GEOLOGICAL SURVEY OF CANADA
OPEN FILE 6947

Deglacial history of the Champlain Sea basin and implications for urbanization;

Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25–27, 2011;
Fieldtrip guidebook

H.A.J. Russell, G.R. Brooks, and D.I. Cummings
(Editors)

2011

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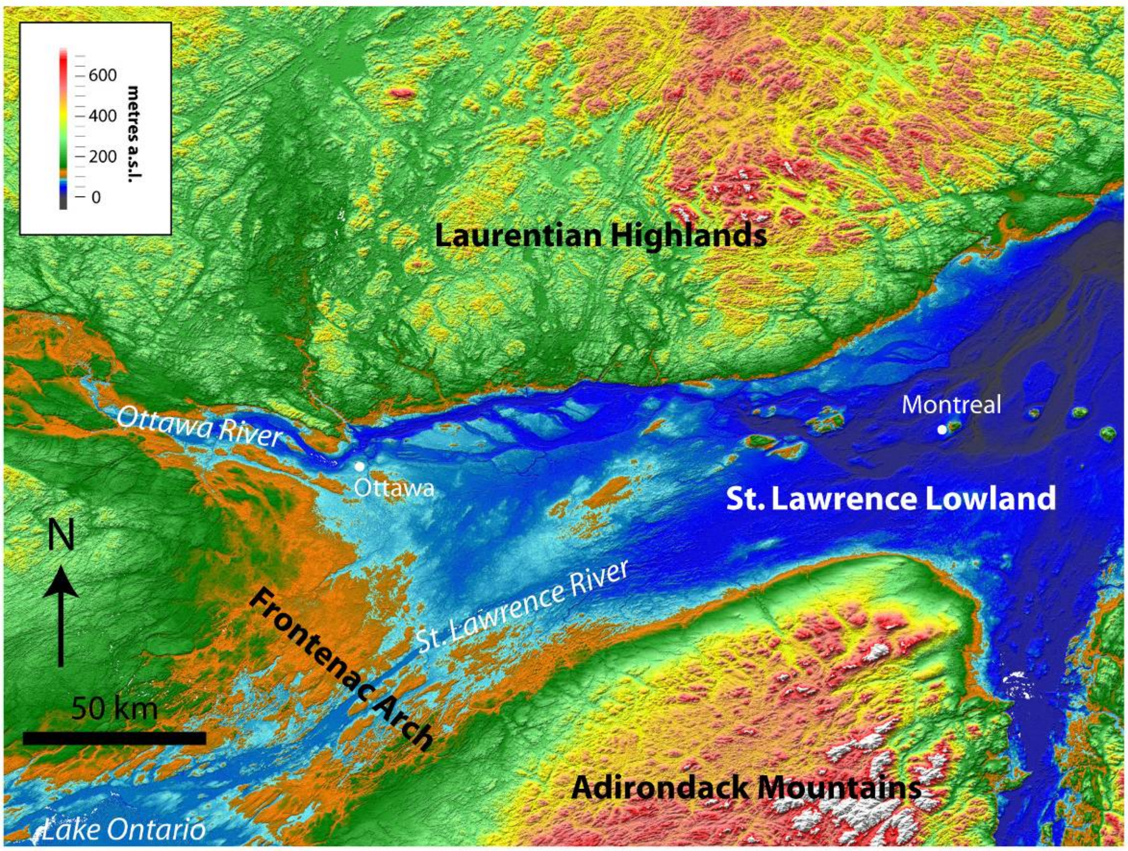
Publications in this series have not been edited; they are released as submitted by the author.

Deglacial history of the Champlain Sea basin and implications for urbanization

Joint Annual Meeting of the
Geological Association of Canada, the Mineralogical Association of Canada,
the Society of Economic Geologists and the Society for Geology Applied to Mineral Deposits

May 25–27, 2011, Ottawa, Ontario



Hazen A.J. Russell, Gregory R. Brooks and Don I. Cummings
(fieldtrip guidebook editors)

With contributions from
Sam Alpay, Jan M. Aylsworth, Heather Crow, Marc J. Hinton, James A. M. Hunter, Ross Knight,
Maurice Lamontagne, Ted Lawrence, Charles Logan, Barbara E. Medioli, Didier Perret, André Pugin,
Susan E. Pullan and David R. Sharpe

Abstract

The Champlain Sea was an inland arm of the Atlantic Ocean that inundated the St. Lawrence Lowland following retreat of the Laurentide Ice Sheet. The fine-grained sediments deposited in this sea have important implications for urbanization (e.g. slope stability, foundation design, seismic hazard assessment) of the National Capital Region. This two-day field trip reviews aspects of the deglacial landforms and deposits of the area, the Champlain Sea deposits, and reviews the societal implications from the perspectives of hydrogeology and natural hazards. On day one, a visit to the Vars–Winchester esker and environs provides the setting for discussing eskers and Champlain Sea deposits and the importance of these features with respect to regional hydrogeological issues. On day two, the focus shifts to natural hazards, with visits to a cluster of earth-flow scars at Breckenridge, Quebec, and an earth flow near Notre-Dame-de-la-Salette, Quebec, that was triggered by an earthquake on June 23, 2010. Discussion will focus on the seismicity of the earthquake and issues of seismic amplification due to variations in local geology.

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 River conservation authorities. Data support by the Ontario Ministry of Natural Resources and the Ottawa–Carleton Regional Municipality is acknowledged. The research that fostered this guide book was supported by the Groundwater and Public Safety geoscience programs, Earth Sciences Sector, Natural Resources Canada.

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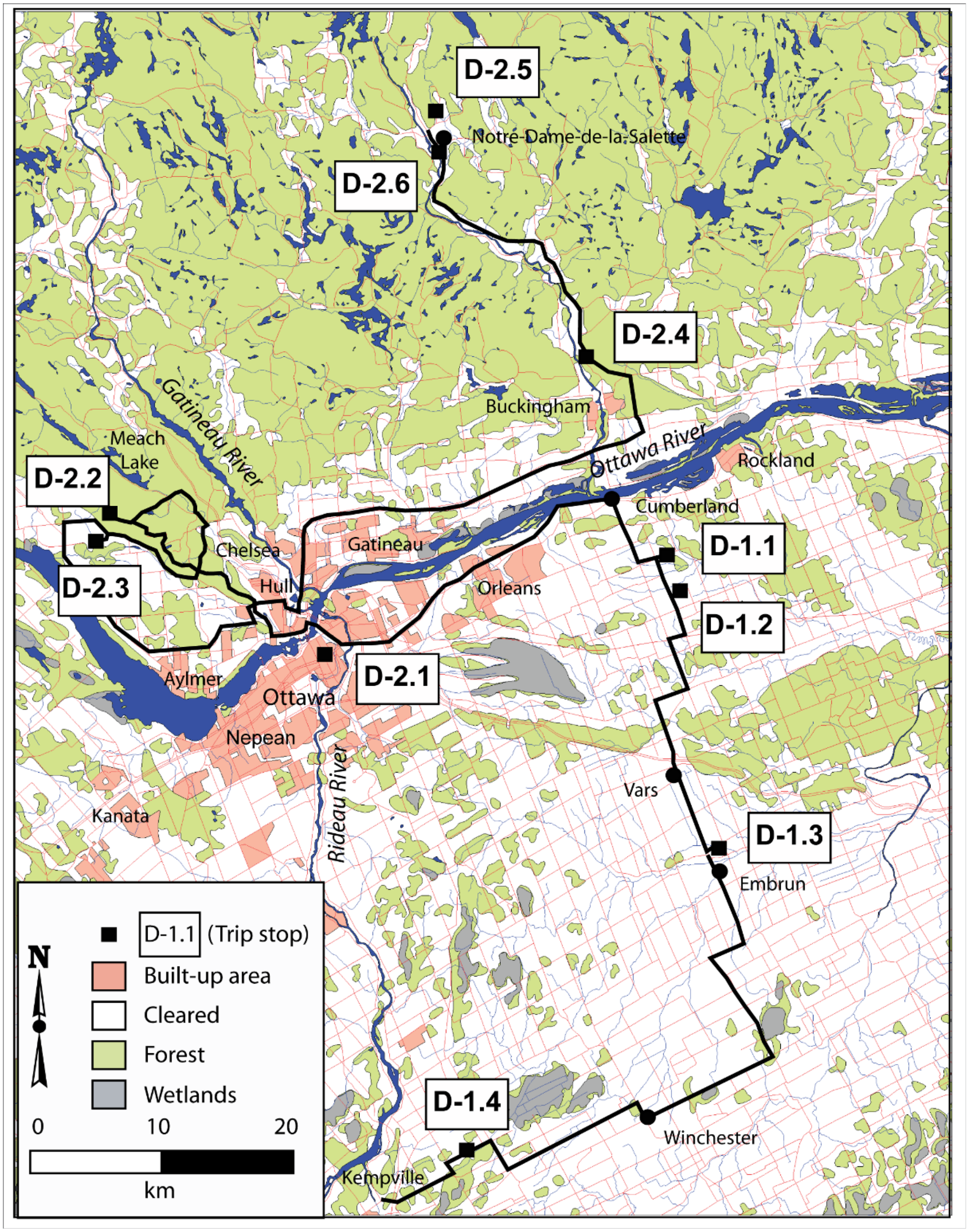
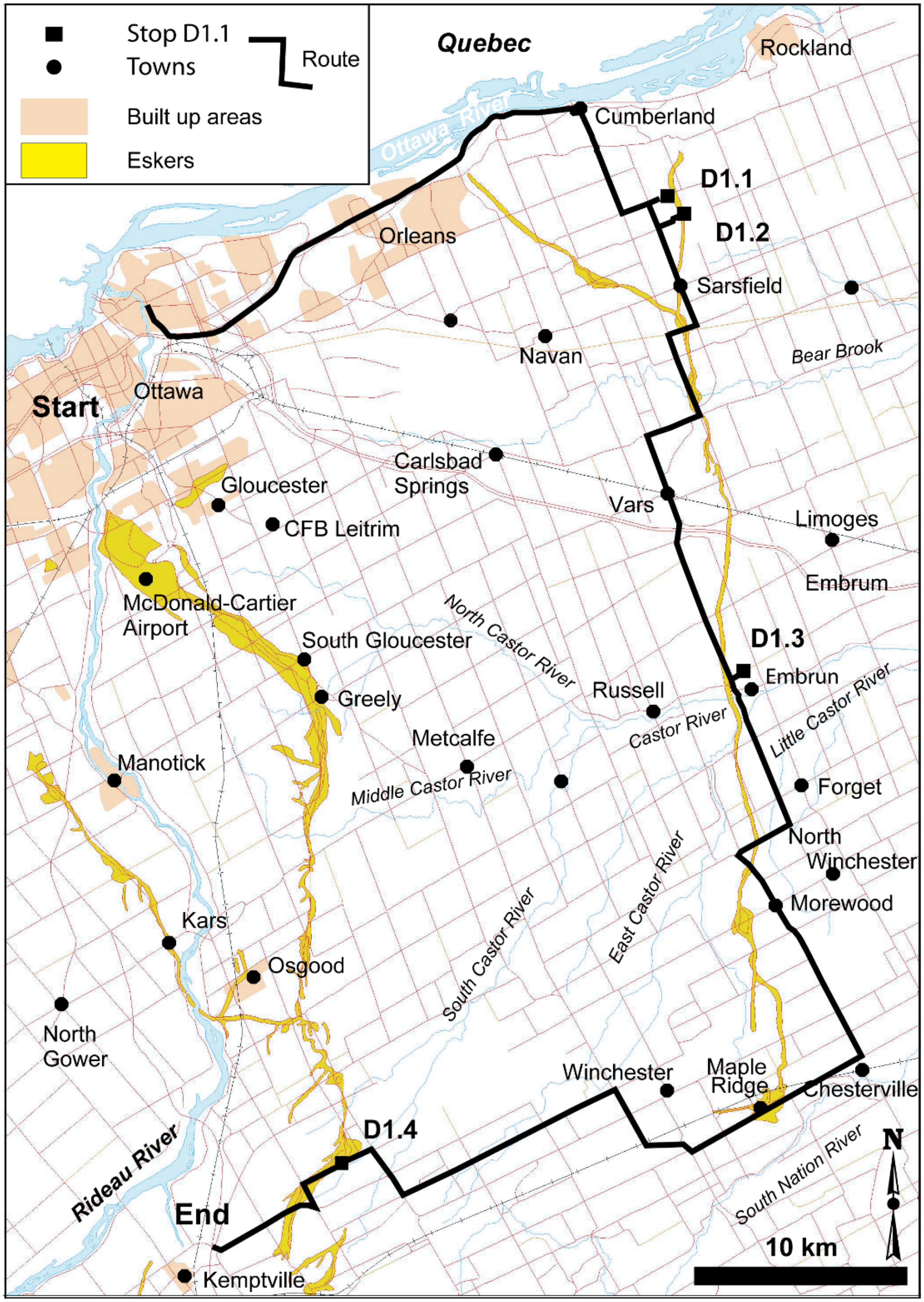
Table of Contents

Field Trip Route Map



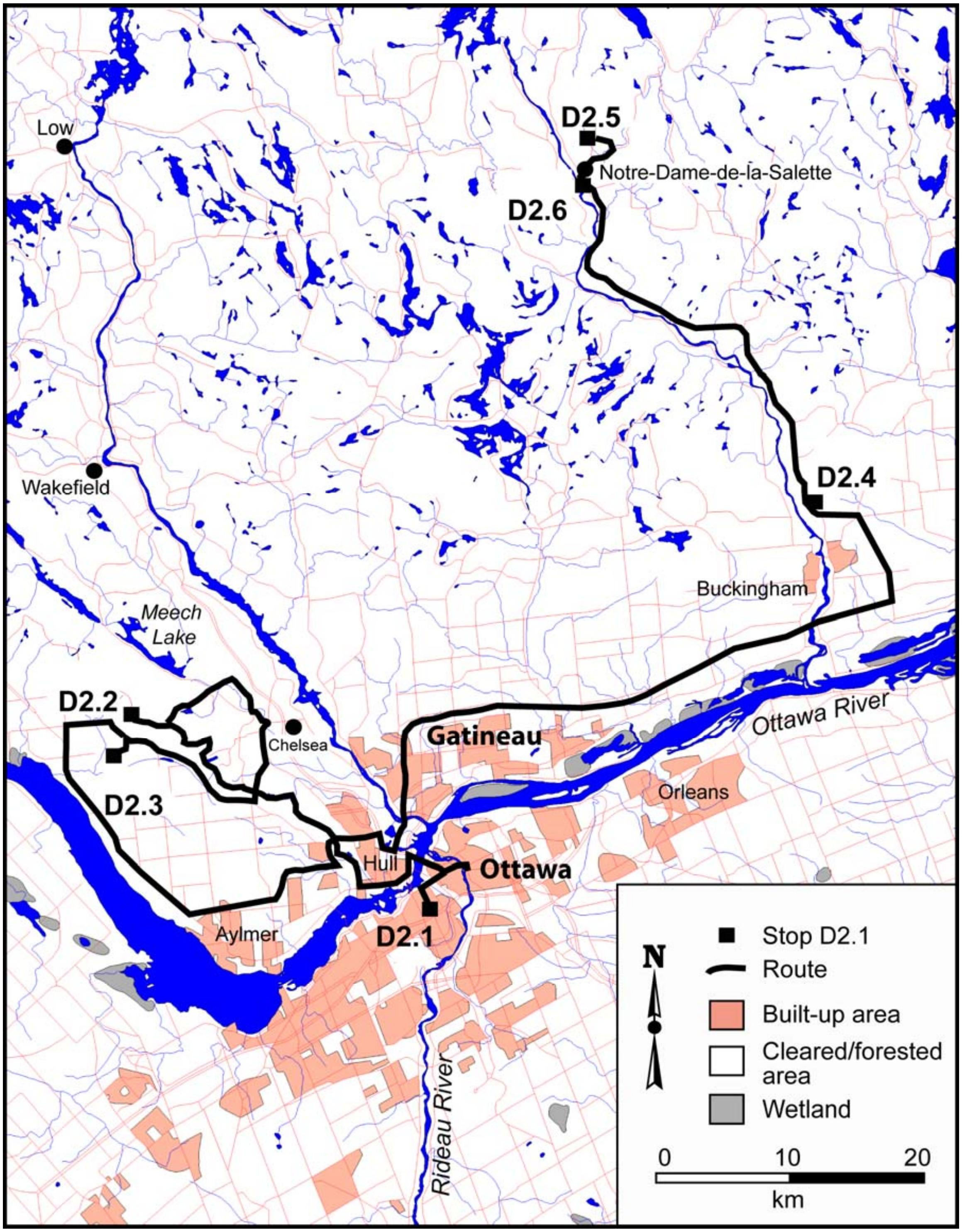
Map 1.Field trip route. Forested areas north of Ottawa are underlain by Precambrian bedrock. Areas east and south of Ottawa are predominantly underlain by Champlain Sea muds and Palaeozoic sedimentary rocks.

Field Trip Route Map: Day One



Map 2. Route map of day one of field trip southeast of Ottawa to the Vars–Winchester esker and Kemptville area.

Field Trip Route Map: Day Two



Map 3. Route map of day two of the field trip to landslide areas and sites of interest in west Quebec.

FIELD TRIP SCHEDULE

DAY ONE

Stop 1: French Hill Road.
1a. Geological overview (Cummings)
1b. Seismic reflection techniques and buried valley (Pugin and Pullan)

Stop 2: Regimbald Road pit
2a. Esker sedimentology (Cummings)
2b. Hydrogeological Review (Hinton et al.)

Lunch at

Stop 3: Route 300
Seismic transect of esker and core (Pullan & Cummings)

Stop 4: Kempville subaqueous fan pit (Russell and Cummings)

DAY TWO

Stop 1: Museum of Nature
Ottawa-Gatineau microzonation mapping

Stop 2: Champlain Lookout
Champlain Sea deposits in western of Ottawa-Gatineau

Stop 3: Breckenridge Valley
3a. Landslides and the buried Breckenridge buried valley
3b. Mapping buried bedrock basins - Tromino demonstration
3c. 2008 earth flow – modern failure and landslide dam
3d. Prehistoric landslide scars

Lunch at Breckenridge Valley

Stop 4: Buckingham Subaqueous fan Deposit
Hyperconcentrated flow deposits

Stop 5: Earthquake triggered earth flow near Notre-Dame-de-la-Salette
5a. Val-des-Bois earthquake of June 23, 2010
5b. Earthquake-triggered landslide at Notre-Dame-de-la-Salette

Stop 6: 1908 Notre-Dame-de-la-Salette earth flow
6a. 1908 Notre-Dame-de-la-Salette landslide

Introduction

The Champlain Sea was an inland arm of the Atlantic Ocean that inundated the St. Lawrence Lowland following retreat of the Laurentide Ice Sheet (Figs. 1, 2). The sea existed from about 12 000 to 10 000 yr BP (Fulton and Richard, 1987), its level falling continuously as the crust rebounded isostatically. Although both glacier and sea are now long gone, the deposits they left behind preserve a detailed record of the deglacial event history and influence many aspects of modern society. The deposits are farmed extensively, mined for aggregate, and used as a substrate for waste disposal. Buried eskers host abundant supplies of potable groundwater. The fine-grained Champlain Sea sediments are prone to rapid retrogressive slope failure, settlement under loading, and amplified shaking during earthquakes.

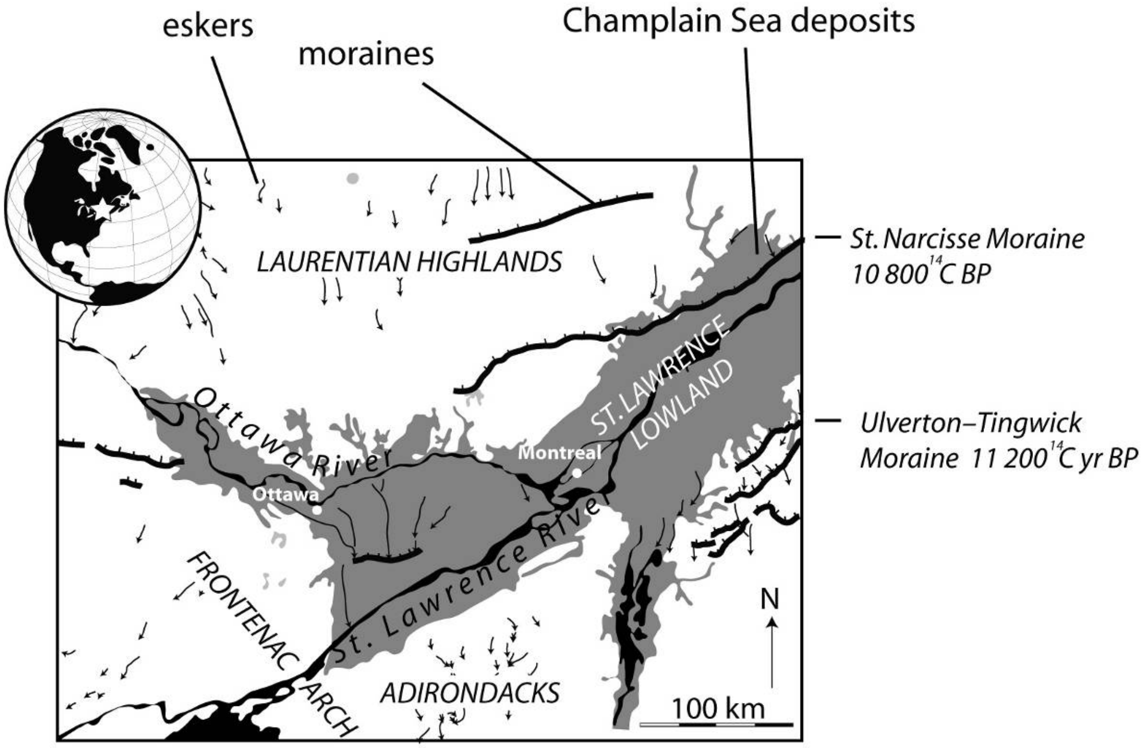


Figure 1. The Champlain Sea basin, a glaciomarine-mud depocenter that existed for a short period of time following deglaciation in the isostatically rebounding St. Lawrence Lowland. Aerial extent of fossiliferous Champlain Sea deposits (grey, mostly mud) is from Gadd et al., (1993). Eskers and moraines are from Parent and Occhietti (1988), Barnett (1988), Gorrell (1991), and Simard et al (2003). The diamicton ridge (grounding line moraine) southeast of Ottawa (black east-west line) along which several eskers terminate is a newly identified feature (Cummings et al., 2011).

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, will touch upon key controversies surrounding the deglacial event history of the basin, and will highlight aspects of the local natural hazards. Fundamental hypotheses on the 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.

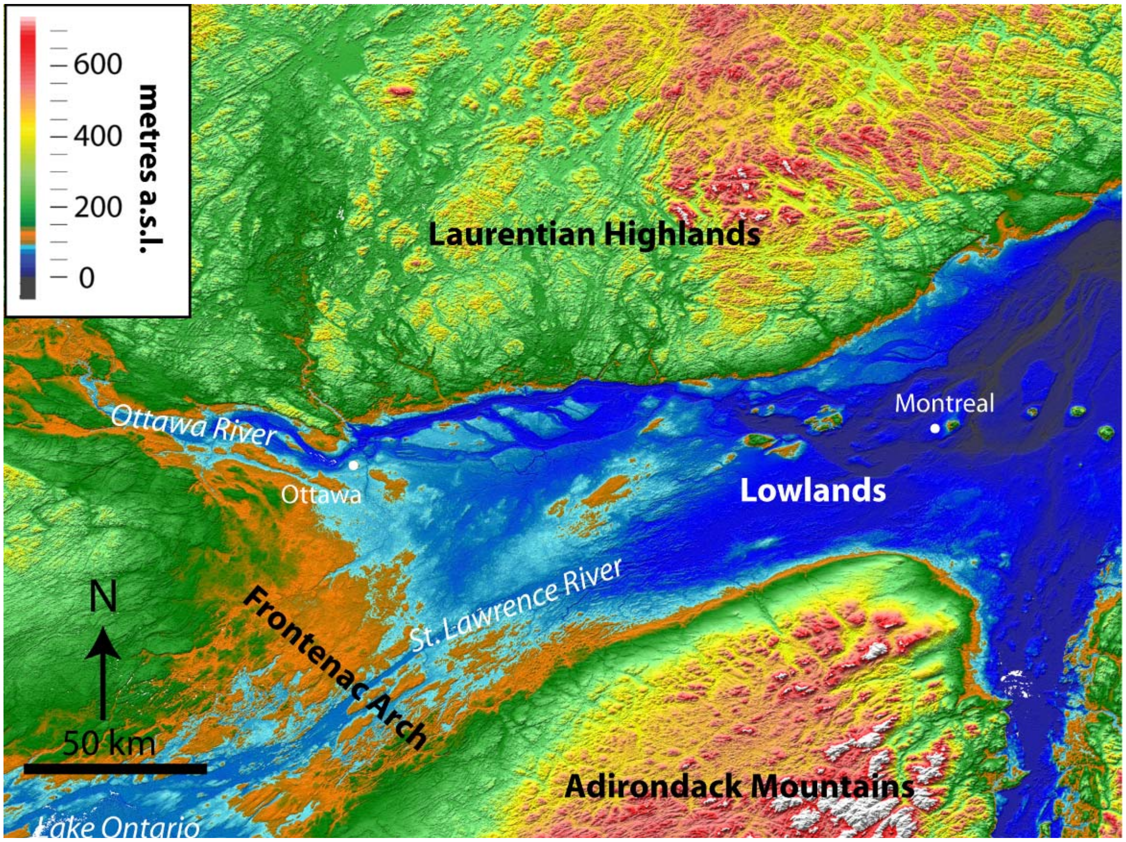


Figure 2. Landscape around the Champlain Sea basin, eastern Ontario and western Quebec. Uplands consist of Precambrian igneous and metamorphic rock covered by a veneer of sandy, carbonate-poor till. Sediments are much thicker in the Lowland (average ~ 10 metres, locally > 170 metres), due primarily to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

Geological Setting and Previous Work

Don Cummings[1]

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

[1] Cummings, D.I., 2011. Geological Setting and Previous Work; *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 02-05. Geological Survey of Canada, Open File 6947.

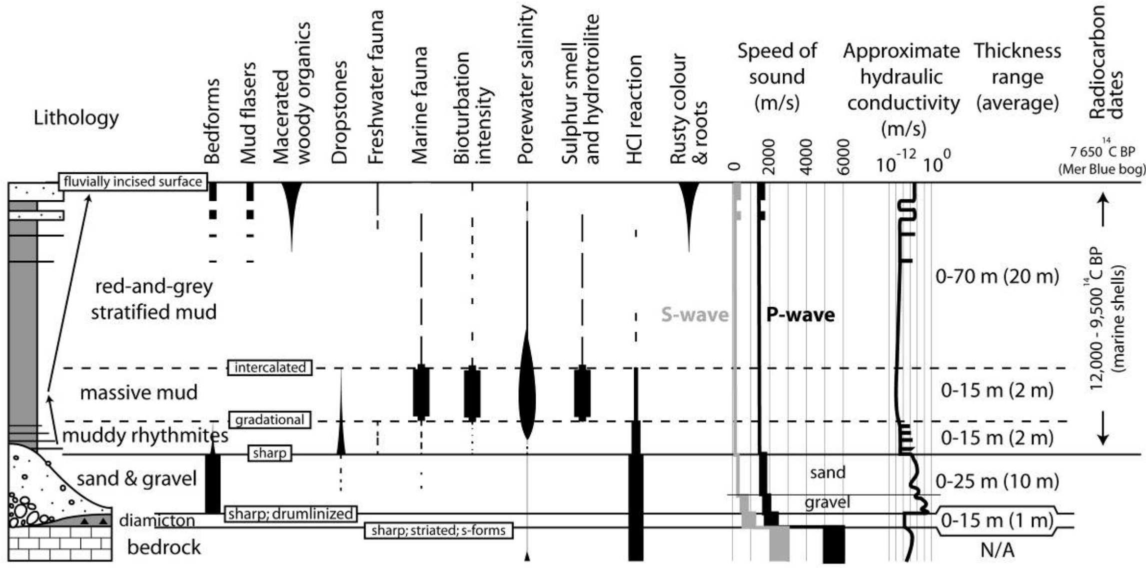


Figure 3. Physical, chemical and biological attributes of Quaternary strata in the St. Lawrence Lowland near Ottawa (idealized) (Cummings et al., (2011) and references therein).

