DEGLACIATION OF THE CHAMPLAIN SEA BASIN, EASTERN ONTARIO

72nd FRIENDS OF THE PLEISTOCENE REUNION

June 6–7, 2009, Ottawa, Ontario

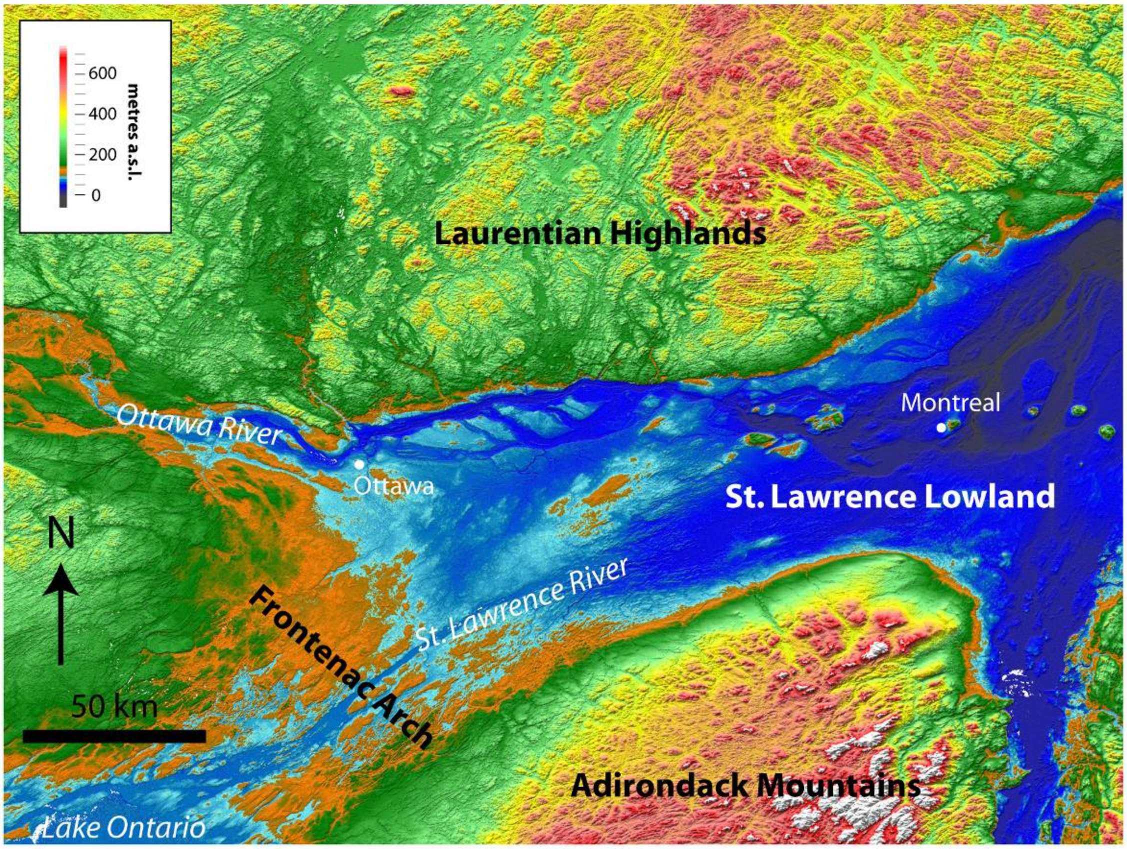

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Table of Contents

Abstract

The Champlain Sea was an inland arm of the Atlantic Ocean that invaded the St. Lawrence Lowland following retreat of the Laurentide Ice Sheet. This fieldtrip reviews a number of aspects of the deglacial landforms and deposits of the area, discusses the Champlain Sea deposits and reviews the societal implications of the deposits from a geotechnical and hydrogeological perspective. Day one of the two day trip is spent on the Vars - Wincehester esker which provides an opportunity to discuss esker and Champlain Sea deposits and to highlight the geotechnical and hydrogeological issues associated with these deposits. Day two of the trip visits the Cantley quarry and discusses the evidence for and against subglacial meltwater erosion for the sculpted forms at the site.
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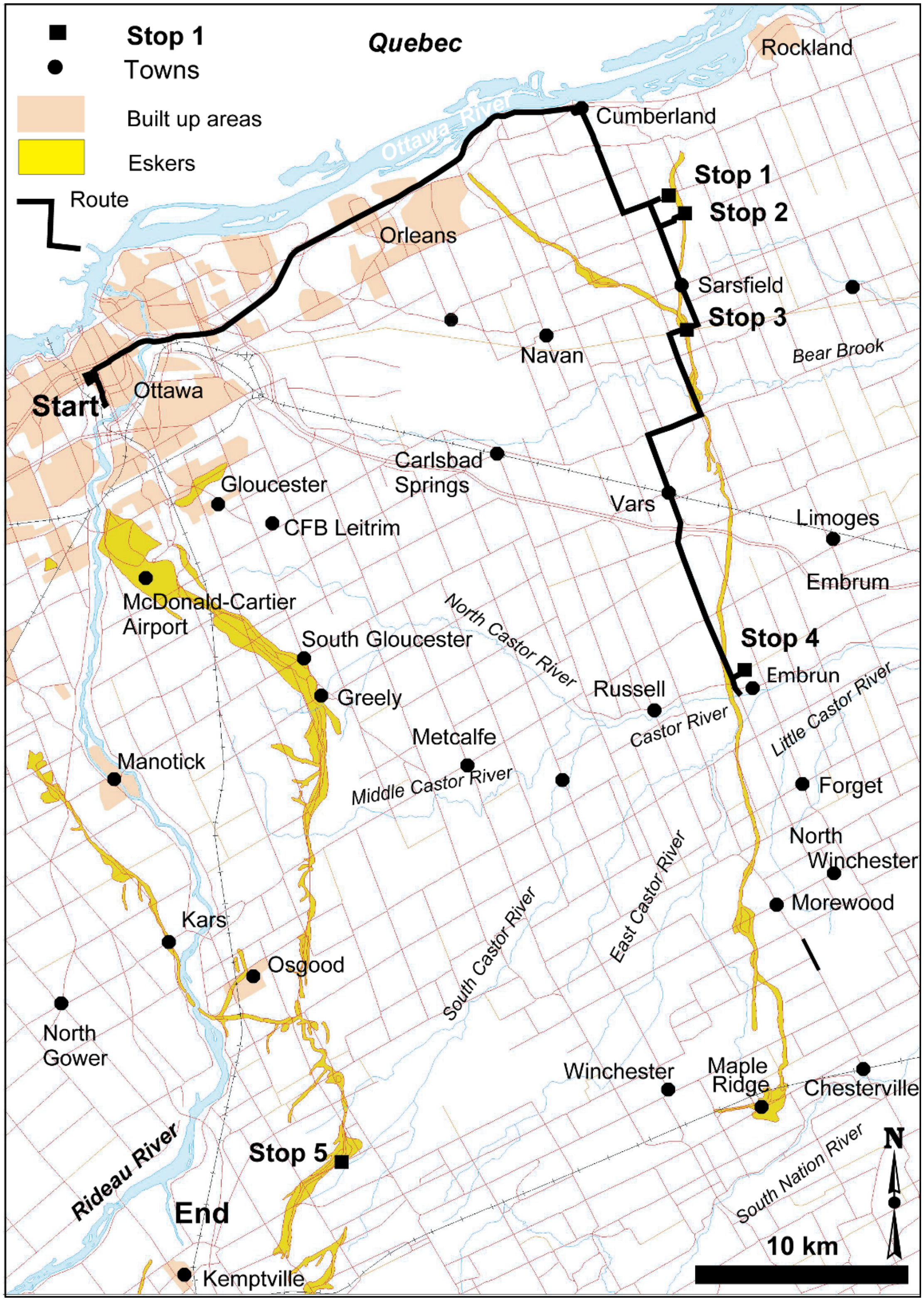
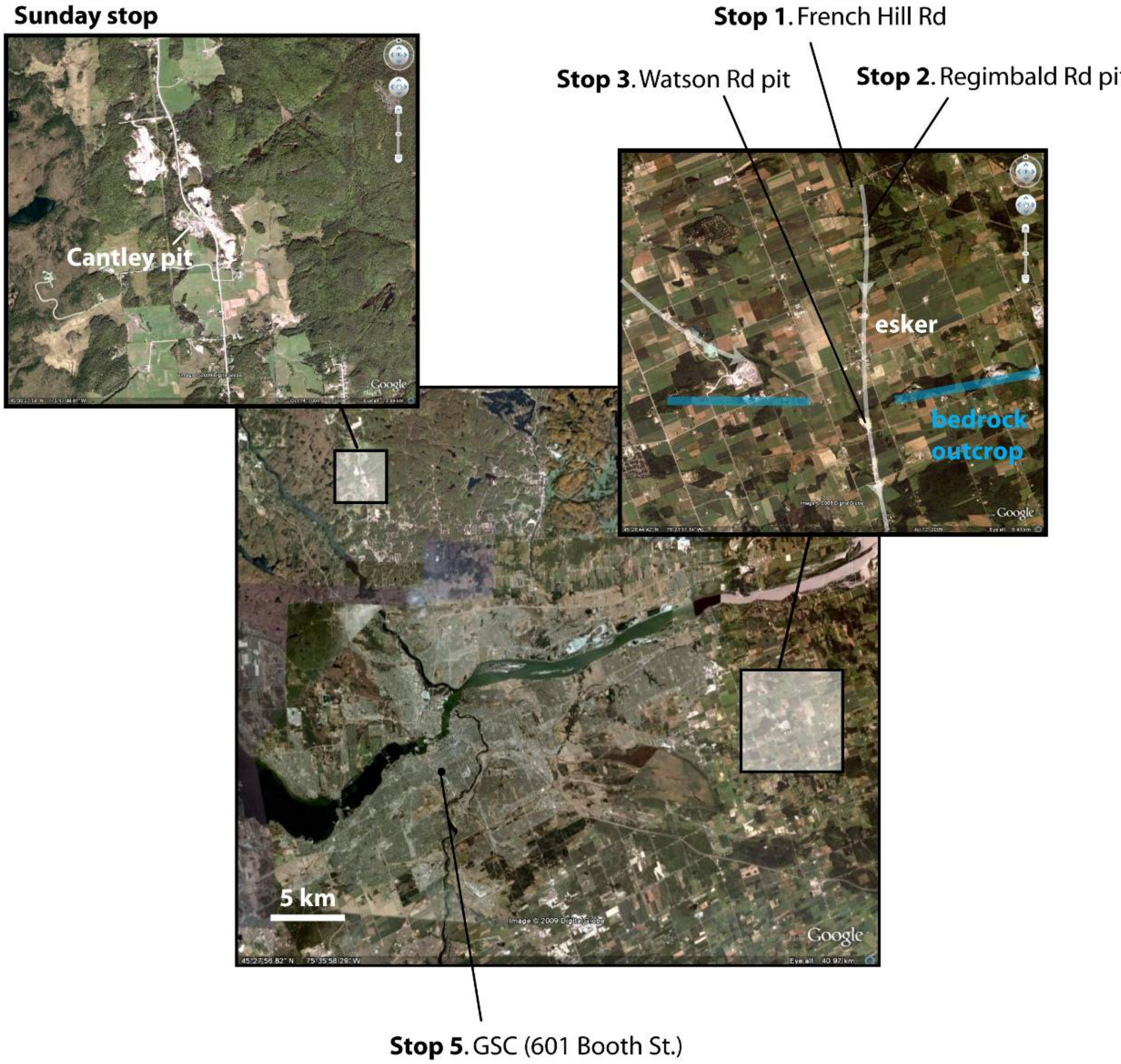
FIELD TRIP STOPS





Acknowledgements

Much of the work presented on the Vars–Winchester esker was funded by a collaborative agreement between the Geological Survey of Canada and the South Nation and Raisin rivers conservation authorities. Data support by the Ontario Ministry of Natural Resources and the Ottawa–Carleton Regional Municipality is acknowledged, as is the permission of Laurent Leblanc Inc. for access to their pit for field-trip stops. D. Ponomarenko calculated paleodischarge for the Ottawa River incised valley and analyzed air photos. André Martel of the Canadian Museum of Nature graciously identified several shell fragments collected from the esker. Alain Plouffe (GSC-Ottawa) is thanked for his constructive review. This publication is a contribution of the Geological Survey of Canada Groundwater Program, Natural Resources Canada.

2009 FRIENDS OF THE PLEISTOCENE MEETING, June 6-7, Ottawa, Ontario

SIMPLIFIED AGENDA

Friday June 5
7:00 PM Icebreaker at the new student residence, 90 University Priv., University of Ottawa campus (the location where most of you are staying). Once inside lobby, follow signs to icebreaker.

Saturday June 6
7:45 AM Meet at bus in front of 90 University Priv. Depart for field.
5:00 PM Return to 90 University Priv
6:30 PM Banquet supper, Room 512, 90 University Priv

Sunday June 7
8:20 AM Meet at bus in front of 90 University Priv. Depart for field.
12:00 PM Lunch
1:30 PM Return 90 University Priv.

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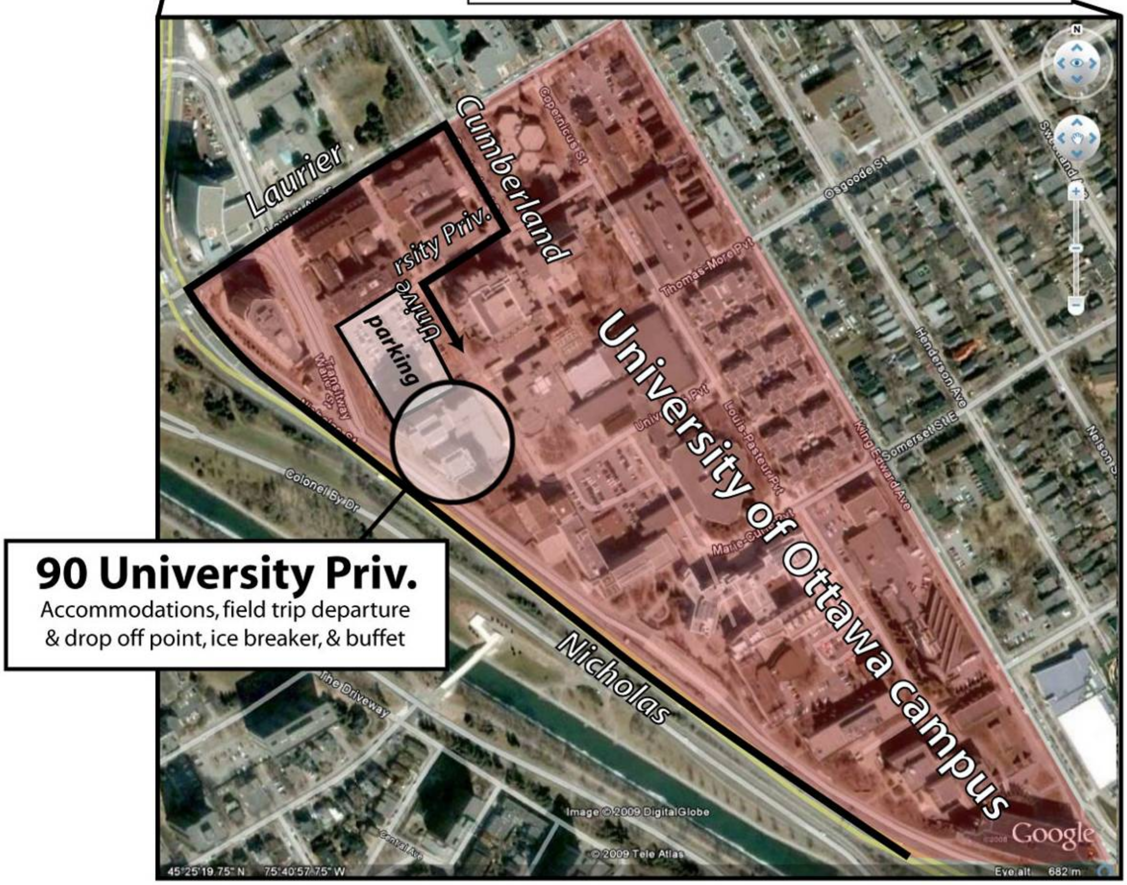
To get to the new student residence (90 University Priv.) from Queensway (Hwy 417)
1) Take Nicholas exit. 2) Right on Laurier. 3) Right on Cumberland. 4) Right on University Priv. You can park in lot adjacent to 90 University. (You will have to pay for parking…there is really no way around it given the downtown location.)



FIELD TRIP SCHEDULE

SATURDAY

7:45 AM Meet at bus in front of new student residence (90 University Priv, U. of Ottawa)

8:00 AM Bus leaves university.
Proceed via Hwy 417 east toward Orleans changing to Route 17. Travel along the Ottawa River shoreline to Cumberland. Proceed south on rue Cameron, Market and Dunning. Proceed south on Dunning for ~5 km. Turn east on French Hill Road. Stop on flat field just east of 3233 French Hill Road. (~40 km).

Stop 1: French Hill Road.
- 1A. Cummings: Geological overview
- 1B. Pugin and Pullan: Seismic reflection techniques and buried valley
- 1C. Aylsworth: Hazards of the Leda clay

Stop 2: Regimbald Road pit
- 2A. Cummings: Esker sedimentology
- 2B. Hinton: Hydrogeological Review

Stop 3: Watson Road pit
- 3A. Cummings: Esker sedimentology
- 3B. Brooks: Seismic microzonation hazard mapping in the Ottawa area

Lunch at Watson Road pit

Stop 4: Route 300
- Pullan & Cummings: Seismic transect of esker and core

Stop 5: Geological Survey of Canada, 601 Booth St.
- Cummings & Guilbault: Core display & micropaleontology of Champlain Sea

Arrive back at University of Ottawa, Parking Lot K (before 17:00).

6:30 PM Banquet at 90 University St.
Invited speaker Ian Clarke (University of Ottawa)

SUNDAY

8:20 AM Meet at bus in front of new student residence (90 University Priv., U. of Ottawa)

8:30 AM Bus leaves university

Stop 1: Cantley pit – Sharpe: sculpted bedrock

Introduction

The Champlain Sea was an inland arm of the Atlantic Ocean that invaded the St. Lawrence Lowland following retreat of the Laurentide Ice Sheet (Figs. 1, 2). The sea existed for about two thousand years around the turn of the Holocene, its level falling continuously as the crust rebounded isostatically. Although both glacier and sea are now gone, the sediment they left behind preserves a detailed record of the deglacial event history and remains integral to life in the Lowland. It is farmed extensively, mined for aggregate, and used as a substrate for waste disposal. Buried eskers host abundant supplies of potable groundwater and Champlain Sea mud is prone to slope failure.

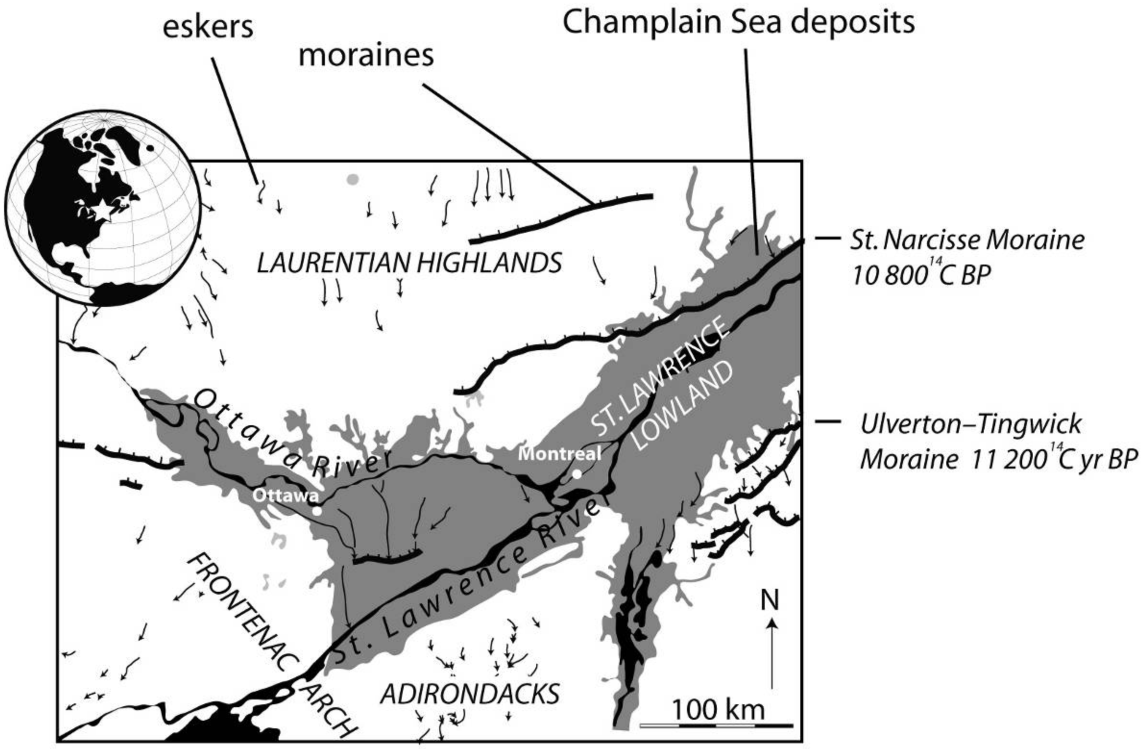


Figure 1. The Champlain Sea basin, a major inland post-glacial mud depocenter in eastern Canada. Aerial extent of fossiliferous Champlain Sea deposits (mostly mud) is from Gadd et al. (1993). Eskers and moraines are from Parent and Occhietti (1988), Barnett (1988), Gorrell (1991), and Simard et al (2003). The diamicton ridge (grounding line moraine) southeast of Ottawa along which several eskers terminate is a newly identified feature.

The Geological Survey of Canada has worked in the Champlain Sea basin for over 100 years, accumulating an extensive body of outcrop, core, and seismic data in the process. Field trip stops will draw from this collective experience, and will touch upon key controversies surrounding the deglacial event history of the basin. Fundamental hypotheses on the origin of sculpted-bedrock forms, eskers, drumlins, and mud-rich glaciated basin fills will be discussed. The nature of the natural hazards and groundwater systems particular to the geological setting will be described. Classic field stops will be visited, including the world-class Cantley sculpted bedrock site.

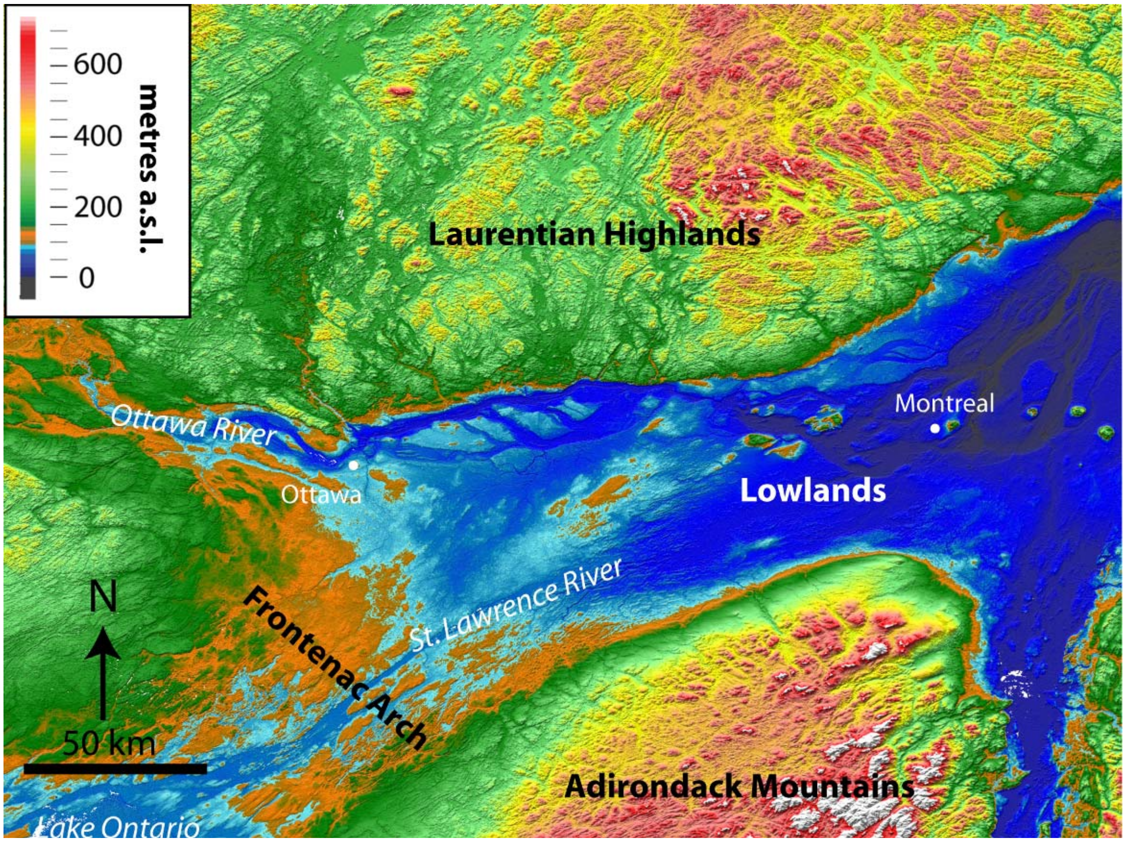


Figure 2. Landscape around the Champlain Sea basin, eastern Ontario. Uplands consist of Precambrian igneous and metamorphic rock covered by a sandy, carbonate-poor till veneer. The sediment is much thicker in the Lowland (average ~10 metres, locally >170 metres), due primarily to the enormous supply of carbonate-poor mud to the basin after ice had retreated into the uplands.

Following the pioneering work of Johnston (1917), most workers have identified three main stratigraphic units in the Champlain Sea basin near Ottawa: drumlinized till, north-south–trending eskers, and Champlain Sea mud with minor sand near the bottom and/or top (Fig. 3). Early work focussed on the mud. De Geer (1892) mapped its distribution, Dawson (1893) studied its macrofossil content, and Antevs (1925) described rhythmites (“varves”) at its base. Subsequent workers investigated the lithostratigraphy of the mud package (Gadd, 1961, 1986; Shilts, 1994; Ross et al., 2006), in addition to its porewater salinity (Torrance, 1988), seismic stratigraphy (Shilts, 1994; Ross et al., 2006), microfossil content (Anderson et al., 1985; Rodrigues, 1988, 1992; Guilbault, 1989; Shilts, 1994; Ross et al., 2006), and geotechnical properties (Fransham and Gadd, 1977; Douma and Nixon, 1993; Aylsworth et al., 2000, 2003). Starting in the 1970s, and continuing until the late 1980s, Brian Rust and his students, along with several additional workers, studied most of the eskers in aggregate pits throughout the basin (Rust and Romanelli, 1975; Rust, 1977; Cheel and Rust, 1982, 1986; Burbridge and Rust, 1988; Sharpe, 1988; Gorrell and Shaw, 1991; Spooner and Dalrymple, 1993). Gorrell (1991) extended this work considerably by mapping all the eskers in both surface and subsurface using outcrops, aerial photos, and uncored water wells (Fig. 4). Kettles and Shilts (1987) studied the till east and north of Ottawa, and MacPherson (1968) and Catto et al. (1982) investigated the incised valley within which the modern Ottawa River sits. Richard played a major role in mapping surficial sediment in the region (e.g., Richard, 1982a,b). A number of key papers can be found in Gadd (1988).

