water to the total discharge (e.g. Fountain and Walder 1998; Tweed and Russell 1999). Furthermore, Shoemaker (1991) has discussed the controls for explaining the origin of lakes beneath the Laurentide ice-sheet. Verification of subglacial reservoir models is emerging from systematic aerial surveys of the Antarctic icecap (e.g. Siegert et al.

1996). Beneath this icecap 5-10% of the surveyed region is estimated to be underlain by subglacial lakes, with individual lakes ranging from 1.3-241 km wide and possibly several hundred metres deep (Siegert et al. 1996; Siegert and Ridley 1998). Additionally, subglacial storage has been implicated in outburst discharges from alpine glaciers (e.g. Rothlisberger and Lang 1987), Alaskan piedmont glaciers (e.g. Fleisher et al. 1997), and for formation of erosional s-forms on bedrock below Antarctic outlet glaciers (Sawagaki and Hirakawa 1997). Consequently, even though the geothermal gradients and /or climatic conditions of these examples differ from that inferred for the Laurentide ice sheet, there is significant evidence to support the probable former existence of lakes beneath the Laurentide ice-sheet.

Supraglacial lakes are another possible source of water for major discharge events. On modern glaciers supraglacial lakes form during the early melt season as areas of slush and ponded water (Muller 1962; Rothlisberger and Lang 1987; Weidick 1988). Although the depth and areal extent of these lakes has been poorly documented, in two examples from Greenland, Thomsen (1989) and Russell (1993) described lakes up to 10 m deep and ~1500 m wide that drained through moulins. For alpine glaciers with low surface slope (< 1.5°) deeper lakes form in areas where the roofs of shallow englacial conduits have collapsed (Kirkbride 1993). Drainage from supraglacial lakes is controlled by the continuity of conduit connections between the surface and englacial or subglacial systems. These studies provide support for interpretations of Laurentide supraglacial meltwater accumulation and drainage. In fact, Shaw (1996) used regional landform and sedimentological evidence to support a model of supraglacial meltwater accumulation as a source of regional jökulhlaup drainage from the Laurentide Ice-sheet.

A third alternative for catastrophic meltwater discharge is high intensity precipitation events, either in the spring (Rothlisberger and Lang 1987) or during the later part of the meltwater season (Cowan et al. 1988). Such events augment existing water storage within or on a glacier, and thereby help provide a sudden peak in meltwater discharge. For a spring precipitation event, Warburton and Fenn (1994) recorded peak discharge of 10 m s^{-1} and estimated that 22% of the total annual sediment load was deposited during a 4 day period. Additionally, a series of discharge events were recorded that caused the sediment flux to pulsate at scales ranging from 5 minutes to several days. Such fluxes in discharge are common to both precipitation and outbreak events (Russell and Knudsen 1999) and are connected to the evolution of the subglacial drainage system. Late season events also produce peak discharge events with sediment fluxes of 5-6 times normal seasonal values (Cowan et al. 1988). For late season events supraglacial storage is probably less significant and rapid drainage of meltwater via a well established and snow-free ice-terminus is more likely cause of a rapid increase in discharge.

The proposed jökulhlaup that constructed the fan described in this study possibly originated from one of several individual reservoirs of either subglacial or supraglacial position, or possibly a combination of sources including precipitation. Studies of spring events in alpine glaciers indicate a tripartite contribution to peak discharges of, i) subglacial water pockets, ii) existing supraglacial storage, and iii) precipitation input (Warburton and Fenn 1994). Routing of and fluctuations in discharge are influenced by changes in subglacial cavity form and connectivity, changes in the contribution from individual meltwater sources, and development and collapse of conduits. These controls produce flow pulsations that influences the discharge magnitude, and sediment load. Discharge at the terminus, regardless of the exact source(s) of the meltwater, was subglacial, or if englacial was delivered at the same elevation as the fan surface.

4.4.2 Sedimentary events

Depositional setting

Cross-bedded sand and gravel within the Oak Ridges Moraine have been interpreted as either braided fluvial, deltaic or subaqueous fan deposits (Barnett et al. 1998; Duckworth 1979; Paterson and Cheel 1997). Additionally, incision of underlying subaqueous fan deposits and infill has been used to support an interpretation of falling lacustrine water levels, subaerial exposure and subsequent rising lake levels

(Paterson and Cheel 1997). The similarity of sedimentary structures and facies produced by unidirectional currents is a common interpretative problem, whether in submarine fan (Hein 1984), glacifluvial esker or subaqueous fan deposits (Rust and Romanelli 1975). In this study a subaqueous fan setting interpretation is favoured for the following reasons: rapid coarse-grained sediment aggradation; overlying basinal silt-clay facies; and similarity with other subaqueous fan deposits. The inferred rapid bed aggradation of these deposits indicates deposition in a relatively deep subaqueous environment unconstrained by base level and accommodation space. There is also extensive evidence of rapid flow expansion and rapid suspended sediment deposition from both climbing dune-scale cross-strata and climbing small-scale cross-laminae. Rapid bed aggradation and flow expansion are conditions more characteristic of subaqueous fans than subaerial braided fluvial deposits. Additionally, the entire fan is overlain by a 2-3 m succession silt-clay interpreted to be glacilacustrine deposits. Thus in the absence of evidence for water level fluctuations the entire succession is interpreted to be deposited in a glacilacustrine environment. Finally, although the shallow channel association has similarities to a braided fluvial system, similar facies associations have been observed in glacigenic subaqueous fans (Burbidge and Rust 1988) and deep water submarine fans (Hein and Walker 1982) and submarine canyon fills (Arnott and Hein 1986).

Flow regime

Sediment was transported to the fan and deposited during a single, semi-continuous flow event. The fan-core, flanking fan-core and steep-walled scour facies associations were deposited from a high-energy unsteady, non-uniform flow within the ZEF and ZFT. Deposition was near peak discharge or during early waning flow conditions. By contrast, the shallow-channel facies association indicates a period of low-energy steady flow and the deposition of cross-bedded sand and small-scale cross-laminae. Temporal variations in flow conditions, and hence the nature of sedimentation, were related to changes in the fluid flux caused by changes in the meltwater conduit system and the stage of flood discharge. Broad spatial flow variations are related to flow expansion. Local flow variations were probably caused by fan morphology and consequent flow thinning and thickening.

Non-volcanic jökulhlaups are commonly characterized by gradual waxing flow, as the conduit system enlarges, followed by rapid flow waning (e.g. Maizels 1993; Tweed and Russell 1999). This asymmetry in discharge is related to the increased capacity of the discharge system with time. As the meltwater reservoir empties, a threshold is reached where fully pressurized flow can no longer be maintained and flow conditions change from confined to unconfined. This transition is accompanied by a marked decrease in the maximum sustained flow energy and a corresponding decrease in sediment transport capacity. Assuming a reservoir source, an approximate discharge volume can be estimated. An empirical relationship developed by Clague and Mathews (1973) relates peak discharge, Q_{max}, to maximum lake volume, V_{max}:

$$Q_{max} = 75(V_{max})^{0.67}$$

By using the observed facies distribution within the jet efflux model, the diameter of the conduit can be estimated to be ~10 m. A paleoflow estimate of between 1-5 m s^{-1} can be obtained using the maximum gravel clast size of 10 cm (cf. Williams 1983). The technique of Williams was based on open channel flow calculations, however, for pipe flow the entrainment threshold velocity is 2.5-3 times greater than for open-channel flow (Powell 1990). Consequently, velocities estimates are probably in the range of 2.5-10 m s^{-1}. These estimates would indicate that Q_{max} was between 250 and 1200 m^3 s^{-1} and V_{max} for the discharge event was from 6.0×10^6 to $62.7 \times 10^6 m^3$. This volume is of the same order as a number of ice-marginal lake jökulhlaup volumes reported by Clague and Mathews (1973) but is an order of magnitude greater than meltwater volumes potentially available from documented supraglacial lakes (e.g. Russell 1993). It is important to point out that this empirical approach, assumes that discharge was uninterrupted and from a single reservoir source. Consequently the calculated volume will be underestimated in the case of incomplete reservoir drainage, or where other meltwater sources including precipitation input were involved.

Plane-wall jet fan sedimentation

A plane-wall jet forms where a supercritical flow debouches from a conduit and undergoes rapid flow expansion (Fig. 4.1). Such flows are inherently characterized by the formation of a hydraulic jump and intense turbulent mixing of the jet efflux and ambient basin water (Fig.4.18a, Gorrell and Shaw 1991). The plane-wall jet efflux is interpreted to have produced a wide range of sediment textures from high-energy sediment charged flows. Within the zone of flow establishment (ZFE), only heterogeneous and planar-stratified gravel were deposited (Fig. 4.18). These strata were deposited from high-energy flows that partially or wholly eroded underlying lower-energy deposits of cross-stratified sand and gravel. Deposition during these high-energy events resulted from a combination of hyperconcentrated dispersion that transported even pebbles in suspension and an overlying fluidal flow.

Downflow and lateral to the ZFE the zone of flow transition (ZFT) developed over a distance of 4 to 8 times the conduit diameter (4-8 D). Assuming a conduit diameter of 10 m this equates to a zone 40-80 m downflow of the conduit mouth. Here the flow would have expanded, and ambient basin water would be rapidly entrained inward toward the jet axis accompanied by extensive turbulent mixing. Initial flow entrainment of basin water would have further concentrated sediment within the basal part of the flow. Increased sediment concentration within this zone could have allowed a basal hyperconcentrated zone to develop within a stratified flow. Deposition from the basal hyperconcentrated traction carpet would cause the bed to rapidly aggrade, emplacing massive or poorly bedded heterogeneous gravel deposits 2-3 m thick. Rapid downflow evolution from a hyperconcentrated dispersion to more fluidal flow conditions is indicated by the lateral transition from massive or poorly bedded heterogeneous gravel to plane-bed and cross-bedded deposits over a distance of ~5 m (Fig. 4.4). These deposits have a striking similarity with coarse sand and gravel rhythmites described from recent jökulhlaups in Iceland (Russell and Knudsen 1999). The 8-10 m thick Icelandic succession was deposited over a maximum of 17 hours and consists of sharp-based graded couplets of sandy gravel deposited by unsteady flow pulses that ranged from seconds to minutes in length (Russell and Knudsen 1999).

As sediment concentration decreased and flow turbulence concomitantly increased, the jet would have expanded rapidly. Initiation of hydraulic jumps would cause extensive local erosion and downflow deposition. Where relatively low-energy undular or weak hydraulic jumps developed along the side of the fan-core quasi-planar stratified sand was deposited (Fig. 4.5). Under higher energy conditions and oscillatory hydraulic jumps (2.5< Fr < 4.5), more extensive erosion affected the flanking fan-core association and formed steep sided scours. Vertical oscillation of the jet rapidly scoured and entrained large quantities of sediment. Reduced transport capacity as the flow expanded resulted in flow stratification. A basal hyperconcentrated layer developed and sediment was deposited by a variety of mechanisms that evolved downflow from en masse suspension sedimentation, to deposition from laminar shearing surge flows and traction carpets to eventually deposition from fluidal flows. Locally, the highest concentrations of sediment were deposited as heterogeneous sandy gravel with intraclasts (Fig. 4.9b, 4.14b). Elsewhere, 5-10 cm thick diffusely graded sand beds infilled scours during multiple stages of scour and deposition (Fig. 4.14a). The more dilute overlying fluidal flow region continued basinward as subcritical underflows, resulting in deposition of a succession of cross-stratified and small-scale cross-laminated medium and fine sand beds in the zone of established flow (Fig. 4.15). Downflow of the hydraulic jump, migrating dunes constructed a 4-6 m thick 250 m long gently dipping accretionary wedge in the proximal ZEF (Fig. 4.10,4.15). Within the upflow 100 m of this zone, sets of dune-scale cross-bedding fines rapidly downflow to mostly ripple-scale cross-lamination. The basinward accretion of the fan is recorded by these low-angle bed sets (Fig. 4.15).

Migration of the conduit mouth or a change to a lower-energy plane jet is defined by the abrupt development of the shallow-channel facies association (Fig. 4.17). Rapid migration of conduit position due to ice-front collapse or conduit roof collapse is common where jökulhlaups have occurred (e.g. Fleisher et al. 1997; Russell and Knudsen 1999) and would result in a consequent shift in the ZFT toward the earlier fan apex. These deposits represent the final stage of efflux from the conduit before the conduit flow either migrated laterally, was captured by a lateral system, or the entire meltwater system shutdown. Alternatively, a shift to a lower-energy plane jet occurred in response to depletion of the meltwater reservoir and a change from pipe-full to partly filled conduit discharge. The resultant loss of flow energy resulted in a loss of transport competence and capacity, and a marked change in the dynamics of the jet efflux. Consequently, sedimentation on the fan was controlled by subcritical bed-load transport and development of a subaqueous braided channel network.

The complete fan is overlain by the 2-4 m thick silt-clay facies. Silt rhythmites in this facies probably record the distal sedimentation signal of efflux discharge elsewhere in the basin and graded Bouma divisions T_c-T_d. Where only graded T_d silt-clay couplets occur, deposition was predominantly by suspension sedimentation unrelated to jet efflux. The rhythmites thus record the end of fan development.

4.5 Discussion

In the subaqueous fan model review (section 4.2) two different types of jet efflux were considered, plane-jet and plane-wall jet along with different efflux geometries, an elongate parabola or a stubby fan-shaped parabola (Fig. 4.1). The geometry of the jet efflux and consequently the fan deposits are related to the relative magnitude of inertial energy and interfacial shear. In the case of high interfacial shear, the fan is shorter in axial length and more hemispherical in shape. For a plane-wall jet, interfacial shear will be particularly high as turbulent mixing is increased by the action of the hydraulic jump. Consequently, the geometry of the subaqueous fan deposited from a plane-wall jet probably would be shorter and broader than the idealized parabolic form discussed in detail by Bates (1953) and is better modelled as a friction-dominated jet efflux (Powell 1981; Wright 1977). Distribution of gravel facies suggests that the ZFE and ZFT have similar lengths for both the plane jet and plane-wall jet. For the ZFE this is expected as little mixing occurs in this proximal efflux region. For the ZFT the relationship is less well-constrained and lengths may vary depending upon sediment concentration, flow velocity and flow thickness. For the ZEF, however, the downflow length is likely to be shorter for the plane-wall jet, because the turbulence generated in the hydraulic jump dissipates the inertia of the jet efflux much more efficiently. This increases proximal sedimentation rates, which in turn reduces the downflow transport of coarse sediment. This produces a marked contrast in grain-size between the proximal fan ZFE and ZFT deposits and those in the more distal ZEF zone. For example, in this study, gravel was absent in the western 200 m of exposed section. An attempt can be made to estimate the percentage of the total fan length that has been described in this study. Due to the number of variables involved in estimating boundary conditions for the friction-dominated jet, no penetration distances are calculated. Instead, estimates are made for the simple condition of an inertia-dominated plane-jet efflux. In this case the basic information required is an estimate of conduit diameter. Assuming a conservative 10 m diameter conduit the ZFE should have an approximate length of four times the diameter (4D), or 40 m long. This is approximately double the length of observed gravel exposure, which suggests either upflow continuation of the fan core association or discharge from a smaller conduit. The ZFT has a similar length of 2-4D for an additional 20-40 m extension downflow. Using the extent over which diffuse-graded sand was observed as a proxy for the ZFT provides a distance of ~30-40m, which corresponds closely with the theoretical estimate. Using an ideal inertia-dominated plane-jet model the 300 m long section observed in this study would correspond to 1.5% of the total 20,000 m distance of jet penetration into the basin. As previously noted for friction-dominated jets the penetration distance is greatly reduced. Nevertheless, even with a reduction of 50-75% or a distance of 10-5 km, the 300 m of streamwise exposure correspond to only ~3-6% of the total potential distance of jet efflux discharge into the basin during fan construction.

Downflow facies changes within this fan have a sharp break between gravel-rich sediment upflow of the steep-walled scour association and downflow where gravel is absent. Furthermore, this break also corresponds with a change in observed sedimentary structures, grain size, and inferred flow strengths. Upflow of the steep-walled scour association there is extensive evidence of supercritical flow conditions with planar-stratified gravel and quasi-planar stratified sand. Downflow and up-section of this association sediment was deposited under critical or supercritical flow conditions. As a result, the steep-walled scour association marks a zone of rapid flow expansion and development of a hydraulic jump. A similar interpretation has been previously proposed by Gorrell and Shaw (1991) for steep-walled scours in a subaqueous fan deposit near Ottawa, Ontario. The lateral and vertical facies associations described in this study hold additional evidence for the formation of such scours by fluid turbulence associated with hydraulic jumps.

Discriminating between subaqueous fan strata deposited by episodic jökulhlaups versus a seasonal steady-state meltwater regime is crucial to the understanding of late Laurentide deglaciation. From Icelandic sandurs, Maizels (1993) has recognized facies and facies associations that can be used to differentiate seasonal diurnal flood deposits from outbreak jökulhlaup deposits of both non-volcanic and volcanic origin. Key elements include, succession thickness, recognition of deposits of traction verses hyperconcentrated flow origin, and cyclicity. Based on these and other criteria related to supercritical flow conditions, the subaqueous fan of this study have been interpreted to have accumulated during a single jökulhlaup. In this case, however, the jökulhlaup is not necessarily the consequence of an outbreak release, but may also have been triggered by a large rainstorm. There are no salient sedimentological characteristics that permit differentiation of precipitation triggered jökulhlaups from outbreak events.

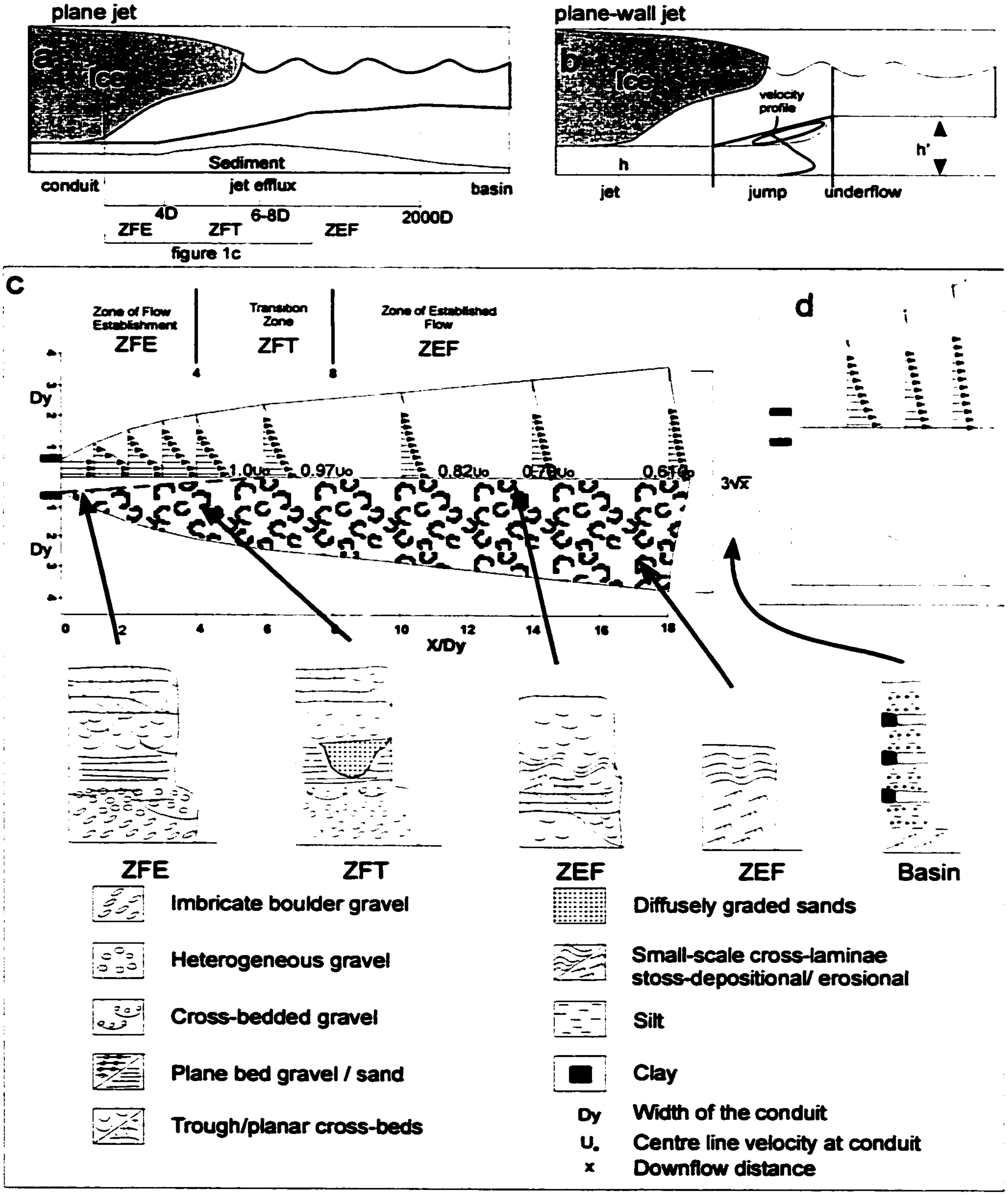
Resolving meltwater sources for such events has important implications for understanding genesis of moraines which have generally been interpreted to be ice-marginal deposits recording long term occupation and seasonal sedimentation (e.g. Hillaire-Marcel et al. 1981). Recent detailed sedimentological studies (Barnett et al. 1998; Brennand and Shaw 1996; Sharpe and Cowan 1990), however, have challenged this concept and suggest that large moraines are composed of deposits from episodic outbreak jökulhlaups. The fan deposit described here is only a small element in the much larger Oak Ridges Moraine. Nevertheless, recent work has suggested that the Oak Ridges Moraine consists of low-energy seasonal basinal sedimentation (Gilbert 1997), seasonal proximal fan deposition (Paterson and Cheel 1997) and mixed seasonal and higher energy episodic flood deposition (Barnett et al. 1998). Data from this study provide additional support for short-lived episodes of voluminous sediment flux to the moraine by high-energy flows.

4.6 Conclusion

Eight facies and five facies associations have been described from 6-8 m high vertical sections cropping out over a distance of 250-300 m subparallel to the general paleoflow direction. The facies and facies associations can be explained in terms of into a jet efflux subaqueous fan model. Deposition is interpreted to have occurred within an evolving single, semi-continuous jet efflux with three distinct stages of flow development: i) Zone of Flow Establishment (ZFE), ii) Zone of Flow Transition (ZFT), iii) Zone of Established Flow (ZEF, Fig. 4.1). A coarse fan-core composed of gravel facies marks the ZFE. This is flanked laterally and downflow by quasi-planar stratified and diffuse-graded sand facies of the ZFT. Sedimentological interpretations suggest transitions from hyperconcentrated flows to fluidal flows over distances of only 2-5 m were caused by rapid flow expansion and loss of transport competence. Within the steep-walled scour association of the ZFT, the scours and gradded/massive infill were deposited downflow of hydraulic jumps that formed as the flow changed rapidly from supercritical to subcritical flow. Here rapid suspension sedimentation was due to a loss of flow capacity and development of hyperconcentrated dispersions and deposition from surge and/or traction carpets producing diffuse-graded sandy rhythmites. Downflow the transition to the ZEF is marked by accreting foresets of planar cross-bedded sand deposited predominantly by 2-D dunes. Strata fine downflow and eventually grade into small-scale cross-laminated sand. Waning jet discharge is recorded by the overlying shallow channel association and cross-stratified medium sand and small-scale cross-laminated fine sand interpreted to be a subaqueous braided system. Fan development and aggradation was followed by a period of basinal suspension sedimentation that overlies the fan and marks fan abandonment.

Discussion of fan facies and architecture within a framework of jet efflux dynamics provides an improved understanding of subaqueous fan sedimentation. This fan succession suggests that the late stages of the Oak Ridges Moraine formation were strongly influenced by local conduits along the margins of a glacilacustrine basin. At least locally, fans were controlled by jökulhlaup discharges that built individual fans during a single meltwater season. High-energy deposition is suggested by rapidly aggrading facies associated with supercritical flow and the flow transition to subcritical conditions. Deposits characteristic of such conditions include: diffuse-graded facies, quasi-planar-stratification, and steep-sided scours. Recognition of jökulhlaup deposits within subaqueous fan deposits and as components of moraine deposits has important implications for interpretation of moraine genesis. For example, moraines may be constructed rapidly over short time periods by episodic meltwater discharges unrelated to climate-modulated ablation and seasonal meltwater production.

Fig. 4.1. Summary of glacilacustrine subaqueous fan model. (a) plane jet. (b) plane-wall jet (modified from Gorrell and Shaw, 1993). (c) Plan view of inertia-dominated jet-efflux development into basin (modified from Bates, 1953) and examples of vertical lithofacies. (d) schematic of friction-dominated jet modified from Wright, (1977).



plane jet
a
Ice
Sediment
conduit
jet efflux
basin
4D
6-8D
2000D
ZFE
ZFT
ZEF
figure 1c
plane-wall jet
b
Ice
velocity profile
h
h'
jet
jump
underflow
c
Zone of Flow Establishment
ZFE
Transition Zone
ZFT
Zone of Established Flow
ZEF
Dy
1.0Uo
0.97Uo
0.82Uo
0.61Uo
3√x
X/Dy
d
ZFE
ZFT
ZEF
ZEF
Basin
Imbricate boulder gravel
Heterogeneous gravel
Cross-bedded gravel
Plane bed gravel / sand
Trough/planar cross-beds
Diffusely graded sands
Small-scale cross-laminae stoss-depositional/ erosional
Silt
Clay
Dy Width of the conduit
U. Centre line velocity at conduit
x Downflow distance

Fig. 4.2. (a) Location of study area in southern Ontario and areal extent of exposed Oak Ridges Moraine sediment. (b) Location of sections described, photo from Ontario Hydro, flown 1991. Paleoflow data presented as frequency percent, outer circle is 10 percent. Only measurements from cross-beds plotted.

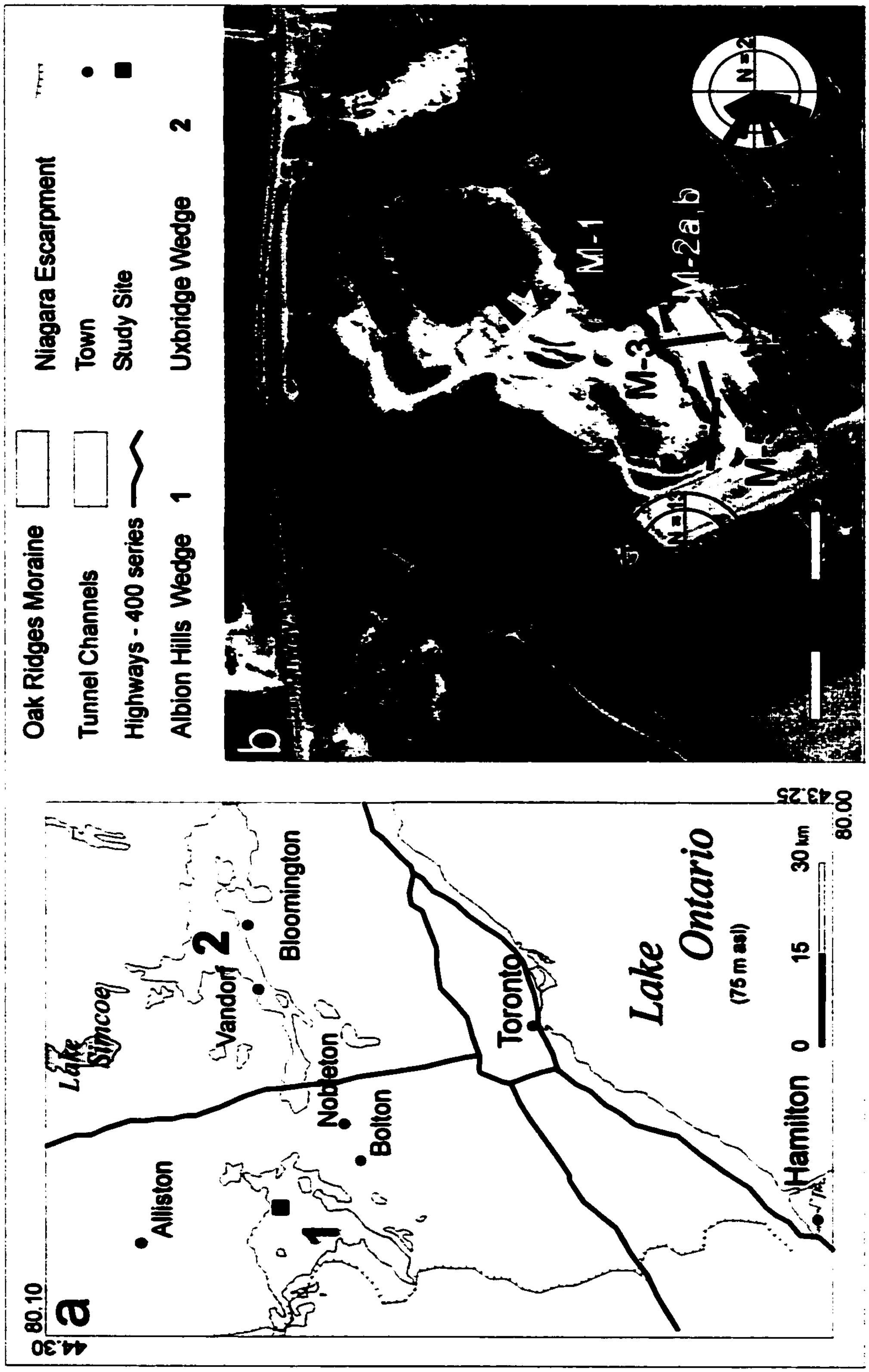

a
44.30
80.10
Alliston
1
Lake Simcoe
Vandorf
2
Bloomington
Nobleton
Bolton
Toronto
Lake Ontario
(75 m asl)
Hamilton
0 15 30 km
43.25
80.00
Oak Ridges Moraine
Tunnel Channels
Highways - 400 series
Albion Hills Wedge 1
Niagara Escarpment
Town
Study Site
Uxbridge Wedge 2
b
M-1
M-2a,b
M-3

Fig. 4.3. Photo mosaic and line drawing of section M-2a. Lower poorly exposed beds consist of cross-stratified gravel and minor planar-stratified gravel. The upper left side consists of poorly-bedded gravel of the heterogeneous gravel facies overlain by cross-stratified sand (scale is 1.5 m at arrow). The right side of face consists of cross-stratified sand and diffusely-graded/massive sand infilling a scour into minor small-scale cross-laminated fine sand. Flow is from left to right and is approximately parallel to section face. Boxes indicate approximate location of subsequent figures. Facies codes are defined in table 4.2.

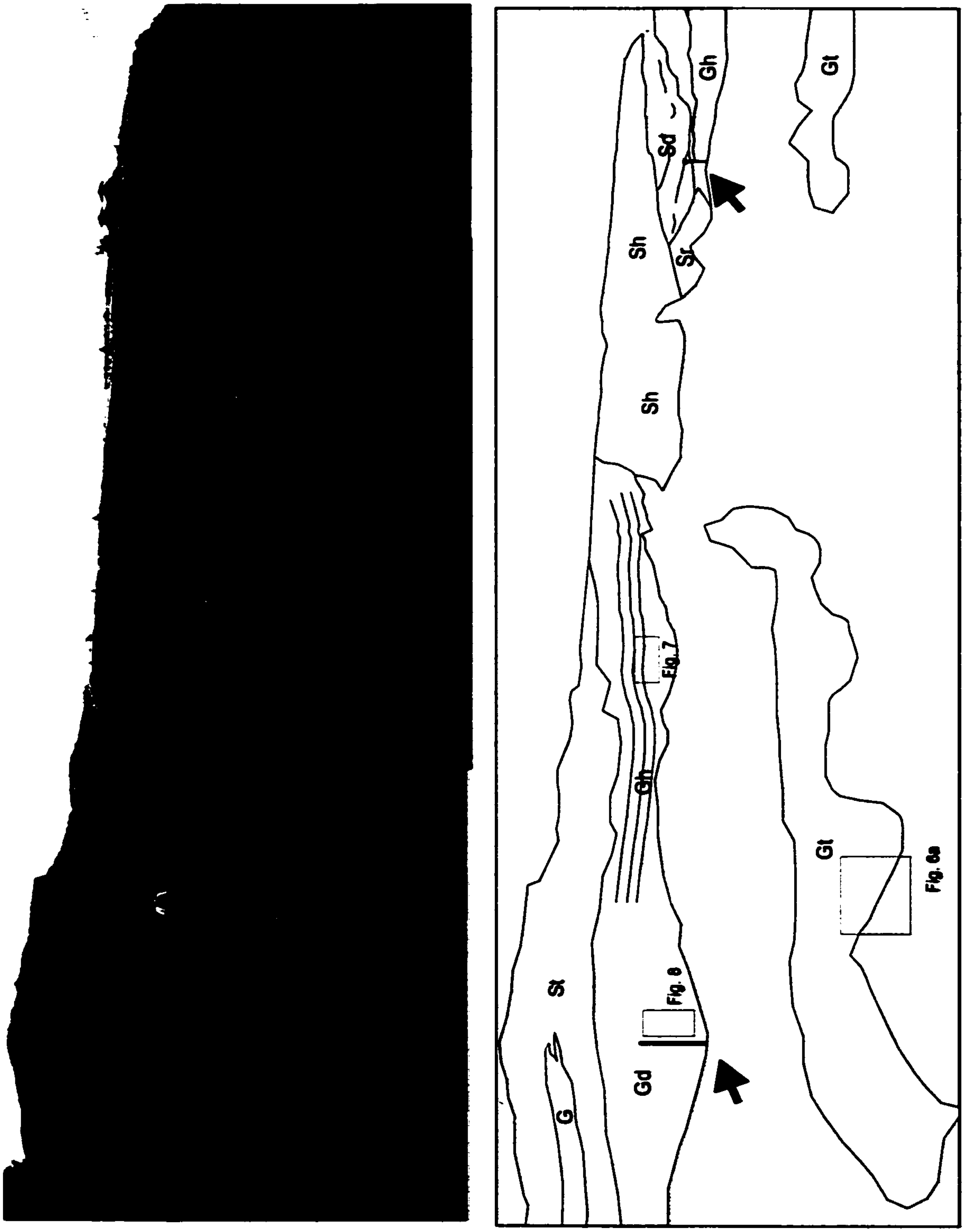

Gh
Gt
Sd
Sh
Sr
Sh
Fig. 7
Gh
Gt
Fig. 6a
St
Fig. 8
Gd
G

Fig. 4.4. Photo mosaic and line drawing of section M-2b. Section is subparallel and approximately 10 m to the south of section 2a. Note large intraclast in the heterogeneous gravel subfacies and rapid downflow transition to interbedded plane-bed and cross-bedded medium sand, flow is from left to right. Shovel is 1 m long.

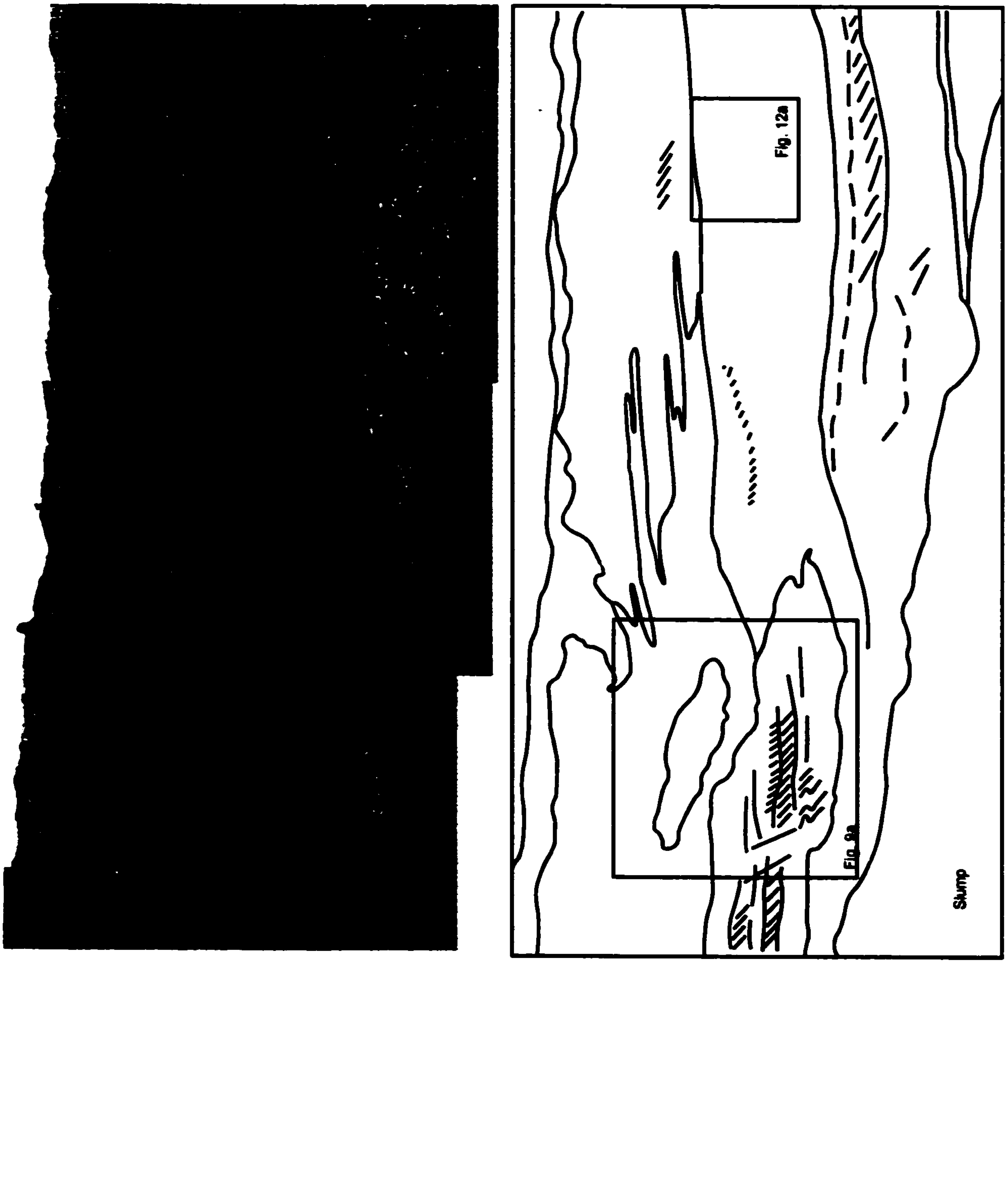
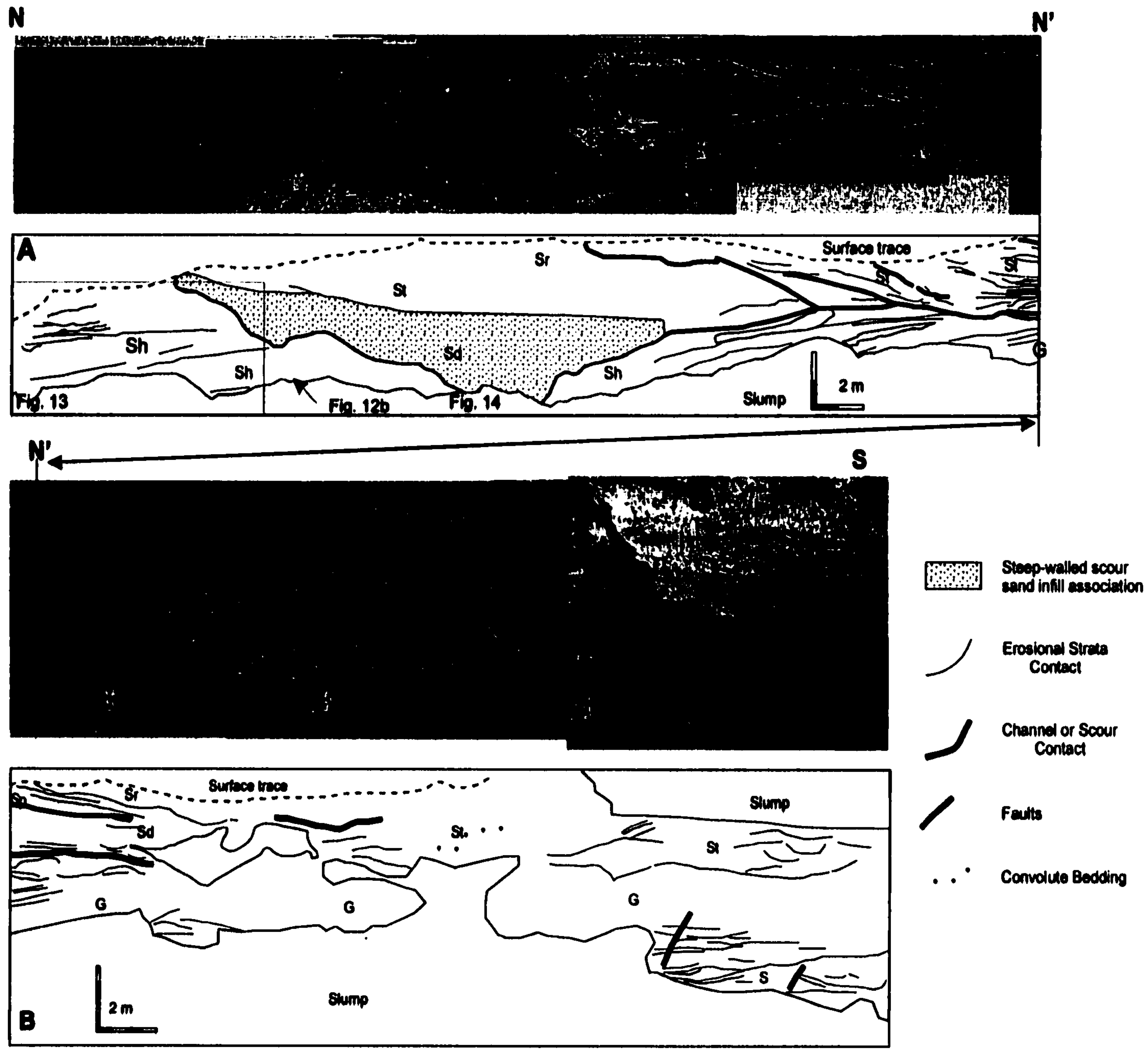


Fig. 12a
Fig. 9a
Slump

Fig. 4.5. Photo mosaic and line drawing interpretation of section M-3. Strata are dominated by quasi-planar stratified (Sh), diffusely-graded/massive Sd), trough cross-stratified (St), and small-scale cross-laminated sand (Sr). Figure numbers on line drawings refer to figures showing details. Flow is obliquely out of the page toward the left.