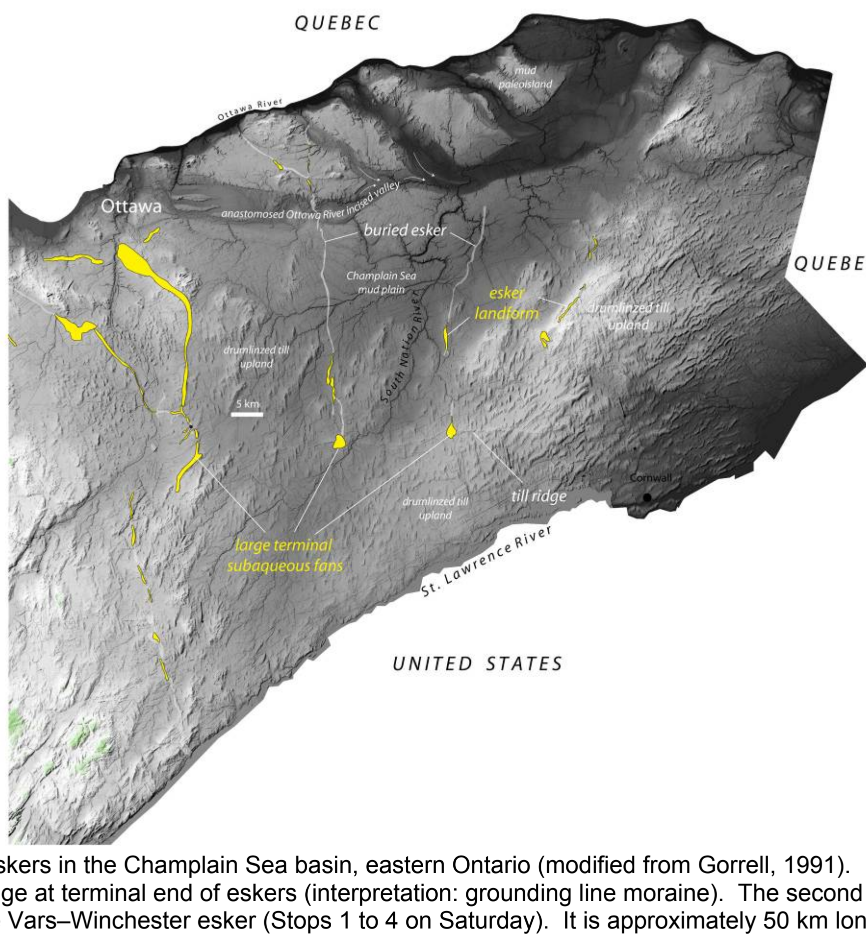


Figure 4. Eskers in the Champlain Sea basin, eastern Ontario (modified from Gorrell, 1991). Note the diamicton ridge at terminal end of eskers (interpretation: grounding line moraine). The second esker from the left is the Vars–Winchester esker (Stops 1 to 4 on Saturday). It is approximately 50 km long.

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), Catto et al. (1982), and Shilts (1994) 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).

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 are 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 (e.g., Shreve, 1985b; Hooke and Fastook, 2007).

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 stratigraphy in the glaciated, mud-rich Champlain Sea basin?

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?

Geotechnical Characteristics of Champlain Sea Deposits

***Gregory R. Brooks and Jan M. Aylsworth*[2]**

The glaciomarine and prodelta silt, clayey silt, silty clay and clay sediments in the Ottawa Valley–St. Lawrence Lowlands region collectively form the Champlain Sea deposits, which are also known informally as 'Leda Clay' (see Gadd, 1975). Typically, the deposits consist of silt- and clay-sized grains that are predominantly composed of quartz, feldspar, amphibole, illite and chlorite minerals, which represent glacial rock flour rather than clay minerals (see Torrance, 1988). These fine-grained particles flocculated within the saline-to-brackish water of the Champlain Sea, producing a sediment structure with a porous, loose structural framework that can retain a high natural moisture content.

Champlain Sea deposits can be geotechnically 'sensitive', a condition whereby the remoulded shear strength is significantly lower than the undisturbed shear strength. In cases where remoulding yields a very low residual strength, the disturbed sediment may liquefy and flow because the collapse of the loose structural framework. The sensitivity of sediment is defined quantitatively by the ratio of undisturbed to remoulded shear strength (at the same moisture content). A sensitivity exceeding a ratio of 30 denotes a 'sensitive' clay, while a 'quick' clay exceeds 50 (Broms and Stal, 1980). In the National Capital Region (NCR), sensitivities commonly range from 20 to 100 (Eden et al., 1971), but locally values of up to 175 have been recorded (see Mitchell and Markell, 1974). Sensitivity generally relates to the high natural water content in the deposits, flocculated fabric of the fine-grained particles, low electrical attraction between the particles, low overburden pressures during deposition, and post-depositional leaching of salts from the clays (Torrance, 1988; Carson and Bovis, 1989). The latter factor is a particularly important factor controlling the development of sensitivity and generally conditions the presence of sediment sensitivity at a given location.

Sensitive 'Clay' Earth Flows

Failure in Champlain Sea sediments can generate large-scale, retrogressive earth flows, lateral spreads, or some combination of the two. Modern earth flows in the region have consumed up to 40 hectares of land (see Michell and Markell, 1974), but there are examples of prehistoric scars that are much larger (see Aylsworth et al., 2000). Liquified earth flow spoil can flow along a valley bottom in excess of a thousand metres, rafting intact blocks of soil, vegetation and anything else engulfed in the failure. For example, spoil from the 1993 Lemieux landslide flowed ~1600 m both upstream and downstream along the valley of the South Nation River (Evans and Brooks, 1994). When a failure occurs along a relatively narrow valley, the spoil will obstruct drainage and cause flooding upstream. Residual impoundments from these landslide dams can persist for months and even several years after the spoil is overtopped.

Earth flow failures can be triggered, for example, by river erosion of the toe of the slope; loading of the top of the slope by a perched water table; elevated porewater pressures associated with the rapid drawdown of the water table during spring runoff; vibrations caused by nearby construction activity; and seismic shaking (Crawford, 1961; Mollard, 1977). As an illustration, the most recent example of a significant earth flow in the area at the time of writing occurred near Notre-Dame-de-la-Salette, north of Buckingham, and was triggered by the Mw 5.0 Valle de Bois earthquake of June 23, 2010. The epicenter of this earthquake was located only ~14 km from the landslide (see Perret, this volume). With many other failures, however, the specific triggering mechanism *per se* is speculative and commonly is the net effect of several contributing factors.

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[2] Brooks, G.R., and Aylsworth J.M., 2011. Geotechnical characteristics of Champlain Sea Deposits; *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 06-07. Geological Survey of Canada, Open File 6947.

Over 250 earth flows scars and associated deposits have been identified in the NCR (see Fransham et al., 1976; Fransham and Gadd, 1977). They occur along river valleys of varying scale as well as the margins of paleochannels and terraces of the Ottawa River. The earth flows occur in relative isolation, within distinct clusters, and within complexes of coalesced failures. The majority of the failures are prehistoric and of unknown age. Dating of 15 large-scale earthflow features along the Ottawa River paleochannels in the areas of Hammond, Bourget, Rockland, and Plantagenet, east of Ottawa, revealed a common age of about 4.5 ka BP which has been interpreted to be indicative of the occurrence of a prehistoric seismic event (Aylsworth at al., 2000; see below). The most significant historic earth flow in terms of consequences occurred in the 1908, causing thirty-three deaths at the village of Notre-Dame-de-la-Salette, Quebec (see Lawrence and Brooks, this volume).

Earthquake History of the Western Quebec Seismic Zone

Maurice Lamontagne[3]

The Ottawa-Gatineau region is located within the Western Quebec Seismic Zone (WQSZ), which includes parts of eastern Ontario and the Adirondacks Mountains in northern New York State, the Ottawa Valley from Montreal to the Temiscaming area, and the Laurentian Mountains north of the Ottawa River (Basham et al., 1982). The extent of the WQSZ is based on the distribution of measured small magnitude earthquakes as well larger historical events of M 5.5 to 6.2 (Fig. 5; Lamontagne et al., 2008). Most earthquakes in the WQSZ, however, occur within two areas: a northwest-southeast band from Cabonga reservoir in the upper Gatineau River watershed to Montreal and a less active sub-zone underlying the Ottawa–Bonnechere graben (Lamontagne et al., 1994). Earthquake epicentres are located in the mid- to upper crust and represent reverse faulting on pre-existing faults of the Grenville Geological Province. No surface rupture has ever been reported in the WQSZ.

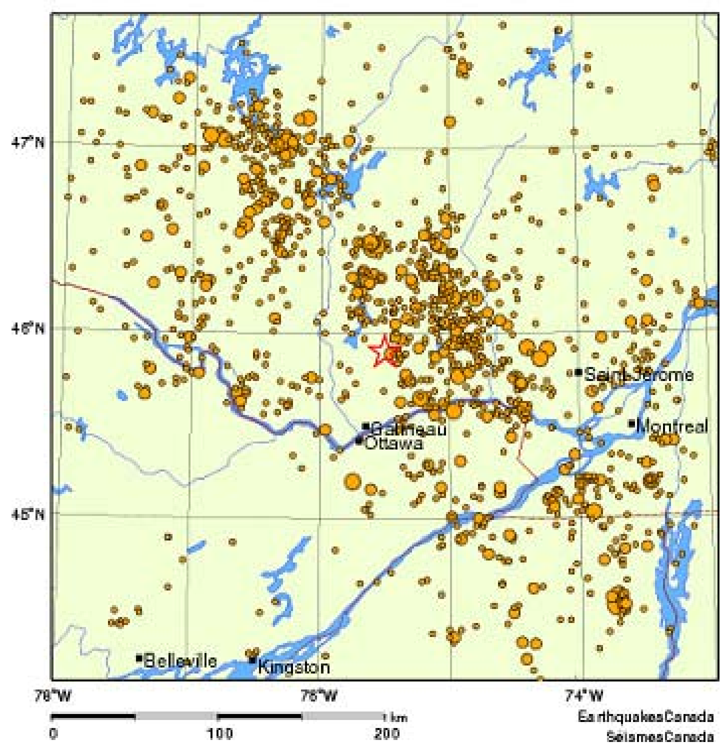

Figure 5. Earthquakes with a magnitude of 2.0 and larger, AD 1980-2010, within the Western Quebec Seismic Zone. The main NW-SE trend of earthquake epicentres extends from the Montreal to Cabonga reservoir areas, while a less active sub-zone follows the Ottawa–Bonnechere graben. The red star marks the location of the June 23, 2010, Val-des-Bois earthquake (Courtesy of Stephen Halchuk, NRCan).

Current knowledge of the earthquake activity in the WQSZ is based on less than 400 years of reported felt events and approximately 100 years of instrumental recordings. Historical events

[3] Lamontagne, M., 2011. Earthquake history of the Western Quebec Seismic Zone; *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 08-10. Geological Survey of Canada, Open File 6947.

are known from written accounts (see Lamontagne et al., 2008). One of the largest events within the WQSZ is the AD 1732 Montreal earthquake, estimated at magnitude 5.8, which damaged about 300 houses in the Montreal region. Other significant events include the AD 1935 M 6.2 Timiscaming earthquake and the AD 1944 M 5.8 Cornwall-Massena earthquake.
During its 200 year history, the Ottawa-Gatineau region experienced moderate shaking from earthquakes in the M 5.0 to 6.2 range, located at less than 300 km epicentral distance in AD 1856, 1861, 1908, 1909, 1913, 1914, 1917, 1924, 1933, 1935, 1944, 1983 and 2010 (Fig. 6; Lamontagne, 2010). More distant earthquakes causing moderate shaking in Ottawa occurred in AD 1870, 1924 and 1925 (events in the Charlevoix Seismic Zone at about 450-525 km epicentral distance), in AD 1929 (an event near Attica, N.Y. at about 350 km epicentral distance), and in AD 1988 (an event in the Saguenay region at about 470 km epicentral distance; Fig. 7). Of these earthquakes, fourteen events reached or exceeded the Modified Mercalli Intensity (MMI) V level (e.g., some dishes and windows broken) between AD 1830 and 2010.

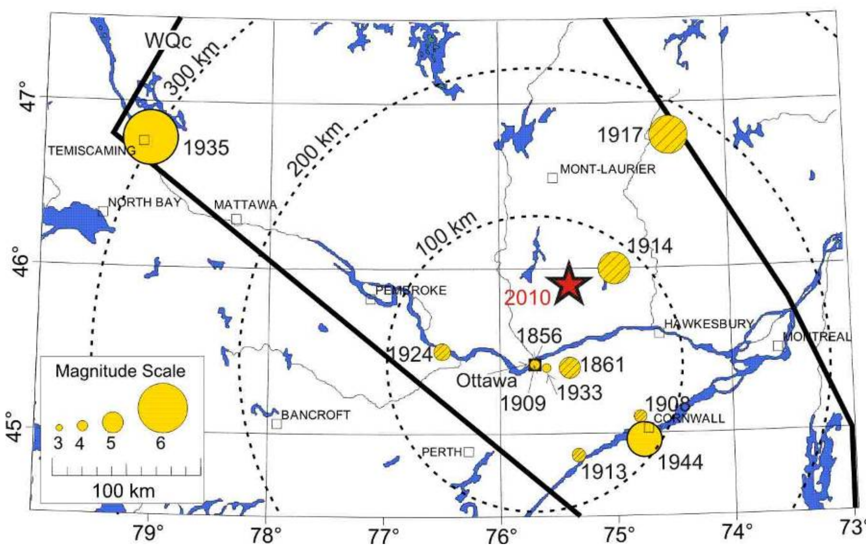

Figure 6. Location of the epicentres of the Val-des-Bois earthquake (star) and of historical earthquakes that had caused some damage in the Ottawa-Gatineau region. Shaded circles represent earthquakes with approximate epicentres and magnitudes.

Historical accounts, however, indicate that certain parts of the city have experienced higher levels of ground motion than others during the larger earthquakes. For example, Lamontagne et al. (2008) found that most reported examples of shaking in the MMI VI to VII range (e.g., chimneys broken, instances of fallen plaster) are centered in the immediate area of the Canadian Museum of Nature, which overlies a bedrock basin containing soft Champlain Sea sediments 10 to 50 m thick (see Crow et al., this volume). In contrast, areas underlain with a thin veneer of unconsolidated deposits over bedrock, such as Parliament Hill and some portions of the Hull area of Gatineau, had almost no occurrences of damage. Most recently, an earthquake of M 5.0 occurred near Valle de Bois, Quebec, on June 23, 2010, the epicenter of which was located 60 km from downtown Ottawa (see Lamontagne, this volume). During the Valle de Bois earthquake, MMI of up to VII were experienced in the NCR, based on felt reports submitted to the earthquakescanada.ca website (Halchuk, 2010). The felt report intensities were found to be generally higher on areas of soft soil than bedrock (Pal and Atkinson, in press).

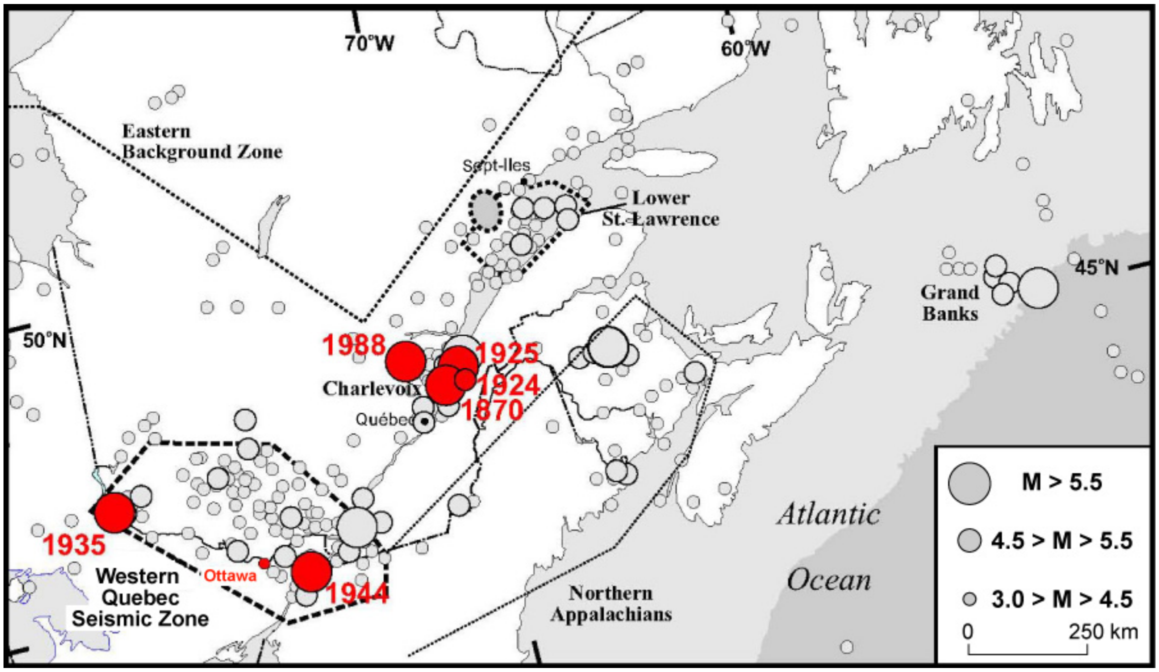

Figure 7. Map depicting 19th and 20th century seismicity in eastern Canada (after Lamontagne, 2008). The larger events that had some impact in the Ottawa-Gatineau region are shown in red. Note, the location of the Western Quebec Seismic Zone in the region of southeastern Ontario and southwestern Quebec.

One, and possibly two, significant pre-historical earthquakes have also shaken the area since the regression of the Champlain Sea. Evidence of the younger event is based on a common age of about 4550 yr BP of 15 earth flows, as mentioned above (see Aylsworth et al., 2000). Three large-scale disturbed terrains in the areas of Lefaivre, Treadwell and Wendover, Ontario, that have a common age of about 7000 yr BP are evidence of a possibly second event (see Aylsworth et al., 2000; Aylsworth and Lawrence, 2003); Brooks unpublished data). These undulating terrains exhibit widespread deformation and disruption in the subsurface both underlying alluvial terrace surfaces and in the modern river channel (see Douma and Aylsworth, 2001; Brooks unpublished data). All three disturbed terrains are located over thick suites of soft Champlain Sea sediment which are sites where amplified seismic shaking may have occurred. Magnitudes of these ancient earthquakes probably exceeded M 6.5 (Aylsworth et al., 2000) and possibly exceeded M 7 (Adams and Halchuk, 2004; Aylsworth and Hunter, 2004).

Overall, during historical time, only minor impacts have been caused by earthquakes in the Ottawa area with the damage resulting from moderate events (M 5.0-5.5) within 200 km or larger (M 5.5-6.5) more distant ones. Nevertheless, the Ottawa-Gatineau area ranks third after Vancouver and Montreal for Canadian urban seismic risk, based on its earthquake hazard and population (Adams et al., 2002).

Fieldtrip Stops Saturday, Day One

Stop 1-1A: Geological Overview

Don Cummings[4]

This section summarizes the results of a large-scale subsurface study of Quaternary strata carried out between 2006 and 2008 in the vicinity of the Vars–Winchester esker, a high yield aquifer east of Ottawa (2nd esker from the left in Fig. 4) (Cummings et al., 2011). 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 primarily Paleozoic limestone that is typically buried but 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. 1-1). 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 striae are oriented nearly north-south (Sharpe, 1979). Near Montreal and on the north face of the Adirondacks, a second population of striae that trend northeast-southwest is observed. Most authors argue that north-south striae are related to regional ice-flow during the last glacial-maximum, whereas younger striae 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 striae data, however, is muddled by inconsistent cross-cutting relationships; for example, northeast-southwest striae commonly cross-cut north-south striae, but also locally appear to be cross-cut by them.

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 are identical to forms that 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 impinging on the bed has to be turbulent. Flows in the atmosphere and hydrosphere are

[4] Cummings, D.I., 2011. Stop 1.1A. Geological Overview; *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 11-24. Geological Survey of Canada, Open File 6947.

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.

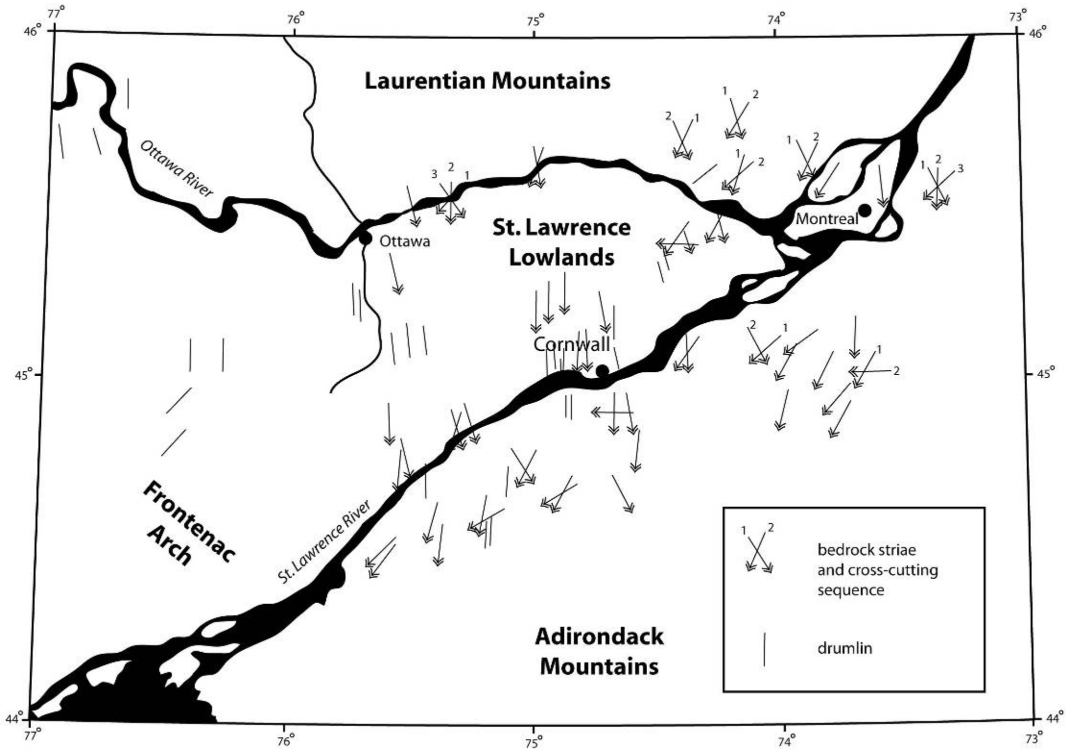

Figure 1-1. Orientations of drumlins and selected striae on bedrock. Modified from Ross et al., (2006).

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. 1-2). 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 pre-existing bedrock unconformity (Wilson, 1903; Amrbose, 1964) near the Ottawa River over multiple glaciations, accentuated the structural grain locally, and generated a complex, composite, irregular erosional surface.

Diamicton (till).

In wells, stiff diamicton (average 1–3 m thick) was intersected locally above bedrock (Figs. 7, 8). 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. 1-5). 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.

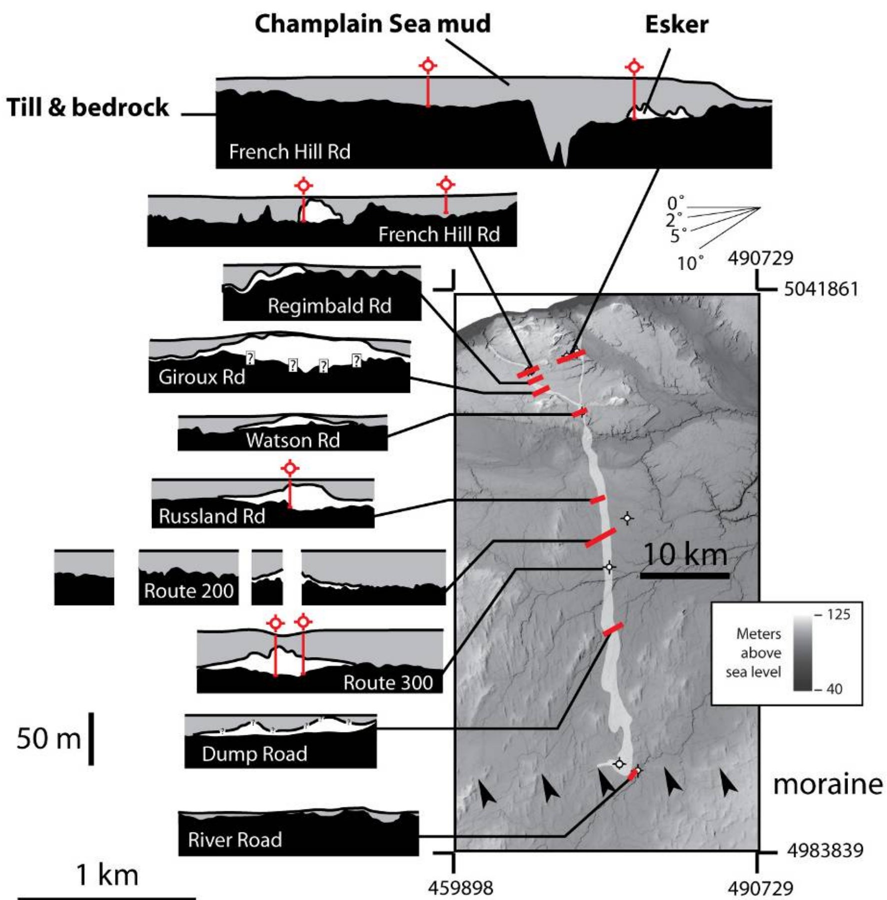

Figure 1-2. Interpreted seismic transects from the Vars–Winchester esker. Data were collected, processed, and interpreted by André Pugin and Susan Pullan..

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 km 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 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).

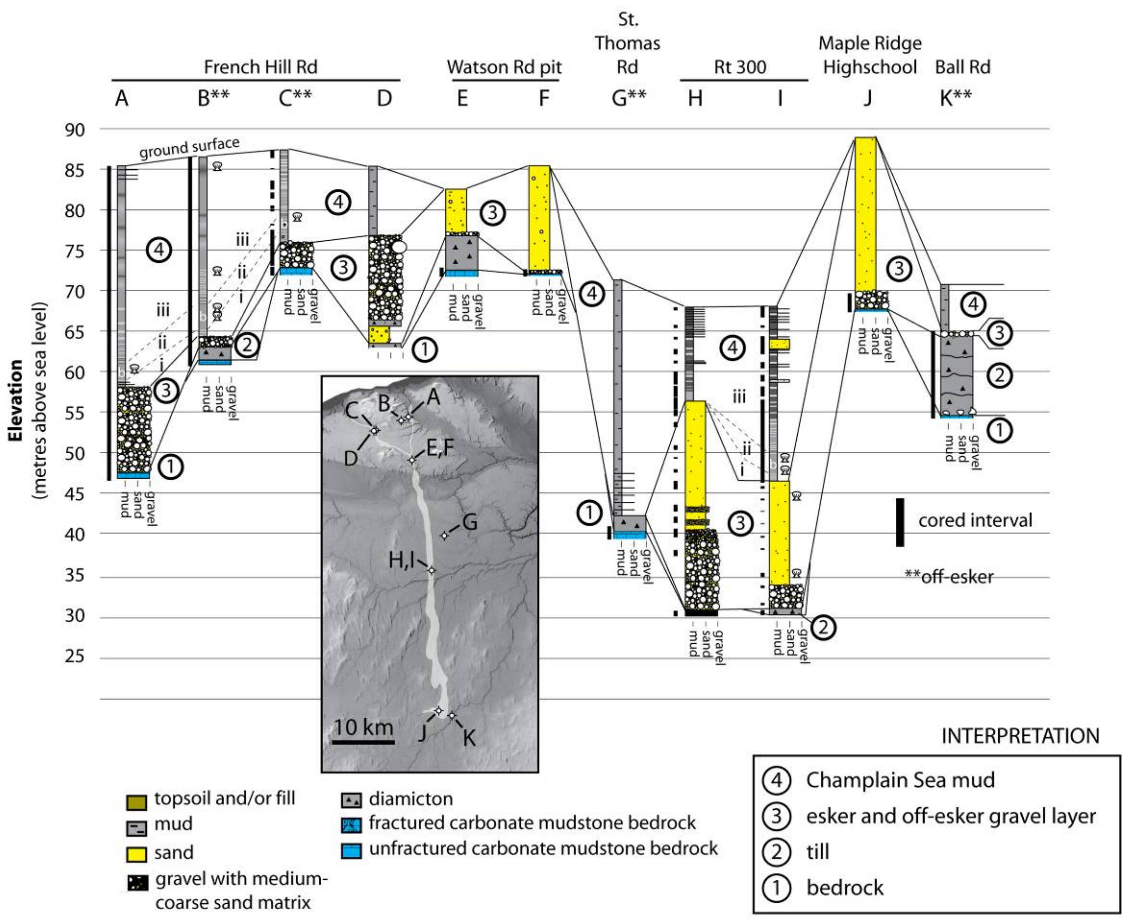

Figure 1-3. Wells drilled in the vicinity of the Vars–Winchester esker between 2006 and 2007. i = grey carbonate-rich rhythmites; ii = massive bioturbated mud; iii = rhythmically stratified, carbonate-poor red and grey mud; b = bioturbation. Shells indicate locations where *Portlandia arctica* shells were observed.