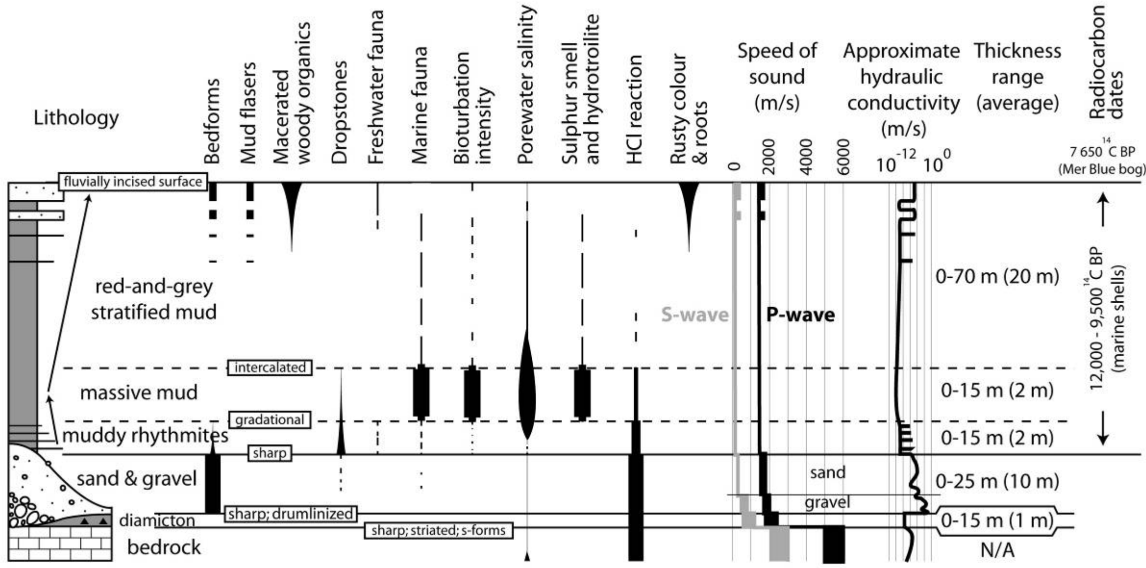


Figure 3. Physical, chemical and biological attributes of Quaternary strata in the St. Lawrence Lowland near Ottawa (idealized). Based on data from Johnson (1917), Richard (1982a,b), Gadd (1986), Rust (1987), Torrance (1988), Rodrigues (1988), Guilbault (1989), Douma and Nixon (1993), Gorrell (1991), Shilts (1994), Aylsworth et al. (2000), Aylsworth et al. (2003), INTERRA (2005), Hunter et al. (2007) and Cummings and Russell (2007).

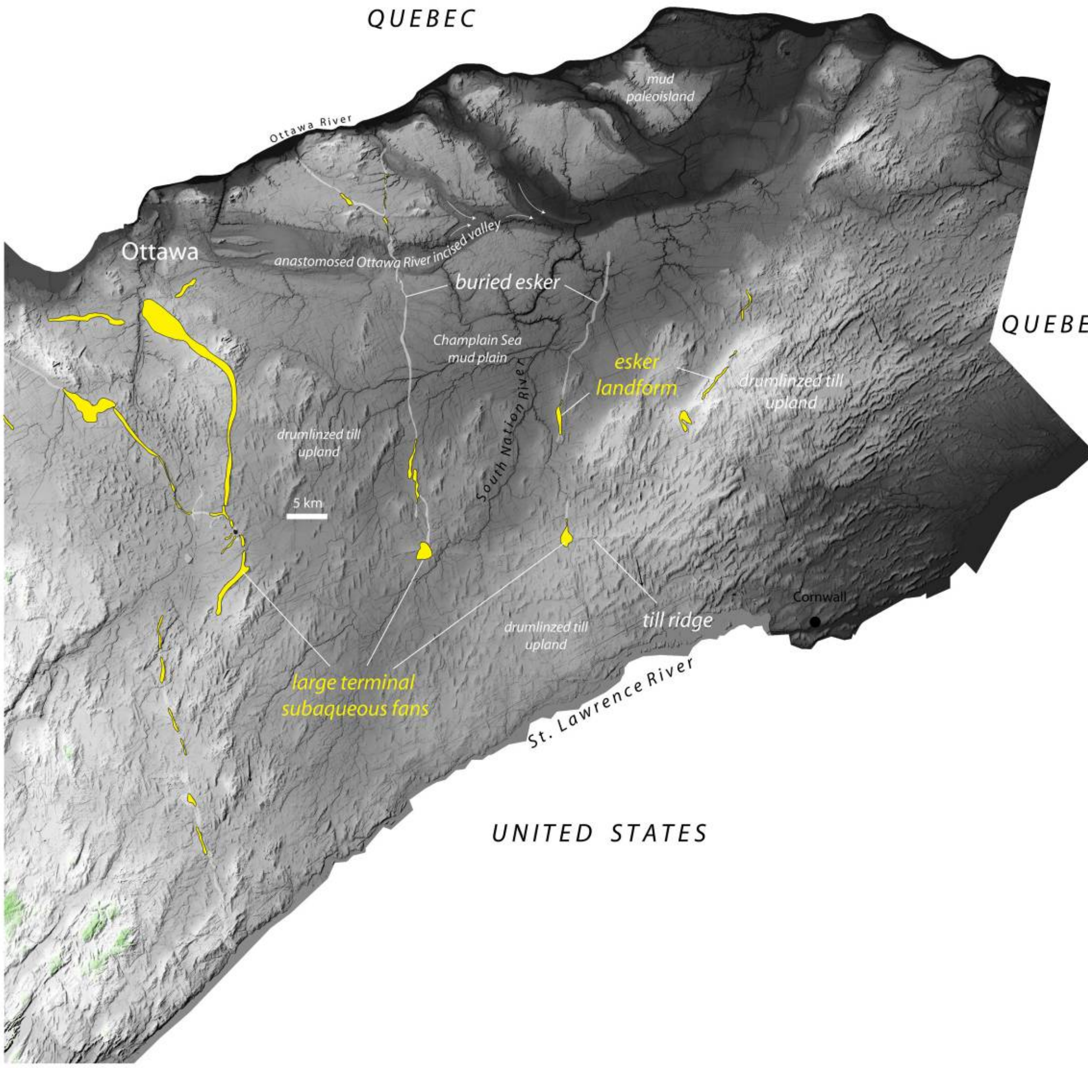


Figure 4. Eskers in the Champlain Sea basin, eastern Ontario (modified from Gorrell, 1991). Note the diamicton ridge at terminal end of eskers (interpretation: grounding line moraine). The second esker from the left is the Vars–Winchester esker (Stops 1 to 4 on Saturday). It is approximately 50 km long.

Several key hypotheses emerge from this body of work, as listed below. Some are controversial, some less so. As new data have been collected over the past few years, several of these hypotheses have been questioned and new hypotheses developed. This fieldtrip affords an opportunity for participants to assess the data supporting these hypotheses and to propose and discuss alternative interpretations.

SUBGLACIAL MELTWATER AS A GEOMORPHIC AGENT

Glacial meltwater was an important geomorphic agent during deglaciation. For example, eskers in the basin are universally viewed as being meltwater-generated. Meltwater is also inferred to have modified the substrate beneath eskers: till is commonly absent and s-forms (e.g., Cantley pit) commonly ornament the bedrock surface. By contrast, the effects of subglacial meltwater in off-esker locations is more controversial. For example, did regional meltwater events (subglacial sheet floods) erode drumlins and s-forms (e.g., Shaw and Gilbert, 1990)?

ICE RETREAT PATTERN

Most workers believe that the ice front retreated northward across the basin during deglaciation like a window blind. However, moraines demarcating this retreat have not been previously identified between Ottawa and Montreal (but see below). Gadd (1988) proposed an alternative interpretation, namely that a calving bay extended up the St. Lawrence River to a position near Ottawa, effectively unzipping the ice sheet in two. Kettles and Shilts (1987) document abundant sand stringers in till west of Ottawa, which they suggest may indicate downwasting. How did the ice retreat?

PROGLACIAL WATER BODY: LAKE OR SEA?

The existence of an ice-contact water body in the Lowland at the time of ice retreat is universally recognized in the literature. Initially, the water body is believed to have been a glacial lake, then later, when ice retreated from a topographic constriction at Quebec City and the lake drained, the Champlain Sea. The timing and ice front position during the lake drainage event is controversial because no deposits or geomorphic features—moraines, spillways, or otherwise—have been identified. Evidence is largely based on the presence of *Candona subtriangulata*, a benthic freshwater ostracode, in rhythmites interpreted to be glaciolacustrine varves at the base of the Champlain Sea mud package. Others have argued that a lake did not precede the sea near Ottawa, and suggest the rhythmites may be glaciomarine deposits (Gadd, 1988; Sharpe, 1988).

ESKER PALEOHYDRAULICS

Eskers were deposited in the basin as the ice retreated. Several of them terminate along a newly identified diamicton ridge interpreted to be a grounding line moraine. What instigated moraine building and esker deposition? What discharges and conduit diameters were involved in esker deposition? Note that previous estimates on discharge and diameter for similar eskers elsewhere vary by several orders of magnitude.

STRATIGRAPHIC PATTERNS IN THE BASIN FILL

Distinct litho-, chemo-, bio-, and seismic stratigraphic patterns are observed in the basin fill succession (Fig. 3). Can they all be attributed to sediment supply and accommodation space changes that accompanied ice-margin retreat? What about autocyclic events? Shoreline translation is the main control on stratigraphic packaging in non-glacial basins—was it important in generating the known stratigraphy?

PROGLACIAL MELTWATER DISCHARGE

The Ottawa–St. Lawrence River corridor has long been interpreted to be a major continent-to-ocean meltwater drainage pathway. Surprisingly, the fact that the Ottawa River sits in a huge incised valley (MacPherson, 1968) has been virtually ignored. Was this valley carved by a large, rapid meltwater pulse during deglaciation? Alternatively, was it carved by a smaller river over a longer period of time? When was it carved? Is there evidence for similar, earlier events in the Champlain Sea mud package?

CHAMPLAIN SEA MUD AND SEISMIC ACTIVITY

Sensitive Champlain Sea mud has long been recognized to have a pivotal role in retrogressive landslide activity in the area. More recently the role of relatively steep-sided bedrock basins, thick unconsolidated fills, and seismic activity has been implicated in more widespread terrain disruption. The complex relationship between bedrock topography, sediment cover and thickness, and seismic amplification is motivating extensive work on microzonation in the basin. What are the key factors that control slope instability and earthquake shaking in this geological setting?

Fieldtrip Stops
Saturday

Stop 1A. Geological Overview

Don Cummings[1], Geological Survey of Canada

This section summarizes the results of a recent large-scale subsurface study of Quaternary strata in the vicinity of the Vars–Winchester esker, a high yield aquifer east of Ottawa (2nd esker from the left in Fig. 4). It is designed to provide an overview of the basin stratigraphy and to give participants context for the field stops. Results are based on an integrated dataset of seismic transects, cored and uncored wells, outcrops, and aerial images.

Bedrock. Bedrock in the vicinity of the Vars–Winchester esker consists of Paleozoic limestone that crops out in east-west trending ridges north of French Hill Road. North of Watson Road pit, wells intercepted non-fissile carbonate mudstone that commonly contains skeletal fragments and wispy shale layers (0.1–2 cm). South of the Watson Road pit, wells intercepted carbonate mudstone that is typically massive and devoid of fossils or terrigenous material. Shale was encountered in one well just north of Embrun.

Bedrock surface. The bedrock surface was not examined during the study, but has been investigated by previous workers in adjacent areas (e.g. Sharpe, 1979; Ross et al., 2006). Smooth, unweathered bedrock surfaces are commonly striated (Fig. 5). Striations are spaced millimetres apart, are decimetres to metres in length, and are less than one millimetre in depth. Most workers believe that they form by differential movement of asperities (hard clasts) in basal ice over bedrock, and that different populations of striations record different ice-flow directions.

In the immediate vicinity of the Vars–Winchester esker, bedrock striations are oriented nearly north-south (Sharpe, 1979). Near Montreal and on the north face of the Adirondacks, a second population of striations that trend northeast-southwest is observed. Most authors argue that north-south striations are related to regional ice-flow during the last glacial-maximum, whereas younger striations record topographically steered flow after the onset of ice-sheet thinning (e.g., Ross et al., 2006). A clear reconstruction of the ice-flow event sequence based on striations data, however, is muddled by inconsistent cross-cutting relationships; for example, northeast-southwest striations commonly cross-cut north-south striations, but also locally appear to be cross-cut by them.

[1] Cummmings, D.I., 2009. Stop 1A. Geological Overview; *in*. Russell, H. A. J. and Cummings, D. I. (compilers), Deglaciation Of The Champlain Sea Basin, Eastern Ontario 72nd Friends of the Pleistocene Field Guide, June 6 – 7, 2009, p. 13–26.

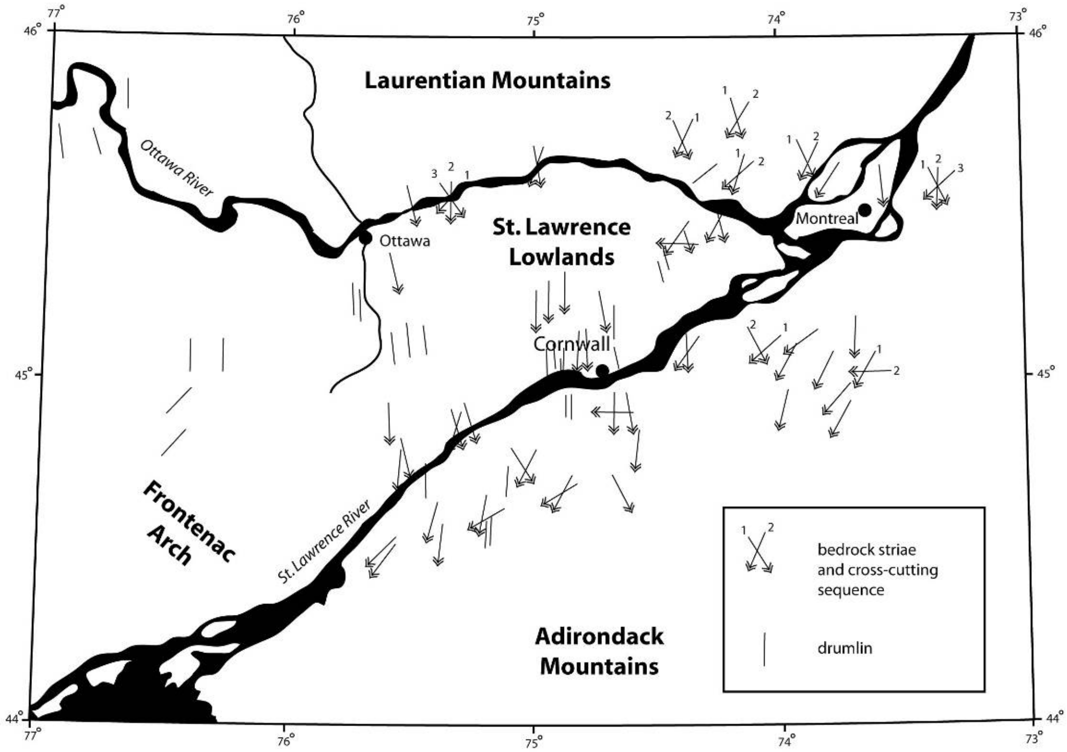


Figure 5. Orientations of drumlins and selected striations on bedrock. Modified from Ross et al (2006).

The bedrock surface beneath eskers near Ottawa is commonly sculpted into various forms, such as potholes, flutes, cavettos, sichelwannen and muchelbruchen (Henderson, 1988; Sharpe and Shaw, 1989). (The bedrock surface beneath the Vars–Winchester has never been observed because of the high groundwater table.) These sculpted forms, or *s-forms*, which commonly occur in bedrock-floored rivers, are also commonly observed in off-esker locations, even on higher ground (e.g., Gilbert, 2000). In order for s-forms to be generated, particles in the flow must spontaneously move at high angles to the mean flow over short distances (centimetres to metres) without the aid of pre-existing obstacles (e.g., bedrock asperities). In other words, the flow has to be turbulent. Flows in the atmosphere and hydrosphere are almost invariably turbulent because of the high inertia-to-viscosity ratios (Reynolds numbers) of naturally flowing air and water. Glacier ice, however, deforms in an extremely slow, laminar (non-turbulent) fashion (as would any subglacial "till slurry") because of the extremely low inertia-to-viscosity ratio of flowing ice. Macroscopic transverse motion may occur, but only if instigated by a pre-existing obstacle. As such, s-forms are interpreted to have been eroded by flowing meltwater at the base of the glacier.

The bedrock surface near the Vars–Winchester esker generates a laterally continuous, high-amplitude seismic reflection that is typically well resolved, except in some places where thick gravel overlies bedrock (Fig. 6). In locations some distance from the Ottawa River (south of the Watson Road pit), the bedrock reflection is relatively flat. By contrast, closer to the Ottawa River (north of the Watson Road pit), the bedrock reflection becomes highly irregular. Here, bedrock ridges (5–10 m high) oriented parallel the Ottawa River (and parallel to the structural fabric) protrude locally through the mud plain, and bedrock cliffs up to 10 m high outcrop on the river-facing sides of paleo-islands in the Ottawa River incised valley. In addition, Sharpe and Pugin (2007) interpret that a ~ 60-metre-deep bedrock valley extends north-south across the area based on seismic and well log data (see Stop 1). Given these observations, it seems probable that a number of flows (ice *and* meltwater) coming from both the north (when the area was subglacial?; see Sharpe and Shaw, 1989; Sharpe and Pugin, 2007) and down ancestral courses of the Ottawa River (when the area was ice free?; see Teller (1988) and below) sculpted the bedrock near the Ottawa River over multiple glaciations, accentuated the structural grain locally, and generated a complex, composite, irregular erosional surface.

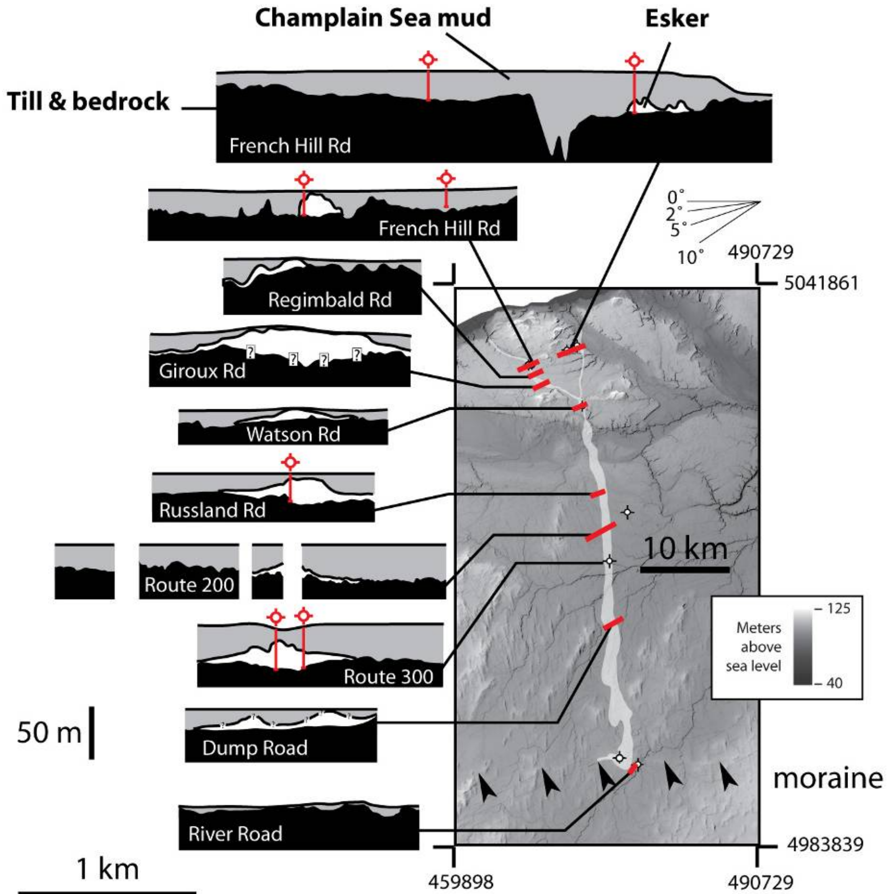


Figure 6. Interpreted seismic transects from the Vars–Winchester esker. Data were collected, processed, and interpreted by André Pugin and Susan Pullan.

Diamicton (till). In wells, stiff diamicton (average 1–3 m thick) was intersected locally above bedrock (Figs. 7, 8). Based on photos, it contains on average 10% gravel and 90% matrix, and the matrix (i.e., < 2 mm fraction) consists of on average of 40% sand, 50% silt and 10% clay (Fig. 9). In core, the diamicton is typically grey except at the Watson Road pit where it is orange-brown. It is typically massive, and shows no obvious upward change in texture or lithology. Carbonate mudstone clasts of local derivation predominate (> 90%), with subordinate amounts of igneous clasts that were likely ultimately derived from the Precambrian Shield to the north, and sandstone clasts that may have ultimately been derived from the Nepean Formation (Cambrian), which outcrops to the north (e.g., Buckingham) and east (Williams, 1991). Clasts are sub-angular to angular. Where greater than a few meters thick, the diamicton unit generates multiple, parallel, high amplitude reflections that can mask the underlying bedrock reflection. Where sufficiently thick, the diamicton therefore in some cases defines the top of acoustic basement.

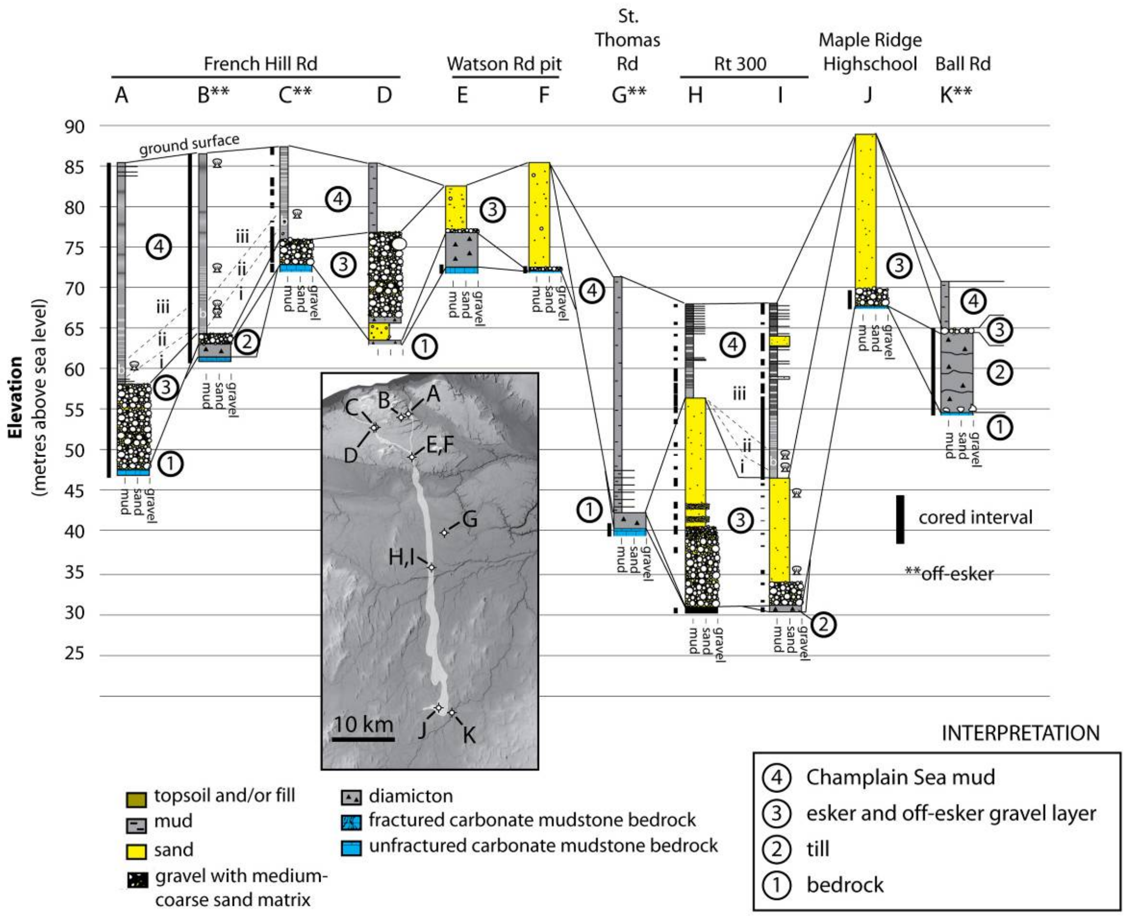


Figure 7. Wells drilled in the vicinity of the Vars–Winchester esker between 2006 and 2007. i = grey carbonate-rich rhythmites; ii = massive bioturbated mud; iii = rhythmically stratified, carbonate-poor red and grey mud; b = bioturbation. Shells indicate locations where *Portlandia arctica* shells were observed.