N
N'
A
Surface trace
Sr
St
Sd
Sh
G
Slump
2 m
Fig. 13
Fig. 12b
Fig. 14
N'
S
B
Sd
St
S
Steep-walled scour sand infill association
Erosional Strata Contact
Channel or Scour Contact
Faults
Convolute Bedding

Fig. 4.6. Cross-stratified gravel. (a) Poorly sorted trough-cross stratified gravel, scale is 8 cm long. Flow is from left to right. (b) Normally-graded medium-scale cross-bedded pebbly gravel. Flow is from right to left. Knife is 23 cm long.

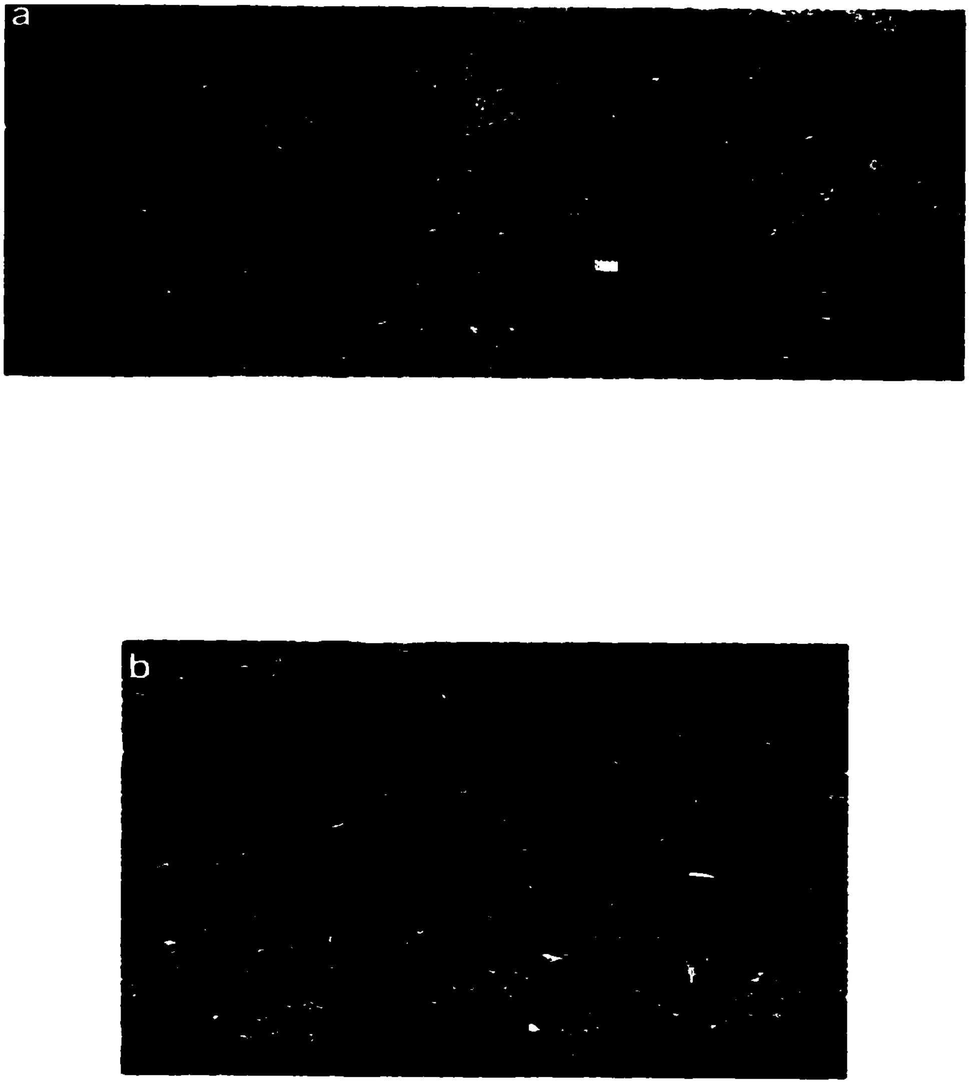

a
b

Fig. 4.7. Planar stratified gravel facies showing upward-fining couplets, clast clusters, and open-work pebble beds. Flow is from left to right.


20 nm

Fig. 4.8. Heterogeneous gravel facies. (a) Faintly bedded heterogeneous gravel showing local clast clusters and upward change in bedding style. Increments on scale are 10 cm. (b) close-up of area outlined in (a). Flow is from left to right. Coin is 1.8 cm diameter.

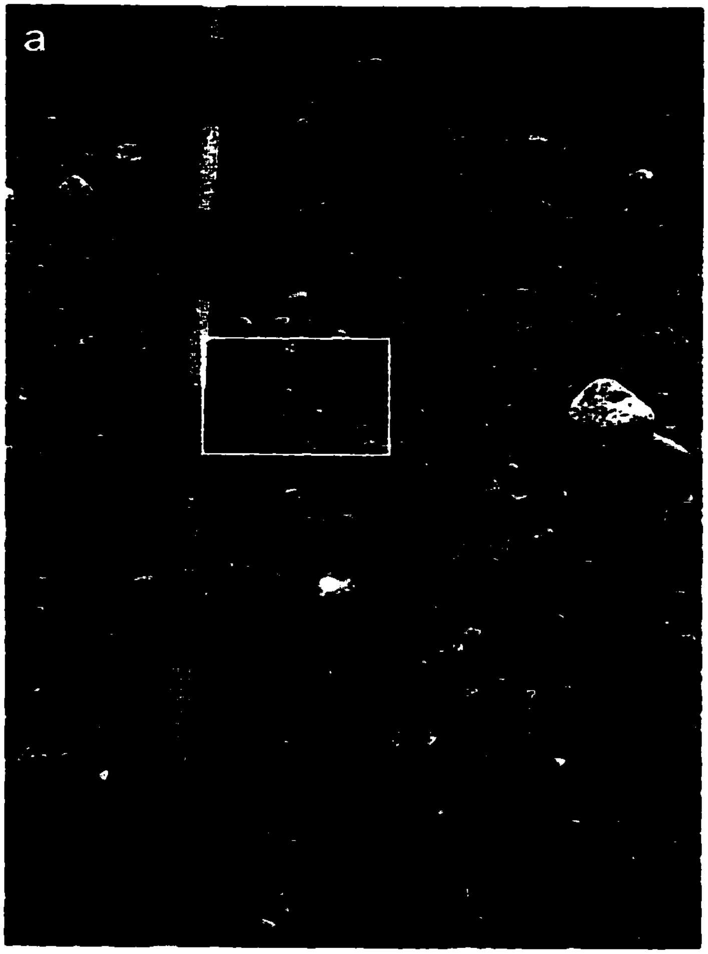
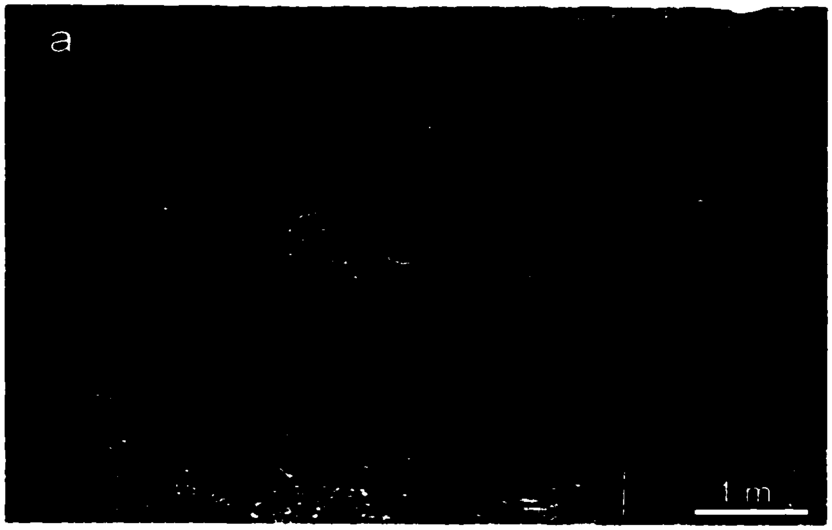

a

Fig. 4.9. Laminated unconsolidated intraclasts in the heterogeneous gravel subfacies. (a) Large intraclast in massive heterogeneous gravel. Shove handle is 1 m long. (b) Intraclast in sandy gravel with fine matrix haloes. Knife is 23 cm long. Flow is from left to right in both photos.


a
1 m

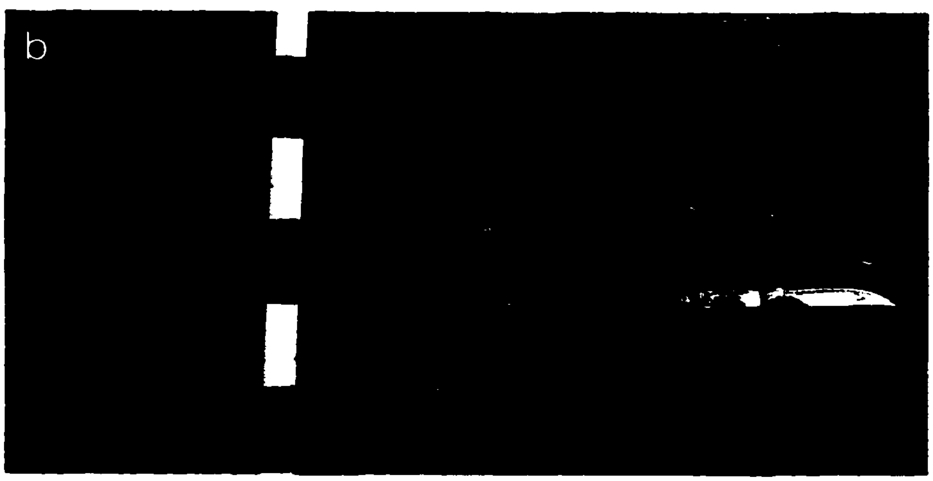

b

Fig. 4.10. Photo mosaic and line drawing interpretation of section M-4. Note predominance of planar cross-bedded sand (Sp) in the lower part of the section and overlying trough cross-bedded(St) and small scale cross-laminated fine sand (Sr) in the upper part. Lower section consists of the gently inclined bed set association, whereas, the upper part of the face (arrow) forms part of the sandy shallow-channel association. Flow is from left to right. Scale is 1 m long.

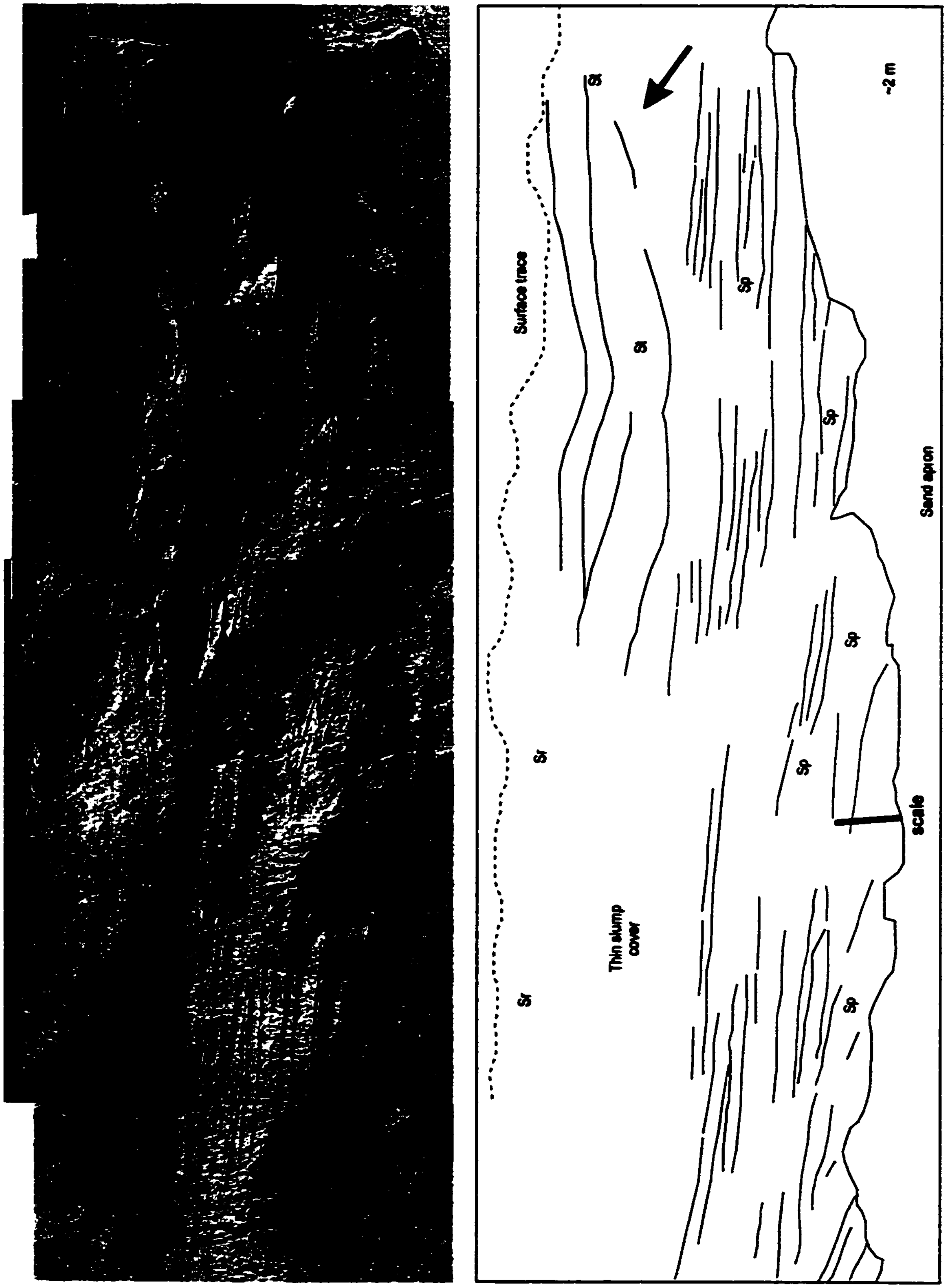
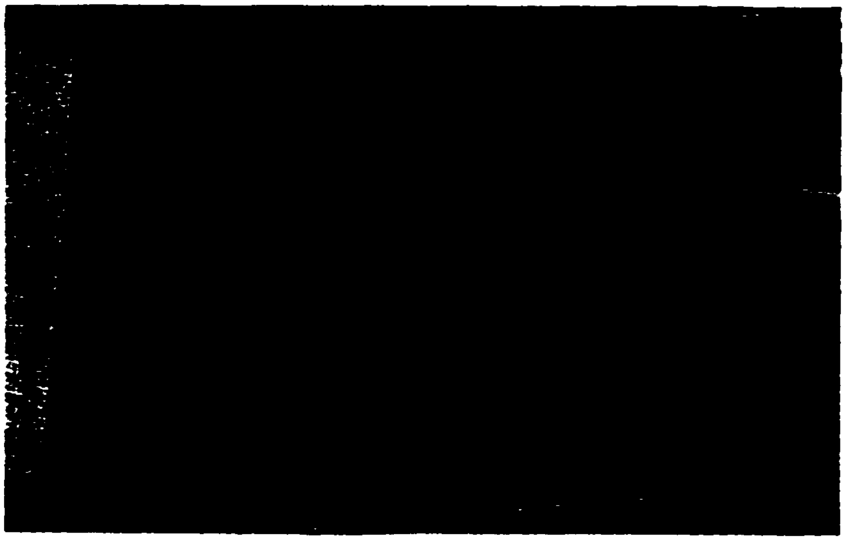

Surface trace
St
St
Sp
Sp
Sp
Sp
Sp
Sr
Sr
Thin slump cover
Sand apron
scale
~2 m

Fig. 4.11. (a) Planar cross-bedded medium sand fining upward to small-scale cross-laminated fine sand, flow is from right to left. Notebook is 17 cm long. (b) Climbing medium-scale cross-strata, gravel lens is slump from overlying strata, flow is from left to right. Scale increments are in 10 cm.



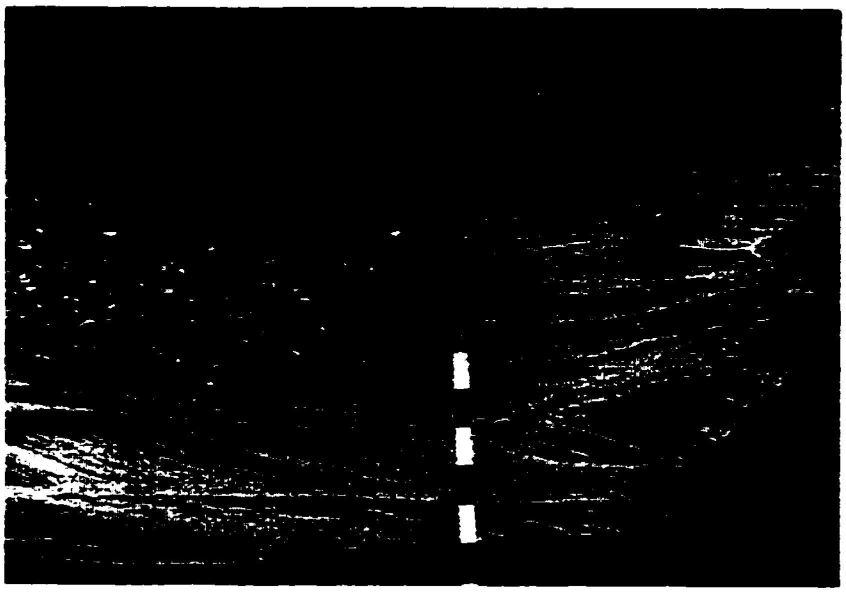


Fig. 4.12. Quasi-planar laminated medium sand facies. (a) Interbedded plane-bed sand and cross-stratified sand, flow is from left to right. (b) Low-angle cross-strata with multiple low-angle erosional surfaces beneath steep-walled scour, note smaller scours with massive sand fill. (c) close-up of (b) showing undulations in beds, onlapping of beds on basal contact, and at top beds parallel to underlying erosion surface (scale is 23 cm long).In b and c flow is obliquely out of photo from right to left.

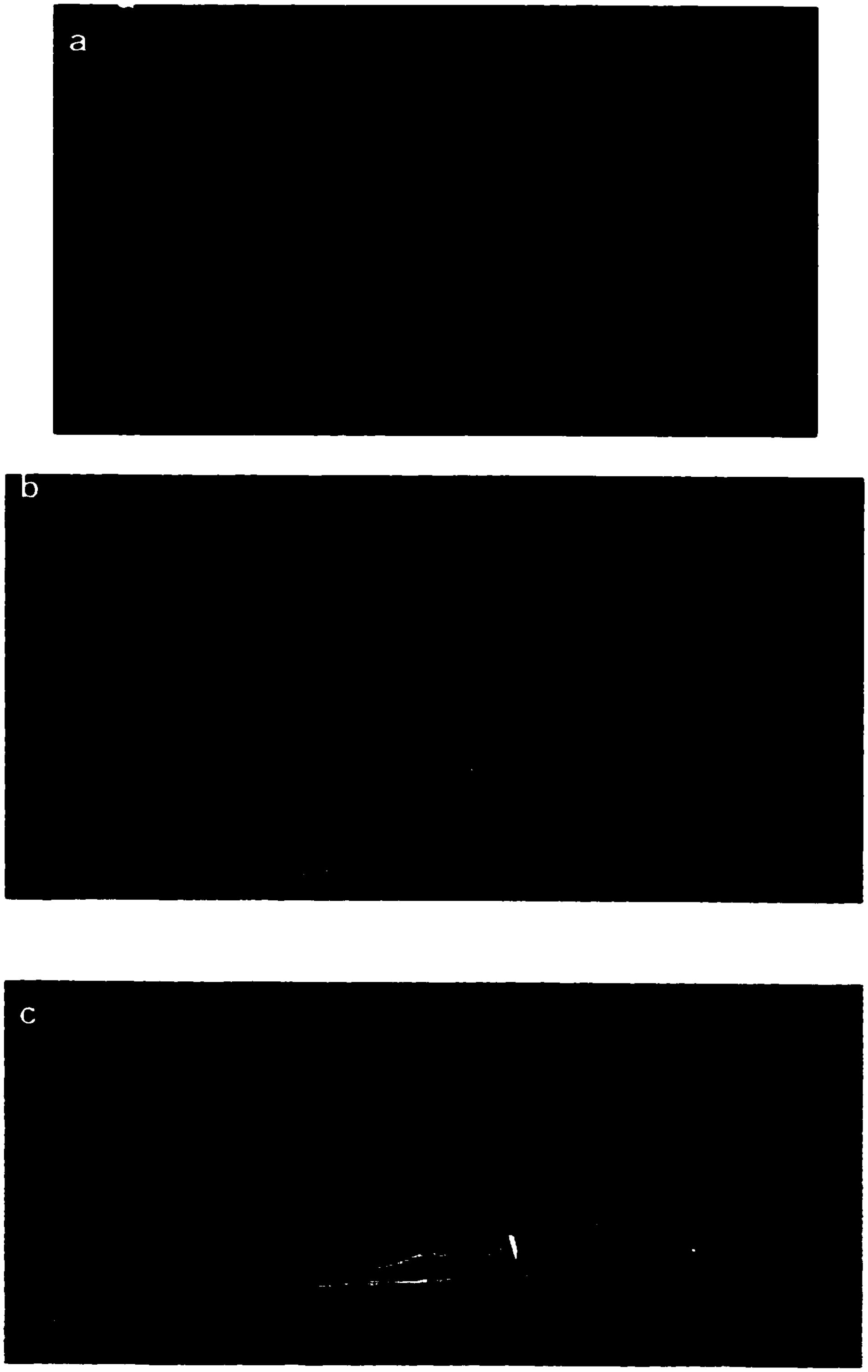

a
b
c

Fig. 4.13. Quasi-planar laminated medium sand subfacies, showing bedding relationships with scour surfaces. Note the onlapping bed relationships with erosional surfaces and down-flow transition to conformable succession of beds, change in bed dip and thickness vertically, and small scours infilled with massive sand (arrow). Steep-walled scour truncates beds to right. Succession is interpreted as antidune stratification formed beneath in-phase waves of an undular hydraulic jump recording rapid bed agradation.

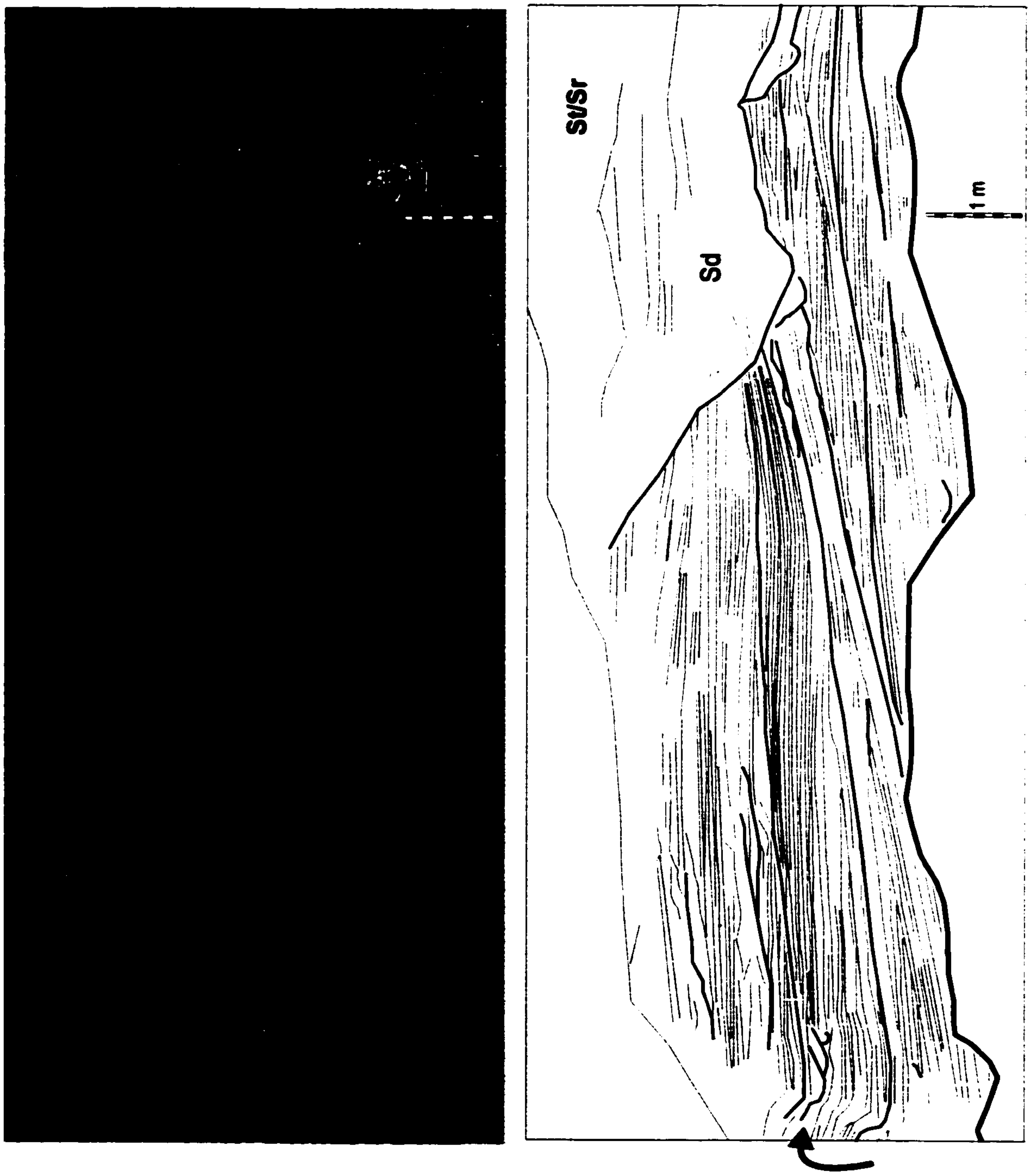
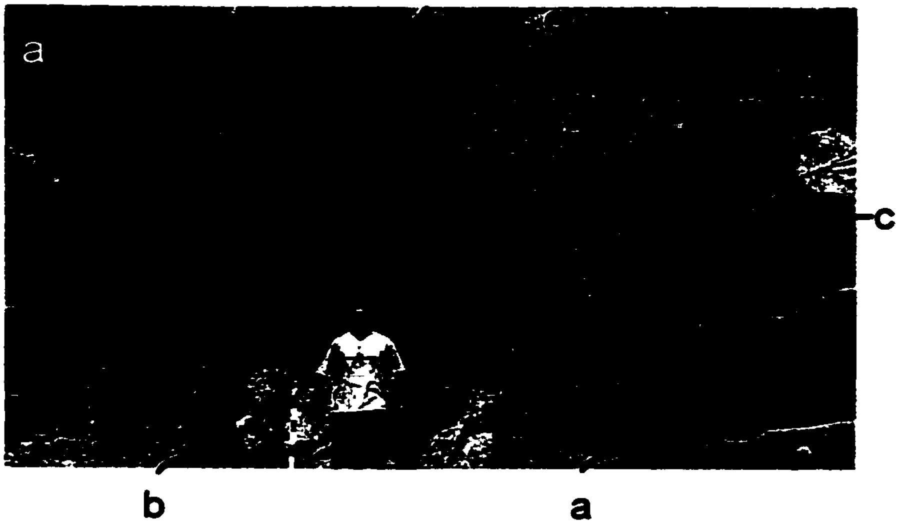

St/Sr
Sd
1 m

Fig. 4.14. Diffusely graded/massive medium sand infilling a steep-walled scour. (a) Note details of scour margin and infill, a) local steep angle of scour margin, b) pebbles near base of fill, c) internal scour margin, d) variable continuity of graded bedding, e) abruptly overlying trough cross-bedded sand of similar grain size but with surface veneer of silty sand. (b) Continuation of scour to left of photo a, a) locally overhanging scour margin, b) apparently massive sand lateral to diffusely-graded sand, c) localized disruption of strata beneath scour margin suggesting local slumping. The scale is 1 m.


a
c
b
a

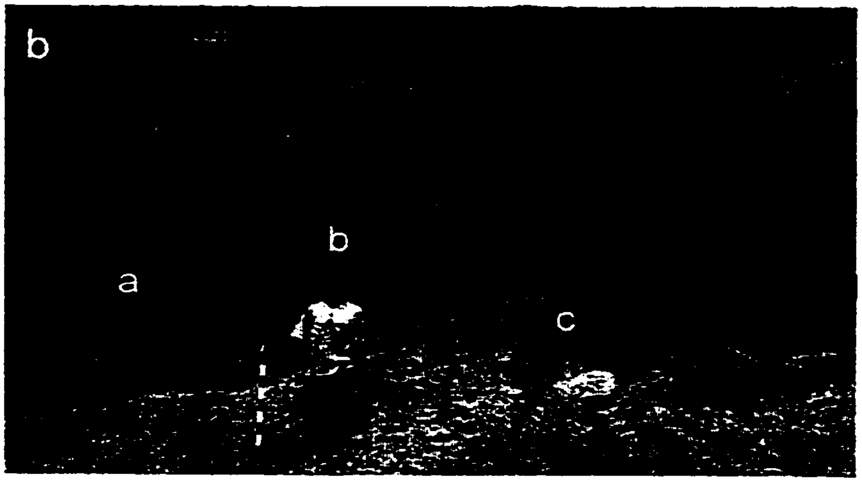

b
a
b
c

Fig. 4.15. Photo mosaic of section M-5. Section is dominated by planar cross-stratified sand and small-scale cross-laminated sand, with minor trough cross-bedded sand. Note change in angle of dip from nearly horizontal to ~15-20°. Dominant facies association is gently-inclined bedsets. Person for scale is 1.8 m.

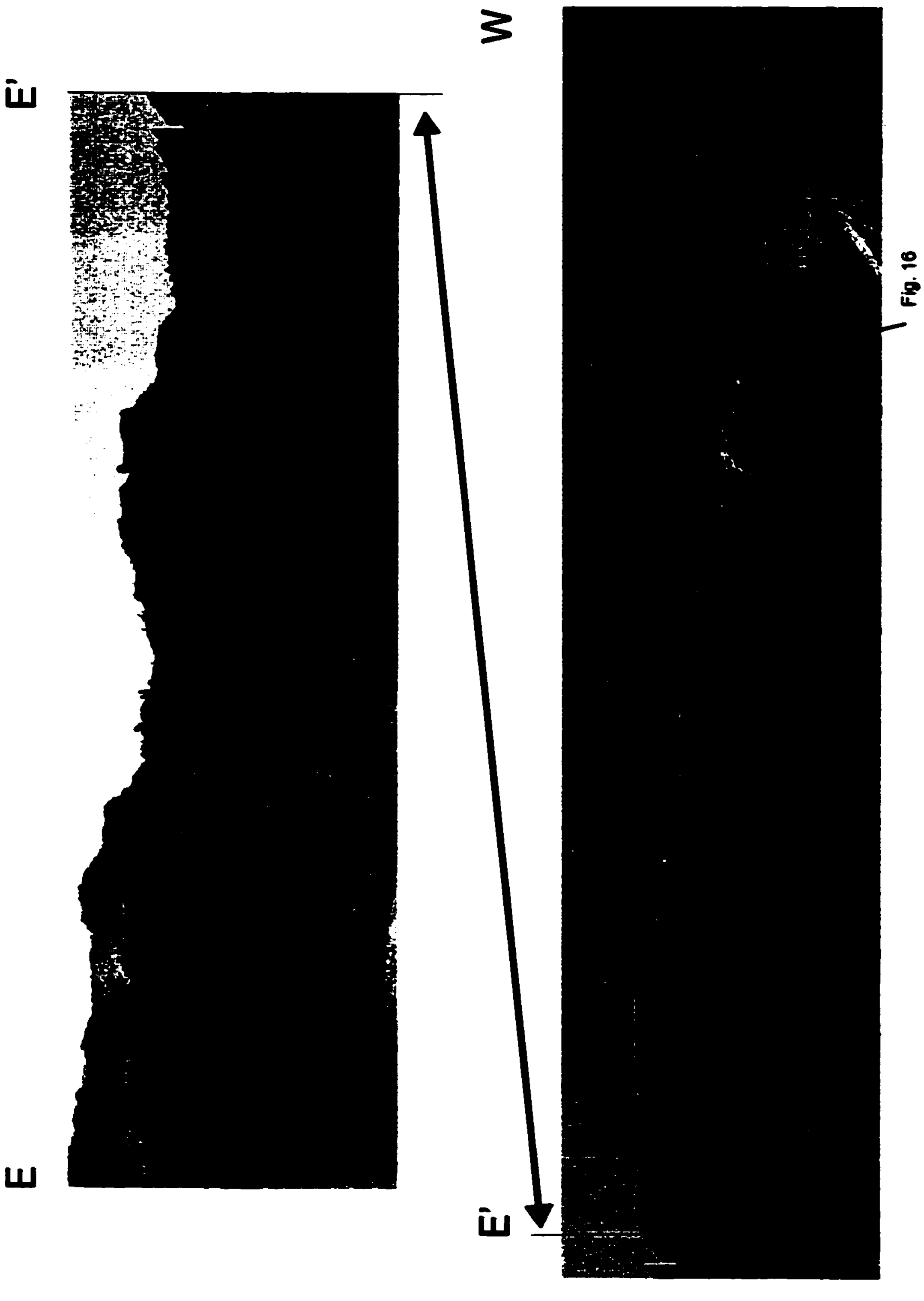

E'
E
W
E'
Fig. 16

Fig. 4.16. Climbing small-scale cross-laminated fine sand (Sr), commonly referred to as ripple drift cross-lamination. Paleoflow is from left to right. Note false bedding produced by silt-rich layers suggestive of large cross-strata. Pencil scale is 13.7 cm long.

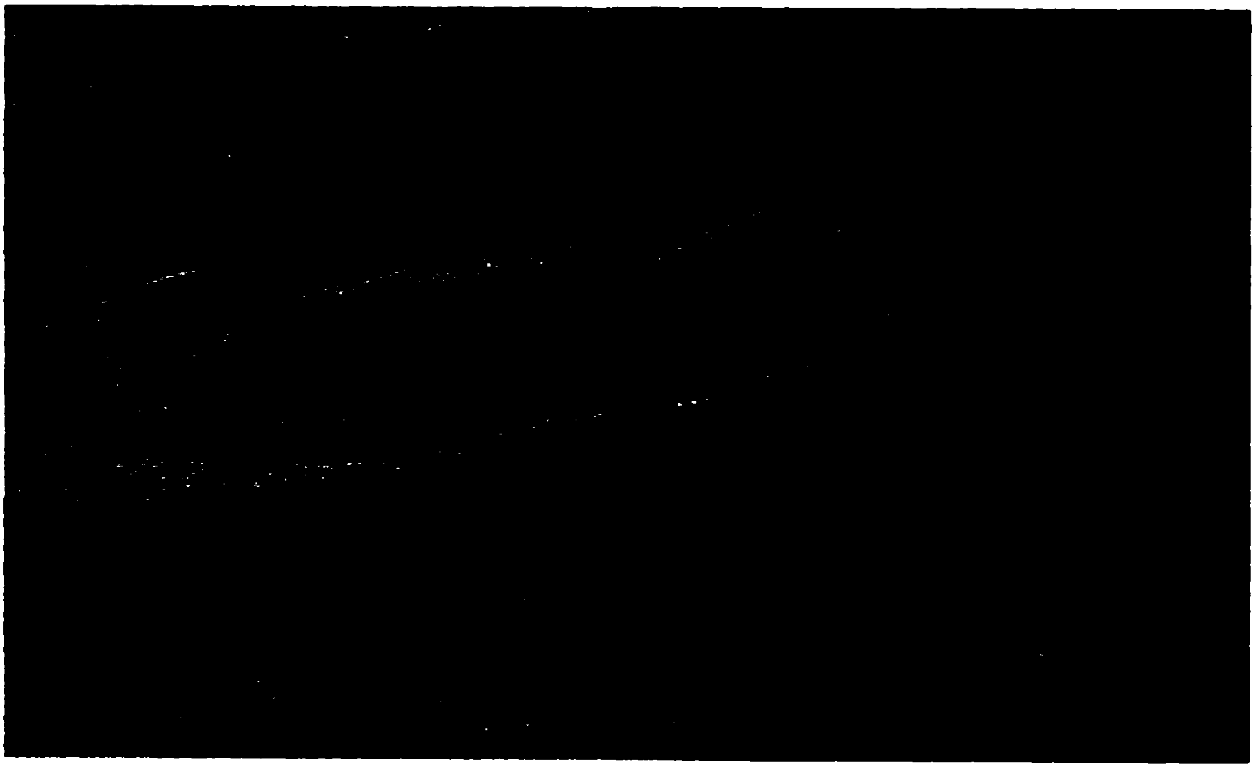


Fig. 4.17. Fence diagram of three sections and relationship of facies associations showing both downflow and vertical transitions. Inset figure is from Figure 2, shown upside down because sections are viewed from the north.