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.

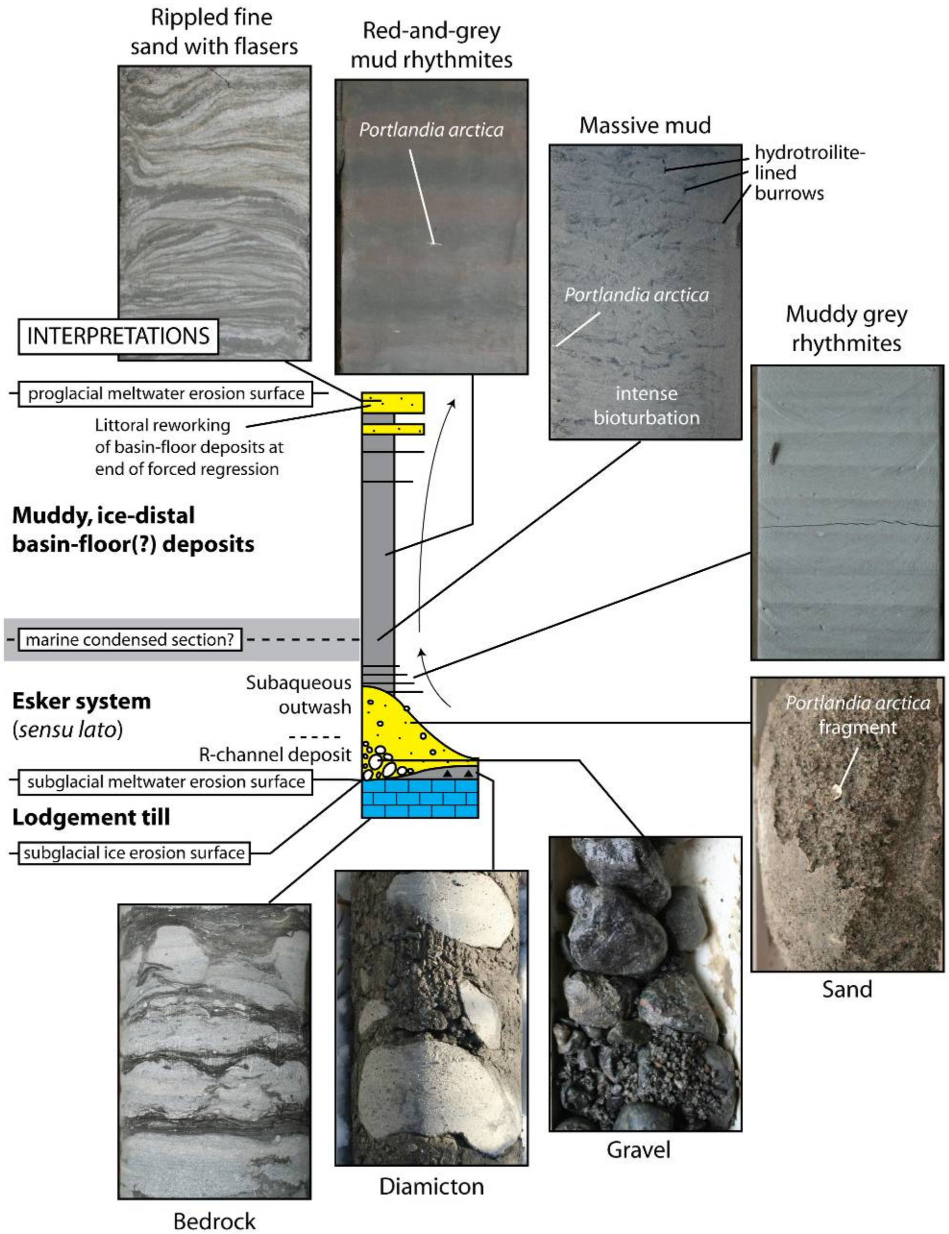

Figure 1-4. Representative photos of sediment facies observed in core and interpreted depositional environments.

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 (e.g., Sharpe, 1979) 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; Cummings et al., 2011), 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). It is of note that similar, permeable off-esker gravel layers exist elsewhere in the Champlain Sea basin (Ross et al., 2006), in addition to mud-rich glaciated basins elsewhere (Veillette et al., 2008).

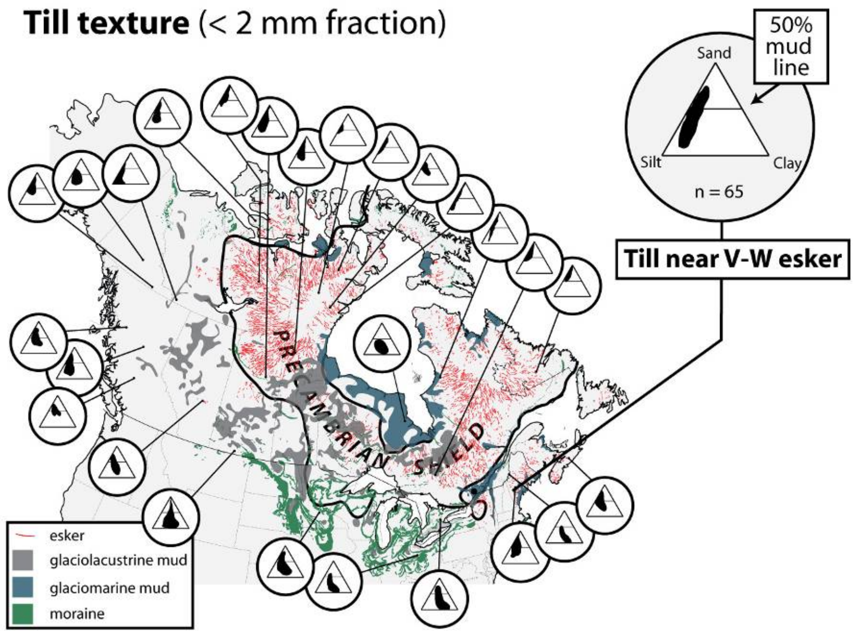

Figure 1-5. Till texture near the Vars–Winchester esker plotted relative to till texture throughout North America. Data from Shepps (1953), Flint et al., (1959), Thompson (1975), Scott (1976), Sharpe (1979, 1992) and references therein, Clague (1984), Dredge et al., (1986), Dyke and Prest (1987), Dyke (1990), Weddle (1991) and references therein, Jackson (1994), 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).

Sand and gravel (esker).

Subsurface data suggest that the Vars–Winchester esker consists of two key elements, a **gravelly central ridge** and a **sandy fan carapace** (Fig. 1-5, 1-6). 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 3).

The gravelly central ridge of the Vars–Winchester esker is 2 to 20 m high (average 15 m) and 100 to 200 m 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. 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).

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. 1-6). 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[5] and that the R-channel may have been several 10s of metres high[6]. This in turn suggests discharges of 1000s of cubic metres per second[7]. 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 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. 1-5). 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.

[5] Calculated using the Helley method (Table 3 in Costa, 1983).
[6] Calculated using the method outlined in Leclair and Bridge (2001).
[7] Assuming a flow width of 125 m, flow speed of 2 m/s, and a (conservative) flow depth of 20 m.
[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).

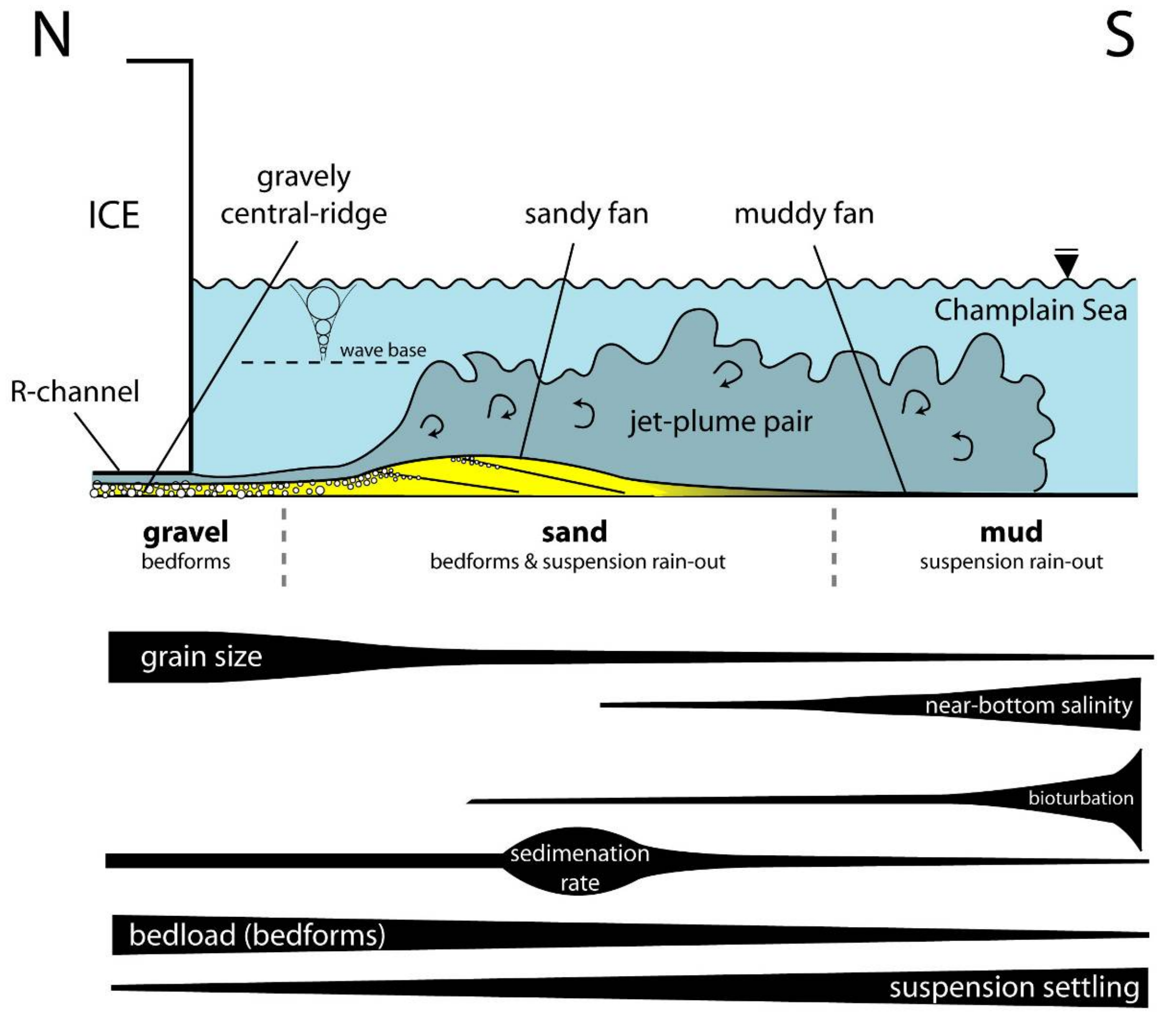

Figure 1-6. Facies model for the Vars–Winchester esker system. The gravelly central ridge is interpreted to have been deposited in a subglacial meltwater channel (R-channel), and the sandy fan carapace is interpreted to have been deposited where the R-channel debouched into deep, standing, proglacial, 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. 1-7), 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 m and 11 m below the top of the sandy fan carapace in one of two long cores collected just north of Embrun (Well I in Fig. 1-3). The esker "container" is clearly resolved in the seismic data at this location (see Stop 3). 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. 1-5). 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?

<u>Mud with minor sand near bottom and/or top (Champlain Sea deposits).</u> 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).

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. 1-4). Couplets tend to be sharp-

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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).

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 m 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. 1-4). 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 odour. 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. 1-4). 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. 1-7D). It is possible that one or more rapid meltwater-events may have punctuated this gradual, ice-mediated sedimentation pattern (e.g., Shilts, 1994).

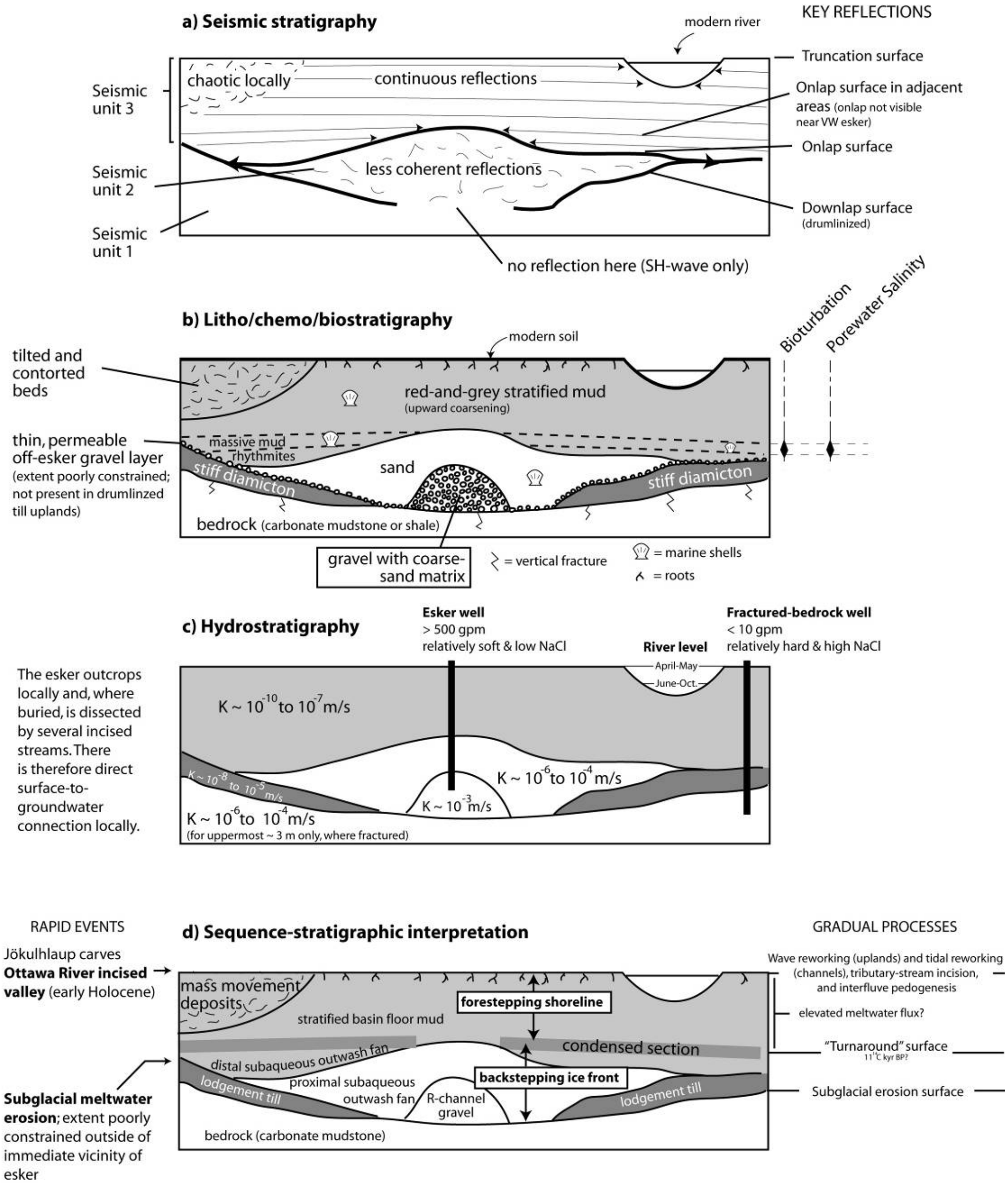

Figure 1-7. Stratigraphy near the Vars–Winchester esker, and sequence stratigraphic interpretation.

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 backstepping is the carbonate-rich upward-fining succession that starts with the esker gravel and ends with the massive bioturbated mud. Backstepping 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 upward-coarsening and upward-thickening 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 upward-coarsening and upward-thickening 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.

<u>Top of Champlain Sea deposits (modern landscape).</u> 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 River, by contrast, is not incised upstream of Montreal.) The Ottawa River incised valley is 15–30 m deep, 5–20 km 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

[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.

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 m 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. 1-8). 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.

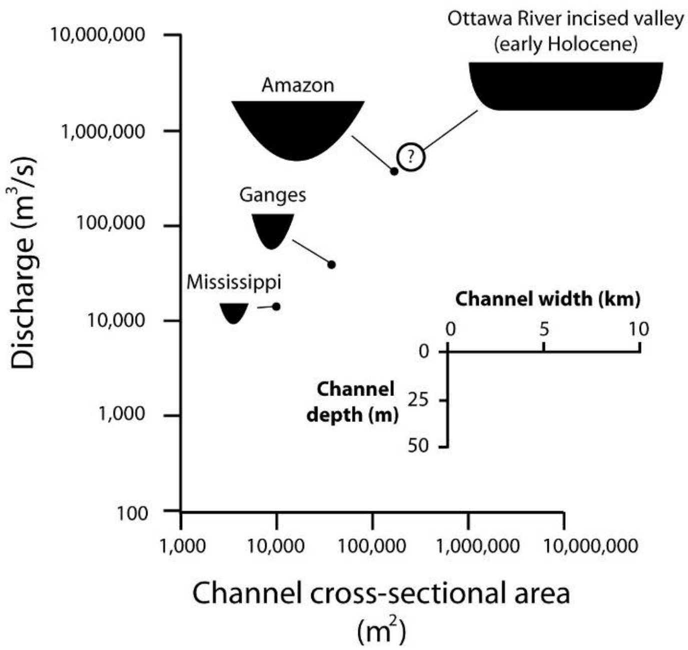

Figure 1-8. Simplified cross-sections of large modern rivers versus that of the Ottawa River incised valley.

Stop 1-1B: Seismic Section French Hill Rd East

André Pugin and Susan E. Pullan[17]

Objective.

Stop 1B provides an overview of shallow shear-wave (SH-wave) reflection seismic profiling and its application in the South Nation study. Geophysical data were collected to better delineate the location and architecture of buried segments of the Vars–Winchester esker. Other interesting features, such as the buried valley beneath this stop, were also discovered by this work. These shear wave reflection surveys were some of the first production surveys carried out with the GSC minivib/landstreamer data acquisition system.

Setting overview.

The study area consists of an upland region of bedrock outcrop, low relief till mounds, faint esker relief and extensive low relief muddy 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. This northern portion of the study area is essentially flat terrain with little geomorphic expression.

Description.

This stop is located on a ~2750 m long, east–west seismic line that images the underlying bedrock topography as well as a number of seismic facies within the overlying surficial sediment secession (Fig. 1-9). Boreholes drilled on the section (Wells A and B in Fig. 1-3) 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. 1-9).

<u>Bedrock signature.</u> The bedrock interface is evident on the seismic sections as a strong high-amplitude reflection. It 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. In the deepest part of the bedrock valley the bedrock SH-wave reflection is weak to absent.

<u>Till on bedrock.</u> Overlying the limestone bedrock along much of the seismic line is a boulder lag and/or a 2–3 m thick till layer. The boulder lag is interpreted on the basis of parabolic refractions, distinct features observed on 'migrated envelope' representations of the seismic sections, and the existence of a boulder field exposed on bedrock in the area. Strong parallel reflections are interpreted to be produced from the top of till and the top of the underlying limestone bedrock.

<u>Valley fill.</u> 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).

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[17] Pugin, A., and Pullan, S.E. 2011. Stop 1.1B. Seismic Section French Hill Rd East;. *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 25-27. Geological Survey of Canada, Open File 6947.

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. This increase in velocity and an associated increase in relative density has also been observed in geophysical borehole logs.

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. 1-9). 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, one of which will be visited at the subsequent stop.

Additional Reading:

Pugin, André J-M., Pullan, Susan E., and Hunter, James A. 2009. Multicomponent high-resolution seismic reflection profiling. The Leading Edge, **28**, 1248-1261.

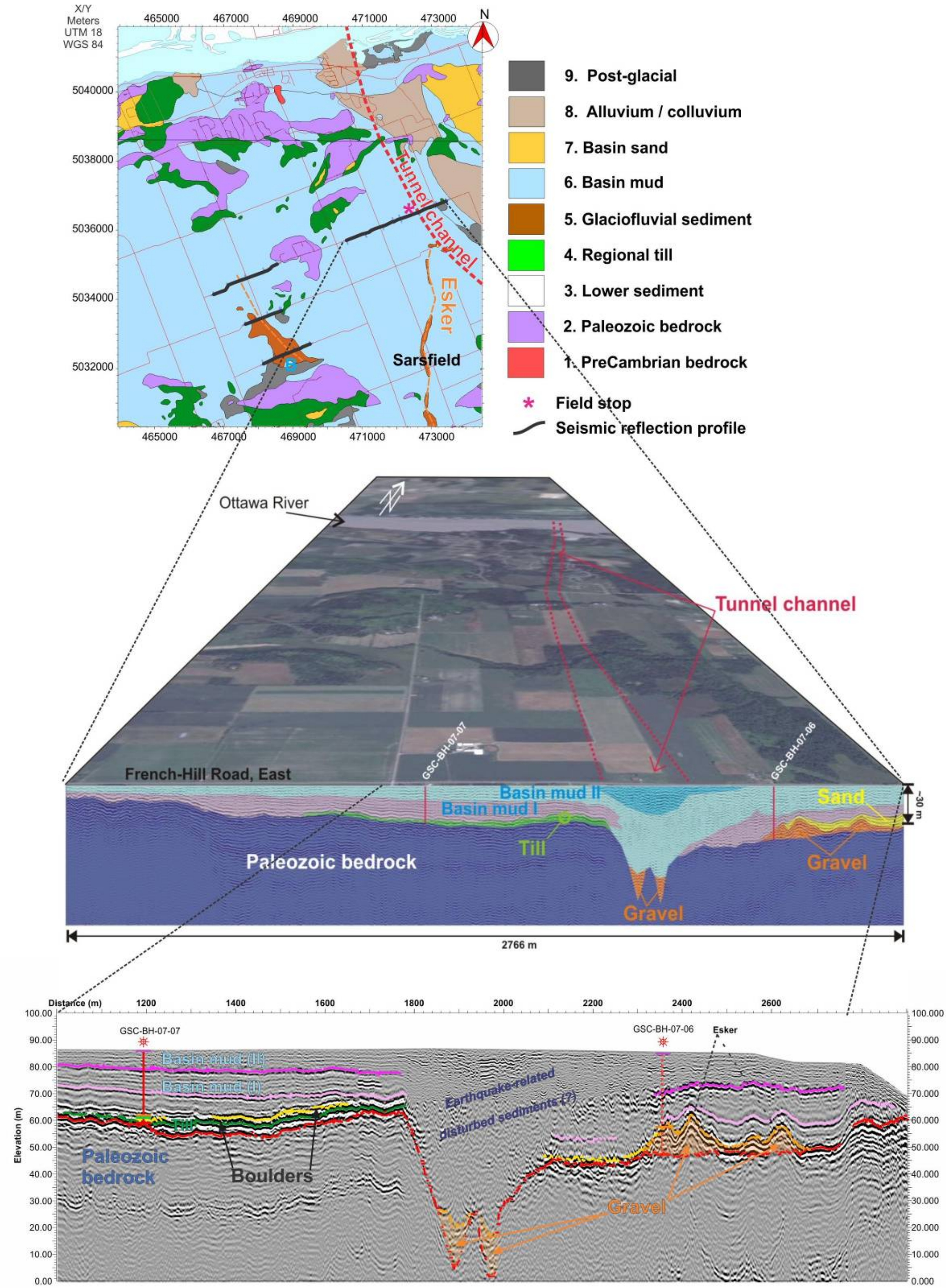

Figure 1-9. Upper panel: Surficial geology map showing location of field stop on French Hill Road and the seismic sections acquired in the area. Middle panel: Satellite image plotted with interpreted seismic section as subsurface corss section. Lower panel: Portion of the shear wave seismic reflection section that images the buried bedrock valley and the blanketing marine sediment that forms a low relief landscape.

Stop 1-2A: Regimbald Road Pit

Don Cummings[18]

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. 1-10). 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 m 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. 1-11) and on a metre-scale (several reactivation-surface bound, downflow-fining packages of cross-strata within a single cross-set).

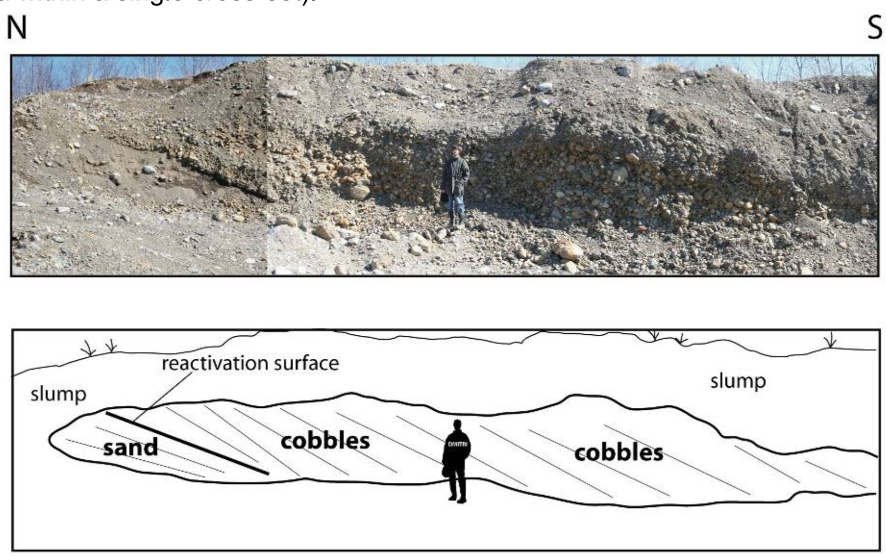

Figure 1-10. One of several thick, stacked, high-angle cross-sets exposed in the Regimbald Road pit. See field trip stop map at start of guidebook for pit location. Note the reactivation surface separating sandy cross-strata from cobble cross-strata. Cross-strata dip roughly southward, parallel to the long-axis of the esker. Dmitri "Kung-fu" Ponomarenko for scale.

[18] Cummings, D.I., 2011. Stop 1.2A. Regimbald Road pit; *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 28-29. Geological Survey of Canada, Open File 6947.

Figure 1-11. 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 1-2B: Hydrogeology

Marc J. Hinton, Sam Alpay, and Charles Logan[19]

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

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[19] Hinton, M. J., Alpay, S., and Logan, C., 2011. Stop 1.2B. Hydrogeology; *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 30-35. Geological Survey of Canada, Open File 6947.

permeability than the contact zone aquifer but may be sufficiently permeable, particularly where it is exposed and weathered, to allow for some groundwater recharge.