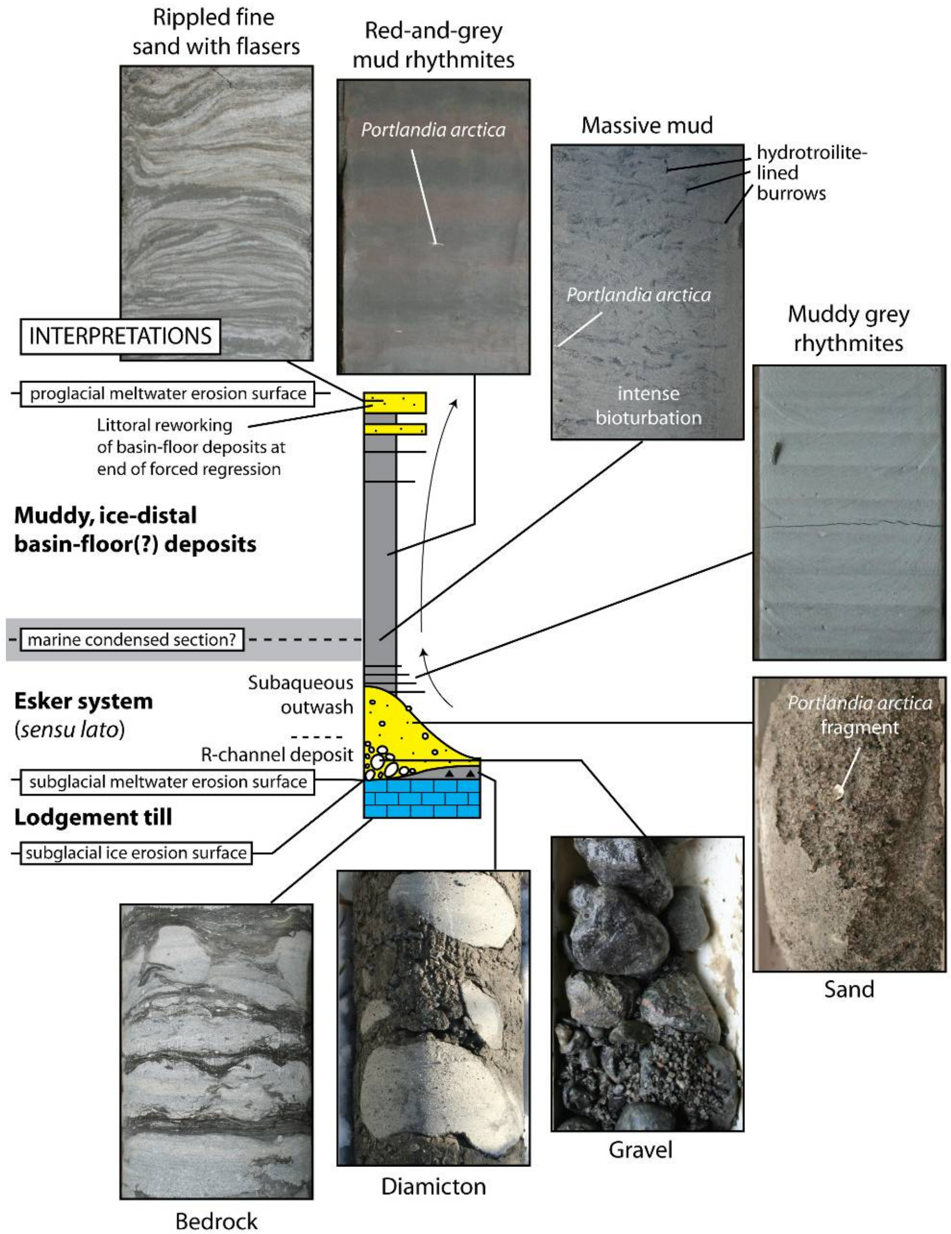


Figure 8. Representative photos of sediment facies observed in core and interpreted depositional environments.

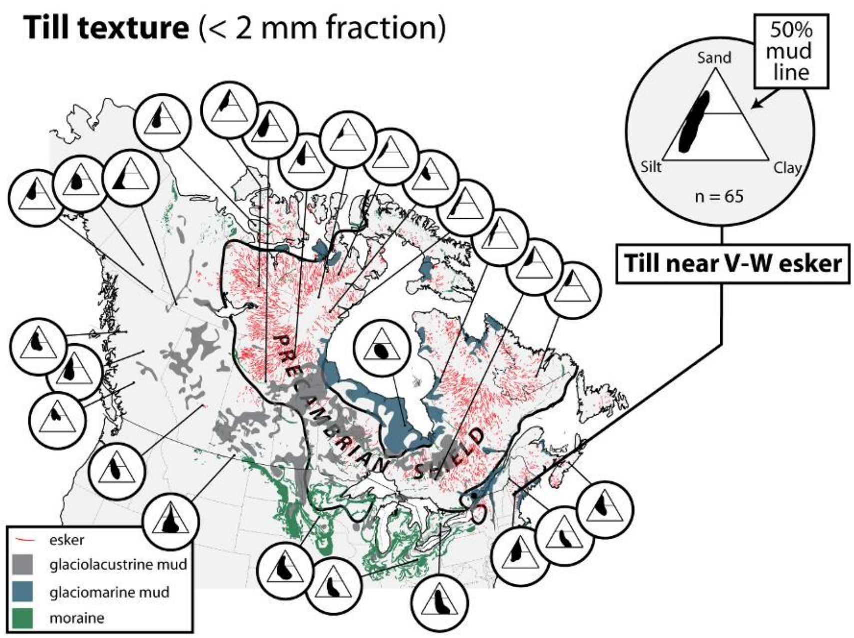


Figure 9. Till texture near the Vars–Winchester esker plotted relative to till texture throughout North America. Data from Shepps (1953), Flint et al. (1959), Thompson (1975), Scott (1976), Sharpe (1979) and references therein, Clague (1984), Dredge et al. (1986), Dyke (1990), Weddle (1991) and references therein, Jackson (1994), Fulton et al. (1995), Metzger (1994), Parent and Paradis (1994, 1995), Dredge et al. (1999), Bednardski (1999), Henderson (2000), Plouffe (2000), McMartin (2000), Klassen (2002), McMartin et al. (2003), McMartin et al. (2006), Thorleifson et al. (2007), and McMartin et al. (2008).

Regionally, well and seismic data suggest the diamicton forms a thin veneer that drapes the bedrock surface; it does not obviously thicken into bedrock lows or thin over bedrock highs. Locally, it is sculpted into north–south oriented drumlins, which are most easily visible on the diamicton-covered uplands that rise above the Champlain Sea mud plain. Near Winchester, a previously unidentified, 3 meter high by 45 kilometer long ridge is superimposed on the drumlins (Fig. 4). Drumlins are not different in appearance or orientation on either side of the ridge. Cores suggest the ridge consists of diamicton that is lithologically and texturally similar to the diamicton veneer to the north, although it contains several mildly deformed interstratified fine sand layers, each less than 2 cm thick. Three eskers, including the Vars–Winchester esker, terminate along this ridge in broad sandy fans (Fig. 4).

Previous authors interpret the diamicton to be a subglacial till (e.g., Johnston, 1917; Sharpe, 1979; Richard, 1982a,b; Kettles and Shilts, 1987), an interpretation that is supported by the observed angular clasts, massive appearance, overconsolidation and sheet-like geometry. Massive diamictons are thought to form by two sometimes-coeval processes, subglacial deformation (shear homogenization) of pre-existing sediment (Elson, 1961; Hart, 1995; Cummings and Occhietti, 2001), and lodgement ("plastering") of sediment from the glacier onto the substrate (Chamberlain, 1895). In the case of the massive till near the Vars–Winchester esker, lodgement seems like the most viable mechanism since deposition appears to have been preceded by erosion. Deformation seems less likely to have generated the till given the presence of rare, mildly deformed thin sand layers. If the till indeed formed by lodgement, abundant meltwater must have been produced because debris-rich basal ice tends to consist predominantly (i.e., typically over 90% and at least 50%) of ice (Kirkbride, 1995). The rare, thin, meltwater-deposited sand layers may in part attest to this. Their general paucity, however, in addition to the overconsolidated nature and massive appearance of the diamicton, suggests that till deposition did not occur by wholesale, regional melt out of debris from stagnant, stratified basal ice (cf. Kettles and Shilts, 1987).

The diamicton ridge along (and over) which the eskers terminate is interpreted to be a grounding line moraine. Its apparent superposition on drumlins suggests the drumlins formed earlier, when the ice front lay farther to the south. The exact timing of ridge deposition is uncertain, though it presumably formed sometime after the nearest moraine to the south (12 000 ^{14}C BP) and before the St. Narcisse Moraine, the nearest moraine to the north (10 800 ^{14}C BP) (Richard and Occhietti, 2005). Ridge (2004) speculates that the ice front had reached this location by 11.5 ^{14}C ka BP.

Top of diamicton (drumlins). The top surface of the regional till, where exposed, is sculpted into elongate, parallel, streamlined ridges that are oriented north-south, and are on average 3–10 m high, 0.5 km wide and 1 km long (Fig. 4). Although this surface was not investigated in detail, we concur with previous authors that the streamlined forms are drumlins (e.g., Sharpe, 1979). Given that s-forms commonly ornament the bedrock surface where till is absent, both beneath eskers (Henderson, 1988) and adjacent to eskers (Sharpe and Shaw, 1989; Gilbert, 2000), and that a thin (< 2 m) gravel sheet occurs above the regional till and beneath Champlain Sea mud in the north part of the study area (Sharpe and Pugin, 2007), the meltwater hypothesis, which argues that drumlins were carved by meltwater floods beneath the ice sheet (e.g., Shaw, 1996), cannot be ruled out. Arguments presented by proponents of the meltwater hypothesis (Sharpe and Shaw, 1989; Shaw et al., 1996; Sharpe and Pugin, 2007) therefore need to be weighed against those of workers who believe the drumlins formed by the action of ice (Ross et al., 2006). Although we find the physics-based arguments of proponents of the meltwater hypothesis (e.g., Shaw, 1996) in general more convincing than those presented by supporters of the ice hypothesis, and argue that large lakes do exist and can drain catastrophically beneath ice sheets (e.g., Wingham et al., 2006; Siegert et al., in press), in reality, the facies architecture relative to the form of the drumlins needs to be investigated before a final conclusion can be drawn. However, the thin, highly permeable gravel sheet present locally off-esker over till and below Champlain Sea mud is difficult to explain unless a sheet-like subglacial discharge of meltwater is invoked[2].

Sand and gravel (esker). The new subsurface data suggest that the Vars–Winchester esker consists of two key elements, a **gravelly central ridge** and a **sandy fan carapace** (Fig. 9, 10). Neither element generates coherent seismic reflections internally, but the boundaries of the esker "container" are commonly well resolved in seismic transects, an exception being where thick gravel obscures underlying sediment and/or bedrock (see Stop 4).

The gravelly central ridge of the Vars–Winchester esker is 2 to 20 metres high (average 15 m) and 100 to 200 metres wide (average ~150 m). Its cross-sectional area varies, but does not systematically increase or decrease along the esker. The flanks of the gravelly central ridge dip between 10° and 30°. Clast lithologies are similar to the adjacent till: carbonate mudstone predominates, with minor percentages of igneous (granite), sandstone, and (surprisingly friable) shale clasts. Where intersected in wells, the gravelly central ridge typically overlies bedrock[3]. Locally, it appears to bifurcate and rejoin over several kilometres. Bifurcation and local widening of the gravel ridge is especially apparent in the north end of the study area, on northward-dipping bedrock surfaces. Although the gravelly central ridge is interpreted to be present in all new seismic transects that cross the esker, the continuity (and therefore hydraulic connectivity) of the gravels between seismic transects is difficult to ascertain, leaving open the possibility that breaks may be present. Where exposed, the gravelly central ridge consists of well-rounded pebbles and cobbles organized into thick (1.5–2.5 m) high-angle (dune) cross-stratified beds that dip southward, parallel to the esker axis. Thick (1–2 m) sand layers are present locally (see Stop 2).

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[2] Note that up to 2 metres of bouldery gravel was also intersected off-esker between till and Champlain Sea mud in long cores collected west of Montreal (see Ross et al. 2006). Barnett (1988) identifies off-esker gravel beneath till and Champlain Sea mud east of Ottawa. Veillette et al (2007) observe permeable off-esker gravel in the Abitibi clay basin.

[3] Gorrell (1991) notes that the largest clasts (commonly rounded boulders) tend to occur at the base of the gravelly central ridges. A subtle fining upward trend was noted in one of the Route 300 cores, but it is unclear if such trends are present everywhere along the gravelly central ridge.

The gravelly central ridge is interpreted to have been deposited in a meltwater conduit (R-channel) that was thermally eroded (and corraded?) into ice at the base of the glacier (Fig. 10). It is considered to be the most proximal element of the esker system. Lithologic similarity of its clasts and clasts in the till suggests the gravelly ridge was sourced in part from the till; basal ice likely contributed sediment also, given volumetric considerations. Bifurcation and subtle widening of the ridge on northward-facing slopes may reflect a decrease in the wall-melting–to–viscous-heating ratio in the R-channel associated with the steeper, adverse bedrock inclines (Röthlisberger, 1972; Shreve, 1972, 1985). The flows must have been relatively fast and deep, given that they carried boulders up to 1 m in diameter and deposited high-angle cross-beds between 1.5 and 2.5 m in height. If scaling relationships for fluvial environments are assumed to apply, the flows likely traveled at speeds of several metres per second[4] and that the R-channel may have been several 10s of metres high[5]. This in turn suggests discharges of 1000s of cubic metres per second[6]. Discharge may have therefore been similar to that of the modern Ottawa River (~ 2000 m^3/s) and up to two orders-of-magnitude greater than R-channel discharge in modern (small) alpine glaciers during melt season (e.g., Østrem, 1975; Hooke et al. 1985).

Sandy fan deposits with variable amounts of pebble gravel sharply overlie the gravelly central ridge locally, forming a carapace that is relatively wide (0.5–2 km) and gentle-flanked (1–5°). The longitudinal extent of the sandy fan carapace is poorly constrained, but based on seismic data it appears to be present in most places along the esker south of Watson Road pit. Core data suggest the distal parts of the sandy fan carapace commonly overlie till. In outcrop, the sandy fan carapace is composed of multiple, mound-shaped, upward-coarsening units (“lobes”) that are stacked compensationally[7] on top of each other; sharp-based units, commonly gravelly, interrupt this motif locally (see Stop 4)[8]. Climbing ripples, dunes (which also rarely climb), low-angle (antidune) cross-strata, and diffusely laminated channel fills are common. In a continuous core from Route 300, marine shell fragments (mostly *Portlandia arctica*)[9] were observed at two intervals in the sandy fan carapace (see Stop 4).

The sandy fan carapace is interpreted to have been deposited by sediment-laden jet–plume pairs (e.g., Fischer et al., 1979; Syvitski, 1989; Powell, 1990; Hoyal et al., 2003; Russell and Arnott, 2003) that were issued from the R-channel into standing, highly freshened water of the Champlain Sea (Fig. 9). The general impression gained from the sedimentology of outcrop exposures is one of rapid sedimentation from rapidly decelerating unidirectional flows: the presence of bedforms attests to the tractive nature of sediment transport, whereas climbing bedforms (ripples *and* some dunes) and diffuse stratification attest to high rates of suspended-sediment rain-out (e.g., Arnott and Hand, 1989). The upward-coarsening, mound-shaped depositional elements at the Watson Road pit (see Stop 3) are interpreted to be progradational fan lobes. Compensational stacking of these sand bodies is interpreted to have been produced by avulsion of the jet–plume depocenter[10]. Sharp-based, ungraded to upward-fining units either reflect progradation of proximal fan channels over distal fan deposits, or abrupt increases in discharge. In either case, these upward-fining units likely correlate downflow to fan lobes.

[4] Calculated using the Helley method (Table 3 in Costa, 1983).
[5] Calculated using the method outlined in Leclair and Bridge (2001).
[6] Assuming a flow width of 125 m, flow speed of 2 m/s, and a (conservative) flow depth of 20 m.
[7] Compensational stacking refers to the tendency for sedimentary systems to preferentially fill topographic lows through deposition,
[8] In adjacent eskers, where more outcrops along the esker exist, architectural elements in the sandy fan are commonly sharp-based and ungraded to upward-fining. It is likely that similar variability exists in fans of the Vars–Winchester esker, but that this variability is simply not observed due to limited exposure.
[9] André Martel from the Canadian Museum of Nature has identified the shells as being from the Yoldidae and Nuculidae families. Using photos for comparison, many of the fragments appear to be *Portlandia arctica*.
[10] The small size of these lobes at the Watson Road pit (Stop 3) may suggest deposition under lower-discharge conditions than at the Kemptville pit (Stop 5) (e.g., Powell, 1990).

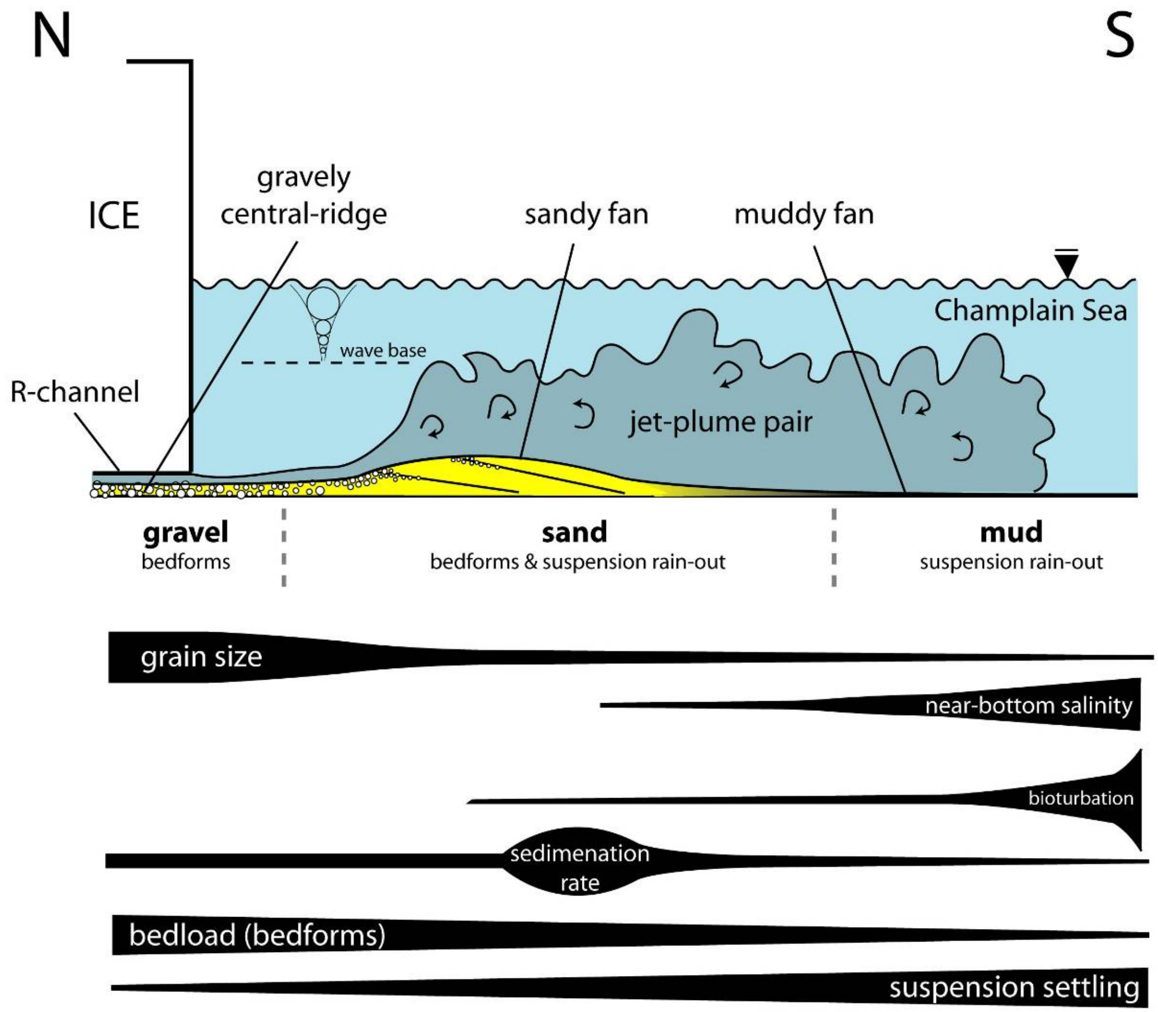

Figure 10. Facies model for the Vars–Winchester esker system. The gravelly central ridge is interpreted to have been deposited in a subglacial meltwater channel (R-channel), and the sandy fan carapace is interpreted to have been deposited where the R-channel debouched into deep, standing, proglacial, marine-influenced water. To fully understand the Vars–Winchester esker, this facies model—a snapshot of the depositional system in time—needs to be integrated with the sequence stratigraphic model (Fig. 11), which takes into consideration how longer-term changes in sediment supply and accommodation space generated the observed stratigraphic succession.

We understand our suggestion that the Champlain Sea existed proglacially during esker deposition will be met with controversy. Most researchers over the past 25 years have argued that a proglacial *lake* existed during esker deposition, and that the basin only became marine after the lake drained into an inland arm of the Atlantic Ocean, which is thought to have encroached up the St. Lawrence River valley through the glacier to a position near Quebec City (Thomas, 1977). The proglacial lake is believed to have been large: it is thought to have extended from Quebec City (Parent and Occhietti, 1988) to Lake Ontario (Clarke and Karrow, 1984), and from south of the Canada–US border up the Ottawa River (Pair and Rodgrigues, 1993). Support for this hypothesis is derived from the presence of rhythmically laminated sediment (“varves”) at the base of the Champlain Sea mud succession that commonly contain only *Candona subtriangulata*, a benthic freshwater ostracode (Anderson et al. 1985; Parent and Occhietti, 1988; Rodrigues, 1992; Pair and Rodrigues, 1993). Some authors disagree with this hypothesis, suggesting instead that the “varves” may be Champlain Sea deposits (e.g., Sharpe, 1988). Despite this, almost all workers agree that the “varves” are likely distal subaqueous outwash fan deposits (e.g., Naldrett, 1988; Gadd, 1988; Pair and Rodrigues, 1993; Ross et al., 2006).

Given the above compelling arguments for the existence of a proglacial lake during esker deposition near Ottawa, we discuss our findings in greater detail. Marine shell fragments were observed 2 metres and 11 metres below the top of the sandy fan carapace in one of two long cores collected just north of Embrun (Well I in Fig. 7). The esker "container" is clearly resolved in the seismic data at this location (see Stop 4). The shell-bearing strata are not reasonably interpreted as having slumped from the esker crest following deposition (11 m is too deep), and are not beach deposits formed during isostatic rebound (the esker is buried by 10 m of mud). The shells were also not introduced during the coring process because the core collected from 11 m below the top of the esker is pristine. It is also very unlikely that the shells were reworked from older sediments because they are *extremely* thin and fragile, and their periostracum (organic coating) is typically preserved. As such, we argue that the most reasonable interpretation, especially given that *Portlandia arctica* is known to thrive in modern, freshwater-influenced prodeltaic environments in Canadian high-arctic fjords (e.g., Syvitski, 1989; Aitken and Gilbert, 1996), is that the marine bivalves simply lived on the subaqueous-outwash fan.

Candona subtriangulata can tolerate high turbidity (Rodgrigues, 1992), but the species has never been identified in modern brackish or marine environments, and is known to tolerate only minor levels of sodium (1–14 mg/L), chloride (1.2–15 mg/L) and sulphate (0.1–27 mg/L) (L. Denis Delorme, personal communication, 2007). However, we believe that Champlain Sea microfaunal assemblages speak for themselves: in addition to occurring by itself, in some places *Candona subtriangulata* is found with or below benthic marine foraminifera in unbioturbated, undisturbed strata (e.g., Cronin, 1977; Hunt and Rathburn, 1988; Rodrigues, 1988, 1992; Guilbault, 1989)[11]. Given that the Champlain Sea was a restricted, inland water body that received an enormous yet highly variable supply of meltwater from the ice-sheet (Marshall and Clarke, 1999), it is not unreasonable to think that near-bottom salinity was low or even fresh in parts of the basin for much of the time, that the salinity front fluctuated significantly in response to astronomic (diurnal, seasonal) and episodic (flood) forcing[12], and that *Candona subtriangulata* was simply able to colonize the seafloor as a result.

As a final note, the (apparent) absence of mud in the subaqueous-outwash fans is striking. However, it is argued that mud should actually be the *dominant* grain-size in the esker system *sensu lato* because mud is the dominant component of the till, the source of esker sediment (Fig. 9). R-channels should have been extremely efficient at transferring mud to the basin—possibly up to 10 times more so than a fluvial system—because they lack floodplain-like sediment storage sites[13]. It is therefore highly likely that the sandy subaqueous outwash fans exposed in the pits correlate to mud distally. The question is, how much?

Mud with minor sand near bottom and/or top (Champlain Sea deposits). Champlain Sea mud buries the esker locally, and forms the surficial sediment unit throughout much of the study area. Three units are identified within the mud succession (bottom to top): 1) rhythmically laminated mud and sand ("varves"), 2) massive mud, and 3) stratified mud, locally with sand layers near or at the top. These three units are commonly stacked on top of each other in "complete" Champlain Sea successions between Ottawa and Montreal (Gadd, 1961, 1986; Shilts, 1994; Aylsworth et al., 2003; Ross et al., 2006).

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[11] Brackish-water ostracodes are observed locally in post-glacial sediments in the Lake Ontario basin (L. Denis Delorme, personal communication, 2007). Unlike freshwater-ostracode eggs, marine-ostracode eggs cannot dry out. They can only migrate into water bodies that are ocean-connected. This suggests the Champlain Sea was likely once confluent with Lake Ontario (e.g., Sharpe, 1979).

[12] The salinity front in estuaries/deltas commonly moves 10s to 100s of km seasonally (e.g., Shanley et al., 1992). In the Amazon River, it is pushed 150 km onto the shelf during high discharge (Geyer et al., 2004). Meltwater discharge into the Champlain Sea may have been similar to that of the Amazon (200,000 cubic metres per second; Marshall and Clarke, 1999), but would have been *much* more seasonal.

[13] 30–90% of a river's sediment load can be trapped in floodplains (Goodbred and Kuehl, 1999).

Where intersected in core, the rhythmically laminated unit consists of thin (< 1 cm), alternating layers of light grey silt or very-fine sand and dark grey mud (Fig. 8). Couplets tend to be sharp-based and normally graded. Bioturbation levels are low to moderate. Rare dropstones are present. The unit fines upward, and reacts strongly with dilute HCl. It sharply overlies the esker. The unit is less than 2 metres thick, which is near the limit of resolution of the seismic data.

Previous studies identify a similar rhythmically laminated unit in the same stratigraphic position throughout the western Champlain Sea basin (e.g., Gadd, 1986; Pair and Rodrigues, 1993). These studies suggest the rhythmites may contain only *Candona subtriangulata* (e.g., Pair and Rodrigues, 1993; Ross et al., 2006), only marine–brackish water fauna (e.g., Shilts, 1994; Ross et al., 2006), or a mix of both (Cronin, 1977; Hunt and Rathburne, 1988; Rodgigues, 1988, 1992). Porewater salinities are commonly low to moderate and may gradually increase upward (e.g., Shilts, 1994; Torrance, 1988).

The rhythmically laminated unit passes gradationally upward into dark grey massive mud that is intensely bioturbated (Fig. 8). The unit commonly reacts strongly with HCl, but less so than underlying rhythmites. *Portlandia arctic* shells are common. Black, vertical to horizontal "squiggles" (0.1–2 mm wide, < 1 cm long) are visible on freshly cut surfaces; these disappear after several hours of exposure to the atmosphere. Freshly cut surfaces have a subtle sulphurous odor. Though not always the case, in nearby wells porewater salinity in the massive mud unit can approach that of the modern ocean (Torrance, 1988). Like the rhythmites it overlies, the massive mud unit appears to be present throughout the western Champlain Sea basin (e.g., Gadd, 1986; Pair and Rodrigues, 1993).

A red and grey stratified mud unit overlies, and commonly intercalates with, the massive mud unit (Fig. 8). In core, the unit consists predominantly of gradationally based couplets that grade from light-grey mud to dark-grey or pinkish-red mud. Black residue, similar to that in the massive-mud unit, is observed in light-grey bands near the base of the succession, and also rarely occurs as discrete layers in pinkish-red bands. Gadd (1986) suggests the light grey bands are coarser than the dark-grey/red bands. Couplets increase in thickness upward, from < 1 cm near the base of the unit to several 10s of centimetres near the top of the unit. The thickest couplets in French Hill Road cores reach 35 cm thickness, whereas the thickest couplets in Route 300 cores are 15 cm thick. Successive light grey mud bands may also become sandier upward. *Portlandia arctica* shells are present, but very rare. In comparison to underlying mud units, the stratified mud unit reacts less strongly with dilute HCl, and commonly not at all, with red layers reacting slightly more than light-grey layers. The upper 1–5 m below ground surface is orange-brown, is stiffer and dryer than underlying mud, and may contain root traces, joint-like structures, and very fine sand beds (< 20 cm thick) that consist of stacked current ripple cross-sets.