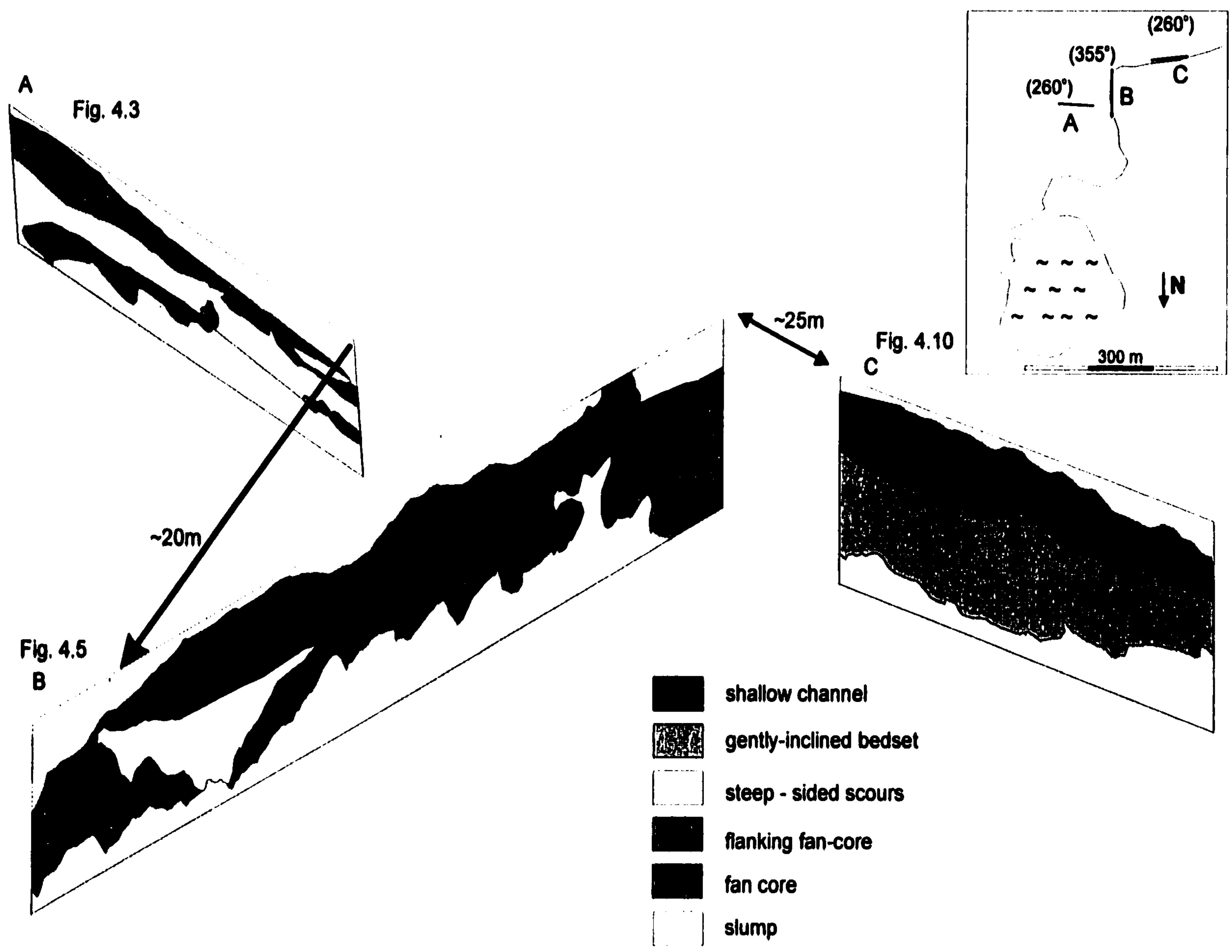
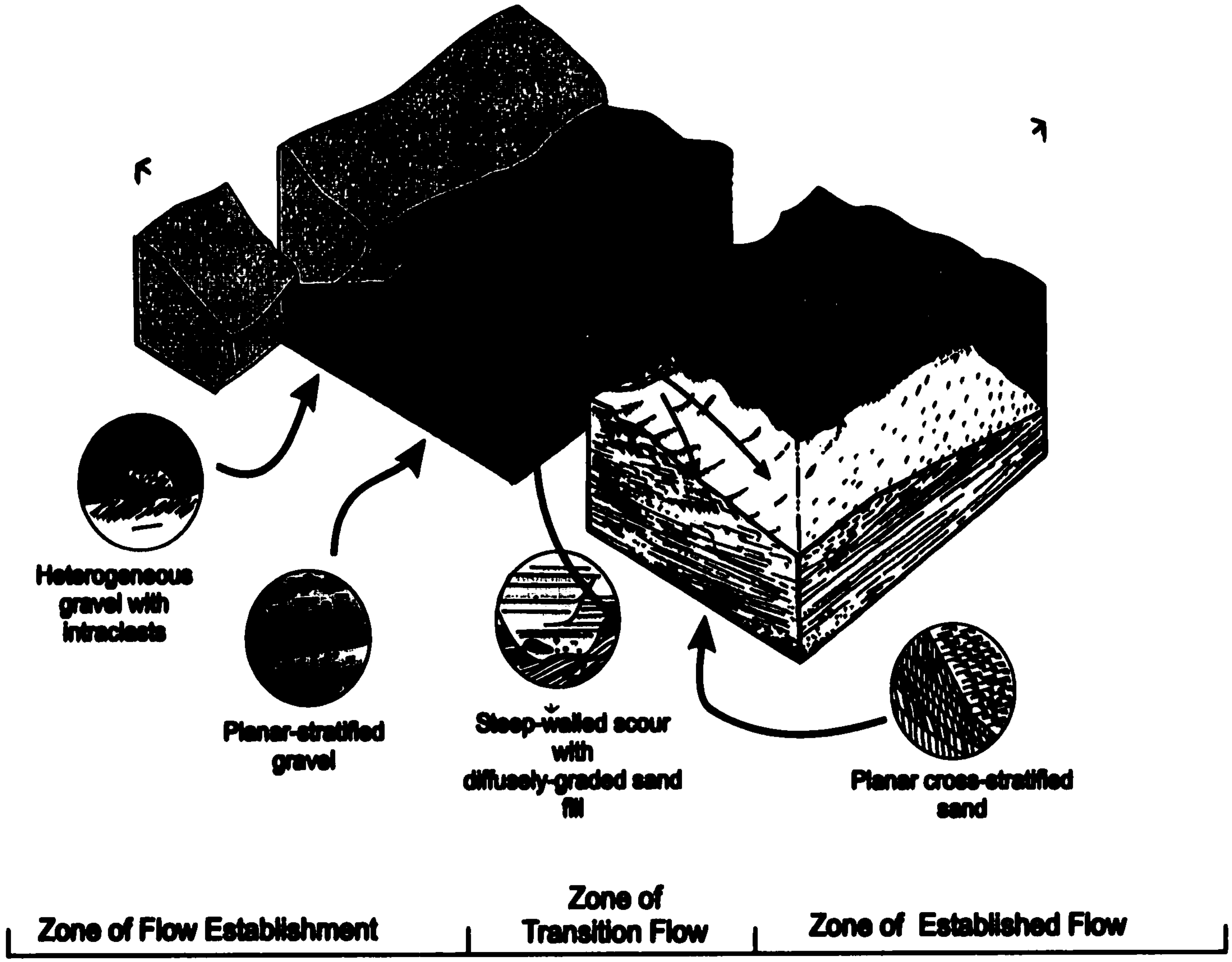

A
Fig. 4.3
~20m
Fig. 4.5
B
~25m
Fig. 4.10
C
(260°)
(355°)
(260°)
A
B
C
N
300 m
shallow channel
gently-inclined bedset
steep - sided scours
flanking fan-core
fan core
slump

Fig. 4.18. Schematic block diagram of proximal depositional fan model showing subaqueous fan lithofacies and lithofacies associations.



Heterogeneous gravel with intraclasts
Planar-stratified gravel
Steep-walled scour with diffusely-graded sand fill
Planar cross-stratified sand
Zone of Flow Establishment
Zone of Transition Flow
Zone of Established Flow

Table 4.1: Summary of selected studies of subaqueous fans in different environments. There is some overlap between categories; however, references are only assigned to one class.

Class	References
Valley / bedrock control	(Brodzikowski and Van Loon 1991; Spooner and Dalrymple 1993)
Continental, Isostatic	(Diemer 1988; Fyfe 1990; Gorrell and Shaw 1991; Sharpe and Cowan 1990; Sharpe et al. 1992)
Continental, Ice confined/ topographic	(Barnett et al. 1998; Henderson 1988; Paterson and Cheel 1997; Warren and Ashley 1994)
Esker / fan	(Banerjee and McDonald 1975; Brennand and Shaw 1994)
Subglacial cavity	(Brennand 1994; Dardis and Hanvey 1994; McCabe and O'Cofaigh 1994)
Glacimarine	(Burbidge and Rust 1988; Hunter et al. 1996; Lonne 1995; Plink-Bjorklund and Ronnert 1999; Powell 1981; Powell 1983; Powell 1990; Rust 1977; Rust 1988; Rust and Romanelli 1975; Sharpe 1988)

Table 4.2: Summary of subaqueous fan lithofacies

Facies	Facies code	Type Section	Character	Thick-ness (m)	Interpretation
cross -bedded gravel	Gt	M-2 M-1	sharp based, trough cross-bedded and planar cross-bedded, sandy pebble to pebble gravel, rare faults	3	2 and 3-D dunes and scour fill within the ZFE
planar stratified gravel	Gh	M-3	sharp-based, graded-beds, heterogeneous to well-sorted, open framework to framework	2	from bed-load sheets under critical flow conditions within the ZFE
heterogeneous gravel	Gd	M-2	normal and reverse graded, clast clustering, sand intraclasts	2-3	hyperconcentrated dispersions, within the ZFE and ZFT, from axial flow and beneath a hydraulic jump
cross -bedded sand	Sp, St	M-3 M-4	sharp-based, planar-tabular (Sp) and trough (St) cross-stratified, medium to coarse sand, minor pebbly sand, rare convolute bedding	5-7	by 2-D and 3-D dunes, within the ZFE and ZFT and axial ZEF,
quasi - planar stratified sand	Sh	M-3	sharp-based, minor undulation, horizontal to low-angle dip	3-5	beneath in-phase waves of an undular hydraulic jump and upper-flow-regime conditions, within ZFT
diffusely graded sand	Sd	M-3	sharp-based, irregular scours, massive to bedded, amalgamated beds, sand intraclasts, water escape structures	3	from hyperconcentrated dispersions beneath a hydraulic jump, within ZFT,
small scale cross-laminated fine sand-silt	Sr	M-4 M-5	sharp-based, stoss-erosional to stoss-depositional, minor micro-faults	5-8	under depletive flow and combined suspension bed-load transport, within the ZEF
silt - clay	F	M-4	lower contact sharp and graded, fining upward, micro-laminae to beds < 5 cm thick,	1-3	basinal underflow and suspension sedimentation

Table 4.3: Summary table of facies associations.

Facies Association	Width / length m	Thickness m	Facies Association (For codes refer to Table 2 and text)	Interpretation
fan-core	30/ >20	7	facies Gt, Gh, Gd	ZFE of jet, ice-proximal, deposition from highest energy flow,
flanking fan-core	10/?	~5	facies Sh, minor Sd St	jet margin within ZFE and ZFT
steep-sided scours	10 /?	4	incised into facies Sh filled with facies Sd	hydraulic jump within ZFT, deposition from hyperconcentrated dispersions
Gently inclined bedsets	?/ >150	6-8	predominantly facies Sp and Sr	accretionary fan foreset in the ZEF
shallow channels	>30/ >300	2-3	incised into facies St filled with St and Sr, minor Sd	late stage and more distal braided channel system in ZEF

Chapter V

Sedimentology of the Western Oak Ridges Moraine, Humber River Watershed, Southern Ontario

5.1 Introduction

Moraines are the largest constructional landform in a glacial landscape, extending for 10s to 100s of km in length and reaching thicknesses in excess of 160 m (Barnett et al. 1998). Nevertheless, the sedimentology of continental Laurentide moraines is poorly understood, in part because most previous work has focused on diamicton and debris dominated moraines deposited by alpine glaciers (e.g. Boulton and Eyles 1979; Post and LaChappelle 1971) or continental moraines formed along the southern margin of the Laurentide ice-sheet in the north-central United States (e.g. Mickelson et al. 1983). For example, Mickelson et al. (1983) defined "a true end moraine" as being composed of till; an opinion perpetuated by Karrow and Occhietti (1989) for moraines in southern Ontario. Additionally, the term moraine is subjected to linguistic confusion; for example, in German the word means till (Schlüchter 1979). This misconception is gradually being resolved as detailed sedimentological studies are completed and moraine genesis re-evaluated (e.g. Barnett et al. 1998; Brennand and Shaw 1996; Fyfe 1990). A number of studies have demonstrated that large continental Laurentide moraines are dominated by glacifluvial-glacilacustrine deposits (e.g. Burbidge and Rust 1988; Hillaire-Marcel et al. 1981; Kaszycki and DiLabio 1986; Sharpe and Cowan 1990) and that some interlobate moraines may, in fact, be eskers (Brennand and Shaw 1996; Veillette 1986). Some moraines are predominantly controlled by ice-sheet dynamics (e.g. Hillaire-Marcel et al. 1981; Sharpe and Cowan 1990) with the morphology, thickness and sedimentology related to the regional event-stratigraphy, whereas others have been related to climatic variables or bed conditions (Nemec et al. 1999).

The need for more integrated studies of large polygenetic landforms is being increasingly recognized (Blair and McPherson 1994; Warren and Ashley 1994). For example, Warren and Ashley (1994) state specifically the need for integrated landscape and sedimentological studies to better understand eskers and moraines, two glacial deposits with an ice-contact origin. Using examples from Ireland, they noted that a purely descriptive and mainly geomorphological approach to interpreting ice-contact ridges has led to equivocal interpretations of moraine origin and deglacial history. The Oak Ridges Moraine provides a unique opportunity to study a large moraine integrating both geomorphology and sedimentology (Fig. 5.1). Encroaching urbanization has resulted in the development of extensive aggregate operations and subsurface data sets, which in turn permit the moraine to be studied from three perspectives: terrain analysis, sedimentology, and subsurface mapping. In this study the synthesis of these data have permitted the development of a depositional model for moraine formation that consists of three principal stages: tunnel channel fill, low-energy glacilacustrine sedimentation, and subaqueous fan sedimentation. Each of these depositional stages can be correlated with elements of the regional event stratigraphy. The moraine is interpreted to have formed in response to regional subglacial meltwater discharge in a subglacial and/or subaerial glacilacustrine setting. Sedimentation was from both large magnitude catastrophic subglacial discharge and low energy annual glacilacustrine processes. The moraine is thus related to regional deglacial ice-sheet dynamics rather than a single short-term climatic event.

5.2 Geologic Setting

At the Late Wisconsin maximum ~20,000 BP, the southern Laurentide ice margin extended across southern Ontario into north-central United States (Barnett 1992). Subsequent deglaciation occurred across southern Ontario between 20,000 and 13,000 BP. Control on the ice-margin configuration and flow paths was largely influenced by bedrock topography in the Great Lake basins with ice lobes developing in each. An ice-margin re-entrant formed between ice masses centred in Lake Ontario-Erie and Lake Huron. Within this interlobate region atop the Niagara Cuesta a number of discontinuous moraines were deposited, including: the Dorchester, Waterloo, and Orangeville moraines (Chapman and Putnam 1984). Additionally, ice flowing out of each lake basin deposited a series of terminal moraines. In the case of the Lake Ontario lobe, deposition of terminal moraines was influenced by topographic control of the Niagara Escarpment.

The traditional deglacial chronology of southern Ontario has been based largely on till sheet stratigraphy, ice-marginal glacilacustrine deposits and moraine relationships (Karrow 1974) but has provided little insight into the potential mechanisms that caused the ice-marginal configuration. More recently Shaw and Gilbert (1990) suggested that erosion and deposition associated with large jökulhlaup discharges (outbreak flood) were important in the Great Lakes region. Drumlins in the Peterborough drumlin field and beyond have been interpreted as deposits of regional subglacial meltwater sheet floods that may have extended from Georgian Bay east at least to the Kingston area (Shaw and Gilbert 1990). Alternatively, these drumlins have been interpreted to have been formed by streams of deforming till (Boyce and Eyles 1991). This drumlinized surface was subsequently dissected by tunnel channels that are best developed north of the Oak Ridges Moraine. The tunnel channels are oriented predominantly northeast-southwest are 10s km long, and are up to 4 km wide (Barnett 1990). Locally channels have surface relief up to 65 m, but in the subsurface exceed 160 m in depth and locally extend to bedrock

(Barnett 1990; Pugin et al. 1999). The continuation of tunnel channels beneath the Oak Ridges Moraine has been defined by borehole and seismic studies (Barnett et al. 1998; Pugin et al. 1999). The tunnel channels have been interpreted to be the product of erosion by episodic subglacial meltwater floods (Barnett 1990; Brennand and Shaw 1996). The drumlins and tunnel channels are interpreted to form part of a regional Late Wisconsin unconformity (Shaw and Gilbert 1990), which in turn has significant implications for understanding ice-sheet geometry and subglacial depositional environments. Common to both the episodic meltwater flood and deforming till stream interpretations for drumlin formation was a reduction in basal shear stress and consequently a southward surge of the ice-sheet. Furthermore, the episodic post-surge discharge of meltwater along tunnel channels would have input a large volume of meltwater into the Lake Ontario basin. This has led Shoemaker (1999) to postulate that under conditions of a thinned, post-surge ice-sheet, a subglacial lake would have formed in Lake Ontario.

The location, form and composition of the Oak Ridges Moraine may thus have been related to this succession of deglacial events during the Late Wisconsin. It is ~160 km long and extends from the Niagara Escarpment eastward to Trenton along the northern margin of the Lake Ontario basin. The moraine consists of four broad wedges of sediment connected by narrow ridges, or separated by short gaps (Chapman and Putnam 1984). The extent of the moraine has been defined by elevation (Hunter 1996), outcrop (White 1973), and by surface roughness (Gwyn and Cowan 1978). This has resulted in considerable confusion as to the areal extent of the moraine, which has been further complicated by subdivision of the moraine into different elements (e.g. Palgrave, White 1975). The moraine is up to 160 m thick and consists predominantly of gravel, sand, and silt with minor interbedded diamicton (e.g. Barnett et al. 1998; Gilbert 1997; Paterson and Cheel 1997; Russell et al. 1998a). Depending on stratigraphic position within the moraine succession, deposition occurred in an ice-bounded basin that may have been subglacial (Barnett et al. 1998), glacilacustrine (Chapman 1985) or a subaerial braid plain environment (e.g. Duckworth 1979). Deposition of the Oak Ridges Moraine has been commonly correlated with climatic warming during the Mackinaw interstadial between 13.3 -15 ka BP (e.g. Duckworth 1979; White 1975). By contrast, Barnett et al. (1998) attributed the formation of the moraine to changes in meltwater flux related to ice-sheet dynamics and local topographic control. In this model four major depositional stages have been identified: i) subglacial, ii) subaqueous fan, iii) subaqueous fan to delta, and iv) ice-marginal sedimentation (Barnett et al. 1998). Stratigraphic evidence of each of these stages occurs in one or more of the four moraine segments, although the most complete sequence occurs in the western end of the moraine, an area that includes the current study area. Subsequently, Shoemaker (1999) suggested that the moraine may have been formed at the grounding-line of a subglacial lake that was similar to or larger than modern-day Lake Ontario.

The study area is located in the Humber River watershed in the western part of the Oak Ridges Moraine. This area is adjacent to the Niagara Escarpment (Fig. 5.1) and overlies the Laurentian Channel, which is a prominent buried bedrock valley that extends from Georgian Bay to Lake Ontario (Brennand et al. 1997; Eyles et al. 1985). The stratigraphy of the area has been established from outcrop and borehole studies (Eyles et al. 1985; White 1975). The oldest Quaternary deposits are the Illinoian York Till (White 1975). These deposits are unconformably overlain by the Sangamonian Don Formation, an organic-rich interglacial deposit (White 1975). These deposits are then overlain by the sand and silt of the Lower Wisconsin Scarborough Formation (Kelly 1994; White 1975). In the cores from drillhole DH-Nob detrital rich Scarborough sand is overlain by silt-clay rhythmites of the Thorncliffe Formation (Russell and Pullan 1998). These Lower Wisconsin strata are overlain by the Upper Wisconsin Newmarket Till that forms a regional till sheet (Sharpe et al. 1999b). Unconformably overlying the Newmarket Till are strata of the Oak Ridges Moraine and the Halton Till (Russell and White 1997; White 1975). This area has been mapped at 1:50 000 by the Ontario Geological Survey (OGS) (Cowan 1976; Gwyn and White 1973; White 1975) and more recently in collaboration with the Geological Survey of Canada (GSC) (Sharpe et al. 1997). For the most part these maps are based on surficial sediment at a depth of ~1 m and consequently provide limited insight into the stratigraphy of the moraine. Since 1993, however, a basin analysis study has been underway at the GSC with the principal objective of mapping the key stratal units defined in the regional stratigraphic model (Sharpe et al. 1999a). This work has focused on the collection of new sedimentological and seismic data and the integration of archival data (Pugin et al. 1999; Pullan et al. 1994; Russell et al. 1996; Sharpe et al. 1996).

5.3 Methodology

This study focused on better understanding the lithofacies and lithofacies-association of the moraine from outcrop and borehole studies within the framework of a basin analysis project (e.g. Potter and Pettijohn 1963). Stratal terminology and relative thickness descriptions follow the classification of McKee and Weir (1953). Archival or supplemental data used in this study include: a digital elevation model (Kenny et al. 1999), shallow field mapping sites (Russell and White 1997), glacial geology mapping (Barnett and Gwyn 1997; Russell and White 1997; Russell and Dumas 1997; Sharpe and Barnett 1997), archival borehole data from the MOE water well database and geotechnical records (e.g. Russell et al. 1996), seismic profiles (Pugin et al. 1999) and drillhole cores (Fig. 5.2). Field observations have been synthesized by lithofacies based on texture and sedimentary structures (e.g. Eyles et al. 1983; Miall 1984; Miall 1996). Archival data have been coded into five simple sediment categories: diamicton, gravel, sand, silt or clay (Russell et al. 1998b). Lithofacies-association studies are based on photo-mosaics, borehole and seismic profile correlations (e.g. Miall 1996). These data were complemented with paleoflow measurements and laboratory grain-size measurements. Grain-size data were analyzed by the sedimentological laboratory at the Geological Survey of Canada (Ottawa) using standard sieve techniques for sediment >4phi and a Brinkman particle size analyzer for finer sediment (Lindsay and Shilts 1995). Additional data were obtained from reports of the Interim Waste Authority (Fenco-MacLaren 1994; Golder and Associates

1994). These data were collected according to the ASTM (1994) procedure using sieves and a hydrometer. No replicate analyses were completed on the <4 phi fraction to in order to quantify differences between data obtained from the Brinkman particle size analyzer or hydrometer tests (see Syvitski 1991). Paleocurrent measurements are presented as rose diagrams following the methodology outlined by Nemec (1988).

5.4 Geomorphology and Surficial Geology

Traditionally sedimentologists describe strata from bottom to top. Because moraines are defined on the basis of geomorphology, it is appropriate to first describe the geomorphology and exposed surficial geology. A digital elevation model (DEM) and six south-north topographic profiles ~25 km long and spaced 4.5-8 km apart are used to characterize the general moraine morphology (Fig. 5.3-5.4). The area indicated on cross-sections to be Oak Ridges Moraine is based on geological mapping, surface roughness, and elevation.

On the digital elevation model four main physiographic elements are identified by topography, surface roughness, and orientation of roughness elements: i) the Niagara Escarpment and cuesta to the west, ii) drumlinized Newmarket Till and steep-walled tunnel channels to the north, iii) Oak Ridges Moraine, and iv) smooth Halton Till plain to the south (Fig. 5.3). The Oak Ridges Moraine forms an east-west ridge 3-24 km wide that has a maximum elevation of ~420 m asl where it onlaps the Niagara Escarpment, and relief of ~100-150 m above the Halton Till plain to the south (Fig. 5.4). Adjacent to the Niagara Escarpment the Oak Ridges Moraine consists of disorganized terrain and most of the exposed sediment is sand (Fig. 5.4). This area consists of closed, internally drained depressions and fluvially dissected topography. Farther to the east the moraine is extensively incised where sand crops out (Fig. 5.4b), but is significantly less dissected where Halton Till is exposed (Fig. 5.4d,e). Using the 100 m grid cell of the DEM, the steepest moraine slopes are ~14°, with ~97% <5°, steep slopes commonly occur where Halton Till crops-out along stream-valley sides. The moraine has a generally asymmetric cross-section with a steeper, ~3° northern flank that is commonly <5 km long and a gentler southern flank that extends for 10-20 km.

The Oak Ridges Moraine rises above the surrounding landscape and has traditionally been interpreted to have been deposited in an ice-supported setting (Chapman 1985; Chapman and Putnam 1984; Duckworth 1979). The disorganized terrain has been interpreted to be hummocky terrain formed by melting of buried ice (White 1975). The depositional form of the moraine has been strongly modified by post-glacial fluvial dissection during the Holocene. Direct ice-contact support of the ridge was probably not present because the slopes are invariably <14°. For comparison, unsupported depositional slopes on debris and water-lain sand and gravel alluvial fans are commonly <22° (e.g. Blair and McPherson 1994). Nevertheless, sediment landform relationships indicate that the northern side of the moraine was most probably ice-supported as underlying Newmarket Till crops out along the moraine margin.

5.5 Basal Contact Geometry and Moraine Thickness

Integration of the archival data and field data permitted the general morphology of the base of the moraine and its thickness to be mapped (Fig. 5.5). The lower surface is generally irregular with a number of anomalously deepened basins and channels (Fig. 5.5a). These localized deepened areas are up to 50-70 m below adjacent terrain, up to 1-2 km wide and 5-7 km long. The best examples occur south of DH-Nob and southeast of Mount Wolfe where elongate north-south depressions occur. Other deepened areas generally lack adequate length to define any preferred orientation. One topographically low area trends northeast from Kleinburg to Bolton, and coincides with the modern Humber River valley. The moraine is up to 150 m thick but is <50 m thick over three-quarters of the area mapped as moraine sediment (Fig. 5.5b). Areas of thick moraine sediment are associated with the localized deepened regions rather than with topographically high areas (Fig. 5.5). Moraine thickness decreases dramatically where it onlaps the Niagara Escarpment and also south of the Humber River where bedrock or lower deposits crop-out along river valleys.

The ORM overlies an irregular surface that is correlated with an exposed regional unconformity formed of tunnel channels and the upper surface of drumlinized Newmarket Till to the north of the moraine (e.g. Sharpe et al. 1999b; Shaw and Gilbert 1990). This unconformity has also been interpreted from subsurface geophysical logs north and south of the study area (Eyles et al. 1985). Based on the southward continuity from exposed tunnel channels to the north of the moraine (Fig. 5.1, Russell and Dumas 1997), character of strata in continuous core (chapter 3), seismic data (Pugin et al. 1999), and trends in isolated locally deepened areas (Fig. 5.5a) three buried tunnel channels are recognized: King City, Holland Marsh and Mount Wolfe (Fig. 5.5c). A north-south cross-section of the buried part of the Holland Marsh Channel shows the variability in channel depth, the general southward shallowing trend and southward rise in the lower surface. This profile does not follow the channel thalweg and may thus fail to accurately represent the streamwise morphology of the channel. Where the modern surface topography rises, the lower stratigraphic surface can have a parallel increase in elevation(Fig. 5.5d). Furthermore, in the southern part of the moraine, thickness is closely controlled by modern river valleys and the location of shallow data points (Fig. 5.5e).This sympathetic relationship between the topographic surface and lower stratigraphic surface in the vicinity of stream valleys is an interpolative artifact that was produced as a result of the small number of data points that penetrate to the underlying stratigraphic unit (Logan, inpress). The presence of these artifacts in the lower stratigraphic surface constrain any interpretations about changes in the surface morphology or moraine thickness, except in the vicinity of high quality data. Nevertheless, the surface does provide a first approximation of general trends in the buried unconformity and of moraine thickness.

5.6 Sedimentology

The Oak Ridges Moraine is composed predominantly of silt, sand and gravel with only minor clay and diamicton (Barnett et al. 1998; Duckworth 1979; Gilbert 1997; Paterson 1995; Russell et al. 1998a; White 1975). Detailed facies descriptions of gravel and sand in tunnel channel fill and subaqueous fan deposits were presented in chapters 3 and 4. In this chapter, gravel and medium-coarse sand facies are only described as new data warrant, and are summarized in tables 5.1 and 5.2. Facies descriptions presented here focus on diamicton, fine sand, silt and clay. A diamicton and glacilacustrine-glacifluvial unit, here collectively termed the Halton Till complex stratigraphically overlies the Oak Ridges Moraine and therefore is not discussed (see Russell and Arnott 1997), although it has been suggested that these deposits may represent a late-stage episode of moraine sedimentation (Barnett et al. 1998).

5.6.1 Lithofacies composition

Sediment from 600 m of core from 11 sites (Fig. 5.6, 5.7, 5.8) and ~150 m of outcrop were described in bed-by-bed detail. These strata have been grouped into ten lithofacies of which six have been described previously (Table 5.1, 5.2). Graded fine-sand to silt facies forms ~56 % of the stratigraphy observed in core and the small-scale cross-laminated facies forms ~17 % (Fig. 5.9). Locally other lithofacies are proportionately more common; for example, DH-Nob consists of ~32 % gravel facies, DH-V-158 intercepts ~30 % sand facies and DH-P-14 consists of ~32 % small-scale cross-laminated facies. Five sections measured from photo mosaics of the Gormley site illustrate the rapid facies variability present locally (Fig. 5.8c).

In order to integrate archival and field data all the data, were coded to a simplified sediment coding system consisting of five units (diamicton, gravel, sand, silt, clay). MOE water well data are considered unreliable for mapping lithological trends within the moraine. For example clay units up to 30 m thick are described in water well descriptions whereas data from nearby field study locations indicate that clay layers >4 cm thick are rare. The variability in apparent moraine composition as described by various archival datasets is plotted in Fig 5.9. For example, the clay facies forms generally <2 % of moraine strata in the current study (Fig. 5.9a) but is commonly >5 % for the archival data and can be >30 % in the MOE water well data (Fig. 5.9b). This apparent disparity can be corrected by merging silt and clay facies and recognizing that most of this unit corresponds with the graded fine-sand to silt lithofacies. For other facies, such as diamicton, it is more difficult to reconcile the reason for these significant discrepancies. The largest group of archival data and the dataset with the most extensive depth distribution throughout the moraine is provided by the MOE water well data. The sediment descriptions in this dataset are suspect particularly for fine-sand, silt, clay, and diamicton (e.g. Russell et al. 1998b). For this reason, only the distribution of gravel is mapped across the moraine (Sect. 5.7).

5.6.2 Diamicton

Only minor diamicton occurs within the Oak Ridges Moraine. In this study only 8 beds with a combined thickness <1 m were observed (Fig. 5.6). Beds are sharp based and overlie either small-scale cross-laminated or normal-graded fine sand-silt. It is important to note that significantly more diamicton is reported in archival descriptions, for example up to ~18 % in the MOE water wells (Fig. 5.9). Diamicton consists of fine sand-silt matrix with 1-3 % subangular carbonate gravel size clasts. Beds are 1 to 45 cm thick and massive or normally graded.

Interpretation: On the basis of abrupt lower contacts, texture, stratal thickness, and the isolated occurrence within a thick succession of small-scale cross-laminated sand and graded silt, diamicton strata are interpreted to be deposits of small debris flows (Mulder and Cochonat 1996) or ice rafted debris (Gilbert 1990). In core it is difficult, if not impossible, to definitively interpret the emplacement mechanism, because settling of ice-rafted debris within the water column can sort sediment and produce textures similar to debris flow deposits (e.g. Gilbert 1990). At least one bed is interpreted to be a winter basinal deposit, because it forms an interlaminae within a clay bed. The discrepancy in the amount of diamicton observed in core (<0.2%) compared with archival data (~18%) is considered to be the result of the wash-boring technique (MOE water well data) as compared to continuous coring (this study), and recoding of original sediment descriptions from archival data (e.g. Russell et al. 1998b). Moreover, other field studies have found little diamicton within the Oak Ridges Moraine (e.g. Barnett et al. 1998; Duckworth 1979; Gilbert 1997; Paterson and Cheel 1997; White 1975). Consequently, the abundant diamicton reported from archival data is considered to be an unrepresentative estimate of diamicton in the moraine.

5.6.3 Gravel lithofacies

Although gravel crops-out at the Gormley, BS&G, an old quarry in the Sandhill area, and is present in core DH-Nob it is generally uncommon in the study area. In contrast, it is commonly reported in MOE water well records comprising up to ~7 % of the total thickness and forming units up to ~30 m thick. In outcrop, basal contacts are sharp but due to poor core recovery in gravel strata are not observed in core. Gravel strata consist of open-framework pebble to heterogeneous sandy gravel with sand intraclasts. Cobble-size clasts are uncommon and boulder clasts are rare. Sedimentary structures consist of trough and planar cross-stratification, planar-stratification, reverse and normal grading, and massive strata (Fig. 5.10). Beds are generally <30 cm thick; however, some massive beds are up to 2 m thick. Deformation is not common but consists locally of low-angle 10-15° dipping beds and high-angle, 70-80° normal faults with minor 10-15 cm offset. Paleocurrent data from clast fabrics and cross-bedding at the Gormley site indicate paleoflow toward the west to west-northwest.

Interpretation: On the basis of texture, sedimentary structures and intraclasts three gravel facies have been interpreted (Table 5.1; see chapter 3 and 4 for detailed descriptions and interpretations). These facies record deposition from subcritical and supercritical fluid flows and hyperconcentrated dispersions. Locally deposits were emplaced under antidune flow conditions. Elsewhere deposition occurred rapidly during flow expansion and flow transition from hyperconcentrated to fluidal flows.

5.6.4 Medium to coarse sand lithofacies

Well-sorted medium to coarse sand and less commonly pebbly sand crop out extensively at two sites north of Bolton and less commonly in Humber River sections and cores (Fig. 5.11). The facies is 0.2 to 25 m thick, and where the lower contact is observed abruptly overlies the gravel or small-scale cross-laminated facies. Sedimentary structures consist of trough and planar cross-stratification, low-angle cross-stratification, planar-bedding, diffuse-grading and massive strata (Fig. 5.11b). Deformation is minor, being restricted to small-scale normal and reverse faults with 1-5 cm displacements and convolute bedding. Locally large convolute structures, 2-3 m wide and 1-2 m high occur where strata are overlain by Halton Till. Minor organic detritus is present in both outcrop and drillcore.

Grain-size was analyzed in 19 samples collected from the cross-stratified and diffusely-graded / massive sand facies at the Gormley and DH-V-158 sites (Fig. 5.12a,b). Three samples from cross-bedded sand are moderately to moderately-well sorted, fine-skewed, leptokurtic upper-fine sand. By contrast, the 16 samples from the diffusely-graded/massive sand are poorly sorted to moderately well-sorted, near symmetric to fine-skewed, mesokurtic to leptokurtic upper-silt to medium-sand. Clay content in both subfacies is <0.5 %, being slightly lower in the cross-bedded sand.

Paleoflow directions were measured at two locations situated on opposite sides of the moraine crest and ~8-10 km apart (Fig. 5.1). At the northern Gormley site, paleocurrents have a bimodal distribution with modes toward the southwest and west-northwest (Fig. 5.8, 3.1). To the south at the BS&G site, paleoflow directions are generally toward the west.

Interpretation: On the basis of texture, sedimentary structures and abundance of intraclasts, three sand facies have been identified and interpreted to have been deposited by fully turbulent fluid flows or hyperconcentrated dispersions (Table 5.2; see chapter 3 and 4 for detailed descriptions and interpretations). Low-angle cross-stratified sand was deposited under high energy, antidune conditions formed in undular hydraulic jumps (chapter 4). Diffusely-graded / massive sand was deposited from hyperconcentrated flows emerging from confined subglacial conduits and deposited a short distance downflow of hydraulic jumps (channel 2 and 3). Cross-stratified sand was deposited from bedload transport. Climbing cross-stratification locally record high rates of suspension sedimentation (Ashley et al. 1982; Daub 1996). Locally both vertical and lateral facies transitions from diffusely-graded to cross-stratified sand suggest deposition from nonuniform, unsteady flows.

5.6.5 Small scale cross-laminated sand

Small-scale, cross-laminated sand crops out extensively across the area and is commonly observed in core. Locally it abruptly overlies gravel, sand, normally graded fine-sand to silt or clay. The facies is up to 7 m thick and consists of silt, fine sand and rare clay-silt laminae (Fig. 5.13a). The small-scale cross-laminae form repetitive successions of sharp-based, upward-fining stoss-erosional or stoss-depositional sets <7 cm thick, with cosets up to 3 m thick (Fig. 5.13c). Within individual cosets the angle of climb increases upward from 40° to 85°. In other cases the angle of climb changes little but the amplitude and grain size decrease upward (Fig. 5.13d). Clay interlaminae, intraformational clay clasts, and shallow scours are uncommon. Deformation consists of both micro-normal faults with 1-2 mm displacement and larger faults with <20 cm displacement. Grain size was analysed for seven samples from DH-V-158 and are moderately well sorted to poorly sorted, near symmetric to strongly fine skewed and meso kurtic to leptokurtic silt (Fig. 5.13c).

A total of 267 paleoflow measurements were made at three sites separated by ~1-6 km along the Humber River (Fig. 5.8). Although local measurements show a preferred orientation, overall the paleoflow directions are highly variable (Fig. 5.9b). At three sites Paleoflows in the stratigraphically lowest strata are toward the west, northwest or north. In contrast measurements in the higher strata at sites 186 and PG are toward the southeast, and toward the northwest at site PG-II(Fig. 5.9b).

Interpretation: This facies is interpreted to have been deposited from quasi-continuous or discontinuous underflows (e.g. Ashley et al. 1982). Because of rapid deceleration flows became overloaded with suspended silt and fine sand resulting in combined traction and suspension sedimentation. This style of sedimentation is indicated by the moderate to poor sorting and fine skewness of the sand. Where thick cosets of rhythmic climbing stoss-depositional cross-stratification are observed deposition is interpreted to have been from a quasi-continuous flow (e.g. Shaw 1975). In a study of climbing current ripples Ashley et al. (1982) estimated aggradation rates of 5-15 cm h^{-1} as an upper limit for stoss-depositional strata. Consequently, similar successions that are up to 3 m thick were possibly deposited in as little as 1-3 days. Although the presence of scour surfaces would significantly lengthen this estimate, the consistent angle of climb within each set suggests relatively steady flow and depositional conditions. Where climb angles show greater variability and steepen with height above the base of the bed, it probably was deposited from a continuously waning flow. Normal faults with 1-2 mm offset and limited vertical continuity are interpreted to be syndepositional. Thinner less rhythmic successions of cross-laminae grading upward into normally graded fine-sand to silt record deposition from discontinuous underflows or turbidity currents (e.g. Gilbert 1975; Gustavson 1975). Individual sets of upward-fining, small-scale cross-laminated fine sand-silt to silt-clay laminae form T_{cd} and possibly T_{cde} turbidites (e.g. Walker 1992).