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; Fig. 1-12). 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 (Fig. 1-13). 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. However, the area of exposed esker is a small proportion of basin area. Areas underlain by the Champlain Sea mud (Area 3) have very low recharge or recharge that only contributes to shallow unconfined flow systems.

Groundwater level monitoring illustrates the distinct nature of groundwater recharge for these three areas (Fig. 1-14). Water level response in an area of exposed esker (*monitoring well* WAT2, Area 2) shows large responses to spring melt and much smaller responses to rainstorms or mid-winter thaws. Monitoring well 2930-FH is located approximately 2 km from a bedrock outcrop and is screened within thin gravel, till and the upper bedrock surface below more than 20 m of Champlain Sea mud. Despite the overlying aquitard and the distance to the bedrock outcrop, the well responds significantly to both spring melt and many smaller recharge events. The larger response and the more rapid decline in comparison to WAT2 is suggestive of a much smaller specific yield which is consistent with recharge from an area of exposed bedrock (Area 1). Monitoring well STTHM is screened in the contact zone below 29 m of Champlain Sea mud and is more distant to any aquifer outcrop or the esker. The well exhibits rapid short term (daily) fluctuations due to barometric pressure changes since it is effectively confined. The longer term seasonal fluctuations are very small and gradual. Therefore, this well is more representative of Area 3 in which there is little hydraulic connection to sources of recharge.

Because eskers, till and bedrock are stratigraphically lower than the Champlain Sea mud, they can only discharge groundwater where the mud is not present. This can occur either above the contact with the mud (Fig. 1-13, arrow A) or where the aquifers are in direct contact with surface water (Fig. 1-13, 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 Fig. 1-13, arrow B) along the South Nation River.

Regional groundwater flow is generally perpendicular to topographic contours in the areas of 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.

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 into and 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).

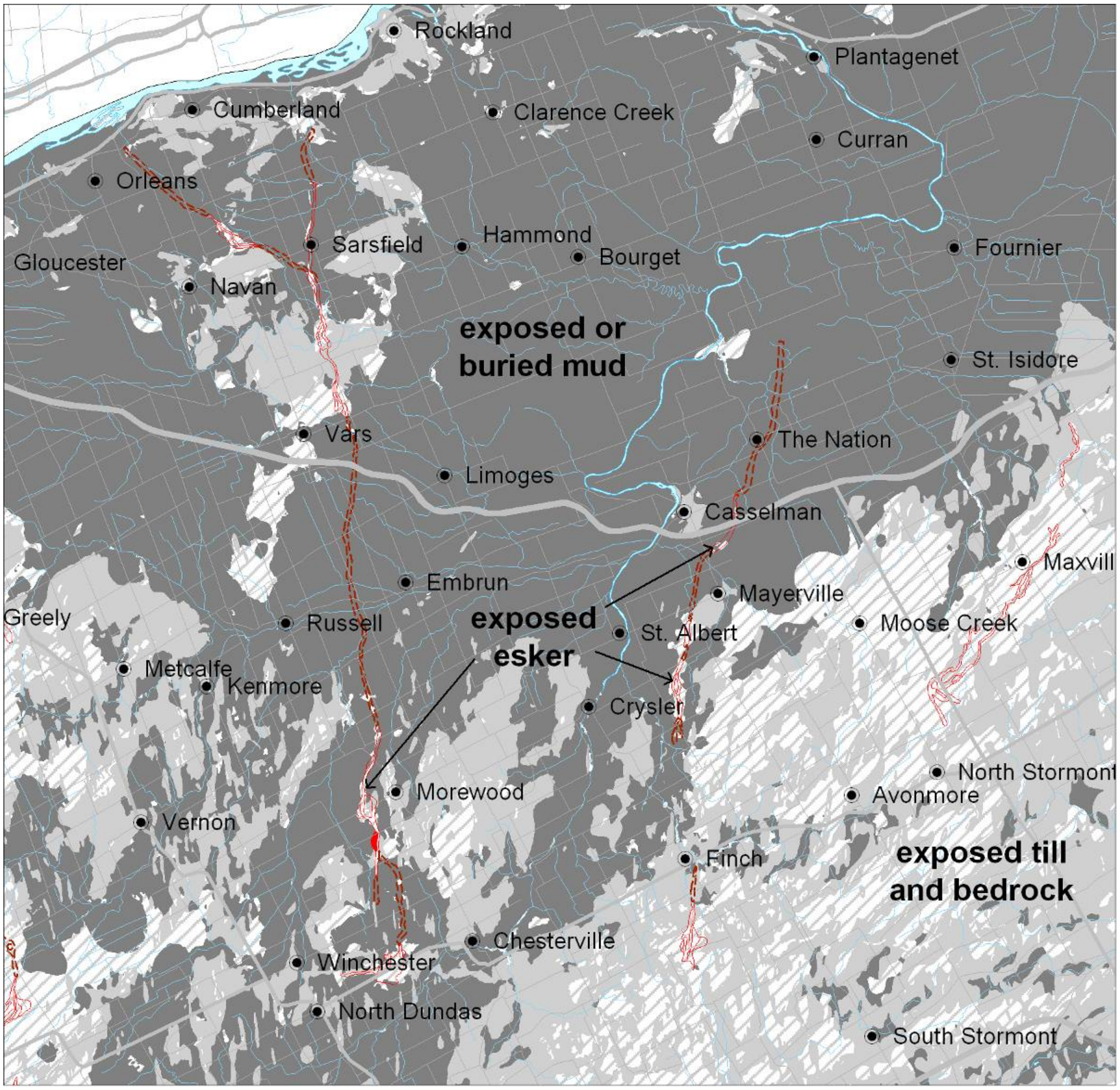

Figure 1-12. Three main hydrogeological settings in the western Champlain Sea basin. The Champlain Sea mud is represented in dark grey. Light grey areas are till and bedrock; cross-hatched light grey areas are till or bedrock overlain by sand. Dashed lines indicate approximate extent of buried eskers; solid lines indicate eskers exposed at the ground surface. Geology modified from OGS.

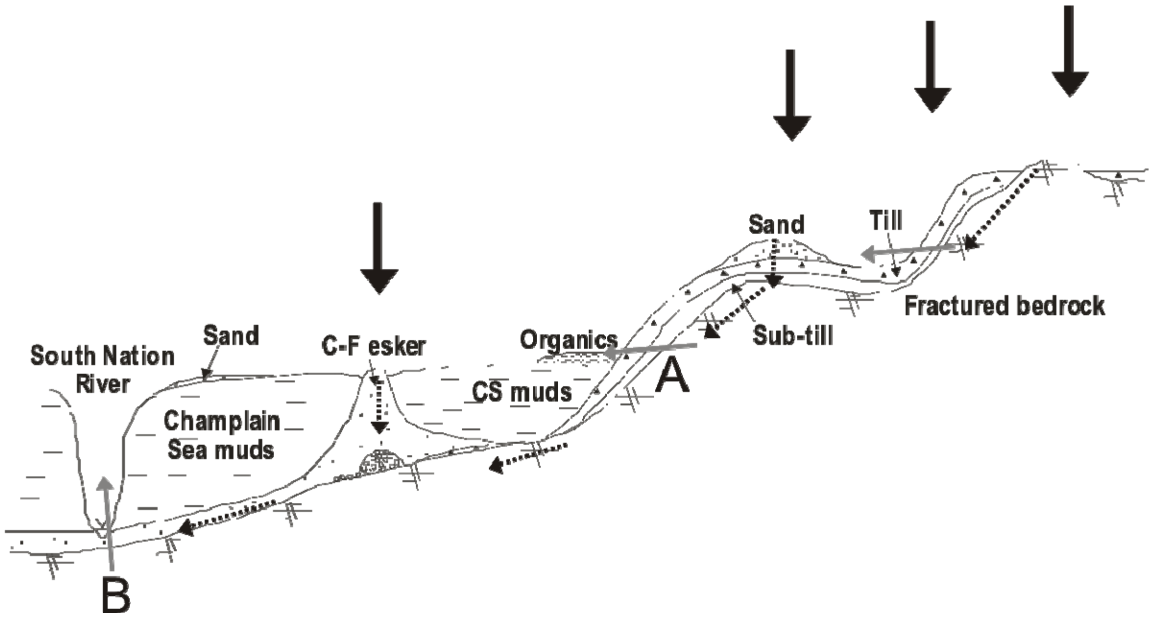

Figure 1-13. Conceptual cross-section across the Crysler–Finch esker. Black arrows indicate recharge areas to the regional flow system, dashed arrows indicate groundwater flow directions and gray arrows indicate discharge areas. Discharge areas A and B are identified in the text. Groundwater flow may 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.

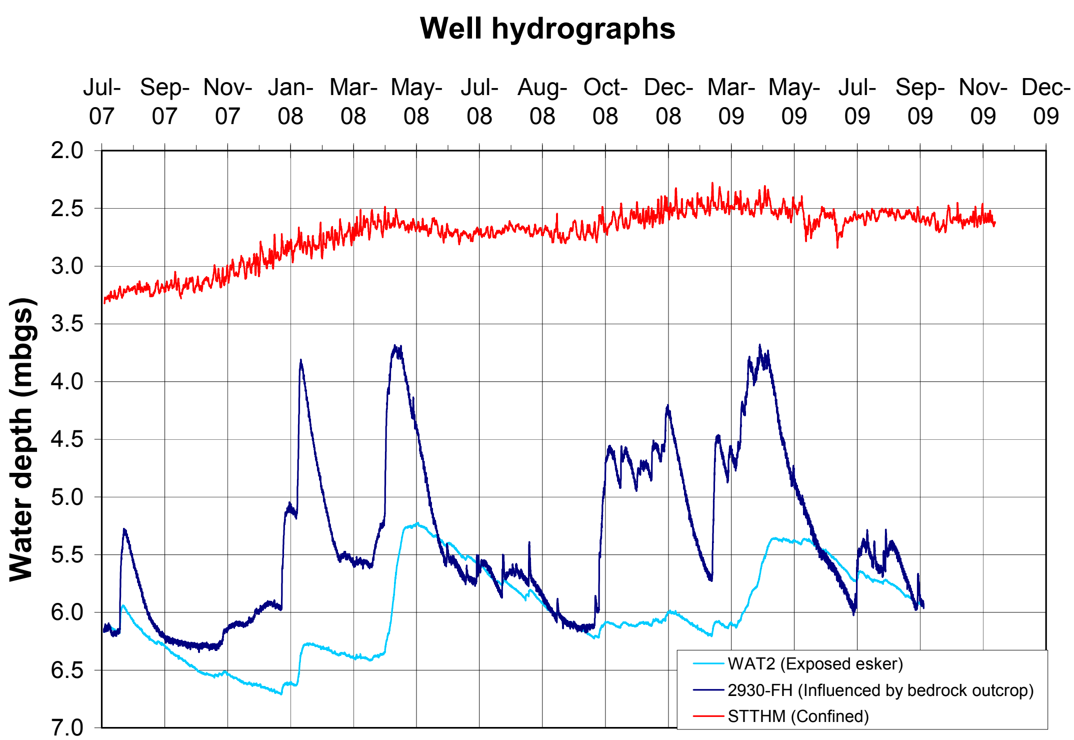

Figure 1-14. Groundwater level fluctuations in three monitoring wells in distinct hydrogeological settings.

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. 1-15). 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.

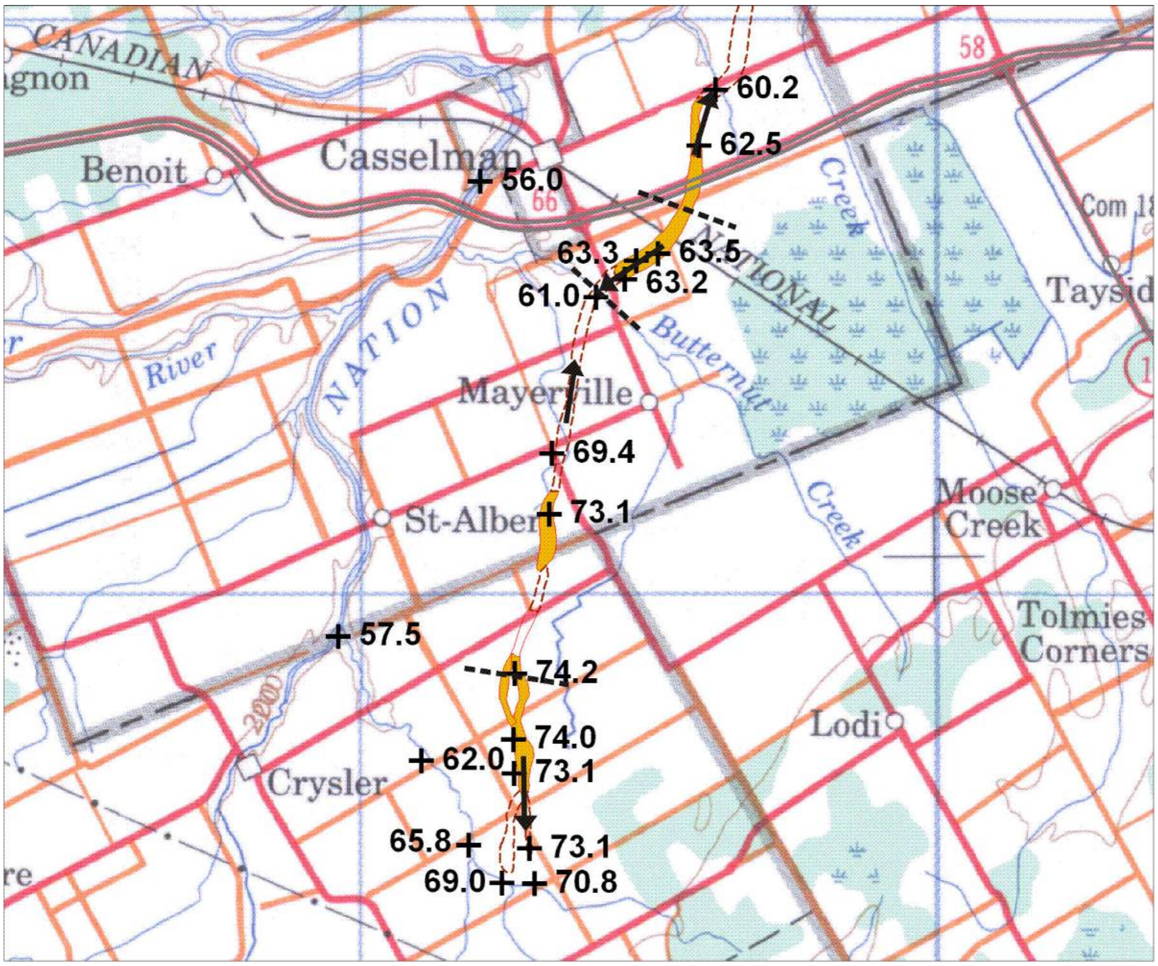

Figure 1-15. Example of <u>segmented</u> flow along the esker. The figure shows the surface water (gravel pits, streams and rivers) and groundwater elevations (in meters) along a portion of the Crysler–Finch esker. Dashed lines indicate the approximate groundwater flow divides along the esker; arrows 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 1-3: Route 300 Seismic Profiles and Cores

Susan E. Pullan, André Pugin, and Don Cummings[20]

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. 1-16), 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 the data acquisition techniques and the minivib-landstreamer system can be found Pullan et al., (2007a,b).

Along this road (Route 300), both P- and SH-wave profiles have been acquired, and both clearly delineate the buried esker and its cross-sectional architecture (Fig. 1-16). Both sections show 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” sediment directly above the esker which suggests that there may be enhanced hydraulic connection between the surface and the esker in this region.

<u>Seismic stratigraphy:</u> 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 m 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. 1-16). 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. 1-16A, and at common mid-point (CMP) 640 - Fig. 1-16B) truncates reflections at the top of this unit. At ~ 10–14 m depth (~ 55 m above sea level), a higher-amplitude reflection is observed within this unit. This reflection parallels overlying and underlying reflections.

<u>Correlation of core and seismic data:</u> The three seismic units described above correlate with shale 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

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[20] Pullan, S.E., Pugin, A, and Cummings, D.I., 2011. Stop 1.3: Route 300 seismic profiles and cores; *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 36-40. Geological Survey of Canada, Open File 6947.

generated by the till or bedrock surface and by the top of the esker. Stiff diamicton (till) was intersected by the borehole 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 provide 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. 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. 1-3). (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 gradationally overlain by the sandy carapace, which is 7 m thick where it overlies the crest of the gravelly central ridge. The gradational contact may help explain the lack of a distinct seismic reflection. Its top contact generates the high-amplitude, mound-shaped seismic reflection. Marine shells (mostly *Portlandia arctica*) are observed 2 m and 11 m 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. 1-3). 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.

<u>Acquisition and processing:</u> Figure 1-17 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. 1-17c).

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. 17b). 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 m asl, while the SH-wave data show significant reduction in reflected signal below coarse-grained units (see Fig. 1-17b, 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. 1-17a, 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. 1-17b, 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.

Additional reading:

Cummings, D.I., Gorrell, G., Guilbault, J.-P., Hunter, J., Logan, C., Ponomarenko, D., Pugin, A., Pullan, S., Russell, H.A.J. and Sharpe, D.R. 2011. Sequence stratigraphy of a glaciated basin fill, with a focus on esker sedimentation. Geological Society of America Bulletin, http://gsabulletin.gsapubs.org/content/early/2011/01/26/B30273.1.abstract

Pugin, André J-M., Pullan, Susan E., Hunter, James A., and Oldenborger, Greg A. 2009. Hydrogeological prospecting using P- and S-wave landstreamer seismic reflection methods. Near Surface Geophysics, **7**, 315-327.

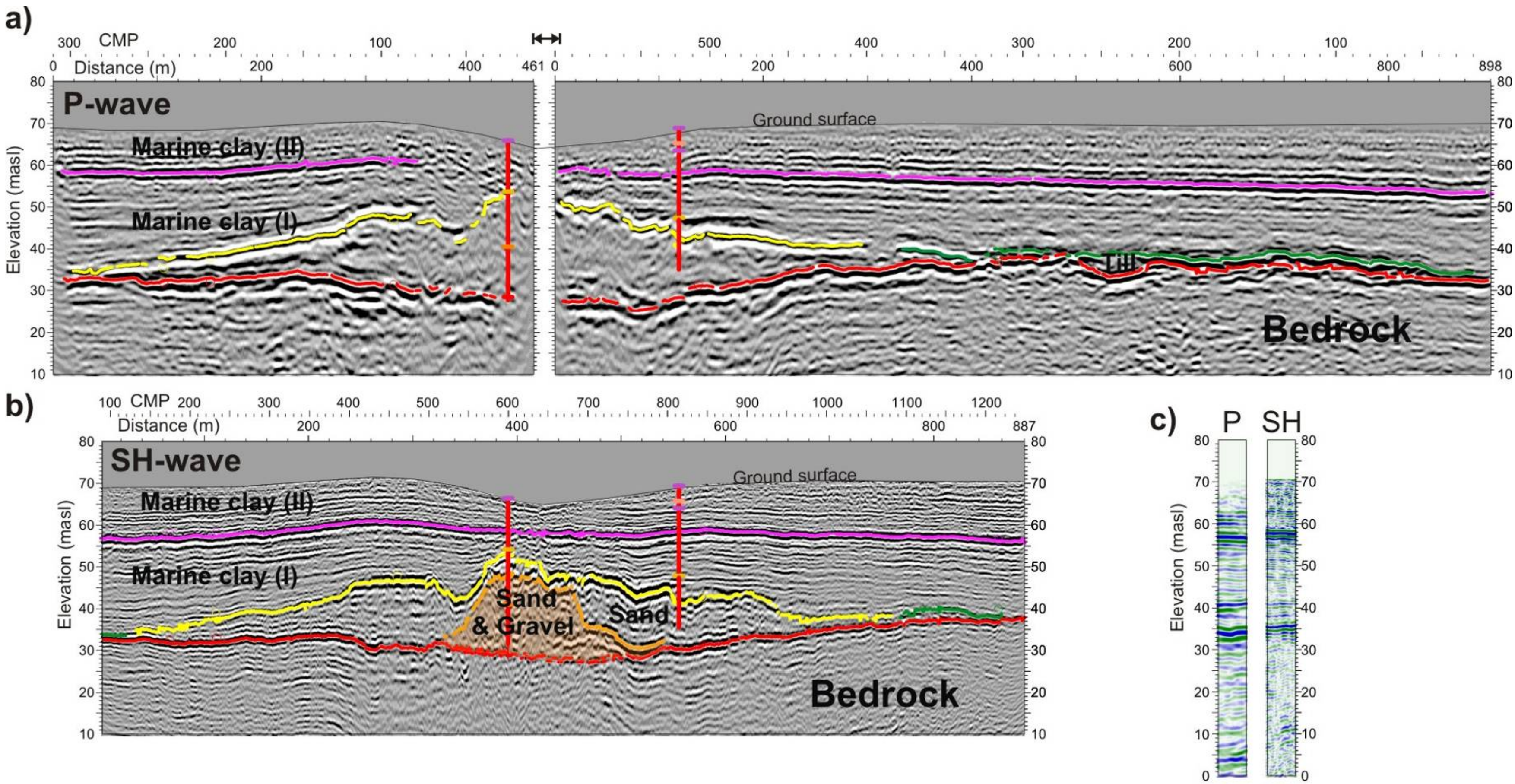

Figure 1-16. Comparison of a) P-wave and b) SH-wave processed seismic reflection sections (vert. exg.= 3x) over buried esker north of Embrun, Eastern Ontario. See discussion in text. The locations of the two boreholes drilled along this line in March 2007 are indicated by the red lines. c) Direct side-by-side comparion of short sections extracted from a) and b) showing higher resolution obtained with SH-waves, particularly in the upper 10–20 m.

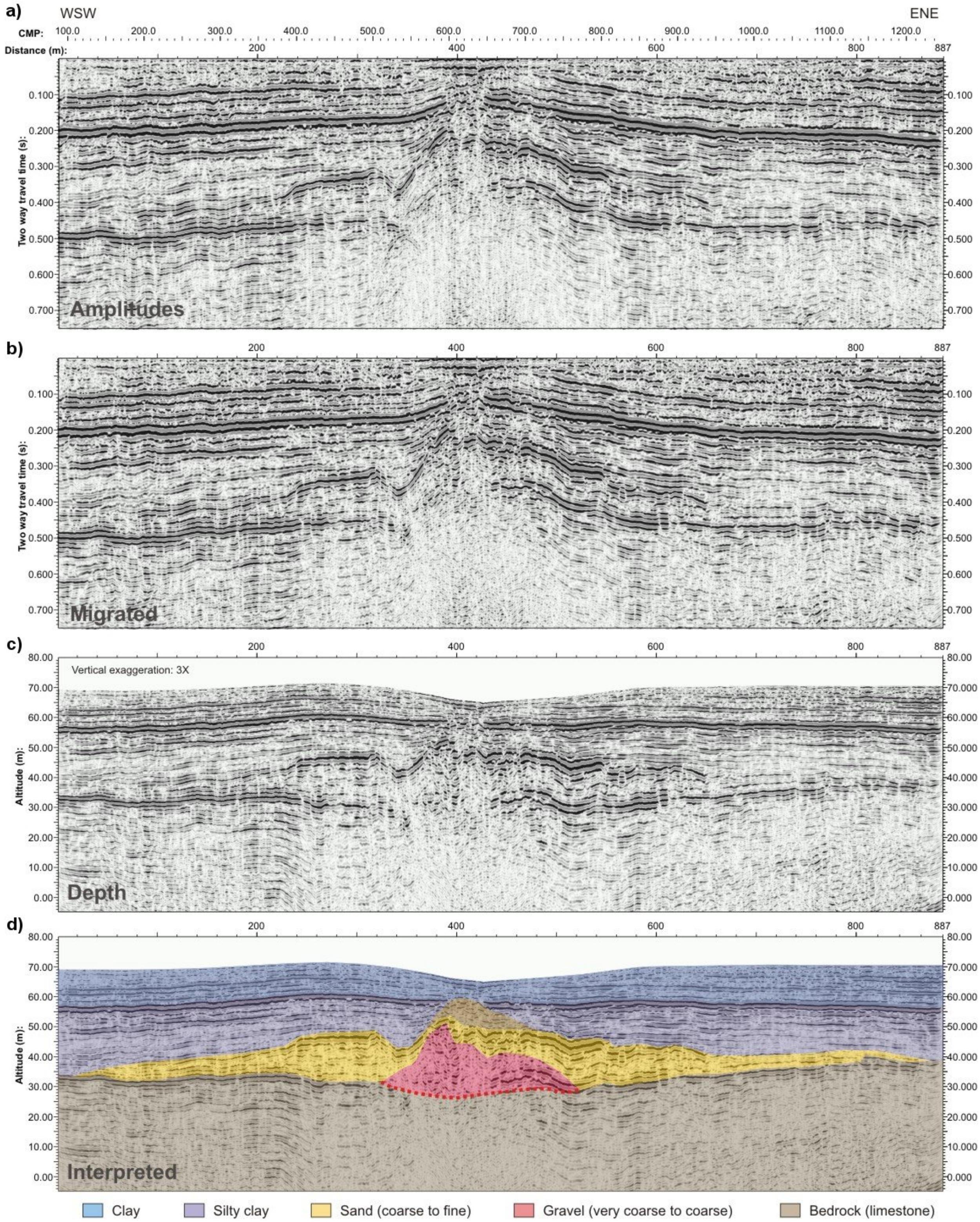

Figure 1-17. Processed SH-wave seismic reflection section over buried esker: a) amplitude section in two-way travel time; b) section a after phase-shift migration; c) section b converted to depth and with topographic corrections (vert. exg.= 3x); d) interpreted section.

Stop 1-4: Kemptville/Loughlin Ridge Esker – Subaqueous Fan

Hazen A. J. Russell and Don I. Cummings[21]

Objective

Stop 1-4 provides an overview of a much larger subaqueous fan deposit associated with the esker west of the Vars–Winchester esker, and provides insight into the vertical succession and lateral continuity of the sedimentary facies. Deposits at the two sites (Lafarge pit and Tackaberry pit) demonstrate the contrast in sediment caliber and hence hydraulic conductivity between the proximal fan setting and the more distal mid fan. The fan at this site is likely an analogue for the similar sized but less extensively exposed deposit at the southern terminus of the Vars–Winchester esker at Maple Ridge (Cummings et al., 2011).

Setting overview

The ridge is mapped as ice-contact stratified drift with flanking littoral sand (Richard, 1974). The deposit is part of an esker extending from the St Lawrence River northward to the Ottawa International Airport. Parts of this esker have been studied by investigators interested in esker – subaqueous fan sedimentology (Rust and Romanelli, 1975, Rust, 1977, 1988; Cheel and Rust, 1986; Gorrell, 1991; Gorrell and Shaw, 1991).

The subaqueous-fan exposed at Stop 1-4 is the source of ~40% of the concrete sand in the Ottawa area. To wash the sand, 100s if not 1000s of gallons per minute of groundwater are used.

Description of Tackaberry pit

This extensively worked pit currently (2007) has a ~ 500 m long, 6 m high, north trending face along the western margin of the operation. More than 80% of the face was free of slump in the summer of 2006. The principal architectural elements are tabular bedsets of diffusely graded and plane-bed medium sand overlain by dune-scale cross-stratified medium sand and ripple-scale cross-stratified fine sand. Locally scours of diffusely graded sand are prominent. Cross-strata have climbing forms at both the ripple- and dune- scale. A general fining trend occurs from northeast to southwest. The primary architecture has been modified locally by deformation features that include convolute bedding, load structures and fluidization pillars (Fig. 1-18). The underlying silty sand unit is commonly intact; however, local rupture and disturbance of the underlying unit is also observed (Fig. 1-19). Deformation is concentrated in the south part of the section, immediately next to the road, over a lateral distance of 50 to 75 m.