Previous studies identify a similar red and grey stratified mud unit in the same stratigraphic position throughout the northern part of the western Champlain Sea basin (e.g., Gadd, 1961, 1986; Aylsworth et al. 2003; Ross et al. 2006). The unit is apparently absent in the south half of the basin (e.g., Pair and Rodgrigues, 1993). Based on numerous cores from various localities, Fransham and Gadd (1977) conclude that the stratified mud unit is finer grained than the underlying massive mud unit.

We now interpret the Champlain Sea succession using standard sequence stratigraphic concepts and principles. Sequence stratigraphy is process-oriented sedimentology at the largest scale; its goal is to relate basin-scale patterns in sedimentary strata to long-term changes in two key variables, sediment supply (*S*) and accommodation space for sediment (*A*) (Curray, 1964; Vail et al. 1977; Posamentier and Vail, 1988; Van Wagoner et al.,1990; Posamentier and Allen, 1999). In non-glacial sedimentary systems, horizontal translation of the shoreline forced by sea-level change is the main process that mediates *S* and *A* over long ("Milankovitch-scale") time scales. Shoreline movement is assumed to be slow and gradual. Glacial systems are different in two ways. First, the key interface that mediates gradual changes in *S* and *A* is the ice-margin, not the shoreline (although shoreline translation is also important). Second, rapid events (meltwater floods) are common, and generate rapid changes in *S* and *A*. These two characteristics—that ice-margin translation causes gradual change in *S* and *A*, and that meltwater events punctuate this gradual change—are believed to be the hallmarks of glacial sedimentary systems.

Although the “complete” Champlain Sea succession between Ottawa and Montreal consists of three main lithostratigraphic units (“varves”–massive mud–stratified mud; e.g., Gadd, 1961), we argue that the succession is best subdivided into two genetically related sediment packages, one deposited as the sediment source (the ice-margin) backstepped northward through and out of the Champlain Sea, and one deposited as a new sediment-source, the ice-distal, meltwater-fed shoreline, moved back into the basin as the result of isostatic rebound (Fig. 11D). It is possible that one or more rapid meltwater-events may have punctuated this gradual, ice-mediated sedimentation pattern.

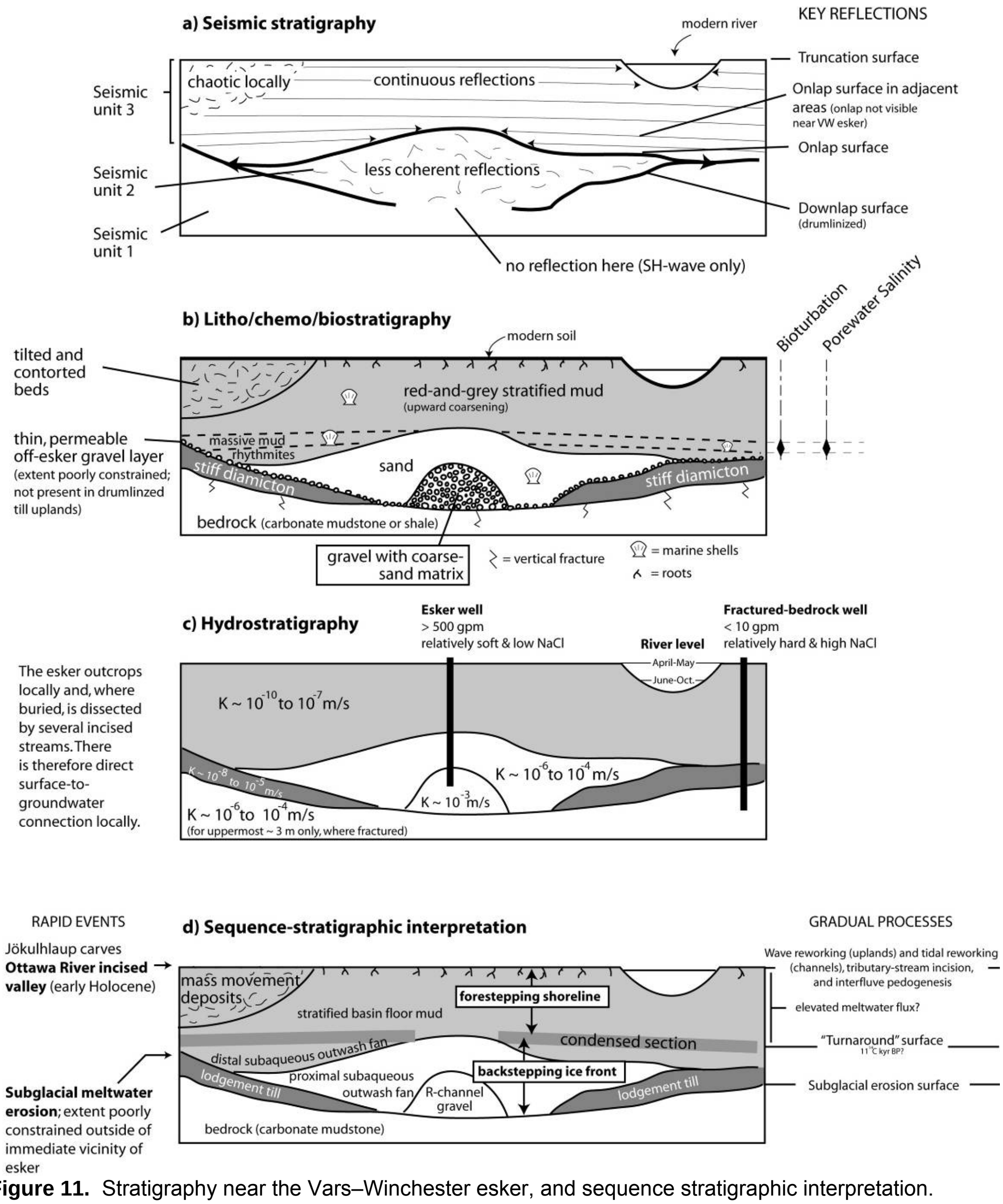

Figure 11. Stratigraphy near the Vars–Winchester esker, and sequence stratigraphic interpretation.

1) Backstepping system
Initially, the ice was in contact with the Champlain Sea. Retreat of the ice front through the sea caused a huge jump in accommodation space when the environment switched from subglacial to proglacial[14]. Continued backstepping of the ice-margin gradually reduced the caliber and supply of sediment to the seafloor, with an associated reduction in the environmental stress on benthic organisms (e.g., salinity stress, turbidity). The sedimentary result of this gradual retreat of the glacier is the carbonate-rich fining-upward succession that starts with the esker gravel and ends with the massive bioturbated mud. Retreat of the ice through marine water adequately explains both the outcrop-scale characteristics of the lower mud package (change in fauna, grain-size, geochemistry, bioturbation level, porewater salinity) and its regional-scale characteristics (deposition of a mud “blanket” throughout the western Champlain Sea, which would be difficult to do if the sediment source was stationary and far away; see Dalrymple and Cummings, 2005; Cummings et al. 2005).

2) Forestepping system
Meltwater flux into to the Champlain Sea is interpreted to have increased abruptly at some point as the ice was retreating onto the Shield, causing unbioturbated, carbonate-poor red-and-grey stratified mud to deposit and locally onlap the massive intensely bioturbated mud unit. Data in Ross et al. (2006) suggest this may have started around 11 ^{14}C kyr BP[15]. Thinning of the stratified mud unit southward, and its apparent absence south of the St. Lawrence River (Pair and Rodrigues, 1993) suggests that sediment was likely supplied by rivers that drained into the northern Champlain Sea (e.g., Ottawa River, smaller rivers that drain the Laurentian Highlands). Forced regression as a result of isostatic rebound is interpreted to have generated coarsening-upward and thickening-upward trends in the unit. If each red-and-grey couplet is a varve (e.g., Gadd, 1986), the sedimentation rate was initially millimetres to centimetres per year, and increased with time, reaching ~ 15 cm/yr along Route 300 and ~ 35 cm/yr along French Hill Road. The low level of bioturbation and paucity of shells support the inference that sedimentation rate was high (MacEachern et al. 2005). Isolated sand beds in the stratified mud unit (< 1 m thick), some of which are associated with higher-amplitude seismic reflections (e.g., Route 300), may have been deposited during meltwater outbursts into the basin.

Gadd (1986) interprets the stratified-mud unit to be a deltaic deposit. This is indeed reasonable based on the core data that were available to him, given the coarsening-upward and thickening-upward trends. However, no obvious downlap is observed in any of the seismic transects. Rather, the stratified mud unit appears to onlap the massive mud unit (e.g., Fig. 4A in Ross et al. 2006). This would not be expected if the unit was deposited by a delta: deltaic deposits consist of clinoforms, albeit commonly very low angle ones (< 1°) if the system is mud-rich (Orton and Reading, 1993). It is possible that seismic transects are not long enough to image the low-angle clinoforms, or that they are oriented obliquely to the progradational axis of the delta(s). An alternative and perhaps more reasonable explanation, given the above observations, is that the red and grey stratified mud unit was deposited on the basin floor by aerially extensive meltwater plumes after ice had receded from the basin and onto the Shield. Resolution of this question may be possible following acquisition of regional seismic data.

Top of Champlain Sea deposits (modern landscape). Where underlain by Champlain Sea deposits, the modern landscape tends to be relatively flat and nearly horizontal, except where streams incise the ground surface (Figs. 1, 3). Two types of incision are observed, one associated with a break of slope, and one that is not.

The only incised valley that lacks an obvious downstream break in slope is the Ottawa River incised valley, within which the modern Ottawa River flows (Fig. 2). (Note that the St. Lawrence

[14] There was no “transgression” in the standard (non-glacial) sense of the word. Rather, water depth was maximum immediately following ice retreat, then decreased with isostatic rebound. Technically, therefore, Champlain Sea sediments are all forced regressive (falling stage) deposits.
[15] Dates from marine shells.

River, by contrast, is not incised upstream of Montreal.) The Ottawa River incised valley is 15–30 metres deep, 5–20 kilometres wide, and is anastomosed just downstream of a bedrock constriction at Ottawa (The modern Ottawa River is an order-of-magnitude smaller.). The downflow extent of the incised valley past Montreal is unknown, but it can be traced 300 km upstream of Montreal to Fort Coulonge, and maintains a relatively constant cross-sectional area along this distance. (A general lack of surficial sediment over bedrock hinders its identification upstream of this.) It is carved primarily into Champlain Sea mud. Paleoislands in the anastomosed section are also composed primarily of mud, with bedrock protruding through locally. Deep, elongate, flow-parallel scours (up to 20 m relief) eroded into Champlain Sea mud are present on the floor of the modern river in a lacustrine-like reach just upstream of Ottawa (Shilts, 1994), and streamlined, flow-parallel mud ridges are present on the incised valley floor south of the first mud paleoisland downstream of Ottawa (Fig. 4). Streamlined boulder ridges have been reported on the floor of the incised valley in Gatineau (Hull sector), just north of Ottawa, that erosively overlie mud with marine fossils and contain angular upstream-imbricated limestone slabs up to 3 metres in length (Keele and Johnston, 1913). In general, however, large clasts and/or bedforms are absent on the floor of the incised valley. A basal date from the Mer Bleu bog suggests incision occurred prior to 7650 ^{14}C yr BP (GSC-681) (see Aylsworth et al., 2000 for sample location).

The second type of stream incision occurs where tributaries of the modern Ottawa River (e.g., South Nation and Rideau rivers) cross the edge of the Ottawa River incised valley (Fig. 3). Depth of incision is greatest at the break in slope, and decreases gradually upstream. An incised tributary that crosses the centre of the Route 300 seismic line truncates near-horizontal Champlain Sea mud reflections (see Stop 4); no channel deposits are observed.

Interfluves outside of the incised valleys and large portions of land within the incised valleys have undergone pedogenic alteration to a depth of one to several metres. Modern and historic landslides occur on the flanks of both types of incised valley. For most historic landslides, wood samples collected within or below the landslide material yield dates that cluster tightly around 4550 ^{14}C yr BP (Aylsworth et al., 2000).

The Ottawa River incised valley is huge: it is an order-of-magnitude larger than most modern rivers (only the Amazon comes close), and is similar in width to the Lake Missoula outburst-flood channel (Fig. 12). Given that no obvious break in slope occurs downstream, and that the St. Lawrence River shows no comparable incision, the Ottawa River incised valley did not likely form by knickpoint migration upstream from a slope-break or by entrenchment related to isostatic rebound. Rather, the valley likely formed when water discharge down the Ottawa River was greater than today. The Ottawa River has long been suspected as having acted as a major continent-to-ocean meltwater conduit during the last deglaciation (MacPherson, 1968; Broecker et al., 1989; Teller, 1988; Teller et al., 2004). Teller (1988) estimates early Holocene discharges of up to 200,000 cubic metres per second from Lake Agassiz outburst floods. (Modern discharge is ~ 2,000 cubic metres per second.) However, if the simple scaling relationship outlined in Figure 12 is assumed to apply, the channel-forming discharge may have been upwards of 800,000 cubic metres per second, almost four times that predicted by Teller (1988). Similar or larger values are obtained using the slope–area method of Dalrymple and Benson (1967) (Dmitri Ponomarenko, personal communication, 2007). In any case, the incised valley is very large, and discharge must have been accordingly very high. Incised tributary streams likely formed differently, by knickpoint nucleation at the edge of the Ottawa River incised valley (e.g., Leeder and Stewart, 1996)[16]. Radiocarbon dates indicate that most landslides postdated erosion of the Ottawa River incised valley by several thousand years. Aylsworth et al (2000) argue that the cluster of landslide dates around 4550 ^{14}C yr BP indicates an earthquake trigger.

[16] Similar yet less pronounced knickpoint incision is visible where tributaries cross a break in slope and enter the modern Ottawa River.

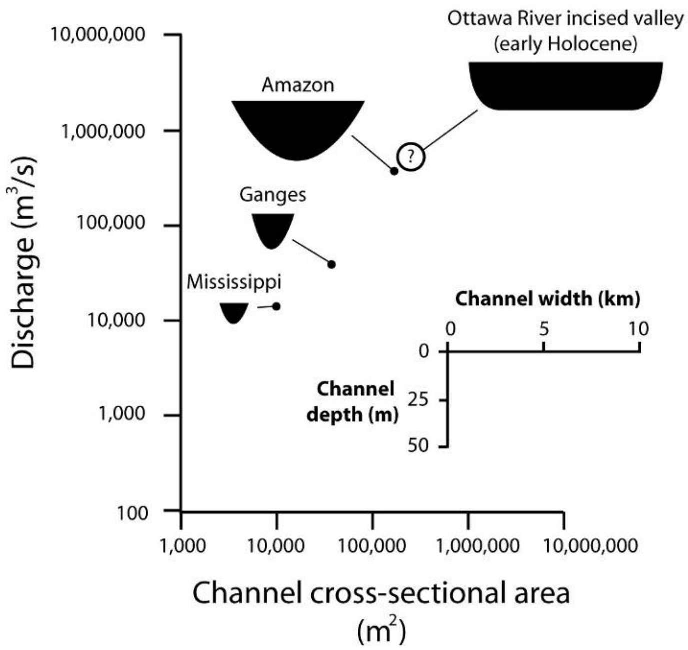

Figure 12. Simplified cross-sections of large modern rivers versus that of the Ottawa River incised valley.

Stop 1B. Seismic Section French Hill Rd East

André Pugin[17] and Susan Pullan, Geological Survey of Canada

Objective. Stop 1B provides an overview of shear-wave (SH-wave) reflection seismic technique applied in the South Nation study by the GSC. Data collection was initiated to better delineate the location of a buried segment of the Vars–Winchester esker.

Setting Overview. Data collection was completed in an upland region that consists of bedrock outcrop, low relief till mounds, faint esker relief and extensive low relief muddy sediment. Facing the Ottawa River the upland has an escarpment of Paleozoic bedrock. The eastern flank of the upland is bordered by a paleo-channel and the bank consists of sandy surficial sediment. Water wells in the upland penetrate to bedrock, are locally flowing, and have yields of up to 200 gallons per minute. The upland forms a paleo-island in the incised-valley of the Ottawa River.

Description. The ~2750 m long, east–west seismic line images bedrock and a number of seismic facies within the overlying surficial sediment secession. Boreholes drilled on the section (Wells A and B in Fig. 7) provide ground truth for the seismic interpretation. The section consists of three distinct entities: I) a shallow western portion of 1800 m length, ii) a 500 m wide bedrock valley, iii) an eastern portion of shallower bedrock (Fig. 13).

Bedrock signature. The bedrock interface slopes gently eastward from the beginning of the line to 1800 m. At this location the bedrock descends abruptly forming the western margin of an asymmetrical bedrock valley of 500 m width with a deep 200 m wide western trough and a 300 m wide shallower eastern shoulder. The valley has 80 m of relief along the axial trough where the bedrock reflection is weak to absent.

Till on Bedrock. Overlying the limestone bedrock there is a boulder lag and/or a 2–3 m thick till layer. The boulder lag is interpreted on the basis of parabolic refractions. Strong parallel reflections are interpreted to be produced from the top of till and the top of the underlying limestone bedrock.

Valley Fill. Fill of the buried bedrock valley consists of an 80 m thick succession characterized by a diffractive chaotic seismic facies. The shallower part is less chaotic, with discontinuous reflections forming a large trough shaped feature. The chaotic facies may be a signature of disturbed sediment that has been subject to liquefaction triggered by earthquake events. Similar signatures have been described elsewhere in the Ottawa basin. These disturbances occur when earthquake energy is amplified by bedrock depressions filled by thick basin mud sequences (Aylsworth et al. 2000).

Basin Mud. The lower basin mud is characterized by continuous reflections that can be subdivided into two subunits separated by a strong reflection. The two units have different velocity/density characteristics. Based on surface seismic data. the lower "Basin Mud I" unit has an average interval SH-wave velocity of ~ 160 m/s and is more transparent, whereas the upper more reflective unit "Basin Mud II" has an average interval velocity of 125 m/s. More definitive velocity measurements will come from geophysical borehole logging surveys.

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[17] Pugin, A., 2009. Stop 1B. Seismic Section French Hill Rd East; in. Russell, H. A. J. and Cummings, D. I. (compilers), Deglaciation of The Champlain Sea Basin, Eastern Ontario; 72nd Friends of the Pleistocene Field Guide, June 6 – 7, 2009, p. 28–29.

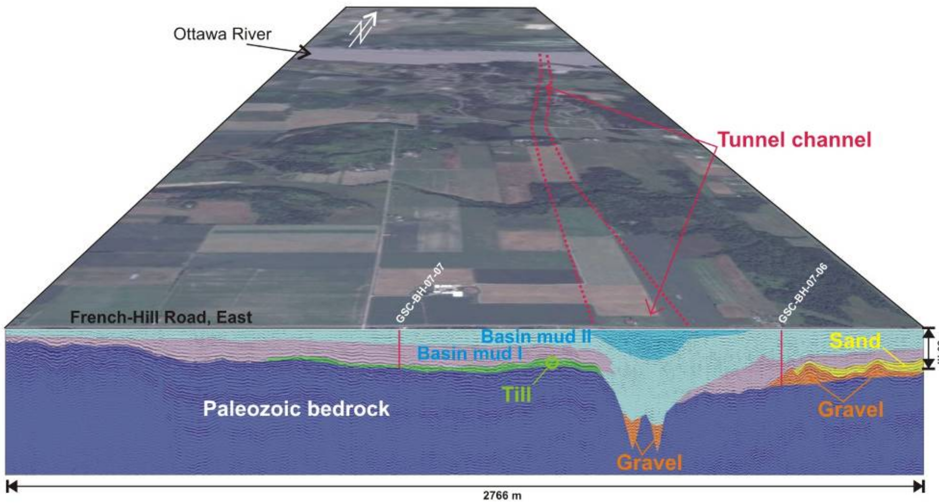

Figure 13. Seismic reflection section from the east-end of French-Hill road. Note the buried bedrock valley and the blanketing marine sediment that forms a low relief landscape.

Discussion. The seismic profile provides information on a number of points on the geological history of the area that are also of hydrogeological significance (Fig. 13). The most unexpected development was the discovery of a > 80 m deep bedrock valley. The 500 m wide valley appears to be intercepted in a number of deep boreholes near the Ottawa River and to the south of the section. From this sparse borehole control, the valley is interpreted to extend north-south with a slight sinuosity in its course. Based on the seismic facies of the valley fill and the interpreted stratigraphic context, the feature is interpreted as tunnel valley that likely formed by the reoccupation of an existing bedrock valley. Gravel mounds on the eastern end of the seismic section are interpreted to be the Vars–Winchester esker. The esker appears to intersect the valley where it curves eastward. Immediately to the south of the valley, the esker is exposed in a number of small aggregate pits, two of which form subsequent stops.

Stop 1C. Hazards of the Leda Clay

Jan M. Aylsworth[18], Geological Survey of Canada

The Ottawa region, like much of the Champlain Sea basin, is susceptible to several hazards associated with the thick deposits of glaciomarine clay and fine silt (informally known as Leda clay), that cover much of the region. These clays are prone to differential settlement under weight loading conditions, are susceptible to landslides in the form of large rapid retrogressive earthflows and lateral spreads, and experience amplified ground motion during seismic events. These hazards present ongoing engineering and planning problems in the Champlain Sea basin.

Leda clay is not formed of clay minerals, but rather it is glacial rock flour. Flocculation of the fine grained particles in the saline-to-brackish water resulted in a structure characterized by a highly porous, loose framework capable of retaining a very high natural moisture content. Lacking the strong interparticulate attractive forces of true clays, the strength of Leda clay relies on salinity of the porewater. In many locations, long term leaching of salt from the pore water of the Leda clay has since reduced its structural strength. Generally, large earthflows occur where the salt content of pore water is less than 2 g/l, an order of magnitude less than in unleached clay (Torrance 1988; Carson 1981). Because of its inherently weak structure and high moisture content, sensitive clay will fail at its natural moisture content. If disturbed, these sensitive sediments can quickly lose strength and collapse, becoming a viscous mud that behaves like a liquid.

Much, although not all, of the Leda clay is geotechnically sensitive. In these cases, the shear strength of remolded clay is dramatically lower than the undisturbed strength. Sensitivity, the ratio of undisturbed to remolded strength at the same moisture content, commonly lies between 20 and 100 and a local high value of 168 was recorded near Casselman. A value exceeding 30 denotes a ‘sensitive clay’; a ‘quick clay’ exceeds 50 (Broms and Stal 1980). Plastic limits range between 20 and 30 and liquid limits are highly variable, resulting in a wide range (4–50) for the plasticity index (Eden and Crawford 1957). The moisture content (25–90% by weight) commonly exceeds the liquid limit of the clay and the liquidity index commonly equals or exceeds 1 (unstable). Sensitivities are directly related to the value of the liquidity index.

Leda Clay Landslides

If disturbed, sensitive clay may liquefy, collapsing into rapid earthflows that can erode retrogressively into flat land above the failed slope and the viscous debris can flow away from the slope for great distances at very low angles. The ‘disturbances’ that can trigger failure include river erosion of the toe of the slope; loading of the top of the slope; porewater pressures; rapid drawdown of the water table associated with snowmelt and the spring runoff; construction activity on or near the slope; and earthquake shaking. The resulting landslide fails as a retrogressive earthflow. Non-sensitive clay slopes may fail as slumps or retrogressive slides and are more contained in area.

A retrogressive earthflow can rapidly destroy extensive areas of flat land and the liquefied debris may flow great distances from the original failure. In many cases, failure begins as a small rotational slide or slump at the bottom of the slope. The initial slump destabilizes sensitive clay behind it, and a major, backward collapse can occur as the sensitive clay liquefies and flows. This process continues in a domino-like fashion, rapidly eating back into the flat land lying behind the failed slope. The sequentially failing scarp may collapse as a series of retrogressive rotational slumps or as a lateral spread and subsidence. At times, large areas may detach and collapse simultaneously, moving in a large flake. It has also been suggested that, instead of a domino-like series of failures, the entire failure plane may actually pre-exist prior to any movement (Quinn et

[18] Aylsworth, J.M., 2009. Stop 1C.Hazards of the Leda Clay; *in*. Russell, H. A. J. and Cummings, D. I. (compilers), Deglaciation Of The Champlain Sea Basin, Eastern Ontario 72nd Friends of the Pleistocene Field Guide, June 6 – 7, 2009, p. 30–34.

al., 2007). Regardless of how the failure occurs, as the mud flows away from the scarp it rafts intact pieces of the overlying, stiffer material. In some landslides, many of these rafted blocks are preserved; in other landslides, most of the sediment ultimately liquefies.

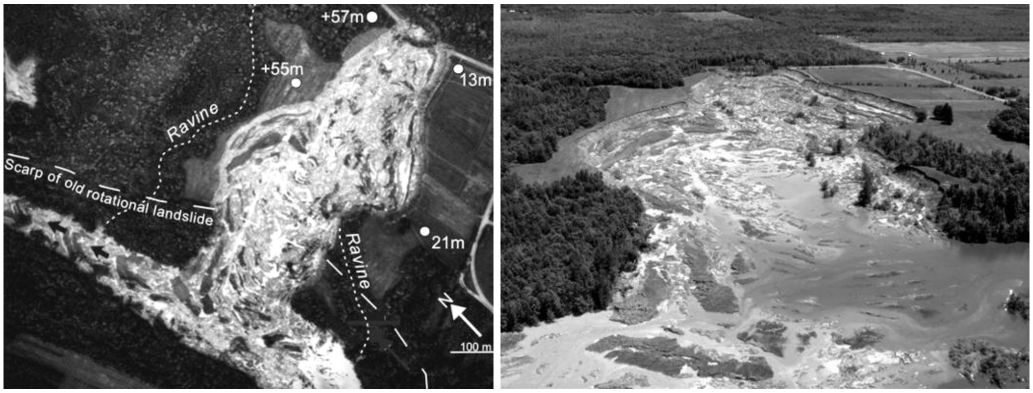

Figure 14. Lemieux Landslide, 1993. Landslide retrogressed 680 m from a 24 m-high bank in under 1 hour; most material was evacuated in about 15 minutes. Oblique photo taken 4 days after event, as floodwaters are about to overtop the landslide dam. (photo: SG Evans)

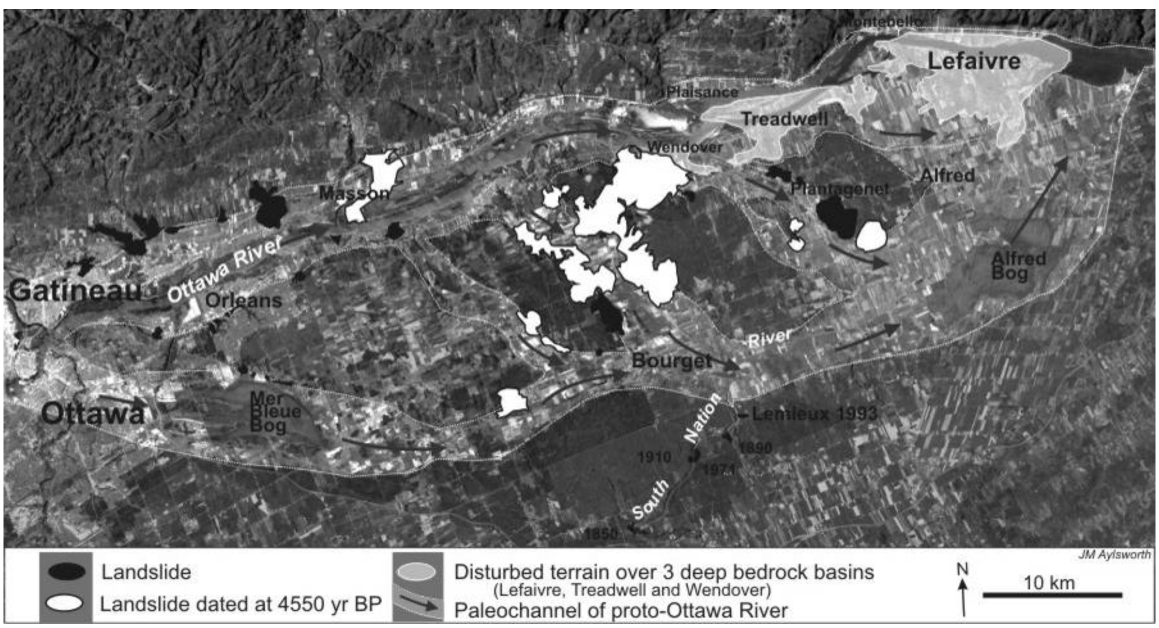

Figure 15. Earthquake induced landslides and severely disturbed terrain in Ottawa River paleochannels.