5.6.6 Normal-graded fine-sand to silt

This facies is up to 1 m thick and commonly overlies small-scale cross-laminated sand or clay. The lower contact is gradational with underlying small-scale cross-laminae and less commonly clay strata. The normal-graded fine-sand to silt couplets range in thickness from <0.1 cm planar-laminae to 4 cm thick beds that form cosets <90 cm thick (Fig. 5.14a,b,c). Small-scale cross-lamination is rare, and sets are generally <1-2 cm thick. Bioturbation is rare and restricted to 1-2 mm wide irregular oxidized horizontal traces on bedding planes in silt-rich sediment. Grain size was analysed in 18 samples. These samples are moderately well-sorted to moderately-sorted, coarse to strongly-fine skewed and playtykurtic to leptokurtic silt (Fig. 5.12d). Of these samples 50% consist of >80% silt. Clay content was generally low, being <1% in 70% of the samples. Deformation is generally limited to minor normal microfaults, convolute bedding, and dewatering structures.

Interpretation: The graded planar-strata may have been deposited by one of three mechanisms: upper-flow-regime plane-bed, migration of low-relief, long wavelength ripples, or suspension sedimentation. Based on the complete absence of any upward transition to small-scale cross-laminae observed in >100 fine sand strata deposition under upper-flow-regime plane-bed conditions is considered improbably. Beds 1-2 cm thick are interpreted to have been deposited by traction currents where the lower part of the strata was probably deposited by low-relief, long wavelength ripples (e.g. McBride et al. 1975). Laminated layers that are usually ~2-3 mm thick may have been deposited by small migrating bedforms or directly from suspension by weak underflow or turbidity currents (e.g. Weirich 1986). Depositional currents were probably generated by underflows or turbidity currents initiated during peak diurnal meltwater discharge that waxed and waned rhythmically as the daily meltwater discharge fluctuated (Gilbert 1997; Gustavson et al. 1975; Shaw 1975). The daily meltwater and sediment flux of underflows for a daily ablation regime is well documented by a number of studies in glacilacustrine settings (e.g. Gilbert 1975; Weirich 1986). Furthermore, a daily rhythm of increasing discharge up to an early afternoon peak followed by a subsequent decrease through the evening has been shown to cause pulsations in continuous underflows (Weirich 1986). Abrupt changes in the thickness or grain size within this facies may record episodic surge events with sediment concentrations 2-3 times or more compared to the continuous current (e.g. Weirich 1986). The common upward transition to thinner, finer graded strata suggests a long-term decrease in flow energy and sediment flux.

5.6.7 Silt - Clay

The silt-clay facies is best observed in core, but it has also been described from outcrop across the area. The lower contact gradationally overlies normally graded fine sand or silt (Fig. 5.14d). Of ~ 300 clay layers measured in core the maximum thickness is 19 cm, the average thickness is 2.5 cm and 94% of the layers are <6.0 cm thick. The facies is ubiquitously normal-graded with minor silt partings <0.01 cm thick. Bioturbation along bedding is made visible by disrupted clay-silt laminae in the lower parts of beds. For three samples the grain size measured at whole phi intervals indicates a moderately to poorly-sorted, fine to strongly-fine skewed, mesokurtic silt (Fig. 5.12). In these samples the clay content is <10 %. In a second grain size dataset that reports only the sand-silt-clay ranges, the clay content is 25-40 % (Fig. 5.6, 5.7).

Interpretation: The graded to massive, fine grained texture, general absence of silt-sand interbeds and graded transition from underlying sand-silt strata suggest that the facies was deposited under conditions of low-energy suspension sedimentation. The general absence of interbeds suggests deposition during a period of basin quiescence and an absence of density underflow or turbidity currents. Deposition is interpreted to have occurred following the end of the summer meltwater season and through the winter season (e.g. Banerjee 1973; Gilbert and Shaw 1981; Gustavson 1975). Rare fine-sand interbeds record late-season meltwater underflows or less probably deposition from slump-induced turbidity currents. The relatively low clay content of these strata, <10 % as compared to other varve studies in the region (Gilbert

1997), may be the result of the definition of the clay grain size used in this study or the analytical technique employed. In this study the ASTM definition of <2 microns has been used for the upper clay grain size limit (ASTM 1994) rather than the 3.9 micron size used in the Wentworth scale (e.g. Gilbert 1997). Additionally, comparing grain size results from different analytical techniques may give equivocal results. For example, the Brinkman particle size analyzer may under-sample the clay size compared to other methods (Lindsay and Shilts 1995), which in turn may explain the difference between data from this study and archival data that were analyzed using standard hydrometer techniques (ASTM 1994).

5.7 Gravel Distribution

The distribution of gravel was mapped using archival data, principally the MOE water well records (Fig. 5.2). Of the total thickness of Oak Ridges Moraine strata drilled, ~7% consists of gravel. The distribution of this gravel is strongly skewed with ~45 % directly overlying the lower boundary and a further ~20% occurring within 10 m of the lower contact. The remainder is distributed throughout the moraine succession. The distribution of gravel for the elevation range 100-400 m asl has been plotted on six maps with each map displaying five ten-metre intervals (Fig. 5. 15). These plots show that buried gravel occurs ~1-2 km south of DH-Nob and ~10 km farther south near the Humber River along tunnel channels. Surprisingly, <5 % of the gravel in the moraine strata occurs along the base of tunnel channels. Rather ~30 % occurs between 190-220 m and can be mapped from the southern part of the watershed northward across the central moraine region to approximately Nobleton (Fig. 5.15d,e). Secondary corridors of gravel are mapped east of King City (Fig. 5.15e) and along the southwestern margin (Fig. 5.15c). An additional area of gravel occurs along the buried Niagara Escarpment where it parallels the escarpment trend. Gravel in this area is interpreted to be part of the Gibraltar and Paris moraines that in the subsurface would be difficult to differentiate from strata of the Oak Ridges Moraine. Elsewhere in the study area, for instance at the Gormley site, gravel has been observed high in the moraine succession (Fig. 5.8c, chapter 4) and occurs immediately below fine grained lacustrine sediment. In several hundred shallow <4 m deep sections across the area (Fig. 5.2) gravel is generally absent in outcrop.

The abundance of gravel above the regional unconformity in archival data correlates well with observed gravel deposits in continuous core (Fig. 5.8a) and interpreted from seismic data (c.f. Pugin et al. 1999). At DH-Nob the gravel overlies the unconformity at the top of the Thorncliffe Formation whereas in seismic profiles to the east, gravel overlies drumlinized Newmarket Till . Gravel overlying the regional unconformity atop interfluves is interpreted to have been deposited by the waning stage of regional sheets floods prior to or during initial tunnel channel incision. Similar gravel deposits exposed atop the interflueve between the Holland Marsh and Alliston tunnel channels are interpreted as being similar deposits(Fig. 5.1). Across this interflueve, 2-5 m thick, fining-upward gravel and sand abruptly overlies Newmarket Till (Russell and Dumas 1997).

5.8 Facies Associations

The Oak Ridges Moraine has been interpreted to be a polygenetic feature consisting of tunnel channel fills (Barnett et al. 1998), basin sediment (Gilbert 1997), subaqueous fan (Paterson and Cheel 1997)and deltaic deposits (Duckworth 1979). In earlier chapters two of these facies associations have been described: buried tunnel channels (chapter 3) and a subaqueous fan (chapter 4). This section summarizes the lithofacies-associations of these two settings in addition to a previously undescribed basin rhythmite association.

5.8.1 Tunnel Channel Fill

Beneath the Oak Ridges Moraine tunnel channels are up to 20 km long, <5 km wide and <170 m deep (Pugin et al. 1999). Channels are asymmetric in cross-section and longitudinally are locally overdeepened (Fig. 5.5a). Channels are infilled by four lithofacies that have a distinctive spatial organization. Gravel occurs along the elevated tunnel channel margin at DH-Nob and forms a broad tabular deposit up to 0.5 km wide and ~ 20 m thick. This unit is interpreted to consist of >4 m high mesoforms and low-relief gravel sheet deposits (chapter 3) emplaced by flows from the north. Along the tunnel channel axis only limited gravel deposits have been identified using water well records. Unfortunately, no textural information is available from this dataset on these buried deposits. Seismic data to the east from the Scugog Lake area record 25 m high clinoforms interpreted to have been formed by large dunes (Sharpe et al. 1999c). Deeply buried gravels that are 15-20 m thick in water well records may be similar but thinner deposits. The most common types of strata in the channel axis have a chaotic seismic facies signature, west of the DH-Nob site tese strata are 80-100 m thick and extend laterally across the channel for 3-4 km (Pugin et al. 1999). This seismic facies has been correlated with diffusely-graded / massive and interbedded cross-stratified sand forming a 50 m thick tunnel channel fill ~ 8 km to the southeast (chapter 3). These strata occur in the terminal reaches of a regional channel network. Deposition probably occurred under conditions of rapid flow expansion, possibly involving a hydraulic jump and rapid loss of transport capacity (chapter 3). These changes are manifested in the abrupt streamwise transition from diffusely graded sand to medium and small-scale cross-strata with abundant climbing cross-stratification.

5.8.2 Subaqueous Fan

Subaqueous fan deposits consist of medium-scale cross-stratified, horizontally bedded and massive sand and gravel and small-scale cross-stratified fine sand, silt and rare clay laminae. Strata fine rapidly in a streamwise direction, and fine upward. The proximal part of the fan is lithologically diverse and consists of five facies-associations (chapter 4). The most proximal association, the fan-core association, consists of massive heterogeneous and horizontal-bedded cobble and pebble deposits. Vertical facies changes from cross-stratified to massive or planar-stratified gravel are commonly abrupt. Streamwise facies changes, although more gradational, are rapid with massive gravel grading in to cross-stratified sand over distances of <10 m. Lateral to the fan-core the flanking fan-core association consists of downlapping low-angle cross-strata and cross-stratified sand. Locally low-angle cross-strata are incised by steep-walled scours up to 10 m wide and 3 m deep filled with diffusely-graded sand forming the steep-walled scour association. Steep-walled scours define an important hydraulic boundary in the streamwise facies arrangement beyond which no gravel was transported. Downflow the gently-inclined bedset association is up to 8 m thick and consists of gently-dipping 5-10° surfaces overlain by planar cross-stratified medium sand and small-scale cross-laminated fine sand. Locally climbing medium-scale cross-strata are common, and more distally small-scale cross-lamination predominates; medium-scale cross-stratified sand generally fines upward. The lower fan succession is overlain locally by the shallow-channel association, which consists of broad (depth-width <1:5) nested channels that truncate underlying sandy strata. These channels are mostly infilled with planar and trough cross-stratified medium sand. Subaqueous fan deposits are commonly capped by silt and clay strata.

5.8.3 Basin Rhythmite Association

In core this facies association forms an expansive subsurface element of the moraine. It occurs between 170-270 m elevation with >70 % occurring between 190 and 230 m. In outcrop, units can be traced for tens of metres laterally with only minor changes in facies. The facies association consists of the rhythmic succession of small-scale cross-laminated, graded sand-silt and clay facies. Commonly the coarsest strata occur in the lower part of the rhythmite and form one or more upward-fining successions overlain by a clay layer. Cross-laminated and graded sand form > 95% of the association and commonly range in thickness from 1cm to 4-5 m with maximum thickness >16 m. Two different rhythmite assemblages are observed: one dominated by graded normally-graded sand-silt strata (Fig. 5.14a,b,e) and a second less common assemblage with thicker deposits including massive, normally-graded sand-silt and small-scale cross laminated sand (Fig. 13c). Of the 301 rhythmites measured in core, ~80% are <1 m thick. In 75 % (n=51) of rhythmites >1 m thick some core is missing. Rhythmite successions at different sites have been correlated on the basis of stratal thickness trends. The most confident correlations have been made where the largest number of rhythmites can be matched. The best example is between two cores DH-P-20a and DH-P-17a that are separated by ~ 2 km. In these two cores, stratal thickness trends can be matched for 22 rhythmites (Fig. 5.16), and can be recognized even where rhythmite thickness varies by more than an order of magnitude. Elsewhere correlating over distances as short as 500 m is extremely tenuous, and correlation is generally based on similarities in general thickness trends for a small number of rhythmites. The stacked, graded fine sand and silt strata that form the largest part of the rhythmites are interpreted as Bouma T_{bc} turbidites with the overlying clay interpreted as T_d turbidites. As discussed previously (Sect. 5.6.7) the clay facies has been interpreted to be deposited under conditions of low-energy suspension sedimentation during the winter. Consequently, in the absence of erosion, each clay strata may represent a single year of sedimentation and form a varve (Ashley 1975; Banerjee 1973; Gilbert 1997; Gustavson 1975). Consequently, these varves record low energy sedimentation that in the southwest part of the watershed lasted for <40 years and at the Vaughn site 20 km to the northeast lasted for ~70 years.

5.9 Depositional Model

A depositional model for the western Oak Ridges Moraine must integrate the sedimentology, regional physiography, surficial geology, and deglacial history. Lithofacies and lithofacies associations constrain interpretations of the local depositional processes and meltwater dynamics, and paleoflow measurements provide information on the direction of local subglacial meltwater flow. Elements of the regional physiography provide a framework for the depositional and erosional events during the Late Wisconsinan.

Landform and sediment associations also constrain the basin architecture. The following depositional model of the Oak Ridges Moraine consists of three stages (Fig. 5.17), that describe the change in meltwater and sediment discharge during the formation of the moraine. The stages generally form a depositional continuum; however, there is some overlap in the depositonal processes between elements of stage II and stage III.

5.9.1 Stage I. Subglacial Channel Fill

North of the study area, exposed shallow tunnel channels are partially eroded into drumlinized Newmarket Till (Fig. 5.1). Deeper tunnel channels are incised into lower deposits and extend 10s m into the subsurface, for example in the Holland Marsh valley (Fig. 5.1). From the northern edge of the Oak Ridges Moraine this channel network can be traced northward on the surface for up to 50 km to the Canadian Shield. Exposed tunnel channels are up to 3-4 km wide and 40 m deep. Beneath the Oak Ridges Moraine, tunnel channels with similar widths have been mapped but are up to 170 m deep (Pugin et al. 1999). Tunnel channels and drumlinized Newmarket Till form part of a regional unconformity that has been mapped from Georgian Bay east 240 km to the Kingston region and south across Lake Ontario (Lewis et al. 1999; Shaw and Gilbert 1990). Tunnel channels in this area have been interpreted to have evolved into an anabranched network following collapse of a regional north to south subglacial jökulhlaup sheet flood (Fig. 5.17a, Brennand and Shaw 1994; Shaw and Gilbert 1990). Initially channels were eroded partly into Newmarket Till. Subsequently, locally enhanced erosion of Newmarket Till occurred by spatial and temporal variations in turbulent flow structure and large-scale erosional vortices (Fig. 5.17b, Shaw 1996) or hydraulic jumps. Once formed, tunnel channels would have been preferentially reoccupied by subsequent meltwater discharge(Shoemaker 1999). Consequently, the final form and fill of tunnel channels is likely the product of multiple discharge events.

North of the Oak Ridges Moraine 1-4 m thick upward-fining gravel and sand deposits have been mapped overlying a broad tunnel channel interfluve of Newmarket Till northwest of the Holland Marsh (Fig. 5.1, Russell and Dumas 1997). Beneath the Oak Ridges Moraine, 10-15 km to the south, gravel deposits up to 20 m thick overlie the regional unconformity (Pugin et al. 1999) and occur on the raised shoulder(s) of deep tunnel channels (Fig. 5.8a). Furthermore ~65% of gravel in water well records classified as Oak Ridges Moraine strata occur within 10 m of the regional unconformity. Much of this gravel extends ~20 km southwestward across the central part of the moraine between 190-230 m asl. (Fig. 5.15).On the basis of defined tunnel channels this area of extensive gravel is a tunnel channel interfluve (Fig. 5.5). Gravel occurring in this interfluve area is interpreted to have been deposited during the waning stages of the north to south jökulhlaup flow that sculpted the regional drumlinized unconformity. Deposition occurred when water initially released as a broad sheet flood was localized into discrete channelized flows that eroded the tunnel channels (e.g. Brennand and Shaw 1994; Shaw 1996). This suggests that these gravels predate deposition of the moraine sensustricto. Gravel on raised channel margins was deposited by fluidal flows within still relatively shallow, <30 m deep, tunnel channels prior to erosion of >100 m deep channel axes (Fig. 3.2). On the basis of seismic reflection data (chapter 3) this channel gravel was deposited by dunes at least 5 m high. Regions of intense local erosion completely removed Newmarket Till and scoured into less competent underlying silt and sand of the lower deposits, for example at the DH-Nob site (Fig. 5.8). Once exposed, local rates of erosion were greatly accelerated. Erosion was further aided by over pressurized sandy aquifers that caused groundwater to flow toward lower pressure areas that coincided with newly formed scours (e.g. Piotrowski 1997). Accordingly, extensive sediment piping along the scour margins introduced additional sediment into the flow and significantly accelerated vertical and lateral scour growth. Locally these zones of intense scour erosion may have occurred at the margins of subglacial cavities, channel confluences or at other sites where large vortice developed. Large-scale vortices that formed in the scour entrained and transported sediment downflow. Where the sediment load was high the flow most probably became stratified with an upper fully turbulent region and an underlying thinner hyperconcentrated region termed a traction carpet (Sohn 1997). Depending upon the rate of sedimentation and the thickness of the traction carpet, massive, or inverse- or normal-graded strata were deposited (Fig. 5.7b, 5.17c). Similar strata have been interpreted from seismic data from the Holland Marsh channel (Pugin et al. 1999). In the eastern study area, extensive diffusely-graded sand fills are interpreted to have been deposited within the King City tunnel channel that was flooded by water of a subglacial lake in the Lake Ontario basin that had water levels of ~180 m (Fig. 5.7b, 5.5c). Interbeds of cross-laminated sand within the diffusely graded sand succession indicate that flow energy fluctuated temporally and/ or spatially between traction carpet and bedload depositional conditions. Thick accumulations of climbing, small-scale cross-laminated sand at sites PG and PG-II, ~ 10 km downflow to the southwest of site DH-V-158, were probably deposited by the same events that deposited the diffusely-graded and reverse-graded sand in the more proximal part of the Holland Marsh and King City tunnel channels.

The apparent divergent paleoflow directions, northwest, north, southwest at sites PG and PG-II may be explained by the radially expanding flow within a subglacial basin, deposition from deflected flows or deposition within a reverse flow eddy. Sections 0.5 to 4 km west and southwest of sites PG and PG-II form the margins of the tunnel channel and consist of lower deposits, specifically Scarborough sands and bedrock (White 1975). The deposits at the PG and PG-II sites are located near the western margin of the tunnel channel and may record deflected flow, or variable paleoflow directions related to local reverse flow eddied at the convergence of the Holland Marsh and King City tunnel channels. Alternatively, variable paleoflow directions are a common element of many subaqueous fan deposits (e.g. Gorrell and Shaw 1991) and submarine turbidite systems (e.g. Pickering et al. 1989) which in turn complicates the interpretation of paleoflow measurements and necessitates the integration of these data into a more regional context.

Downflow, an additional 20-30 km offshore within Lake Ontario, thick fills of glacilacustrine sand and silt are interpreted from seismic data to infill the Humber and Dundas valleys (Lewis et al. 1997a). Sediment fills within these two valleys are 50-60 m thick of which a substantial portion was emplaced following formation of the regional unconformity (Lewis et al. 1997a; Shaw et al. 1998). These deposits may be distal deposits of channel fills observed in the study area. Stratigraphically upward the diffusely-graded sand of stage I is intercalated with small-scale cross-laminated fine sand and the first development of clay strata that define the rhythmite association. This transition records a decrease in sedimentation rate and flow energy and a probable reduction in meltwater discharge along the tunnel channel system.

5.9.2 Stage II. Glacilacustrine Varve Sedimentation

In cores within well defined tunnel channels (DH-Nob, DH-V-158) the stage I channel fill sediment is overlain gradationally by thick coarse-grained rhythmites (Fig. 5.7, 5.17d). Additionally, ~15 km to the southwest of DH-V-4-158, rhythmites are present in the southwest part of the watershed along the buried southern flank of the moraine in 50-60 m deep bedrock valleys (Fig. 5.6). Initially the transition consists of small-scale cross-laminated sand that fines upward and is progressively overlain by the fine-sand to silt facies and silt-clay rhythmites. In DH-V-158 it consists of ~60 varves that thin and fine upward for 25 m and then rapidly coarsen upward. Locally, the upward-fining trend is punctuated by thicker rhythmites interpreted to record seasonal variation in discharge, or localized sediment input (e.g. slumping) and local order of magnitude increases in sedimentation (Fig. 5.17). The gradational transition and development of rhythmites is interpreted to record the collapse of the stage one jökulhlaup and transition to seasonal meltwater discharge and sedimentation in a glacilacustrine setting. Graded laminae were most probably deposited by unsteady underflows related to the diurnal meltwater flux. Similar laminated strata have been interpreted to record days of high ablation and increased meltwater discharge (Gilbert 1997; Weirich 1985).

Varves have only been observed in topographically low parts of the basin. This is most clearly illustrated in the southwest part of the study area (Fig. 5.6). Here the rhythmite association forms a 30-50 m thick fill within bedrock valleys but is absent or thin (<8 m thick) beyond the valley margins. Varves in the northwest part of the study area within tunnel channels are interpreted to have a similar relationship beyond the channel margins. Correlation of rhythmites across the study area is not possible. Consequently, rhythmite successions in the bedrock valleys and the tunnel channels may not be coeval and furthermore may have been deposited in isolated basins or cavities instead of in a single contiguous glacilacustrine setting. Regional models of the ice-sheet margin for this period do not support formation of isolated basins caused by low subaerial lake levels at elevations of ~230 m (e.g. Barnett 1992). Furthermore, subaerial exposure would have exposed easily eroded unconsolidated sediment overlying the intrabasin highs and provided proximal coarse-grained sediment to the basins; however, no such proximal deposits occur within the rhythmite succession. Consequently a subglacial setting is interpreted where ice reattached to the substrate across interfluves and bedrock highs. Deposition was then restricted to tunnel channels, bedrock valleys, and ponded water within the Lake Ontario basin.

If the rhythmites are interpreted to be record annual sedimentation these varves indicate a period of relatively low meltwater discharge for ~70 years along tunnel channels and 40-50 years in basin bedrock lows. Varve formation is not attributed to a change in sediment sources but instead to a regional reduction in meltwater discharge as upflow reservoirs recharged and only seasonal meltwater flowed into the basin.

In the southwest part of the watershed the rhythmite association of stage II is overlain by Halton Till, indicating sustained grounded-ice conditions during the remainder of moraine deposition. Elsewhere stage II rhythmites are gradationally overlain by coarser grained strata deposits of stage III.

5.9.3 Stage III. Esker - Subaqueous fan

In core an upward-thickening and coarsening of the varve succession is interpreted to record an increase in the meltwater discharge and the change from stage II to stage III sedimentation (Fig. 5.17e). This stage of moraine formation was in an ice-bounded basin with the Niagara Escarpment to the west. On the basis of ridge morphology (Fig. 5.3, 5.4), extent of exposed surficial sediment(Fig. 5.4), and areal variation in moraine thickness (Fig. 5.5b) deposition during stage III was restricted predominantly to the area of the moraine ridge in a basin that was ~30-40 km long (east-west) and ~20 km wide.

Evidence of the location of the basin margin and nature of drainage into the basin is based on previous studies north and south of the moraine. To the north of the moraine, Newmarket Till crops out along the edge of exposed moraine sediment (Fig. 5.3, Sharpe et al. 1997) indicating the approximate position of the grounding line and northern basin margin. North of this interpreted grounding line discontinuous glacifluvial deposits (c.f. Russell and Dumas 1997) indicate partly preserved conduit fills, or possibly, fill of a distributed cavity meltwater system (e.g. Fyfe 1990). Paleoflow measurements in these deposits indicate flow to the southwest or west (Russell unpublished data). Along the southern margin of the moraine the position of the grounding line is defined approximately by changes in the moraine thickness (Fig. 5. 5), surface slope in the vicinity of the 250m contour (Fig. 5.3), and the Brampton esker (Fig. 5.1). Relatively thin moraine sediment beneath the Halton Till plain supports the interpretation that ice was generally grounded below this elevation. A minimum width for grounded ice is provided by the length of the Brampton esker which is interpreted to be coeval with the Oak Ridges Moraine and to have been deposited in a R-channel (channel eroded into the ice, see Brennand, 2000 for a review of eskers) prior to deposition of the Halton Till. Consequently grounded ice conditions were required along the 30-40 km length of the esker and provide a minimum distance over which ice was grounded south of the Oak Ridges Moraine. A northward hydraulic gradient in this part of the glacier is indicated by paleoflow directions from strata of the Brampton esker that are generally northerly, toward the moraine (Saunderson 1976).

The depositional basin within which the moraine ridge was deposited may have been subglacial or proglacial (Fig. 5.17e). Presently there is no sedimentological or landform evidence that unequivocally supports either interpretation. Ponded water levels in this basin were controlled by the location at which ice impinged on the Niagara Escarpment and the ice thickness. Maximum water levels of ~420 m are indicated by the highest meltwater channel at Caledon(Fig. 5.1) (e.g. Chapman 1985). Fluctuations of the basin water depth would have been related to changes in the rate of meltwater discharge into the basin and corresponding rates of drainage from the basin. Basin drainage would have been closely related to the thickness of ice along the Niagara escarpment and the ice-sheet profile. A thin, low profile ice-sheet would have been more susceptible to buoyancy induced lift-off and subglacial drainage or erosion of ice-marginal channels along the escarpment that connected to drainage systems that flowed to the south (Fig. 5.1) (e.g. Chapman 1985; Tweed and Russell 1999).

Information on the glacihydraulic regime is limited. Based on the described esker sediment (Saunderson 1975) and moraine strata (e.g. Gilbert 1997; Paterson and Cheel 1997) a fully developed seasonal glacihydraulic system of supraglacial meltwater streams and englacial connections via moulins to a subglacial meltwater conduit system is envisioned (e.g. Brennand 2000; Gustavson and Boothroyd 1987). These associations of massive gravel, diffusely-graded strata, low-angle cross-stratified sand and steep-walled scours provide evidence that this system was locally supplemented by episodic jökulhlaup discharges. These jökulhlaup discharges may have drained supraglacial or subglacial reservoirs or alternatively may have been triggered by anomalous precipitation events (chapter 4, Cowan et al. 1988; Gilbert 1997; Maizels 1997). Consequently, stage III of moraine formation is characterized by esker-subaqueous fan sedimentation marginal to and within a small, ice-bounded glacilacustrine setting. Esker deposits are rare within the moraine and therefore the following discussion focuses on the subaqueous fan deposits. Subaqueous fan deposits record two contrasting but streamwise continuous depositional settings: the high-energy coarse grained proximal jet efflux and the low-energy fine-grained distal jet-efflux. The subaqueous fan sedimentation may have been controlled by seasonal meltwater discharge or local jökulhlaup discharge.

Jet Efflux, Proximal: Two types of jet efflux geometries are possible as the conduit debouched into a glacilacustrine environment: a subcritical plane jet or a supercritical plane-wall jet with an associated hydraulic jump (Fig. 4.1). Depending on conduit diameter, flow velocity, flow density and interfacial friction the jet may extend for up to 10s km into the basin. Where interfacial friction is low the jet has a downflow elongate parabolic form (e.g. Powell 1990), however, where friction is high the jet will have a broader more semicircular form (e.g. Wright 1977). Where deposition was from a plane-wall jet, the deposits grade rapidly downflow from massive clast-supported and matrix-supported gravel to plane-bed gravel. The abrupt loss of flow competence and capacity deposited rapidly aggrading strata that in a streamwise direction show rapid changes from massive to plane-bed and then cross-bedded coarse sand over distances as short as 10 m. Where flow energies were comparatively lower, sand beds containing antidune cross-stratification were deposited beneath undulatory jumps (chapter 4). In contrast, higher-energy flow conditions formed steep-walled scours that were infilled by diffusely graded sand beneath or downflow of oscillatory jumps. Steep-walled scour and fill record the most rapid rates of erosion and sedimentation; for example, overhangs that formed in unconsolidated sand along the scour margins. Downflow of the jump medium-scale planar cross-strata predominate and locally show moderate (15°) angles of climb.

Jet Efflux, Distal: More distal jet efflux deposits are commonly characterized by small-scale cross-lamination (Gorrell and Shaw 1991; Rust and Romanelli 1975) or graded fine-sand to silt. These strata are interpreted to be intermediate or distal subaqueous fan deposits. Where small-scale cross-laminated facies form thick metre-scale cosets, deposition is interpreted to have been from jökulhlaup discharges with internal flow pulsation. On the other hand, where thinner sets occur and fine upward to graded fine-sand to silt strata, deposition from surge-type turbidity currents is interpreted. In this case the stratal assemblage is similar to the T_{bc} and locally T_d Bouma turbidite divisions.

Stage III thus marks a second major change in the style of ORM sedimentation. Following the low sedimentation rates and meltwater discharge of stage II, there was a significant but gradual increase in energy, and a change in the style of meltwater delivery. It is postulated that discharge into the basin was from lateral conduits to the north and south and from an axial east-west conduit. Within the study area the best example of this lateral discharge is from the Gormley study site (chapter 4). Evidence of axial discharge is more equivocal, and is based on westerly paleoflows measured in coarse grained deposits 10-30 km east of the study area (e.g. Barnett 1997). Within the present study area the strata most likely related to this discharge consist of distal fine sand and silt deposits. The dimensions of the ice-bounded basin probably changed significantly as the ice-margins responded to thermal, and/or mechanical erosion of the meltwater discharge and buoyancy generated by increased water levels. During periods of high discharge, the basin grew larger as ice was either eroded or locally was lifted off the bed. During periods of low discharge, the basin shrank as ice became grounded and sedimentation was confined to the area of the ORM ridge.

5.10 Discussion

A number of depositional models have been proposed for the formation of the Oak Ridges Moraine. The longest standing model suggests that the moraine is an interlobate deposit formed between the Lake

Ontario ice lobe and the Simcoe or main Laurentide ice lobe (e.g. Chapman and Putnam 1984). Other authors have suggested that the moraine is not interlobate but rather is, at least partly, a subaerial braid plain deposit (Duckworth 1979; Gwyn and DiLabio 1973). More recently, however, detailed landform and sediment studies complemented by seismic profiles permitted Barnett et al. (1998) to interpret the moraine as a polygenetic landform of subglacial and ice-marginal origin with predominantly glacifluvial - glacilacustrine deposition. Subsequently, it has been suggested that the moraine may have formed subglacially at the margin of a large subglacial lake (Shoemaker 1999).

The sedimentological results of this study provide further support for the polygenetic origin of the moraine and also provide evidence that the moraine was deposited, at least in part, in a subglacial setting. The following discussion addresses conflicts between the four stage model of moraine formation recently proposed by Barnett et al. (1998) and the depositional history proposed in the current study. In addition, it reviews the current understanding of subglacial lakes and the suggestion by Shoemaker (1999) that formation of the moraine was controlled by a subglacial lake centred in the Lake Ontario basin.

The current study provides additional details from the western Oak Ridges Moraine to refine the four-stage depositional model proposed by Barnett et al. (1998) (Table 5.4). The channel-fill lithofacies that defines stage I in the current study corresponds with stage I in the Barnett et al. model. Deposition was from episodic subglacial jökulhlaup flows along north-south trending tunnel channels. Furthermore, strata were deposited at a zone of rapid flow expansion and commensurate rapid and voluminous sedimentation within a subglacial glacilacustrine setting. This is followed by stage II, which in this study corresponds to a period of low-energy meltwater discharge and varve formation with thin normally-graded fine-sand to silt laminae recording diurnal meltwater pulsation. Locally subaqueous fan sedimentation may have been active in stage II (this paper); however, it is not characteristic of the regional sediment and meltwater flux associated with this interval of moraine formation in the Humber River watershed. This differs from the Barnett et al. definition of stage II as subaqueous fan sedimentation. Consequently, rapid and voluminous sedimentation associated with subaqueous fan processes is more appropriately placed at the base of stage III, which Barnett et al. termed subaqueous fan to delta. Previous descriptions of stage III are consistent with the stage III strata in the current study except that deltaic deposits were not observed in the Humber River watershed. This might be expected, as deposition of subaqueous fans or deltas are commonly controlled by the depth of water at the ice-front or the position of the jet efflux from the glacier (i.e. subglacial, englacial or supraglacial). Consequently for the same melt water discharge variation in basin geometry, depth and ice-sheet drainage will result in different stratal architectures (e.g. Fyfe 1990). Subaqueous fan and deltaic deposits, although lithologically similar can commonly be differentiated on the basis of differences in stratal architecture, as well as lateral and vertical facies associations. Subaqueous fans generally do not have well developed clinoforms, or foreset beds and where present generally have a much lower angle of dip (Chapter 4). Furthermore, streamwise facies changes in subaqueous fans are generally much more abrupt and generally fine rather than coarsen upward. Stage IV of the Barnett et al model described ice-marginal sedimentation of the gradationally overlying Halton Till which has not been included in this study.

The depositional model presented by Barnett et al. (1998) provided only limited discussion concerning mechanisms that may have controlled the location of moraine deposition. The location of the Oak Ridges Moraine has traditionally been attributed to the development of an interlobate re-entrant formed due to ablating and thinning ice along the ice-margins of the Lake Ontario and Lake Simcoe ice lobes. This re-entrant has been suggested to have formed during the Mackinaw interstadial (Duckworth 1979). Recent developments in the understanding of ice-streams (e.g. Hart 1999), surging glaciers (Kamb et al. 1985), and subglacial lakes (Shoemaker 1999) indicates that the origin of the moraine may be more closely controlled by such events. Using a regional subglacial meltwater model, Shoemaker (1999) suggested that a subglacial lake developed within the Lake Ontario basin and that the moraine may have formed at the grounding line of this lake. Contrary to the suggestion of Shoemaker (1999), evidence presented in this study suggests that only the lowest tunnel channel fill succession (Stage I) was controlled directly by this inferred lake. Nevertheless the control on formation of the moraine by a subglacial lake is sufficiently significant and controversial that a brief review of the evidence for subglacial lakes is presented.

Subglacial lakes are known to occur beneath the Antarctic ice-sheet and to have areal dimensions similar to or larger than modern-day Lake Ontario (Dowdeswell and Siegert 1999). Furthermore, it is recognized that the Antarctic lakes exist under radically different glacial climatic conditions than the deglacial conditions of the Laurentide ice-sheet. Antarctic subglacial lakes have no supraglacial recharge and are sustained by subglacial meltwater generated by geothermal heat flux and extremely thick ice producing a warm based glacial regime. Consequently, the development of subglacial lakes during Late Wisconsin Laurentide deglaciation is plausible in light of its generally warm based character, large volumes of supraglacial meltwater, and interconnected englacial meltwater system (e.g. Shaw 1996). In fact, sedimentological studies are beginning to recognize evidence for the existence of such lakes not only beneath the Laurentide ice-sheet (Munro-Stasiuk 1999) but also beneath other ice-sheets (McCabe and O'Cofaigh 1994). Formation of a subglacial lake in the Lake Ontario basin most probably post-dated the formation of regional drumlin fields that occur north and south of modern Lake Ontario (Boyce and Eyles 1991; Shaw and Sharpe 1987) and also have been mapped on the bottom of Lake Ontario (Lewis et al. 1997b). Irrespective of whether the drumlins were formed by streams of deforming till (Boyce and Eyles 1991) or as a result of regional meltwater floods (Shaw 1996; Shaw and Sharpe 1987; Shoemaker 1999) a later glacial surge and consequent thinning of the ice-sheet occurred. Glacial surges have been shown to coincide with increased basal meltwater, subsequent meltwater discharge and thinning of the ice-sheet (e.g. Kamb et al. 1985; Patterson 1997). Evidence of extensive meltwater flow following the regional drumlin-forming events is further supported by the development of tunnel channels that traverse and truncate the drumlin fields (e.g. Barnett 1990; Brennand and Shaw 1994; Mullins and Hinchey 1989; Mullins and Eyles 1996). Tunnel channels are interpreted to form subglacially either by steady state meltwater discharge or due to episodic, catastrophic meltwater discharge (e.g. Cofaigh 1996). In the area north of Lake Ontario, tunnel channel formation has been interpreted to be by episodic jökulhlaup discharges (Barnett 1990; Brennand and Shaw 1994). Irrespective of the formative tunnel channel meltwater regime, the width, depth and fill of the channels (Pugin et al. 1999; Shaw and Gorrell 1990) indicate that large quantities of meltwater discharged southward along tunnel channels into the Lake Ontario basin. Channelisation of meltwater within tunnel channels would have coincided with the end of surging glacial flow conditions and the development of a thinned region of ice across the Lake Ontario basin (e.g. Patterson 1997; Shoemaker 1999). This change in the ice-sheet thickness would have produced conditions favourable for the accumulation of meltwater in the Lake Ontario basin and development of a subglacial lake (e.g. Shoemaker 1991; Shoemaker 1999). The development of this subglacial lake is supported by the absence of strongly developed channels south of the Oak Ridges Moraine and farther south within Lake Ontario (c.f. Kenny 1998; Sharpe et al. 1996). Consequently, the abrupt change in landform styles of the Oak Ridges Moraine from north to south suggests a major change in the glacial hydraulic regime in the area of the moraine. This change was unrelated to a glacial margin because ice was well to the south of the moraine area prior to formation of glacial Lake Iroquois (e.g. Karrow and Occhietti 1989). Glacial Lake Iroquois is generally interpreted to have been formed ~500 years after deposition of the Oak Ridges Moraine.

A subglacial lake the size of Lake Ontario would have significantly influenced the surface topography of the overlying ice-sheet (Shoemaker 1999) in a fashion similar to that documented from the Antarctica ice-sheet (Siegert and Ridley 1998). For example, Lake Vostok, the largest documented subglacial lake beneath the Antarctic ice-sheet, has a similar surface area to present day Lake Ontario, and is overlain by ice with a surface topography of 0.004° (0.25 m km^{-1}) (Siegert 2000). Consequently, regional ice-sheet dynamics during Laurentide deglaciation established three important conditions for formation of the Oak Ridges Moraine: a thinned ice-sheet, a subglacial lake within the Lake Ontario basin, and a low ice-sheet gradient. The thinned ice-sheet would have permitted more water to accumulate in Lake Ontario as the lithostatic pressure of the confining ice was reduced. Furthermore, the low ice-sheet gradient over a subglacial lake results in a lower hydraulic gradient in the area and reduced subglacial meltwater pressures outward from the centre of the Lake Ontario basin to the grounded margins (e.g. Brampton esker). Once established, a subglacial lake would have been impossible to drain in the absence of a considerable increase in ice thickness (Shoemaker 1999). A thinned ice-cover over the lake would also have been more susceptible to changes in water depth within the basin. The influence of changing water depth in subglacial lakes on the overlying ice-lid is poorly understood. It is plausible that the ice may have fractured extensively, forming crevasses around the lake margin, in a fashion similar to that in ice-covered reservoirs with falling water level (e.g. Stander et al. in prep). The number and width of the crevasses would be closely related to the thickness of the ice-cover, the underlying basin geometry and topography, the rate of change in water level, and the total change in water level. Although such crevassing of lake ice may involve the whole thickness of the ice-cover, it is unlikely that this would be the case with a glacier ice cover where the ice was 100s m thick. Dry crevasses in temperate glaciers generally only extend to a depth of 25-30 m because at this depth the horizontal stress in the ice-sheet generates rapid ice flow and closure of crevasses (Paterson 1981). An exception to this occurs where the crevasses fill with water and in this case crevasses may extend to the bottom of the glacier (Paterson 1981). With a low gradient ice-cover and supraglacial ponding of water it is probable that the crevasses formed vicinity of the grounding line would have filled with water and provided an additional control on the location of the Oak Ridges Moraine. Crevassing may have localised thermal and meltwater erosion and consequently controlled development of the ice-walled glacilacustrine environment into which stage III of the Oak Ridges Moraine was deposited. Shoemaker (1999) postulated that the Oak Ridges Moraines may have formed at a grounding line where discharge into subglacial lakes underwent flow expansion and consequently lost transport competence.