Discussion

<u>Scours and fills</u> Cheel and Rust (1982) describe three styles of scour fills from a nearby subaqueous fan in the Stittsville area: i) horizontally stratified medium sand with imbricate pebbles and cobbles along the base and on bedding planes within the fill; (ii) massive sands; and (iii) horizontally stratified sands. Documented scours in other Champlain Sea fans are up to 10 m deep and 40 m across (e.g. Rust, 1977; Burbidge and Rust, 1988). In the Tackaberry pit, scours are generally relatively small, < 2 m deep and < 10 m across. Fills consist of diffusely graded and planar bedded fineto medium sand. In the western section no pebbles are observed; however, more proximally, fills are more texturally diverse. Generally some degree of faint stratification is present; massive fills are rare.

[21] Russell, H.A.J. and Cummings, D.I. 2011. Stop 1.4. Kemptville/Loughlin Ridge esker – subaqueous fan ; *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 41-45. Geological Survey of Canada, Open File 6947.

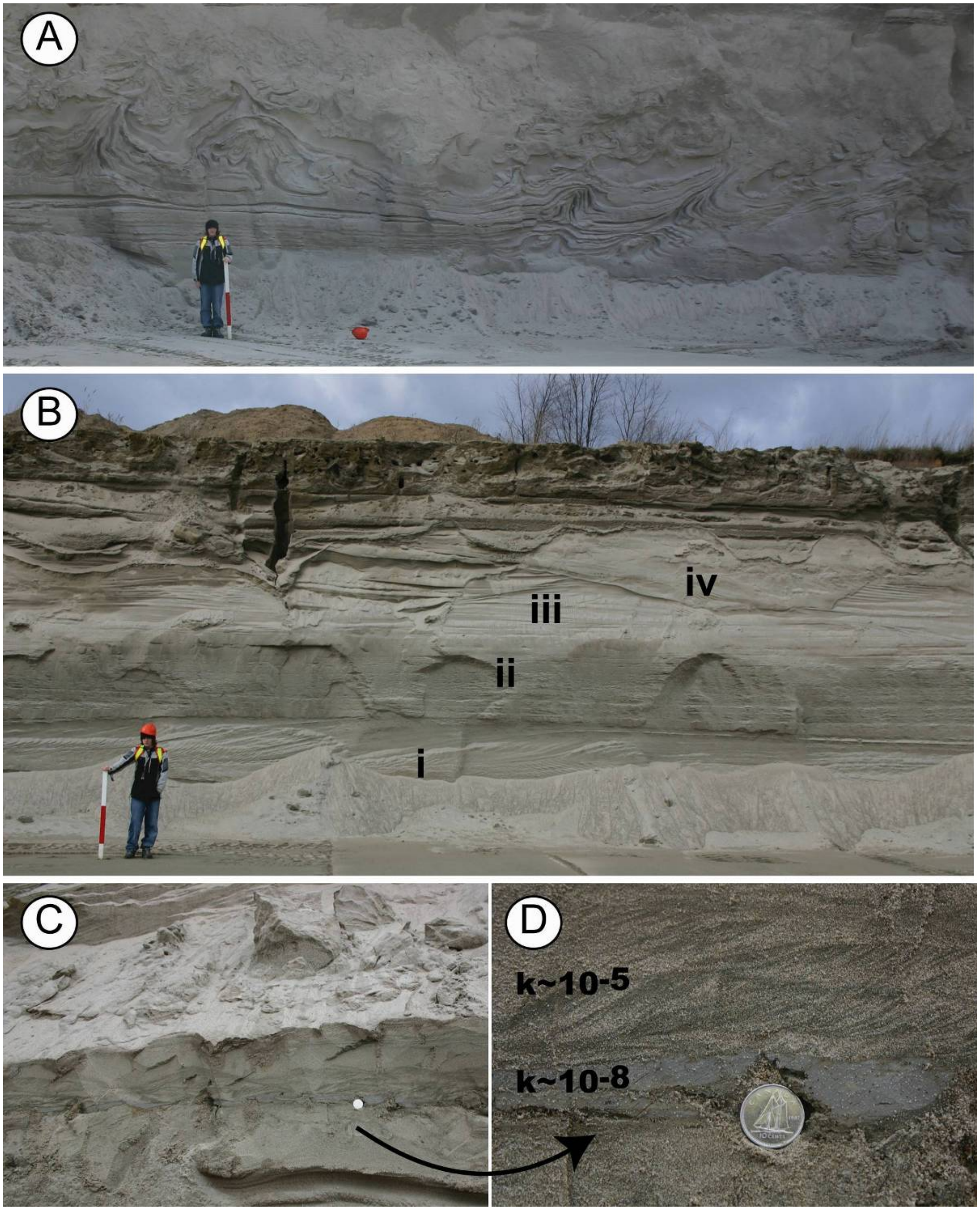

Figure 1-18. Representative images of the western exposure in the Tackaberry pit at Loughlin Ridge. A) Convolute bedding with diaperic structures. B) Typical stratigraphic section of i) dune-scale cross-stratification, ii) climbing ripple-scale cross-lamination, iii) low-angle and dune-scale cross-stratification, iv) scours with diffusely graded fills. C) Toe-set climbing ripples with isolated mud laminae. K values are calculated from grain size data using the Hazen method using D10 of the fine fraction (Freeze and Cherry, 1979). Dime scale is 1.8 cm. Metre stick in (A) and (B) is 110 cm long.

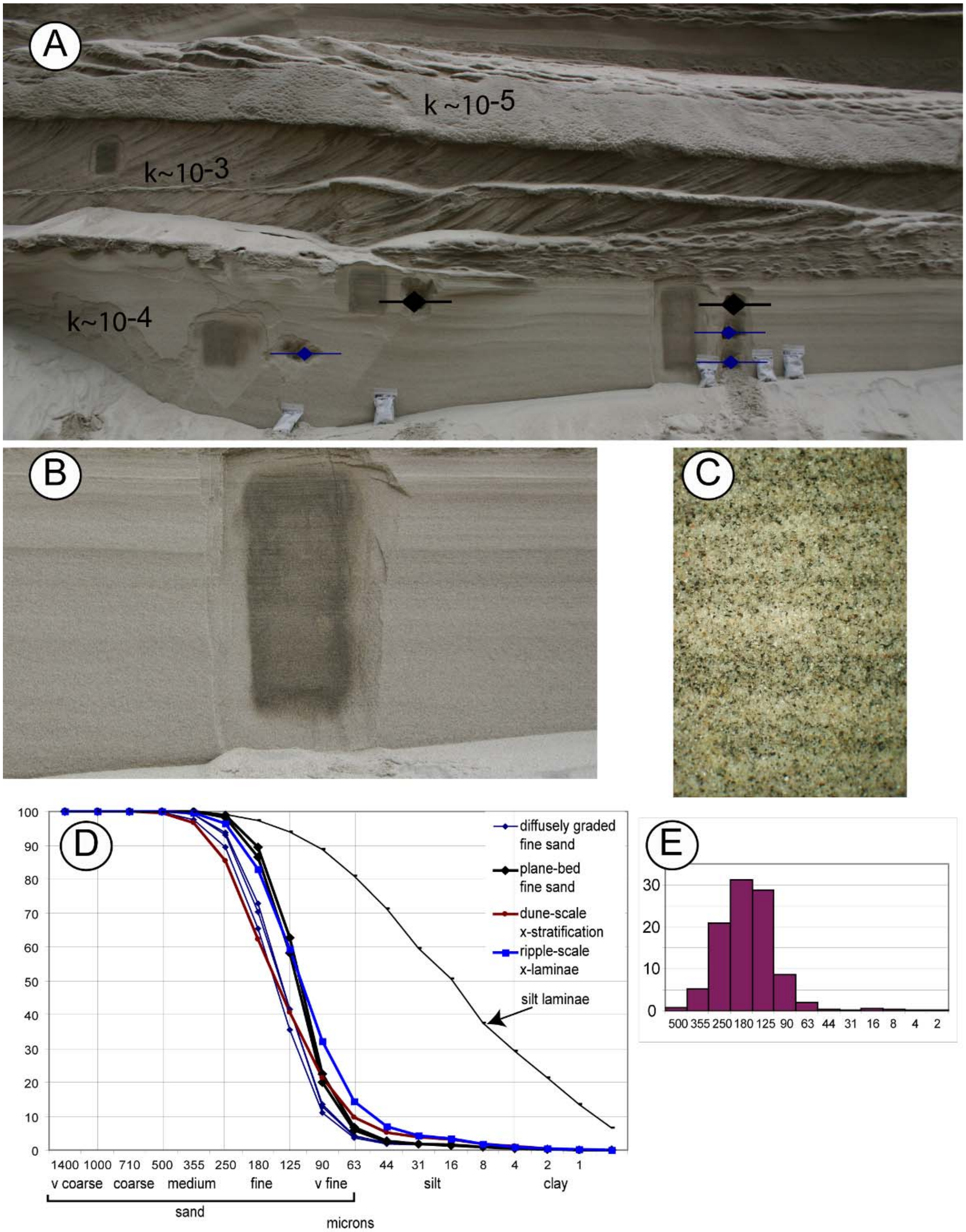

Figure 1-19. A) Section of diffusely graded sand overlain by climbing dune-scale cross-stratification and climbing ripple-scale cross-stratification. Sample locations indicated with bars. B) Close-up of diffusely graded sand fining upward to plane-bedded sand. C) Microscopic image of sediment peel from diffusely graded sand. Note distinct graded bands of grading and heavy mineral concentration. D) Grain size data for sample locations shown in (A). E) Histogram of grain size distribution of a sample of diffusely graded sand. All K values calculated from grain size data by the Hazen method (Freeze and Cherry, 1979).

The predominant fill is diffusely graded fine and medium sand that can locally be traced laterally and/or vertically to plane bed. Previous authors have attributed the scour and fills to slumping (e.g. Rust 1977; Cheel and Rust, 1982) or alternatively to scour beneath hydraulic jumps (Gorrell and Shaw, 1991). The scours and fills at the Tackaberry pit are interpreted to have formed by the latter mechanism, where large vortices within a hydraulic jump impinged on the basin floor (e.g. Long et al. 1990). The absence of slump scars, the low depositional slope, and the upward gradation from diffusely graded to plane bedded sand suggest the scour fills were deposited from sediment laden, turbulent flows rather than from laminar, slump generated sandy debris flows. In flume experiments, Leclaire and Arnott (2005) found that parallel lamination formed at bed-aggradation rates up to 4 mm s^{-1} and bedload-layer sediment concentration up to 0.36. Consequently, this suggests diffusely graded sediment is deposited under even higher aggradation rates and sediment concentrations.

Climbing bed-forms. Climbing small-scale cross-stratification is a ubiquitous element of many glacilacustrine and subaqueous fan deposits (e.g Jopling and Walker, 1968). Climbing dune-scale cross-stratification is less commonly reported. The development of this scale of climbing cross-strata further documents the high-rates of both suspension and bedload sedimentation in the subaqueous fan environment.

Deformation. Deformation features have been described by Cheel and Rust (1986) in subaqueous fan deposits in the nearby Stittsville area. They observed convolute bedding, ball and pillow structures and dish structures. Deformation is attributed to two mechanisms, i) rotational slumping, and ii) dewatering or fluidization due to melt-out of underlying buried ice. By contrast, given the abundant evidence for high sedimentation rates,, we interpret convolute bedding in the Tackaberry pit to be related to dewatering of underlying units as a result of rapid sedimentation.

Groundwater implication Based on a small number of samples, average hydraulic conductivity (K) values are 10^{-3} to 10^{-5} m sec^{-1} for diffusely graded sand, dune-scale cross-stratified fine sand, and ripple-scale cross-laminated fine sand (Fig. 1-19). Local mud laminae are less permeable, with K values of 10^{-8} m sec^{-1}. Mud laminae in the mid fan setting are rare and relatively discontinuous. The sandy fan may therefore provide significant reservoir capacity for coarser gravel deposits inferred to be buried beneath the fan, as they are observed to be elsewhere in Champlain Sea eskers (e.g. Route 300).

Depositional model**.** Sediment exposed in the Tackaberry pit is interpreted to have been deposited on a subaqueous fan beneath a jet downflow of the hydraulic-jump (Fig. 1-20). Hydraulic jumps are characterized by high suspended sediment loads and intense turbulent fluidal flow. Depositional evidence for this is provided by diffusely graded sand and climbing dune- and ripple- scale cross-stratification. Diffusely graded sand may be deposited from the hyperconcentrated the basal layer of a thicker flow (Pierson and Scott, 1985; Best, 1992; Russell and Arnott, 2003). Climbing bedforms have been reported from flume studies in association with hydraulic jumps (Daub, 1996). Additional work needs to be completed at the Tackaberry pit to identify whether the low-angle stratification common to the diffusely graded sand facies was deposited by antidunes. Improved knowledge on this element of the bedding would further constrain the depositional model.

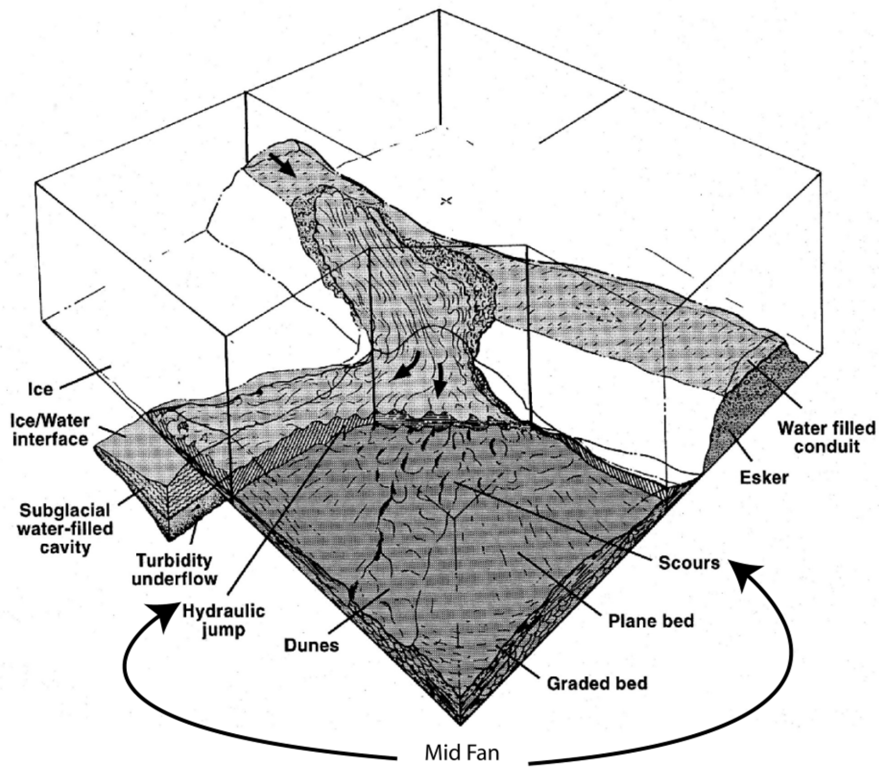

Figure 1-20. Depositional facies model of a grounding line fan with differential ice contact with the bed and lateral breaching of the conduit (Gorrell and Shaw, 1991). Sediment images in figures 1-18and 1-19 illustrate the character of the central part of the model.

Field Trip Stops Sunday, Day Two

Stop 2-1: Seismic Microzonation Hazard Mapping in the Ottawa Area

Heather Crow, Jim A.M. Hunter, Gregory R. Brooks, and Dariush Motazedian22

Introduction

The nature of earthquake seismic waves propagating through the earth depends on the source mechanism, the depth of the source, and the character of the rock types along the travel path to a particular site. The character of the shaking at a specific location on the ground surface (amplitude, frequency and duration) 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 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:

- <u>impedance contrast amplification</u>, caused by shear-wave velocity contrasts between rock and overlying soil layers;
- <u>resonance amplification</u>, caused by the repeated reflection of seismic energy within a soil layer between the ground surface and the boundary with underlying stiffer strata;
- <u>basin curvature effects</u>, caused by focusing-defocusing effects of the subsurface geometry; and;
- <u>basin edge effects</u>, caused by the generation of Rayleigh and Love waves across the surface of a sediment-filled basin.

The net effect is that the ground motion often is greater on areas of softer soil than on stiffer soil or bedrock with the result that damage tends to be greater in areas of soft soil (BSSC, 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).

Seismic site classification

The influence of ground motion amplification for building design is recognized in the 2010 National Building Code of Canada (2010 NBCC) (see Finn and Wightman, 2003). The code uses 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 2-1, 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 2-1. This classification scheme is adopted directly from the system developed by NEHRP (National Earthquake Hazard Reduction Program) for the United States (BSSC, 1994).

[22] Crow, H., Hunter, J.A.M., Brooks, G.R. and Motazedian, D. 2011 Stop 2-1: Seismic microzonation hazard mapping in the Ottawa area. In: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 46-51. Geological Survey of Canada, Open File 6947.

Table 2-1 Seismic site categories as defined in the 2010 National Building Code of Canada (NRC, 2010)

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.

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 2010 NBCC). In the 2010 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 2-2 and 2-3), 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 increasing/decreasing spectral accelerations. For building design, spectral accelerations for the design ground motion are defined in the 2010 NBCC in terms of class C and the amplification factors in tables 2-2 and 2-3 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 2-1). The variations in spectral acceleration of the Ottawa design ground motion for classes A to E are depicted in Fig. 2-1, for shaking periods of 0.2, 0.5, 1.0 and 2.0 s, based on the amplification factors in Tables 2-2 and 2-3. 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.

Ottawa microzonation maps

Seismic microzonation maps have been compiled for the Ottawa area that depict: (i) the spatial distribution of the 2010 NBCC seismic site classes (Fig. 2-2), and (ii) a map of fundamental site period ranges (Fig. 2-3). These maps are the product of research undertaken between 2006 and 2009 within the Eastern Canada Geohazards Assessment Project, Reducing Risk from Natural Hazards Program, in partnership with Carleton University (see Hunter et al., 2009; Hunter et al., 2010). The purpose of this research was to develop and demonstrate cost-effective geophysical techniques to contribute to the creation of seismic microzonation maps (Hunter et al., 2007, Pugin et al., 2007, Benjumea et al., 2008, Crow et al., 2007, Hunter et al., 2009, Hunter et al., 2010).

Table 2-2. Short period amplification factors (F_a) as a function of site class and spectral accelerations at 0.2 s period (NRC, 2010).

Site Class	Values of F_a				
	$S_a(0.2) \le 0.25$	$S_a(0.2) = 0.50$	$S_a(0.2) = 0.75$	$S_a(0.2) = 1.00$	$S_a(0.2) \ge 1.25$
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 2-1.

Table 2-3. Long-period amplification factors (F_v) as a function of site class and spectral acceleration at 1.0 s period (NRC, 2010).

Site Class	Values of F_v				
	$S_a(1.0) \le 0.1$	$S_a(1.0) = 0.2$	$S_a(1.0) = 0.3$	$S_a(1.0) = 0.4$	$S_a(1.0) \ge 0.5$
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 2-1.

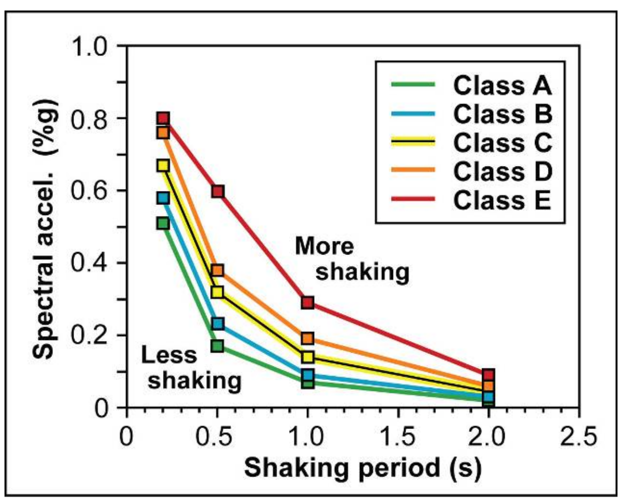

Figure 2-1. Spectral accelerations (ground motion) by site class for the 2010 NBCC design ground motion for Ottawa at shaking periods of 0.2, 0.5, 1.0 and 2.0 s (NRC, 2006), as explained in the text. The design ground motion has a 2% in 50 yr probability of exceedance (or once in 2475 years).

The maps were compiled using subsurface geology derived from 20,000 boreholes, and shear wave velocity (V_s) data from 685 reflection/refraction sites, 25 km of shallow seismic landstreamer data, and nine downhole V_s logs.

The seismic site categories map (Fig. 2-2) has a borehole database at its core, and was compiled by applying V_s-depth functions derived from shear wave reflection surveys to post-glacial deposits, and V_s ranges for glacial deposits and bedrock. These relationships allow for each borehole record to be converted into a shear-wave velocity profile, whereby the travel-time-averaged V_s for the upper 30 m of the ground surface (V_{s30}) could be determined for each borehole site.

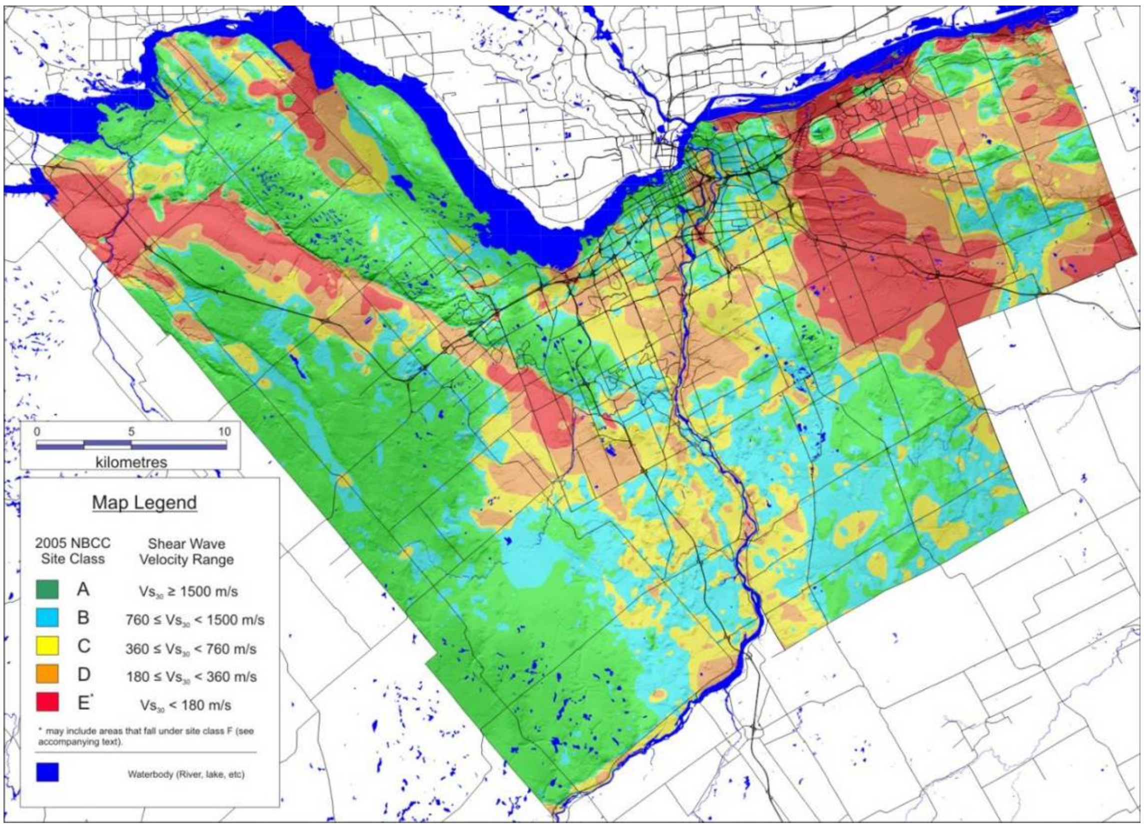

Figure 2-2. 2010 National Building Code of Canada seismic site class map of Ottawa. Refer to Table 2-1 for details on these classes.

As is apparent in Fig. 2-2, 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 Champlain Sea sediment or ‘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, Hunter et al., 2008).

The fundamental period predicts the period, T_o, (or frequency, $f_0 = 1/T_o$), at which resonance amplification will occur. The fundamental period map (Fig. 2-3) was compiled using the 2010 NBCC equation of $4*h/V_s$, where h is the depth to ‘firm ground’, or the primary impedance contrast in the stratigraphic sequence, and V_s is the average shear wave velocity to this depth. The periods increase with increasing thickness of soft soil over firm ground (glacial deposits or bedrock).

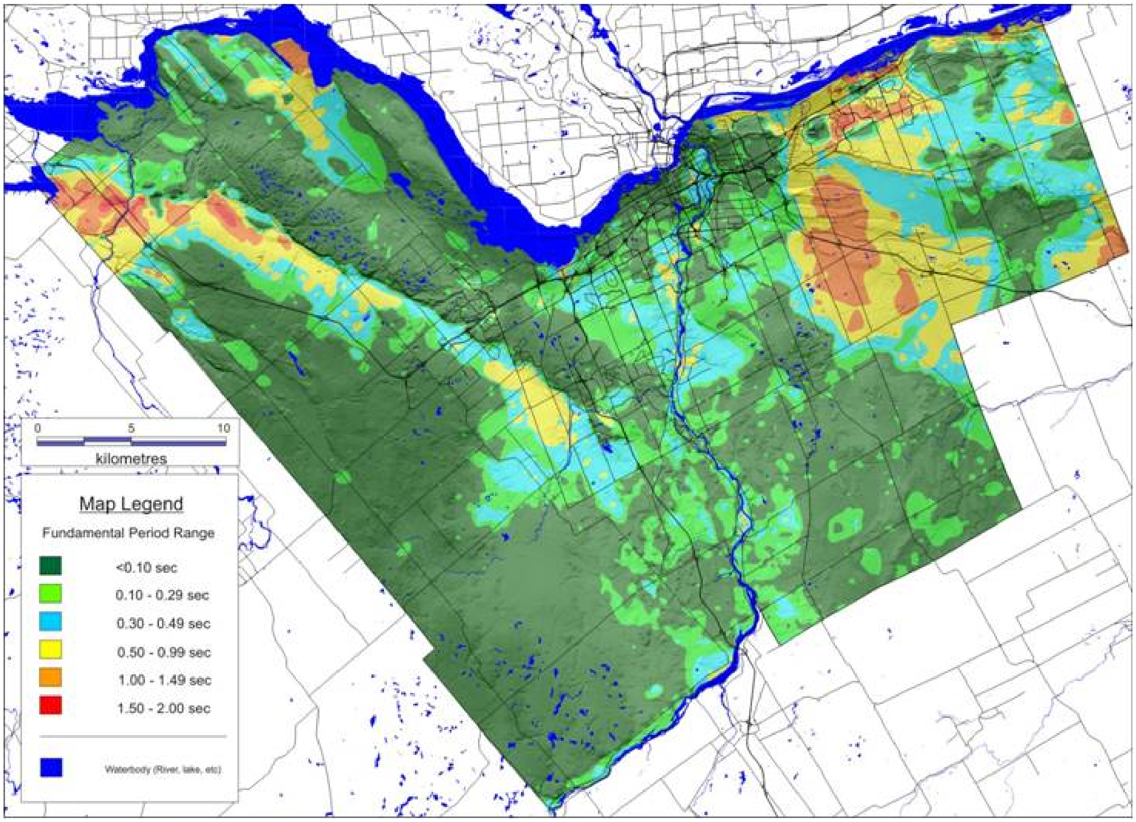

Figure 2-3. Fundamental site period map of the Ottawa area using 2010 NBCC 4h/Vs method.

The motions experienced by a structure during an earthquake can be amplified significantly when the natural resonance period of the structure matches (or approaches) that of the underlying ground. This phenomenon of 'double resonance' can result in greater damage to the structure than might have been expected to occur based solely on the ground motion acceleration. Double resonance has been invoked to at least partially account for observed patterns of damaged buildings in the well-known case of the 1985 Mexico City earthquake, where buildings between 8 and 15 stories high were heavily damaged preferentially (Anderson, 1986).

Map applications

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 maps in Figs. 2-2 and 2-3 allow for 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 the reflect variability of local seismic hazards (Clark and Khadilkar, 1991; Smolka and Berz, 1991).