The significant impact of these earthflows is a function of their size, the speed with which they develop, and the fact that the St. Lawrence Lowlands are heavily populated and developed. Over 250 large landslides or landslide complexes, both ancient and recent, have been mapped within a 60 km radius of Ottawa. Thirty-three deaths occurred in the 1908 landslide at Notre-Dame-de-la-Salette, Quebec. Elsewhere in the region, homes and property have been affected or destroyed and transportation and communications routes have been interrupted. On occasions, landslides have dammed rivers, sometimes for considerable time, creating flooding and siltation problems. Remedial measures have been implemented and homes moved in response to landslide hazard. The most recent large event (Fig. 14) occurred in 1993 at the townsite of Lemieux, where 17 ha of flat land were destroyed as 2.8 million cubic metres of clay, silt and sand flowed into the South Nation River, choking the valley for 3.3 km and causing flooding upstream for 25 km, ultimately costing 12.5 million dollars (Evans and Brooks 1993). The town had been evacuated prior to the event.

Soft Sediment Response to Earthquake Shaking

Although only smaller earthquakes have been recorded historically in the Ottawa segment of the Ottawa–St Lawrence seismic zone, there is ample geological evidence that this region has experienced two of the most geologically-destructive paleo-earthquakes known to have occurred in eastern Canada (Aylsworth *et al.* 2000; Aylsworth and Lawrence 2003). One, at 4550 yr B.P., induced widespread massive landsliding in sensitive marine clays (Bourget area, Fig. 15). The other, at 7060 yr B.P., caused severe, irregular surface subsidence, lateral spreading, and sediment deformation over an area of 46 km^2 in thick deposits of marine clay and sand fill of a deep bedrock basin at Lefaivre (Fig. 15). The terrain remains a fossil landscape of this earthquake disturbance.

On the basis of descriptions in the historical record and ground response modeling, an earthquake of a minimum magnitude of 6.5 on the Richter scale, with an epicentre distance within ~ 50 km is required (Keefer 1984; Benjumea *et al.* 2003). If the epicentre is more removed the magnitude must be greater. John Adams (pers. comm., 1999, 2006) estimates a 6.8 to 7.2 magnitude for these events. These two paleo-earthquakes are evidence that historically “quiet” parts of the Ottawa–St. Lawrence seismic zone, such as the Ottawa region, are still subject to occasional high-magnitude earthquakes and, also, that area underlain by thick deposits of soft sediment experience greatly amplified seismic response.

Bourget area: 4550 yr BP earthquake

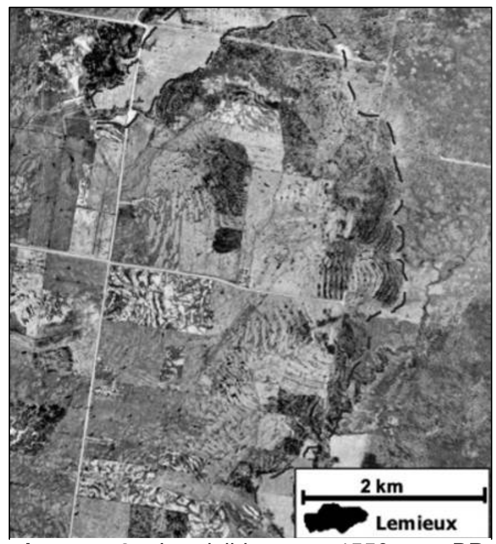

Figure 16. Landslide ca. 4550 yr BP. Lemieux *landslide as scale.*

Several broad cross-cutting paleochannels of the proto-Ottawa River lie east of Ottawa (Fig. 15). These abandoned channels are steep-sided, about 20 to 25 m deep, and, in width, range from 1.5 km to 11 km. The paleochannel walls are scarred by many huge earthflows that are an order of magnitude larger than any historic landslides (Fig. 16). Sizes of individual earthflows range from 1 to 5 km^2 in area (≥ 55 million m^3 volume) and, in places, large, overlapping and coalescing slope failures form huge landslide complexes, up to 32 km^2 in area. The earthflows retrogressed up to 1.5 km into the terrace behind and ran out across the floor of the paleochannels up to 2 km. Radiocarbon dates on buried organics in these landslides cluster at 4550 yr B.P. Most significantly, these dates are several thousands of years younger than the time of paleochannel abandonment established by extensive bogs in the channels (*minimum ages: Mer Bleue Bog, ca. 7650 ± 210 yr B.P.; Alfred Bog, ca 7100 ± 100 yr B.P.*). Evidence that most paleo-landslides occurred simultaneously, ca. 4550 yr BP, long after channel abandonment, has been interpreted as slope response to a massive earthquake (Aylsworth *et al.* 2000; Aylsworth and Lawrence 2003).

Although the large earthflows are widely distributed over an area of 50 km by 25 km, some slopes did not fail. Electromagnetic surveys demonstrated that both paleo- and modern earthflows coincide with zones exhibiting higher resistivity values. In these marine sediments, high electrical resistivity (low conductivity) values are a reliable indication of low porewater salinity, which is associated with higher geotechnical sensitivity in clay, and thus, unstable conditions. In addition, P-wave refraction and reflection seismic data were used to establish the regional thickness of overburden. It was found that a strong peak at 43 m thickness existed along the scarps of the paleo-landslides, in contrast to a regional average of 21 m. This suggests that thickness of overburden may be a critical factor in earthquake-induced landsliding and that not only broad band amplification in thick soft sediment but also critical resonance amplification is important (Aylsworth and Hunter 2004).

Lefaivre area: 7060 yr BP earthquake

Slightly further east, at Lefaivre, a large area of disturbed terrain in an otherwise flat erosional clay plain is attributed to a major earthquake ca. 7060 yr BP. Seismic surveys have mapped the existence of a small but deep (up to 180 m) bedrock basin underlying the disturbed area and have shown that the confining basin walls rise sharply. Surface topography ranges from undulating to hummocky, with local relief of 2 to 10 m and numerous small closed depressions Fig. 17). Surface relief is greatest where soft soils are the thickest. In ditches and borehole logs, sediments have been seen to be severely deformed – ranging from brittle shear to plastic deformation and liquefaction of the Leda clay, liquefaction of sand layers, and intrusion of numerous sand dykes and veins from lower sand units. Sub-bottom profiling confirmed that deformed sediment extends offshore under the bed of the Ottawa River (Douma and Aylsworth 2001).

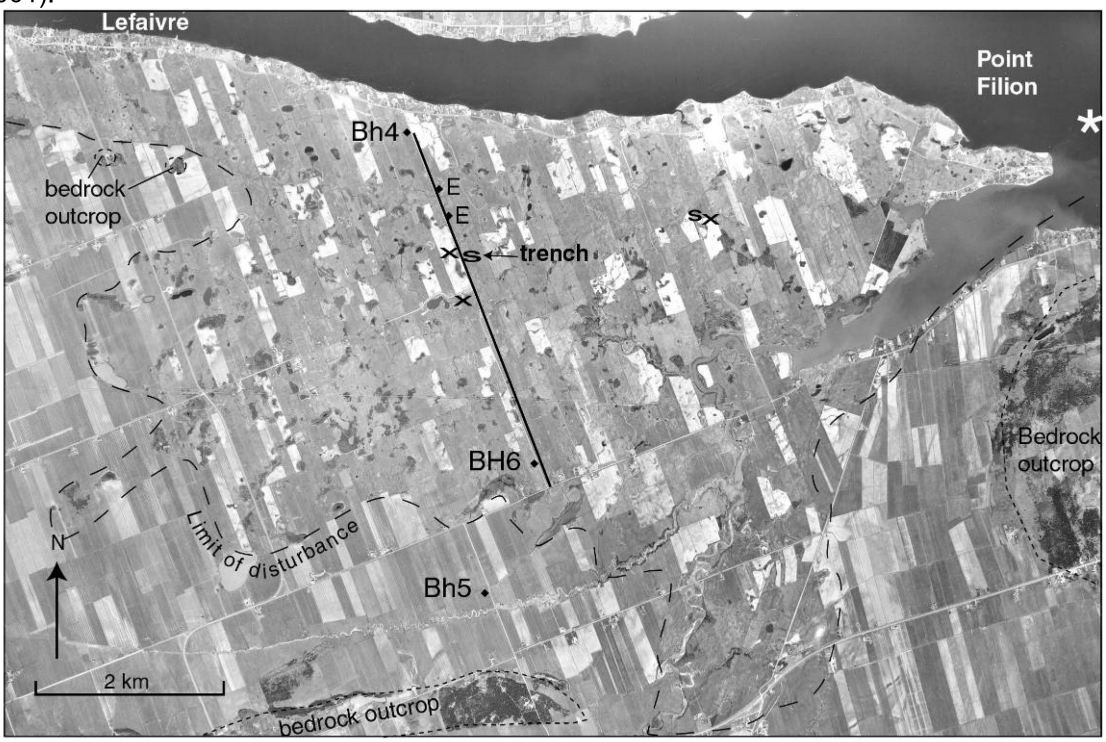

Figure 17. Lefaivre area of disturbed terrain and deformed sediments, interpreted to result from an earthquake that occurred 7060 yr BP.

Continuous core was collected from 3 boreholes (shown on Fig. 4) that were drilled through the Quaternary sediments to bedrock. Testing revealed that the clay is not geotechnically sensitive, probably because of the high pore water salinity. Core from the deepest borehole (BH4, 150 m of overburden) revealed the presence of two thick layers of saturated fine to medium sand at depth within the clay sequence and sediment deformation occurred to an amazing depth of 50 m. It is likely that liquefaction of the sand units played a key role in the overall disturbance. Buried sand layers were minor and deformation was shallower in borehole BH6 (70 m of overburden) near the edge of the disturbed ground, and neither sand nor deformation was present in borehole BH5 (40 m of overburden) outside the disturbed area (Aylsworth and Lawrence 2003).

Both surface surveys and downhole velocity logs confirmed the possibility of earthquake ground motion amplification and resonance effects in the Lefaivre area and also in the other two areas of severe surface disturbance (Wendover and Treadwell, Fig. 2), which also overlie deep bedrock basins infilled with soft sediment (Aylsworth and Hunter 2004; Benjumea *et al.* 2003). In all three of these areas, during earthquake shaking, thick soft sediments (low shear wave velocity) promote broad-band amplification of ground motion due to large near-surface shear wave velocity gradients within the soil column. Although attenuation of high frequency spectral components may occur, significant low frequency amplification may result (over a 4 month monitoring program, x 6 amplification has been recorded for small quakes, M. Lamontagne, pers. comm. 2002). In addition, because of the substantial thickness of the sediments (especially in basin areas) and low shear wave velocities, resonance amplification of earthquake ground motion in the 0.3 to 1 Hz range could also occur (> 8 times amplification at the fundamental site period).

Stop 2A. Regimbald Road pit

Don Cummings[19], Geological Survey of Canada

Objective. Observe and interpret the sedimentology of the esker where it is narrow and composed entirely of gravel.

Setting. The pit is excavated into a tributary-like arm of the esker (see field trip stop map at start of guide book). At this location, the esker is partially overlain by Champlain Sea mud. In a nearby cored well, the esker rests directly on carbonate mudstone bedrock.

Description. The coarsest (and presumably most proximal) sediment in the esker is exposed in the Regimbald Road pit. In the exposure, the gravel is organized into thick (1.5–2.5 m) high-angle cross-stratified beds that are stacked on top of each other (Fig. 18). Cross-strata dip at a high-angle (25–30°) towards the south, which is parallel to the esker long-axis. Clasts are well rounded and are typically pebble to cobble size, although boulders up to ~ 1 metre in diameter are present. Clast lithology is similar to that of the regional till: carbonate mudstone predominates, with minor percentages of igneous (granite), sandstone and shale. Within the cross-sets, cyclicity is observed on both a centimetre-scale (alternation of sandy cross-strata with gravelly cross-strata; Fig. 19) and on a metre-scale (several reactivation-surface bound, downflow-fining packages of cross-strata within a single cross-set).

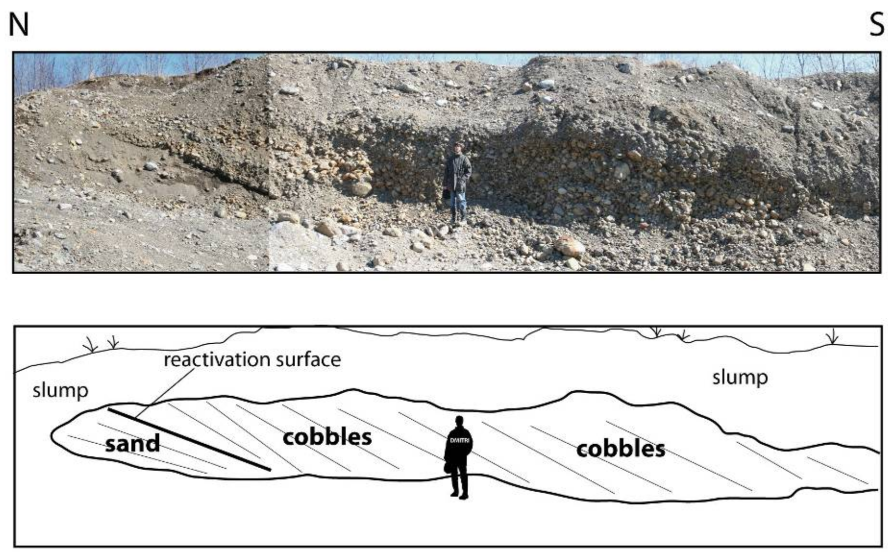

Figure 18. One of several thick, stacked, high-angle cross-sets exposed in the Regimbald Road pit. See field trip stop map at start of guidebook for pit location. Note the reactivation surface separating sandy cross-strata from cobble cross-strata. Cross-strata dip roughly southward, parallel to the long-axis of the esker. Dmitri "Kung-fu" Ponomarenko for scale.

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[19] Cummings, D.I., 2009. Stop 2A. Regimbald Road pit; *in*. Russell, H. A. J. and Cummings, D. I. (compilers), Deglaciation Of The Champlain Sea Basin, Eastern Ontario 72nd Friends of the Pleistocene Field Guide, June 6 – 7, 2009, p. 35–36.

Figure 19. Close-up of centimetre-scale rhythmicity in the high-angle cross-strata.

Discussion

- What was the source of the gravel—how did it “get into” the esker?
- Were the cross-sets deposited by dunes or bars?
- Are these subglacial deposits?
- Why is the sandy fan carapace absent?
- What is the significance of the two scales of rhythmicity?
- How deep was the flow?
- How fast was the flow?
- How long did it take to deposit the esker at this location? Days? Years? 100s of years?
- With respect to the groundwater system, are we justified in upscaling observations from this pit to the whole esker-aquifer?

Stop 2B. Hydrogeology

Marc Hinton, Sam Alpay, and Charles Logan[20]; Geological Survey of Canada

An important control on the hydrology and hydrogeology of the Champlain Sea basin is the regionally extensive low permeability mud, which results in low groundwater recharge and high surface runoff. Water resources in this intensely farmed region are also indirectly affected by extensive tile drainage, fertilizer application and manure management. Where the mud is extensive, the bedrock contact zone and esker aquifers are particularly important water resources.

Water balance and water supply

Watersheds in the Champlain Sea basin have a large annual moisture surplus; however there is a moisture deficit from May to August. For example, one estimate of the water surplus in the South Nation River watershed is 554 mm/yr, the difference between the average precipitation of 971 mm/yr and the estimated actual evapotranspiration of 417 mm/yr (South Nation Conservation and Raisin River Conservation Authority, 2007). The extensive cover of mud in low lying areas, coupled with till and bedrock in uplands, rapidly convert much of this surplus into stream runoff. Rapid stream runoff is especially significant during the spring freshet (March and April), when roughly 60% of the annual surface water flow occurs (CH2MHill, 2001). In contrast, streamflow conditions are very low from summer to early autumn. Summer stream discharge averages on the order of 0.2–0.3 mm/day and decreases to less than 0.02 mm/day during dry conditions (7Q20, seven-day low flow with a recurrence interval of 20 years). These low summer flows can represent a significant constraint to the development of surface water supplies, particularly when considering in-stream needs (e.g. ecological, water quality and waste dilution) and additional summer demand. The low summer flows are also indicative of low groundwater recharge and discharge and a limited groundwater resource. Given the limited water resources, protection of existing water supplies is important, particularly since extensive and intensive agriculture, increasing rural population and commercial activities all contribute to increasing water demand and to water contamination. The extensive Champlain Sea mud is an effective barrier to groundwater recharge and flow. Therefore, it is a contributing factor both to impairment of surface water quality from surface runoff and to groundwater protection in underlying confined aquifers.

Hydrostratigraphic units

The western Champlain Sea Basin has up to ten major hydrostratigraphic units of which five are bedrock, four are Quaternary sediment and one is the contact zone aquifer (Table 1). Extensive faulting and displacement in the Paleozoic bedrock results in significant variation in the type, distribution, thickness and hydraulic properties of the bedrock formations. In general, water quantity and quality in the bedrock are best in early Paleozoic formations (Nepean, March, Oxford) and decrease upward and eastward into the younger, more shale-rich formations. The uppermost fractured or weathered portion is frequently the most permeable and accessible bedrock zone where the majority of domestic wells are completed. Sand and gravel deposits can directly overlie the fractured bedrock and contribute to its productivity. Collectively these sand, gravel and fractured bedrock units are referred to as the contact zone aquifer.

The diamicton (till) that overlies bedrock throughout much of the basin is discontinuous locally, especially beneath the eskers where it was eroded by subglacial meltwater and in bedrock uplands where it was eroded by wave action during the Champlain Sea regression. It has lower permeability than the contact zone aquifer but may be sufficiently permeable, particularly where it is exposed and weathered, to allow for some groundwater recharge.

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[20] Hinton, M., Alpay, S. and Logan, C., 2009. Stop 2B. Hydrogeology; *in*. Russell, H. A. J. and Cummings, D. I. (compilers), Deglaciation Of The Champlain Sea Basin, Eastern Ontario 72nd Friends of the Pleistocene Field Guide, June 6 – 7, 2009, p. 37–41.

The eskers are composed of strata with a wide range of grain sizes and hydraulic properties. The narrow gravel esker core, interpreted to be a subglacial stream deposit, is the most permeable and is the main target for high capacity wells. The esker core is draped by a broad sandy cover, interpreted to be subaqueous outwash, which can extend outward hundreds of meters beyond the esker landform beneath the Champlain Sea mud. This sandy exterior can provide the esker core with a much broader area from which to draw water either as a natural drain or when it is pumped. The eskers are locally exposed at the surface where they are important areas for recharge. Where buried beneath Champlain Sea mud, eskers are confined aquifers and conduits for groundwater flow.

As previously mentioned, the Champlain Sea mud has very low permeability and is a key aquitard. Groundwater velocities are so low in unweathered mud that diffusion dominates over advection with respect to solute transport as supported by several boreholes that still have preserved remnant salinity from the Champlain Sea (Torrance, 1988; Desaulniers and Cherry, 1989). Where it is sufficiently thick or unweathered, this unit effectively prevents groundwater recharge and discharge.

Surficial sand and silt units include: fluvial deposits; near-shore, deltaic and estuarine deposits of the Champlain Sea; and glaciofluvial deposits that have been reworked by waves during regression of the Champlain Sea. They are of variable thickness and are underlain by the Champlain Sea mud such that flow is likely unconfined and occurs within local flow systems.

Table 1. Major hydrostratigraphic units in the Champlain Sea basin, Eastern Ontario.

Hydrostratigraphic unit	Description	Hydrogeologic function
Surficial sand and silt	fluvial, marine regression and reworked older deposits	unconfined surficial aquifer
Champlain Sea mud	marine clay and clayey silt	regional aquitard
Eskers	gravelly core (subglacial stream deposit) with a broad sandy cover (subaqueous outwash)	local highly productive aquifer, surficial or confined
Regional till	diamicton of variable composition and compaction	regional leaky aquitard/ domestic aquifer
Contact zone	fractured bedrock, basal sand and gravel	regional aquifer
Billings-Carlsbad-Queenston	shale	aquitard/ domestic aquifer
Ottawa	limestone and shale	aquifer/aquitard
Rockcliffe	sandstone and shale	aquitard/ domestic aquifer
Nepean-March-Oxford	sandstone and dolostone	productive aquifer
Precambrian	igneous and metamorphic	aquitard

Regional groundwater flow system

Three distinct areas are significant for the regional groundwater flow system of the Western Champlain Sea basin: 1) areas of exposed till and bedrock (includes areas with overlying sand); 2) areas of exposed eskers; and 3) areas of Champlain Sea mud (either exposed or underlying surficial sand and silt; Figure 1). Exposed till and bedrock areas occur mostly in uplands and in the southern portions of the region. Areas with Champlain Sea mud are most extensive in lowlands, but also occur in some upland areas in the northern portion of the region. Exposed eskers occur in both upland and lowland areas.

Areas 1 and 2 are the dominant areas of groundwater recharge to flow systems that extend beneath the mud. Recharge rates in exposed bedrock and till (Area 1) are likely variable and may be limited by the permeability of the till or shallow bedrock and the topographic slope. The large areal extent of exposed till and bedrock means that these areas are the most important recharge areas. Recharge rates in exposed eskers (Area 2) are very high since they are likely not limited by geology but rather are controlled by climate and vegetation (Fig. 20). However, the area of exposed esker is a small proportion of basin area. Areas underlain by the Champlain Sea mud have very low recharge or recharge that only contributes to shallow unconfined flow systems.

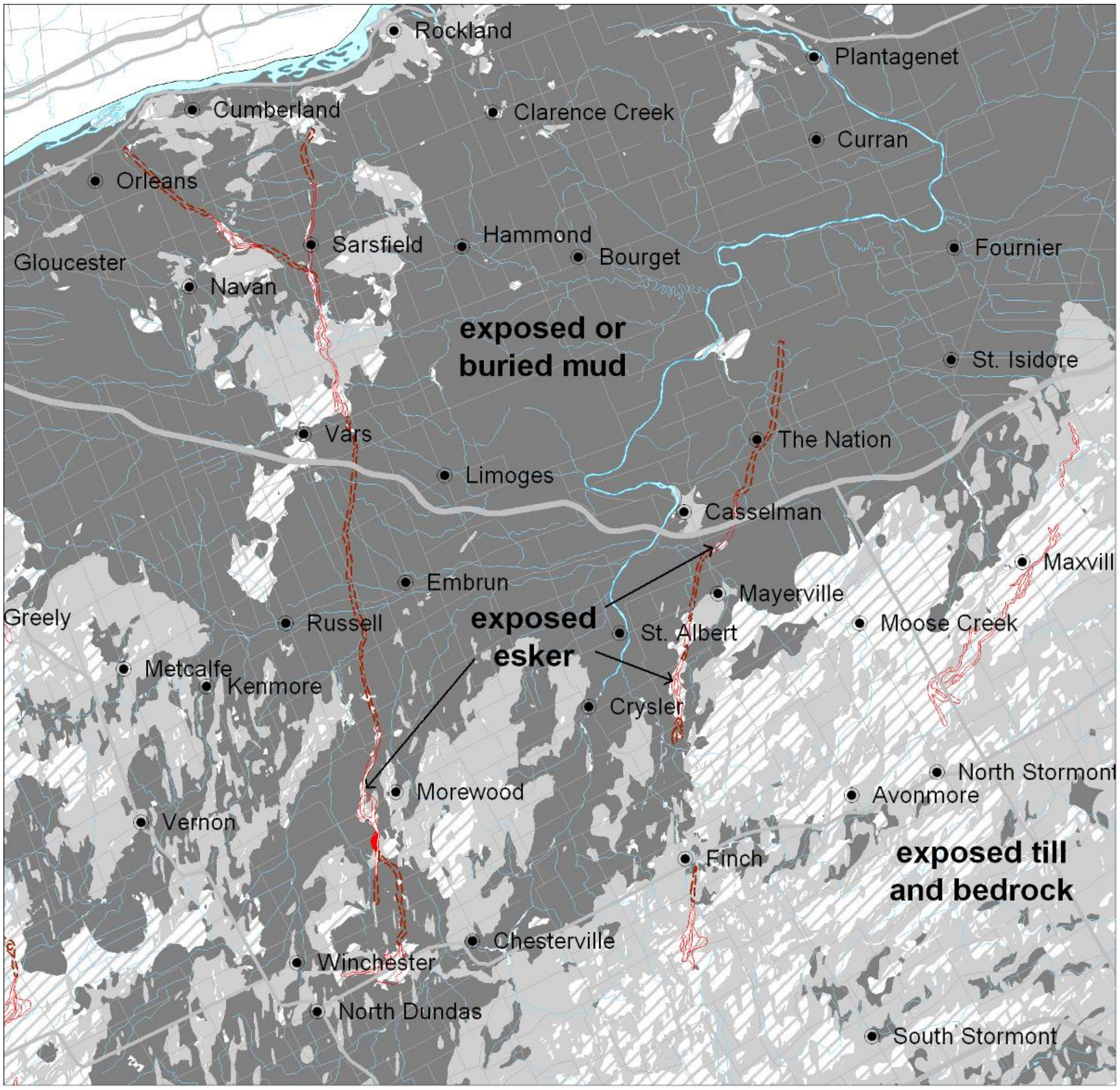

Figure 20. Three main hydrogeological settings in the western Champlain Sea basin. The Champlain Sea mud is represented in dark grey. Light grey areas are till and bedrock; cross-hatched light grey areas are till or bedrock overlain by sand. Dashed lines indicate approximate extent of buried eskers; solid lines indicate eskers exposed at the ground surface

Because eskers, till and bedrock are stratigraphically lower than the Champlain Sea mud, they can only discharge where the mud is not present. This can occur either above the contact with the mud (Fig. 21, arrow A) or where the aquifers are in direct contact with surface water (Fig. 21, arrow B). Upland till and bedrock often produce seepage areas at the base of hillslopes since the hydraulic gradients in the lowlands decrease, thereby lowering their capacity to transmit groundwater flow. Within the lowlands, discharge occurs mostly where there are "hydraulic windows" in the mud. These may occur along exposed esker, drumlin, and bedrock ridges, and along river beds where mud has been eroded. However, there appear to be few hydraulic windows (as in Figure 21, arrow B) along the South Nation River.