Observations of sediment fill from tunnel channels beneath the western Oak Ridges Moraine provides some support for the hypothesis that at least the lower part of the Oak Ridges Moraine was deposited subglacially. The lowermost channel fill strata of diffusely-graded to massive sand that grade downflow into climbing small-scale cross-laminated sand are interpreted to have been deposited beyond the grounding line within a subglacial cavity - possibly a subglacial Lake Ontario. Based on the elevation of these strata the water level in this subglacial lake would have approached ~180 m asl as compared to the current Lake Ontario level of ~74 m asl.. This subglacial lake may have been ~300 m deep and 10s km wider than modern-day Lake Ontario. Additionally, evidence of rapid sedimentation at a zone of flow expansion is provided by the chaotic seismic facies in the Holland Marsh channel, interpreted to be correlative with strata from the subparallel King City tunnel channel located ~8 km to the east (Fig. 5.5). These two tunnel channel fills are interpreted to be the initial stages of moraine formation, possibly deposited by the last large flow along the tunnel channels before reorganization of the ice-sheet hydrology and development of lower-energy axial flow along the Oak Ridges Moraine, accompanied by secondary lateral inflow.

5.11 Summary/Conclusion

The western segment of the Oak Ridges Moraine is predominantly a glacifluvial-glacilacustrine deposit that accumulated in an ice-controlled environment. The moraine overlies an irregular erosional unconformity with relief of 80-100 m. As a result, moraine thickness is highly variable ranging from <10 m to >150 m. Regional models of landform development for both drumlins and tunnel channels suggest that the Late Wisconsin Laurentide ice-sheet surged and as a consequence thinned. Regardless of whether the tunnel channels formed in response to episodic and catastrophic meltwater discharge or steady state discharge, large volumes of water obviously was discharged south along the channels.

The absence of tunnel channels south of the Oak Ridge Moraine suggests that the position of the moraine marks a significant change in hydraulic regime in relation to the ice-sheet, and that a subglacial lake, such as that recently proposed by Shoemaker (1999) must have formed in the Lake Ontario basin. Presently the extent and depth of this lake is poorly constrained, but it was probably larger than Lake Ontario and as much as ~300 m deep. This subglacial lake, in conjunction with tunnel channels, most probably provided the primary control on formation of the Oak Ridges Moraine.

Moraine formation has been interpreted to have occurred in three stages each reflecting changes in meltwater discharge, location and orientation of subglacial or proglacial basins, conduit efflux location, and ice-cover. Deposition during stage I was from waning jökulhlaup discharges through subglacial tunnel channels. At this time, enhanced headward erosion of the tunnel channel(s) upflow from the grounding line was probably related to supercritical discharge into subglacially ponded water in Lake Ontario. The largest meltwater discharges were focused at the mouth of these tunnel channels where hydraulic jumps developed as the flow underwent rapid flow expansion and mixing, resulting in transition to subcritical flow. Meltwater discharge was from the north and the predominant channel fill lithofacies consisted of diffusely-graded / massive sand deposited from hyperconcentrated flows. Stage II marked a change to annual meltwater flux and varve formation in a subglacial environment. Sediments produced during stage II consist mostly of normal-graded fine-sand to silt deposited from low-energy diurnal discharge. The location of the Oak Ridges Moraine and Stage III sedimentation may have been strongly controlled by crevassing along the grounding line of the thinned ice-cover over the lake as water levels within the subglacial lake fell below ~140 m asl.. Stage III represented a gradual return to higher sediment flux and the development of multiple lateral sediment point sources into an axial meltwater system that flowed westward. Coarse-grained sediments accumulated predominantly within subglacial conduits and proximal parts of subaqueous fans. Distal deposits consist entirely of medium and fine sand commonly with climbing cross-stratification.

The Oak Ridges Moraine is interpreted to have formed in response to a complex sequence of events related to the Lake Wisconsin Laurentide deglaciation. Regional events, possibly involving broad regional catastrophic meltwater floods that caused enhanced glacial flow conditions, possibly including a glacial surge, culminated in the thinning of the ice-sheet and excavation of a tunnel channel network discharging meltwater into the Lake Ontario basin. Ponding of meltwater in this subglacial lake provided the primary control on location of the moraine through development of a grounding line, and subsequently during lake drainage along a zone of grounded ice and extensive crevasses. These regional scale jökulhlaup discharges caused rapid and voluminous sedimentation and deposition of the moraine, possibly in as little as 100 years.

Fig. 5.1. a) Digital Elevation Model of the Oak Ridges Moraine with the four moraine segments labeled. Northern part of the study area is shown. Letters in italics indicate selected landscape elements, i) tunnel channels, ii) Peterborough drumlin field, iii) Niagara Escarpment. The 270 m contour is used to outline the approximate extent of the Oak Ridges Moraine. Note tunnel channels defined by northeast-southwest trending valleys with low surface roughness north of the Oak Ridges Moraine and intervening drumlinised terrain.

b) Location of the Humber River watershed and the Oak Ridges Moraine northwest of Toronto. Principal drill core, river section and aggregate pit sections used in this study are shown. Background DEM shows contrasting landscape of area with tunnel channels (TC) and drumlinised Newmarket Till uplands (D) north of moraine and Halton Till plain (HT) south of moraine. Selected towns are indicated by the letters in circles, N- Nobleton, B- Bolton, W-Woodbridge.

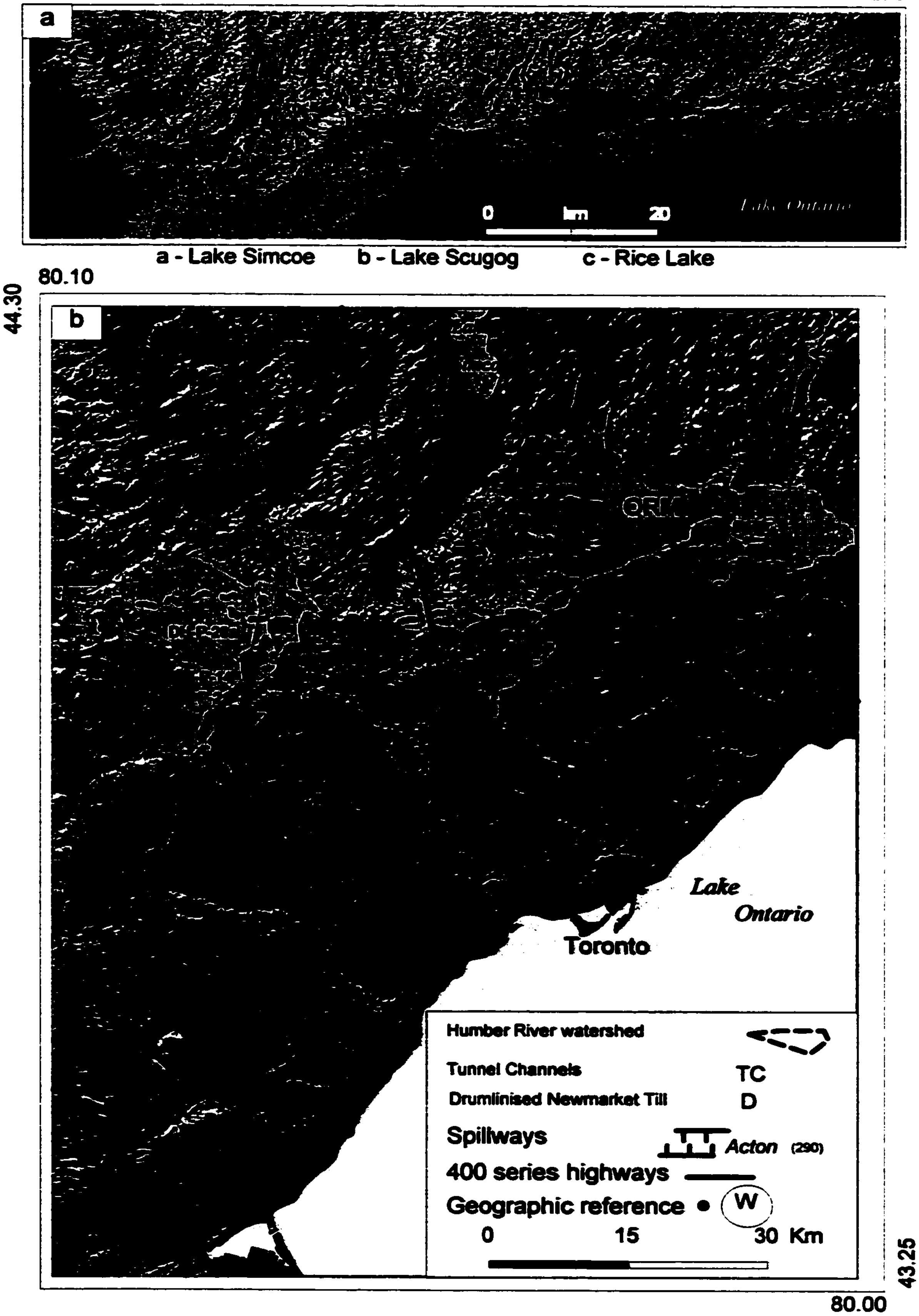

a
0 km 20
a - Lake Simcoe
b - Lake Scugog
c - Rice Lake
80.10
44.30
b
Lake Ontario
Toronto
Humber River watershed
Tunnel Channels TC
Drumlinised Newmarket Till D
Spillways Acton (290)
400 series highways
Geographic reference W
0 15 30 Km
43.25
80.00

Fig. 5.2. Archival datasets used primarily to create the Oak Ridges Moraine structural and isopach maps. (a) Surficial mapping sites of Russell and White (1997). (b) Hydrogeology borehole data and location of two Interim Waste Authority sites in study area. (c) Geotechnical borehole sites in the study area and location of two seismic profiles. (d) Selected MOE water well data based on deepest hole per 500 m grid. (e) Matrix showing relationship between lithofacies of this study (left column) and

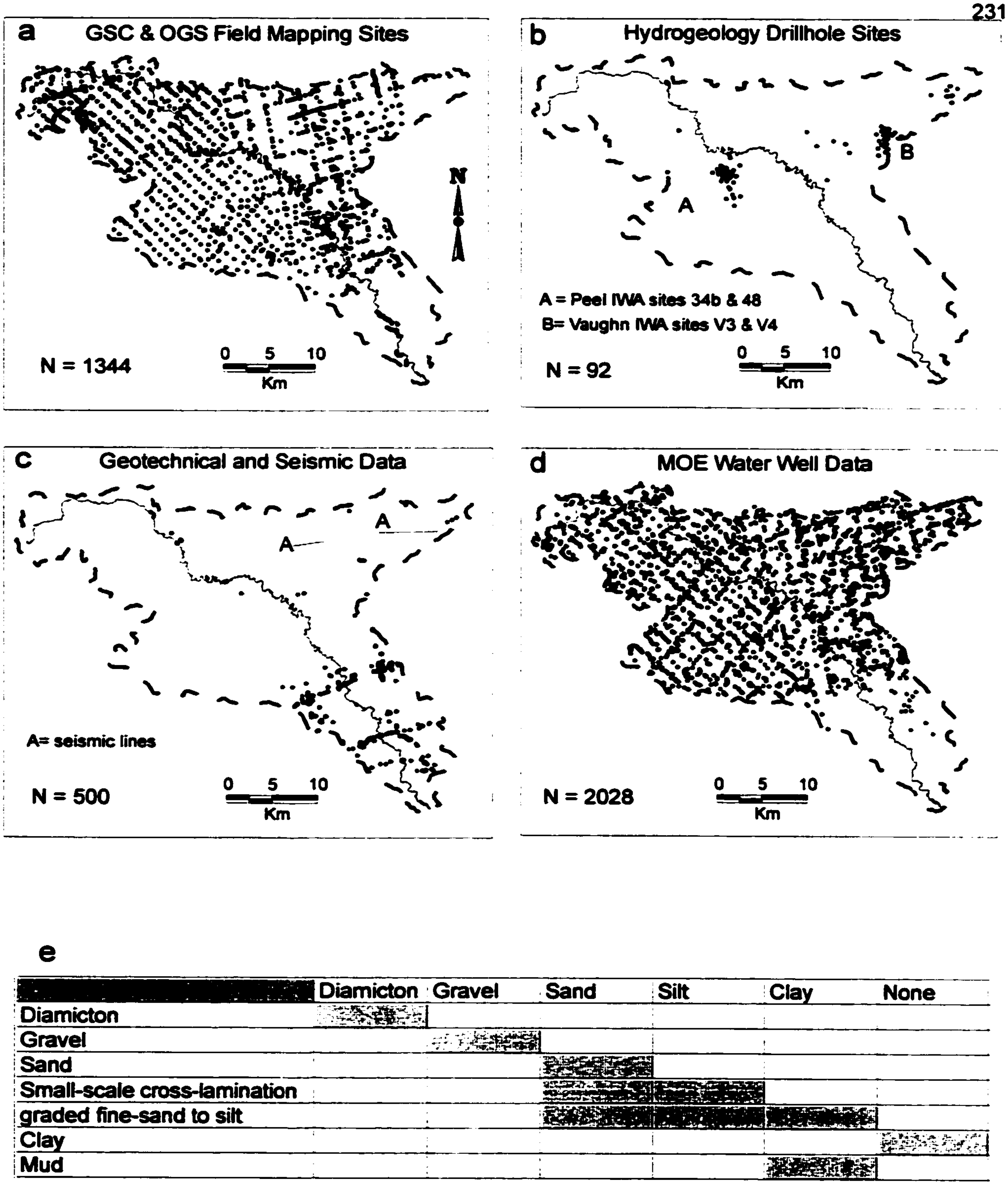

a
GSC & OGS Field Mapping Sites
N
N = 1344
0 5 10
Km
b
Hydrogeology Drillhole Sites
A
B
A = Peel IWA sites 34b & 48
B= Vaughn IWA sites V3 & V4
N = 92
0 5 10
Km
c
Geotechnical and Seismic Data
A
A
A= seismic lines
N = 500
0 5 10
Km
d
MOE Water Well Data
N = 2028
0 5 10
Km
e
Diamicton
Gravel
Sand
Silt
Clay
None
Diamicton
Gravel
Sand
Small-scale cross-lamination
graded fine-sand to silt
Clay
Mud

Fig. 5.3. DEM of the Humber River watershed indicated by the dashed line (Kenny et al. 1999) with location of 25 km long south-north cross-sections. Note irregular topography of the moraine between sections A and C. Steeper slopes on the southern moraine flank are indicated by deeper drainage incision. Smoother topography of moraine occurs where Halton Till is exposed at the surface. Elements of the regional physiography are labeled in italics: i Niagara Escarpment, ii tunnel channels, iii Oak Ridges Moraine, and iv Halton Till. Interpreted northern and southern position of Oak Ridges Moraine glacilacustrine basin during formation of moraine ridge is shown by lines a and b respectively. Approximate minimum extent of grounded ice during stage III of deposition of the moraine and Brampton esker is shown by line c.

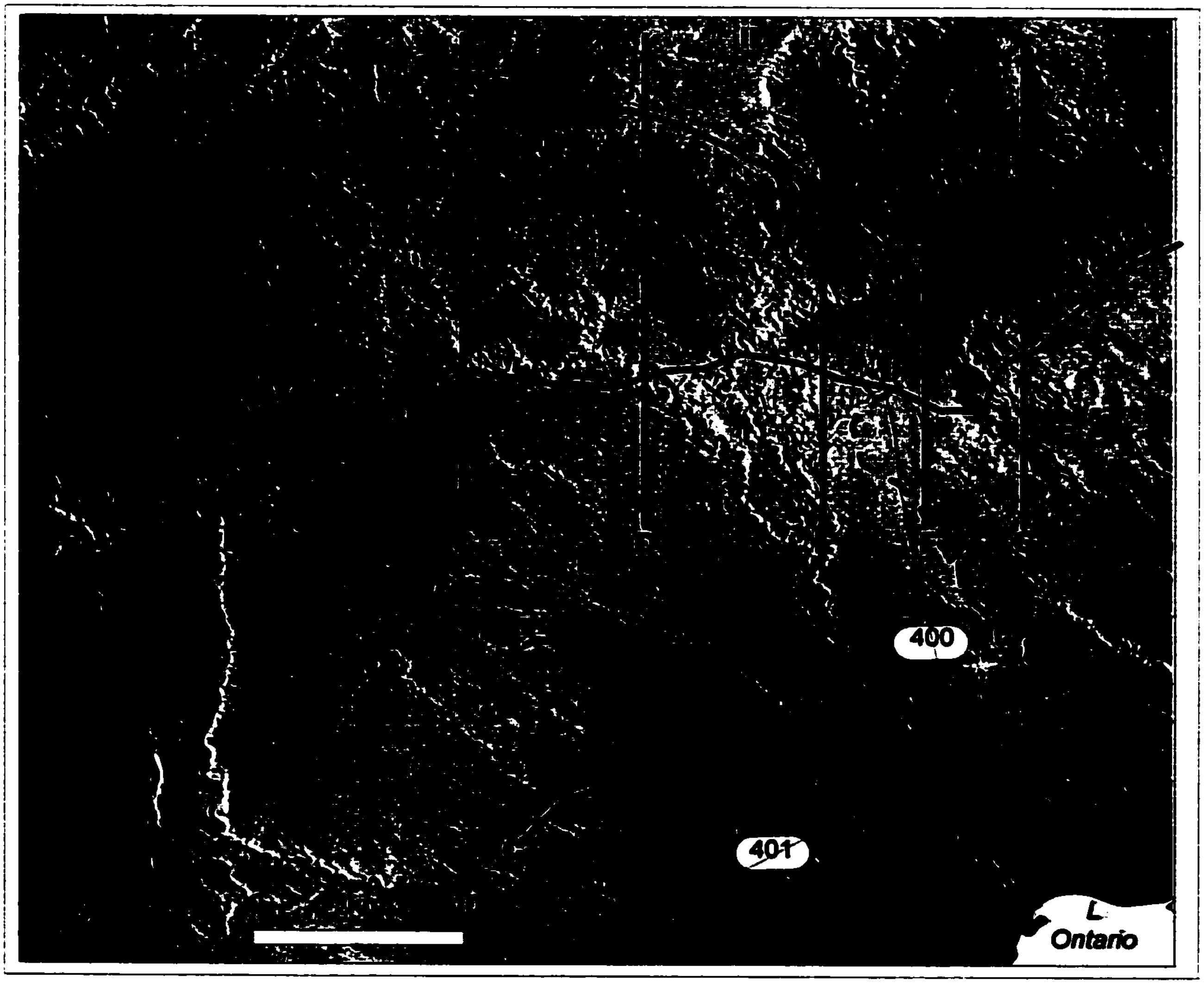
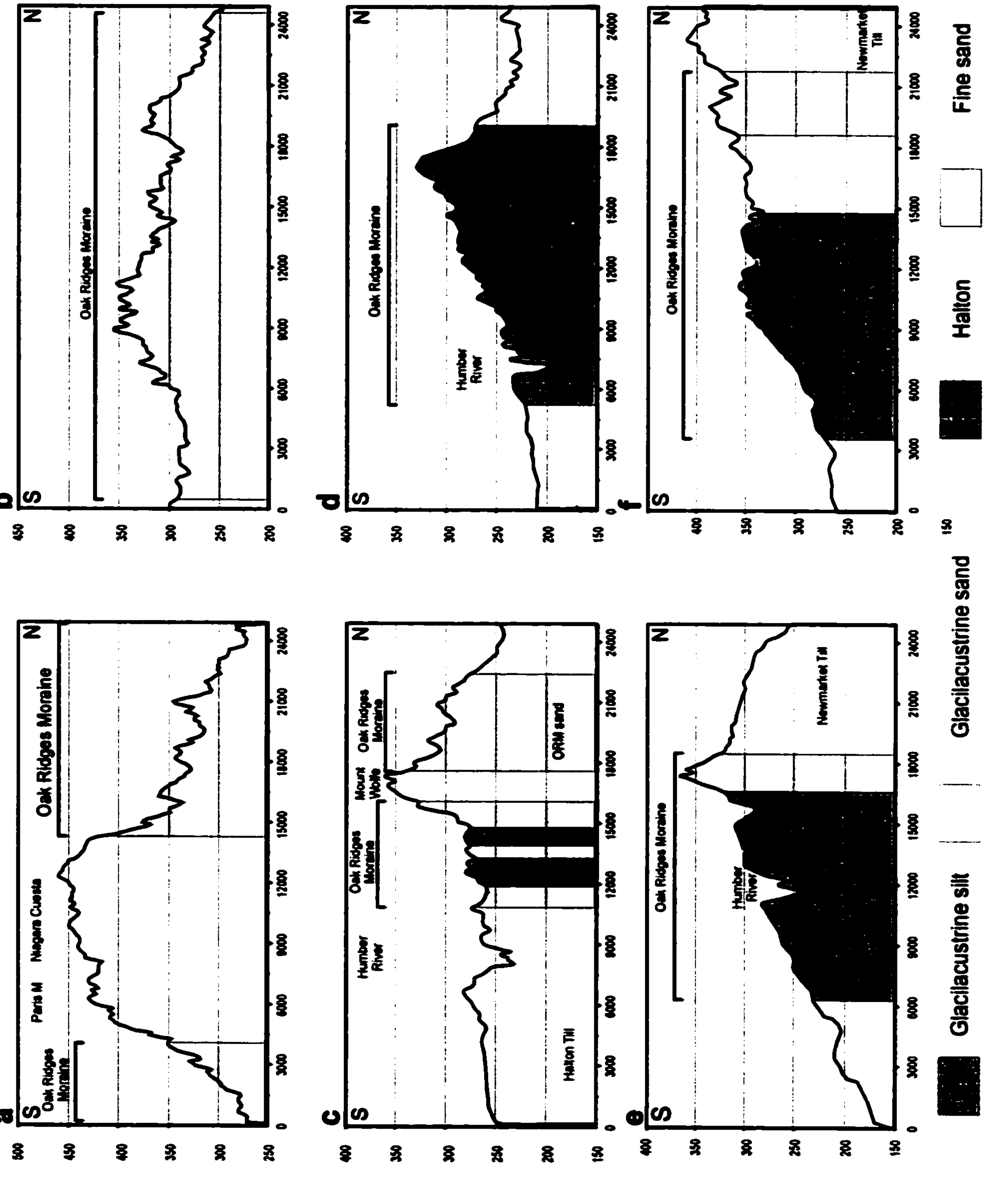
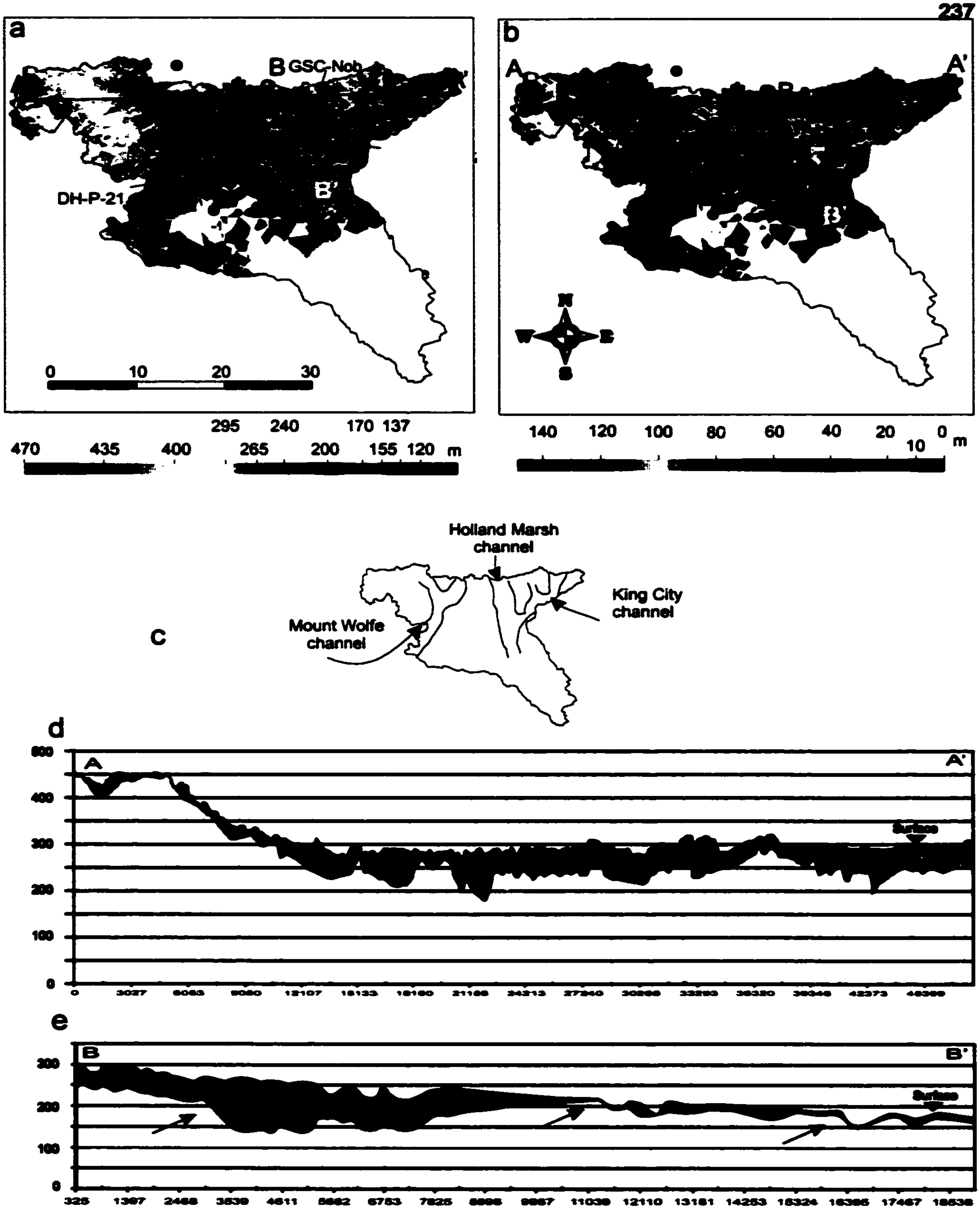

400
401
L.
Ontario

Fig. 5.4. Six 25 km long north-south topographic cross-sections of the Oak Ridges Moraine spaced 4-8 km apart (see fig 5.3 for location). Lower shaded graph indicates surface sediment composition of the moraine Note the moraine is composed predominantly of moraine sand and Halton Till. Vertical exaggeration is ~23 times.



a
b
c
d
e
f
S
N
Oak Ridges Moraine
Paris M
Niagara Cuesta
Humber River
Mount Wolfe
Halton Till
ORM sand
Newmarket Till
Glaciacustrine silt
Glaciacustrine sand
Halton
Fine sand

Fig. 5.5. (a) Colour coded elevation model for the base of the Oak Ridges Moraine. Note overdeepened areas. (b) Isopach map of Oak Ridges Moraine. (c) Interpretation of tunnel channel locations. (d) West to east cross-section of the moraine showing top and bottom surfaces. (e) North to south cross-section of the moraine showing top and bottom surfaces. Arrows point to artifacts discussed in text.



a
B GSC-Nob
DH-P-21
B'
0 10 20 30
295 240 170 137
470 435 400 265 200 155 120 m
b
A
A'
B'
N
W
E
S
140 120 100 80 60 40 20 0 m
10
Holland Marsh channel
King City channel
Mount Wolfe channel
c
d
A
A'
Surface
0 100 200 300 400
0 3027 6053 12107
e
B
B'
Surface
0 100 200 300
325 1397 2468 3539 4611 6753 7825 11039 12110 13181 14253 15324 17467 18538

Fig. 5.6. Graphic sedimentological logs of continuous core from the Peel IWA sites, southwestern Humber River watershed. Halton Till overlies Oak Ridges Moraine in this area. Note how Halton Till drapes underlying deposits and how varves are generally restricted to the bedrock low.

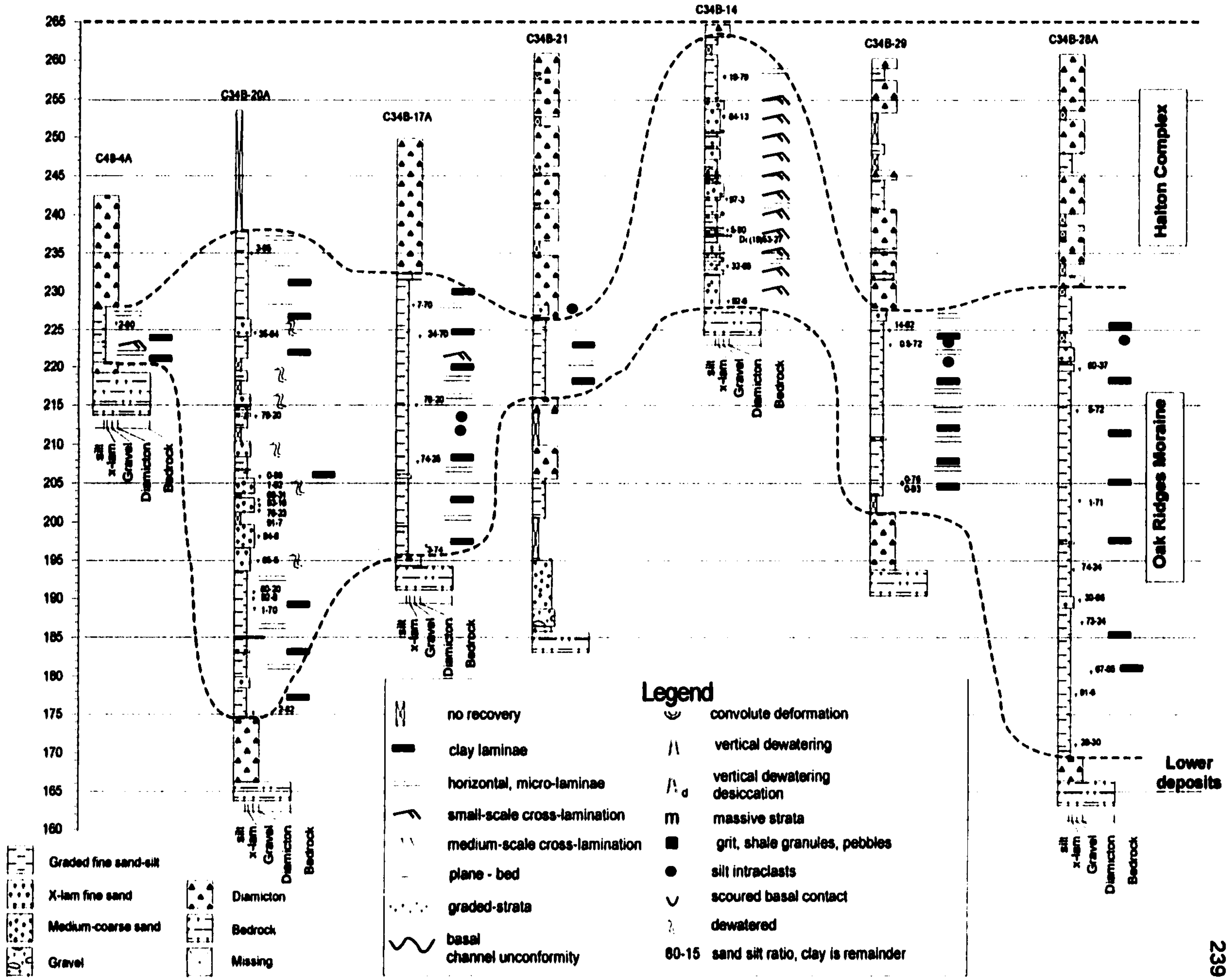
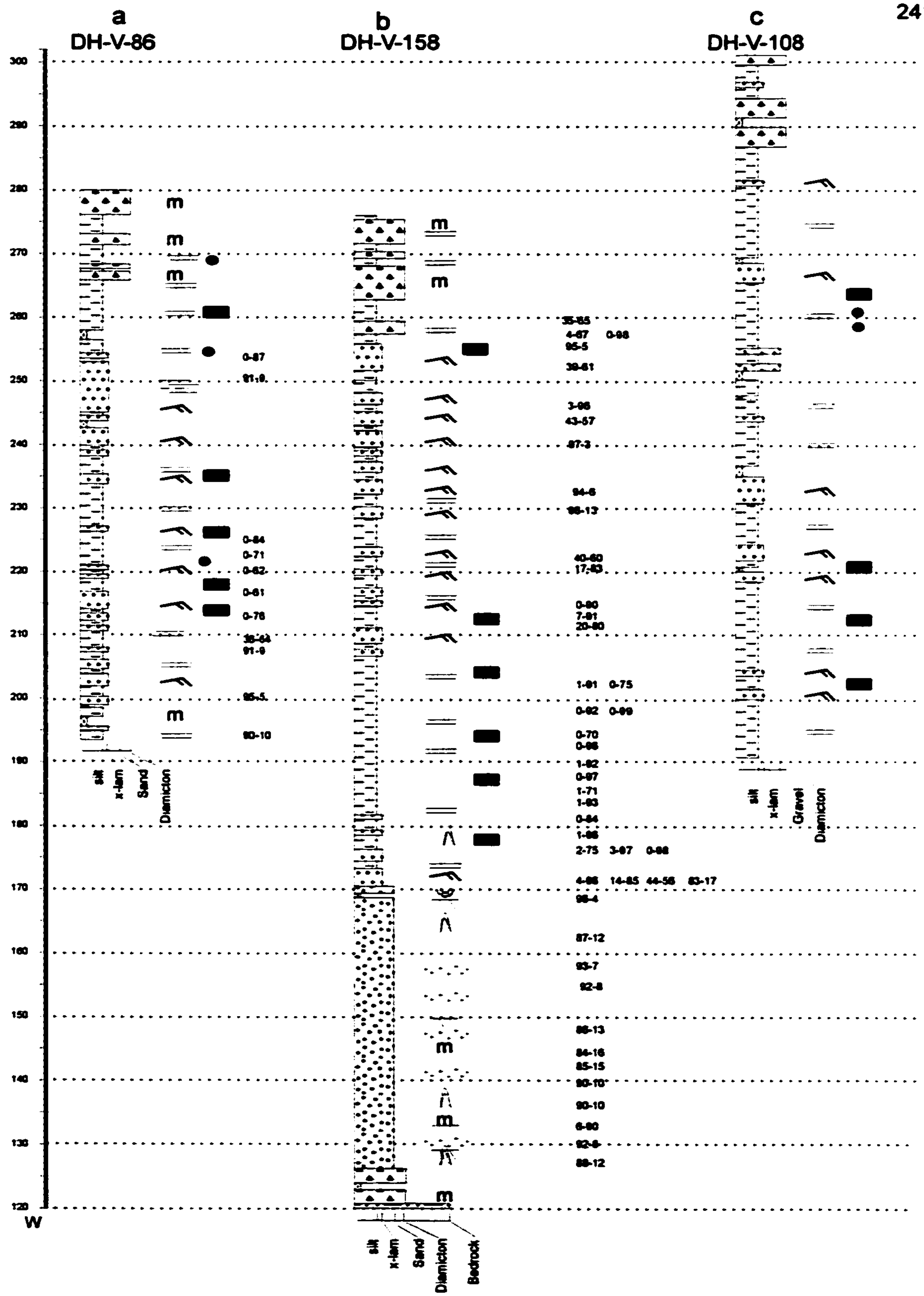

C4B-4A
C34B-20A
C34B-17A
C34B-21
C34B-14
C34B-29
C34B-28A
Halton Complex
Oak Ridges Moraine
Lower deposits
silt
x-lam
Gravel
Diamicton
Bedrock
Graded fine sand-silt
X-lam fine sand
Medium-coarse sand
Gravel
Diamicton
Bedrock
Missing
Legend
no recovery
clay laminae
horizontal, micro-laminae
small-scale cross-lamination
medium-scale cross-lamination
plane - bed
graded-strata
basal channel unconformity
convolute deformation
vertical dewatering
vertical dewatering desiccation
massive strata
grit, shale granules, pebbles
silt intraclasts
scoured basal contact
dewatered
80-15 sand silt ratio, clay is remainder

Fig. 5.7. Graphic sedimentological logs of continuous core from the Vaughn IWA site, northeastern Humber River watershed. At this location, thin interbedded Halton Till overlies Oak Ridges Moraine sediment. Note unique occurrence of diffusely-graded sand in lower part of DH-V-158. See Figure 5.6 for description of symbols.



a
DH-V-86
b
DH-V-158
c
DH-V-108
300
290
280
270
260
250
240
230
220
210
200
190
180
170
160
150
140
130
120
W
silt
x-lam
Sand
Diamicton
Bedrock
Gravel
m

Fig. 5.8. Graphic sedimentological log of a) GSC-Nob core, b) Humber River sections, and c) Gormley pit (see figure 5.1 for location). See Figure 5.6 for description of symbols. Lithofacies for the Gormley site are shown.