Copies of the seismic site class map (Fig. 2-2; Hunter et al., 2009) were presented to both the Office of Emergency Management and the Planning and Infrastructure Department at the City of Ottawa, where they are contributing to improved recognition of seismic hazards in the city. The Office of Emergency

Management used the map in a 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.

The maps and shear wave datasets have also been used for research applications by the Geological Survey of Canada and its academia partners within the NSERC-funded Canadian Seismic Research Network. Site categories were evaluated against the "Did You Feel It" reports resulting from the June 23, 2010, magnitude 5.0 earthquake, which struck about 60 km from downtown Ottawa (http://earthquake.usgs.gov/earthquakes/dyfi/). A positive correlation was found between the relative levels of ground motion people experienced and the amplification levels predicted by the seismic site classes.

When converted to GIS layers, the maps can be used with other GIS datasets to assess urban risk and structural vulnerability. Mathematical analysis of potential losses from an earthquake event can be carried out 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.

Stop 2-2: Champlain Sea Deposits West of Ottawa-Gatineau: Kinburn Borehole

Barbara E. Medioli, Heather Crow, Don I. Cummings, Ross Knight, Charles Logan, André Pugin, and David R. Sharpe[23]

Champlain Lookout is situated atop the Eardley Escarpment, a fault that separates Precambrian bedrock uplands of the Gatineau Hills from lowlands in the Ottawa–Bonnechere graben underlain primarily by Paleozoic carbonate rocks (Kay, 1941). The uplands are covered by a patchy veneer of sandy carbonate-poor till. By contrast, the lowland is blanketed by a thick succession of sediment that includes, from bottom to top, carbonate-rich till, eskers, and thick Champlain Sea mud (Johnston, 1917; Cummings et al., 2011). The objective of this stop is to describe the litho-, bio-, chemo-, and seismic stratigraphic patterns observed in an anomalously thick part of the Champlain Sea basin fill near Kinburn, Ontario (Fig. 1). Implications of the results with respect to the deglacial event-sequence are discussed.

Kinburn seismic data and cored borehole

A number of seismic transects were collected in 200X to image what was suspected to be an anomalously thick part of the Champlain Sea basin fill near Kinburn, Ontario. The data confirmed the initial hypothesis and revealed a deep, irregular bedrock basin filled by a thick accumulation (120 m) of sediment. A mound-shaped reflection package is visible at the base of the basin fill, which in turn is overlain by a thick package of continuous horizontal reflections (Fig. 2-5). Given previous experience in the basin (e.g., Pullen et al., 2007), the mound was interpreted to be an esker, and the horizontal reflections were interpreted to be associated with Champlain Sea mud.

A 97 m deep, continuously cored borehole was drilled in 200X to groundtruth the seismic data and test the seismic-based interpretation (Fig. 2-6). The borehole penetrated most of the package of horizontal reflections, confirming that they were indeed generated by Champlain Sea mud deposits. The underlying mound-shaped reflection package, however, was not penetrated because of concerns that pressurized conditions would be encountered, as observed in nearby wells. A piezometer was installed; a suite of downhole geophysical well-logs were collected; and the core was sampled on site for porewater chemistry, grain size, and drop-cone tests, then brought back to the lab where it was split, photographed, and described in detail.

Borehole stratigraphy

The sedimentary succession observed in the Kinburn core can be subdivided into five distinct units (Fig. 2-6). These units are described and interpreted below, from bottom to top.

<u>Unit A – Interstratified mud and sand</u>*:* Unit A consists of of interstratified silty clay and sand core (92 to 97 m; Fig. 2-6). The unit fines upward from interstratified coarse sand and mud with dropstones into silty clay with thin, fine-sand laminae. The unit reacts moderately to strongly with dilute HCl. It contains a sparse monospecific assemblage of *Candona subtriangulata*, a benthic freshwater ostracode.. The apparent conductivity is low, and in-situ porewaters are fresh. The unit appears to correlate with a package of roughly horizontal seismic reflections that onlap the mound-shape seismic package (Fig. 2-5).

Given these observations, Unit A is interpreted to be distal subaqueous outwash strata deposited during northward backstepping of the ice margin across the basin. The high carbonate content suggests that sediment in the meltwater plumes was sourced from the underlying till, which tends to be carbonate-rich

[23] Medioli, B.A., Crow, H., Cummings, D.I., Knight, R., Logan, C., Pugin, A.J.-M., Russell, H.A.J. and Sharpe D.R., 2011. Stop 2-2: Champlain Sea deposits west of Ottawa-Gatineau: Kinburn borehole. *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 52-56. Geological Survey of Canada, Open File 6947.

throughout the western Champlain Sea basin (e.g., Cummings et al., 2011). The presence of *Candona* indicates that the depositional environment was highly freshened during deposition (Guilbault, 1989; Rodrigues, 1992), an interpretation supported by the low apparent conductivity values and fresh in-situ porewaters. Most workers interpret similar units at the base of the Champlain Sea mud package to have been deposited in a glacial lake that existed in the lowland prior to Champlain Sea invasion (e.g., Anderson et al., 1985; Parent and Occhietti, 1988). Others have questioned this interpretation (e.g., Gadd, 1986; Sharpe, 1988), in part because benthic marine foraminifera are observed in places with *Candona* in the basal parts of the mud package (Cummings et al., 2011; see also Rodrigues, 1988), and in part because underlying subaqueous outwash fans locally contain rare *Portlandia arctica* shells (Cummings and Russell, 2007). These observations bring into question the commonly held idea that *Candona*-only assemblages, such as that observed in Unit A of the Kinburn core, are incompatible with marine-disconnected conditions.

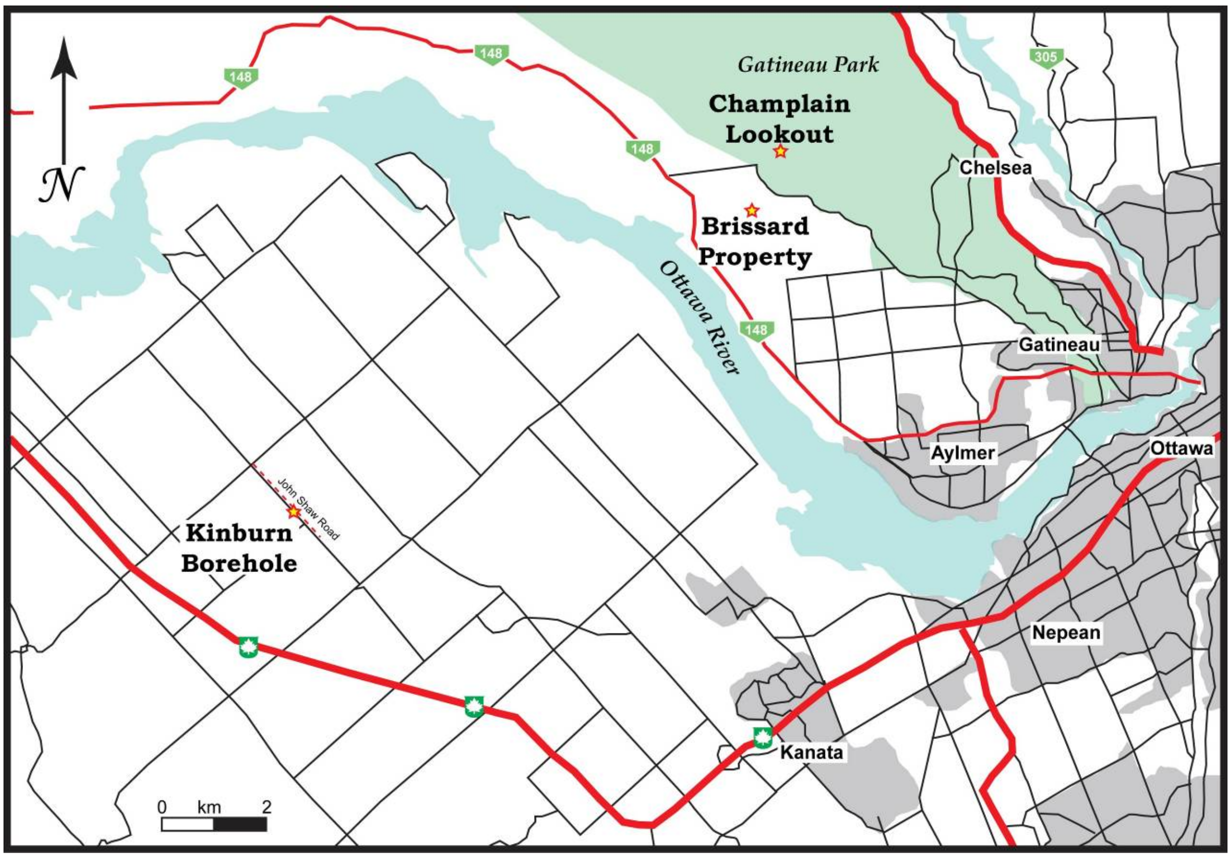

Figure 2-4. Map showing the location of the Kinburn borehole, Champlain Lookout, and the Kinburn seismic reflection profile (red dotted line).

<u>Unit B – Grey rhythmites</u>*:* Unit B (92 to 62 m depth) consists of massive silty clay with sandy or silty partings that grades up into rhythmites defined by alternating bands of finer, dark grey and coarser, light grey mud Unit B gradationally overlies Unit A. Rhythmite thickness ranges from 1-2 mm up to 10 cm and become less distinct near the top of the unit. Apparent conductivity is low and increases upward; in-situ porewater salinity displays a similar trend. The unit reacts moderately to strongly with dilute HCl. Rare shells (suspected to be poorly preserved *Portlandia arctica*, the most common shell type observed in Champlain Sea mud in nearby cores; see Cummings et al., 2011) and rare benthic foraminifera are observed and increase in abundance up core. Rare intensely bioturbated intervals are also observed with increasing frequency upward.

Given the above observations, Unit B is interpreted to have been deposited in a similar cold, deep water depositional setting as Unit A. However, the depositional conditions were not identical: the rare foraminifera and rare Portlandia arctica provide the first unequivocal evidence that the basin had become marine-connected. The proglacial water body was therefore the Champlain Sea, albeit likely still highly freshened at this location. The depositional conditions also appear to have changed slightly over the course of deposition of Unit B: increasing bioturbation intensities and shell abundances moving upward suggest that stress on benthic organisms decreased over time, possibly due to a decrease in turbidity, an increase in salinity or food supply, or some combination of these factors. Ice-front retreat provides a viable unifying mechanism to explain these observations. It also explains the upward-fining trend observed through Units A and B.

<u>Unit C – Massive mud</u> Unit C (19 to 62 m depth) consists of massive grey mud with rare intervals of vague horizontal stratification (Fig. 2-6). Its basal contact is cryptic, possibly gradational. Shells (suspected to be poorly preserved *Portlandia arctica*; e.g., Cummings et al., 2011) reach maximum abundances in Unit C. Bioturbation intensity also reaches a maximum, as does microfossil diversity and density, apparent conductivity, and in-situ porewater salinity. A black substance, possibly hydrotroilite, that disappears after several hours of exposure to air is commonly observed on freshly cut core surfaces. Freshly cut core surfaces also commonly have a subtle sulphorous odour. Unit C commonly reacts with dilute HCl, but typically less vigorously than observed with Units A and B. .Unit C is interpreted to have been deposited in a similar cold, deepwater setting as Units A and B, but at a time when the near-floor salinity of the Champlain Sea had reached a maximum and when stress on the benthic biotic community had reached a minimum.

<u>Unit D – Red & grey mud rhythmites</u> Unit D consists of nine metres (10-19 m depth) of red and grey rhythmites with a sharp basal contact . Rhythmite beds range in thickness from 5-50 cm and generally thicken upward. Each rhythmite consists of a basal red band that grades into a grey clay-rich band. The top of the grey bands is sharp. Black mottling occurs throughout the unit and can cross cut the rhythmite boundaries in places. Rare shells (likely *Portlandia arctica*) are observed. As a whole, the unit exhibits several notable trends: grain-size, apparent conductivity, in-situ porewater salinity, and microfossil diversity and abundance all decrease moving upwards.

It is possible that the rhythmites may actually be varves which represent fluctuating, seasonal deposition (see Aylsworth, 2007). A drop in salinity vertically within the in the declining diversity and abundance of foraminifera and ostracodes, on the apparent conductivity log, and by the pore water conductivity (see Medioli et al., in press). Facies D is interpreted as a transitional sequence recording the final retreat of the Champlain Sea from the Kinburn area (see Fulton et al., 1985; Gadd, 1986) where the water column shifted from marine to estuarine salinities.

<u>Unit E – Laminated mud</u> Unit E consists of 8 m of bioturbated, brown clayey silt. Some laminations are present in this facies, primarily in the form of silt laminae. The sediments are devoid of shells and microfossils but contain organic detritus. The facies E deposits are interpreted to represent sediments aggraded within the proto-Ottawa River that became established after 10 ka BP, as the Champlain Sea receded due to differential uplift (Fulton, 1987). This fluvial deposition occurred in one of the many paleo-channels that are present throughout the NCR.

Summary

The Kinburn borehole provides an important stratigraphic representation of the de-glacial and post-glacial histories in the western portion of the NCR. The borehole penetrated almost 100 m of undisturbed late Pleistocene sediments spanning the deglaciation of the region, the inundation of the western arm of the Champlain Sea, and the eventually recession of the sea and establishment of a proto-drainage network. The suite of sediments record the salinity fluctuations and changing sediment provenance that characterize the Champlain Sea. Data collected from the Kinburn site enhance our knowledge of this period and improve the understanding of the groundwater and seismic characteristics of late Quaternary basin fills within the Ottawa-Bonnechere graben system.

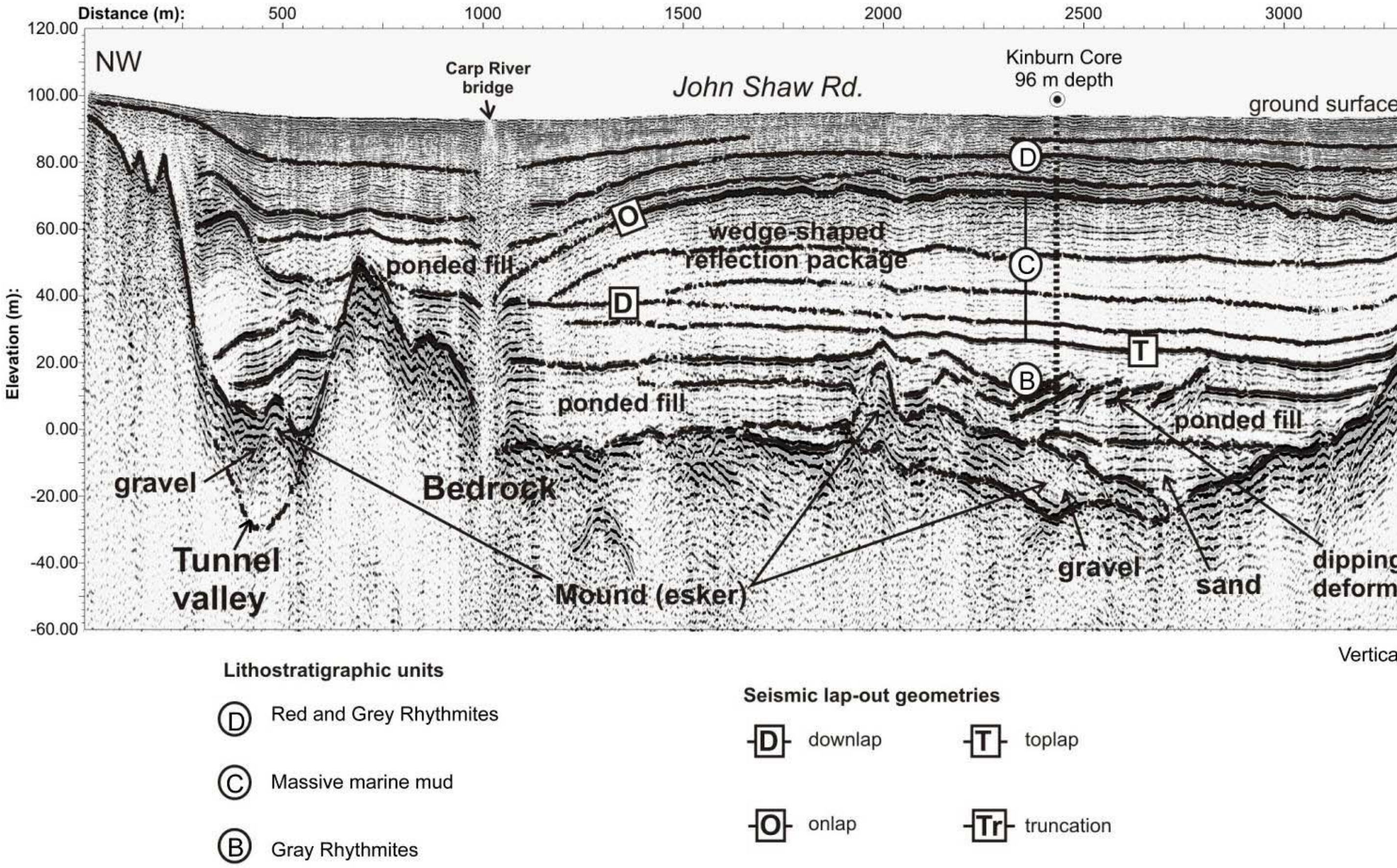

Figure 2-5. Interpreted seismic reflection profile along John Shaw Road (Kinburn, Ontario).

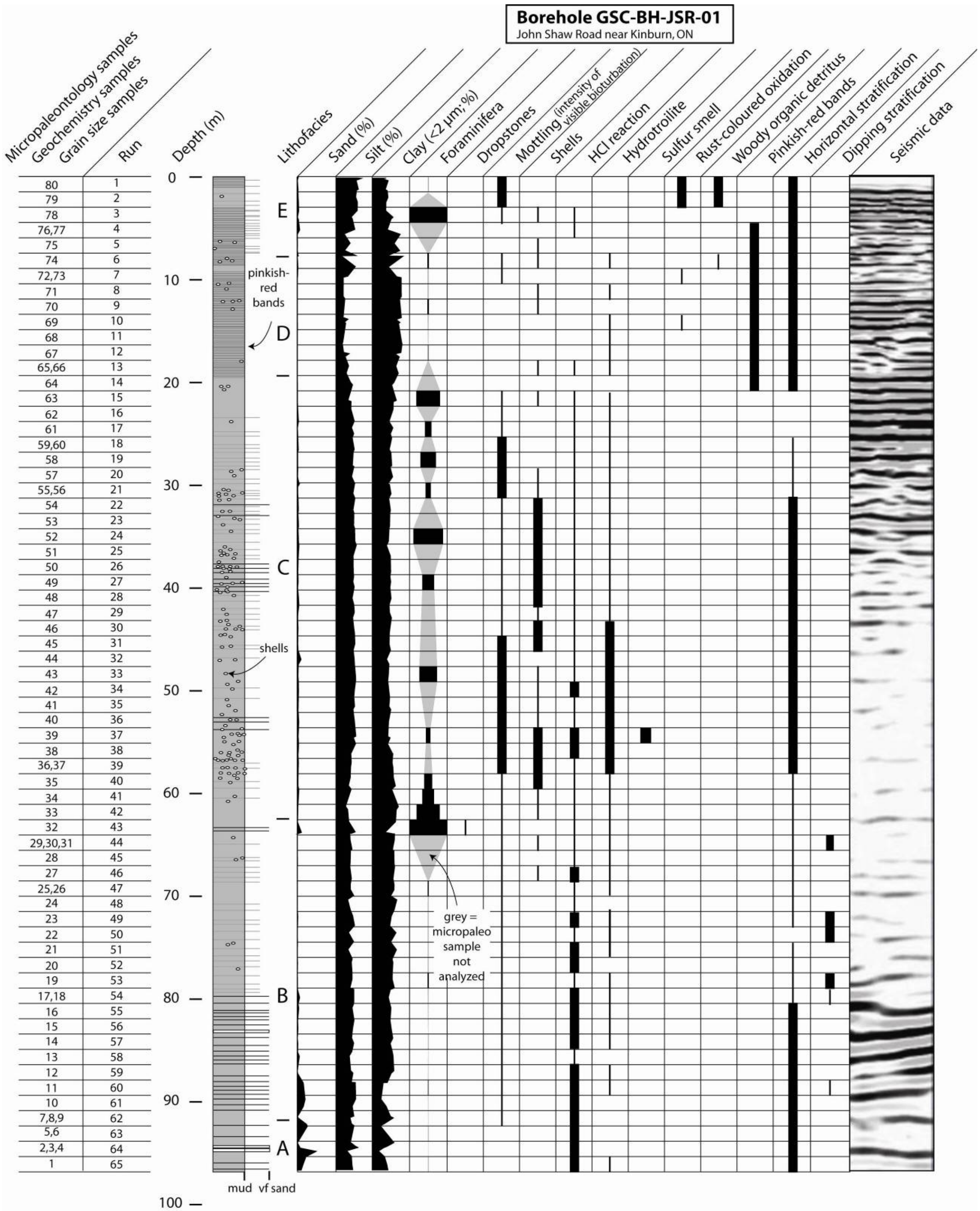

Figure 2-6. Summary stratigraphic log of the Kinburn borehole.

Stop 2-3A: Earth Flow Scars of Breckenridge Valley

Gregory R. Brooks and Barbara E. Medioli[24]

Earth flow scars

Breckenridge Creek is a tributary of the Ottawa River, located approximately 14 km northwest of Aylmer, Gatineau, Quebec (Map 3). Draining ~66 km^2, the creek watershed encompasses i)Precambrian bedrock, which forms the steep western side of the Gatineau Hills, ii) a quasi-flat plain underlain by fine-grained Champlain Sea deposits, and, along its lower 2 km, iii) an erosional terrace of the Ottawa River (Fig. 2-7). Across the surfaces of the Champlain Sea plain and the terrace, the stream network has eroded steep-sided valleys that are incised up to about 30 m deep (shallower on the terrace) into Champlain Sea sediments. Bedrock underlying these surfaces is composed of Paleozoic sedimentary rock, which is present in the area extending to the south of Breckenridge Creek, and Precambrian bedrock that forms small hills protruding through the plain to the north of the creek (see Richard, 1982c). The boundary between the two basic rock types is not exposed and its specific location is uncertain. Along the incised course of Breckenridge Creek, bedrock (Paleozoic) is exposed only along two short reaches located immediately downstream of the Lac Mountains reservoir (Fig. 2-7).

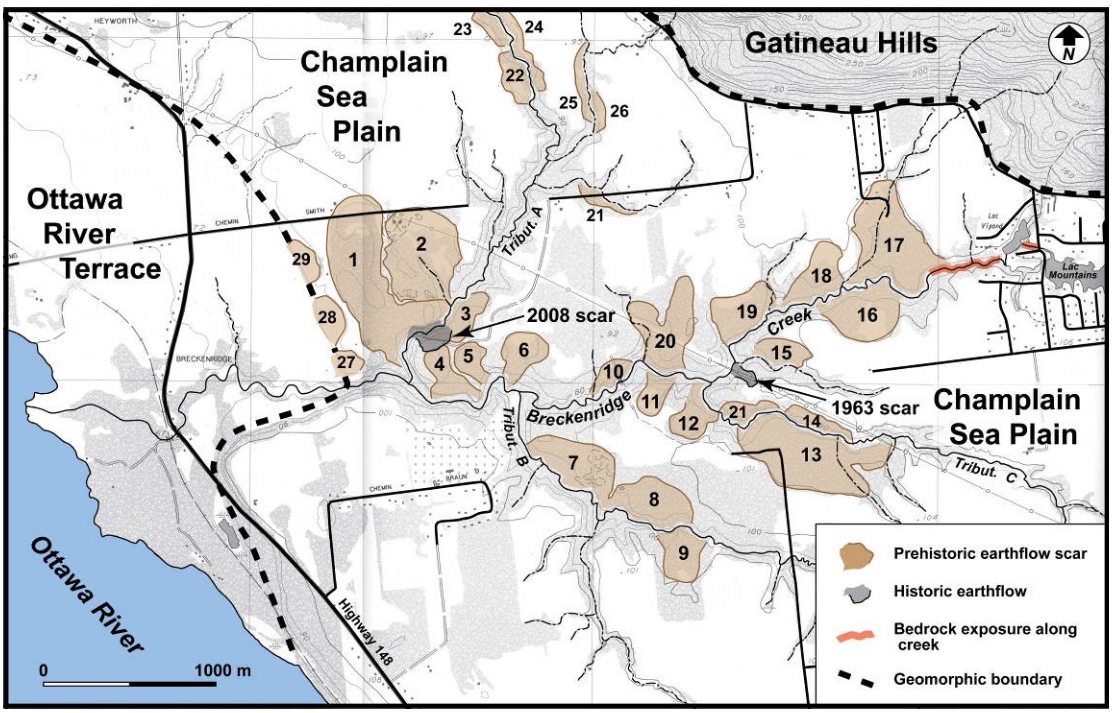

Figure 2-7. Map showing the location of prehistoric and historic earth flow scars along the incised course of Breckenridge Valley, where the stream network crosses the plain composed of Champlain Sea sediments.

Breckenridge Valley is particularly notable because of the large clustering of earth flow scars along the stream network where it crosses the Champlain Sea plain. Twenty six of these scars

[24] Brooks, G.R. and Medioli, B.A., 2011. Stop 2-3: Earth flow scars of Breckenridge Valley. *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 57-61. Geological Survey of Canada, Open File 6947.

(numbered 1 to 26 in Fig. 2-7) represent large prehistoric failures that occurred along Breckenridge Creek or one of its larger tributaries. The scars range from 13 000 to 252 000 m^2 and have retrogression distances of 50 to 920 m. The morphology of the scars is variable and includes several well-defined bottleneck bowls (e.g., scars 5 and 6) and others where the scar width (W) greatly exceeds the retrogression distance (R) (i.e., R/W is less than 0.5, e.g., scars 3, 10, 13, 14, 15, 24 and 25). Three scars (numbered 27 to 29) occur along the side of the fluvial terrace just to the west of scar 1 (Fig. 2-7). The deposits of scars 28 and 29 are truncated at the edge of the terrace because of erosion by the ancestral Ottawa River. The locations of two smaller earth flows that occurred in April 1963 (see Eden et al., 1965; Mitchell, 1970), and April 29, 2008 (R. Brizard, personal communication, April 21, 2010), are also shown on Fig. 2-7. Not shown, however, are numerous shallow slumps that occur regularly along the valley sides of the creek and its tributaries.

As depicted in Fig. 2-8, the earth flow scars are concentrated within an area of about 11 km^2. This map shows contour lines that represent depth from the ground surface to an impedance layer, which is a surrogate for the thickness of soft sediment (see Hunter, this volume). These contours reveal that the scars are situated over a buried valley filled with up to 90 m of soft sediment (Fig. 2-8; see Hunter, this volume). The logs of two water well records in the local area record this material as 'clay', which is interpreted to be Champlain Sea sediments.