Regional groundwater flow is generally perpendicular to topographic contours in the till and bedrock uplands. In the lowlands, groundwater flow directions are less predictable because topographic and hydraulic gradients are smaller and possible locations of groundwater discharge are limited.

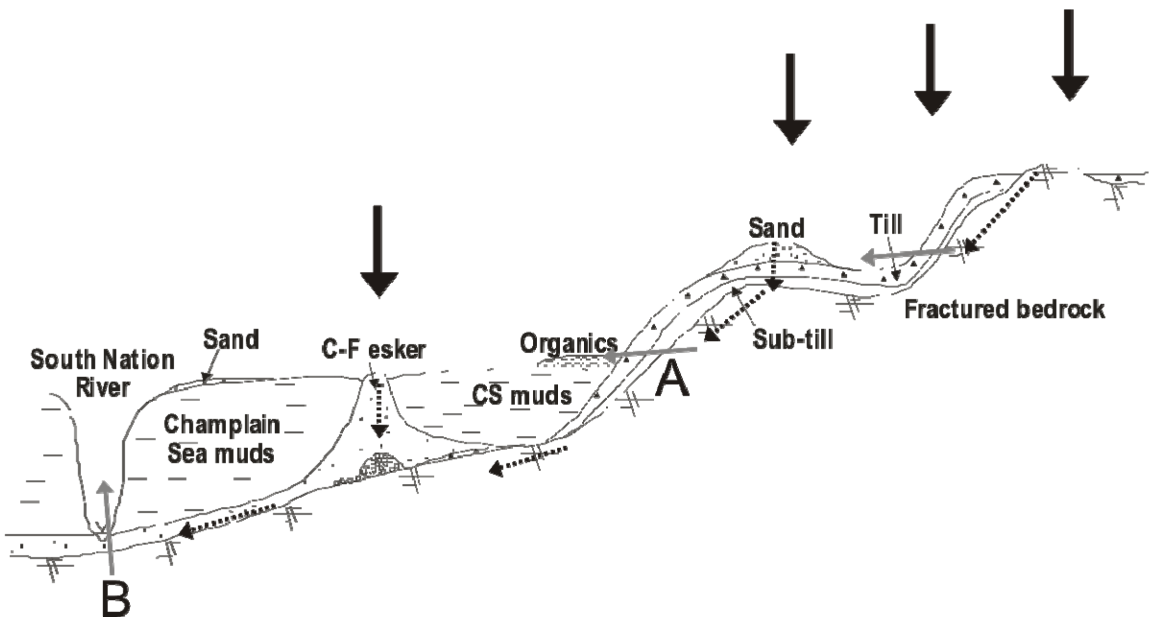

Figure 21. Conceptual cross-section across the Crysler–Finch esker. Black arrows indicate recharge areas to the regional flow system, dashed arrows indicate groundwater flow directions and gray arrows indicate discharge areas. Discharge areas A and B are identified in the text. Groundwater flow may also occur perpendicular to the section along the esker and may discharge where the esker intersects surface waters. The cross-section is approximately 20 km long.

Groundwater flow along eskers

Groundwater flow along an esker will vary according to several factors which include: the area of exposed esker; the hydraulic gradient along the esker; the continuity and transmissivity of the esker core; and the degree of hydraulic connection to the surface in the discharge area. Eskers transect both upland and lowland areas and cross several sub-watersheds. Consequently, groundwater flow along eskers is segmented into multiple, small local flow systems. Mapping groundwater flow along the eskers requires local and regional investigations. Surface water–groundwater investigations can provide insight on recharge into and discharge out of eskers. For example, temperature and electrical conductivity anomalies measured at the sediment-water interface along a reach of the East Castor River successfully identified a groundwater discharge zone where the Vars–Winchester esker intersects the river (Bustros-Lussier et al., 2007).

The Crysler–Finch esker is oriented approximately perpendicular to regional groundwater elevations (equipotentials); the role of this esker in local and regional groundwater flow is still poorly understood but investigations are beginning to assess its significance. Elevations of surface water and select monitoring wells help provide a preliminary understanding of the segmentation of flow along the esker (Fig. 22). Geological, geophysical and hydrogeological data suggest that the esker transmissivity is likely variable from sediment facies changes along the esker. In areas of lower esker transmissivity, the transmissivity of the fractured bedrock may be higher than in the esker; consequently, flow in the contact zone aquifer may be greater than flow within the esker. Another important consideration in the Crysler–Finch esker is that the connectivity of the esker system to surface waters may be limited in some areas and groundwater flow directions may be controlled by the locations of groundwater discharge.

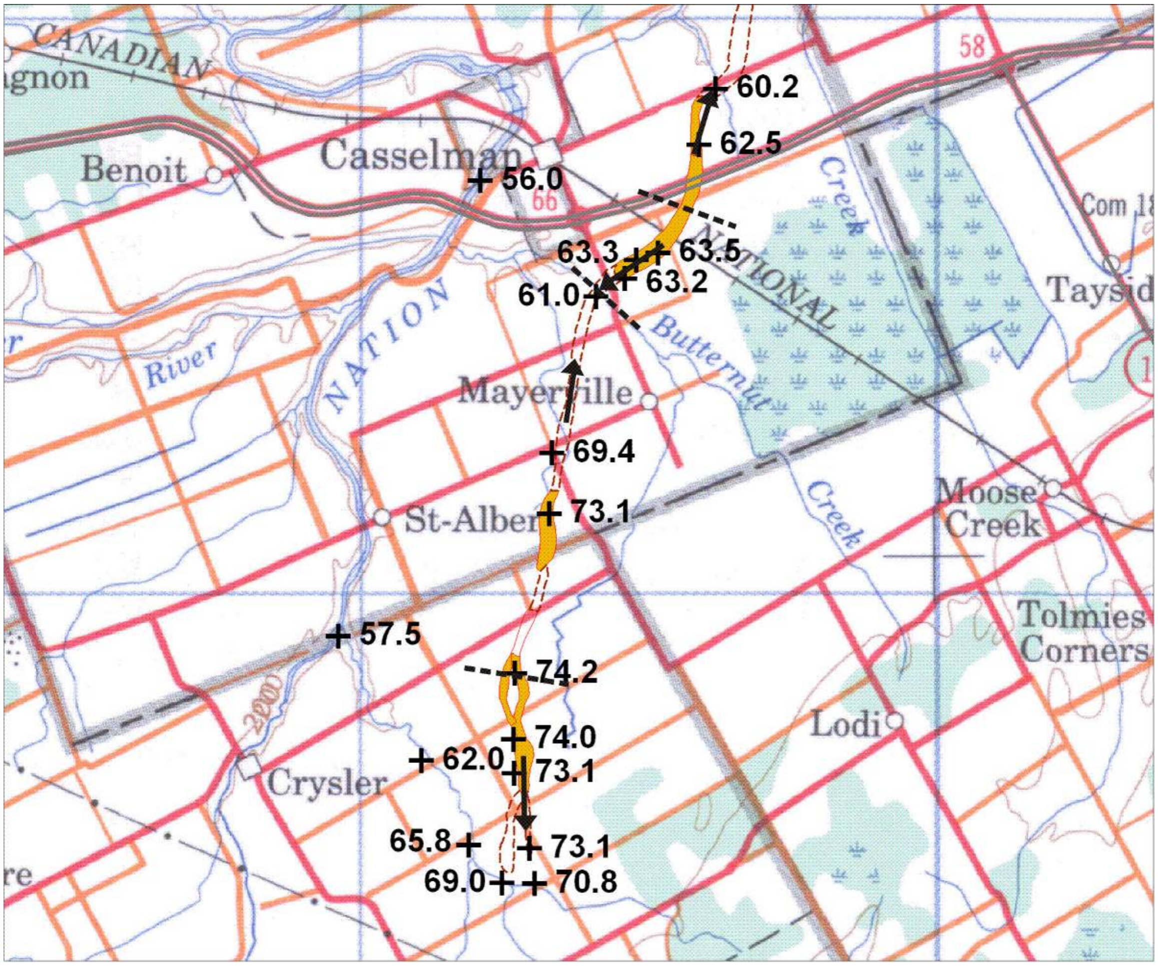

Figure 22. Example of segmented flow along the esker. The figure shows the surface water (gravel pits, streams and rivers) and groundwater elevations (in meters) along a portion of the Crysler–Finch esker. Dashed lines indicate the approximate groundwater flow divides along the esker; arrows indicate the direction of the component of groundwater flow along the esker. The actual direction of groundwater flow may not be oriented along the esker. Elevations were obtained from a 10 m grid DEM (Ontario Ministry of Natural Resources, 2006).

Stop 3A. Watson Road pit

Don Cummings[21], Geological Survey of Canada

Objective. Observe and interpret the sedimentology and stratigraphy of the esker where it is wider, composed of a gravelly central ridge with a broad sandy fan carapace, and partially buried by fossiliferous mud.

Setting. The Watson Road pit is located immediately downflow of an apparent confluence between two tributary-like esker ridges (see field stop map at start of guidebook). The esker doubles in width at this location. It is flanked by Champlain Sea mud and crops out along an axial zone of 100–200 m width.

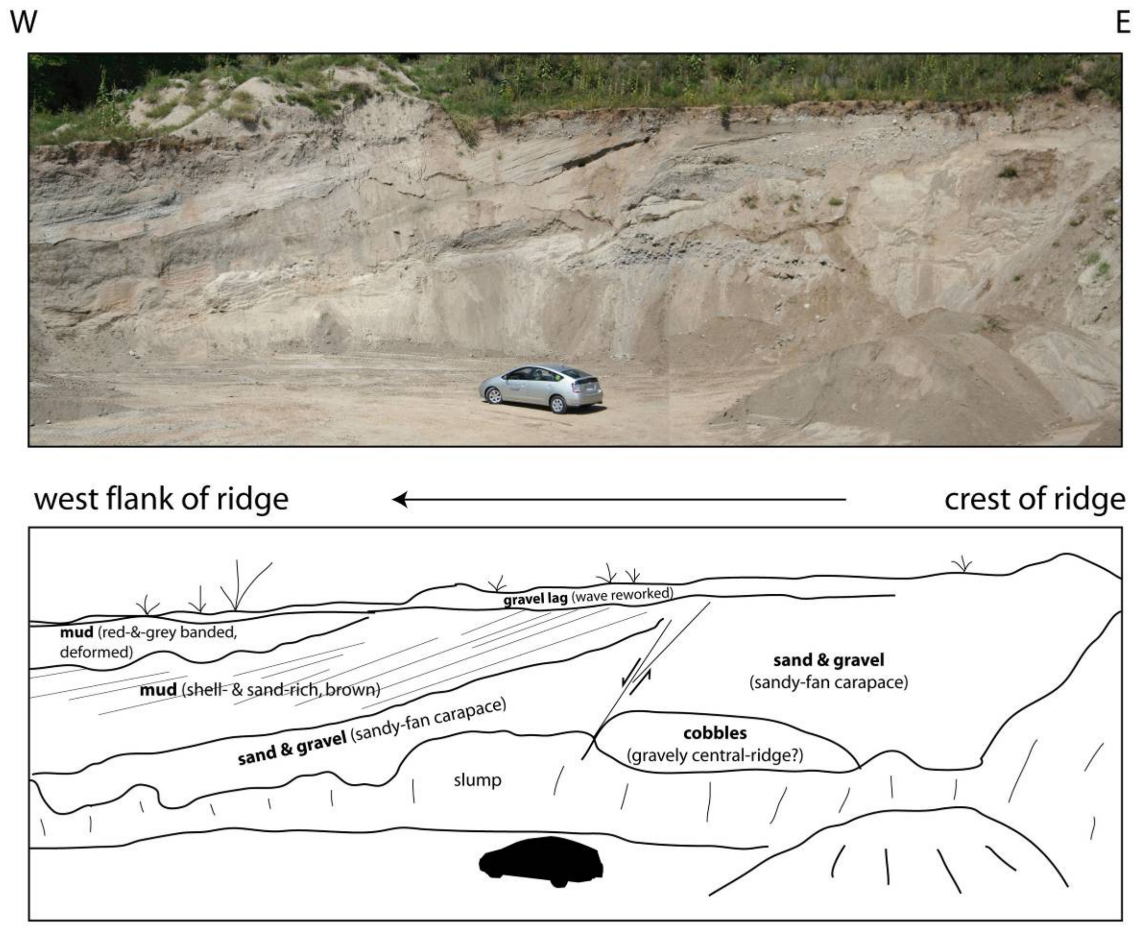

Figure 23. West flank of the esker ridge exposed in cross-section at the Watson Road pit. Note that succession fines up from cobble gravel (gravelly central ridge?) to sand and gravel (subaqueous-outwash fan) to Champlain Sea mud.

[21] Cummings, D.I., 2009. Stop 3A. Watson Road pit; *in*. Russell, H. A. J. and Cummings, D. I. (compilers), Deglaciation Of The Champlain Sea Basin, Eastern Ontario 72nd Friends of the Pleistocene Field Guide, June 6 – 7, 2009, p. 42–44.

Description. The Watson Road pit is the only outcrop where all components of the esker system are exposed—the gravelly central ridge, the sandy fan carapace *and* the mud that overlies the esker (Fig. 23).

At the base of the western side of the pit, a poorly-exposed, mound-like cobble gravel unit is exposed. Given its coarse grain-size and mound-like morphology, it is interpreted to be the gravelly central ridge, or possibly the most proximal portion of the subaqueous-outwash fan.
The sandy fan carapace sharply overlies the gravelly central ridge. In the eastern half of the pit, the fan is well exposed, and is composed of mound-shaped, coarsening-upward sand and gravel units that are stacked compensationally on each other (Fig. 24). Sharp-based beds, commonly gravelly, interrupt this motif locally. Climbing current-ripples and diffusely-laminated sand beds are common, suggesting deposition from tractive unidirectional flows with abundant sediment rain-out from suspension.

Champlain Sea mud overlies the sandy fan carapace (Fig. 23). Two mud units are recognized. The lowermost unit is brownish and contains numerous sand beds, some rich in marine shells (mostly *Hiatella arctica*). This unit downlaps the sandy fan carapace at a very low angle; its clinoforms prograde outward at a normal/oblique angle from the crest of the landform. The second, overlying mud unit consists of red-and-grey laminated mud. Where exposed, its lower contact is sharp and truncates the underlying mud unit at a low angle. Strata are highly folded and contorted close to the crest of the landform, but undeformed off the crest of the landform. Shells are less common in this unit. A thin layer of gravel caps the succession.

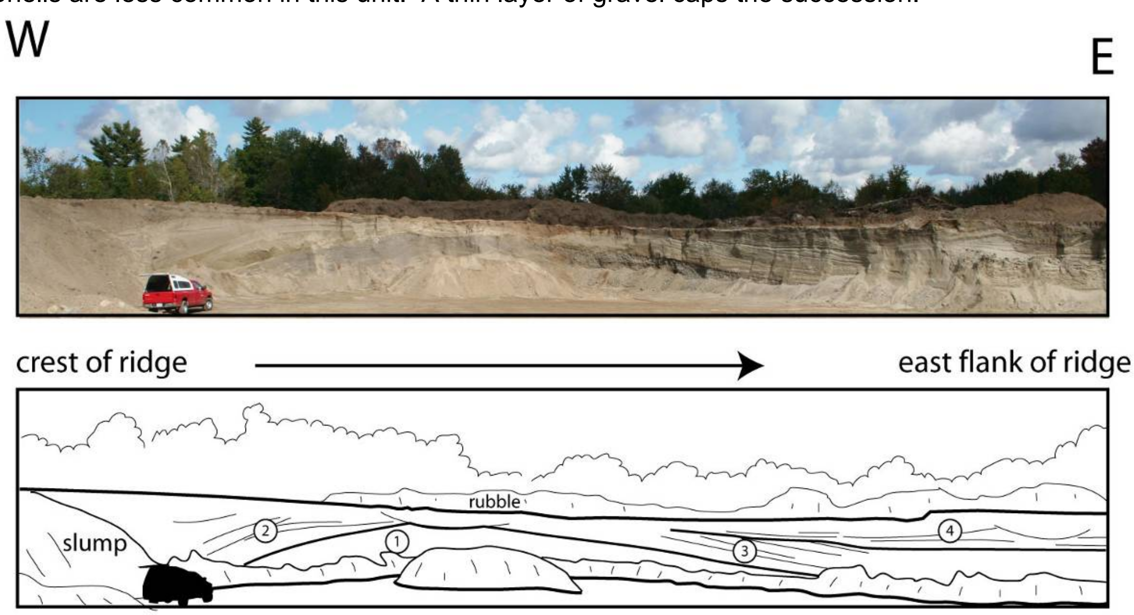

Figure 24. East flank of the esker ridge exposed in cross-section at the Watson Road pit. Paleoflows are out of the page (southward). Note that the sandy fan carapace here is composed of several sharp-based, mound-shaped, coarsening-upward units (numbered 1 to 4). After several months of pit excavation in an upflow direction (northward), mound 1 became noticeably narrower and coarser over several tens of meters.

Discussion

- Why is sandy fan carapace present here, but not at previous stop?
- Note that sandy fan carapace is composed of smaller, coarsening-upward, mound-shaped architectural elements. What are they? How did they form?
- How was downlap relationship between Champlain Sea mud and esker generated?
- Note that two mud units exist, a lower sandy/shelly one and an upper, locally deformed, shell-less, red-and-grey banded one. Why two units? Why are they so different? Why is their contact sharp?

- What succession of events deposited the observed stratigraphy, taking into account the nature of the contacts between and within them?
- Does water percolate through the mud into the esker aquifer? Does this mud layer prevent recharge? If so, how does water get into esker?

Stop 3B. Seismic microzonation hazard mapping in the Ottawa area

Greg Brooks, James Hunter, Heather Crow[22]; Geological Survey of Canada

Dariush Motazedian; Carleton University

The nature of earthquake seismic waves propagating through the earth is dependent on the source mechanism, the depth of the source, and the character of the rock types along the travel path to a particular surface site. The character of the shaking at a specific location on the ground surface (amplitude, frequency and duration), however, is additionally affected by the geologic materials over the last few hundred meters (or less) of the propagation path. This phenomenon arises because seismic energy is modified by differing amounts by areas of dissimilar sub-surface geology and soil conditions as it propagates towards the ground surface. The energy can be either amplified or attenuated by a number of factors that may act in combination depending on the local setting, including: shear-wave velocity contrasts between rock and overlying soil layers (impedance contrast amplification); the repeated reflection of seismic energy within a soil layer between the ground surface and the boundary with an underlying stiffer strata (resonance amplification); focusing-defocusing effects caused by the subsurface geometry (basin curvature effects); and the generation of Rayleigh and Love waves across the surface of a sediment-filled basin (basin edge effects). The net effect is that the ground motion often is greater on areas of softer soil than on stiffer soil or bedrock with the consequent result that damage tends to be greater in areas of soft soil (Building Seismic Safety Council, 1995). The amplification of earthquake shaking has been a major factor influencing damage distribution and severity during many earthquake disasters, for example, the 1985 Mexico City, 1989 Loma Prieta, 1994 Northridge and 1995 Kobe events (see Seed et al., 1988; Borcherdt and Glassmoyer, 1992; Holzer, 1994; Olshanky, 1997; Ishikawa et al., 2000).

The influence of ground motion amplification for building design is recognized in the 2005 National Building Code of Canada (2005 NBCC) (see Finn and Wightman, 2003). This document introduced a seismic site classification system that characterizes the underlying geologic materials at a given location for the purpose of defining amplification factors that modify the design ground motion for building design. As shown in Table 2, five of the six site categories (or classes) correspond approximately to hard rock (A), rock (B), soft rock or very dense soil (C), stiff soil (D), and soft soil (E). The classes are defined based on the average stiffness in the upper 30 m of the ground using either shear-wave velocity (Vs), standard penetration resistance (blow count), or undrained shear strength. The latter two properties, however, can only be used to define classes C, D and E, since these are soil tests that are unsuitable for bedrock materials. The sixth class (F) is a special case and is defined based on more site specific characteristics, as listed in Table 2. This classification scheme is adopted directly from the system developed by NEHRP (National Earthquake Hazard Reduction Program) for the United States (BSSC, 1994).

Amplification of earthquake energy is variable depending on shaking period (s) and the magnitude of the ground motion (g) (the latter is referred to as spectral accelerations (S_a) when pertaining to buildings, as is the case in the 2005 NBCC). In the 2005 NBCC, short-period (F_a) and long-period (F_v) amplification factors are summarized in look-up tables for periods of 0.2 and 1.0s, respectively. In these tables (see Tables 3 and 4), amplification is expressed relative to class C, whereby amplification for classes A and B is 'deamplified' relative to C, and that for classes D and E is 'amplified'. As is apparent in both tables, the degree of amplification is non-linear both between classes for a given spectral acceleration and within a class for

[22] Brooks, G., Hunter, J. M., Crow, H., and Motazedian, D., 2009. Stop 3B:.Hazards fo the Leda Clay; *in*. Russell, H. A. J. and Cummings, D. I. (compilers), Deglaciation Of The Champlain Sea Basin, Eastern Ontario 72nd Friends of the Pleistocene Field Guide, June 6 – 7, 2009, p. 45–50.

increasing/decreasing spectral accelerations. For building design, spectral accelerations for the design ground motion are defined in the 2005 NBCC in terms of class C and the amplification factors in tables 2 and 3 are used to modify these accelerations based on the specific site class of a construction site (as defined based on the stiffness of the upper 30 m of the ground; see Table 1). The variations in spectral acceleration of the Ottawa design ground motion for classes A to E are depicted in Fig. 25, for shaking periods of 0.2, 0.5, 1.0 and 2.0 s, based on the amplification factors in Tables 3 and 4. This figure shows that up to four times more 'shaking' is expected for the spectral accelerations between zones A and E reflecting differences in amplification between hard rock (A) and soft soil (E) sites

Table 2. Seismic site categories as defined in the 2005 National Building Code of Canada (NRC, 2005)

Site Class	Ground Profile Name	Average Properties in Top 30 m, as per Appendix A		
		Average Shear Wave Velocity, $\bar{V}_s$ (m/s)	Average Standard Penetration Resistance, $\bar{N}_{60}$	*Soil* Undrained Shear Strength, s_u
A	Hard *rock*	$\bar{V}_s > 1500$	n/a	n/a
B	*Rock*	$760 < \bar{V}_s \leq 1500$	n/a	n/a
C	Very dense *soil* and soft *rock*	$360 < \bar{V}_s < 760$	$\bar{N}_{60} > 50$	$s_u > 100$ kPa
D	Stiff *soil*	$180 < \bar{V}_s < 360$	$15 \leq \bar{N}_{60} \leq 50$	50 kPa $< s_u \leq 100$ kPa
E	Soft *soil*	$\bar{V}_s < 180$	$\bar{N}_{60} < 15$	$s_u < 50$ kPa
		Any profile with more than 3 m of *soil* with the following characteristics: • plasticity index: PI > 20 • moisture content: $w \geq 40\%$, and • undrained shear strength: $s_u < 25$ kPa		
F	Other *soils*(1)	Site-specific evaluation required		

(1) Other *soils* include:
(a) liquefiable *soils*, quick and highly sensitive clays, collapsible weakly cemented *soils*, and other *soils* susceptible to failure or collapse under seismic loading,
(b) peat and/or highly organic clays greater than 3 m in thickness,
(c) highly plastic clays (PI > 75) more than 8 m thick, and
(d) soft to medium stiff clays more than 30 m thick.

Table 3. Short period amplification factors (F_a) as a function of site class and spectral accelerations at 0.2 s period (NRC, 2005)

Site Class	Values of F_a				
	$S_a(0.2) \leq 0.25$	$S_a(0.2) = 0.50$	$S_a(0.2) = 0.75$	$S_a(0.2) = 1.00$	$S_a(0.2) \geq 1.25$
A	0.7	0.7	0.8	0.8	0.8
B	0.8	0.8	0.9	1.0	1.0
C	1.0	1.0	1.0	1.0	1.0
D	1.3	1.2	1.1	1.1	1.0
E	2.1	1.4	1.1	0.9	0.9
F	(1)	(1)	(1)	(1)	(1)

(1) see class F in Table 1.

Table 4. Long-period amplification factors (F_v) as a function of site class and spectral acceleration at 1.0 s period (NRC, 2005)

Site Class	Values of F_v				
	$S_a(1.0) \leq 0.1$	$S_a(1.0) = 0.2$	$S_a(1.0) = 0.3$	$S_a(1.0) = 0.4$	$S_a(1.0) \geq 0.5$
A	0.5	0.5	0.5	0.6	0.6
B	0.6	0.7	0.7	0.8	0.8
C	1.0	1.0	1.0	1.0	1.0
D	1.4	1.3	1.2	1.1	1.1
E	2.1	2.0	1.9	1.7	1.7
F	(1)	(1)	(1)	(1)	(1)

(1) see class F in Table 1.

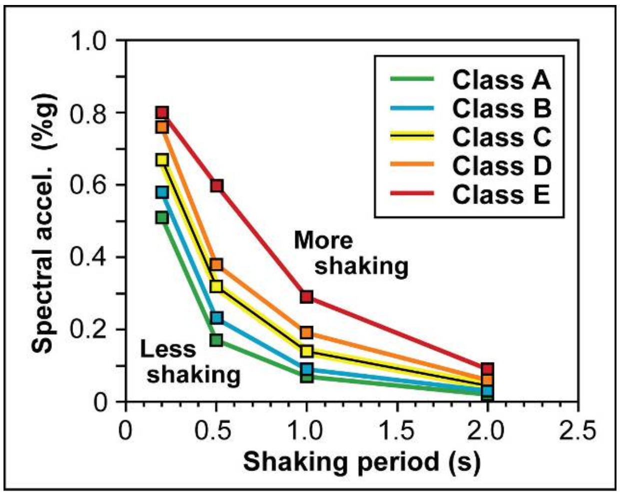

Figure 25. Spectral accelerations (ground motion) by site class for the 2005 NBCC design ground motion for Ottawa at shaking periods of 0.2, 0.5, 1.0 and 2.0 s (NRC, 2006), as explained in the text. The design ground motion has a 2% in 50 yr probability of exceedence (or once in 2475 years).