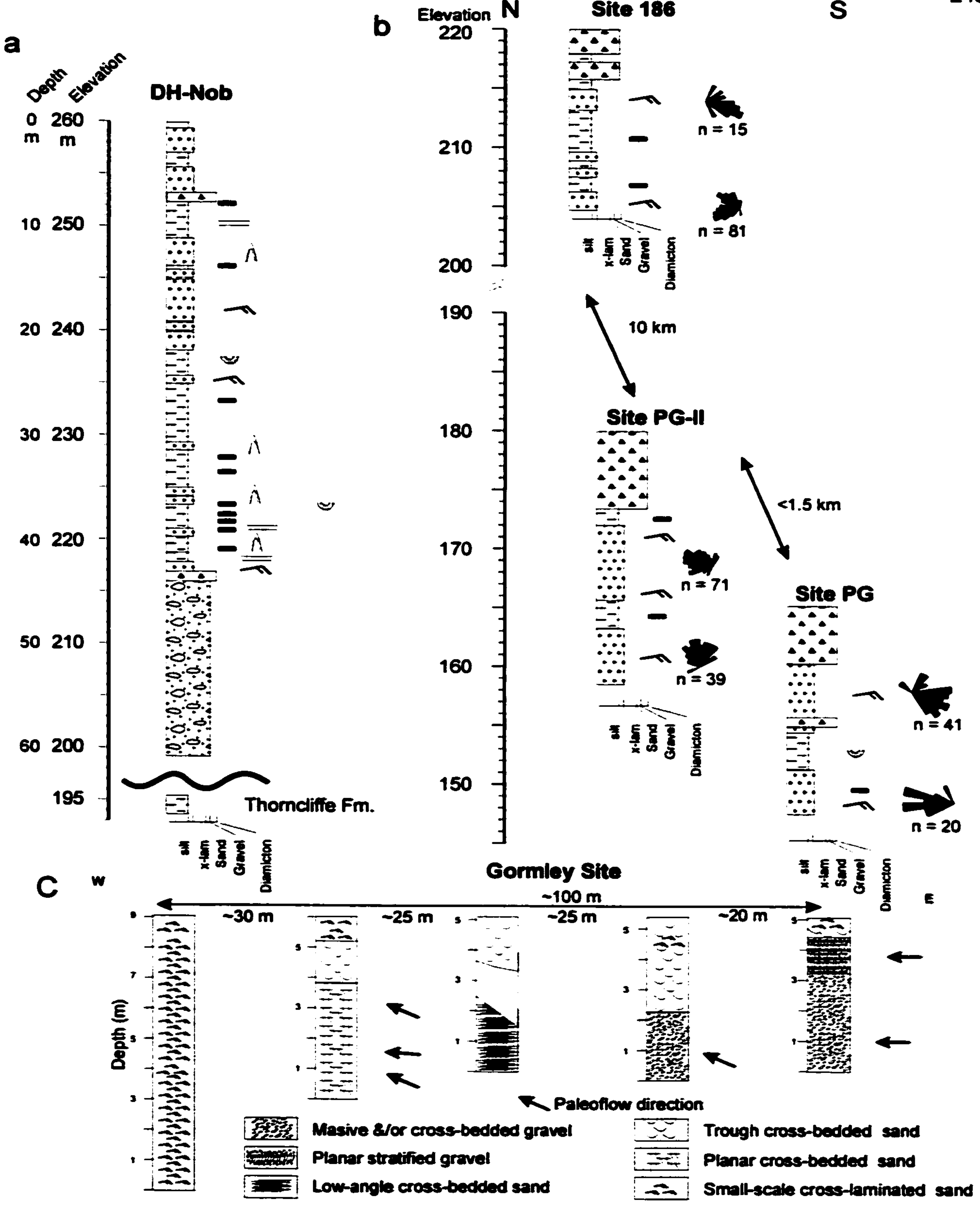
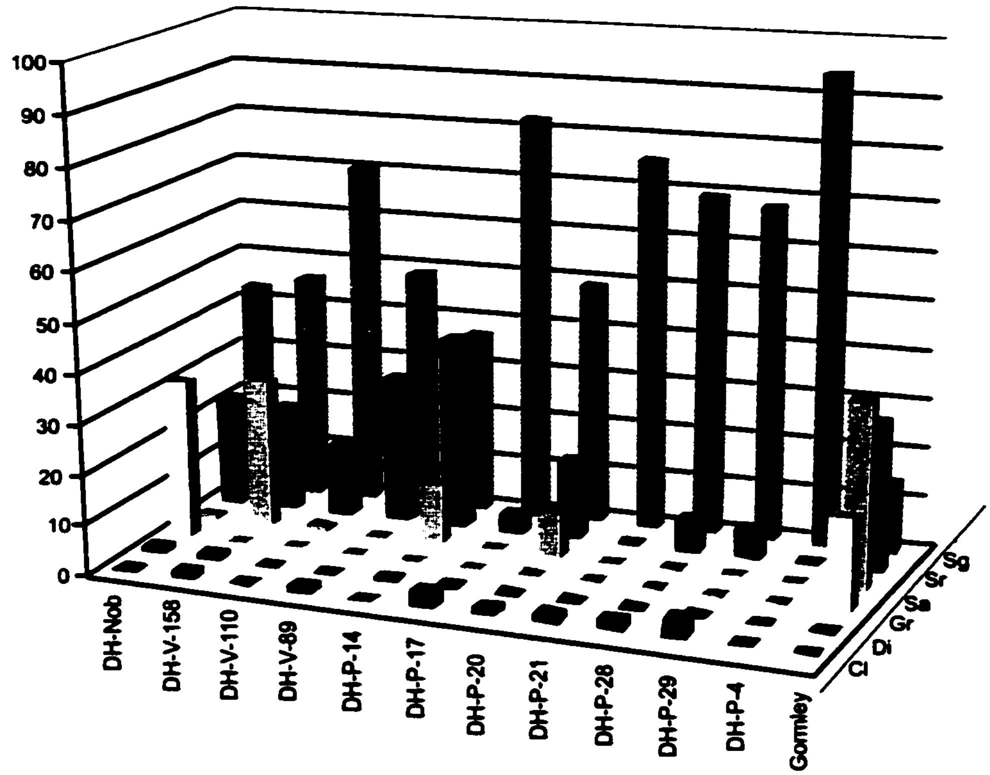

a
Depth
Elevation
DH-Nob
0 m
260 m
10 250
20 240
30 230
40 220
50 210
60 200
195
Thorncliffe Fm.
silt
x-lam
Sand
Gravel
Diamicton
b
Elevation
N
Site 186
S
220
210
200
190
180
170
160
150
n = 15
n = 81
10 km
Site PG-II
<1.5 km
n = 71
n = 39
Site PG
n = 41
n = 20
C
W
Gormley Site
~100 m
~30 m
~25 m
~25 m
~20 m
E
Depth (m)
Paleoflow direction
Masive &/or cross-bedded gravel
Planar stratified gravel
Low-angle cross-bedded sand
Trough cross-bedded sand
Planar cross-bedded sand
Small-scale cross-laminated sand

Fig. 5.9. (a) Graph of lithofacies by borehole site. Note predominance of graded fine-sand to silt lithofacies. (b) Graph of lithofacies from MOE, water wells, hydrogeology and geotechnical boreholes. Vertical axis is percent. Cl - clay, Di - diamicton, Gr - gravel, Sa- medium-coarse sand, Sr- small-scale cross-laminated fine sand, Sg- graded fine sand to silt. Abbreviations on horizontal axis of b are, MTO - Ministry Ontario Transport, RMP - Regional Municipality of Peel, RMY - Regional Municipality of York, GSC-Bh - Geological Survey of Canada (data in a recoded), IWA - Interim Waste Authority, MOE - Ministry of Environment water well records.

a


100
90
80
70
60
50
40
30
20
10
0
DH-Nob
DH-V-158
DH-V-110
DH-V-89
DH-P-14
DH-P-17
DH-P-20
DH-P-21
DH-P-28
DH-P-29
DH-P-4
Gormley
Sg
Sr
Sa
Gr
Di
Cl

b

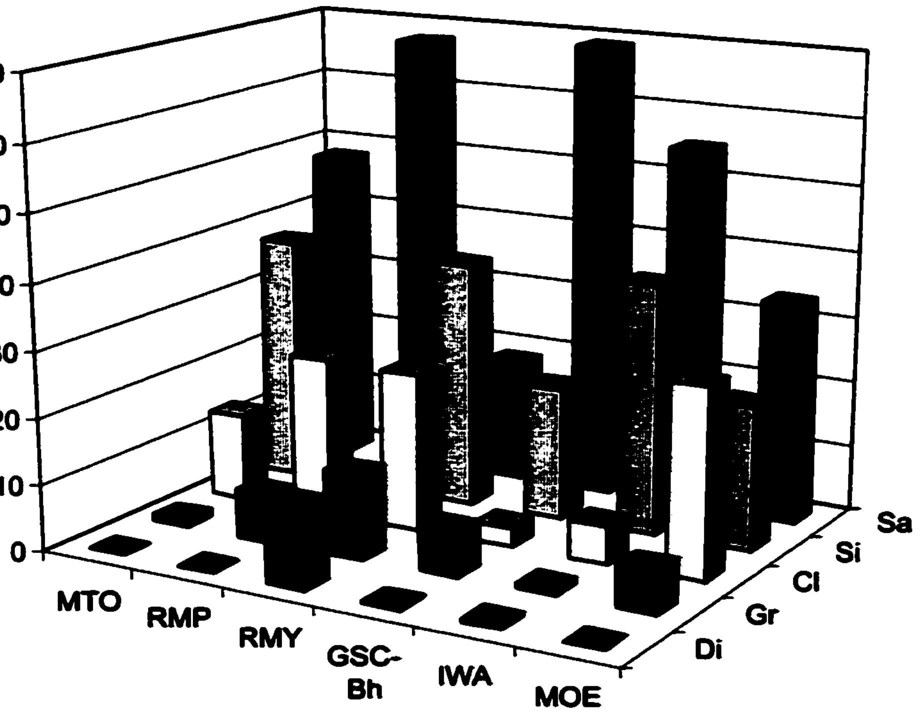

70
60
50
40
30
20
10
0
MTO
RMP
RMY
GSC-
Bh
IWA
MOE
Sa
Si
Cl
Gr
Di

Fig. 5.10. Photos of gravel subfacies. (a) Poorly sorted cross-stratified gravel. Paleoflow was obliquely from left to right. Scale is 8.5 cm long, squares at top of scale are 1 cm. (b) plane-bed gravels with basal open-framework pebble, upward-fining beds, clast clusters, and amalgamated coarse sand beds. Paleoflow was from left to right. (c) Massive heterogeneous gravel with sand intraclast. Paleoflow was from left to right. Intraclast is ~ 2 m long. (d) Sandy gravel with sand and silt intraclasts, reverse-normal matrix grading, and fine grained matrix envelopes on larger intraclasts. Paleoflow was from left to right. Intervals on metre stick are 10 cm.

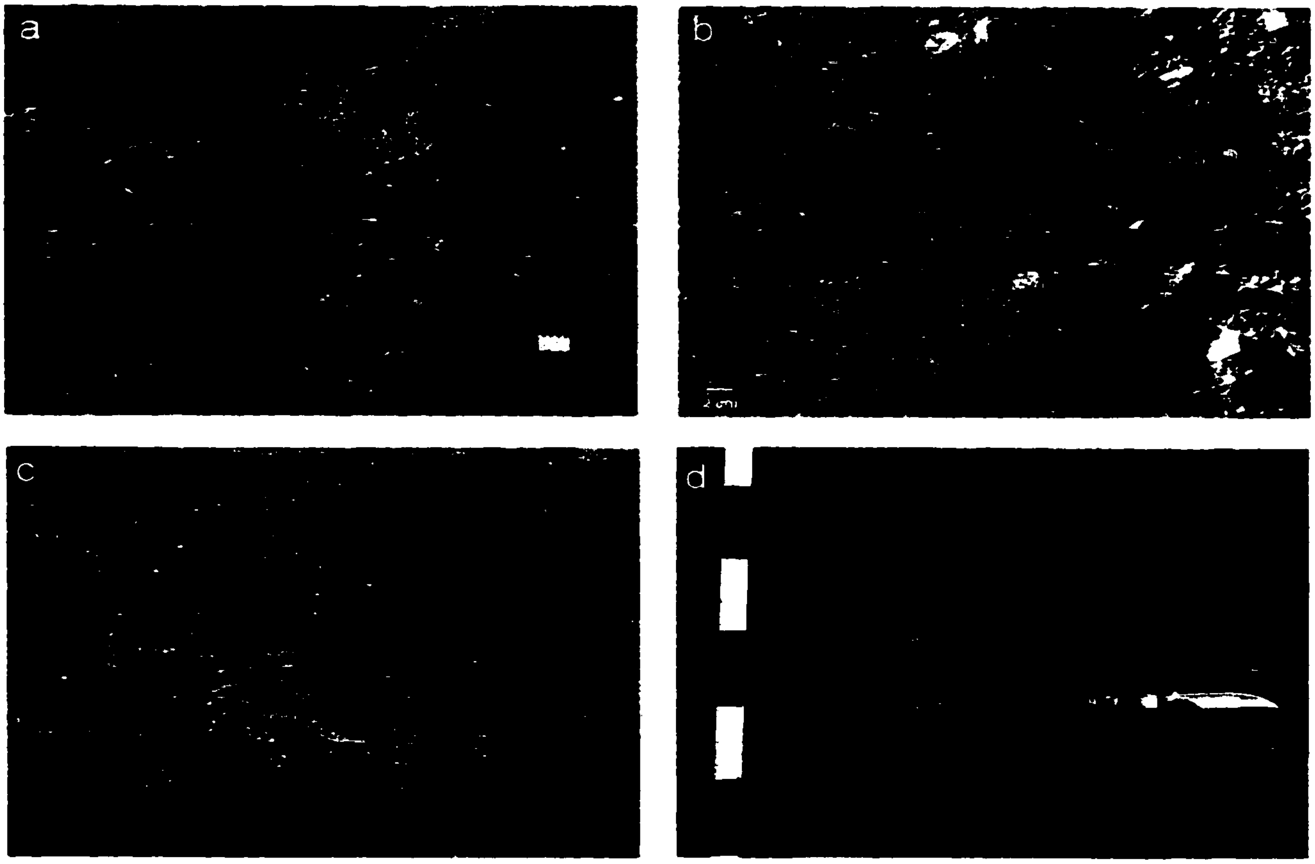


Fig. 5.11. Photos of sand facies. (a) Planar cross-beds overlain by small-scale cross-laminations. Paleoflow is from right to left. Notebook is 17 cm long. (b) Climbing medium scale cross-stratification. Paleoflow is from left to right. Pencil is 15 cm long (c) Low-angle cross-stratification of the quasi-planar-laminated subfacies. Paleoflow is obliquely out of the photo from right to left. Metre stick with 10 cm intervals (d) Steep-wall scour infilled with diffusely-laminated sand. Paleoflow is out of the

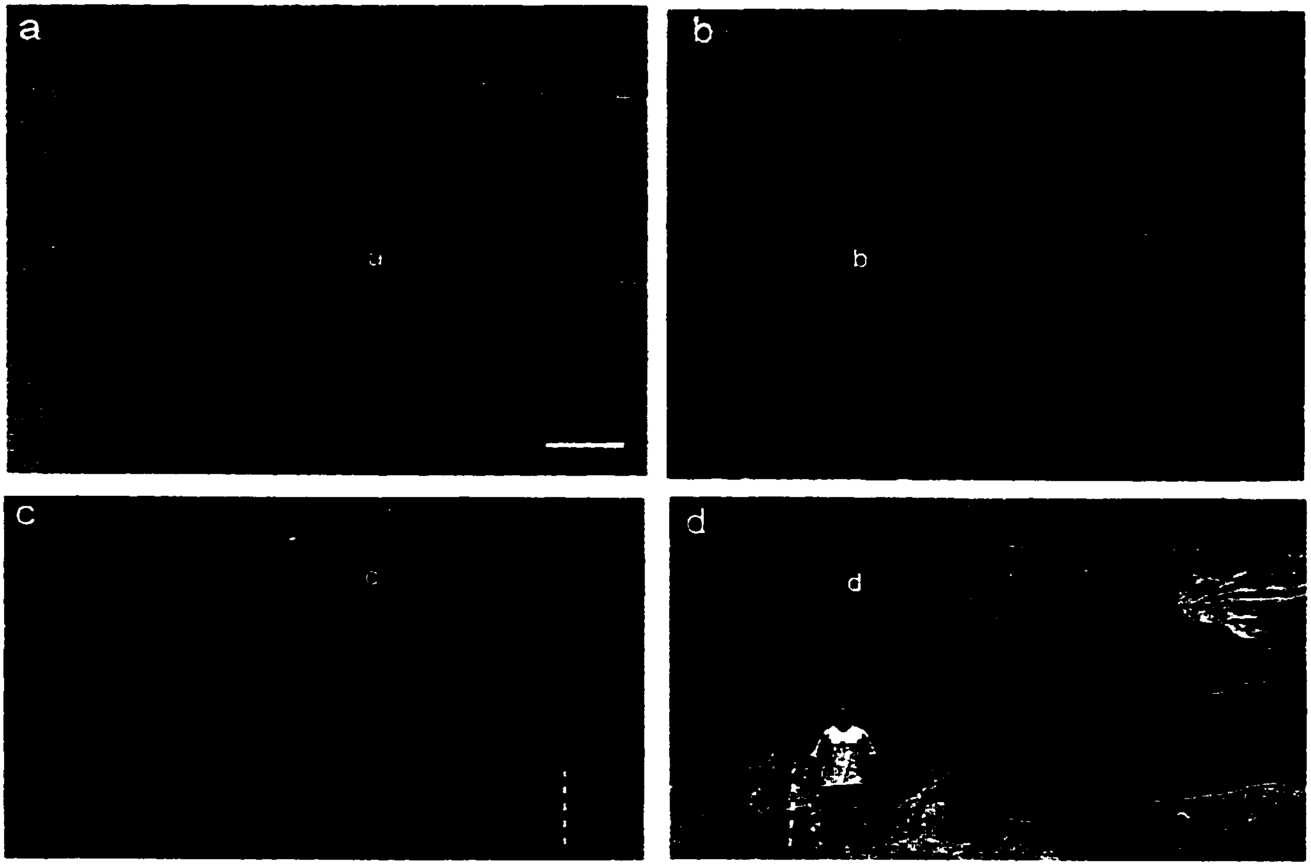

a
b
c
d

Fig. 5.12. Grain size analysis (a) Cross-stratified sand facies. (b) Diffusely-graded / massive sand. (c) Small-scale cross-laminated facies. (d) Graded fine-sand to silt facies. (e) Clay facies.

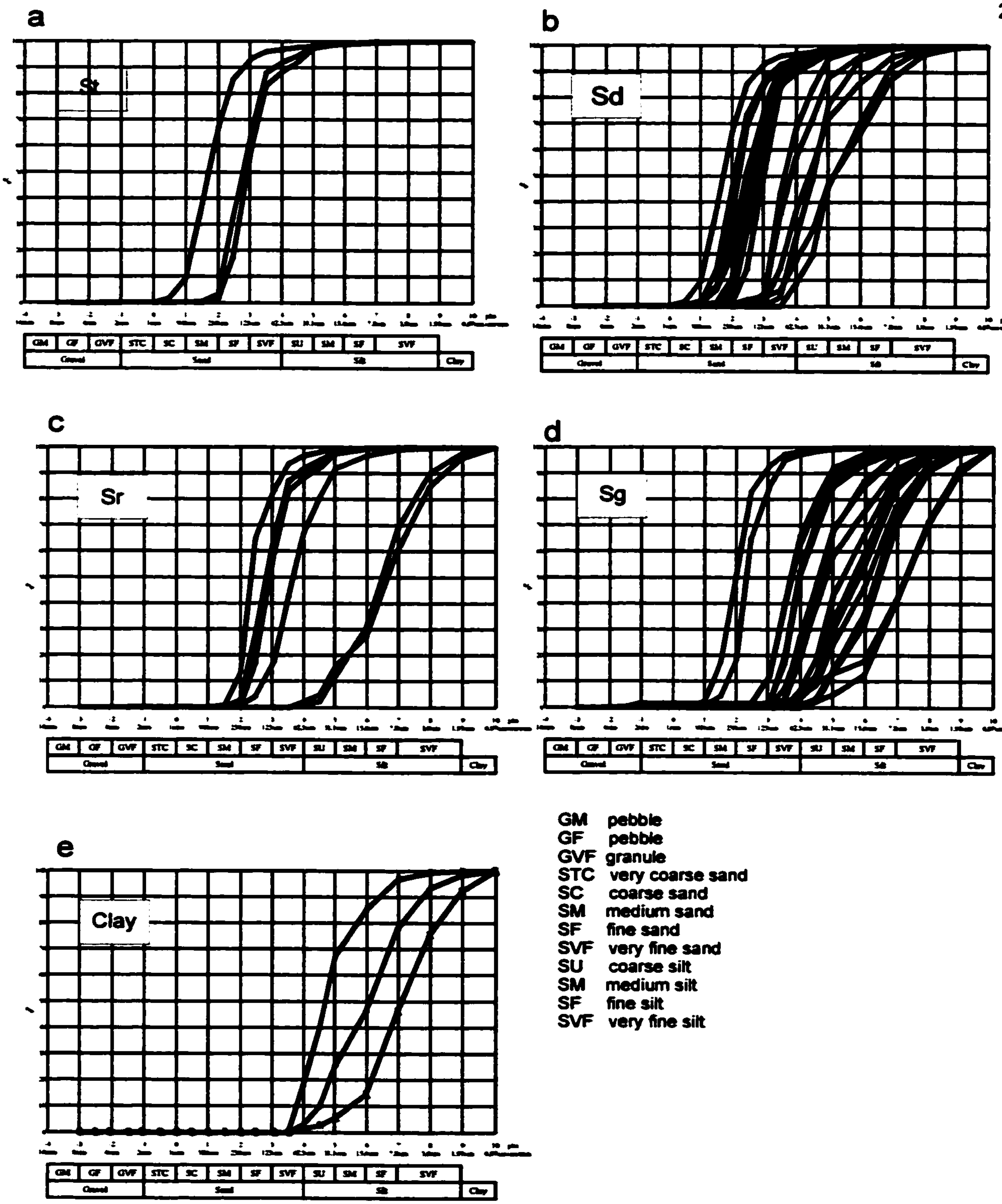

a
St
b
Sd
c
Sr
d
Sg
e
Clay
GM GF GVF STC SC SM SF SVF SU SM SF SVF
Gravel Sand Silt Clay
GM pebble
GF pebble
GVF granule
STC very coarse sand
SC coarse sand
SM medium sand
SF fine sand
SVF very fine sand
SU coarse silt
SM medium silt
SF fine silt
SVF very fine silt

Fig. 5.13. Photos of small-scale cross-laminated facies. (a) 2.5-3 m thick coset of climbing small-scale cross-laminations. Paleoflow direction is from right to left. Metre stick is in 10 cm intervals. (b) Upward-fining climbing stoss-erosional to stoss-depositional fine-sand cross-laminae to graded silt. Succession is overlain by a clay laminae. Paleoflow direction is from left to right. Scale intervals are 10 cm long. (c) Climbing small-scale cross-laminations in fine sand. Paleoflow direction is from right to left. Coin is ~2 cm. (d) Climbing stoss-erosional cross-laminae. Paleoflow direction is from left to right. Pencil is 14.5 cm long.

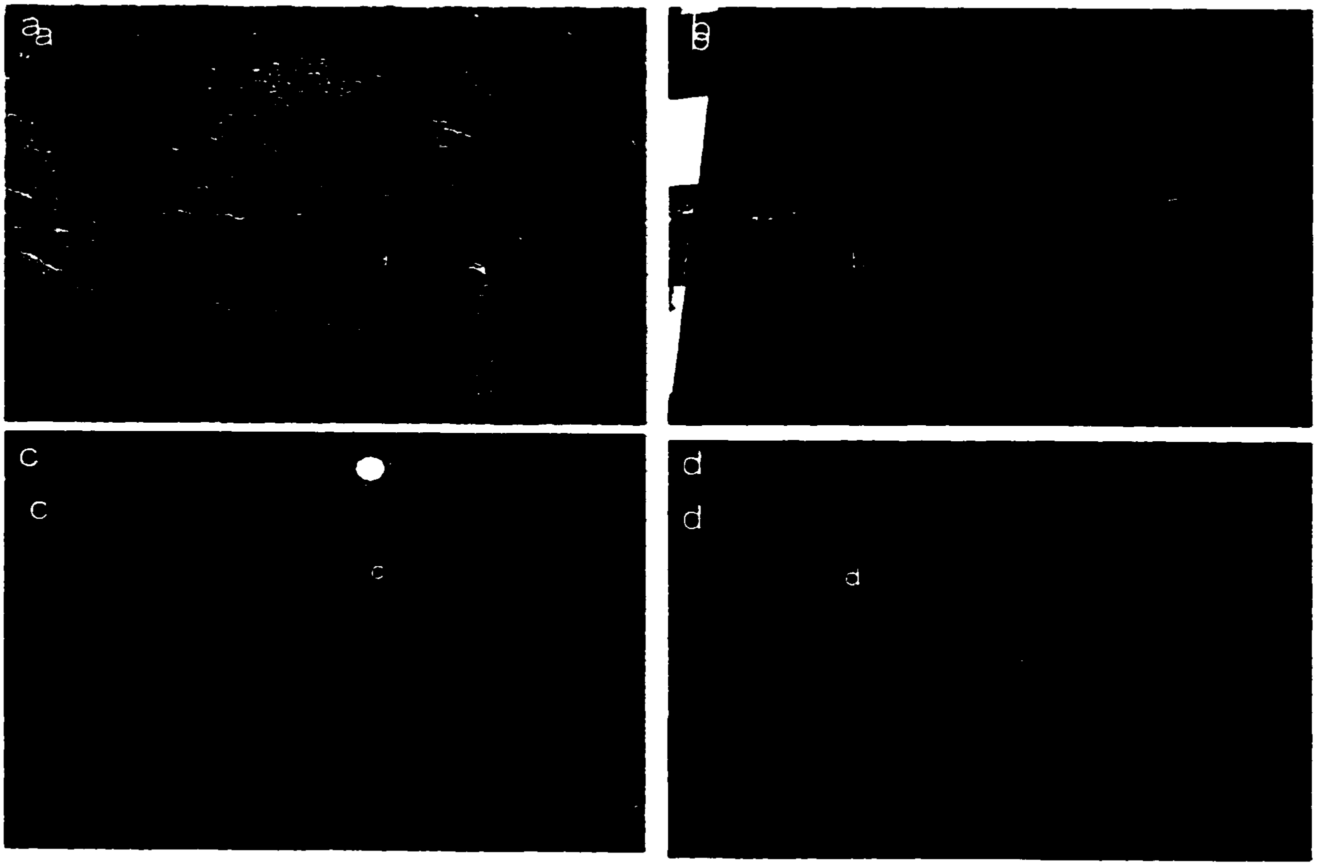

a
b
c
d

Fig. 5.14. Photos of graded fine-sand to silt and clay facies. (a) normal-graded fine sand to silt with minor coring induced deformation. (b)normal-graded fine sand to silt at site PG. (c) micro-laminae of normal-graded fine-sand to silt facies. (d) bioturbated silt and overlying clay facies (e) thick clay strata overlying normal-graded fine-sand to silt facies.

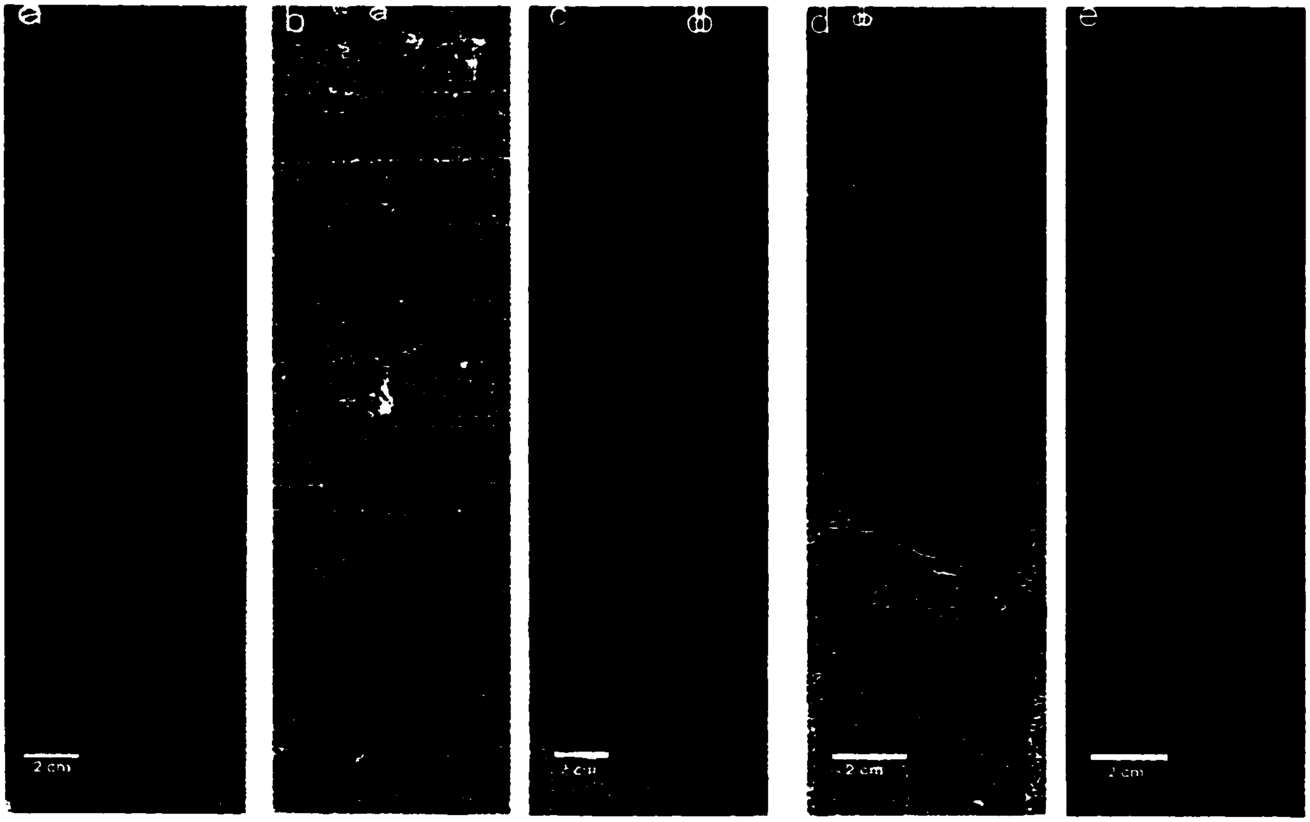

a
b
c
d
e
2 cm
2 cm
2 cm
2 cm
2 cm

Fig. 5.15. Six plots of gravel by 10 m elevation range from 100 to 400 m elevation. Note relatively small amount of gravel in tunnel channels. Note that data are plotted on a colour-coded elevation surface of the lower Oak Ridges contact.

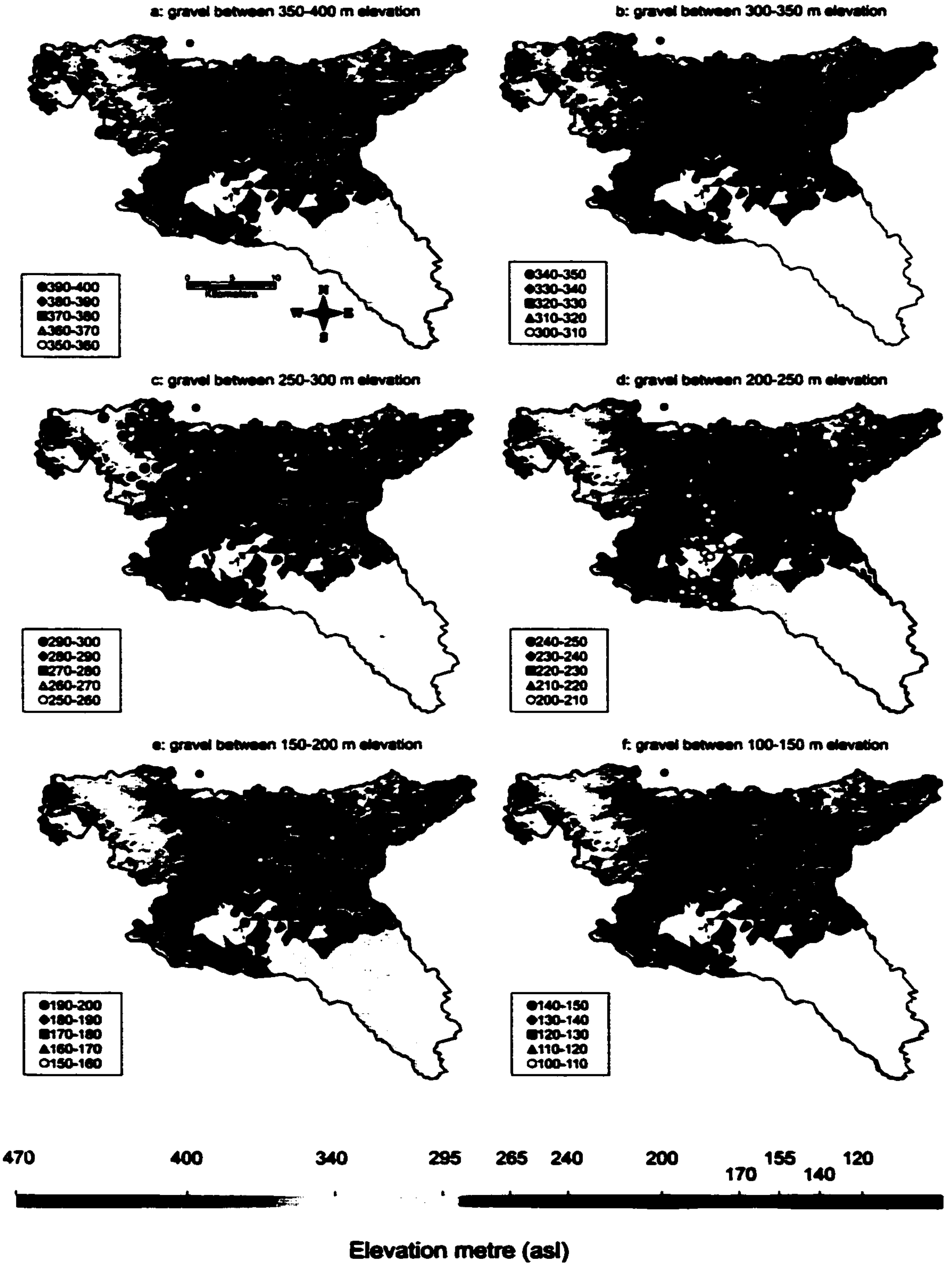

a: gravel between 350-400 m elevation
390-400
380-390
370-380
360-370
350-360
Kilometers
N
W
E
S
b: gravel between 300-350 m elevation
340-350
330-340
320-330
310-320
300-310
c: gravel between 250-300 m elevation
290-300
280-290
270-280
260-270
250-260
d: gravel between 200-250 m elevation
240-250
230-240
220-230
210-220
200-210
e: gravel between 150-200 m elevation
190-200
180-190
170-180
160-170
150-160
f: gravel between 100-150 m elevation
140-150
130-140
120-130
110-120
100-110
470
400
340
295
265
240
200
170
155
140
120
Elevation metre (asl)

Fig. 5.16. Varve diagram for core data showing varve thickness and varve count from top down. Arrows indicate probable correlative varves. Note thick varves in C34-20 as compared to C34-17.

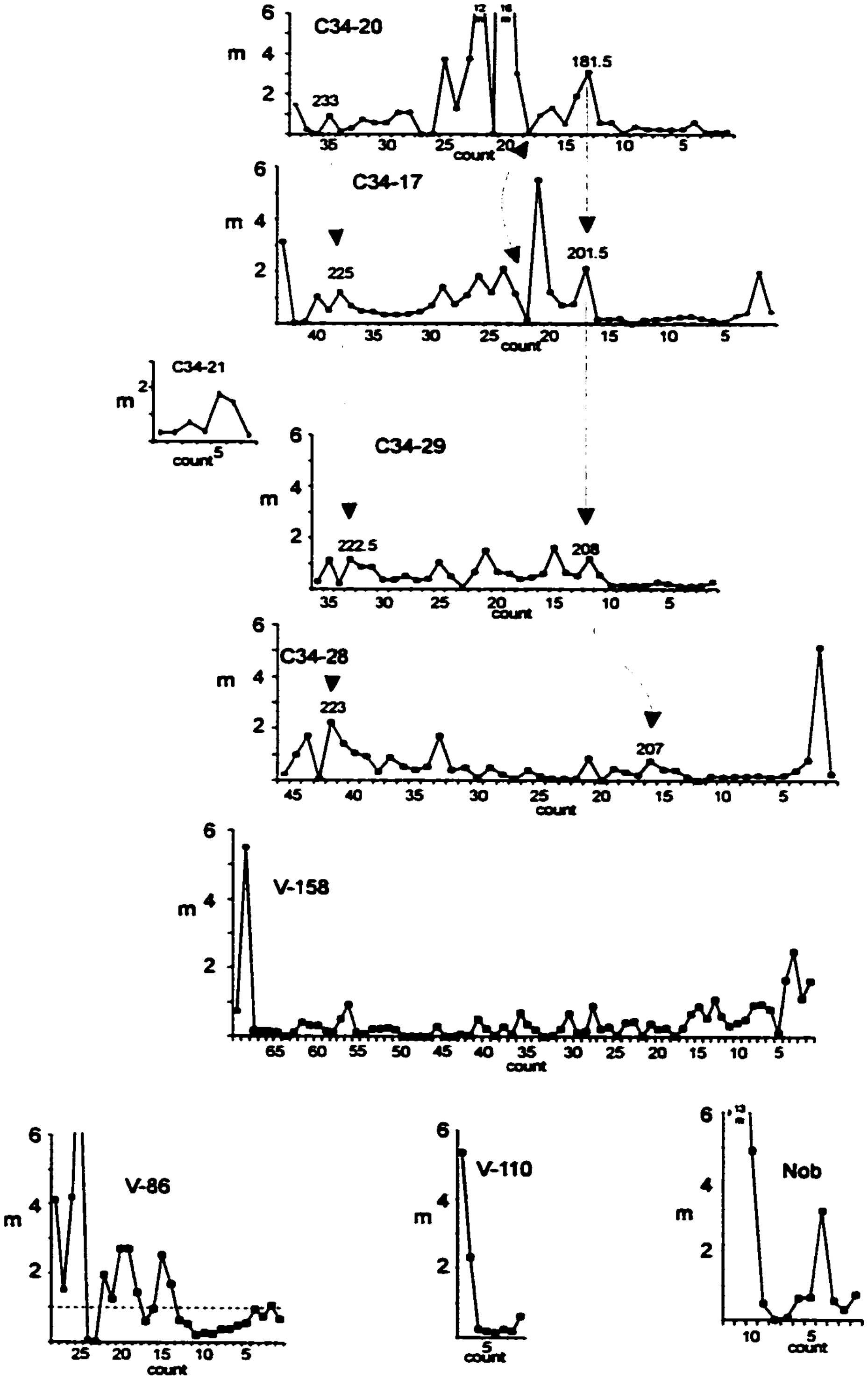

C34-20
181.5
233
count
C34-17
225
201.5
count
C34-21
count
C34-29
222.5
208
count
C34-28
223
207
count
V-158
count
V-86
count
V-110
count
Nob
count
m

Fig. 5.17. Line diagrams outlining the depositional history of the Oak Ridges Moraine. (a) Sculpting of Newmarket Till into drumlins by regional sheet floods. (b) Tunnel channel erosion by episodic jokulhlaup floods. (c) Stage I channel fill showing diffusely-graded sand along channel axis and gravel deposits along channel margin. (d) Stage II rhythmite deposition from seasonal meltwater discharge into subglacial flooded tunnel channels. (e) Outline of the proglacial basin scenario for stage III sedimentation into the vicinity of the moraine ridge. (f) Subaqueous fan sedimentation during stage III from combined seasonal and jokulhlaup discharge from subglacial conduits.