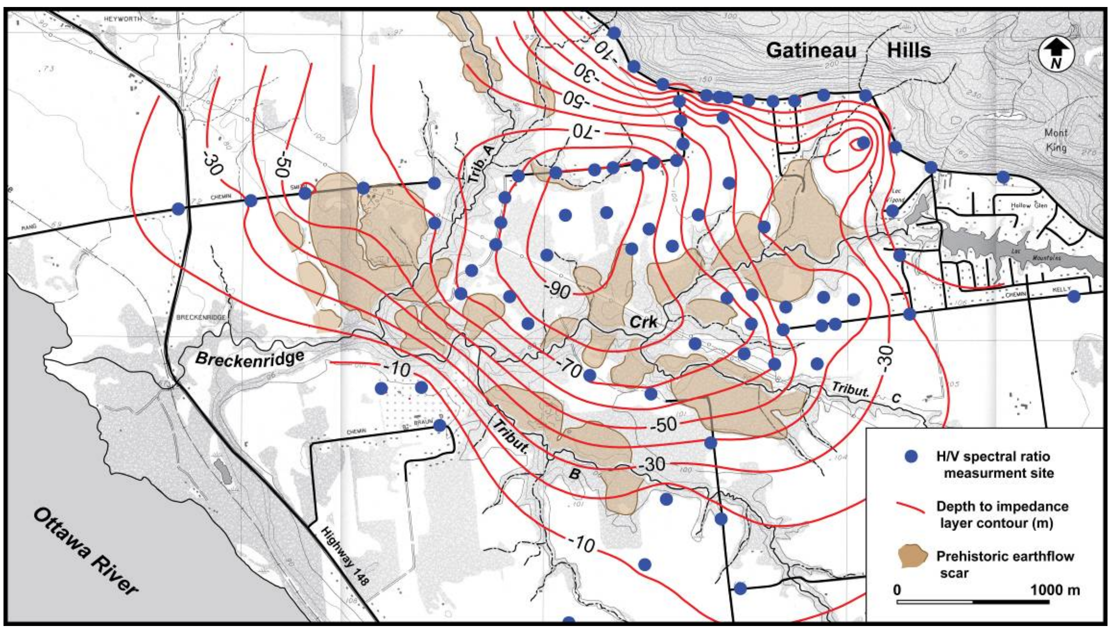

Figure 2-8. Map showing the locations of prehistoric earth flow scars along Breckenridge Valley superimposed on depth to the impedance layer contour lines that represent the thickness of soft sediment. These contour lines reveal the presence of a buried valley that is overlain by a suite of soft Champlain Sea sediment (see Hunter, this volume).

The clustering of earth flow scars over buried valleys has been noted previously in the St. Lawrence Lowlands region (see La Rochelle et al., 1970). The soils over buried valleys have been found to exhibit lower soil strengths and higher sensitivities than adjacent areas. These characteristics are attributed to the leaching of salts due to ground water movement, which occurs preferentially in areas underlain by buried valleys. Although data are not available for the Breckenridge Valley area to show a similar pattern, geotechnical data at the site of the 1963 earth flow reveal sensitivities in excess of 100, which are consistent with occurrence of leached sediments (see Eden et al., 1965).

2008 Earth flow

The most recent earthflow in Breckenridge Valley happened on April 29, 2008, (B. Brizard, personal communication, April 21, 2010) along Tributary A just upstream of the confluence with Breckenridge Creek (Figs. 2-7 and 2-9). Spoil from the failure obstructed Tributary A flow forming an impoundment that flooded the valley upstream. The failure was not directly observed, but became apparent after local residents noticed flooding along Tributary A and another small creek crossing Smith-Lenard Road and investigated the source (McLean, 2008).

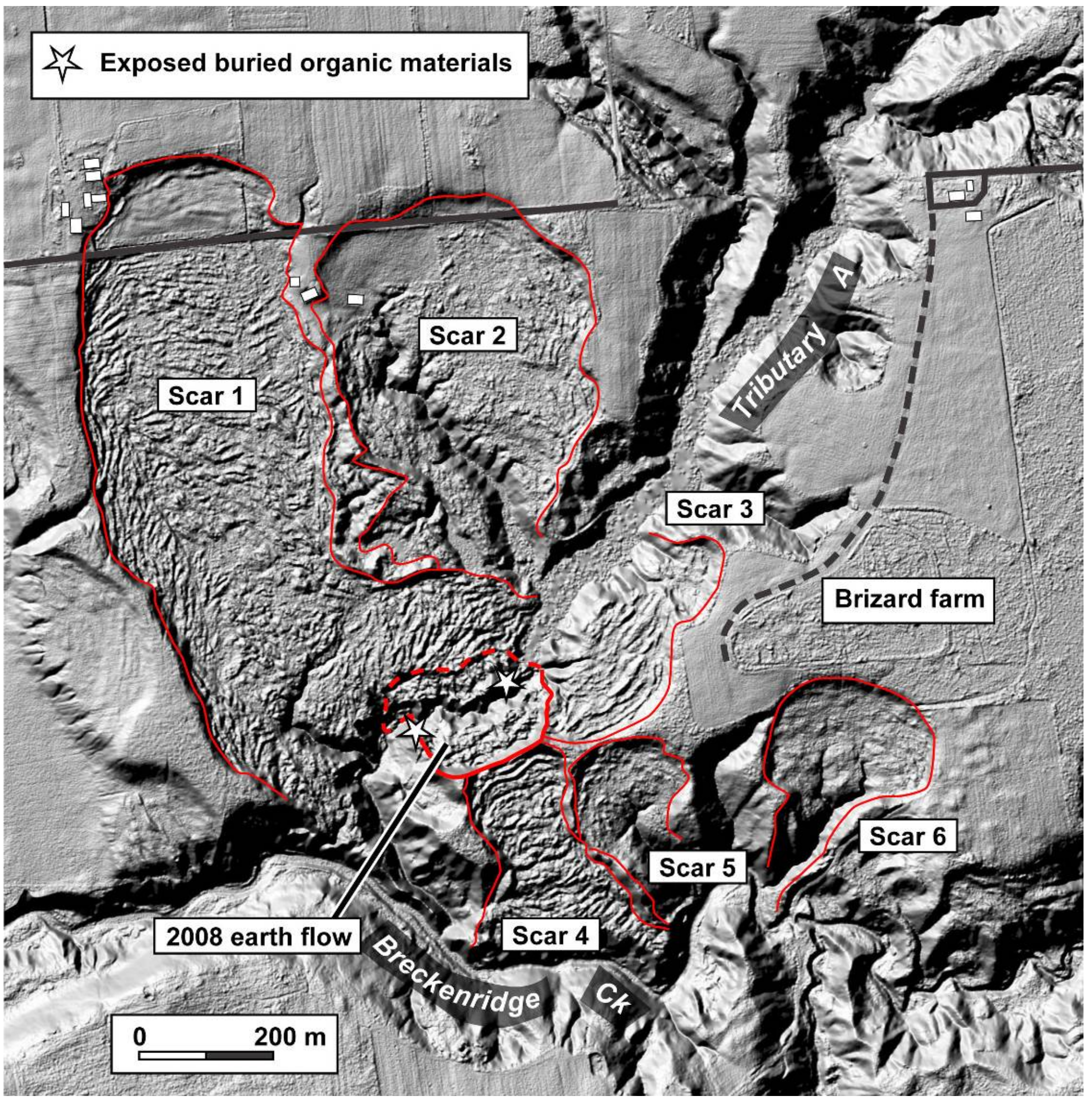

Figure 2-9. LiDAR image showing the location of the 2008 earth flow backscarp (heavy solid red line) and limit of spoil (dashed red line) within a cluster of six prehistoric earth flow scars (light solid red lines) in the area of the confluence of Breckenridge Creek and Tributary A. Also shown are two white stars that mark the locations of buried organic material exposed along the backscarp or within spoil of the 2008 failure. Note, the backscarp of the 2008 failure truncates portions of scars 3 and 4. (Reproduced with permission of Ministère des Ressources naturelles et de la Faune Québec)

The crater forming the source area of the failure is about 190 m wide and retrogressed about 90 m from the pre-failure edge of tributary A. The distance from the backscarp to the outer edge of the spoil on the opposite side of Tributary A is about 160 m (Fig. 2-9). Spoil traveled downstream along the valley of Tributary A about 40 m, but the travel distance upstream is unknown because of the continued impoundment of Tributary A. While sizeable, the crater of the 2008 failure is small in comparison to the prehistoric earth flow scars in the immediate area, particularly scars 1 and 2 on the opposite side of the valley (Fig. 2-9).

The spoil impounding Tributary A was overtopped several days after the failure and has since been incised up to 5-7 m by the creek through the middle and lower portions of the stream course, as of September 2010 (Fig. 2-10). A residual impoundment persisted through 2010, albeit lower by 2- 3 m than the initial water level. This incision evidently occurred progressively rather than catastrophically, as there is no geomorphic evidence of a mud flow downstream along Tributary A or Breckenridge Creek. There were initial concerns by local authorities that a mud flow traveling down the creek might damage the new bridge along Highway 148 near the mouth of the valley.

Figure 2-10. Photograph of the incised creek course through the spoil of the 2008 failure, taken in early May 2010. On the surface behind the stream course, note, the weathered appearance to the spoil compared to that shown in Fig. 2-11.

Since April 2008, the spoil has dried, cracked and crumbled resulting in the loss of primary morphological features associated with the movement of the failure (Fig. 2-11). Nevertheless, interesting features still persist including exposures of intact Champlain Sea deposits along the portions of the backscarp (particularly beneath the pathway leading up to the scar), back-rotated slide blocks, and freshly-exposed spoil at depth within the failure deposits along the incised stream course. Champlain Sea deposits undoubtedly liquefied at depth during the failure, but these have not yet been exposed along the incised stream course (as of September 2010). Instead, primarily tilted, intact beds are seen within the Champlain Sea deposits along the immediate level of the creek course. The presence of these tilted beds, and the limited run out of the spoil downstream (and presumably upstream) along the valley of Tributary A are consistent with the 2008 failure being a lateral spread. A lateral spread is the product of liquefaction along a relatively thin layer at the failure plane (or zone) and experiences subsidence as it moves, as opposed to a retrogressive failure that occurs as a rapid series of successive rotational failures (Cruden and Varnes, 1996; e.g., the failure at stop 5; see Perret, this volume).

Figure 2-11. Photograph of the 2008 failure taken in early May 2008 showing the fractured Champlain Sea sediments that form the surface of the spoil. Note, the leaning trees in the background that have been rafted by the failure. An arrow marks the location of a person for scale. (Photograph: Gaétan Lessard, Ministère de la sécurité publique du Québec)

The headscarp of the 2008 failure retrogressed into both scar 3 and the immediate headscarp area of scar 4, exposing spoil within both older scars (Fig. 2-9). A layer of organic material buried beneath prehistoric-aged spoil can be found about 5 m below the top of the bank along the downstream edge of the backslope (Fig. 2-9). Three radiocarbon dates from this layer range between 1115 to 1205 yr BP (Brooks unpublished data). A second buried layer of organic material, including large logs, can be found at the upstream end of the spoil on the north side of the creek (Fig. 2-9). This layer is inclined and can be followed discontinuously diagonally from near the top of the bank down to creek level. The ages of two wood samples from near the top of the bank range from 7000 to 7105 yr BP (Brooks unpublished data) and are evidently part of the same layer that contained wood aged 7050 ± 80 (GSC-6233), 7030 ± 70 (GSC-6243) and 6980 ± 80 (GSC-6246) yr BP that was collected by the GSC in 1996. The presence of these two buried organic layers of markedly different age within the spoil and the backscarp reveal that the 2008 failure occurred through the remobilization of older earth flow spoil and that one of the older failures occurred at about 7000 yr BP. Ongoing research is examining the chronology of the prehistoric earth flows scars in Breckenridge Valley to elucidate possible triggering mechanisms for these ancient failures.

Stop 2-3B: A Passive Noise-Monitoring Geophysical Method to Detect Buried Valleys

Jim A.M. Hunter[25]

In areas where the thickness of soft soil (with low shear wave velocities) exceeds about 10 m and overlies competent bedrock or firm soil (with high shear wave velocities), there is a possibility of ground resonance due to ambient seismic noise. Resonance is important for earthquake engineers because it estimates the band of frequencies at which earthquake energy will be amplified. However, the resonant period (or frequency) can also be used on a reconnaissance level to map the thickness of soft sediment because the period of the ground resonance is governed by the shear wave velocity of the soft soil and the thickness of this layer.

Nakamura (see SESAME, 2004) introduced a method to estimate the fundamental site period T_0 (or fundamental site frequency F_0 - the inverse of T_0) utilizing ambient seismic noise in the frequency range 0.1 to 20 Hz. The ambient seismic noise is thought to be a combination of vertically-travelling tele-seismic shear wave noise from microseismic activity as well as the contribution of Rayleigh and Love surface waves from local sources, such as, wind, flowing water, vehicle traffic (SESAME, 2004).The horizontal-to-vertical spectral ratio (HVSR) of ambient ground motion can indicate a peak period equivalent to the resonant period of the ground given by:

(1) $T_0 = 4H / Vs$

Vs - average shear wave velocity through soil layer
H - thickness of the soil

The working model for the HVSR-resonant period phenomenon suggests that only horizontal components of shear motion are amplified through a low shear wave velocity surface layer.

An instrument, such as the Micromed Tromino shown in Fig. 2-12, is a high resolution three-component ground motion measurement device that has been designed specifically for attaining HVSR measurements. The unit is self-contained, readily deployable and can be used on road-side or remote terrain as well as in both rural and densely urban environments. It comes with user-friendly interpretation software that allows T_0 and F_0 to be easily determined.

An example of a HVSR plot of ground motion obtained from the Tromino seismograph is shown in Fig. 2-13. This plot depicts a well defined peak that represents T_0 and F_0 at this specific site. The shape of this plot is typical of that produced where a strong seismic impedance boundary occurs between a 'thick' deposit of soft sediments and underlying firm ground or sound bedrock.

In the Ottawa area, the predominant seismic impedance boundary is the base of the soft Champlain Sea sediments overlying glacial sediments (ice-contact or glaciofluvial sediments) or, Paleozoic or Precambrian bedrock (if glacial sediments are absent). This stratigraphy typically exhibits a strong impedance contrast and yields a well defined sharp peak to the HVSR plot, such as that depicted in Fig. 2-13. HVSR responses, and hence the shape of the peak, however, become altered by rapid lateral changes to the impedance depth which can occur in the subsurface, particularly at basin edges.

Direct measurements of soft soil thickness from boreholes and seismic reflection profiling, and HVSR measurements of T_0 and F_0 in the Ottawa area at 107 locations have produced the

[25] Hunter, J.A.M. 2011. Stop 2-3B: A Passive noise-monitoring geophysical method to detect buried valleys. Russell, H.A.J., Brooks, G.R. and Cummins, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 62-65. Geological Survey of Canada, Open File 6947.

Figure 2-12. Tromino three-component noise monitoring seismograph (on right).

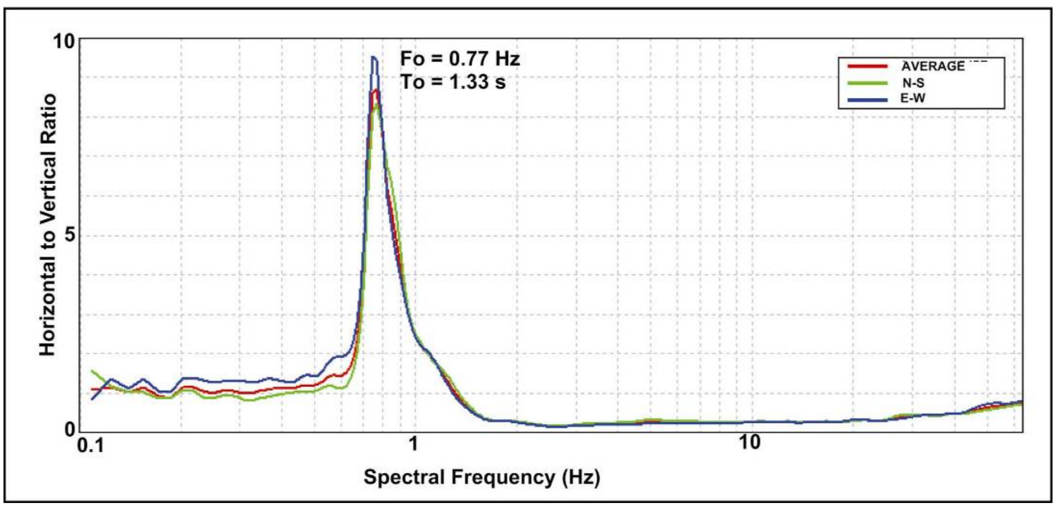

Figure 2-13. Typical processed HVSR field record from a site where a strong impedance contrast occurs between a thick deposit of soft Champlain Sea sediments and underlying firm ground or sound bedrock.

empirical relationship shown in Fig. 2-14, which relates depth to the impedance boundary (Z) to fundament site period (T_o). This relationship is summarized by the function:

(2) $Z = 56.7 * T_o^{1.48} \pm 6.1$ (1σ range)

Equation 2 thus allows HVSR measurements of T_o to be used to estimate the thickness of the soft soil layer at a given location on the ground surface. HVSR measurements at multiple sites within a given area allow the lateral thickness of the soft soil layer to be mapped. Ground-truth observations of depths to the impedance boundary, however, are necessary within the survey area to aid in the interpretation and to improve the reliability of depth estimates. Ground truthing can be done using two-dimensional shallow geophysics, borehole data, or both.

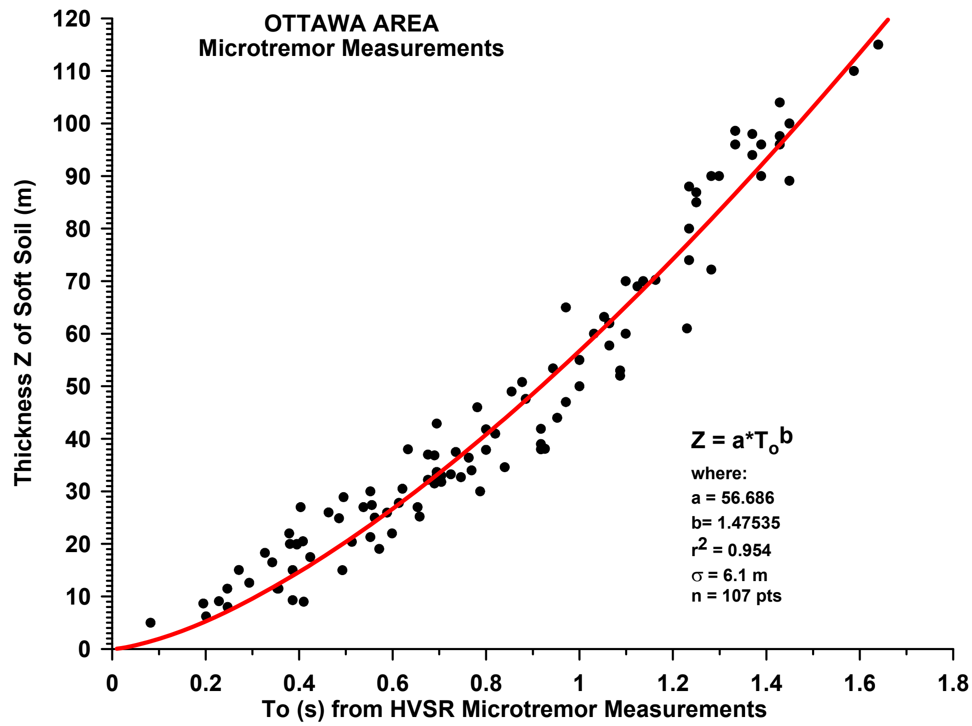

Figure 2-14. Empirical relationship between fundamental site period (To) and layer thickness (Z) for Champlain Sea sediments in the Ottawa area.

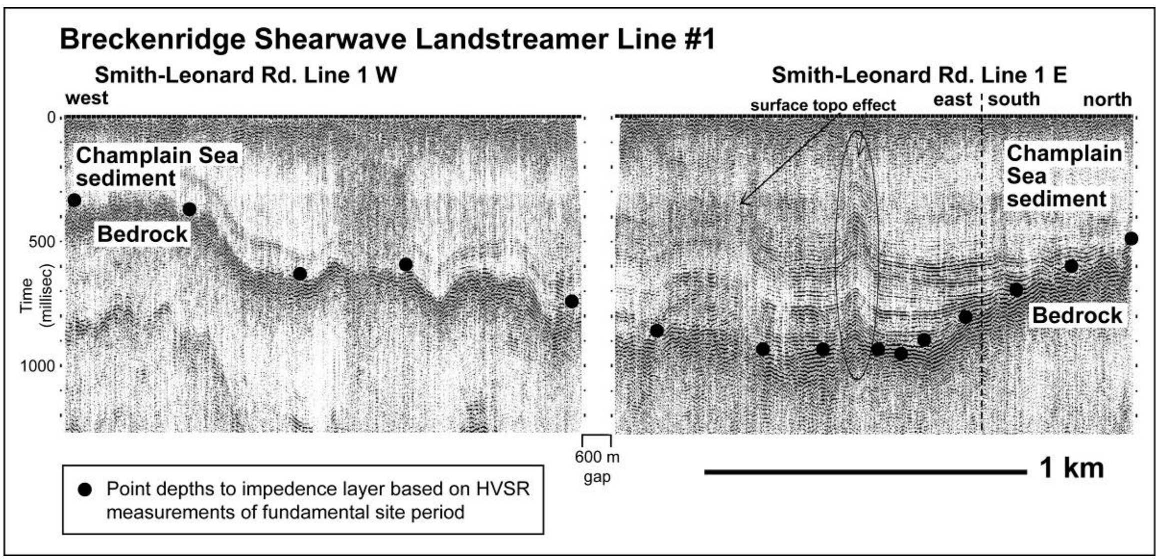

Figure 2-15. Landstreamer shearwave line obtained along Smith-Leonard Road, Breckenridge Valley, showing the estimates of depth to the impedance boundary based on HVSR measurements, as explained in the text.

HVSR measurements of T_o have been collected at over 71 points in the area of Breckenridge Valley, Quebec, to map the depth to the impedance boundary (see Fig. 2-8). To verify the data, the depth estimates obtained along Smith-Leonard Road are superimposed upon the returns obtained from a landstreamer shear wave line (see Pugin et al., 2007), as shown in Fig. 2-15. As is readily apparent, the depth estimates fall on or very close to the reflector marking the interface between soft Champlain Sea sediments and bedrock. The presence of Champlain Sea sediments is confirmed by local water well and geotechnical boreholes. It is apparent in Fig. 2-15 that the impedance boundary is formed at the interface between Champlain Sea sediments and bedrock; glacial sediments are absent (or too thin to be readily obvious) at least in this portion of the buried valley. As discussed further in Brooks and Medioli (this volume), Breckenridge Valley overlies a buried valley containing soft sediment up to 90 m thick.

Overall, the HVSR technique is a cost-effective method for obtaining depth to impedance layer measurements in areas where subsurface information may be sparse or restricted to road ways. Reliable estimations of a buried interface can be obtained where shear wave impedance contrasts are large (e.g. where soft sediments overlying firm ground or bedrock).

Stop 2-4: Buckingham Subaqueous Fan and Champlain Sea Deposits

Hazen A.J. Russell and Don I. Cummings[26]

Glacial sedimentology studies in the Ottawa area have historically focused on larger aggregate operations south and west of the city (e.g., Rust and Romanelli, 1975; Gorrell, 1991). Many of these deposits are now worked out. By contrast, urbanization in Gatineau has led to the enlargement of aggregate operations on the Quebec side of the Ottawa River. Consequently, the Buckingham pit has only recently come to the attention of geologists and has had little detailed study.

The Buckingham pit is located on the Precambrian Shield in an area of hummocky terrain within the du Lièvre River valley. Bedrock outcrops immediately to the east. Striated bedrock is exposed locally at the base of the pit. Strata in the pit are well exposed (Fig. 2.16A, B) and exhibit net upward fining (Fig. 2.16A). Three units comprise the stratigraphic succession: (i) a ~25 m thick sand and gravel unit overlying bedrock, (ii) a ~3 m thick gradationally based laminated mud unit, and (iii) a < 2 m thick, sharp-based layer of oxidized sand and gravel.

The sand and gravel unit has a pronounced southwestward (downflow) fining: crudely stratified gravel is abundant in the main pit face, whereas sand is abundant on the opposite side of the pit, to the south and southwest. The sand is diffusely graded. Dune- and ripple-scale cross-stratification is rare (Fig. 2.16). Diffusely graded sand beds are laterally extensive, in places extending several tens of meters. Scours are rare. Sand beds in the south part of the pit dip roughly southward, but dip directions are highly variable, fanning nearly 180° from east to west (Fig. 2.16C). Rare boulder-sized dropstones are observed. High-angle faults are present locally.

Discussion

<u>Diffusely graded bed.</u> This pit is exemplary for its diffusely graded beds of medium to coarse sand, which are commonly interpreted as being hyperconcentrated flow deposits (Gorrell and Shaw, 1991). The diffusely graded sand does not appear to be related to large, steep-walled scours, as it commonly observed in sites south and west of Ottawa (e.g. Rust, 1977; Cheel and Rust, 1982), but rather forms laterally continuous beds.

<u>Dune- and ripple cross-stratification—why so scarce?</u> Climbing-ripple cross-stratification is a ubiquitous element of many subaqueous fan and delta deposits (e.g., Jopling and Walker, 1968). In fact, few other,depositional environments produce similar thicknesses of climbing bedforms. At this site, however, climbing ripple cross-stratification is minor in extent. Climbing ripples are typically associated with *high* sedimentation rates; however, might their absence at this site indicate sedimentation rates that were in fact *too high*? This argument is commonly invoked to explain lack of ripple- or dune cross-strata in high-energy deposits, the reasoning being that increased near-bed sediment concentrations increase near-bed effective viscosity, which stifles the flow separation needed to initiate and maintain lower-regime bedforms. (e.g., Best, 1992).

<u>Depositional mechanism.</u>. A commonly invoked mechanism to explain the generation of stratified hyperconcentrated flows in subaqueous fan settings is a submerged hydraulic jump (Gorrell and Shaw, 1991). At the Buckingham pit, features commonly taken as evidence of hydraulic jumps (e.g., scours, intraclast rip-ups, rapid transition from upper-to lower flow-regime bedform

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[26] Russell, H.A.J. and Cummings, D.I. 2011. Stop 2-4: Buckingham subaqueous fan and Champlain Sea deposits *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 66-68. Geological Survey of Canada, Open File 6947.

stratification) have not been observed. Consequently, although flows are interpreted to have been hyperconcentrated, the exact mechanism behind the generation of these flows is unclear. The vertical and lateral extent of beds and the bedding style of the diffusely graded sand at this site are noteworthy features because they are uncommon, even in landform elements and aggregate excavations orders of magnitude larger (e.g. in Oak Ridges Moraine). They suggest that depositional conditions at the Buckingham pit location were somewhat different than that inferred for many subaqueous outwash fans in southern and eastern Ontario.

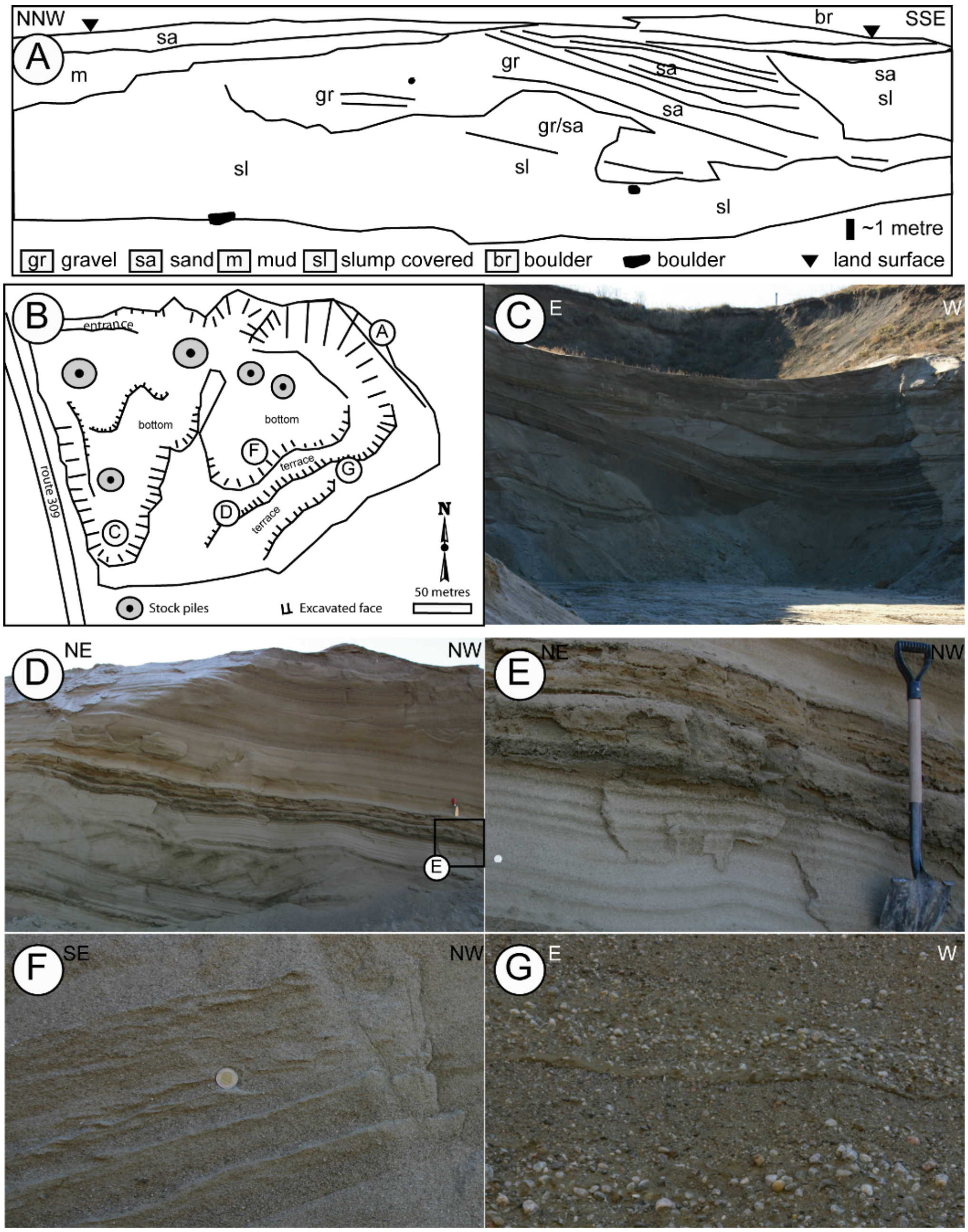

Figure 2-16. Overview of the Buckingham subaqueous fan deposit. A) Line drawing of northeastern pit face illustrating proximal–distal fining of sand and gravel foresets. Note Champlain Sea mud toward upper northeast part of section. B) Sketch map from Google Earth image (April 2010) of pit floor. C) Example of dipping sand strata at base of pit. D) Illustrative sand facies, dune scale cross-strata, laminated mud facies, and diffusely graded sand. E) Close-up of diffusely graded sand and mud facies. F) Faint poorly sorted

climbing ripples in a succession of diffusely graded sand. G) Massive to diffusely graded gravel with sand rich beds. Note gravel clusters, weak imbrication and change in matrix caliber.