A seismic microzonation hazard map is being created for the Ottawa area that uses the 2005 NBCC seismic site categories; a draft version is shown in Fig. 26. This map has been compiled using a digital borehole database that consists of 20,000 water well and 1910 engineering records, most of which extend to bedrock. For map compilation purposes, the stratigraphy of the borehole records was summarized into three units based on Vs characteristics: 1) de-glacial/post-glacial sediments (consisting of glaciomarine, deltaic and fluvial deposits), 2) glacial sediments (diamicton and glaciofluvial deposits), and 3) bedrock. Hunter et al. (2007) found that the Vs of the de-glacial/post-glacial sediments is reasonably represented by a linear, travel-averaged Vs/depth function (Vsav/D; Fig. 27), while Benjumea et al. (2008) found that the glacial sediments and bedrock are reasonably represented by Vs average values of 503 and 2380 m/s, respectively. Using the Vsav/D function and the two average Vs values, the stratigraphy of each borehole record was converted into a Vs profile allowing Vsav for the upper 30 m of the ground surface (i.e., Vs30) to be determined for each borehole site. In turn, Vs30 allows the NBCC seismic site class to be determined using Table 2. The spatial pattern of the seismic site classes depicted in Figure 26 has been generated by extrapolating Vs30 between the borehole sites using a kriging interpolation. This map (at the time of writing) is currently being updated to incorporate 685 site measurements of Vs profiles obtained using refraction-reflection geophysical techniques, Nine additional down-hole borehole measures of Vs, and ~ 25 km of landstreamer shear-wave reflection data that allows improved definition of the geometry of key buried bedrock valleys (see Crow et al., 2007; Pugin et al., 2007; Hunter et al., 2008).

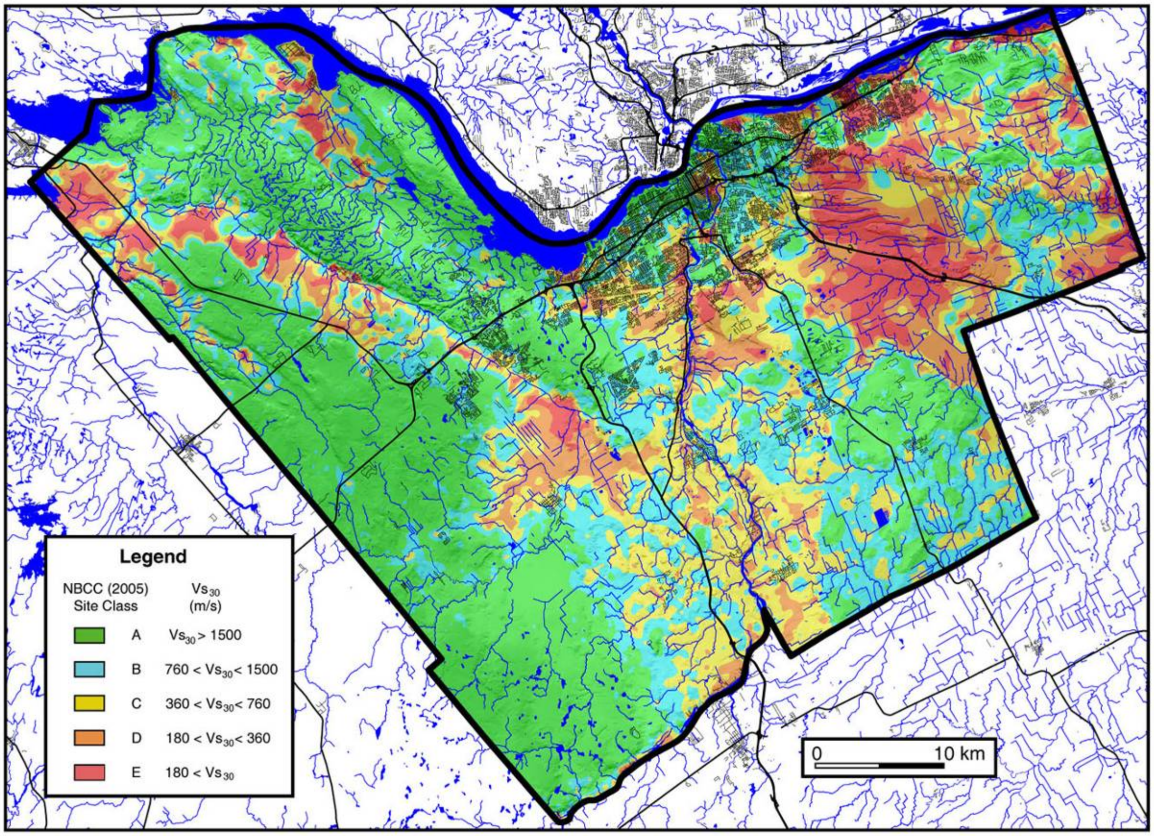

Figure. 26. Seismic site class map of Ottawa based on Vs_{30} – the average shear wave velocity over the upper 30 m of the ground surface. Refer to Table 2 for details on these classes.

As is apparent in Fig. 26, all five of the NBCC site classes are present in the Ottawa area indicating there are significant differences in seismic hazard spatially. In particular, the map reveals that class D and E areas, representing the more amplification-susceptible terrains, are present beneath some of the built-up areas of the city. The occurrence of classes D and E primarily relates to the presence of thicker deposits of 'soft' glaciomarine sediment (aka Leda clay) that locally are up to 100 m thick. These areas of thicker sediment are situated overtop of buried bedrock valleys that in many instances have negligible to weakly-defined surface expression. Reflecting the steeply-sloped margins of the buried valleys, the transitions from classes A to E locally can occur over distances of less than 0.5 km, for example, in the east Ottawa suburb of Orleans (see Motazedian and Hunter, 2008).

The distribution of the seismic site classes is directly relevant to emergency response planning and seismic mitigation strategies because the built-up urban areas overlying the D and E class areas are likely to experience more damage than the A and B areas during a significant earthquake, all factors being considered equal. The map in Fig. 26 allows a qualitative assessment to be made of the vulnerability of linear utilities (e.g., gas lines, water mains, and power lines) and linear transportation corridors (e.g., railways, highways), prioritization of structures for seismic retrofitting, and the siting of new critical infrastructure. The map is also relevant to the insurance industry for better assessing their exposure to earthquake risk and to aid in calculating 'fair' premiums that better reflect variability of local seismic hazards (Clark and Khadilkar, 1991; Smolka and Berz, 1991).

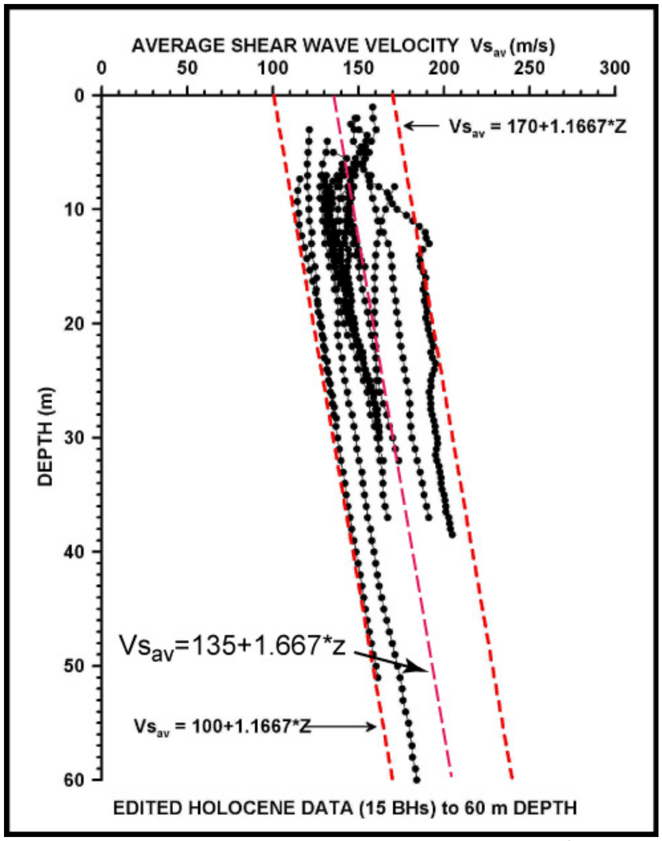

Figure 27. Plot depicting the average shear-wave velocity versus depth for 18 boreholes in the Greater Ottawa area (from Hunter et al., 2007). The plot is represented by a linear regression as well as upper and lower boundaries that define the distribution of points.

A more sophisticated approach to applying the map, however, is to utilize it within a mathematical analysis of potential losses from an earthquake event using risk assessment software, such as HAZUS-MH (see FEMA, 2009). These analyses take into account many variables to portray the vulnerability of a study area, such as, the characteristics and location of the built environment (buildings, essential facilities, transportation and utility lifelines), population densities, and building occupancy (based on the time of day). The loss estimate can be modeled for selected scenarios of earthquake magnitude, spectral characteristics and distance from the study area.

A copy of the draft seismic site class map was presented to both the Office of Emergency Management and the Planning and Infrastructure Department, City of Ottawa, and is contributing to improved recognition of seismic hazards in the Ottawa area. The Office of Emergency Management will be using the map in an upcoming review of the City's comprehensive vulnerability analysis as well as in prioritizing the City's response to the affected neighbourhoods during a significant earthquake event. Building Code Services of the Planning and Infrastructure Department, use the map on a regular basis as part of a screening process for assessing potential developments. As well, the information on the map can lead to the stipulation by the City for an increased level of geotechnical analysis for building permit approval. The seismic site class information on the map, however, is not intended for building design which should be based on site-specific geotechnical data collected by the developer or their consultants. The map contains a disclaimer clearly stating this fact.

Stop 4. Route 300 seismic profiles and cores

Susan Pullan, André Pugin, and Don Cummings[23]; Geological Survey of Canada

Objective

Stop 4 provides an opportunity to discuss and interpret compressional-wave (P-wave) and horizontally polarized shear-wave (SH-wave) seismic profiles collected along Route 300 (Fig. 6), and to compare these data with two continuous cores. Acquisition and processing methods, and the difference between P- and SH-wave data, will be outlined.

Setting Overview

Route 300 crosses an area of flat, low-relief agricultural land where the esker is completely buried by Champlain Sea mud.

Description

In 2006, the Geological Survey of Canada collected approximately 20 line-km of shallow seismic reflection data in order to delineate the three-dimensional structure of the buried Vars–Winchester esker-aquifer and the surrounding stratigraphy. P-wave data were collected using planted geophones and an in-hole shotgun source. SH-wave data were collected using a landstreamer and minivib source. Additional information on seismic reflection profiling and the minivib-landstreamer system can be found in the methodology section of this field guide.

Along this road (Route 300), both P- and SH-wave profiles have been acquired, and clearly delineate the buried esker and its cross-sectional architecture (Fig. 28). Both sections show the relatively flat-lying, fine-grained Champlain Sea deposits overlying the lower-frequency and less-coherent reflections related to the esker deposits. The esker deposits are observed to be at least 20–25 m thick at the crest, and the central core of the esker is ~ 200 m wide. The flanks extend an additional > 200 m on each side. Both sections display some evidence of “disturbed” (sandier?) sediment directly above the esker which suggests that there may be enhanced hydraulic connection between the surface and the esker in this region.

Seismic stratigraphy

Three main reflection-packages are observed in the Route 300 seismic profiles. 1) At the base of the profiles, a package of high- to moderate-amplitude reflections is observed at ~ 30–35 m depth (~ 35 metres above sea level). This represents the deepest interpretable seismic signal on the record. 2) A mound-shaped reflection package is present above this. It is relatively symmetric, ~ 800 m wide and ~ 20 m thick at its apex. Its flanks dip at low angles (2–5°). Reflections within the mound are generally low–moderate amplitude, often discontinuous, and somewhat chaotic. 3) At the top of the profiles, a package of low-amplitude, continuous, nearly-horizontal reflections is present (Fig. 28). The package varies in thickness from ~35 m off the mound to ~ 12 m over the crest of the mound. Reflections onlap the mound-shaped unit. A stream that crosses Route 300 (at break in P-wave profile – Fig. 28A, and at common mid-point (CMP) 640 - Fig. 28B) truncates reflections at the top of this unit. At ~ 10–14 m depth (~ 55 metres above sea level), a higher-amplitude reflection is observed within this unit. This reflection parallels overlying and underlying reflections.

Correlation of core and seismic data

The three seismic units described above correlate with carbonate-mudstone bedrock (or bedrock plus a thin overlying till), the buried Vars–Winchester esker, and Champlain Sea mud, respectively. High-amplitude reflections between seismic units are generated by the till or bedrock surface and by the top of the esker. Stiff diamicton (till) was intersected by the borehole

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[23] Pullan, S., Pugin, A, and Cummings, D.I., 2009. Stop 4: Route 300 seismic profiles and cores; *in*. Russell, H. A. J. and Cummings, D. I. (compilers), Deglaciation Of The Champlain Sea Basin, Eastern Ontario 72nd Friends of the Pleistocene Field Guide, June 6 – 7, 2009, p. 50–54.

midway along the esker flank but not beneath the esker crest. However, given the unit's thinness (less than several metres), this reflection is not easy to differentiate from the bedrock reflection at this location. The shallow higher-amplitude reflection within the Champlain Sea deposits is generated at a mud-on-mud contact (see below).

The seismic data clearly delineate the esker surface, and provides some information about its internal architecture (Figs. 28, 29). Core data indicate that the esker consists of two elements at this location, a gravelly central ridge and a broad sandy carapace. The gravelly central ridge is 12 m thick in the borehole near the esker crest. The seismic data suggest that it may extend laterally >100 m, though its top surface does not generate a distinct, coherent reflection. Clasts observed are predominantly carbonate-mudstone pebbles and cobbles, with minor percentages of granite and sandstone. A subtle fining-upward trend is observed in the uppermost several metres, which may help explain the lack of a distinct seismic reflection. The gravelly central ridge directly overlies bedrock in the borehole drilled at the esker crest, whereas in the borehole along the esker flank it overlies a thin till-sheet (Fig. 7). (The latter core did not extend to bedrock, but based on the seismic section, the till unit is likely to be very thin.) The cores indicate that the gravelly central ridge is sharply overlain by the sandy carapace, which is 7 m thick where it overlies the crest of the gravelly central ridge. Its top contact generates the high-amplitude, mound-shaped seismic reflection. Marine shells (mostly *Portlandia arctica*) are observed 2 metres and 11 metres below the top of the sandy carapace.

Champlain Sea mud that overlies the esker can be subdivided into three lithostratigraphic units, 1) moderately bioturbated, fining-upward rhythmites, which pass gradationally upward into 2) massive bioturbated mud, which is in turn sharply overlain by 3) coarsening-upward red-and-grey stratified mud (Fig. 7). The higher-amplitude reflection observed in the seismic data apparently correlates to a sand bed in the stratified-mud unit. Stratified mud overlying the sand bed is soft ("buttery") relative to the stratified-mud below the sand bed.

Acquisition and processing
Figure 29 demonstrates the effect and importance of the final processing steps (migration, topographic correction and depth conversion), using the SH-wave profile as an example. Topographic corrections related to a small creek crossing are significant (note the flattening of shallow reflectors in Fig. 29c).

Significant differences exist between the P- and SH-wave sections. The SH-wave data have a significantly higher vertical resolution, particularly in the shallow subsurface (0–20 m depth; see Fig. 29b). Continuous, coherent reflections are observed in the SH-wave data at depths of less than 5 m, whereas the P-wave section contains little information in this depth range (in part this is a limitation of the larger geophone spacing and source offsets used in P-wave survey). Deeper in the section, the P-wave data show a more coherent and higher-amplitude reflection from the interpreted bedrock surface at ~ 30–35 masl, while the SH-wave data show significant reduction in reflected signal below coarse-grained units (see Fig. 29b, CMPs 450–800, where bedrock reflection is very weak). As well, there are significant differences between the P- and SH-data in the relative amplitudes of some reflectors; for example, the amplitude of the esker surface along the flanks of the feature remains high in the case of P-waves (see Fig. 29a, CMPs 150–250 on west end of profile and CMPs 350–500 on east end), whereas the amplitude of this reflection drops considerably on the SH-section (see Fig. 29b, CMPs 150–350, 800–1200). These differences may be significant in terms of understanding the lithological and geotechnical information that can be gleaned from the seismic data, and need to be further investigated through borehole sampling and logging.

Conclusion
Seismic reflection techniques have produced excellent high-resolution data that have significantly improved the understanding of the cross-section form of the buried eskers and surrounding basin stratigraphy. The new knowledge on esker dimensions permits better estimation of the aquifer scale and groundwater reserves. The seismic facies observed also provide an indication of the

extent of a coarse-grained esker core, of the flanking esker sands, and of “mixed” zone above the esker crest. The SH-wave landstreamer-minivib system yielded higher-resolution data and has the added advantages of much faster data acquisition (2–3 times the data acquisition rate of the P-wave system), and fewer data gaps in survey lines (P-wave data cannot be obtained where shotholes cannot be drilled or where geophones cannot be planted; e.g. road/driveway/creek crossings, buried utilites etc.). The P-wave system provided better definition of the bedrock surface, as SH-waves do not seem to penetrate through overlying coarse-grained units (gravel or coarse-grained till?). However, the differences in reflection character between the P- and SH-data may provide important information related to the lithology or physical properties of the subsurface sediment. Further investigation and integration of groundtruth data, continuous core and downhole geophysics are required to understand these differences.

Discussion

- Why is till absent beneath the gravelly central ridge?
- What is the significance of marine shells in the sandy carapace of the esker?
- What is the significance of the high-amplitude reflection within the Champlain Sea mud package?

Additional reading:

Pullan, S.E., Pugin, A.J-M, and Hunter, J.A., 2007. Shallow seismic reflection methods for the delineation and hydrogeological characterization of buried eskers in Eastern Ontario. In Proceedings, SAGEEP’07 (Symposium on the Application of Geophysics to Engineering and Environmental Problems), April 1-5, 2007, Denver, CO, CD-ROM edition, 9 p.

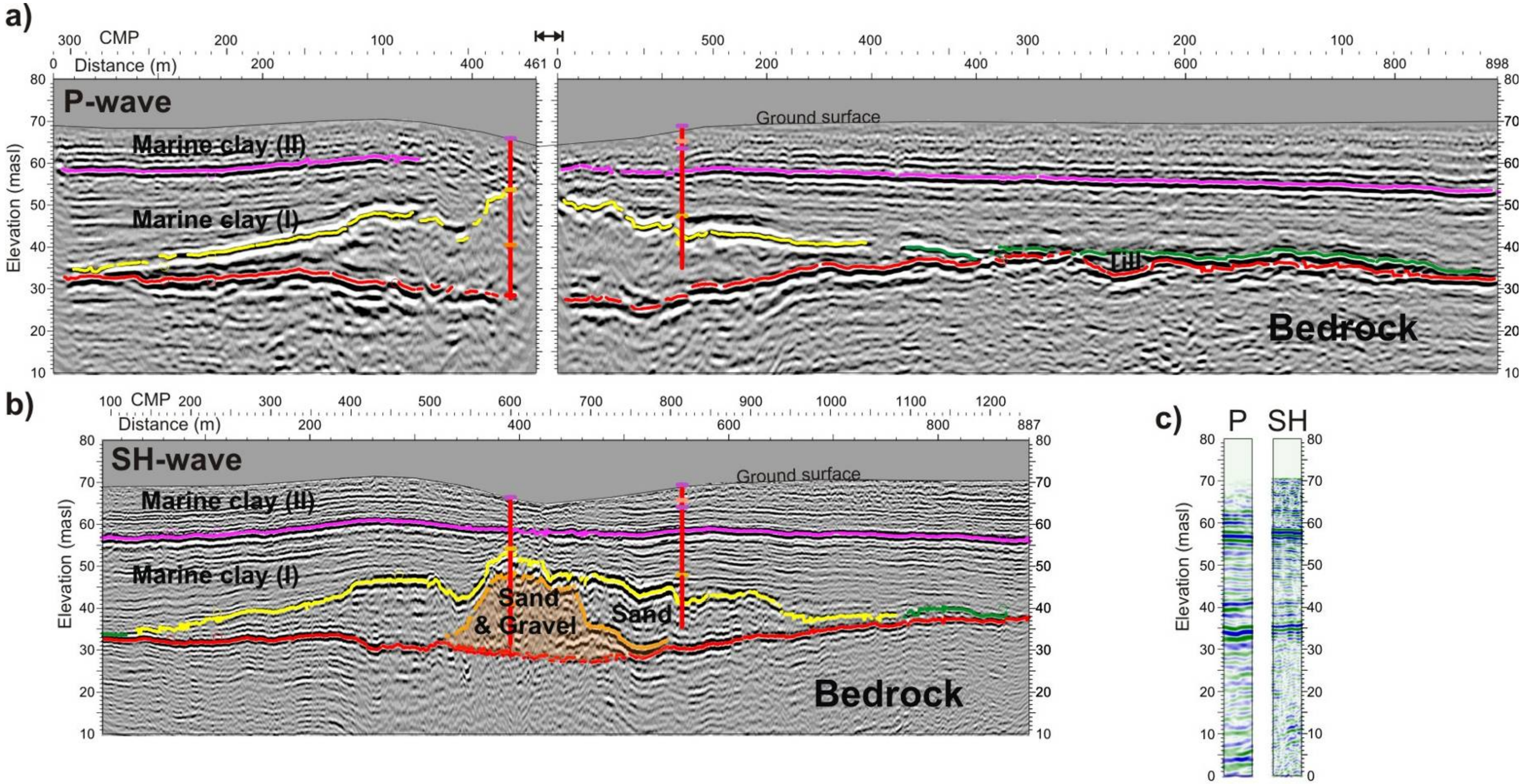

Figure 28. Comparison of a) P-wave and b) SH-wave processed seismic reflection sections (vert. exg.= 3x) over buried esker north of Embrun, Eastern Ontario. See discussion in text. The locations of the two boreholes drilled along this line in March 2007 are indicated by the red lines. c) Direct side-by-side comparion of short sections extracted from a) and b) showing higher resolution obtained with SH-waves, particularly in the upper 10–20 m.

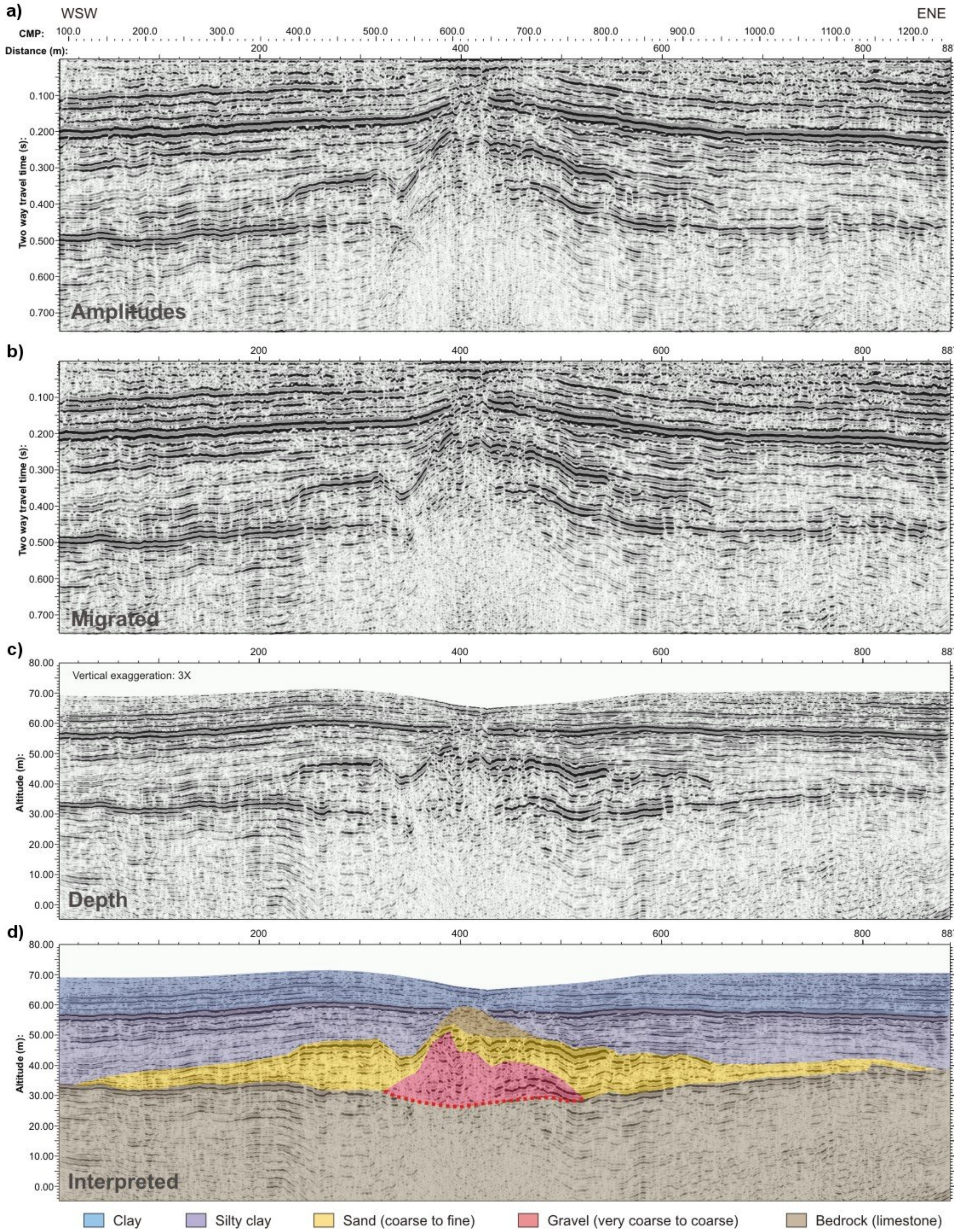

Figure 29. Processed SH-wave seismic reflection section over buried esker: a) amplitude section in two-way travel time; b) section a after phase-shift migration; c) section b converted to depth and with topographic corrections (vert. exg.= 3x); d) interpreted section.

Stop 5. Paleoenvironmental implications of the microfauna from the 3233 French Hill Road core

Jean-Pierre Guilbault*[24]*, BRAQ-Stratigraphie

We collected 24 samples for micropaleontological analysis from a core collected from 3233 French Hill Road near Rockland (Well A in Fig. 7). The top of the borehole is at 86 m elevation. From the bottom up there was 11 m of esker gravel (not sampled), 0.5 m of rhythmites (4 samples), 1.7 m of massive marine mud (3 samples), and 24.8 m of red-and-gray stratified mud (17 samples).

In the rhythmites, we found either minimal numbers of the freshwater ostracod *Candona*, or many *Candona* accompanied by foraminifera (Fig. 30). This widespread but as yet unexplained association of freshwater and saltwater elements is here particularly difficult to explain in that the foraminifera, though few, were suggestive of polyhaline conditions.