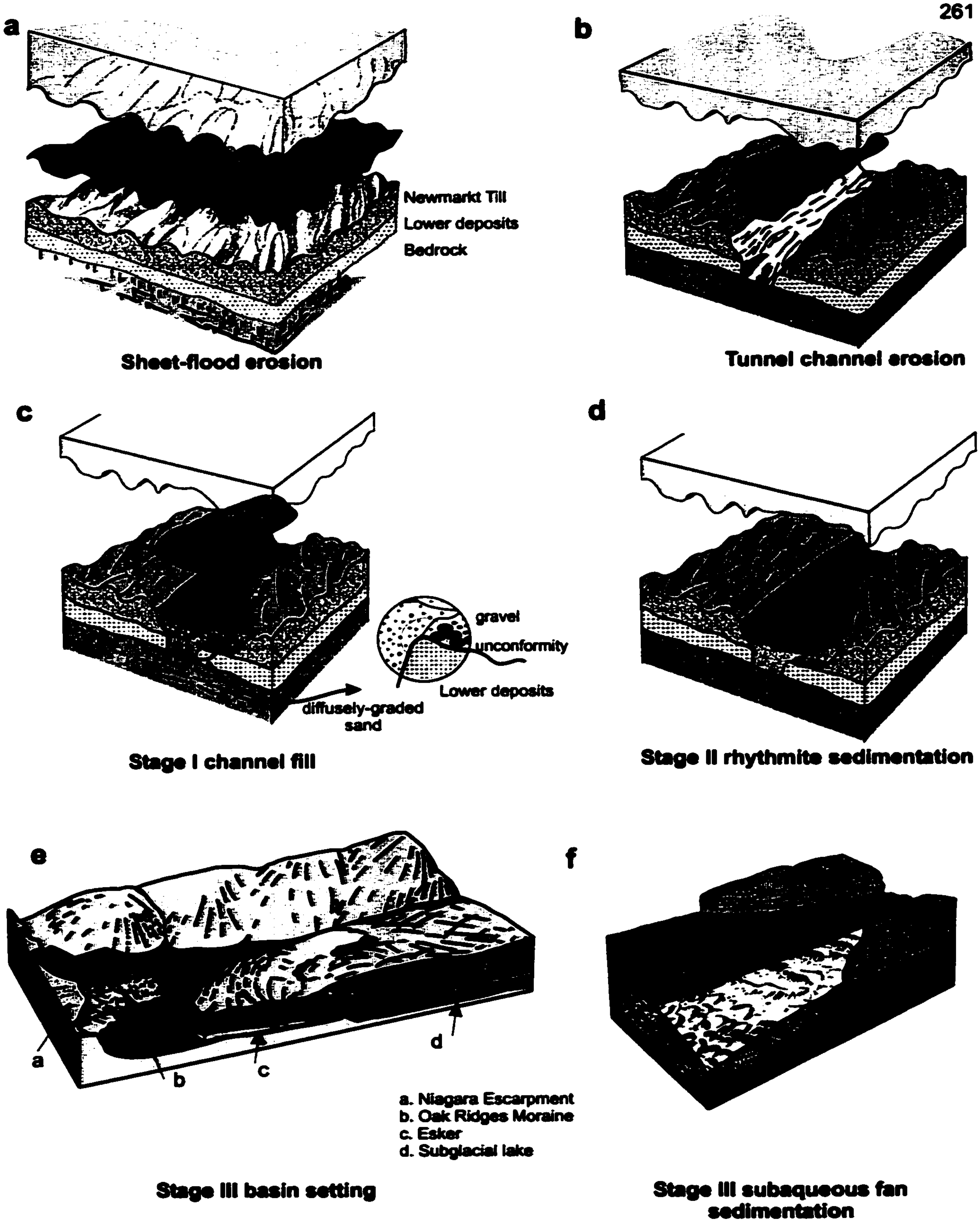

a
Newmarkt Till
Lower deposits
Bedrock
Sheet-flood erosion
b
Tunnel channel erosion
c
gravel
unconformity
Lower deposits
diffusely-graded sand
Stage I channel fill
d
Stage II rhythmite sedimentation
e
a
b
c
d
a. Niagara Escarpment
b. Oak Ridges Moraine
c. Esker
d. Subglacial lake
Stage III basin setting
f
Stage III subaqueous fan sedimentation

Table 5.1: Summary of diamicton and gravel lithofacies in the western Oak Ridges Moraine

Facies	Sub-facies	Texture	Sedimentary Structures	Interpretation
Diamicton		dispersed clasts in a silt to sand-silt matrix	sharp or gradational base, massive,	
Gravel G	cross-bedded[1] Gp, Gt	sandy pebble to pebble gravel	planar and trough cross-beds, sharp based, rare faults	from bed-load sheets and dunes, scour fill
	horizontal stratified[1] Gh	heterogeneous to well sorted	sharp based, graded beds, open framework to matrix supported, clast clusters	from bed-load sheets under critical flow conditions
	heterog-eneous[1] Gd	heterogeneous pebble gravel	normal and reverse graded, clast clustering, faint horizontal bedding, matrix or clast supported, intraclasts of sand	from hyperconcentrated dispersions, related to hydraulic jump scour fills

Table 5.2: Summary of sand lithofacies in the western Oak Ridges Moraine

Facies	Sub-facies	Texture	Sedimentary Structures	Interpretation
Medium-coarse sand S	cross - bedded[1] Sp, St	medium coarse sand, rare pebbly sand	sharp based, rare convolute bedding, minor faults	by 2-D and 3-D dunes and macroforms
	quasi - planar[1] Sh	medium sand	sharp based, minor undulation, horizontal to low angle dip, multiple internal erosional surfaces	antidunce and chute and pool conditions of supercritical flow
	diffusely graded[1] Sd	medium sand, minor intraclasts, and granules	sharp based, irregular scours, massive to bedded, amalgamated beds, sand intraclasts, water escape structures	from hyperconcentrated dispersions

Table 5.3: Summary of sand-silt and clay lithofacies in the western Oak Ridges Moraine

Facies	Sub-facies	Texture	Sedimentary Structures	Interpretation
Small-scale cross-laminated sand Sr		silt and fine sand	sharp based, stoss-erosional to stoss-depositional, minor micro-faults	waning flow and combined suspension/bed-load transport
Graded fine sand Silt Si		silt and fine sand	lower contact sharp and graded, fining upward, micro-laminae and beds < 5 cm thick	density underflows
Clay Cl		silt and clay	gradational lower contact, normal graded, minor bioturbation	basinal underflow and suspension sedimentation

Table 5.4. Four stage depositional model from Barnett et al. (1998) expanded and refined for the western Oak Ridges Moraine, Humber River Watershed. NS - not studied

Stage		Position	Setting / Environment	melt-water input	Q	Description	Depositional Process
4 (NS)		subglacial - proglacial	glacilacustrine	lateral	T	Halton Till (not studied)	ice-marginal - glacilacustrine
3	b	subglacial - proglacial	delta	lateral / axial	T	not observed in the study area	
3	a	subglacial - proglacial	subaqueous fan	lateral / axial	T	massive, plane-bed, cross-bedded gravel steep-walled scours, diffusely graded sand, low angle cross-stratified sand, planar and trough cross-bedded sand	plane-wall jet efflux, hydraulic jump scouring and infill
2	b	Subglacial	basin	lateral / axial	S	varves of graded sand, silt and clay, upward-fining succession of 60-70 varves	turbidity currents and suspension sedimentation
2	a	Subglacial	subaqueous fan	lateral	T	small-scale cross laminae, graded sand silt, thick varves,	plane-jet and plane-wall jet
1	b	Subglacial	tunnel channels - north-south	lateral	J	massive-diffusely graded sand,	Hyperconcentrated flows along tunnel channels
1	a	Subglacial	tunnel channels - north-south	lateral	J	low relief gravel mesoforms and dunes	tractional bedload deposits

Q = discharge, S = seasonal meltwater discharge, J = Jokulhlaup discharge, T = transitional or composite.

CHAPTER VI

Summary and Conclusion

6.1 Introduction

This thesis has described the lithofacies and stratal geometry of the Oak Ridges Moraine in southern Ontario and addressed the three objectives set out at the beginning of the thesis:

i) *define and interpret the lithofacies of channel fill sediment beneath the moraine,*

ii) *detail the lithofacies and stratal geometry of a subaqueous fan deposit in the moraine,*

iii) *integrate objectives i and ii into an understanding of the meltwater discharge flux responsible for formation of the moraine.*

Additionally, the study has presented a detailed depositional model of moraine genesis for the western Oak Ridges Moraine. The following sections summarize and discuss the principal chapters and the implications of this study for the Oak Ridges Moraine and for moraine genesis in general.

6.2 Tunnel Channel Erosion and Fill (Chapter 3)

6.2.1 Channel fill

Two buried tunnel channels beneath the western Oak Ridges Moraine are infilled with gravel and diffusely-graded / massive and cross-stratified sand. Gravel is ~ 17 m thick and consists of dune-scale mesoform and low-relief gravel bedload sheet deposits emplaced by turbulent tractional flows prior to an episode of significant channel deepening. In deeper areas of tunnel channels the fill consists predominantly of diffusely-graded / massive and cross-stratified sand in the lower 50 m of a 100 m deep channel. The channels are interpreted to have developed following collapse of a regional sheet flood that formed drumlins and subsequently eroded a regional tunnel channel network (Brennand and Shaw 1994; Shaw 1996). Channels were overdeepened of developed by differential erosion of Newmarket Till, and unroofing of less consolidated silt and sand strata. Exposure of these sediment allowed the rate of local channel incision to increase significantly. In addition, high-pore fluid pressure in sand beds caused horizontal groundwater movement toward low-pressure zones that coincided with the scours. This groundwater flow initiated sediment piping and transport of sand into the channel. This sediment became part of the sediment load entrained by the turbulent flow. Sedimentation proceeded rapidly downflow of the grounding line of the subglacial lake where a hydraulic jump occurred as the flow expanded and went from supercritical to subcritical. Nonuniform unsteady flows deposited inverse-normal graded or massive sand strata during periods of rapid, voluminous sedimentation and cross-stratified sand was deposited during periods of low discharge and low sediment flux. Stratigraphically upward, the transition to sand-silt-clay rhythmites indicates a decrease in fluid and sediment discharges, probably due to a substantial reduction in meltwater production following the end of the jökulhlaup discharge and a return to seasonal, diurnal conditions.

6.2.2 Implications for buried tunnel channels

Tunnel channels have a broad distribution across North America and northwestern Europe in both continental and offshore regions (Cofaigh 1996). In many areas the channel fills have not been investigated or have been studied primarily by a variety of reflection seismic techniques (e.g. Boyd et al. 1988; Mullins, et al. 1996). Where drilling has been completed, it commonly has not penetrated the lowest part of the channel fill succession. In many areas these basal strata have similar seismic facies as those recently described by Pugin et al. (1999). The diffusely-graded / massive and cross-stratified sand facies association of this study is correlated with the low-amplitude reflectors and chaotic seismic facies described by Pugin (1999) from the thick, deeply buried channel fill west of DH-Nob. The sand facies association is interpreted to be analogous to comparable low-amplitude seismic reflector facies reported from deep channel fills in eastern North America (Boyd et al. 1988; Mullins et al. 1996) and the North Sea (Wingfield 1990). For the first time this correlation provides direct sedimentological evidence for the origin of this common but previously enigmatic, channel fill seismic facies.

6.3 High-energy Subaqueous Fan Processes (Chapter 4)

6.3.1 Plane-wall jet model

The discharge of a jet into a standing water body has been extensively studied and applied to deltas (Bates, 1953) and subaqueous fans (Powell, 1990). Most of this work has focused on the plane-jet efflux with little discussion of plane-wall jet efflux, except in laboratory flow experiments (Rajaratnam and Subramanyan 1986) and much less commonly in subaqueous fan models (Gorrell and Shaw 1991). A key element of the plane-wall jet efflux is the hydraulic jump which forms at the rapid transition from supercritical to subcritical flow. Due to the intense turbulence generated by the hydraulic jump, the plan geometry of the plane-wall jet efflux is likely to be similar in shape to a plane-jet efflux with high interfacial shear. The longitudinal facies distribution, however, is likely to be markedly different due to the rapid change in flow characteristics at the hydraulic jump and the consequent change in sediment transport competence. Eight facies and five facies associations form the subaqueous fan deposit of an inferred plane-wall jet efflux at the Gormley site. Deposition is interpreted to have occurred within a single, quasi-continuous jet efflux that in a streamwise direction had three distinct zones of flow development: i) Zone of Flow Establishment (ZFE), ii) Zone of Flow Transition (ZFT), iii) Zone of Established Flow (ZEF, Fig. 4.1). A coarse fan-core composed of massive - plane-bed gravel facies association makes up the ZFE. This is flanked laterally and downflow by quasi-planar-stratified and diffusely-graded sand facies of the ZFT. Rapid changes in lithofacies suggest that flows changed from hyperconcentrated flows to dilute fluidal flows over distances of only 2-5 m. This rapid transition was caused by rapid flow expansion at the mouth of the conduit and resultant loss of transport competence. Within the ZFT, the steep-walled scour infill association of diffusely-graded / massive infill was deposited rapidly downflow of the hydraulic jump that formed as the flow changed from supercritical to subcritical. Here rapid suspension sedimentation was due to a loss of flow capacity and development of hyperconcentrated dispersions that deposited sediment rapidly from a laminar shear layer and/or traction carpets. Downflow the transition to the ZEF is marked by accreting foresets (master beds) that form set boundaries for medium-scale planar cross-bedded sand deposited by 2-D dunes. Deposits fine downflow along these large foresets and eventually grade into small-scale cross-laminated fine sand. Waning jet discharge is recorded by the overlying shallow-channel association of cross-stratified medium sand and small-scale cross-laminated fine sand deposited by a subaqueous braided system that incised into the upper part of the fan deposits.

The distinctive downflow organization of lithofacies is interpreted to have been produced by a plane-wall jet efflux. All gravel and pebbly sand occur upflow of the region of the steep-walled scour and diffusely-graded sand infill, deposits interpreted to define a region of flow-transition from supercritical to subcritical conditions and a hydraulic jump. Downflow of this area only subcritical fluidal flow deposits occur. Consequently, it is interpreted that the hydraulic jump formed an extremely efficient sorting mechanism that caused all sediment coarser than ~2 mm to be deposited immediately, whereas finer sediment continued to be transported farther basinward.

6.3.2 Hydraulic jump scours

Steep-walled scours infilled with diffusely-graded sand have been traditionally considered a characteristic facies association of subaqueous fans (Rust 1977). This association has previously been interpreted commonly to be the result of sediment slumping (Cheel and Rust 1982; Postma et al. 1983; Rust 1988) and deposition by grain flows (Cheel and Rust 1982; Postma et al. 1983). Alternatively, Gorrell and Shaw (1991) suggested that similar deposits were the product of hydraulic jump scouring. The mechanics and downflow facies transitions from hydraulic jumps have been studied in submarine fan-turbidite systems(Daub 1996; Garcia 1993; Hand 1974; Komar 1971). These studies, however, have provided little insight into stratal characteristics within the hydraulic jump region, although hydraulic jumps have been intensively studied by engineers (e.g. Olufemi 1996; Rajaratnam 1981). The intensity of scouring is closely related to the flow energy upflow of the hydraulic jump and consequently scouring is likely to be most extensive under high energy direct hydraulic jumps. Four aspects of the steep-walled facies association described from the Gormley site are interpreted to be evidence for a hydraulic jump origin. Firstly the abrupt vertical facies transition from planar-laminae or shallow-dipping strata to diffusely-graded scour fill sand to cross-stratified and small-scale cross-laminated sand suggests a rapid evolution in flow conditions from critical or supercritical antidune flow to hydraulic jump scouring and infill to subcritical traction flow deposition. Secondly, deposits adjacent to the scour margin show little or no evidence of deformation related to sediment failure; the single exception being the presence of a locally slumped or partially liquified sediment along a ~11 m section of the scour margin. This suggests that slumping was not a significant mechanism in forming the scours. Thirdly, the well defined sediment sorting in the streamwise direction of the fan (see above). Finally, the locally steep-walled form with rare overhangs suggests a turbulent erosional mechanism and rapid, near simultaneous infill. On this basis, the data are interpreted to support the plane-wall jet and hydraulic jump model first proposed for some other steep-walled scour and fill deposits by Gorrell and Shaw (1991).

6.4 Implications for the Oak Ridges Moraine

The Oak Ridges Moraine has been traditionally interpreted to be an interlobate moraine. This interpretation has been largely based on landform analysis and observation in shallow, surface outcrops

(Chapman and Putnam 1984). Where previous studies have integrated borehole records (Duckworth 1979) the data commonly lacked adequate resolution to resolve detailed sedimentological characteristics. Furthermore, other studies with access to continuous core and hence detailed sedimentological descriptions (Fenco MacLaren 1994; Gilbert 1997; Golder and Associates 1994) were site specific and in most cases relied heavily on the existing geological framework (e.g. Paterson and Cheel 1997). More recently a more complex polygenetic origin has been proposed for the moraine that includes discrete episodes of subglacial glacifluvial, subaerial ice-bounded glacilacustrine, or locally subaerial glacifluvial deposition (Barnett et al. 1998). In this model meltwater flow to the moraine involved both lateral and axial pathways. Furthermore, Shoemaker (1999) has recently hypothesized that during deposition of the moraine, the Lake Ontario basin contained a subglacial lake, and that the Oak Ridges Moraine may have formed subglacially along the margin of the lake. Additionally, as indicated by Barnett et al. (1998), the moraine overlies a regional unconformity that was probably formed by large-magnitude Late Wisconsin flood discharges that formed regional drumlin fields (Shaw and Gilbert 1990; Shaw and Sharpe 1987) and later tunnel channels (Barnett et al. 1998; Brennand and Shaw 1994). With these new developments in mind how does the sedimentological data presented in this study support or refute these various perspectives of moraine genesis?

6.4.1 Meltwater discharge

In this study the most complete record of sedimentation and characterization of the depositional environments in the Oak Ridges Moraine is provided by DH-V-158. This core intercepts 158 m of upper Wisconsin sediment of which ~130 m is interpreted to be Oak Ridges Moraine. Regional analysis indicates the core occurs within a northeast-southwest trending tunnel channel. Three distinct stages of sedimentation have been recognized, each corresponding to one of the stages in the depositional moraine model described in chapter V. Each stage is interpreted to record changes in meltwater and sediment flux to the moraine. The basal 50 m of moraine sediment consists of the diffusely-graded /

massive and cross-stratified fine sand association (Stage I). These strata are gradationally overlain by the varve association consisting of distinct upward-fining successions of small-scale cross-laminae and the graded fine-sand to silt facies overlain by thin clay strata (Stage II). At DH-V-158 the varve association forms an upward-thinning and fining succession of ~70 varves. These varves are in turn overlain by an upward coarsening and thickening varve succession (Stage III). These strata are overlain by a succession of small-scale cross-laminae and graded fine-sand to silt with only rare clay laminae (Stage III). As a result, on the basis of lithofacies and lithofacies associations, the meltwater-sediment flux is interpreted to have changed from waning subglacial jökulhlaup discharge (Stage I) to seasonal, diurnal discharge (Stage II) and back to a mixed regime of minor jökulhlaup and seasonal meltwater discharge (Stage III). The interpretation of jökulhlaup discharge during stage III is supported by the high-energy deposits at the Gormley site. At this site, high energy deposits of supercritical flow and hyperconcentrated dispersions were emplaced rapidly and locally incised steep-walled scours in unconsolidated sand (see section 6.3). Thus, the moraine is interpreted to have formed in response to a complex set of sedimentological variables that changed significantly in both space and time related to: changes in the ice-sheet profile, modification of the ice-sheet hydrologic profile and pathways, reservoir storage, episodic discharge, reservoir recharge, seasonal meltwater flux, and possibly also large precipitation events.

6.5 General Implication to Laurentide Landforms and Deglaciation

6.5.1 Moraine genesis

The genesis of ice-contact landforms, of which moraines are one element, has long been based on landform analysis and analysis of regional ice-flow indicators. In addressing this problem, Warren and Ashley (1994) specifically highlight the need for integrated landscape and sedimentological studies to better understand eskers and moraines. Using examples from Ireland, they noted that a purely descriptive and mainly geomorphological approach to interpreting ice-contact ridges has led to problems in understanding their origin and the implied glacial history. In Canada there are a number of examples of conflicting interpretations with respect to some Laurentide ice-sheet landforms. Example include the Harricana moraine or esker (Brennand and Shaw 1996; Veillette 1986) and the Leaf Rapids interlobate moraine (Kaszycki and DiLabio 1986) or esker (Ringrose 1982). This problem is further highlighted by studies of the Oak Ridges Moraine. Detailed sedimentological studies provide critical insight into moraine genesis that other techniques are unable to resolve, notably: the nature of depositional meltwater discharge, paleoflow directions of meltwater discharge, the polygenetic nature of moraines and associated depositional environments. This and other recent studies of the Oak Ridges Moraine have contributed much to resolving the origin of the Oak Ridges Moraine. Variations in meltwater flux have been interpreted based on deep, continuous core and description of vertical outcrop sections. In this study, deposits emplaced by variations in discharge between jökulhlaup and seasonal diurnal flow conditions have been described. Paleoflow measurements have constrained interpretations of meltwater discharge to the moraine. In the Humber River watershed paleoflow data indicate flow from both the north and south along with probably axial flow from the east during moraine deposition. Analysis of lithofacies-associationsallowed identification of a variety of depositional environments and processes during moraine formation. Parts of the moraine have been deposited as tunnel channel fills, eskers, subaqueous fans, turbidite basinal deposits and low-energy suspension deposits. By analogy, other less completely studied moraines may also consist of similar depositional elements and have been deposited by variable meltwater regimes.

6.5.2 Moraine Composition

As introduced in chapter I, the composition of moraines is commonly misunderstood due to the shallow depth of investigations (Karrow 1973) or due to inappropriate moraine analogues derived from modern alpine or arctic glaciers. Through the combined description of continuous core and shallow outcrop sections it has been possible to characterize the lithofacies variability of the moraine. Analysis of the lithofacies indicates that the moraine is composed predominantly (~70%) of small-scale cross-stratified fine sand and graded fine-sand to silt. Additionally, the rapid nature of lithofacies transitions, not only lateral to flow, but also longitudinally has been documented. Downflow lithofacies successions may change from 75 % gravel to 90 % fine sand over distances of ~100 m.

6.6 Conclusion

Moraines of the Laurentide ice-sheet offer considerable research opportunity for improved understanding of glacial deposits and deglacial Laurentide ice-sheet processes. Large Laurentide moraines have been shown to consist of a diverse suite of lithofacies deposited by meltwater discharges of highly variable energy. Deposits of similar nature have more commonly been interpreted from deep-marine submarine-fan deposits rather than ice-marginal environments. Consequently, the high-energy and hyperconcentrated deposits of ice-marginal landforms are of considerable sedimentological interest as their unconsolidated nature provides the opportunity to study such deposits using techniques generally unsuitable for lithified sediment. Furthermore, this study demonstrates the value of detailed sedimentological research in conjunction with traditional investigative techniques of glacial landscapes for the continued advancement of deglacial processes.

References

Albertson, M.L., Asce, J., Dai, Y.B., Jensen, R.A., Rouse, H., and Asce, M., 1948, Diffusion of submerged jets: Transactions American Society of Civil Engineers, v. 115, p. 639-667.

Anderson, M.P., 1989, Hydrogeologic facies models to delineate large-scale spatial trends in glacial and glaciofluvial sediments: Geological Society of America Bulletin, v. 101, p. 501-511.

Aravena, R., and Wassenaar, L.I., 1993, Dissolved organic carbon and methane in a regional confined aquifer, southern ontario, Canada: Carbon isotope evidence for associated subsurface sources: Applied Geochemistry, v. 8, p. 483-493.

Arnott, R.W., and Hein, F.J., 1986, Submarine canyon fills of the Hector Formation, Lake Louise, Alberta: Late Precambrian syn-rift deposits of the proto-Pacific miogeocline: Bulletin of Canadian Petroleum Geology, v. 34, p. 395-407.

Arnott, R.W.C., and Hand, B.M., 1989, Bedforms, primary structures and grain fabric in the presence of suspended sediment rain: Journal of Sedimentary Petrology, v. 59, p. 1062-1069.

Ashley, G.M., 1975, Rhythmic sedimentation in glacial Lake Hitchcock, Massachusetts-Connecticut, *in* Goldthwait, R.P., ed., Glacial deposits: Stroudsburg, Pa., United States, Dowden, Hutchinson & Ross, Inc., p. 450-455.

Ashley, G.M., Shaw, J., and Smith, N.D., 1985, Glacial Sedimentary Environments: Tulsa, Oklahoma, Society of Economic Paleontologists & Mineralogists Short Course 16, 246 p.

Ashley, G.M., Southard, J.B., and Boothroyd, J.C., 1982, Deposition of climbing-ripple beds; a flume simulation: Sedimentology, v. 29, p. 67-79.

ASTM, 1994, Standard Test Method for Particle-Size Analysis of Soils (D422-63), 1994 Annual Book of ASTM Standards, sect. 4, vol 04.08.

Baker, V.R., 1978, Large-scale erosional and depositional features of the Channeled Scabland, *in* Baker, V.R., and Nummedal, D., eds., The Channeled Scabland: A Guide to the Geomorphology of the Columbia Basin, Washington; Prepared for the Comparative Planetary Geology Field Conference: National Aeronautics and Space Administration.

Banerjee, I., 1973, Part A: Sedimentology of Pleistocene glacial varves in Ontario, Canada; Part B: Nature of grain-size distribution of some Pleistocene glacial varves in Ontario, Canada: Geological Survey of Canada, Bulletin, v. 226: Ottawa, Ont., Dept. of Energy Mines and Resources, 60 p.

Banerjee, I., and McDonald, B.C., 1975, Nature of Esker Sedimentation, *in* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and Glaciolacustrine Sedimentation: Tulsa, Oklahoma, Special Publication - Society of Economic Paleontologists and Mineralogists, p. 132-154.

Barnett, P.J., 1990, Tunnel valleys: evidence of catastrophic release of subglacial meltwater, central-southern Ontario, Canada: Abstracts with Programs, Northeastern Section, Geological Society of America, Syracuse, New York, p. 3.

Barnett, P.J., 1992, Quaternary geology of Ontario, *in* Thurston, P.C., Williams, H.R., Sutcliffe, R.H., and Stott, G.M., eds., Geology of Ontario: Toronto, Ontario Geological Survey, Special Volume 4, Part 2, p. 1011-1088.

Barnett, P.J., 1997, TRT sand and gravel pit, *in* Sharpe, D.R., and Barnett, P.J., eds., Where is the Water? Regional Geological/ Hydrogeological Framework, Oak Ridges Moraine Area, southern Ontario: Ottawa, Ontario, Geological Association of Canada, Field Trip A1 Guidebook 17-18 May, p. 34-36.

Barnett, P.J., Cowan, W.R., and Henry, A.P., 1991, Quaternary geology of Ontario, southern sheet: Ontario Geological Survey, map 2556.

Barnett, P.J., and Gwyn, Q.H.J., 1997, Surficial Geology of the Newmarket Area, NTS 31 D/3, southern Ontario: Geological Survey of Canada, Open File 3329.

Barnett, P.J., and Kenny, F.M., 1997, Piping, an important process in the postglacial landscape development of southern Ontario: GAC/MAC Program with Abstracts, Geological Association of Canada, Ottawa '97, May 19-21, 1997, Ottawa, Ontario.

Barnett, P.J., Sharpe, D.R., Russell, H.A.J., Brennand, T.A., Kenny, F.M., and Pugin, A., 1998, On the origins of the Oak Ridges Moraine: Canadian Journal of Earth Science, v. 35, p. 1152-1167.

Bates, C.C., 1953, Rational theory of delta formation: American Association Petroleum Geologists Bulletin, v. 37, p. 2119-2162.

Benn, D.I., and Evans, D.J., 1998, Glaciers and Glaciation: New York, Wiley, 716 p.

Bennett, M.R., and Glasser, N.F., 1996, Glacial Geology Ice Sheets and Landforms: Toronto, Wiley, 364p.

Best, J.L., 1987, Flow dynamics at river confluences: implications for sediment transport and bed morphology, *in* Ethridge, F.G., Flores, R.M., and Harvey, M.D., eds., Recent Developments in Fluvial Sedimentology:, Society of Economic Paleontologists and Mineralogists, Special Publication 39, p. 27-35.

Blair, T., and McPherson, J.G., 1994, Alluvial fans and their natural distinction from rivers based on morphology, hydraulic processes, sedimentary processes, and facies assemblages: Journal of Sedimentary Research, v. A64, p. 450-489.

Blair, T.C., 1987, Sedimentary processes, vertical stratification sequences, and geomorphology of the Roaring River alluvial fan, Rocky Mountain National Park, Colorado: Sedimentary Petrology, v. 57, p. 1-18.

Blair, T.C., 1999, Sedimentary processes and facies of the waterlaid Anvil Spring Canyon alluvial fan, Death valley, California: 46, p. 913-940.

Boothroyd, J.C., and Ashley, G.M., 1975, Processes, bar morphology, and sedimentary structures on braided outwash fans, northeastern Gulf of Alaska, *in* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and Glaciolacustrine Sedimentation: Tulsa, Oklahoma, Special Publication - Society of Economic Paleontologists and Mineralogists, p. 193-222.

Boulton, G.S., and Eyles, N., 1979, Sedimentation by valley glaciers; a model and genetic classification, *in* Schluechter, C., ed., Moraines and varves; origin, genesis, classification: Rotterdam, Netherlands, A. A. Balkema, p. 11-23.

Boulton, G.S., and Hindmarsh, R.C.A., 1987, Sediment deformation beneath glaciers: rheology and geological consequences: Journal of Geophysical Research, v. 92, p. 9059-9082.

Boyce, J.I., and Eyles, N., 1991, Drumlins carved by deforming till streams below the Laurentide Ice Sheet: Geology, v. 19, p. 787-790.

Boyce, J.I., Eyles, N., and Pugin, A., 1995, Seismic reflection, borehole and outcrop geometry of Late Wisconsin tills at a proposed landfill near Toronto, Ontario: Canadian Journal of Earth Sciences, v. 32, p. 1331-1349.

Boyd, R., Scott, D.B., and Douma, M., 1988, Glacial tunnel valleys and Quaternary history of the outer Scotian shelf: Nature, v. 333, p. 61-64.

Brennand, T.A., 1994, Macroforms, large bedforms and rhythmic sedimentary sequences in subglacial eskers, south-central Ontario: implications for esker genesis and meltwater regime: Sedimentary Geology, v. 91, p. 9-55.

Brennand, T.A., 2000, Deglacial meltwater drainage and glaciodynamics: inferences from Laurentide eskers, Canada: Geomorphology, v. 32, p. 263-293.

Brennand, T.A., Moore, A., Logan, C., Kenny, F.M., Russell, H.A.J., Sharpe, D.R., and Barnett, P.J., 1997, Bedrock Topography of the Greater Toronto and Oak Ridges Moraine areas, southern Ontario: Geological Survey of Canada, Open File 3419.

Brennand, T.A., and Shaw, J., 1994, Tunnel channels and associated landforms, south-central Ontario: their implications for ice-sheet hydrology: Canadian Journal of Earth Sciences, v. 31, p. 505-522.

Brennand, T.A., and Shaw, J., 1996, The Harricana glaciofluvial complex, Abitibi region, Quebec; its genesis and implications for meltwater regime and ice-sheet dynamics: Sedimentary Geology, v. 102, p. 221-262.

Bretz, J.H., 1969, The Lake Missoula floods and the Channeled Scablands: Journal of Geology, v. 77, p. 505-543.

Bristow, C.S., and Best, J.L., 1993, Braided rivers: perspectives and problems, *in* Best, J.L., and Bristow, C.S., eds., Braided rivers.:, Geological Society of London; Special Publication, 75, p. 1-11.

Brodzikowski, K., and Van Loon, A.J., 1991, Glacigenic Sediments: Developments in Sedimentology 49, Elsevier, 674 p.

Burbidge, G.H., and Rust, B.R., 1988, A Champlain sea subwash fan at St Lazare, Quebec, *in* Gadd, N.R., ed., The Late Quaternary Development of the Champlain Sea Basin: St. John's, NF., Geological Association of Canada, Special Paper 35, p. 47-61.

Chapman, L.J., 1985, On the origin of the Oak Ridges Moraine, southern Ontario: Canadian Journal of Earth Sciences, v. 22, p. 300-303.

Chapman, L.J., and Putnam, D.F., 1984, The Physiography of Southern Ontario: Ontario Geological Survey Special Volume 2, 270 p.

Cheel, R.J., 1982, The depositional history of an esker near Ottawa, Canada: Canadian Journal of Earth Sciences, v. 19, p. 1417-1427.

Cheel, R.J., and Middleton, G.V., 1986, Horizontal laminae formed under upper flow regime plane bed conditions: Journal of Geology, v. 94, p. 489-504.

Cheel, R.J., and Rust, B.R., 1982, Coarse grained facies of glacio-marine deposits near Ottawa, Canada, *in* Davidson, A.R., Nickling, W., and Fahey, B.D., eds., Research in Glacial, Glacio-fluvial, and Glacio-lacustrine Systems: Proceedings - Guelph Symposium on Geomorphology: Guelph, On, Canada, University of Guelph, p. 279-295.

Cheel, R.J., and Rust, B.R., 1986, A sequence of soft-sediment deformation (dewatering) structures in late Quaternary subaqueous outwash near Ottawa, Canada: Sedimentary Geology, v. 47, p. 77-93.

Chow, V.T., 1959, Open-Channel Hydraulics: McGraw-Hill Civil Engineering Series: New York, McGraw-Hill Book Company, 680 p.

Church, M., 1972, Baffin Island Sandurs; A Study Of Arctic Fluvial Processes: Geological Survey of Canada, Bulletin 216: Ottawa, On., Canada, Geological Survey of Canada, 208 p.

Church, M., and Gilbert, R., 1975, Proglacial fluvial and lacustrine environments, *in* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and Glaciolacustrine Sedimentation: Tulsa, Oklahoma, Special Publication - Society of Economic Paleontologists and Mineralogists, p. 22-100.

Clague, J.J., and Mathews, W.H., 1973, The magnitude of jökulhlaups: Journal of Glaciology, v. 12, p. 501-503.

Clayton, L., Attic, J.I., and Mickelson, D.M., 1999, Tunnel channels formed in Wisconsin during the last glaciation, *in* Mickelson, D.M., and Attic, J.I., eds., glacial Processes Past and Present: Boulder, Colorado, geological Society of America Special Paper 337, p. 69-82.

Cofaigh, C.O., 1996, Tunnel valley genesis: Progress in Physical Geography, v. 20, p. 1-19.

Collinson, J.D., and Thompson, D.B., 1989, Sedimentary Structures: London, Unwin, 207 p.

Costa, J.E., 1988, Rheologic, geomorphic, and sedimentologic differentiation of water floods, hyperconcentrated flows, and debris flows, *in* Baker, V.R., Kochel, R.C., and Patton, P.C., eds., Flood Geomorphology: New York, NY, United States, John Wiley & Sons, p. 113-122.

Cowan, E.A., Powell, R.D., and Smith, N.D., 1988, Rainstorm-induced event sedimentation at the tidewater front of a temperate glacier: Geology, v. 16, p. 409-412.

Cowan, W.R., 1976, Quaternary Geology of the Orangeville Area, southern Ontario: Toronto, Ontario, Ontario Division of Mines, GR141, p. 98.

Cowan, W.R., 1984, Geology of the Caledon Outwash, Ontario, *in* Guillet, G.R., and Martin, W., eds., The Geology of Industrial Minerals in Canada: Special Volume - Canadian Institute of Mining and Metallurgy: Montreal, PQ, Canada, Canadian Institute of Mining and Metallurgy, p. 137-142.

Dardis, G.F., and Hanvey, P.M., 1994, Sedimentation in a drumlin lee-side subglacial cavity, northwest Ireland: Sedimentary Geology, v. 91, p. 97-114.

Daub, G.A., 1996, Experimental study of deep-sea fan evolution and sedimentology [unpublished MSc thesis]: Colorado State University, Fort Collins, CO, 165 p.

Diemer, J.A., 1988, Subaqueous outwash deposits in the Ingraham ridge, Chazy, New York: Canadian Journal of Earth Sciences, v. 25, p. 1384-1396.

Dorsey, R.J., and Falk, P.D., 1998, Rapid development of gravelly high-density turbidity currents in marine Gilbert-type fan deltas, Loreto Basin, Baja California Sur, Mexico: Sedimentology, v. 45, p. 331-349.

Dowdeswell, J.A., and Siegert, M.J., 1999, The dimensions and topographic setting of Antarctic subglacial lakes and implications for large-scale water storage beneath continental ice sheets: Geological Society of America Bulletin, v. 111, p. 254-263.

Duckworth, P.B., 1979, The late depositional history of the western end of the Oak Ridges Moraine, Ontario: Canadian Journal Earth Sciences, v. 16, p. 1094-1107.

Ehlers, J., and Linke, G., 1989, The origin of deep buried channels of Elsterian age in northwest Germany: Journal of Quaternary Science, v. 4, p. 255-265.

Eyles, C.H., and Eyles, N., 1983, Sedimentation in a large lake: a reinterpretation of the late Pleistocene stratigraphy at Scarborough Bluffs, Ontario, Canada: Geology, v. 11, p. 146-152.

Eyles, N., and Clark, B.M., 1988, Last interglacial sediments of the Don Valley Brickyard, Toronto, Canada, and their paleoenvironmental significance: Canadian Journal of Earth Sciences, v. 25, p. 1108-1122.

Eyles, N., Clark, B.M., Kaye, B.G., Howard, K.W.F., and Eyles, C.H., 1985, The application of basin analysis techniques to glaciated terrains; an example from the Lake Ontario Basin, Canada: Geoscience Canada, v. 12, p. 22-32.

Eyles, N., Eyles, C.H., and Miall, A.D., 1983, Lithofacies types and vertical profile models; an alternative approach to the description and environmental interpretation of glacial diamict and diamictite sequences: Sedimentology, v. 30, p. 393-410.

Eyles, N., and McCabe, A.M., 1989, Glaciomarine facies within subglacial tunnel valleys: the sedimentary record of glacio-isostatic downwarping in the Irish Sea Basin: Sedimentology, v. 36, p. 431-448.

Eyles, N., and Williams, N.E., 1992, The sedimentary and biological record of the last interglacial-glacial transition at Toronto, Canada, *in* Clark, P.U., and Lea, P.D., eds., The Last Interglacial-Glacial Transition in North America: Boulder, Colorado, Geological Society of America Special Paper 270, p. 119-137.

Fenco MacLaren, L., 1994, Detailed assessment of the proposed site V4A, Appendices CA to CJ, geology and hydrogeology: Toronto, Interim Waste Authority, prepared by M Gomer, Fenco MacLaren, November 1994.

Fenco-MacLaren, 1994, IWA landfill site search, Metro/York region: Step 6 hydrogeological report sites V3B, V4A, V4D: Toronto, Interim Waste Authority Report, Prepared by M. Gomer, Fenco MacLaren Inc., January 1994.

Fisher, T.G., and Taylor, L.D., 1999, Tunnel channels and drumlins in south central MI. record a readvance over the Kalamazoo and Tekonsha moraines: Geological Society of America, 33 annual meeting, Geological Society of America, Champaign, Illinois, April 22-23, 1999, v. 1, p. A-15.

Fleisher, P.J., Cadwell, D.H., and Muller, E.H., 1997, Tsivat basin conduit system persists through two surges, Bering Piedmont Glacier, Alaska: Geological Society of America Bulletin, v. 110, p. 877-887.

Fogg, G.E., Noyes, C.D., and Carle, S.F., 1998, Geological based model of heterogeneous hydraulic conductivity in an alluvial setting: Hydrogeology Journal, v. 6, p. 131-143.

Folk, R.L., 1974, Petrology of Sedimentary Rocks: Austin, Hemphill Publishing Company, 182 p.

Fountain, A.G., and Walder, J.S., 1998, Water flow through temperate glaciers: Reviews of geophysics, v. 36, p. 299-328.

Fraser, J.Z., 1982, Derivation of a summary facies sequence based on Markov chain analysis of the Caledon Outwash; a Pleistocene braided glacial fluvial deposit, *in* Davidson, A.R., Nickling, W., and Fahey, B.D., eds., Research in glacial, glacio-fluvial, and glacio-lacustrine systems: Proceedings - Guelph Symposium on Geomorphology: Guelph, ON, Canada, University of Guelph, p. 175-199.

Fyfe, G.J., 1990, The effect of water depth on ice-proximal glacilacustrine sedimentation: Salpausslka I, southern Finland: Boreas, v. 19, p. 147-164.

Gerber, R.E., and Howard, K.W.F., 1997, Ground-water recharge to the Oak Ridges Moraine, *in* Eyles, N., ed., Environmental Geology of Urban Areas: St. John's, Geological Association of Canada, Geotexts 3, p. 173-192.

Gilbert, R., 1975, Sedimentation in Lillooet Lake, British Columbia: Canadian Journal of Earth Sciences, v. 12, p. 1697-1711.

Gilbert, R., 1990, Rafting in glacimarine environments, *in* Dowdeswell, J.A., and Scourse, J.D., eds., Glacimarine Environments: Processes and Sediments: Geological Society, London; Special

Publication, No. 53, p. 105-120.

Gilbert, R., 1997, Glaciolacustrine sedimentation in part of the Oak Ridges Moraine: Geographie physique et Quaternaire, v. 7, p. 55-66.

Gilbert, R., and Shaw, J., 1981, Sedimentation in proglacial Sunwapta Lake, Alberta: Canadian Journal of Earth Sciences, v. 18, p. 81-93.

Golder and Associates, 1994, Peel region proposed landfill site C-34b: detailed assessment of the proposed site, appendix C - geology/hydrogeology: , Interim Waste Authority Report, prepared by R. Blair, Golder Associates, November 1994.

Gorrell, G., and Shaw, J., 1991, Deposition in an esker, bead and fan complex, Lanark, Ontario, Canada: Sedimentary Geology, v. 72, p. 285-314.