Stop 2-5A: The June 23rd, 2010 Val-des-Bois Earthquake

Maurice Lamontagne[27]

On June 23rd, 2010, a moment magnitude (Mw) 5.0 earthquake occurred in the Papineau-Labelle Fauna Reserve, near Val-des-Bois, QC (Fig. 2-17). The earthquake was widely felt in eastern Canada (from Thunder Bay, Ontario, to Nova Scotia) and into the northeast United States, and as far south as Kentucky (Hayek et al., 2011). Many instances of damage were reported within 100 km of the epicentre, including many damaged chimneys in Ottawa at about 55 km distance (Fig. 2-18). Two cases of landslides in Champlain Sea deposits were triggered within 20 km of the epicentre (see Perret, this volume). Near field strong motions exceeded extrapolated values for such moderate magnitude events (Atkinson and Assatourians, 2010). The main event was well recorded by strong and weak motion instruments. A maximum PGA of ~8-10 %g was recorded by strong motion instruments in the Ottawa region, and 0.8% g in Montreal (Hayek et al., 2011).

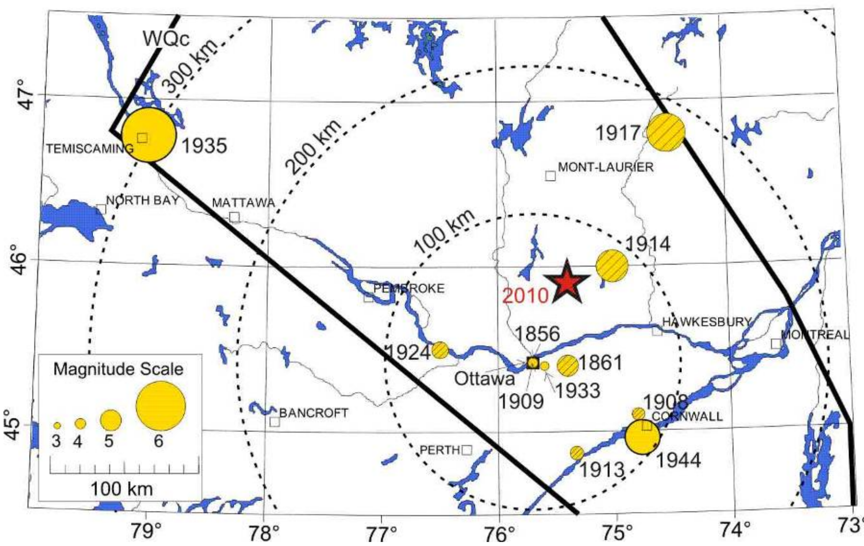

Figure 2-17 Location of the epicentres of the Val-des-Bois earthquake (star) and of historical earthquakes that had caused some damage in the Ottawa-Gatineau region. Shaded circles represent earthquakes with approximate epicentres and magnitudes.

The Val-des-Bois earthquake is located in the NW-SE seismicity band of Western Quebec Seismic Zone (WQSZ) that trends from Montreal to the Reservoir Cabonga (see Introduction). The focal depth of the main shock was determined as 22 km, which is not unusual is eastern Canada, but certainly towards the deepest part of the hypocentre distribution. A predominantly reverse faulting focal mechanism was computed with planes trending northwest-southeast (Hayek et al., 2011) consistent with most other WQSZ earthquakes and other fault reactivation mechanisms in eastern North America.

The aftershock sequence is one of the most active ever recorded in the Canadian Shield for an earthquake of such a moderate magnitude. Between June 23, 2010, and the February 2011, some 25 aftershocks larger or equal to magnitude 2.5 (i.e. that could have been felt) have been

[27] Lamontagne, M. 2011. Stop 2-5A: The June 23rd, 2010 Val-des-Bois earthquake. *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummings, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 69-71. Geological Survey of Canada, Open File 6947.

recorded, with hundreds more of smaller magnitude earthquakes. Altogether, it is more than 100 events that have been detected by the permanent seismograph network, with the largest aftershock reaching a magnitude of 3.3. Field stations that were deployed in the epicentral region within a few hours of the main shock recorded even more events. As typically happens, the rate of earthquake occurrence progressively diminished with time. As mentioned, the number of aftershocks is especially surprising considering the mid-crustal depth of the earthquake. In general, shallower earthquakes are expected to produce more aftershocks compared to events with main shocks at mid-crustal depths.

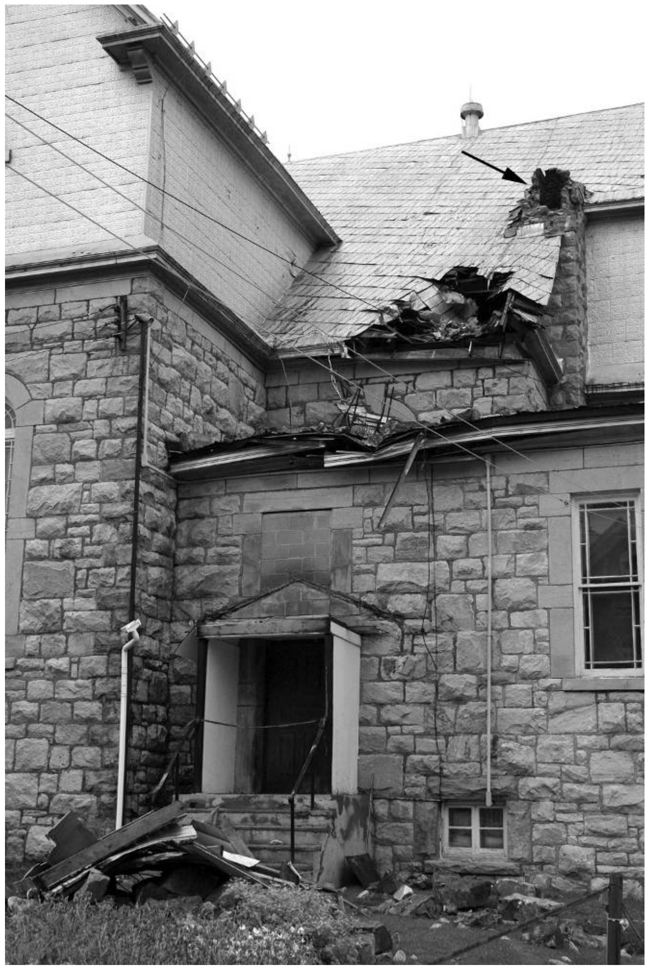

Figure 2-18 Damage to l'Église de la Visitation, Gracefield, Quebec, from the June 23, 2010, earthquake. Arrow marks the location of the chimney which toppled and damaged the roof and an entranceway cover while falling to the ground (Photograph: Greg Brooks).

Felt over a vast area and by millions of people in south-eastern Canada and the north-eastern United States, the earthquake caused considerable public concern. Media interest was high and a large number of radio and TV interviews were given by the GSC personnel. Over 5000 people filled out the "Did-you-feel-it" questionnaire (felt report) on the *EarthquakesCanada.ca* website (see Halchuk, 2010). Since the earthquake occurred during work hours, a very large number of people accessed the *EarthquakesCanada.ca* web site causing it to be unavailable for 1.5 hr after the event. Since then, the server and the procedures have been upgraded to be able to cope with the high level of interest following a future, widely-felt earthquake. However, emergency organizations (Public Safety Canada, power utilities, railroad companies) were notified in a timely fashion according to pre-determined procedures.

The earthquake, however, also exemplified the low-level of earthquake preparedness of the general population in eastern Canada. Many cases of incorrect reactions to the ground vibrations were reported. Examples of these include people running outside during the shaking and people using elevators to evacuate tall buildings. This low-level of earthquake preparedness reflects, at least in part, the decades of time that separate such events which causes public awareness and the perceived need for preparedness to wane.

Stop 2-5B: Binette Road Earth Flow Induced by the June 23, 2010, Val-des-Bois Earthquake

Didier Perret, Rémi Mompin, François Bossé and Denis Demers[28]

The two most important triggering factors of large retrogressive landslides in Eastern Canada lowlands are bank erosion along rivers and human activities like excavating the base of a slope or overloading its crest. These two factors represent 65% and 34% of all cases inventoried since 1840, respectively. Albeit to a much lesser extent, strong earthquakes are also known to have triggered large landslides in Eastern Canada during the Holocene, some of them having a surface area of several square kilometers (e.g. Aylsworth et al., 2000).

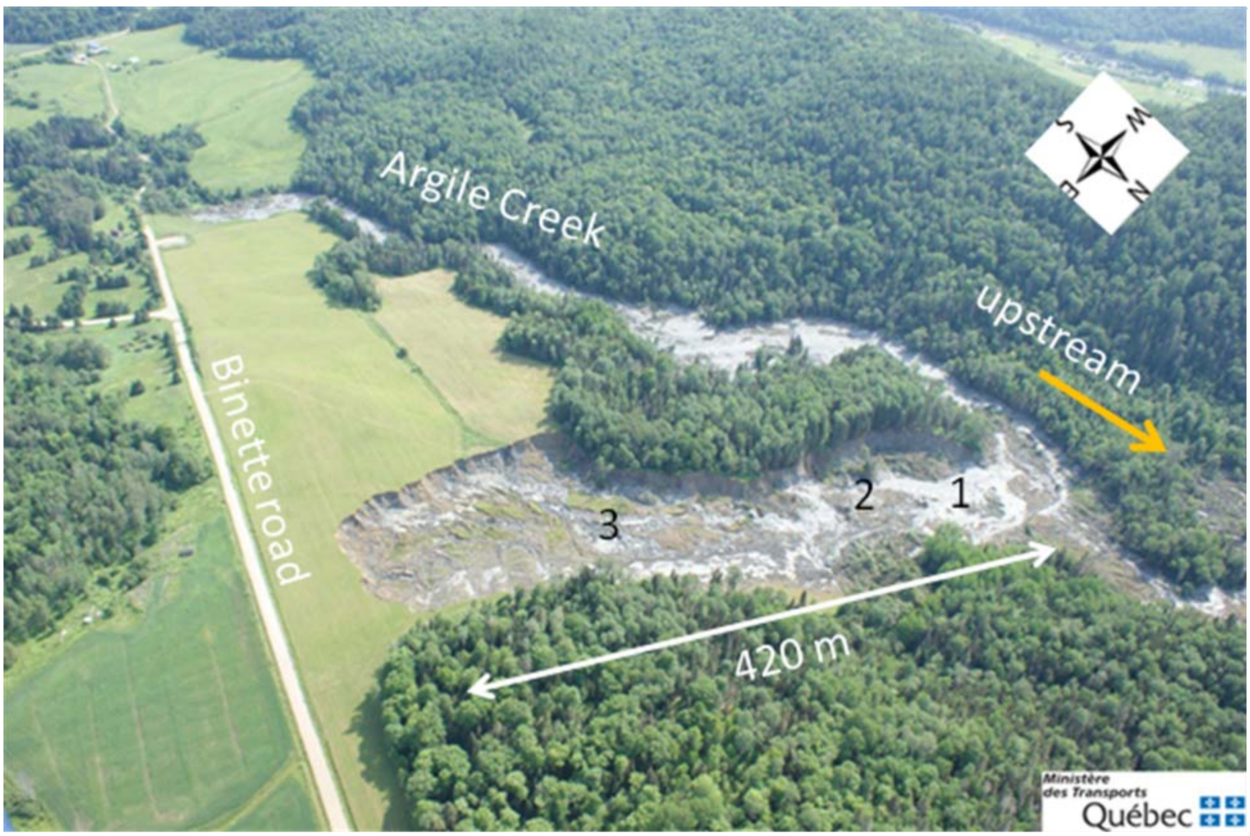

Figure 2-19. Oblique airphoto of the Binette Road earth flow (photo taken on June 26, 2010). Numbers 1 to 3 indicate the approximate location of the three plateaus of the stepped failure surface.

[28] Perret, D., Mompin, R., Bossé, F. and Demers, D. 2011. Stop 2-5B: The Binette road earth flow induced by the June 23rd, 2010 Val-des-Bois earthquake. *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummins, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 72-74. Geological Survey of Canada, Open File 6947.

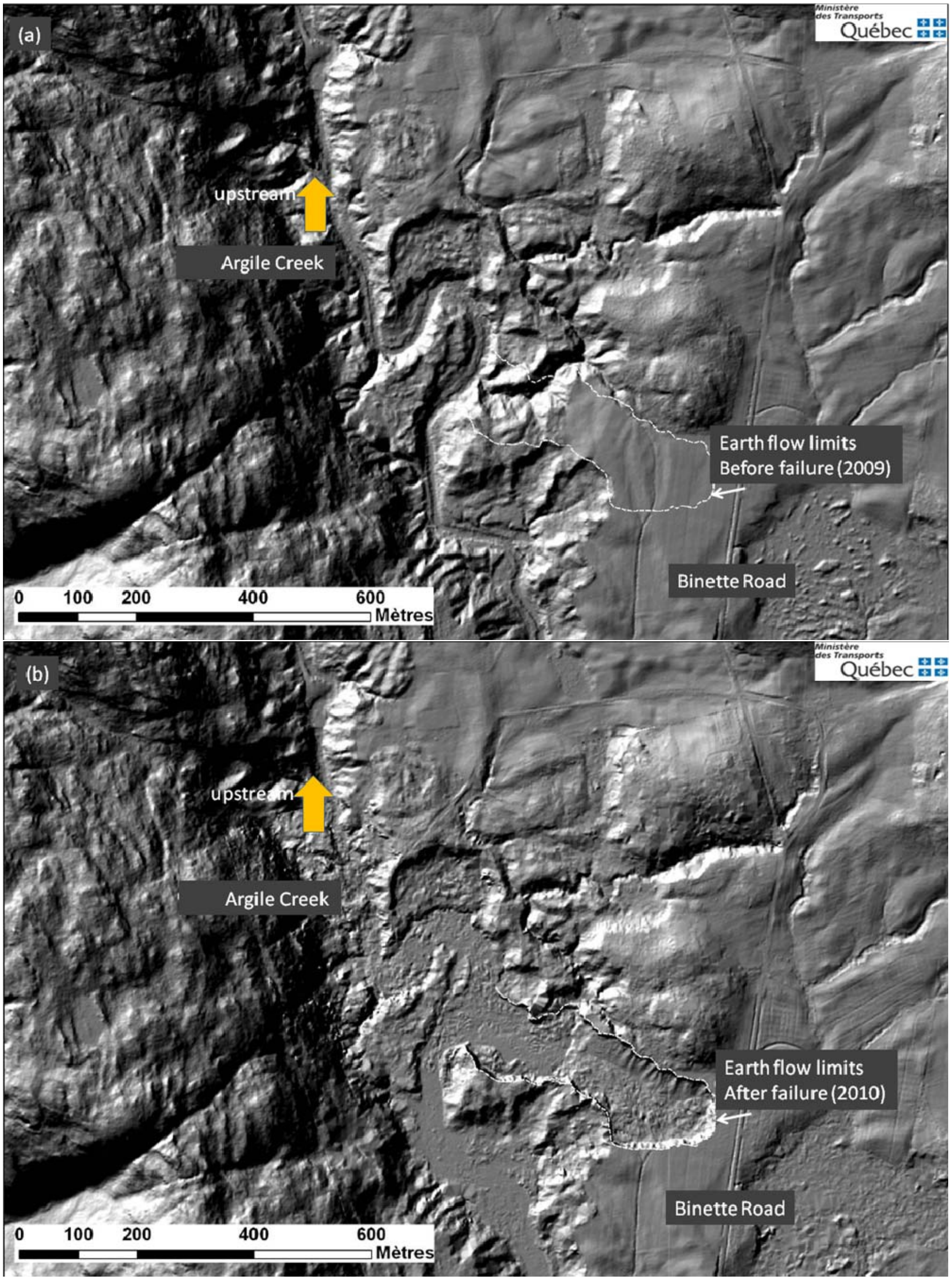

Figure 2-20. LIDAR hillshade of the earth flow area; (a) before failure (2009 flight), and (b) after the landslide occurred (2010 flight).

The Binette Road earth flow (Fig. 2-19) was induced by the moderate Mw5.0 earthquake which shook the Ottawa region on June 23, 2010. This landslide, located at about 14 km south-west of the earthquake epicenter, retrogressed over a distance of 420 m from the Argile (or Clay) Creek (Fig. 2-19), a small tributary of the Lievre River along which a deadly retrogressive landslide occurred in 1908 (see Lawrence and Brooks, this volume). With a width of 150 m, the Binette Road earth flow has a surface area of 5.6 ha and a volume estimated to be 1.1x106 m^3. The landslide shape and lateral extension have been controlled by bedrock outcropping on its north side and by an old earth flow scar on its south side (Fig. 2-20). The elevation of the marine terrace affected by the landslide is 164-167 m. A geotechnical investigation has allowed determining the position of a stepped failure surface, with three plateaus at elevations ranging from 140 m close to the creek to about 152 m near the back scarp (numbered 1 to 3 on Fig. 2-19). The sensitivity of the clay at depths where the failure propagated is larger than 500, the remolded shear strength being lower than 0.07 kPa, the lowest value than can be determined with the Swedish fall cone. This extreme sensitivity explains why debris flowed almost entirely out of the crater (Fig. 2-21). The thickness of the remaining debris inside the scar is about two to three meters only along the central part of the third plateau.

Figure 2-21 Slickenslides observed on the desiccated failure surface exposed near the center of the crater, at the onset of the 3rd plateau, just east of Number 2 on Fig. 1. The pen gives the scale and its grey tip points towards the river (toward viewer). Slickenslides are oriented parallel to the orientation of the pen (photo taken on July 1, 2010).

A reliable eyewitness account indicates that the earth flow did not occur during the earthquake or in the moments that followed but the day after, about 22 hours after the main shock. Several aftershocks have been recorded during this period, the highest magnitude being Mw3.3 which generated a ground motion probably too weak for having a detrimental impact on slope stability in the area. According to a preliminary analysis, the failure mechanism currently proposed is that the main shock triggered a first rotational slide along the bank of the creek or possibly along a thalweg close-by. A delayed retrogressive failure then occurred, as it is commonly observed for earth flows in Eastern Canada.

Stop 2-6: 1908 Notre-Dame-de-la-Salette Earth Flow

Ted Lawrence and Gregory R. Brooks[29]

On April 26 1908, a sensitive clay earth flow occurred on the west bank of the Lièvre River at the village of Notre-Dame-de-la-Salette, Quebec, (Maps 1, 3) that resulted in the death of thirty-three people. This failure caused the greatest loss of life from a landslide in glaciomarine sediment within the Ottawa Valley–St. Lawrence Lowlands region and thus represents an important case study of an eastern Canadian earth flow. A general account of the landslide is presented by Ells (1908), while Lapointe (1974) provides a detailed historical summary of the disaster that is illustrated with interesting historical photographs taken immediately after the event. The historical aspect of this summary is based on these accounts.

The village of Notre-Dame-de-la-Salette is located along the east side of the Lièvre River (Fig. 2-22). In 1908, a portion of the village was situated on a low terrace about four metres above river level. At about 0400 on April 26 1908, an earth flow occurred suddenly on the west river bank opposite the village (Fig. 2-23), while an unbroken layer of ice about 0.5 m thick still covered the river. Three houses and several 'outbuildings' on the west bank were immediately engulfed in the landslide, resulting in the deaths of six people (Fig. 2-24). Upon entering the river channel, which is about 110 m wide at this location, the landslide generated a displacement wave that overwhelmed part of the village on the opposite (eastern) bank. Large blocks of river ice were lifted and carried with the wave, contributing to the destructive effects on structures within the village (Fig. 2-25). Twelve houses and twenty-five 'outbuildings' were destroyed by the wave, including a large, two-storey building, which was located about 12 m from the river bank, housing the post-office, general store, blacksmith's shop, and village inn. Building debris was found up to 15 m above river level. The recession of displaced water swept ice, building debris, and many building occupants into the river.

Twenty seven people were killed in the village, in addition to the six killed on the west side of the river. The wave also travelled upstream and destroyed or damaged buildings on the river bank, including Vieuvenu's Hotel, which was partly wrecked by ice floes, 800 m north of the slide. Spoil from the earth flow impounded the Lièvre River, causing flooding upstream. Several survivors of the earth flow and displacement wave were found stranded on ice floating within the impoundment. The presence of an ice cover on the river was deemed to have contributed greatly to the destructive effects of the displacement wave on the structures located on the east bank and consequently on the large number of people killed in the disaster.

The earth flow retrogressed ~160 m back from the river and created a crater ~460 m wide. The backscarp of the crater is ~20 m high. The volume of the landslide is estimated to be 1.5 x 10^6 m^3. The ratio of scar width to retrogression distance is 0.34. This comparatively low relative retrogression distance suggests that the earth flow was the product of a lateral spread where the liquefaction occurred along a relatively thin layer at the failure plane. The island in the modern river channel adjacent to the earth flow scar is a remnant of the earth flow spoil. Energy head calculations, based on the described run-up of the wave, indicate a minimum wave velocity in the order of 17 m/s.

The specific trigger of the earth flow is not known. The landslide took place at a location where the river channel is comparatively straight and not obviously concentrated against the right river bank (Fig. 2-23). Nevertheless, it is possible that fluvial erosion at the toe of the bank was a contributing factor. Newspaper accounts as well as Ells (1908) report that a crack had developed

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[29] Lawrence, D.E. and Brooks, G.R., 2011. 1908 Stop 2-6: 1908 Notre-Dame-de-la-Salette earth flow. *In*: Deglacial history of the Champlain Sea basin and implications for urbanization. Russell, H.A.J., Brooks, G.R. and Cummins, D.I. (Editors), Joint annual meeting GAC-MAC-SEG-SGA, Ottawa, Ontario, May 25-27, 2011, Field Guide Book, p. 75-77. Geological Survey of Canada, Open File 6947.

several days before the landslide on the ground surface 65 to 100 m back from the river edge. Local geotechnical borehole data reveal the presence of a stiff cap of sediment, 3-4 m thick, overlying softer sediment (J. Aylsworth, unpublished data). Sensitivities of the deposits are variable with depth, with layers of higher sensitivities ranging from 15 to 37.

The 1908 Notre-Dame-de-la-Salette disaster illustrates the impacts that can be caused by the secondary effects of an earth-flow-generated displacement wave, which in this case extended the zone of damage well beyond the immediate limits of the landslide source area and run-out zone.

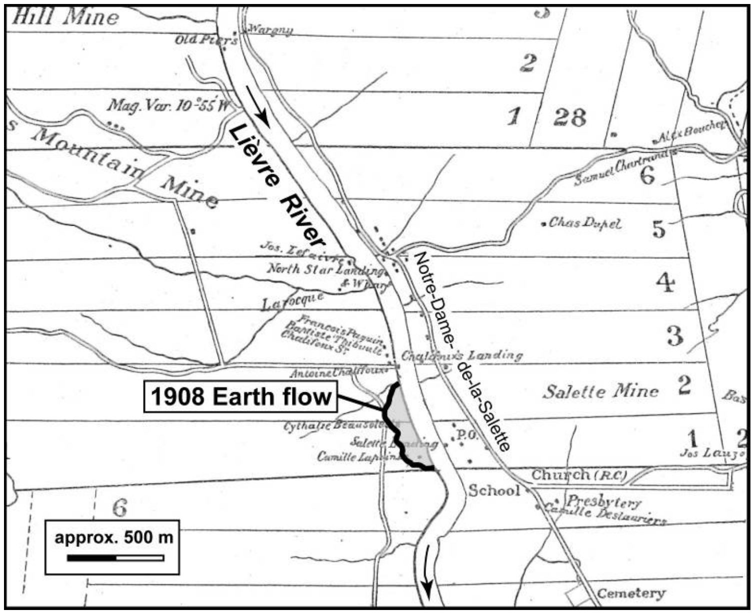

Fig. 2-22. 1908 map showing location of Notre-Dame-de-la-Salette and the earth flow scar (after Ells, 1908).

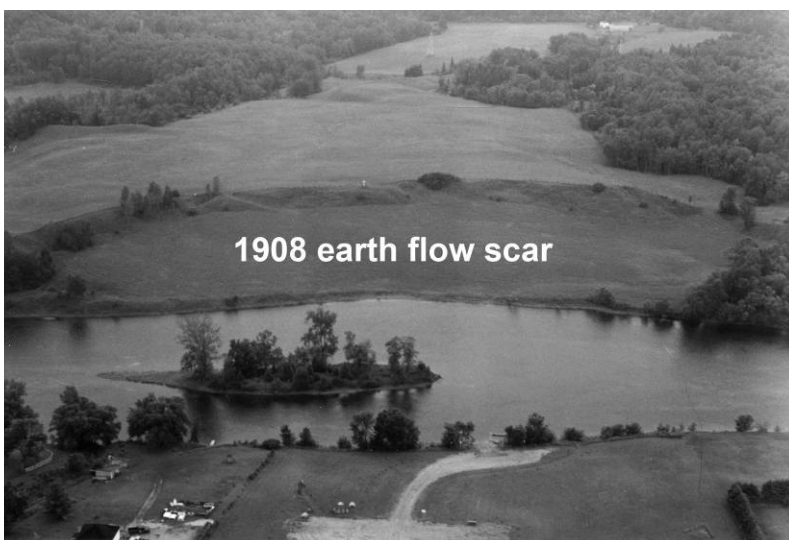

Fig. 2-23. Scar of the 1908 Notre-Dame-de-la-Salette earth flow, viewed from the air in 1996. The island in the river channel is remnant spoil (landslide debris). The river flows from right to left. (Photograph: Ted Lawrence)

Fig. 2-24. Contemporary post-failure view across the Liève River channel into the landslide scar, showing spoil and destroyed buildings within the landslide crater. The river is flowing from right to left. (Photograph: Archives Nationales du Québec)

Fig. 2-25 Damaged and destroyed houses within the village of Notre-Dame-de-la-Salette showing the effects of the landslide-generated displacement wave and ice on wooden structures located on the east bank of the Liève River. Arrow marks the headscarp of the landslide. (Photograph: Archives Nationales du Québec)

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