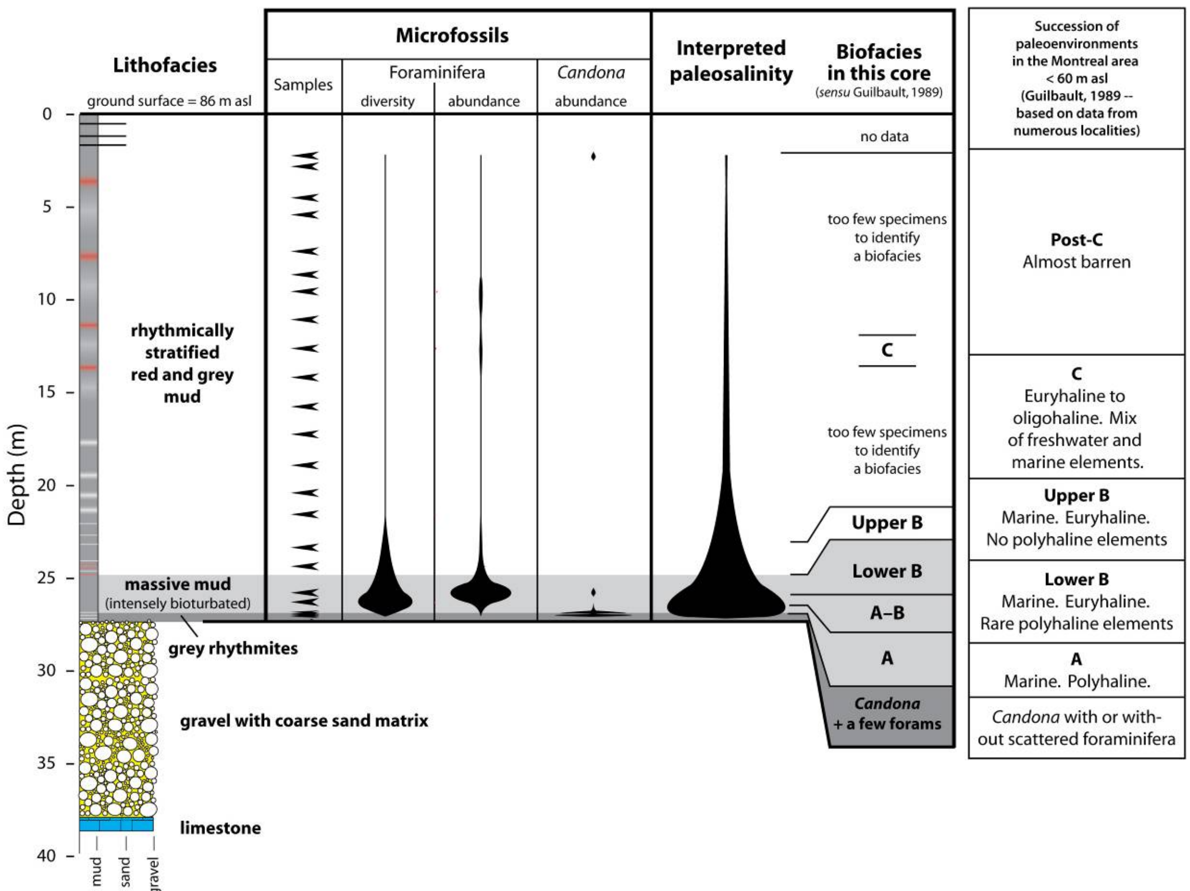

Figure 30. Microfossils observed in the core from 3233 French Hill Road (Well A in Fig. 7), biofacies (*sensu* Guilbault, 1989) and interpreted paleosalinity variations.

The massive mud samples were representative of Zone A and early Zone B of Guilbault (1989) in the central Champlain Sea, around Montreal. A fair proportion of the foraminifera belonged to polyhaline species, indicating salinities above ca. 25 ‰. In 2 of the 3 samples, ostracods were generally

[24] Guilbault, J.-P., 2009. Stop 5: Paleoenvironmental implications of the microfauna from the 3233 French Hill Road core; *in*. Russell, H. A. J. and Cummings, D. I. (compilers), Deglaciation Of The Champlain Sea Basin, Eastern Ontario 72nd Friends of the Pleistocene Field Guide, June 6 – 7, 2009, p. 55–56.

polyhaline but, in the lowermost massive mud sample, the ostracod assemblage consisted almost entirely of *Candona*. According to L. Delorme (written comm.) *Candona* cannot tolerate even the slightest salinity. This apparent contradiction can be explained if we consider that this sample is from a level that is still vaguely rhythmic, and that it includes a large burrow coming from a few centimeters above. The foraminiferal assemblage would be from the burrow and the *Candona*, from the encasing rhythmites. This, plus the result from the rhythmite unit, implies a very close association of freshwater and polyhaline marine waters, perhaps on a seasonal basis.

Above 24.8 m, in the red and gray stratified lithofacies, foraminifera and ostracod species are few, and all species are euryhaline. The number of specimens in general is small and irregular, except at the base of the unit, where the two lowermost samples correspond to the upper part of the B biofacies near Montreal. One sample, at 12.6 m, is representative of the C stage near Montreal. Where there are enough indicators to conclude, we can assume paleosalinities of between 5 and 25 ‰, with maximum values at the base. We could not find the gradual and regular decline in numbers and diversity down to zero reported by Guilbault (1989). The succession is not very different from the microfauna assemblage of the Mer Bleue CRF-21A borehole (Gadd, 1986; Guilbault, 1989). The main difference is that at Mer Bleu, the upward transition from abundant to small-number foraminifera assemblages (50 or less) takes place within the massive mud whereas in the present core, it occurs inside the stratified mud.

Rodrigues (1992) reported a succession of very gradual increase in salinity, mostly in the Ottawa region, starting with *Candona*, followed by gradual replacement by euryhaline ostracods and finally, gradual development of foraminiferal faunas. This is completely different from our results. The explanation may reside in the fact that this section rests on the top of an (just formed?) esker, and better sampling may help us understand the relationship between the eskers and the marine environment.

Sunday

Stop 1: Cantley bedrock and erosional s-forms

David R. Sharpe[25]. Geological Survey of Canada

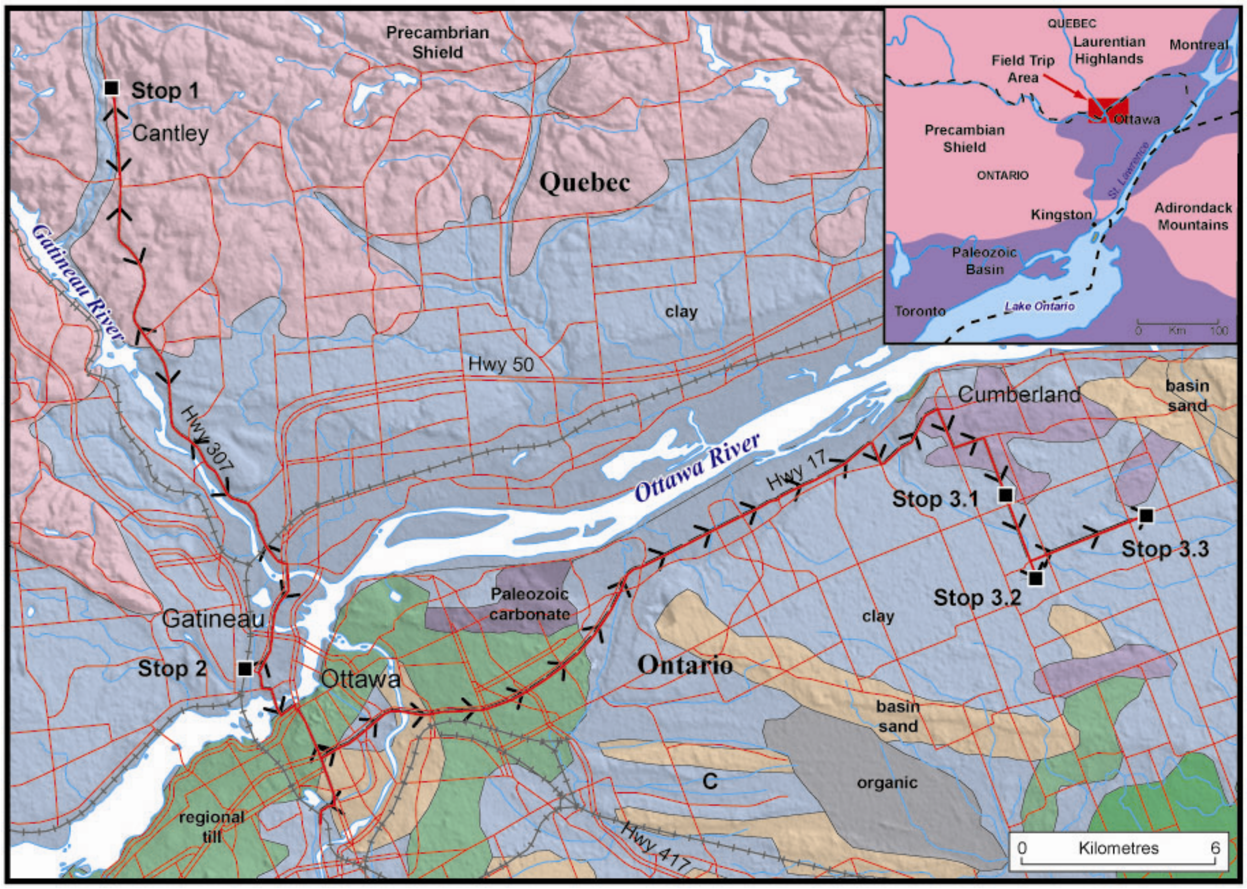

Figure 31. Geology and digital terrain model of the Ottawa area showing fieldtrip stops and route. Inset provides setting of eastern Great Lakes.

Location: Cantley Quebec, NTS 31G/12, UTM 438049; 45 35' N, 75 21' W. Parking is in front of a site gate on the west side of Highway 307 (Fig. 31). The quarry is owned by Les Entreprises Vetel Ltée; access is at your own risk after writing to the owner.

Purpose

The purpose of this stop is to draw attention to the significance and implications of erosion in glaciated terrain. Bedrock exposures within the quarry display subglacial erosional forms that are inferred to be mainly cut by powerful subglacial meltwater flow (Sharpe and Shaw, 1989) within a tributary valley of the Gatineau River (Fig. 31).

[25] Sharpe, D.R., 2009. Stop 1: Cantley bedrock and erosional s-forms; *in*. Russell, H. A. J. and Cummings, D. I. (compilers), Deglaciation Of The Champlain Sea Basin, Eastern Ontario 72nd Friends of the Pleistocene Field Guide, June 6 – 7, 2009, p. 57–64.

Description of site

Bedrock within the quarry consists of Precambrian marble containing resistant granitic or volcanic clasts. The studied outcrop is streamlined parallel to valley orientation and to regional ~ north-south ice flow (Fig. 32). This site was deglaciated prior to 12 000 BP, but until recently, the described features were covered with gravel, sand, and mud deposited rapidly on subaqueous fans formed at the margin of the Champlain Sea (e.g. Rust, 1988; Sharpe, 1988), and below a marine limit of ~ 200 m asl. Similar erosional forms occur on other outcrops in this poorly-exposed valley. Sculpted forms as observed on the bedrock surface at this site diminish and disappear on adjacent outcrops at higher elevation. At the north end of the Cantley site, a 15 m thick sequence of glaciomarine sediments overly the site and consist of a fining upwards sequence of sandy gravel and bedded sand overlain by silt and laminated clay.

Figure 32. Photograph of major outcrops at the Cantley rock quarry, Stop 1. R is rock drumlin, RR is remnant ridge. Outcrop forms are oriented parallel to the valley (~N-S) and in alignment with flow directions on small features in the quarry. See person for scale.

Erosion forms

Forms range in size from centimeter-scale striations and rat-tails, to meter-scale obstacle marks, to forms the size of the outcrop, 50-100 m long, 5-10 m high (Fig. 32), and perhaps larger north and south of the area.

Striations and rat-tails

Portions of the rock surface are planed-off and striated, such that the granitic and volcanic inclusions lie flush with the surface of the surrounding marble. Small rat-tails are present in addition to striations.

Rat-tails are positive forms and indicate differential erosion of the surrounding rock surface, rather than erosion of striations by a tool held in ice.

Obstacle marks and sichelwannen

The most common and distinctive erosion forms at the site are obstacle marks (Fig. 33), which have ridges in the lee of obstacles (Allen, 1982). These forms consist of a proximal, crescentic furrow wrapped around an upstanding obstacle, or resistant bedrock clast (Fig. 33b). The proximal furrow commonly has a sharp leading margin. The arms of the crescentic furrow extend downflow in a pair of furrows that become shallower and wider downflow (Fig. 33a, b).

The furrows are often smooth or less striated than adjacent surfaces outside the furrow (Fig 33a), and they can contain divergent flow features (Fig. 33b). Furrows also extend far downflow (Fig. 33c) to form longitudinal forms (Kor et al., 1991). Remnant ridges, which form behind the obstacle and between the furrows, may occur at several scales (compare Fig. 33a, c). An element (d), divergent from main flow, indicates vortex flow in the upflow furrow (F; Fig. 33c).

Other observed sculpted forms occur without obstacles; they have been called sichelwannen (Ljugner, 1930) or crescentic scours (Dahl, 1965). These are now termed s-forms (Kor et al., 1991) and consist of transverse elements with furrows extending downflow around a central medial ridge (Fig. 32). At this scale, the remnant medial ridges can be considered to be rock drumlins (Fig. 32). A smaller, mussel-shaped transverse form (muschelbruch) occurs with sharp margins but without a medial ridge (see M, Fig. 33a).

Channels or Cavettos and Potholes

Furrows that extend down flow (Fig. 33c) as longitudinal forms result in channels or cavettos (Dahl, 1965). They occur on vertical rock faces either as elongate troughs, or, as winding channels with tight curves (Fig. 34). Cavettos are commonly meters long and centimeters to meters deep, and they may crosscut other features (Fig. 34b). They may have vertical segments similar to truncated potholes (Gjessing, 1967). Vertical segments (Fig. 34c) are considered to be non-directional forms (Kor et al., 1991). At this site, rat-tails indicate that the eroding flow was only upward, unlike classical potholes, in which flow at the outside of the form is downward. Vertical flux implies flow under a very high hydrostatic pressure gradient.

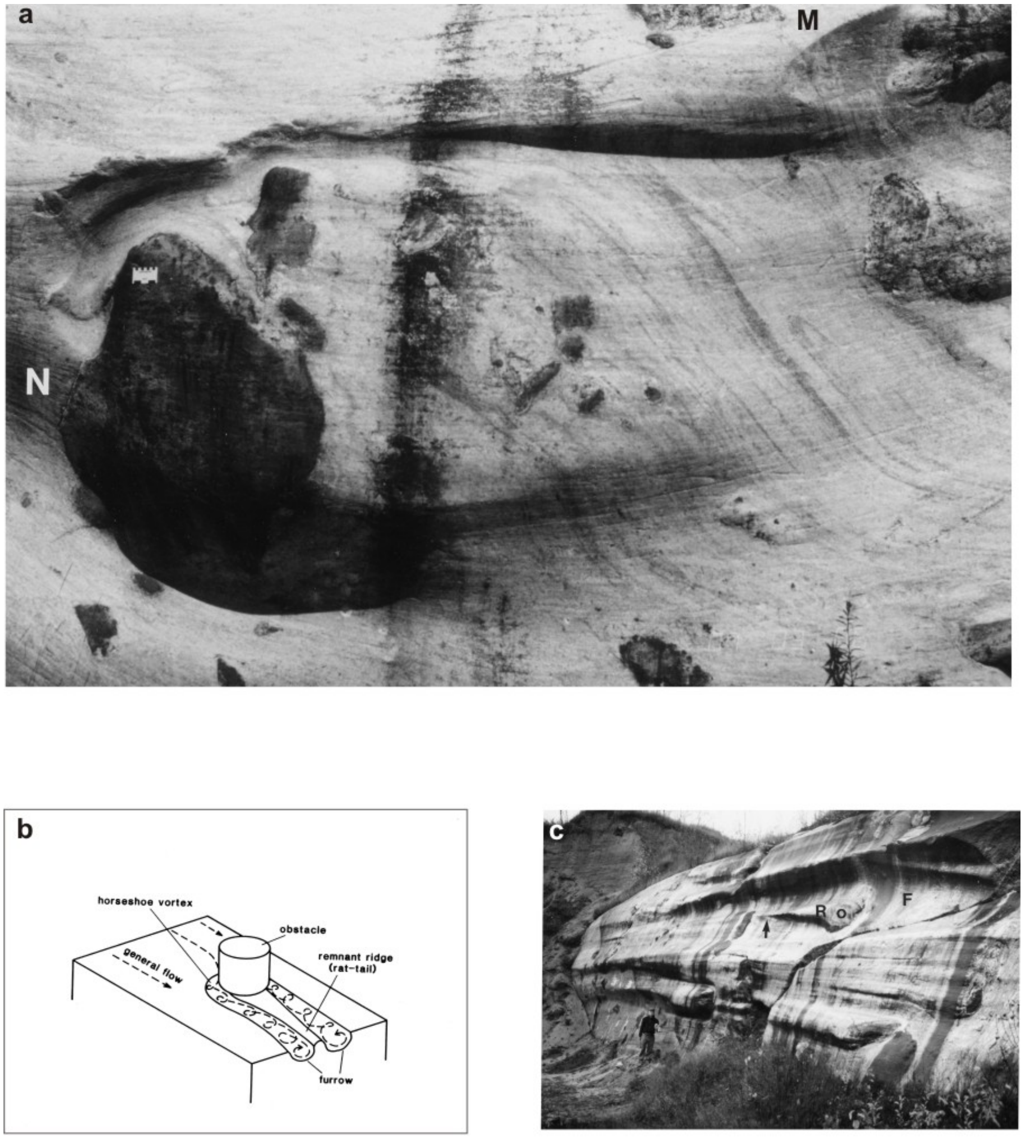

Figure 33. Obstacle marks (transverse s-forms): a) Sculpted forms diverge around dark inclusion (see 8 cm scale card at top) or obstacle within marble, set on a vertical rock face. Sculpted arms, or furrows, become wider and shallower downflow to the right, leaving a remnant ridge in the lee of the obstacle. Note striation or lineation, and rat-tails on the rock surface. Flow left to right. b) Conceptual process model of obstacle mark formation under turbulent flow (Allen, 1971). c) A number of obstacle marks (O) occur on rock surface, some with large furrows (F) upflow of an obstacle and well-defined tapered remnant ridges (R). Note large, divergent rat-tail within the upflow furrow. Flow right to left parallel to outcrop orientation. Note person for scale, lower left.

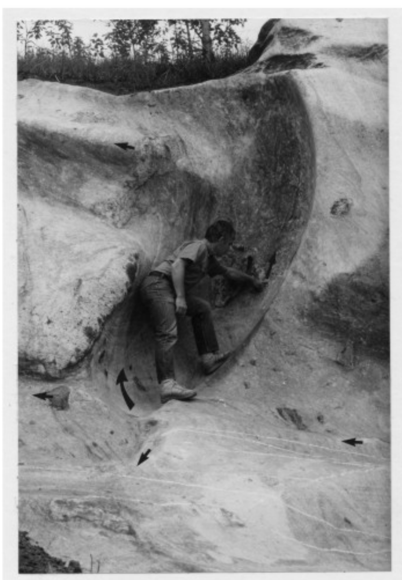

Figure 34. Furrows, cavettos and potholes: a) (top photo) Poorly-developed upflow furrow extends to lengthy longitudinal furrows (~10 m) and remnant ridge. Flow right to left parallel to outcrop orientation. b) (lower left photo) Cavetto or channel-like form with vertical orientation. Flow is upwards as indicated by rat-tails (arrow). This vertical structure had flow directed upwards rather than downward as in potholes. This indicates very high piezometric heads in the subglacial water system. c) (lower right photo) Complex pattern of longitudinal s-forms; furrows and cavettos. Flow left to right parallel to outcrop orientation. Note person lower right for scale.

Other features

Interior surfaces of many s-forms are scored by light scratches. The scratches are short and are generally aligned with the direction of flow indicated by the divergent axes of troughs and associated rat-tails . They also diverge as the sculpted form becomes wider downflow.

Carbonate precipitate is found on polished surfaces in places, particularly in the lee of obstacles, and on remnant ridges.

Pitting can occur on the upflow side of obstacles with remnant ridges. This appears to be incipient development of upflow furrows, perhaps as an emerging rock ridge is exposed to accelerated erosion.

Cross-cutting forms

There are many cross-cut relationships involving striations and sculpted forms on Cantley outcrops (Fig. 33c, 34b, c). For example, striations occur as ornamentations, set at an oblique angle, on top of a transverse s-form (sichelwannen). In some forms, rat-tails readily show divergent flow within s-forms (Fig. 34c).

Interpretation

Glacial Abrasion

From the evidence of striations on the surfaces of some sculpted marks, it is concluded that the erosional forms at Cantley were created subglacially. Striations and planed-off inclusions indicate glacial abrasion. Such abrasive planning is attributed to erosion by debris in the bed of flowing glacial ice (Hallet, 1981). There are fewer planed erosion surfaces than surfaces carrying sculpted forms (s-forms). S-forms, in contrast to striations, likely relate to differential erosion expected in flow systems with secondary, turbulent structures (Allen, 1971). Alternatively, the possibility of glacial formation of obstacle marks considers laminar streaming of debris-rich glacial ice around an obstacle (Boulton, 1974). In addition, Gjessing (1965) proposed that, although some s-forms result from fluvial erosion, others are a product of erosion by a subglacial slurry of saturated till. Others suggest that the saturated till, as a deforming bed, was responsible for erosion of streamlined rock forms (e.g. O'Cofaigh et al., 2005).

Corrasion in turbulent separated flow

Cantley obstacle marks, and other transverse s-forms, are thought to result from glaciofluvial processes (Sharpe and Shaw, 1989; Kor et. al., 1991). S-forms associated with obstacles are likely formed by flow separation and horseshoe vortices (Fig. 33c), resulting from a vertical pressure gradient generated at the upstream face of the obstacle in turbulent flow (Shaw, 1988, 1994). This gradient sets up secondary flow and a pair of oppositely rotating vortices (Fig. 33c). These secondary flows reattached to the bed where high velocity fluid approached at a high angle and caused maximum erosion (Allen, 1971, 1982). The crescentic furrow cut around the leading side of the obstacle and the paired furrows extending downflow are likely products of a horseshoe vortex. These vortices expanded rapidly and thus, became reduced in intensity and erosional power downstream. As a consequence, the furrows become broader and shallower downflow (Fig. 33b, c), and, rat tails, the remnant (rock) ridges between the furrows, becomes narrower and lower. Sichelwannen similarly relate to flow separation and horseshoe vortices (Kor et al., 1991), whereas muschelbruch (Fig. 33a, b) likely relate to impingement of low-angle vortices on the bed (Shaw, 1988).

Some s-forms that are oblique to the general flow direction indicated by a longitudinal element are preserved due to their protected location within a furrow. Most appear to represent secondary flows related to vortices within a primary flow. This type of mechanical erosion at Cantley is considered to be glaciofluvial corrasion and can occur by direct fluid stressing, or, it can be related to erosion by tools carried in flowing water (Allen, 1982).

The smaller scale forms described are also identical to some sculpted fluvial forms in terrain subject to flooding in Australia (Baker and Pickup, 1987). Some larger forms are similar to forms identified within the Channeled Scablands of Washington State (Bretz, 1925; Baker, 1978). These comparisons support the interpretation of formation by water erosion. Allen (1971, 1982) also verified experimentally that erosional forms such as these can be produced by high-velocity, separated, turbulent fluid flow.

Cavitation

Cavitation is the rapid formation and collapse of bubbles due to local pressure gradients in turbulent flow. It becomes a viable erosion process where turbulent flow has high enough velocity (~5-10 m/s; Hjulstrom, 1935; Richardson and Carling, 2005). Damaging shock waves and violent jets of water score turbine blades, and can loosen grains in rocks, pit massive rocks, such as may be inferred in places at Cantley.

Dissolution features

There is some evidence of dissolution of Precambrian marble at Cantley. There are, however, few dissolution features, and carbonate precipitation indicates a non-dissolution regime following main subglacial erosion. It does not appear to be a dominant process, despite the suggestion of Hillaire-Marcel (2005) that the site 'represents the mixed influence of mechanical abrasion by basal ice and of dissolution features due to high pressure subglacial water channeling'. Dissolution features probably relate to minor post-glacial modification of common sculpted forms that has taken place following removal of sand and gravel from the quarry rock surface in recent years.

Discussion

The outcrops at Cantley display s-forms (e.g. obstacle marks, furrows, and cavettos) with sharp rims, divergent flow features, and remnant ridges. They also show ice abrasion forms, striations, and plucked forms such as crescentic fractures. The occurrence of abrasion, pitting, polishing, and carbonate precipitate with meltwater forms suggests that the meltwater flows were subglacial. Lifting of ice from its bed by fast-flowing meltwater suspended glacial abrasion. When ice settled back on the bed, as the meltwater flow subsided, abrasion resumed, rounding sharp edges and lightly striating rock faces, in places at oblique directions.

The association of forms produced both by glaciofluvial erosion and ice abrasion suggests that the glacier was lifted from and let down on the bed during subglacial floods. The assemblage of sculpted features at Cantley is best explained by differential erosion produced by strong vortices. Rapid, sediment-laden, turbulent, subglacial meltwater flows likely produced most forms by corrasion and cavitation erosion (Sharpe and Shaw, 1989). Depositional sequences (sand and gravel) related to these high-energy meltwater outbursts were probably deposited on subaqueous fans in the adjacent Champlain Sea basin (e.g. Rust, 1988; Sharpe, 1988).

Regional process implications

Water-sculpted erosion forms imply subglacial bed conditions with little frictional resistance in the areas of subglacial meltwater flow at the time of such discharge (Shoemaker, 1992). The large inferred discharge rates also require meltwater storage, likely in subglacial reservoirs (e.g. Alley et al., 2006; Evatt etal., 2006), up-ice from the Cantley site. Separation of the glacier from its bed by a subglacial meltwater sheet, as in some Icelandic floods, also involves minimal basal resistance to ice flow (Shoemaker, 1992; 1999), and a flat ice sheet is expected to cover such discharge and storage areas (Wingham et al., 2006). The importance given here to meltwater events corresponds to that applied by others to subglacially deforming till beds (e.g. Boulton and Hindmarsh, 1987; O'Cofaigh et al., 2005). Both process models imply low basal shear stresses, a relatively flat ice-sheet profile, and minimum ice volumes for a given ice-sheet radius. It is important to note that, in the meltwater explanation, these conditions may occur even where the substrate is bedrock. The rapid discharge of meltwater interpreted here is expected to have been accompanied by accelerated ice flow, perhaps surging (Kamb et al, 1985), or ice streaming (Bell et al., 2007). If ice-sheet profiles are to be credibly reconstructed, it becomes critical that subglacial meltwater forms be mapped and their timing be assessed (Sharpe, 2005).

Key questions posed during the Cantley field trip stop

1. Are sculpted erosion forms (s-forms) distinguishable from ice flow indicators (striations etc.)?
2. Do s-forms and striations record fundamentally different processes?

3. Should s-forms and related features be mapped and recorded on maps separately from ice flow indicators (striations)?
4. Is it possible that s-forms and ice flow forms record different manifestations of the same event, or closely timed events?
5. If s-forms are present across a region, is it reasonable to assume/ link a regional event(s) as is the case of with striation mapping.
6. If rapid flow events are inferred in this region, are ice streams responsible for erosional and depositional landforms?
7. What are the processes responsible for rapid flow if ice streams operated in this region?
8. Are ice stream and meltwater-flow features identifiable on this mapping landscape?
9. Could ice streams and meltwater floods occur in closely-timed or the same event sequence?
10. Are the drumlins south of the fieldtrip area erosional or depositional? How can we test/ constrain either case?
11. If drumlins south of the fieldtrip area are erosional, how are striations under a sediment drumlin linked to the landform sediment and its orientation?

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