Gustavson, T.C., 1975, Sedimentation and physical limnology in proglacial Malaspina Lake, southeastern Alaska, *in* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and Glaciolacustrine Sedimentation: Tulsa, Special Publication - Society of Economic Paleontologists and Mineralogists, p. 249-263.

Gustavson, T.C., Ashley, G.M., and Boothroyd, J.C., 1975, Depositional sequences in glaciolacustrine deltas, *in* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and Glaciolacustrine Sedimentation: Tulsa, Special Publication - Society of Economic Paleontologists and Mineralogists, p. 264-280.

Gustavson, T.C., and Boothroyd, J.C., 1987, A depositional model for outwash, sediment sources, and hydrogeologic characteristics, Malaspina Glacier, Alaska: A modern analog of the southeastern margin of the Laurentide Ice Sheet: Geological Society of America Bulletin, v. 99, p. 187-200.

Gwyn, Q.H.J., and Cowan, W.R., 1978, The origin of the Oak Ridges and Orangeville moraines: The Canadian Geographer, v. 22, p. 345-352.

Gwyn, Q.H.J., and Dilabio, R.N.W., 1973, Quaternary Geology of the Newmarket Area, Southern Ontario: Ontario Division of Mines, Preliminary Map 836.

Gwyn, Q.H.J., and White, S., 1973, Quarternary Geology Alliston Area: Ontario Geological Survey, Ministry of Natural Resources, Preliminary map P.835.

Hampton, M.A., 1975, Competence of fine-grained debris flows: Journal of Sedimentary Petrology, v. 45, p. 834-844.

Harms, J.C., Southard, J.B., Spearing, D.R., and Walker, R.G., 1975, Depositional Environments as Interpreted From Primary Sedimentary Structures and Stratification Sequences: Lecture notes for Short Course No. 2: Tulsa, Society of Economic Paleontologists and Mineralogists, 161 p.

Harms, J.C., Southard, J.B., and Walker, R.G., 1982, Structures and Sequences in Clastic Rocks: Society Economic Paleontologists and Mineralogists Short Course No. 9, Society of Economic Paleontologists and Mineralogists, 161 p.

Hart, J., 1999, Identifying fast ice flow from landform assemblages in the geological record: a discussion: Annals of Glaciology, v. 28, p. 59-66.

Hein, F.J., 1984, Deep-sea and fluvial braided channel conglomerates: a comparison of two case studies, *in* Koster, E.H., and Steel, R.J., eds., Sedimentology of Gravels and Conglomerates: Calgary, Canadian Society of Petroleum Geologists, p. 33-50.

Hein, F.J., and Walker, R.G., 1982, The Cambro-Ordovician Cap Enrage Formation, Quebec, Canada; conglomeratic deposits of a braided submarine channel with terraces: Sedimentology, v. 29, p. 309-329.

Henderson, P.J., 1988, Sedimentation in an esker system influenced by bedrock topography near Kingston, Ontario: Canadian Journal Earth Sciences, v. 25, p. 987-999.

Hillaire-Marcel, C., Occhietti, S., and Vincent, J.S., 1981, Sakami moraine, Quebec: A 500-km-long moraine without climatic control: Geology, v. 9, p. 210-214.

Hinton, M.J., Russell, H.A.J., Bowen, G.S., and Ahad, M.E., 1998, Groundwater discharge in the Humber River watershed: Groundwater in a Watershed Context, Canadian Water Resources Association, Burlington, Ontario, v. 2.

Hiscott, R.N., 1994a, Loss of capacity, not competence, as the fundamental process governing deposition from turbidity currents: Journal of Sedimentary Research A, v. 64, p. 209-214.

Hiscott, R.N., 1994b, Traction-carpet stratification in turbidites - fact or fiction?: Journal of Sedimentary Research, v. 64A, p. 204-208.

Hoffmans, G.J.C.M., 1998, Jet scour in equilibrium phase: Journal of Hydraulic Engineering, v. 124, p. 430-437.

Howard, K.W.F., Eyles, N., Smart, P.J., Boyce, J.I., Gerber, R.E., Salvatori, S.L., and Doughty, M., 1995, The Oak Ridges Moraine of southern Ontario: a ground-water resource at risk: Geoscience Canada, v. 22, p. 101-120.

Hunter, G.T., 1996, Executive summary: hydrogeological evaluation of the Oak Ridges Moraine Area (part of background report no. 3 for the Oak Ridges Moraine planning study), prepared by Hunter and Associates with Raven Beck Environmental Ltd.

Hunter, L.E., Powell, R.D., and Smith, G.W., 1996, Facies architecture and grounding-line fan processes of morainal banks during deglaciation of coastal Maine: Geological Society of America Bulletin, v. 108, p. 1022-1038.

Iseya, F., and Ikeda, H., 1987, Pulsation in bedload transport rates induced by a longitudinal sediment sorting: a flume study using sand and gravel mixtures: Geografiska Annaler, v. 69, p. 15-27.

Johansson, C.E., 1976, Structural studies of frictional sediments: Geografiska Annaler, v. 58, p. 201-300.

Johnson, M.D., 1999, Spooner Hills, northwest Wisconsin: High-relief hills carved by subglacial meltwater of the Superior Lobe, *in* Mickelson, D.M., and Attic, J.I., eds., Glacial Processes Past and

Present: Boulder, Colorado, Geological Society of America Special Paper 337, p. 83-92.

Johnson, M.D., Armstrong, D.K., Sanford, B.V., Telford, P.G., and Rutka, M.A., 1992, Paleozoic and Mesozoic Geology of Ontario, *in* Thurston, P.C., Williams, H.R., Sutcliffe, R.H., and Stott, G.M., eds., Ontario Geological Survey, Special Volume 4, Part 2: Toronto, Ontario Geological Survey.

Jopling, A.V., and Walker, R.G., 1968, Morphology and origin of ripple-drift cross-lamination, with examples from the Pleistocene of Massachusetts: Journal of Sedimentary Petrology, v. 38, p. 971-984.

Kamb, B., Raymond, C.F., Harrison, W.D., Engelhardt, H., Echelmeyer, K.A., Humphrey, N., Brugman, M.M., and Pfeffer, T., 1985, Glacier surge mechanism; 1982-1983 surge of Variegated Glacier, Alaska: Science, v. 227, p. 469-479.

Karrow, P.F., 1963, Pleistocene Geology of the Hamilton-Galt Area: Ontario Mines GR16, Accompanied by Maps 2029, 2030,2033, and 2034.

Karrow, P.F., 1967, Pleistocene Geology of the Scarborough Area: Toronto, Ontario Ministry of Natural Resources, 108 p.

Karrow, P.F., 1973, The Waterloo kame-moraine, a discussion: Zeitschrift fur Geomorphologie, v. 17, p. 126-133.

Karrow, P.F., 1974, Till stratigraphy in parts of southwestern Ontario: Geological Society of America Bulletin, v. 85, p. 761-768.

Karrow, P.F., 1991, Quaternary geology of the Brampton area: Toronto, Ontario Geological Survey, Open File Report 5819, p. 136.

Karrow, P.F., and Occhietti, S., 1989, Quaternary geology of the St. Lawrence Lowlands of Canada, *in* Fulton, R.J., ed., Quaternary Geology of Canada and Greenland: Ottawa, Geological Survey of Canda, p. 321-389.

Kaszycki, C.A., and DiLabio, R.N.W., 1986, Surficial geology and till geochemistry Lynn Lake - Leaf Rapids region, Manitoba, Current Research, Part B: Ottawa, Ontario, Geological Survey of Canada, Paper 86-1B, p. 245-256.

Kelly, R., and Martini, I.P., 1986, Pleistocene glacio-lacustrine deltaic deposits of the Scarborough Formation, Ontario, Canada: Sedimentary Geology, v. 47, p. 27-52.

Kelly, R.I., 1994, Results of a Quaternary Geology Hollow Stem Augering Program, Woodbridge, Ontario: Toronto, Ministry of Northern Development and Mines, Open File Report 5887, p. 52.

Kenny, F., 1998, A chromostereo enhanced Digital Elevation Model of the Oak Ridges Moraine Area, southern Ontario: Geological Survey of Canada and Ontario Ministry of Natural Resources, Open File 3423.

Kenny, F.M., Paquette, J., Russell, H.A.J., Moore, A.M., and Hinton, M.J., 1999, A Digital Elevation Model of the Greater Toronto Area, Southern Ontario and Lake Ontario Bathymetry: Ottawa, Ontario, Geological Survey of Canada, Ontario Ministry of Natural Resources, and Canadian Hydrographic Service; Geological Survey of Canada.

Kirkbride, M.P., 1993, The temporal significance of transitions from melting to calving termini at glaciers in the central southern Alps of New Zealand: The Holocene, v. 3, p. 232-240.

Kneller, B.C., and Branney, M.J., 1995, Sustained high-density turbidity currents and the deposition of thick massive sands: Sedimentology, v. 42, p. 607-616.

Komar, P.D., 1971, Hydraulic jumps in turbidity currents: Geological Society of America Bulletin, v. 82, p. 1477-1488.

Kor, P.S.G., and Cowell, D.W., 1998, Evidence for catastrophic subglacial meltwater sheetflood events on the Bruce Peninsula, Ontario: Canadian Journal of Earth Sciences, v. 35, p. 1180-1202.

Kor, P.S.G., Shaw, J., and Sharpe, D.R., 1991, Erosion of bedrock by subglacial meltwater, Georgian Bay, Ontario; a regional view: Canadian Journal of Earth Sciences, v. 28, p. 623-642.

Koster, E.H., 1978, Transverse ribs: their characteristics, origin and paleohydraulic significance, *in* Miall, A.D., ed., Fluvial Sedimentology: Calgary, Canadian Society of Petroleum Geologists, p. 161-186.

LeGrand, H.E., and Rosen, L., 1998, Putting hydrogeological site studies on track: Ground Water, v. 36, p. 193-194.

Levy, A.G., and Ellms, J.I., 1927, The hydraulic jump as a mixing device: Journal of the American Water Works Association, v. 17, p. 1-23.

Lewis, C.F.M., Blasco, S.M., Cameron, G.D.M., and King, E.L., 1999, A Late Wisconsinan regional unconformity beneath eastern Lake Erie: Evidence of significant subglacial erosion events.

Lewis, C.F.M., Cameron, G.D.M., Mayer, L.A., and Todd, B.J., 1997a, Drumlins in Lake Ontario, *in* Davies, T., Bell, T., Cooper, A., Josenhans, H., Polyak, L., Solheim, A., Stoker, M., and Stravers, J., eds., Glaciated Continental Margins: An Atlas of Acoustic Images: London, Chapman and Hall, p. 48-49.

Lewis, C.F.M., Cameron, G.D.M., and Todd, B.J., 1997b, Seismostratigraphy of western Lake Ontario: Quaternary geology and implications for groundwater discharge: GAC/MAC Ottawa'97 Annual Meeting May 19-21, Geological Association of Canada, Ottawa, Ontario, p. A-89.

Liberty, B.A., 1969, Paleozoic Geology of the Lake Simcoe District, Ontario: Ottawa, Geological Survey of Canada, Memoir 355, 201 p.

Lindsay, P.J., Percival, J.B., Tsai, A.C., and Wygergangs, M.H.M., 1998, Investigation of automated particle size analysis techniques, Current Research 1998-E: Ottawa, Geological Survey of Canada, p. 173-182.

Logan, C., Russell, H.A.J., and Sharpe, D.R., 2001, Regional 3-D stratigraphic modelling of the Oak Ridges Moraine area, southern Ontario, Current Research (in press): Ottawa, Geological Survey of Canada.

Loncarevic, B.D., Piper, D.J.W., and Fader, G.B.J., 1992, Application of high-quality bathymetry to geological interpretation on the Scotian Shelf: Geoscience Canada, v. 19, p. 5-13.

Lonne, I., 1995, Sedimentary facies and depositional architecture of ice-contact glaciomarine systems: Sedimentary Geology, v. 98, p. 13-45.

Lowe, D.R., 1976, Grain flow and grain flow deposits: Journal of Sedimentary Petrology, v. 46, p. 188-199.

Lowe, D.R., 1982, Sedimentary gravity flows: II. Depositional models with special reference to the deposits of high density turbidity currents: Journal of Sedimentary Research, v. 52, p. 279-297.

Lowe, D.R., and LoPiccolo, R.D., 1974, The characteristics and origins of dish and pillar structures: Journal of Sedimentary Petrology, v. 44, p. 484-501.

Maizels, J., 1989, Sedimentology, paleoflow dynamics and flood history of jökulhlaup deposits: paleohydrology of Holocene sediment sequences in southern Iceland sandur deposits: Journal Sedimentary Research, v. 59, p. 204-223.

Maizels, J., 1993, Lithofacies variations within sandur deposits: the role of runoff regime, flow dynamics and sediment supply characteristics: Sedimentary Geology, v. 85, p. 299-325.

Maizels, J., 1997, Jökulhlaup deposits in proglacial areas: Quaternary Science Reviews, v. 16, p. 793-819.

MapInfo, 1998, MapInfo Professional: Troy, New York, MapInfo Corporation.

Martin, C.A.L., and Turner, B.R., 1998, Origins of massive-type sandstones in braided river systems: Earth-Science Reviews, v. 44, p. 15-38.

McBride, E.F., Shepherd, R.G., and Crawley, R.A., 1975, Origin of parallel, near-horizontal laminae by migration of bed forms in a small flume: Journal of Sedimentary Petrology, v. 45, p. 132-139.

McCabe, A.M., and O'Cofaigh, C., 1994, Sedimentation in a subglacial lake, Enniskerry, eastern Ireland: Sedimentary Geology, v. 91, p. 57-95.

McDonald, B.C., and Day, T.J., 1978, An experimental flume study on the formation of transverse ribs, Current Research Paper 78-1A: Ottawa, Geological Survey of Canada, p. 441-451.

McDonald, B.C., and Vincent, J.S., 1972, Fluvial sedimentary structures formed experimentally in a pipe, and their implications for interpretation of subglacial sedimentary environments: Geological Survey of Canada; 72-27: Ottawa, Ont, Dept. of Energy Mines and Resources, 30 p.

McKee, E.D., and Weir, G.W., 1953, Terminology for stratification and cross-stratification in sedimentary rocks: Bulletin of the Geological Society of America, v. 64, p. 381-390.

Miall, A.D., 1977, A review of the braided-river depositional environment: Earth-Science Reviews, v. 13, p. 1-62.

Miall, A.D., 1984, Principles of sedimentary basin analysis: New York, Springer-Verlag, 490 p.

Miall, A.D., 1996, The Geology of Fluvial Deposits: Sedimentary Facies, Basin Analysis, and Petroleum Geology: Berlin; New York, Springer, xvi, 582 p.

Mickelson, D.M., Clayton, L., Fullerton, D.S., and Borns, H.W., 1983, The Late Wisconsin glacial record of the Laurentide ice sheet in the United States, *in* Wright, H.E., ed., Late-Quaternary Environments of the United States: Volume 1, The Late Pleistocene: Minneapolis, University of Minnesota, p. 3-37.

Middleton, G.V., 1965, Antidune cross-bedding in a large flume: Journal of Sedimentary Petrology, v. 35, p. 922-927.

Mooers, H.D., 1989, On the formation of the tunnel valleys of the Superior lobe, central Minnesota: Quaternary Research, v. 32, p. 24-35.

MSAccess, 1997, MSAccess: Redmond, Microsoft.

Mulder, T., and Cochonat, P., 1996, Classification of offshore mass movements: Journal of Sedimentary Research, v. 66, p. 43-57.

Muller, F., 1962, Zonation in the accumulation area of the glaciers of Axel Heiberg Island, N.W.T., Canada: Journal of Glaciology, v. 4, p. 302-311.

Mullins, H.T., and Hinchey, E.J., 1989, Erosion and infill of New York Finger Lakes: Implications for Laurentide ice sheet deglaciation: Geology, v. 17, p. 622-625.

Mullins, H.T., Hinchey, E.J., Wellner, R.W., Stephens, D.B., Anderson, W.T., Jr., Dwyer, T.R., and Hine, A.C., 1996, Seismic stratigraphy of the Finger Lakes; a continental record of Heinrich event H-1 and Laurentide ice sheet instability, *in* Mullins, H.T., and Eyles, N., eds., Subsurface Geologic Investigations of New York Finger Lakes; Implications for Late Quaternary Deglaciation and Environmental Change: Special Paper - Geological Society of America 311: Boulder, CO, United States, Geological Society of America, p. 1-35.

Munro-Stasiuk, M.J., 1999, Evidence for water storage and drainage at the base of the Laurentide ice sheet, south-central Alberta, Canada: Annals-of-Glaciology, v. 28, p. 175-180.

Murray, C.J., Lowe, D.R., Graham, S.A., Martinez, P.A., Zeng, J., Carroll, A.R., Cox, R., Hendrix, M., Heubeck, C., Miller, D., Moxon, I.W., Sobel, E., Wendebourg, J., and Williams, T., 1996, Statistical analysis of bed-thickness patterns in a turbidite section from the great valley sequence, cache creek, northern California: Journal of Sedimentary Research, v. 66, p. 900-908.

Nemec, W., 1988, The shape of the rose: Sedimentary Geology, v. 59, p. 149-152.

Nemec, W., Lønne, I., and Blikra, L.H., 1999, The Kregnes moraine in Gauldalen, west-central Norway: anatomy of a Younger Dryas proglacial delta in a palaeofjord basin: Boreas, v. 28, p. 454-476.

Nordin, C.F., 1963, A Preliminary Study of Sediment Transport Parameters Rio Puerco Near Bernardo New Mexico: Geological Survey Professional Paper 462-c: Washington, United States Geological Survey, 21 p.

Olufemi, A., 1996, Contributions to Erosions By Jets [unpublished Phd thesis]: University of Alberta, Edmonton, 177 p.

Pair, D.L., 1997, Thin film, channelized drainage, or sheetfloods beneath a portion of the Laurentide Ice Sheet: an examination of glacial erosion forms, northern New York state, USA: Sedimentary Geology, v. 111, p. 199-215.

Paterson, J.T., 1995, Sedimentology of the Bloomington fan complex, Oak Ridges Moraine, southern Ontario: Brock University, Dept. of Geology, M.Sc. thesis.

Paterson, J.T., and Cheel, R.J., 1997, The depositional history of the Bloomington Complex, an ice-contact deposit in the Oak Ridges Moraine, southern Ontario, Canada: Quaternary Science Reviews, v. 16, p. 705-719.

Paterson, W.S.B., 1981, The Physics of Glaciers, 2nd Edition: New York, Pergamon Press, 380 p.

Patterson, C.J., 1994, Tunnel-valley fans of the St. Croix Moraine, east-central Minnesota, USA, *in* Warren, W.P., and Croot, D.G., eds., Formation and Deformation of Glacial Deposits: Rotterdam, Netherlands, Balkema, p. 69-87.

Patterson, C.J., 1997, Southern Laurentide ice lobes were created by ice streams: Des Moines Lobe in Minnesota, USA: Sedimentary Geology, v. 111, p. 249-261.

Pickering, K.T., Hiscott, R.N., and Hein, F.J., 1989, Deep Marine Environments; Clastic Sedimentation and Tectonics: London, United Kingdom, Unwin Hyman, 416 p.

Pierson, T.C., and Scott, K.M., 1985, Downstream dilution of a lahar: transition from debris flow to hyperconcentrated streamflow: Water Resources Research, v. 21, p. 1511-1524.

Piotrowski, J.A., 1994, Tunnel-valley formation in northwest Germany - geology, mechanism of formation and subglacial bed conditions for the Borhoved tunnel valley: Sedimentary Geology, v. 89, p. 107-141.

Piotrowski, J.A., 1997, Subglacial hydrogeology in north-western Germany during the last glaciation: groundwater flow, tunnel valleys and hydrogeological cycles: Quaternary Science Reviews, v. 16, p. 169-185.

Plink-Bjorklund, P., and Ronnert, L., 1999, Depositional processes and internal architecture of Late Weichselian ice-margin submarine fan and delta settings, Swedish west coast: Sedimentology, v. 46, p. 215-234.

Post, A., and LaChappelle, 1971, Glacier Ice, University of Washington Press, 102 p.

Postma, G., Nemec, W., and Kleinspehn, K.L., 1988, Large floating clasts in turbidites: a mechanism for their emplacement: Sedimentary Geology, v. 58, p. 47-61.

Postma, G., Roep, T.B., and Ruegg, G.H.J., 1983, Sandy-gravelly mass-flow deposits in an ice-marginal lake (Saalian, Leuvenumsche Beek valley, Veluwe, the Netherlands), with emphasis on plug-flow deposits: Sedimentary Geology, v. 34, p. 59-82.

Potter, P.E., and Pettijohn, F.J., 1963, Paleocurrents and Basin Analysis: New York, Academic Press, 296 p.

Powell, R.D., 1981, A model for sedimentation by tidewater glaciers: Annals of Glaciology, v. 2, p. 129-134.

Powell, R.D., 1983, Glacial-marine sedimentation processes and lithofacies of temperate tidewater glaciers, Glacier Bay, Alaska, *in* Molnia, B.F., ed., Glacial-Marine Sedimentation: New York, NY, Plenum Press, p. 185-231.

Powell, R.D., 1990, Glacimarine processes at grounding-line fans and their growth to ice-contact deltas, *in* Dowdeswell, J.A., and Scourse, J.D., eds., Glacimarine Environments: Processes and Sediments, Special Publication, No. 53: London, Geological Society, p. 53-73.

Powell, R.D., 1991, Grounding-line systems as second-order controls on fluctuations of tidewater termini of temperate glaciers, *in* Anderson, J.B., and Ashley, G.M., eds., Glacial Marine Sedimentation; Paleoclimatic Significance: Boulder, Colorado, Geological Society of America Special Paper 261, p. 75-93.

Price, R.J., 1973, Glacial and Fluvioglacial Landforms: Edinburgh, Oliver and Boyd, 242 p.

Pugin, A., Pullan, S.E., and Sharpe, D.R., 1999, Seismic facies and regional architecture of the Oak Ridges Moraine area, southern Ontario: Canadian Journal of Earth Sciences, v. 36, p. 409-432.

Pullan, S.E., Pugin, A., Dyke, L.D., Hunter, J.A., Pilon, J.A., Todd, B.J., Allen, V.S., and Barnett, P.J., 1994, Shallow geophysics in a hydrogeological investigation of the Oak Ridges Moraine, Ontario, *in* Bell, R.S., and Lepper, C.M., eds., Symposium on the Application of Geophysics to Engineering and Environmental Problems: Boston, Environmental and Engineering Geophysical Society, p. 143-160.

Rajaratnam, N., and Subramanyan, S., 1986, Plane turbulent denser wall jets and jumps: Journal of Hydraulic Research, v. 24, p. 281-296.

Rothlisberger, H., and Lang, H., 1987, Glacial hydrology, *in* Gurnell, A.M., and Clark, M.J., eds., Glacio-fluvial Sediment Transfer An Alpine Perspective: Chichester, John Wiley and Sons Ltd, p. 207-284.

Russell, A.J., 1993, Supraglacial lake drainage near Sondre Stromfjord, Greenland: Glaciology, v. 39, p. 431-433.

Russell, A.J., and Knudsen, O., 1999a, An ice-contact rhythmite (turbidite) succession deposited during the November 1996 catastrophic outburst flood (Jökulhlaup), Skeidararjokull, Iceland: Sedimentary Geology, v. 127, p. 1-10.

Russell, A.J., and Knudsen, O., 1999b, An ice-contact rhythmite (turbidite) succession deposited during the November 1996 catastrophic outburst flood (Jökulhlaup), Skeidararjokull, Iceland: Sedimentary Geology, v. 127, p. 1-10.

Russell, H.A.J., and Arnott, W.R.C., 1997, Halton Complex, Humber watershed, *in* Sharpe, D.R., and Barnett, P.J., eds., Where is the Water? Regional Geological/ Hydrogeological Framework, Oak Ridges Moraine Area, southern Ontario: Ottawa, Geological Association of Canada, Field Trip A1 Guidebook 17-18 May, p. 18-22.

Russell, H.A.J., Brennand, T.A., Logan, C., and Sharpe, D.R., 1998a, Standardization and assessment of geological descriptions from water well records: Greater Toronto and Oak Ridges Moraine Areas, southern Ontario, Current Research 1998-E: Ottawa, Geological Survey of Canada, p. 89-102.

Russell, H.A.J., and Dumas, S., 1997, Surficial geology of the Alliston area, NTS 31D/4, southern Ontario: Geological Survey of Canada, Open File 3334.

Russell, H.A.J., Logan, C., Moore, A., Kenny, F.M., Brennand, T.A., Sharpe, D.R., and Barnett, P.J., 1998b, Sediment Thickness of the Greater Toronto and Oak Ridges Moraine NATMAP areas, southern Ontario: Geological Survey of Canada, Open File 2892.

Russell, H.A.J., and Pullan, S., 1998, Stop 2.4. Nobleton borehole in the Laurentian Channel, *in* Sharpe, D.R., Hinton, M., Russell, H.A.J., and Barnett, P.J., eds., Quaternary Geology and Hydrogeology of the Oak Ridges Moraine Area, Southern Ontario: , GSA, Annual Meeting Toronto Canada 1998, Field trip guide number 15.

Russell, H.A.J., Sharpe, D.R., and Arnott, R.W.C., 1998c, Sedimentology of the Oak Ridges Moraine, Humber River watershed, southern Ontario: a preliminary report, Current Research 1998-C: Ottawa, Ontario, Geological Survey of Canada, p. 155-166.

Russell, H.A.J., and White, O.L., 1997, Surficial geology of the Bolton area, NTS 30M/13, southern Ontario: Geological Survey of Canada, Open File 3299.

Rust, B.R., 1977, Mass flow deposits in a Quaternary succession near Ottawa, Canada: Canadian Journal of Earth Sciences, v. 14, p. 175-184.

Rust, B.R., 1988, Ice-proximal deposits of the Champlain Sea at south Gloucester, near Ottawa, Canada, *in* Gadd, N.R., ed., The Late Quaternary Development of the Champlain Sea Basin: St John's, NF., Geological Association of Canada, Special Paper, p. 37-45.

Rust, B.R., and Romanelli, R., 1975, Late Quaternary subaqueous outwash deposits near Ottawa, Canada, *in* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and Glaciolacustrine Sedimentation: Tulsa, Oklahoma, Special Publication - Society of Economic Paleontologists and Mineralogists, p. 177-192.

Sado, E.V., White, O.L., Barnett, P.J., and Sharpe, D.R., 1983, The glacial geology, stratigraphy and geomorphology of the North Toronto area: A field excursion, *in* Mahaney, W.C., ed., Correlation of Quaternary Chronologies: Norwich, Geobooks, p. 505-517.

Sanford, B.V., and Baer, A.J., 1981, Geological Map of southern Ontario: Geological Survey of Canada, Southern Ontario Sheet 30s, 1335A.

Saunderson, H.C., 1975, Sedimentology of the Brampton Esker and its associated deposits; an empirical test of theory, *in* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and Glaciolacustrine Sedimentation: Tulsa, OK, Society of Economic Paleontologists and Mineralogists, Special Publication 23, p. 155-176.

Saunderson, H.C., 1976, Paleocurrent analysis of large-scale cross-stratification in the Brampton Esker, Ontario: Journal of Sedimentary Petrology, v. 46, p. 761-769.

Saunderson, H.C., and Lockett, F.P.J., 1983, Flume experiments on bedform and structures at the dune-plane bed transition: Special Publication of the International Association of Sedimentologists, v. 6, p. 49-58.

Sawagaki, T., and Hirakawa, K., 1997, Erosion of bedrock by subglacial meltwater, Soya Coast, East Antarctica: Geografiska Annaler, v. 79A, p. 223-238.

Schlüchter, C., 1979, Moraines and varves: origin, genesis, classification: proceedings of an INQUA Symposium on Genesis and Lithology of Quaternary Deposits, Balkema, Zurich, 10- 20 September, 1978, p. 441.

Shanmugam, G., 1996, High-density turbidity currents; are they sandy debris flows?: Journal of Sedimentary Research, v. 66, p. 2-10.

Shanmugam, G., 1997, The Bouma Sequence and the turbidite mind set: Earth-Science Reviews, v. 42, p. 201-229.

Sharpe, D.R., 1986, Glaciomarine fan deposition in the Champlain Sea, 1986 joint annual meeting of GAC, MAC and CGU.: Program with Abstracts - Geological Association of Canada; Mineralogical Association of Canada; Canadian Geophysical Union, Joint Annual Meeting: Waterloo, ON, Canada, Geological Association of Canada, p. 126-127.

Sharpe, D.R., 1987, Quaternary geology of Toronto area, Ontario, *in* Roy, D.C., ed., Northeastern section of the Geological Society of America.: Boulder, CO, United States, Geol. Soc. Am., p. 339-344.

Sharpe, D.R., 1988, Glaciomarine fan deposition in the Champlain Sea, *in* Gadd, N.R., ed., The Late Quaternary Development of the Champlain Sea Basin: St. John's, NF., Geological Association of Canada, Special Paper, p. 63-82.

Sharpe, D.R., and Barnett, P.J., 1997a, A Field Guide to the Geology of the Oak Ridges Moraine Area, GAC/MAC field guide, Ottawa, 1997.

Sharpe, D.R., and Barnett, P.J., 1997b, Surficial Geology of the Markham Area, NTS 30M/14, southern Ontario: Geological Survey of Canada, Open File 3300.

Sharpe, D.R., Barnett, P.J., Brennand, T.A., Finley, D., Gorrell, G., Russell, H.A., and Stacey, P., 1997, Surficial geology of the Greater Toronto and Oak Ridges Moraine area, southern Ontario: Geological Survey of Canada, Open File 3062.

Sharpe, D.R., Barnett, P.J., Russell, H.A.J., Brennand, T.A., and Gorrell, G., 1999a, Regional geological mapping of the Oak Ridges Moraine, Greater Toronto Area, southern Ontario, Current Research 1999-E: Ottawa, Geological Survey of Canada, p. 123–136.

Sharpe, D.R., Brennand, Barnett, and Russell, 1999b, Mapping regional hydrostratigraphic units in 3-dimensions: Newmarket Till in the Oak Ridges Moraine area, southern Ontario: Geological Society of America 33rd annual meeting, north-central section,, Geological Society of America, Champaign, Illinois, April 22-23, 1999, v. 1.

Sharpe, D.R., and Cowan, W.R., 1990, Moraine formation in northwestern Ontario: product of subglacial fluvial and glaciolacustrine sedimentation: Canadian Journal of Earth Sciences, v. 27, p. 1478-1486.

Sharpe, D.R., Dyke, L.D., Hinton, M.J., Pullan, S.E., Russell, H.A.J., Brennand, T.A., Barnett, P.J., and Pugin, A., 1996, Groundwater prospects in the Oak Ridges Moraine area, southern Ontario: application of regional geological models, Current Research 1996-E: Ottawa, Geological Survey of Canada, p. 181-190.

Sharpe, D.R., Pullan, S., and Warman, T., 1992, A basin analysis of the Wabigoon area of Lake Agassiz, a Quaternary clay basin in northwestern Ontario: Geographie physique et Quaternaire, v. 46, p. 295–309.

Sharpe, D.R., Russell, H.A., Pullan, S.E., Pugin, A., Brennand, T.A., and Barnett, P.J., 1999c, Channel form and sediment fill of tunnel channels beneath the Oak Ridges Moraine, southern Ontario: GSA Annual meeting, Geological Society of America, Denver, Colorado, October 24-28, 1999.

Shaw, J., 1975, Sedimentary successions in Pleistocene ice-marginal lakes, *in* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation: Tulsa, Oklahoma, Society of Economic Paleontologists and Mineralogists. 23, p. 281-303.

Shaw, J., 1996, A meltwater model for Laurentide subglacial landscapes, *in* McCann, S.B., and Ford, D.C., eds., Geomorphology Sans Frontieres: Chichester, John Wiley & Sons Ltd, p. 181-236.

Shaw, J., and Gilbert, R., 1990, Evidence for large-scale subglacial meltwater flood events in southern Ontario and northern New York State: Geology, v. 18, p. 1169-1172.

Shaw, J., and Gorrell, G., 1990, Subglacially formed dunes with bimodal and graded gravel in the Trenton drumlin field, Ontario: Geographie physique Quaternaire, v. 45, p. 21-34.

Shaw, J., and Kvill, D., 1984, A glaciofluvial origin for drumlins of the Livingstone Lake area, Saskatchewan: Canadian Journal of Earth Sciences, v. 21, p. 1442-1459.

Shaw, J., Munro-Stasiuk, M., Rains, R., B, Sharpe, D.R., Pugin, A., Pullan, S., Lewis, C.F.M., Todd, B.J., and Gilbert, R.G., 1998, Regional landscape unconformities formed beneath the Laurentide ice sheet: GSA Annual Meeting, Geological Society of America, Abstracts with Programs, Toronto, Ontario, v. 30, no. 7.

Shaw, J., and Sharpe, D.R., 1987, Drumlin formation by subglacial meltwater erosion: Canadian Journal of Earth Sciences, v. 24, p. 2316-2322.

Shoemaker, E.M., 1991, On the formation of large subglacial lakes: Canadian Journal of Earth Sciences, v. 28, p. 1975-1981.

Shoemaker, E.M., 1999, Subglacial water-sheet floods, drumlins and ice-sheet lobes: Journal of Glaciology, v. 45, p. 201-213.

Siegert, M.J., 2000, Antarctic subglacial lakes: Earth-Science Reviews, v. 50, p. 29-50.

Siegert, M.J., Dowdeswell, J.A., Gorman, M.R., and McIntyre, N.F., 1996, An inventory of Antarctic sub-glacial lakes: Antarctic Science, v. 8, p. 281-286.

Siegert, M.J., and Ridley, J.K., 1998, An analysis of the ice-sheet surface and subsurface topography above the Vostok Station subglacial lake, central East Antarctica: Journal of Geophysical Research B: Solid Earth, v. 103, p. 10195-10207.

Simons, D.B., Richardson, E.V., and Nordin, C.F., 1965, Sedimentary structures generated by flow in alluvial channels, *in* Middleton, G.V., ed., Primary Sedimentary Structures and their Hydrodynamic Interpretations: Tulsa, Society of Economic Paleontologists and Mineralogists, p. 34-52.

Smith, G.A., and Lowe, D.R., 1991, Lahars: volcano-hydrologic events and deposition in the debris flow - hyperconcentrated flow continuum, *in* Fisher, R.V., and Smith, G.A., eds., Sedimentation in volcanic settings: Tulsa, Oklahoma, Society of Economic Paleontologists and Mineralogists, p. 59-70.

Sohn, Y.K., 1997, On traction-carpet sedimentation: Journal Sedimentary Research, v. 67, p. 502-509.

Spencer, A., 1890, Origins of the basins of the Great Lakes of America: American Geologist, v. 7, p. 86-97.

Spooner, I.S., and Dalrymple, R.W., 1994, Sedimentary facies relationships in esker-ridge/esker-fan complexes, southestern Ontario, Canada: Application to the exploration for asphalt blending sand: Quaternary International, v. 20, p. 81-92.

Stander, E., Verrette, J.L., and Hodgson, M., in prep, Ice stresses in reservoirs: the effect of water level fluctuations.

Straw, A., 1968, Late Pleistocene erosion along the Niagara Escarpment of southern Ontario: Geological Society of America Bulletin, v. 79, p. 889-910.

Sturm, M., and Benson, C.S., 1985, A history of jökulhlaups from Strandlake, Alaska: Journal of Glaciology, v. 31, p. 272-280.

Syvitski, J.P.M., 1989, On the deposition of sediment within glacier-influenced fjords: oceanographic controls: Marine Geology, v. 85, p. 301-329.

Syvitski, J.P.M., 1991, Principles, Methods, and Applications of Particel Size Analysis: New York, Cambridge Universtiy Press, 368 p.

Thomsen, H.H., Thorning, L., and Olesen, O.B., 1989, Applied glacier research for planning hydro-electric power, Ilulissat/Jakobshavn, West Greenland: Annals of Glaciology, v. 13, p. 257-261.

Todd, S.P., 1989, Stream driven, high-density gravelly traction carpets: Possible deposits in the Trabeg Conglomerate Formation, SW Ireland and some theoretical considerations of their origin: Sedimentology, v. 36, p. 513-530.

Tweed, F.S., and Russell, A.J., 1999, Controls on the formation and sudden drainage of glacier-impounded lakes: implications for jökulhlaup characteristics: Progress in Physical Geography, v. 23, p. 79-110.

Veillette, J.J., 1986, Former southwesterly ice flows in the Abitibi - Timiskaming region: implications for the configuration of the late Wisconsinan ice sheet: Canadian Journal Earth Science, v. 23, p. 1724-1741.

Vertical Mapper, 1999, Vertical Mapper: Ottawa, Northwood Geoscience.

Vrolijk, P.J., and Southard, J.B., 1997, Experiments on rapid deposition of sand from high-velocity flows: Geoscience Canada, v. 24, p. 45-54.

Walker, R.G., 1992, Turbidites and submarine fans, *in* Walker, R.G., and James, N.P., eds., Facies Models - response to sea level change: St John's, NF., Geological Association of Canada, p. 239-264.

Wang, D., and Anderson, D.W., 1998, Direct measurement of organic carbon content in soils by the Leco CR-12 Carbon Analyzer: Communications in Soil Science and Plant Analysis, v. 29, p. 15-21.

Warburton, J., and Fenn, C.R., 1994, Unusual flood events from an Alpine glacier: observations and deductions on generating mechanisms: Journal of Glaciology, v. 40, p. 176-186.

Warren, W.P., and Ashley, G.M., 1994, Origins of the ice-contact stratified ridges (eskers) of Ireland: Journal of Sedimentary Research, v. A64, p. 433-449.

Weidick, A., 1988, Greenland, *in* Williams, R.S., and Ferringno, J.G., eds., Satellite Image Atlas of Glaciers of the World: Washington, United States Geological Survey Professional Paper 1386-C, p. 141.

Weirich, F., 1985, Sediment budget for a high energy glacial lake: Geografiska Annaler, v. 67 A, p. 83-98.

Weirich, F., 1986, The record of density-induced underflows in a glacial lake: Sedimentology, v. 33, p. 261-277.

Wellner, R.W., Petruccione, J.L., and Sheridan, R.E., 1996, Correlation of drillcore and geophysical results from Canandaigua Lake valley, New York; evidence for rapid late-glacial sediment infill, *in* Mullins, H.T., and Eyles, N., eds., Subsurface Geologic Investigations of New York Finger Lakes; Implications for Late Quaternary Deglaciation and Environmental Change.: Special Paper - Geological Society of America 311: Boulder, CO, United States, Geological Society of America (GSA), p. 37-49.

White, O.L., 1973, Bolton Quaternary Geology: Ontario Division of Mines, Geological Map 2275.

White, O.L., 1975, Quaternary Geology of the Bolton area, southern Ontario: Toronto, Ontario Division of Mines, Geological Report 117, p. 119.

Williams, G.P., 1983, Paleohydraulical methods and some examples from Swedish fluvial environments: Geografiska Annaler, v. 65, p. 227-241.

Wingfield, R., 1990, The origin of major incisions within Pleistocene deposits of the North Sea: Marine Geology, v. 91, p. 31-52.

Wright, H.E., 1973, Tunnel valleys, glacial surges, and subglacial hydrology of the Superior Lobe, Minnesota, *in* Black, R.F., Goldthwait, R.P. and Willman, H.B., ed., The Wisconsinan Stage: Boulder, Geological Society of America, Memoir 136, p. 251-276.

Wright, L.D., 1977, Sediment transport and deposition at river mouths: A synthesis: Geological Society of America Bulletin, v. 88, p. 857-868.